SUMMARY REPORT
OF THE
SNOW INVESTIGATIONS

SNOW HYDROLOGY

NORTH PACIFIC DIVISION
CORPS OF ENGINEERS, U.S. ARMY
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PREFACE

During the past three decades, increasing demands upon the water resources of this country have resulted in the construction of many projects for the control of our river systems, thus bringing about a need for a better understanding of the natural processes which govern their flow. This knowledge is necessary for both the economical design and efficient operation of engineering works required for river control. In response to this need, the field of hydrology has also experienced a large growth, but most of this increased knowledge has been concerned with the hydrology of rainfall. Methods of computing snowmelt and snowmelt runoff have been largely based on empirical relations derived from very limited data. To promote a more fundamental understanding of snow hydrology for project design and streamflow forecasting, particularly in the western part of the United States, the Corps of Engineers and the U. S. Weather Bureau initiated the Cooperative Snow Investigations. Activation and sustaining support of the program resulted primarily from the efforts of Mr. G. A. Hathaway of the Corps of Engineers and Mr. Merrill Bernard of the U. S. Weather Bureau (deceased). Following the cooperative phase of the investigations, the Corps of Engineers continued the work.

The snow investigation program was organized to meet specific technical objectives in the field of snow hydrology for both agencies. In order to meet these objectives, fundamental research in the physics of snow was needed. An extensive laboratory program was established, and observations were gathered over a period of several years at three headwater locations, having differing conditions of climate and physical environment. Data obtained from the laboratories have been processed and published. Analysis of these data forms the basis for the basic relationships and methods of application derived for the solution of snow hydrology problems. These in turn have been utilized by the Corps of Engineers in specific applications to project design or operation. These applications include: (1) the derivation of maximum probable and standard project floods, which partly form the basis of project design; (2) the development of procedures for forecasting seasonal runoff, which are used primarily in connection with regulation of multiple-purpose reservoirs and appraisal of flood potential; and (3) the formulation of procedures for hydrograph synthesis of snowmelt or rain-on-snow events, which are used as the basis for forecasting streamflow at reservoir projects and river control works, and for flood fight operations.

Some specific developments in snow hydrology which have resulted from the work of the snow investigations are: (1) experimental evaluation of the coefficients of snowmelt, in terms of appropriate meteorological parameters, for each of the several processes of heat transfer to the snowpack; (2) methods of applying thermal-budget indexes of snowmelt to drainage basins; (3) derivation of general snowmelt equations which are applicable to drainage basins according to their physical characteristics; (4) determination of the reliability of snow courses and precipitation gages, as related to their site characteristics; (5) evaluation of each component of the hydrologic balance in areas of snow accumulation, and application of the water-balance technique to procedures for forecasting seasonal runoff volumes; (6) experimental determination of the liquid-water-holding capacity of the snowpack and transmission of heat and water through the snowpack, with methods of application of results to basin hydrologic studies; (7) methods of synthesizing streamflow hydrographs for areas involving snow; (8) investigation of the general features of atmospheric circulation as it affects moisture and energy input to drainage basins, and the use of upper air data in estimating snowmelt; (9) derivation of an index procedure for forecasting spring-season snowmelt runoff by use of low-elevation winter runoff, without recourse to direct measurements of precipitation or snow accumulation. Under the Civil Works Investigations of the Corps of Engineers, work on two projects was accomplished in conjunction with the snow investigations program. Under Project CW-170, a radioisotope-radiotelemetering snow gage was developed which transmits daily readings of snowpack water equivalent by high-frequency radio from remote gage sites to a base receiving station. Under Project CW-171, a training program for engineers of the Corps was organized, whereby methods being developed within the investigations could be put to use prior to the completion of formal research papers and a general summary of the investigations. Also under project CW-171, certain features of the investigations were developed, including an electronic storage routing analog which is applicable to general hydrologic use. The results of the individual investigations within the snow program have been reported from time to time in the various technical publications of the program.

This report, which summarizes the work of the Snow Investigations, is intended as a reference on snow hydrology. Although the information in the report was developed mainly from studies of mountainous areas in the western United States, the basic relationships derived are applicable to all regions in which snowfall is of appreciable hydrologic concern. The information is intentionally presented in considerable detail, in order that the practicing hydrologist who has need for it may thoroughly understand the fundamental relationships involved and the derivation of the methods given. Accordingly, the report not only includes technical background material necessary to a general understanding of the subject matter, together with methods and examples of application, but also includes some material not essential to application itself. Also, there is some duplication of material among chapters to provide completeness of presentation for individual subjects. Use of the report as a simple handbook or manual of procedure is not intended, and little attempt has been made toward the condensation and generalization that characterize works of that kind. Work on the report has been accomplished under the general supervision of personnel of the office of the North Pacific Division, Corps of Engineers, U.S. Army, including Mr. F. S. Brown, Head, Engineering Division; Mr. Mark L. Nelson, Head, Water Control Branch; and Mr. Oliver Johnson, Head, Hydraulics and Hydrology Section.

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CHAPTER 1 - INTRODUCTION

1-01. AUTHORITY

1-01.01 This report was authorized by the 1954 annual conference of the Snow Investigations of the Corps of Engineers, U. S. Army, held in offices of the North Pacific Division on 13-14 May 1954. Pertinent sections of paragraph 4 of the notes from that conference, as revised, are quoted for information: "The principal feature of the proposed plan of future work was the completion within the next two years of a publication which would summarize all present knowledge of the Snow Investigations Unit. It was decided that an editorial committee should be formed to prepare an outline and specifications for the report. F. F. Snyder, W. L. D. Bottorf, and D. M. Rockwood (Chairman) were designated as members of that committee. After completion of the outline and specifications, it should be circulated among the participating offices for comment..."

1-01.02 In accordance with the above-quoted directive, an outline of material to be presented in the report was prepared during the summer of 1954 and submitted to the editorial committee for review. The outline was reviewed in detail at the periodic conference, Snow Investigations, held on 4 November 1954 at the North Pacific Division office. Suggested revisions were incorporated in a revised outline, which has been closely followed in this report.

1-02. PURPOSE AND SCOPE OF REPORT

1-02.01 This report is designed to be a reference volume for hydrologists. The Cooperative Snow Investigations, since the time of its organization, has dealt with the analysis of individual snow hydrology problems. Their solutions, when combined and integrated, form the technical background for the report as a whole. It is the intent, therefore, to present pertinent results of investigations accomplished to date and to indicate how they apply in solution of snow hydrology problems that arise in work of the Corps of Engineers. Specific purposes are as follows:

a. To make available to hydrologists a comprehensive report which summarizes all present knowledge of the Snow Investigations Unit with regard to (1) deposition and distribution of the snowpack and the depletion of snow cover, (2) evaluation of the hydrologic water balance of basins where appreciable snow occurs, (3) the physics of snowmelt, (4) the storage and transit of liquid water in the snowpack, and (5) possible methods for estimating rates of streamflow and volumes of runoff in basins where snow affects those quantities.

b. To point out gaps in existing knowledge of snow hydrology and suggest ways of filling them.

1-02.02 It is the intent that the report be suited primarily to an engineering approach to snow-hydrology problems, in order that solutions may be developed for project basins from data commonly available. However, a reasonably sound theoretical background for the hydrologist is believed essential, in order that he may formulate short cuts and approximations without significant departure from fundamental laws. Accordingly, the report deals with both theory and applications, but emphasis is placed on presentation of relationships which may be used by hydrologists in their practical work.

1-02.03 The report embraces all pertinent analyses from prior publications of the unit, as well as its unpublished analyses and certain pertinent work done by other investigators. In general, the extraction of materials is fairly brief, but is in sufficient detail that hydrologists without previous experience in the analysis of snow problems can gain a working knowledge of the field. It is assumed that the reader has access to a set of the previously published reports; accordingly, most basic data and certain detailed analyses contained therein, though considered useful supplementary material, are referred to without the necessity of repetition in this volume.

1-03. INITIATION, OBJECTIVES, AND ACTIVITIES OF THE SNOW INVESTIGATIONS

1-03.01 General. - For several years prior to 1945, problems encountered by the Corps of Engineers in determining spillway design floods and by the Weather Bureau in meeting its responsibilities for streamflow forecasting gave rise to discussions relative to initiating an integrated investigational program in snow hydrology. As a result, in 1945 the Corps of Engineers and the Weather Bureau formulated a joint research program, organized as the Cooperative Snow Investigations and pointed initially toward solution of hydrologic problems pertinent to mountain regions of western United States. Over-all administration was placed with the Division Engineer, South Pacific Division (then Pacific Division), Corps of Engineers, at San Francisco, California, where the office of the Program Director, which included a Processing and Analysis Unit and a Technical Supervisor, was established. Assistance to the program was provided to a limited degree by three other Federal agencies, the Geological Survey, Bureau of Reclamation, and Forest Service. The Snow, Ice, and Permafrost Research Establishment of the Corps of Engineers provided assistance to the program for the years 1950 through 1953, particularly with regard to field observations made at the Central Sierra Snow Laboratory. Occasional assistance was provided by other organizations, such as state and local agencies and private research organizations. Direct participation in the program by the Weather Bureau ended in 1952. Thereafter, the research was continued under the name "Snow Investigations" by the Corps, but the basic program objectives remained unchanged. In June 1953 the administration of the program was transferred from the South Pacific Division to the North Pacific Division office of the Corps of Engineers, and program personnel and records were moved to Portland, Oregon. Reference is made to the Progress Reports (see Appendix I) for detailed descriptions of year-to-year progress of the investigations.

1-03.02 Objectives. - The direction of the work has been focused according to the broad objectives of the program, which were set forth initially as follows:

a. Determination of a practicable and reliable method of evaluating the maximum streamflow which may be produced by a watershed as the result of snowmelt or combined snowmelt and rain.

b. Development of a practicable and reliable method of forecasting seasonal and short-term streamflow, including floods, resulting from snowmelt or combined snowmelt and rain.

c. Expansion of basic knowledge of hydrodynamic and thermodynamic characteristics of snow through a program of fundamental scientific research.

d. Advancement of knowledge of meteorological, climatological, and hydrological phenomena as they influence the above three objectives.

These broad objectives have remained unchanged throughout the duration of the program. However, emphasis on various phases of the work has shifted from time to time. Initially, the emphasis was upon processing and compilation of the basic data from the snow laboratories. During the intermediate period, considerable time was spent on development of fundamental scientific research. More recently, work on application of methods to indicate hydrologic conditions on snow laboratory basins and project basins has received a proportionally larger share of the effort of the unit.

1-03.03 Field operations. - The field operation phase of the program consisted primarily of the operation of three snow laboratories with different environments in the mountains of western United States. Chapter 2 describes in detail the laboratories, the observations made, and the pertinent data gathered and published. The laboratories were operated for periods ranging from 5 to 8 years each, and records were generally concurrent. The purpose of the operations was to determine and measure the physical factors affecting snow hydrology, and also to evaluate variations of certain of these quantities over the laboratory drainage basins, which consisted of relatively small areas, ranging from 4 to 21 square miles. The laboratories also were used for the investigation of special techniques for evaluating and reporting snow conditions from remote mountain areas, and were designed to serve as pilot areas whereby methods of basin application of hydrologic principles could be tested, with sufficient instrumentation to assure reasonable delineation of the variation in basin amounts from point measurements. The operation of each of the laboratories required the provision of living facilities for 5 to 10 men and facilities for maintenance of equipment and instruments. Each laboratory was located in a headwaters area where rugged terrain and severe climate necessitated a large expenditure of effort to meet minimum living requirements and perform periodic visits to the instruments. Over-snow vehicles were provided and were used when feasible, but much of the travel was done on foot. All records, either in the form of automatic recorder charts or spot observations of hydrometeorological elements, were sent directly to the Processing and Analysis Unit of the Cooperative Snow Investigations.

1-03.04 Data processing. - The processing and publishing of records from the laboratories also constituted a major task. In all, there were some 1,500,000 observations published in final form in hydrometeorological logs. This mass of data required the preparation of a manual to insure standard methods of reducing original recorder charts to usable data. At the time of maximum output in the processing unit, 30 individuals were working on the several phases of data reduction. Approximately one-third of the data was entered on individual punch cards for aiding in compilation of tabular data and for expediting analytical studies. All original records were microfilmed. The hydrometeorological logs for each laboratory were published by water years; each log averaged approximately 200 pages. In addition to data tabulations, the logs included brief descriptions of instrumentation, site characteristics, and a basin map. The hydrometeorological logs were given general distribution to cooperating agencies and scientific organizations both in this country and abroad.

1-03.05 Analytical work. - The analytical phase of the Cooperative Snow Investigations actually commenced with the inauguration of the program. Technical Report No. 6-1, dated 10 March 1947, consists of a classified outline of the analytical program. The outline is in five major sections and includes a list of over 200 analytical projects which were proposed on the basis of examination of the literature, consultation, and logical classification of then-known requirements in order to meet program objectives. The analytical program outline was intended as a comprehensive listing of fields of research in problems related to snow hydrology, and served as a guide from which priorities of studies could be established. During the period from 1947 to 1950, many of the analytical projects were investigated in an exploratory manner, but few reports were issued on analytical work. During this time, much of the effort was required for orientation of personnel in snow-hydrology problems, because the field was, for the most part, unexplored. Also, due to the fact that laboratory data reduction was not complete but was proceeding concurrently with the analytical work, much of time was spent in the processing of field data. While many of the studies performed during this time were not fruitful in producing usable relationships, the work was of value in suggesting future lines of approach, and expansion or modification of the laboratory program was made as investigations progressed, to provide bases for improvement of instruments and observational techniques and schedules. Beginning in 1950 the analytical work was directed primarily to problems related to development of methods and relevant criteria for determining maximum-type floods involving snowmelt. In general, these methods are also applicable to problems involved in project operation. The results of these investigations have been published in the form of Research Notes and Project CW-171 Technical Bulletins which have been given limited distribution.

1-03.06 Organization and administration of the Cooperative Snow Investigations. - Formulation and general direction of the program was provided through policy-making conferences--usually held annually, but occasionally more frequently for special requirements--which were attended by participating-office representatives at the several levels of responsibility for both the Corps of Engineers and the Weather Bureau. At these conferences, the work was reviewed, and over-all planning of activities, both technical and administrative, was developed. Specific direction of the program as a whole rested with the Program Director, a Corps of Engineers employee, to carry out the plans formulated by the conference. The Program Director supervised the operation of the field laboratories as well as the Processing and Analysis Unit. Until 1950, a Technical Supervisor was responsible for direction of the technical phases of laboratory operation, conducting of special experiments in the field, and development of instrumentation. Plate 1-1 is an organization chart which shows the channels of authority that existed between the various Weather Bureau and Corps of Engineers offices, and the Cooperative Snow Investigations activities. This organization applied essentially from 1947 through 1950 fiscal years. In 1950 the program was reorganized, and field work was curtailed at the laboratories. Under the reorganization, the Weather Bureau and the Corps of Engineers established analysis units under their respective organizations to pursue problems in their fields of responsibility, but coordination was continued. Special emphasis was directed toward processing and publishing laboratory data in final form. Although after 1950 the large-scale operation of the laboratories was discontinued, an intensive program of special observations was carried on directly by groups of analysts temporarily detailed from the unit to obtain information essential to the analytical work. Key administrative and technical supervisors during the entire period are listed below:

Program Director

W. F. Bingham, Corps of Engineers,	1944-1945
W. C. Cassidy, Corps of Engineers,	1946
F. L. Rhodes, Corps of Engineers	1947-1950
W. L. D. Bottorf, Corps of Engineers,	1950-1953
D. M. Rockwood, Corps of Engineers,	1953-1956

Technical Supervisor

R. W. Gerdel, Weather Bureau, 1946-1950

Processing and Analysis Unit

W. T. Wilson, Weather Bureau, (Chief) 1946-1950
D. H. Miller, Corps of Engineers, (Asst.) 1946-1950

Head, Corps of Engineers Civil Works Investigation Project 171

D. W. Hullinghorst 1949-1951
C. E. Hildebrand 1951-1956

Supervisor of Special Field Observations

P. B. Boyer 1951-1954

Direction of the individual laboratories is described in chapter 2 of this report. Cooperative Snow Investigations personnel (exclusive of laboratory personnel), whose tenure was in excess of one year, are listed in table 1-1.

1-03.07 Coordination with other agencies and research organizations. - Snow research is carried on by various agencies and research groups, and results of studies have been published in technical journals and reports, both in the United States and in foreign countries. The Cooperative Snow Investigations has maintained contact with those organizations with which it is familiar, either directly or through the literature. This coordination falls into three categories, as follows: (1) direct cooperation by Federal agencies or special research groups with the Cooperative Snow Investigations; (2) participation of members of the Snow Investigations in the activities of technical societies whose objectives are directed toward snow or hydrologic research; and (3) informal discussions with individuals or representatives of organizations for exchange of ideas on snow hydrology.

1-03.08 Direct cooperation from Federal agencies began with the initiation of the program. Shortly after formulation of the cooperative program by the Corps of Engineers and the Weather Bureau, the Bureau of Reclamation and the Geological Survey each provided the services of one or two employees for a period of from two to four years. The Forest Service contributed primarily to the laboratory phase of the work, and cooperated on a reimbursable basis in a number of aerial flights for obtaining photographs of snow-covered areas. The Forest Service also supplied information on mountain soils and forest effects on the snowpack. The U. S. Air Force made aerial photographs and provided access to weather data and charts. The Snow, Ice and Permafrost Research Establishment maintained close liaison with the unit in operation of the Central Sierra Snow Laboratory, exchange of data, and consultation on analytical work. A contract was negotiated with the University of California for the purchase of two Gier-Dunkle radiometers for measuring heat transfer to the snowpack by radiation. Members of the university faculty were most helpful in providing technical advice on application of these instruments to snowpack work. Other phases of cooperative effort at the Central Sierra Snow Laboratory were with the U. S. Navy Electronics Laboratory for testing an automatic weather station and with the Landing Aids Experiment Station for studying the performance of lighting equipment to be used for aircraft landings. In connection with the development of the radioisotope-radiotelemetering snow gage, a contract for the radio reporting equipment was negotiated with the Motorola Corporating which gave considerable aid in the radio transmission phase of development of the gage. A more complete discussion of cooperation in the field investigations is contained in chapter 2.

1-03.09 Members of the Cooperative Snow Investigations have, from time to time, participated in technical society functions, either by submitting technical papers or providing formal or informal discussions of work by others. These societies include: (1) American Geophysical Union, (2) American Meteorological Society, (3) American Society of Civil Engineers, (4) International Union of Geodesy and Geophysics, and (5) Western Snow Conference. Many of the staff were members of one or more of the above organizations. In addition to the above, there were miscellaneous discussion groups and conferences through which staff members have contributed information on various activities of the program. These include interagency committees and work groups, educational organizations, and special groups dealing with snow survey work.

1-03.10 Occasionally, representatives of private engineering organizations, commercial enterprises, universities, and hydrologic research groups contacted the unit informally or by letter and sought information on application of some phase of the work to specific problems. The Investigations Unit always attempted to provide the required information, within the limitations of available time.

1-03.11 Bibliographic material. - While the Cooperative Snow Investigations made no special attempt to review or list all published work on snow hydrology, a brief bibliography (Technical Report 13) was prepared, listing pertinent reference material available at the time of publication (1950). Abstracts of work done in this and related fields were reviewed periodically, and copies of papers of particular interest were obtained for review. The Transactions of the American Geophysical Union, Proceedings of the American Society of Civil Engineers, the various publications of the American Meteorological Society, and the Journal of Glaciology were the principal technical society publications utilized by the unit, but other scientific publications, including periodicals from a few foreign societies, were scanned for work done by others in the field of snow hydrology. The annotated Bibliography of Snow, Ice and Permafrost (SIPRE Report 12), published by the Snow, Ice and Permafrost Research Establishment has been particularly helpful for reviewing the literature in the field of snow. The meteorological abstracts published by the American Meteorological Society similarly provided a convenient and comprehensive listing of meteorological literature.

1-04. THE PROBLEM OF SNOW HYDROLOGY

1-04.01 General. - The field of hydrology is concerned entirely with the evaluation of the various components of the hydrologic cycle. Quantitative evaluation of factors affecting each of these components requires a knowledge of certain phases of the sciences of meteorology, hydraulics, thermodynamics, geology, soil mechanics, and plant physiology.

1-04.02 Knowledge of meteorological factors is of primary importance in snow hydrology. In evaluating moisture inflow from the atmosphere, it is necessary to understand the mechanism of precipitation, the characteristics of airmasses and fronts, and general atmospheric circulation patterns. Meteorology is relied upon particularly in physical study of snowmelt, where energy exchange, both with the atmosphere and with the sun, as well as effects of the atmosphere on radiation exchange, must be generally understood. Fundamental knowledge of meteorology, therefore, is essential to understanding the problem of snow hydrology as a whole, since the atmosphere constitutes the source of moisture supply and also regulates the energy exchange within a basin, which in turn governs snowmelt rates.

1-04.03 Just as the atmosphere is a regulating device for the production of water available for runoff, so the snowpack, soil, underlying geologic formations, and forest cover act to retard runoff and affect the water balance of an area. Thus other basic sciences must be utilized to evaluate the effects of all influencing factors throughout the hydrologic cycle. Particular attention must be given to: (1) the hydraulics of flow, both in open channels and through various media, and the time rate of change of flow as described by routing procedures; (2) the effect of geological formations upon ground-water storage and ground-water flow; (3) the capacity of soils to transmit and store water, as well as to transfer heat to the snowpack; and (4) the ability of plants and forests to affect the deposition of snow, to transpire water to the atmosphere, and to influence energy transfer between the atmosphere and the snowpack.

1-04.04 Point relationships. - In applying the basic sciences in snow hydrology, it is necessary first to determine each component of the hydrologic cycle individually at a point, under stated conditions of environment. Such a determination usually requires the derivation of the proper mathematical relationships of the variables affecting each component, and the establishment of constants of proportionality in the form of coefficients. The relative magnitudes of the components must be taken into account, and emphasis must be placed on the more important functions. In this regard, the relationships of snow hydrology may be expressed by rational mathematical procedures.

1-04.05 Areal relationships. - Having determined point values, the next step consists of utilizing these values to determine amounts and distribution over a basin area, with respect to environmental differences, as well as to time. This involves procedures which are much less exact and less rigorous than for point evaluations. It is impractical to attempt a point-by-point analysis in basin application; rather, it is necessary to deal with basin averages in major subdivisions of geography or environment, and also to deal with averages in time. This concept leads to the use of indexes to represent basin averages from individual point measurements. The application of indexes requires the intelligent use of knowledge gained in studies of evaluation of conditions at a point, and consideration of the physical character of the area involved. Since the use of indexes involves the theory of sampling and errors in measurement, statistical procedures may be employed to evaluate the reliability of estimates and also the weightings of individual factors. The random selection of indexes, without particular regard to representation of basin amounts by point measurement in attempting to improve the fit of historical data should be avoided. The relative reliability of indexes may be established, however, through proper utilization of statistical methods, if the factors being evaluated are selected in accordance with the theoretical considerations. This permits use of indirect methods of evaluation of basin amounts from observational data commonly available. The deficiencies of those methods should be recognized, and when the available data are inadequate to represent basin conditions, steps should be taken to obtain more adequate data.

1-04.06 Applications. - Insofar as the Corps of Engineers is concerned, the application of snow hydrology principles is confined to project design and project operation. In each case the problems involved are grouped into three main categories, as follows: (1) the evaluation of water stored in the snowpack, and its relation to the hydrologic balance; (2) evaluation of rates of melt, the physical causes of snowmelt, and methods of application to basin hydrology; and (3) effect of the snowpack on runoff, both from snowmelt and rain on snow. For project design, fixed sequences of meteorologic and hydrologic conditions are usually chosen, the choice depending primarily on the over-all functional requirements of the project. Project operation, on the other hand, requires evaluation of conditions at a specific time and forecasts of streamflow for both long and short periods. The procedures developed must be flexible, to facilitate adjustment for changing weather; the effect of possible meteorological events subsequent to the date of the forecast must be determined and separated from effects of preceding conditions. Also, the limitations imposed by inaccuracies in weather and streamflow forecasting must be taken into account in the derivation of reservoir regulation schedules and long range volumetric streamflow forecasts.

1-05. PRIOR REPORTS BY COOPERATIVE SNOW INVESTIGATIONS

1-05.01 Technical publications of the unit have been issued in the form of Technical Reports, Research Notes, Technical Bulletins for Project CW-171 and Miscellaneous Reports. A listing of all published reports is contained in Appendix I. Technical Reports consist of the hydrometeorological logs for the laboratories, annual progress reports of the Cooperative Snow Investigations, reports on subjects dealing with snow characteristics, brief bibliographies, and terrain characteristics of Central Sierra Snow Laboratory. Research Notes contain reports of studies performed by the analytical unit and deal with a variety of problems; preliminary in nature, they were designed to disseminate technical findings and procedures quickly to participating offices for comment and use. Technical Bulletins for Project CW-171 are informal presentations of analytical work completed under the project (see section 1-06). The subject matter for Technical Bulletins is principally basinwide in application, but there are, in addition, separate studies for the analysis of individual factors at a point. Included in Appendix I under the title (Miscellaneous Reports) are CSSL micrometeorological studies made by researchers of the SIPRE Analytical Unit of the Cooperative Snow Investigations during 1952-53, and three other reports of interest, not covered under the previous classifications.

1-05.02 Information as to availability of prior reports issued by the Cooperative Snow Investigations will be furnished upon request to the Division Engineer, North Pacific Division, U. S. Corps of Engineers, 210 Custom House, Portland 9, Oregon.

1-06. WORK DONE UNDER PROJECT CW-171

1-06.01 The Corps of Engineers Civil Works Investigations Project 171, entitled, "Criteria for Estimating Runoff from Snow Melt," (referred to herein as CW-171) was initiated in 1949 to aid in the rapid prosecution of analytical work to be used in connection with the design and operation of Corps of Engineers projects. The primary purpose of the project was to accelerate the development of criteria needed in estimating standard project floods, spillway design floods and similar engineering determinations, and to make it possible for engineers from Division and District offices to contribute to the overall program while increasing their knowledge of the snowmelt problem. It was recognized that much of the work was within the objectives of the Cooperative Snow Investigations and would ultimately be accomplished under that program. The project was established, however, to disseminate quickly the results of preliminary investigations, prior to their subsequent publishing in more complete form. CW-171 has, as one of its further objectives, a program for training Division and District office hydrology personnel in the procedures developed in the Snow Investigations. The project also served for training the unit personnel in the practical application of these procedures to hydrology problems encountered by the participating offices. Project Bulletin No. I, dated 8 July 1949, describes in detail the objectives and administrative regulations of the project. The work reported on in the 18 Technical Bulletins of CW-171 is summarized in later chapters of this report. Since the inception of CW-171, there have been over 50 visits of individuals from the participating offices. Pertinent information on each of the visits is listed in Appendix II.

1-06.02 Methods developed under CW-171 and the over-all program have been used wholly or in part in the derivation of design floods for the several Corps of Engineers projects, including a standard project flood for the Cougar Project, and spillway design floods in the Libby, Painted Rock, Pine Canyon, Mathews Canyon, Success, and Terminus projects. Problems in connection with development of operational procedures for Corps of Engineers projects have received considerable attention under CW-171. These problems include derivation of forecasting procedures for streamflow from snowmelt, both on a long-term volumetric basis and on a short-term, day-to-day rate basis. Some work has been done in developing operating-rule curves for projects where snow is a factor. Assistance was also rendered in the derivation of an interim operating procedure for Pine Flat Reservoir and the development of procedures for forecasting seasonal runoff volume for Lookout Point and Detroit Reservoirs by the water-balance method.

1-07. ORGANIZATION OF REPORT

1-07.01 In general, the material which follows in this report is presented in order of development of the subject of snow hydrology, namely: (1) review of the basic data available from snow laboratory operation; (2) evaluation of the water stored in the snowpack, and its relation to the hydrologic balance; (3) theoretical considerations of factors affecting snowmelt; (4) methods of application of snowmelt indexes to estimate basin melt; (5) description of snow-cover depletion and approximate relationships between snow cover depletion and ablation of the snowpack; (6) determination of factors affecting storage of liquid water in the snowpack and runoff therefrom; (7) application of techniques for evaluating snowmelt, rainfall, snow cover, and effect of the snowpack on runoff, and reconstitution of streamflow hydrographs; (8) application of techniques to design flood determinations; (9) application of the relationship of snow water and the hydrologic balance to seasonal runoff forecasting; and (10) application of procedures to reservoir regulation.

1-07.02 Reference material. - Since there is frequent reference to the past reports of the Cooperative Snow Investigations, they are treated separately from other source material in this report and are referenced directly in the text by title or number and listed in Appendix I. All reference material other than that published under the Cooperative Snow Investigations or Corps of Engineers Snow Investigations is listed at the end of each chapter under the appropriate chapter heading.

TABLE 1-1

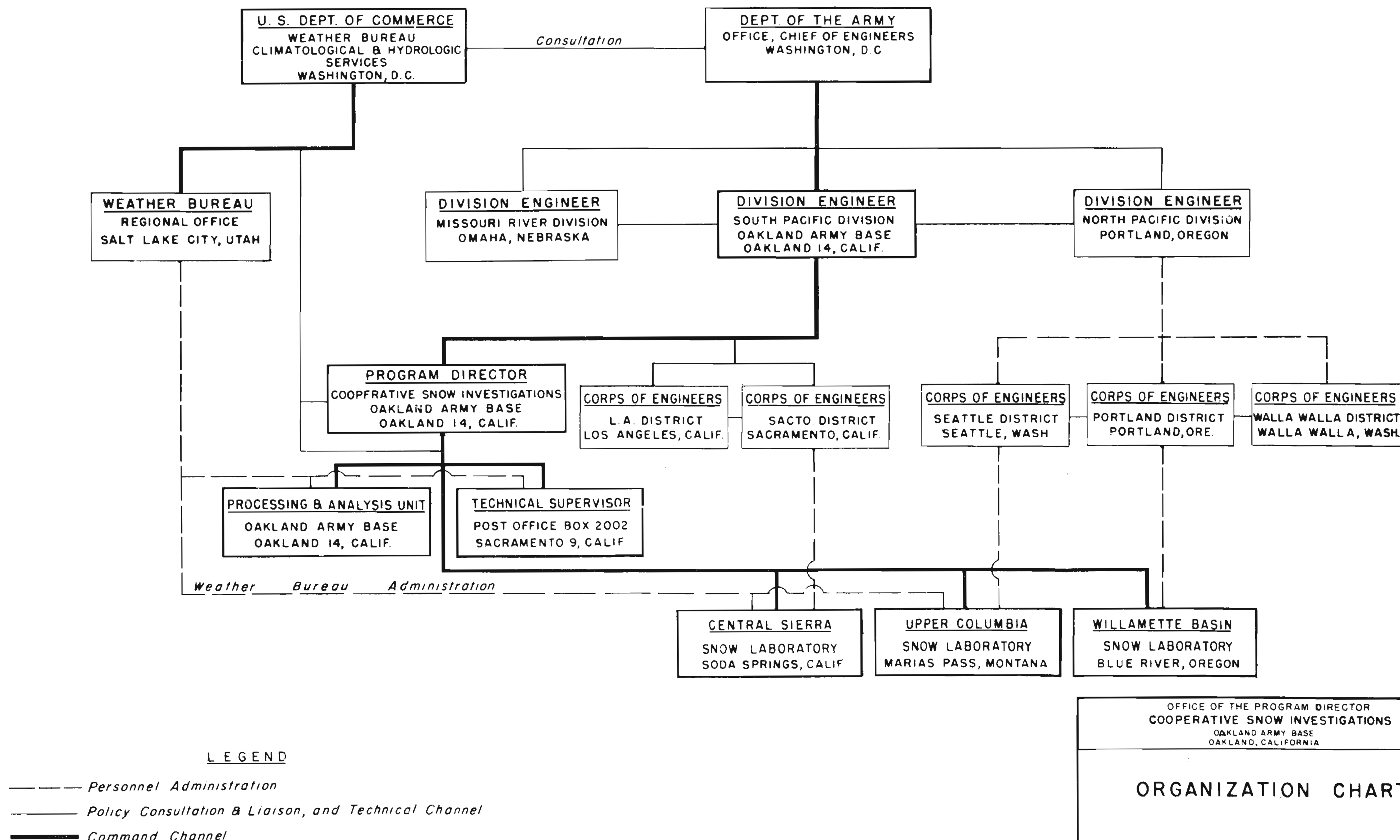
COOPERATIVE SNOW INVESTIGATIONS EMPLOYEES EXCLUSIVE OF LABORATORY PERSONNEL 1/ 2/

Employee	Entered Duty	Separated	Employee	Entered Duty	Separated
Allison, I. D.	1953	1956	Mark, E.	1949	1951
Arnold, B. A.	1951	1953	McClain, M. H.	1951	1956
Bayuk, M.	1948	1951	Merrill, P. +	1954	1956
Berger, P.	1951	1953	Miller, D. H.	1946	1953
Bertle, F. (USBR)	1946	1949	Miller, S.	1953	1956
Bingham, W. F.	1944	1945	Mixsell, J. W.	1949	1953
Bottorf, W. L. D.	1950	1953	Mondrillo, G.	1948	1956
Boyer, P. B.	1949	1956	O'Keefe, M.	1951	1953
Brecheen, K. G.	1948	1952	Pagenhart, T. H.	1952	1956
Cassidy, W. C.	1945	1946	Patton, C. P. (SIPRE)	1951	1952
Czuba, W.	1949	1951	Peasley, P. (USBR)	1949	1951
Daniels, G. E.	1946	1952	Rantz, S. E. (USGS)	1947	1949
Gerdel, R. W. (USWB) +	1946	1950	Rhodes, F. L. +	1947	1950
			Rockwood, D. M.	1953	1956
Hamilton, R.	1949	1950	Roelle, D.	1949	1950
Himmel, J. M.	1947	1952	Sargis, F.	1948	1949
Hildebrand, C. E.	1949	1956	Stark, F. C.	1951	1953
Hullinghorst, D. W.	1949	1951	Summers, B. L.	1947	1951
Humphrey, H. N.	1946 *	1953	Tarble, R. (USWB) +	1950	1952
Humphrey, M. N.	1946 *	1949	Threlkeld, A. F. (USWB)	1951	1952
Jencks, C. E.	1947 *	1956	Walsh, K. J. (SIPRE)	1951	1953
Kaemmerling, W. H.	1950	1952	Weimar, M. B.	1947	1956
Lewis, M.	1950	1953	Williams, C., Jr.	1950	1953
Lopez, E.	1948	1949	Wilson, W. T. (USWB)	1945	1952

1/ Minimum of one year's continuous service. All employees are Corps of Engineers (CSI) personnel, except as indicated.
2/ See table 2-2 for listing of laboratory directors.
* Some discontinuous service.
\+ Additional time spent in laboratory operation not indicated here.

ORGANIZATION CHART
COOPERATIVE SNOW INVESTIGATIONS

1 MARCH 1950



OFFICE OF THE PROGRAM DIRECTOR
COOPERATIVE SNOW INVESTIGATIONS
OAKLAND ARMY BASE
OAKLAND, CALIFORNIA

ORGANIZATION CHART

DATE 1 MARCH 1950 | PD-20-25/1

CHAPTER 2 - SNOW LABORATORY DATA

2-01. INTRODUCTION

2-01.01 At the inception of the Cooperative Snow Investigations in 1944, the selection of snow laboratory areas was given careful consideration by a group of men from both the Weather Bureau and Corps of Engineers, who were at that time working on the problem of snow hydrology research. This committee worked under the general guidance of the instituting conference and was instructed to recommend locations for three laboratories, one in the upper Columbia River Basin, one in the middle Willamette River Basin, and one in the central Sierra Nevada. General specifications were established as to physical, climatological and hydrological features of each of the laboratories. They were intended to be out-of-doors laboratories of about 5 to 20 square miles in extent, located in areas of heavy precipitation and snow accumulation, with freedom from hydrologic uncertainties regarding their physical character. Three types of climatic regions with regard to winter season precipitation were to be sampled: (1) snow, (2) rain and snow with snow predominating, and (3) rain and snow with rain predominating. Accordingly, the locations for the three laboratories were selected, and types of instrumentation were chosen. Each laboratory contained a headquarters area, where studies of physics of snow were made and continuous observations of hydrometeorological events were assured. There were also field stations located at various points throughout the laboratory area at which instruments were checked periodically by field crews and observations were made of precipitation, snow accumulations and related meteorological elements.

2-01.02 In addition to the meteorological variability among laboratories, significant differences in exposure, topography, and forest cover prevail. Inasmuch as the forest influence has particular significance in establishing methods for evaluating snowmelt rates and also in consideration of forest influence on deposition and interception of snow, the amount and type of forest were carefully considered in the selection of laboratory areas. One of the principal aims of the laboratory program was to test methods of basin application in areas which were highly instrumented relative to that generally found in project basins. Differences in environment among laboratories were essential in order to sample effectively the range of conditions that are experienced in project basins.

2-01.03 Observations at the headquarters area of each laboratory were designed to provide continuous record of the hydrometeorological elements, with the equipment serviced at least once a day, according to the requirements of individual instruments. In addition, some special experiments were performed which would provide information on the physics of snow and hydrologic application of the principles as determined by point measurements. These include observations of soil and snow temperatures, soil moisture, radiation exchange, vertical gradients of wind, temperature, and humidity, snow characteristics at surface and within the snowpack, and thermal quality and liquid water content of snow; special tests conducted on impervious snow lysimeters, water holding capacity of snow, temperature and moisture gradients of the air close to the snow, atmospheric moisture transfer experiments, water transmission through the snow, and radiation penetration into the snowpack; and testing of equipment such as precipitation gages, soil moisture blocks, and automatic radio-reporting snow gages. Determination of snow-covered areas during the melt season was considered to be essential and was made at two of the laboratories for various seasons, primarily by aerial photographs.

2-01.04 The laboratory program was designed, therefore, to meet the objectives of the Investigations, both in determining physical relationships at a point and in applying these methods to basin areas, whereby the hydrologist could analyze with assurance, problems of snowmelt runoff for project design and operation, or for river forecasting.

2-02. GENERAL COMPARISON OF LABORATORY CHARACTERISTICS

2-02.01 All laboratories were located in rugged mountainous headwater areas of the western United States on the windward side of principal mountain ranges. All are characterized by rough topography in primarily virgin areas and thereby may be used to sample hydrologic variables in remote regions which ordinarily comprise a large part of the drainages effective in producing runoff for major rivers and tributaries in the West.

2-02.02 The geographical locations of each of the three laboratories (hereinafter designated CSSL, UCSL, and WBSL for Central Sierra Snow Laboratory, Upper Columbia Snow Laboratory and Willamette Basin Snow Laboratory, respectively) are shown on plate 2-1. Topographic maps showing locations of observation stations and general topographic features for each of the laboratories are shown on plates 2-2 through 2-4. Table 2-1 lists comparative data for each of the three laboratories, by which the general features of topography, climate, and hydrology may be evaluated. Detailed descriptions of laboratory characteristics are contained in section 2-04.

2-02.03 Upper Columbia Snow Laboratory. - UCSL was located on Federal lands in the Montana Rockies at the extreme headwaters of the Columbia-Clark Fork-Flathead river system. It lies just west of Marias Pass on the Continental Divide, which forms its eastern boundary, and is partly within the Waterton-Glacier International Peace Park. The basin comprises an area of 20.7 square miles with elevations ranging between 4500 and 8600 feet msl, with an average elevation of 5700 feet. It forms the headwaters of Bear Creek and its tributaries, Skyland Creek and Autumn Creek. It is traversed by the Great Northern Railroad and U. S. Highway No. 2, which cross the Continental Divide at Marias Pass (elevation 5215 msl), and is generally accessible at all times during the year. The laboratory area comprises three sub-basins: those of Skyland Creek, Upper Bear Creek, and Lower Bear and Autumn Creeks. Streamflow is measured so as to separate Skyland Creek basin from the rest of the area. Skyland

Creek basin, comprising 8.1 square miles, has a fairly deep soil mantle and is densely forested with lodgepole pine, while the other sub-basins have less soil cover and timber. The land surface of Skyland basin has an average slope of thirty percent and an average orientation toward the west. The climate in the laboratory region is characterized by a cold snowy winter with mean temperatures during December to March of 15° and short mild summers from July through August with temperatures averaging 55° F. During the year extremes of temperatures range from about 45° below zero to more than 90° above zero. Airmasses affecting the area generally move across the mountains from the Pacific Ocean or from the Arctic plains of Canada during the winter months. Summer airmasses produce appreciable amounts of rainfall. The annual basin precipitation before interception totals approximately 50 inches, or about 40 inches after interception, and it is fairly evenly distributed during the year. Virtually all winter precipitation is snow, which accumulates to an average basin water equivalent (after interception) of about 20 inches toward the end of March, just prior to the melting period. Estimated annual runoff averages 26 inches, of which about 85 percent occurs during the six-month period, April through September.

2-02.04 Central Sierra Snow Laboratory. - CSSL is a 4-square-mile area comprising the basin of Castle Creek, a tributary to the South Fork of the Yuba River, just west of the crest of the Sierra Nevada in California. It was located partly on leased and partly on Federal lands. The area ranges in elevation from 6900 to 9100 feet with a mean elevation of 7500 feet msl. It has a moderately rough surface of granite and lava inclined toward the south-southwest. This surface bears patches of soil and moraine and is sparsely covered with stands of lodgepole pine. The laboratory basin is served by the Southern Pacific Railroad and U. S. Highway No. 40, which cross the Sierra Divide a few miles to the east at Donner Pass. The climate is distinguished by summers that are warm and rainless and winters that are cool and snowy. Winter temperatures average a few degrees below freezing. Most winter precipitation falls as snow, occasionally interspersed with fairly heavy rainfall. The annual precipitation before interception averages 70 inches, 83 percent of which occurs during the six months, October through March. The snow pack begins to accumulate in November and increases by the end of March to an average basin water equivalent (after interception) of approximately 40 inches.

2-02.05 Willamette Basin Snow Laboratory. - WBSL is an 11-square-mile area on Federal land in the dense wet forest of the middle Cascade Range, about 30 miles east of Willamette Valley and 10 miles west of the high lava plateau of central Oregon. It includes the headwaters of the Blue River, a tributary of the McKenzie River, which flows west to join the Willamette River near Eugene, Oregon. Almost all the area is covered with heavy virgin forest of Douglas fir, hemlock, and other conifers. The ground surface has an average slope of 35 to 40 percent and is oriented generally toward the south and southwest. The basin lies between 2000 and 5500 feet msl, with a mean elevation of 3400 feet. Access to the area is by Forest Service roads and trails. The climate is maritime, with a seasonal shift in airflow from southwest in winter to northwest in summer. Precipitation is heavy, totaling over 100 inches per year, about one-half of which falls as snow. Mean air temperatures vary from slightly below freezing during the winter to the high fifties in late summer. Maximum precipitation occurs in early winter and diminishes during late winter and spring to a minimum in late summer, which is nearly dry. The snowpack begins to accumulate during November and continues to increase in depth to an average of about 120 inches (as estimated for headquarters station) near the end of March, with a water equivalent of about 60 inches. Runoff from the basin averages about 75 inches annually, and is concentrated in the winter, the greatest portion occurring in the six months from November through April. Rainfall on the snow is prevalent during the winter and this situation was of importance in original selection of the laboratory site. Ordinarily, flood peak runoff, which may be expected in December and January, is primarily due to rainfall on the snow-covered watershed.

2-02.06 General comparison of topographic and environmental features. - Inspection of data contained in Part I of table 2-1, reveals that the mean elevations of the laboratory basins range from about 3400 feet for WBSL to 7500 feet for CSSL. Considering the effect of latitude differences on climatic elements for western United States, it is estimated that the mean basin elevation of each of the laboratories corresponding to a reference latitude of 45°N would be equivalent to about 6000 feet for CSSL, 3500 feet for WBSL, and 6500 feet for UCSL. The range in elevation within individual laboratories is 2200 feet at CSSL, 3400 feet at WBSL, and 4100 feet at UCSL. Skyland Creek in the UCSL has a range of 2800 feet. All basins are relatively steep, and considering average slopes of each basin area, WBSL, with an average slope of 40 percent, is steepest. Skyland Creek in UCSL is next steepest, with 32 percent average slope, while CSSL has an average slope of 25 percent. The slopes are predominantly south-facing in WBSL and CSSL, while in Skyland Creek, slightly over half of the area faces north. There is a wide range of forest types among laboratories. CSSL is sparsely forested, predominantly with lodgepole pine; only 40 percent of the basin area is forested, and the canopy density on the forested portion is estimated to be 50 percent. WBSL on the other hand, lies in an area of heavy forests of Douglas fir and hemlock-fir types which cover 93 percent of the area. Skyland Creek at UCSL is heavily forested, mainly with immature lodgepole pine, over about 90 percent of the area, and average canopy density is estimated to be 80 percent. In contrast, the remaining area on UCSL is estimated to have only 30 percent of its area forested. Considering the geology of the laboratories, the predominant rock formations range from granite and other non-porous volcanic rocks at CSSL, and relatively non-porous formations of volcanic origin at WBSL, to old sedimentary rocks which have been extensively glaciated in recent times at UCSL. None of the laboratories is known to have significant uncertainties from unknown conditions of ground water outflow from its basin. Soils are thin at CSSL and UCSL, but there are deep mountain soils at WBSL. There are large parts of the Bear Creek drainage in UCSL and at CSSL that are devoid of soils.

2-02.07 General comparison of climatic features. - Part II of table 2-1 lists general comparative climatic data for each laboratory, adjusted to basin means and based on a period representing approximately a 20-year normal ending in 1954. The average annual basin temperature ranges from 34°F at UCSL to 45°F at WBSL. CSSL is intermediate at 38°F. There is little spread in mid-summer temperatures among laboratories and they range between 56°F and 62°F for July averages. Winters, however, are much colder at UCSL than at the other laboratories. The mean January temperature for Skyland Creek is 14°F, while at the WBSL it is 32°F. At CSSL, the January temperature is 23°F.

2-02.08 There is wide variation in amount of total precipitation at each laboratory. Each year, the UCSL normally receives 50 inches, CSSL 70 inches and WBSL 105 inches.* The winter months of October through March are wettest on each of the basins; about 63 percent of the annual amount falls in winter at UCSL, about 80 percent at WBSL, and about 83 percent at CSSL. During April through September, UCSL receives nearly 60 percent more precipitation than CSSL. The average amount of winter precipitation falling as snow is about 90 percent at UCSL and CSSL, and only about 50 percent at WBSL. Interception of snowfall accounts for 10 to 20 percent loss from basin snowfall. Melt during the winter period, including ground melt and melt caused by atmospheric processes, varies in average seasonal amount from 4 to 30 inches, the UCSL being least and WBSL greatest. Solar radiation was measured at UCSL and CSSL, and incident radiation averaged about 30 to 40 percent greater at CSSL than at UCSL during the spring melt months of April, May and June. It is difficult to compare mean windiness on the basis of surface data, but average air flow at the 700 mb level (equivalent to about 10,000 feet) offers some guide to predominant wind directions. Wind data, given in Part II of table 2-1, were obtained from the Normal Weather Charts for the Northern Hemisphere 7/ and actually represent the vectorial sum of wind at a given point. All three laboratories lie in the zone of prevailing westerly wind, but there are seasonal shifts from southwest to northwest. In general, the circulation is somewhat stronger at the more northerly laboratories than at CSSL.

2-02.09 General comparison of hydrologic features. - Part III of table 2-1 lists comparative streamflow data, for a general comparison of hydrologic features of the laboratory areas. The variation in average annual runoff among laboratories corresponds to the variability of annual precipitation. Considering average basin runoff in inches, adjusted to a 30-year normal period ending in 1950, the WBSL, whose average is 77.0 inches per year, is wettest of the three laboratories. The upper portion of WBSL, as shown by the Mann Creek and Wolf Creek gages, has from 10 to 20 percent more runoff than the basin average. CSSL is intermediate among laboratories, in terms of average annual runoff, and normally has about 43 inches. Annual runoff from UCSL averages 26.5 inches, and the Skyland Creek portion has about 8 percent more runoff than the basin as a whole. Seasonal distribution of runoff for UCSL and CSSL is characterized by low winter flows, which average 15 to 22 percent, respectively,

* Beneath the forest crown.

of the annual amount during the six-month period, October through March. At WBSL, on the other hand, from 50 to 60 percent of the runoff occurs during the winter season.

2-02.10 Because of the relatively short periods of record at the laboratory basins, extremes of discharge are not necessarily significant for comparative purposes. The maximum flow at CSSL corresponds to a discharge of 300 cfs per square mile, and occurred in November 1950, as the result of an intense rain on a light snowpack. This flow is considered to represent a near-record maximum for this type of area. At WBSL, the maximum observed discharge of 122 cfs per square mile is one which would be expected once in every few years for its area, on the basis of maximum flow in surrounding streams. At UCSL, the maximum recorded discharge was only 35 cfs per square mile. Minimum discharges range from 0 to 0.26 cfs per square mile. At CSSL, the flow in Castle Creek entirely ceases every summer. In the other two laboratories, there is sufficient flow from ground-water storage to cause a carryover of runoff during periods of no basin moisture input.

2-02.11 In order to define the relative basin time lags resulting from all factors serving to delay runoff other than the snowpack itself, average recession curves were derived for each laboratory basin. These curves define the time rate of change of flow during normal flow recession and thereby provide a measure of the natural time lag characteristics of the basin. From curves derived empirically by methods set forth in chapter 4, it can be stated in general that in Castle Creek at CSSL, the delay of runoff that is due to surface and subsurface storage is about one-half of the corresponding delay for either WBSL or UCSL, throughout the various ranges of unit flows in cfs per square mile. When comparing recession coefficients on the basis of actual magnitude of discharges, CSSL has, at relatively high flows, about one-third to one-fourth the delay that either USCL or WBSL has. This effect is evident from the much greater magnitude of diurnal fluctuation of flows at CSSL, compared with WBSL and UCSL. The recession characteristic integrates a multitude of basin effects, including average slopes over the basin, types of soil, channel lengths and conditions, ground-water geology, forest cover, and many lesser influences, into a single average relationship, and thus becomes a very useful tool in applied hydrology and hydrograph analysis. More detailed comparisons of laboratory recession curves are presented in paragraph 2-04.35.

2-03. LABORATORY ADMINISTRATION

2-03.01 The Cooperative Snow Investigations had, as its first major operation, the establishment of each of the three snow laboratories, including the necessary provisions for laboratory operation and instrumentation. In addition, operational and administrative channels were formulated, in order to provide adequate control of the operation as a whole. Reference is made to plate 1-1, showing channels of command for laboratory operation during the cooperative phase of the work between the Weather Bureau and the Corps of Engineers. The responsibility for selection of instrumentation, type and frequency of observations, and methods of handling the data was held by the program director. Coordination and standardization of observational techniques for the laboratories, as well as supervision of special tests and development of new instrumentation, were effected by the Technical Supervisor. The administration of the physical operation of laboratories, including arrangements for procurement of supplies and equipment, living facilities, transportation and land acquisition, was accomplished by the Corps of Engineers District Offices having supervision of the area in which the laboratory was located. The direct responsibility of laboratory operation, to implement the requirements set forth from both technical and administrative supervision, rested in the laboratory director, who was a Weather Bureau employee at CSSL (up to 1950) and a Corps of Engineers employee at UCSL and WBSL. Both Weather Bureau and Corps of Engineers employees were on permanent duty at CSSL and UCSL, and personnel administration was accomplished by their respective offices. At WBSL, all employees on permanent duty were under the supervision of the Corps of Engineers. Table 2-2 lists the laboratory director and average number of employees, both Corps of Engineers and Weather Bureau, for each year of laboratory operation.

2-03.02 Staffing presented a major problem in laboratory operation. Since the laboratories were located in remote, headwater regions in the mountainous areas, there was considerable difficulty in obtaining properly trained personnel who could adapt themselves to the rigors and isolation required, and at the same time pursue with diligence their normal and sometimes unique duties. Trips to field stations were made weekly or bi-weekly throughout the winter, and instruments were, as a rule, serviced under adverse weather conditions. The WBSL was particularly isolated and a large percentage of the effort of the staff was required for maintenance of minimum living standards.

2-03.03 Agency cooperation. - Besides the Weather Bureau and Corps of Engineers, other Federal Agencies participated in some of the functions of the field observational program, on a cooperative basis. The U. S. Geological Survey installed and operated the stream gaging stations at both WBSL and UCSL, on a reimbursable basis. Aerial photographs of snow cover through the melt season were made in cooperation with the U. S. Air Force at CSSL and U. S. Forest Service at UCSL. Aerial photographs were also obtained by the Forest Service for the purpose of constructing aerial mosaics of each laboratory, from which basin topographic maps could be prepared. The Forest Service also installed soil moisture meters at UCSL and obtained observational data from them. Forest Service cooperation included construction and maintenance of roads and trails, notably at UCSL, and in some cases, use of buildings or shelters.

2-03.04 Cooperation with Snow, Ice, and Permafrost Research Establishment. - Responsibility for snow, ice and permafrost research for the joint benefit of the U. S. Armed Services was assigned to the

Department of the Army and to the Chief of Engineers for operations in 1949. The Snow, Ice and Permafrost Research Establishment (hereinafter referred to as SIPRE) was organized the same year. Its purpose is to perform basic research on properties of snow, ice and permafrost, and the application of basic snow research to military problems.[3] In 1950, the Weather Bureau terminated its participation in the Cooperative Snow Investigations laboratory program, and at approximately the same time, SIPRE was embarking on an observational program in snow research. Since the facilities of CSSL were available, arrangements were made whereby SIPRE could establish its facilities for observations at CSSL, and the observations were then cooperative between SIPRE and the Snow Investigations. While the Snow Investigations Unit was interested primarily in hydrologic application of snow research, and SIPRE was concerned with basic research leading to application for military use, many of the observations could satisfy both requirements. The establishment of a micrometeorological program at the Lower Meadow was one of the major accomplishments done under the direction of SIPRE, and special observations were performed at the request of each of the agencies. The laboratory director was employed by SIPRE, but administrative operation of the laboratory was under the direction of Sacramento District. This phase of the observational work at CSSL closed in June, 1953, when SIPRE moved to its new laboratory in the midwest, and the snow investigations terminated year-round observations.

2-03.05 CSSL was operated during the 1954 melt season under the direction of the Snow Investigations Unit. Observations were confined to special studies in connection with previously constructed snow lysimeters. Runoff for the laboratory basin was measured, and meteorological instrumentation was maintained at the headquarters area and at the Lower Meadow. No basinwide snow surveys or precipitation measurements were made. At the close of the 1954 melt season, CSSL was closed and no further observations were performed by the Snow Investigations Unit. UCSL was closed at the end of the 1951 melt period, and observations at WBSL were terminated in 1952.

2-04. DETAILED DESCRIPTION OF THE LABORATORY AREAS

2-04.01 General. - It was originally the intent of the program to prepare detailed technical reports on the physical and climatological features of each laboratory. Lack of time precluded the completion of those reports, except for Technical Report 4-A, entitled, "Terrain Characteristics, Central Sierra Snow Laboratory." This report presents in detail the features of that basin with regard to topography, geology, vegetation, and drainage, whereby analysis could be made for transferring the hydrologic variables to conditions of known environment on project basins. The work included the delineation of the 4 square mile basin into 20 topographic units of individual characteristics. Such detailed analyses have not been accomplished for the other laboratories, and later studies have indicated that more general classifications of terrain factors affecting snow accumulation and snowmelt are adequate, when considering the relative degree of accuracy of measured amounts in basin application. Therefore, this section presents comparative data generally in less detail than was originally considered necessary in developing hydrologic relationships, but the comparisons are believed to be adequate for the purpose of obtaining qualitative evaluation of methods used in relating measured to basin amounts. Each laboratory is discussed as to its physical landscape and general characteristics of climate and hydrology. Insofar as possible, direct comparisons of laboratory features are presented, with regard to both verbal descriptions and presentation of data by diagrams.

2-04.02 Laboratory access. - The ease of operation of each laboratory and the maximum utilization of personnel on the observational program was dependent to a large extent on the ease of access to the basin area from major communication routes and on the extent and quality of roads and trails within the basin areas. The headquarters area at UCSL was located adjacent to U. S. Highway 2 and the Great Northern Railroad, both of which are all-weather arteries through the Continental Divide at Marias Pass and connect the Pacific Northwest with the midwest of the United States. Although the highway is occasionally blocked by snow during the winter, railroad traffic is interrupted only under very unusual conditions. Travel within the laboratory was accomplished by roads and trails, which could be negotiated by four-wheel drive vehicles during the summer to service outlying gages. In winter, snow tractors were utilized for transportation within the laboratory area whenever conditions would permit, but approximately half of the distance traveled was accomplished on foot, either with skis or snowshoes. Commercial power facilities were not available at the time the laboratory was established, so portable power supplies were used to furnish electrical power at the headquarters area.

2-04.03 CSSL is similar to UCSL with respect to major highway access; U. S. Highway 40 borders the laboratory area on the south, and the Southern Pacific Railroad connecting San Francisco with the East parallels the highway at Donner Pass through the Sierra Nevada. With the exception of short periods following major storms they were kept open throughout the winter season. Within the laboratory area, roads and trails served as access to field stations during the summer, and snow tractors could be utilized over much of the area during winter. About 50 percent of the travel to service outlying stations was accomplished by snow vehicle, and the remainder on foot, either with skis or snowshoes. Commercial power supplied electrical energy to the headquarters area and later to the Lower Meadow micrometeorological observation station.

2-04.04 Access to WBSL was difficult. The closest highway to the headquarters area was 5 miles by forest road, which required that when snow was in the area (usually from October through May or June), transport of all supplies, equipment and personnel be by snow vehicle or on skis. The South Santiam highway, which the forest road joins at Rabbit Camp, was kept open throughout the winter but was occasionally closed following a major storm. Travel within the laboratory area was tedious. The use of snow vehicles was limited to about 10 percent of the total travel to outlying stations, because of many fallen trees over the roads and trails, and because of the steep slopes in the area. There was no commercial electric power, and portable power equipment was utilized at the headquarters area.

2-04.05 Surface configuration. - For the purpose of this description, comparative data for each laboratory are given below for stream systems and profiles, area-elevation relationships, steepness of slopes, and orientation. It is believed that these general classifications of topography are sufficient for defining the features of each laboratory, in order to arrive at a qualitative appraisal of the characteristics of each area. Reference is made to the topographic maps of each laboratory contained on plates 2-2 to 2-4.

2-04.06 UCSL, with 20.7 square miles total area, may conveniently be divided into two subdivisions: (1) Skyland Creek basin and (2) the Intermediate area of Upper Bear, Lower Bear and Autumn Creeks. Skyland Creek drains 8.09 square miles of steeply rolling forested area bounded by the Continental Divide on the east, Challenge Divide on the south, and Mule Ridge on the southwest. The Intermediate area of 12.61 square miles is bordered by the rugged Algonkian Ridge on the northwest, which is a very steep glaciated escarpment about 4 miles long whose summit averages about 2500 feet higher than the gentle slopes of the valley floor. The Blacktail Hills separate Upper Bear Creek drainage from Autumn Creek drainage and are low-lying hills whose summit averages about 6000 feet. Stream profiles for Bear Creek and Skyland Creek are shown on figure 1, plate 2-5. The average slope of Bear Creek is about 140 feet per mile after it emerges from the rugged slopes of Bear Peak. The slope of Skyland Creek averages about 200 feet per mile in its lower reaches and the channel slopes gradually steepen when approaching the headwaters of Elkcalf Mountain. In general, all major tributaries are typical mountain streams consisting of alternate cascades and pools. There are no lakes on the stream courses, but a portion of Upper Bear Creek passes through an area of marshy ground which is slightly less than a mile long. Almost all of the analytical work on UCSL was performed on the Skyland Creek drainage because of its freedom from excessively irregular topography and its greater homogeneity, in comparison with the Intermediate area.

2-04.07 CSSL contains 3.96 square miles of rugged land located just west of the summit of the Sierra Nevada, the eastern boundary of the laboratory forming a segment of the divide. The northern end of the basin is bounded by Castle Peak (elevation 9105 ft.) which forms a sharp escarpment with steep slopes, in places almost vertical, and rises about 1500 feet above the valley floor. On the west, the divide is formed in part by Andesite Peak, whose maximum elevation is 8215 feet. Andesite Ridge projects southeastward into the basin from Andesite Peak, and effectively divides the western half of the basin into two parts. Castle Creek is the only major drainage channel within the basin. It heads on the slopes of Castle Peak, flows southeastward through Willow Valley and

Upper Meadow and swings in around Andesite Ridge, from where it flows southwestward to the basin outlet. Channel slopes average about 200 feet per mile except for the Lower Meadow, where the slope is about 70 feet per mile. The headwaters of Castle Creek on the face of Castle Peak are much steeper. There are no lakes on the main stem of Castle Creek. Grass Lake, located in Euer's Saddle on the eastern edge of the basin has little effect upon runoff because of the very small area contributing to it. Reference is made to Technical Report 4-A for more detailed description of the terrain features of CSSL.

2-04.08 The 11.5 square mile area of WBSL may be divided into three segments. Mann Creek, which drains nearly the entire northern half of the basin, contains 5.12 square miles. Wolf Creek drains a segment of 2.06 square miles on the east side of the basin. Mann Creek and Wolf Creek join to form Blue River. The area draining into Blue River below the confluence and above the laboratory basin outlet is the third segment and is identified as Intermediate Blue River Drainage (D.A. = 4.33 sq. miles). The northern edge of WBSL is formed by the divide separating McKenzie River and Santiam River drainages. Squaw Mountain (elevation 5235 ft.) is the principal peak on this portion of the divide. The eastern boundary is formed by a divide culminating in Carpenter Mountain (elevation 5364 ft.), located in the southeast corner of the basin. Mann Ridge, which varies from about 3000 to 4000 feet msl, serves as the western boundary of the basin. A small butte known as Wolf Rock is a volcanic neck rising some 1000 feet above the surrounding valley and lies in the east central portion of the basin. Mann Creek heads in the vicinity of Squaw Mountain and flows generally southward to confluence with Wolf Creek near the center of the basin. Wolf Creek originates in Wolf Meadow and flows westward to the south of Wolf Rock. Blue River flows generally southwestward to the laboratory basin outlet, which in turn is some 14 miles above its junction with the McKenzie River. Channel slopes on Blue River, within the laboratory basin, average about 200 feet per mile, but slopes of Mann and Wolf Creeks are about 500 feet per mile. The streams are all swift mountain cascades, and there are no lakes of significance within the basin.

2-04.09 Area-elevation relationships. - Area-elevation data were computed for each laboratory, and graphical plots of these relationships are shown on figure 1, plate 2-6. The data were obtained by planimetering zones of elevation from the topographic maps for each laboratory. The curves represent the percentage of area above a given elevation for each laboratory basin and major subdivision. All are plotted on a common scale of elevation and percentage of area, so that direct comparison of elevation characteristics of each laboratory may be made. The following table summarizes the data from this diagram:

Laboratory Drainage	Drainage Area Sq. Mi.	Elevation Feet MSL			Elevation above which lies given percent of area	
		Mean	Max	Min	25%	75%
UCSL (Total)	20.7	5700	8605	4480	5950	5350
Skyland Creek	8.09	5900	7610	4800	6175	5600
Intermediate Area	12.61	5500	8605	4480	5700	5225
CSSL (Total)	3.96	7500	9105	6880	7725	7275
WBSL (Total)	11.51	3433	5364	1959	3925	2950
Mann Creek	5.12	3760	5235	2491	4175	3350
Wolf Creek	2.06	3600	5364	2491	3850	3275
Intermediate Area	4.33	2980	5364	1959	3325	2550

2-04.10 Area-slope relationships. - Area-slope relationships were derived for each laboratory, on the basis of sampling on a topographic map the slope characteristic at the grid-intersection points uniformly spaced over the basin. There were approximately 250 such intersection points on each laboratory sub-basin. The percentage of area for a given slope was accumulated from the steepest slope, downward, and curves were plotted representing the percentage of area whose slope is equal to or greater than a given slope, as shown on figure 2, plate 2-6. A summary of these data is listed in the following tabulation:

Laboratory	Drainage Area Sq. Miles	Slope in Percent*				
		Basin Mean	Values of slopes equalled or exceeded, for given percentage of area			
			10	20	50	80 % of area
UCSL (Skyland Creek)	8.09	32	52	44	30	21
CSSL (Total)	3.96	21	51	26	16	9
WBSL (Total)	11.5	40	63	53	38	25

It is seen that the mean slope of CSSL is about one half of that for WBSL, and that Skyland Creek in UCSL is about midway between the two. There

*Slope is measured as the vertical rise in feet per 100 feet of horizontal distance, averaged for a distance of 500 feet from each intersection point.

are some very steep slopes in all the laboratory areas, as shown by the values given for the steepest 10 percent of the area, ranging from 51 to 63 percent.

2-04.11 Orientation. - The basin orientation is important for two general considerations: (1) its effect on the accumulation of precipitation, both in the form of rain and snow; and (2) its effect on snowmelt rates. The effects may be independent of one another, when considering the prevailing meteorologic conditions during the snow accumulation period and their relationships with terrain, and also, when considering the meteorologic and terrain factors affecting melt. General evaluation of orientation was determined on the basis of basin averages for each of the laboratories, and graphical plottings of the percentage of the basin area facing in a given octant of the compass are shown in figure 3, plate 2-6, for (1) Skyland Creek, UCSL; (2) Mann Creek, WBSL; (3) the entire WBSL; and (4) CSSL. These graphs are presented in such a manner that the area in a given octant on the graph is directly proportional to the percentage of basin area whose orientation is within the octant.

2-04.12 Comparing orientation among laboratories, it is shown that the CSSL and Mann Creek (WBSL) areas have similar average orientation, with nearly 50% of the two areas facing in the quadrant from SE to SW. Skyland Creek, on the other hand, is nearly uniformly distributed between north and south orientation. The following table lists percentages of basin area for given sectors of orientation:

Laboratory Drainage Area	Percentage of area facing quadrant centered on: N	E	S	W
Skyland Creek, UCSL	31	15	31	23
WBSL (entire area)	17	21	34	28
Mann Creek, WBSL	6	22	48	24
CSSL (entire area)	11	17	49	23

2-04.13 Geology. - The evaluation of the effect of surface and subsurface rock formations on transmitting and storing water requires adequate geologic investigation. To accomplish this, detailed geologic field surveys were performed for UCSL and CSSL. No such survey was made for WBSL, but geologists from the Portland District made cursory field examinations of that area; in addition, published geologic descriptions for the Cascade Mountains were reviewed in detail. It is sufficient for the purpose of this report to describe the geology only in general terms, but emphasis is placed on the conclusions from investigations regarding the water permeability of the basins, considering the likelihood of water passing into the basin from surrounding areas or out of the basins through underground channels and thereby not measured as basin outflow. The storage time of delay to runoff by the combined effect of storage and flow through subsurface channels (both through the soil and underlying rocks) is taken into account by streamflow recession analysis, and direct evaluation of water stored in basin aquifers is not necessary. Therefore, the principal consideration of the geologic character of each basin is the determination of how well the basin boundary determined by surface drainage patterns represents the true area contributing to runoff, considering also the possibility of water loss by deep percolation.

2-04.14 The physical condition of UCSL is characterized by a mature stage of stream erosion, with sharp ridges and steep slopes, as shown on the topographic map (plate 2-2). It has been modified somewhat by glaciation, and glacial debris remains on a large part of the area and varies in thickness from a few feet up to 150 feet in the valley bottoms. Rock is exposed practically everywhere around the border except at Marias Pass, and there are rock outcrops scattered throughout the basin. The rocks are sedimentary and consist predominantly of sandstones, limestones, shales, and conglomerates belonging to the Jurassic and Cretaceous periods of the Mesozoic era. On the north end of the basin, the rocks forming the higher portions of Algonkian Ridge belong to a Pre-Cambrian belt consisting of limestone and argillite, and lie above the Lewis Overthrust Fault, which is exposed on the south slopes of Bear Peak. Nearly everywhere the rock strikes about north 60 degrees west and dips about 40 degrees to the southwest. The entire area is intricately faulted but with minor folding, and it is believed the flow of water through fault zones is negligible. Because the entire drainage area of Skyland Creek is bounded by rock ridges, it is believed that the basin neither gains nor loses water underground around the periphery. A careful study of conditions at the junction of Skyland with Bear Creek, where ground water is observed seeping into the creek from the overburden, revealed that there is no loss of water from Skyland Creek near its mouth. The boundaries of the drainage area of Bear Creek in the vicinity of Marias Pass are poorly defined, but since nearly all studies of runoff are confined to Skyland Creek, it is of little consequence.

2-04.15 CSSL is founded on granite which forms part of the Jurassic batholith of the Sierra Nevada. Subsequent to the erosion of the granitic surface, flows of volcanics including rhyolite, andesite, volcanic mud, and basalt covered the area, part of which was later removed by glaciers, which in turn left small patches of moraine. Along Castle Creek are small areas of recent alluvium. The following tabulation lists the proportional amounts of the basin area presently exposed to the above listed formations:

Formation	Percent of Basin Area
Granite, mostly unweathered and tightly jointed	35
Overlying volcanic rocks, some rather porous	55
Glacial moraines, permeable but discontinuous	5
Alluvial deposits	5

Detailed descriptions of the rocks and delineation of the surface geologic pattern are contained in Technical Report 4-A. In general, the lower (SW) quadrant of the basin is predominantly granitic, while the entire upper portion contains volcanic rocks. Alluvial materials have been deposited along the sections of the stream channel with flatter gradients characteristic of the Lower and Upper Meadows. Little is known of the fault structure within the basin proper. Surface drainage boundaries are well defined except in Euer's Saddle, lying on the eastern edge of the basin. Conditions of subsurface flow near the basin boundaries are not known, but it is assumed they conform closely to the surface pattern. Deep percolation, whereby water could leave the basin unmeasured by the stream gage for Castle Creek, is assumed to be negligible, but no definite geologic information is available to substantiate this. When determining the water balance for CSSL as a whole for 5 years of record, (see chapter 4), the measured outflow appears to be too low. This is not sufficient evidence to conclude that there is unmeasured ground-water outflow from the basin, but it does point to that possibility.

2-04.16 No detailed geologic investigation of the WBSL was made, primarily because of the difficulty in identification of rocks and their structure. The heavy soil and forest cover over nearly all of the area effectively obscures the underlying formations, and rock outcrops appear only on Wolf Rock and on some of the peaks bordering the basin. Generalizations of the geologic structure of the basin were determined from published surveys of areas in the mid-Cascades, and by field reconnaissance of the area. The geology is considered to be typical of the middle Cascades area, which was formed from volcanic material ranging from lavas to agglomerates and tuffs. It has gone through a period of faulting and folding, and in recent times may have been subjected to glaciation, but there is no direct evidence of morainal deposits. The basin has been eroded to an early stage of maturity, and stream patterns are well defined except for small meadow areas. Heavy vegetative cover and high precipitation have resulted in deep weathering, and in most places the rock is covered with a thick mantle of relatively impervious soil. The bedrock underneath the soil mantle consists of weathered andesite (a fine grained volcanic rock) together with tuffs and breccias, which help to serve as a storage reservoir to supply summer flows for the numerous perennial streams in the area. The small butte known as Wolf Rock has been identified as a volcanic neck, but so far as is known, it has no particular significance in relation to the basin's ground-water geology. The basin boundary is formed by sharp divides between adjacent drainages, and considering the relatively impervious character of the soil mantle and underlying rocks, there is little reason to suspect underground water loss or gain from adjacent areas. No information exists on the strike and dip of the rocks within the laboratory area. Inspection of streamflow records for adjacent areas and on the entire Blue River (D.A. = 75 sq. miles) during periods of summer ground water recession flow show that unit rates of runoff per square mile are nearly identical for areas in the vicinity. This points to the probability that there is no sizeable water loss by underground flow past the stream gaging station, in view of the homogeneity of the area as a whole.

2-04.17 Soils. - The soil mantle on laboratory basins is important in the analysis of basin water-balance computations as a storage reservoir for the infiltrating water upon which plant growth normally draws in the transpiration process. It also represents a part of the ground-water zone which induces a time delay of runoff from water excesses. The soils at the laboratories are described qualitatively from field observations; however, no comprehensive soil surveys were performed whereby the soil characteristics for the basin as a whole could be evaluated. There are two differing concepts used in defining soils. In agricultural usage the zone of soil consists of that part of the earth's crust penetrated by plant roots; and in engineering usage soil is the total layer of unconsolidated material, thus differentiating between rock and earth materials. Both definitions are important to the hydrologist when considering separately the processes of transpiration and ground-water flow.

2-04.18 Skyland Creek in UCSL is almost completely covered with a thin soil mantle formed primarily by weathering of the sedimentary rocks underlying the basin, and, also, of the relatively small areas of old glacial debris. The remaining Bear and Autumn Creek areas were scoured heavily by glaciation during recent times, and there has been little opportunity for the formation of soils. The soils are thinner and there is a much larger percentage of area bare of soil than on Skyland Creek.

2-04.19 In general, soils at CSSL are thin, and the area is characterized by wide variation in soil conditions over the basin. Post-glacial weathering has not appreciably affected the granite rocks that were scoured by ice during recent glaciation, but the glacial deposits have been weathered to form immature soils. Soil is lacking on many of the steeper lava slopes, but the flatter areas of agglomerate have weathered fairly deeply. The meadows containing principally alluvial material have an overlying soil mantle. For the basin as a whole, it is estimated that about half of the area is soil-covered. The texture of the soil is generally light, and it is classified as sandy loam. There are, however, a few areas where the soil is underlain with clay.

2-04.20 Soils at WBSL are thick and cover nearly the entire basin, with the exception of portions of Wolf Rock and Squaw Mountain. They have resulted from weathering of the underlying volcanic rocks and consist primarily of clayey material. There was no soil probing or sampling program in the area, but generalizations on probable soil conditions can be made from known conditions in adjacent basins of similar topographic conditions. The total depth of earth materials probably ranges from a foot or two to as much as perhaps 50 feet. In general, it is believed that the thickness over the major portion of the area would be in excess of 10 feet. Observations of the soil conditions at the 5 ground-water wells showed uniformity of conditions, with approximately $1\frac{1}{2}$ feet of loam at the surface, and a shallow transitional zone into the underlying clayey material. The wells were not necessarily dug to bedrock and the depths range from 7 to 12 feet. The active zone penetrated by the plant roots is believed to be no more than four feet throughout the basin, and soils below this depth are aquifers for ground-water flow.

2-04.21 Vegetation. - The vegetation cover at each laboratory is an important factor to be considered in the hydrologic comparison of laboratory areas and in the extension of methods of analysis from the laboratory to project basin areas. The effect of grasses and low-growing shrubs on the accumulation and melting of the snow pack is minor, but they would have some influence on transpiration and infiltration rates. Forests, on the other hand, have a major influence on the accumulation of snow, because of differences in snow accumulation between sites in clearings and those beneath the forest crowns, and interception of precipitation in the form of both rain and snow. Forests also play a major role in the process of heat transfer to the snowpack and to the lower layers of the atmosphere. Water loss by evapotranspiration from a snow-covered area is accomplished largely by forest transpiration. In the following descriptions, therefore, the amount and type of forest is given primary consideration.

2-04.22 UCSL is characterized by wide variation in amount of forest from one side of the basin to the other. Skyland Creek, comprising the southeast third of the basin, is heavily forested with conifers, predominantly lodgepole pine (Pinus contorta), over about 90 percent of that area, and the average canopy density is estimated to be 80 percent. In contrast, large sections of Autumn and Upper Bear Creeks are unforested, and those areas which are forested are patchy. The net crown cover for this area as a whole is estimated to be 30 percent. An aerial mosaic of UCSL is shown on plate 2-7, and illustrates the forest cover over the basin. While lodgepole pine is the dominant type, it is intermixed with Engelmann spruce (Picea engelmanni), white fir (Abies concolor), and tamarack (Larix laricina). This forest is not the climax type for this area, and there is evidence of old burns. The climax types would be Douglas fir (Pseudotsuga taxifolia) and western yellow pine (Pinus ponderosa) in the lower portions of the basin, and fir-hemlock species in the higher levels. The trees in the forest at present average between 30 and 50 feet in height; the Skyland Creek stands are dense, and the average distance between tree trunks is about 10 feet. In general, there is little underbrush beneath the dense forest canopy, but in the openings, there is a heavy growth of a wide variety of shrubs. Bear grass grows in the openings at the higher elevations. The bare areas, particularly on the north side of the basin, are rock outcrops or talus slides which cannot support plant growth.

2-04.23 The forest at CSSL is primarily second-growth lodgepole pine and the basin is characterized by a relatively light, open forest. There is considerable range in forest density in different portions of the basin, varying principally with elevation, soil, and exposure. The average timberline in this area is at about 8000 feet, but occasional growths of hemlock and fir occur above this elevation. Plate 2-8 is an aerial mosaic of CSSL taken when the area was bare of snow, and shows the forest density over the basin. For the basin as a whole, about 40 percent of the area is forested, and it is estimated that only 20 percent of the basin is directly beneath the tree crowns. The average height of the pines is from 30 to 50 feet. There are interspersed a few stands of red fir (Abies magnifica), white fir (A. concolor) and western yellow pine, whose heights range up to 125 feet. Various brushes cover about 15 percent of the basin. Grass covers about 6 percent of the area with no other cover, and 40 percent of the basin is bare of any vegetation.

2-04.24 WBSL is heavily forested and is typical of the climax type of the western Cascades in Oregon. Non-timbered areas are few and consist of small mountain meadows which are usually less than one-quarter acre in size. A notable exception, however, is Wolf Rock, whose area is about one-eighth of a square mile. Douglas fir is the dominant species and comprises about 60 percent of the timbered area. The remaining 40 percent of the forest belongs to the hemlock-fir types and consists of Pacific silver fir (Abies amabilis), noble fir (A. procera), white fir, western hemlock (Tsuga heterophylla), Engelmann spruce, and mountain hemlock (T. mertensiana). Old-growth and second-growth trees of all sizes are found in both timber types. The average height of trees is estimated to be about 150 feet. The total forested area is 93 percent of the basin as a whole, and the average canopy density is estimated to be about 90 percent. Plate 2-9 is a panoramic photograph taken looking south from Squaw Mountain over the basin area, showing the character of the dense forest. Beneath the forest crown is an understory of various types of shrubs which in turn cover low-lying ferns and other weeds. The understory is denser in the lower part of the basin and includes small deciduous trees.

2-04.25 Climate. - The climatic regimes of the three laboratories are similar, in that all of the areas are predominantly exposed to Pacific maritime airmasses as the result of the western atmospheric circulation of the middle latitudes. Occasional reversals in circulation, however, cause invasion of continental airmasses, and in some periods the circulation is weak, so that there is established a local climate which is mostly independent of the atmospheric circulation. The average climate of the regions, then, is a function of (1) the relative frequencies of the above listed meteorological patterns, (2) the opportunity for modification of the airmasses, which is primarily a function of distance from the sea coast and extent of intervening topographic barriers, and (3) the latitude and elevation of the basins. Reference is made to plate 2-1, which shows the geographical location of the laboratories with respect to the major topographic features of western United States.

2-04.26 In order to determine climatic averages of basic hydrometeorologic variables for each laboratory, the relatively long-term records of temperature, precipitation, snowpack water equivalent, and runoff for key stations in or adjacent to the laboratory areas were analyzed on the basis of mean monthly amounts. Means were computed for the entire period of record of the individual stations and also for the shorter period of laboratory record, in order to establish the relation between the meteorologic characteristics during the period of laboratory record and long-term climatic means. The following tabulation lists the key hydrometeorologic stations and the total years of record used for each element to represent the long-term hydroclimatic mean at each laboratory.

Snow Lab.	Temperature	Precipitation	Snowpack Water Equivalent	Runoff
UCSL	Summit, Montana 1938-1954	Summit, Montana 1937-1954	Marias Pass, Montana 1936-1955	Middle Fork Flathead River at Essex, Mont. 1940-1952
CSSL	Soda Springs, California 1930-1954	Soda Springs, California 1930-1954	Soda Springs, California 1930-1955	South Fork, Yuba R., near Cisco, Calif. 1943-1952
WBSL	Leaburg, Oregon 1934-1954	Leaburg, Oregon 1934-1954	Santiam Jct., Oregon 1941-1955	Blue River near Blue River, Ore. 1936-1952

Tables 2-3 through 2-5 list mean monthly values of temperatures in °F, precipitation in inches, and runoff in inches over the contributing area for January through December, for the above listed stations. Average snowpack water equivalents are given for the first of each month, from January through May, but for some of those months at certain stations values are omitted because records are incomplete or lacking entirely.

2-04.27 Reference is made to the descriptions of general climatic and hydrologic features of the laboratories contained in paragraphs 2-02.07 through 2-02.11 for summaries of the hydroclimatic characteristics of the basins. The climatic comparisons set forth in tables 2-3 to 2-5 are presented for the purpose of interpreting values of prime hydrometeorologic variables presented on the basis of averages for the laboratory basins and for the period of laboratory record.

2-04.28 During the period of laboratory operations, the mean annual temperature at each laboratory was within one degree F of its long-term mean, but there were some anomalies for individual months. At UCSL, winters averaged somewhat colder than normal, and in extreme, January was nearly 7°F below normal. During the spring months the temperatures averaged slightly above normal. Precipitation at UCSL for the water years 1947 through 1950 was about 10 percent above normal; most of the excess occurred during the period October through March. Spring and summer precipitation was very nearly normal. Water equivalent of the snowpack as measured at Marias Pass averaged a little over 20 percent above normal, reflecting the above normal precipitation and below normal temperatures during the winter. Annual runoff was about 15 percent above the long-period average. WBSL similarly experienced somewhat below normal winter temperatures, and above normal precipitation, water equivalents, and runoff during the period of laboratory operation. Values of temperature, precipitation, and runoff averaged near normal at CSSL during the period of laboratory record. The water equivalent measured on the snow courses at Soda Springs indicated that the snowpack contained only 60 to 80 percent of the normal amount for the early spring months, but it is believed that the apparent departure may be due to above normal percentages of winter precipitation falling as rain during the period of laboratory operation.

2-04.29 Hydrologic comparison. - One of the principal objectives of the laboratory program was to establish with relative assurance the components of the hydrologic cycle in areas of snow accumulation, under varying conditions of climate and environment. This has been accomplished on the basis of basin mean values for each element for each of the laboratories, by monthly increments over the period of laboratory record. The methods used in deriving those values are set forth in chapter 4, and tabulations of amounts by individual months are presented in that chapter. The average amounts for the period of record for each laboratory were summarized for the purpose of presenting in this chapter a hydrologic comparison of the areas. Graphical representation of these summaries is shown on plate 2-10 and include mean monthly values of (1) basin temperature, (2) incident radiation (where available), (3) basin precipitation and basin snowfall, (4) accumulated net basin snowfall (after interception) and basin water equivalent, and (5) generated runoff.

2-04.30 Basin temperatures were computed on the basis of means for the period of record adjusted to the mean elevation, except for CSSL, where elevation effect is small and station 1-B was used to represent basin means directly. Records of insolation were obtained at CSSL and UCSL, but are not available for WBSL. Total basin precipitation represents, as a basin mean value, the gross amount for tree-top level. Basin snowfall similarly represents the average monthly amount above the tree canopy. Separation of rain and snow was made on the basis of detailed studies of the form of precipitation and independent checks by the total water balance computation. The accumulated net basin snowfall represents average basin amounts of the water equivalent of newly fallen snow that arrives at the ground level and, accordingly, interception losses are accounted for in these amounts. The basin water equivalents are given as of the end of the month and represent mean basin amounts of water remaining in storage in the snowpack. Differences between accumulated net basin snowfall and basin water equivalent are the summation of melt to that particular time. The values for generated runoff are the equivalent basin inches of monthly runoff, adjusted for changes in ground-water storage by means of recession curve analysis as described in paragraph 2-02.11.

2-04.31 Plate 2-10 presents a graphical comparison of the characteristics of each laboratory basin with regard to the hydroclimatic features discussed above. These comparisons serve as orientation for detailed analysis of laboratory data presented in subsequent chapters, with regard to snow accumulation, snowmelt, and hydrograph reconstitution. In general, it is seen that UCSL has severe, snowy winters, and warm summers. Precipitation occurs predominantly during the winter season and is almost entirely in the form of snow during that period. There is significant precipitation, largely in the form of rain, during the spring and summer months, and a secondary precipitation maximum occurs in June. On the average, basin snowpack water equivalent accumulates through March of each year, but the heat supply is usually sufficient after April first to produce enough melt to cause a net decrease in water stored in the snowpack. Runoff generated during the winter months is generally very small, but October or November rainfall, or occasional minor periods of snowmelt, may provide water in excess of soil moisture requirements and thereby produce runoff. About 80 percent of the annual runoff is generated in the 3 months of active snowmelt, April through June. Usually, the snow is completely melted by the first of July, but occasionally, a late melt period will cause a small carryover of snowmelt into July.

2-04.32 Winters at CSSL are not as cold as at UCSL, but precipitation is about 50 percent greater. The proportion of precipitation falling as rain during the winter (October through March) is greater at CSSL than at UCSL, and for the period of laboratory record was about 22 percent. It is believed that this value is somewhat above normal for this area, because of unusually heavy rains that fell in November, 1950 and other periods. The average April 1 basin snow accumulation each year at CSSL is equivalent to about 32 inches of water, while it is about 24 inches at UCSL. Seventy-six percent of the annual runoff at CSSL is generated in the three-month period, April through June, resulting from the melting of the snowpack. Summer precipitation is negligible, and Castle Creek usually becomes dry near the first of August. Winter flows are proportional to winter rains and snowmelt, and while they are greater than those experienced at UCSL, the volume of runoff is relatively low. Occasionally at CSSL, a short period of heavy winter rain produces a peak discharge far in excess of that normally experienced during the snowmelt runoff season.

2-04.33 Winters at WBSL are warmer than at either UCSL or CSSL and are characterized by heavy snowfall, occasional rainfall, and heavy runoff. Summers are warm and relatively dry. Mean yearly precipitation (tree-top level) for the period of record was 125 inches, more than twice as much as at UCSL. Precipitation is concentrated in the winter season, more than 80 percent of the yearly total falling in the winter (October through March). Summer precipitation is variable, but usually light, and intense convective showers are infrequent. Rain averaged 40 percent of total precipitation during the winter months, ranging from 26 percent (1948-49) to 47 percent (1947-48 and 1950-51). Though much of the winter runoff resulted from rain on snow, considerable runoff was generated by the relatively mild weather conditions associated with the basin's location and elevation range (2000 - 5500' msl). In general, precipitation catch at WBSL gages showed a marked increase with elevation except for gages at windy sites. Variation in snowpack water equivalent was even more marked. In many storms, the freezing level was located well above the basin's lower boundary (2000' msl). Variation in snowpack water equivalent due to initial differences in amount and form (snow or rain) of precipitation was further aggravated by the higher melt rates characteristic in the lower part of the basin. Average basin water equivalent on 1 April was about 25 inches. In two of the four years of record, snowfall in late spring resulted in seasonal maximum water equivalents during May. In marked contrast to UCSL and CSSL, WBSL is characterized by high winter runoff. Nearly two-thirds of the yearly total generated runoff occurred during October through March; only one-third occurred during the spring snowmelt months of April through June. The percentage of yearly runoff occurring in winter ranged from 54 percent (1948-49) to 78 percent (1950-51). Peak flows resulted from rain on snow.

2-04.34 It is pointed out that the above comparisons of hydrologic characteristics are based on water balances derived for the period of record at each laboratory. Cognizance should be taken of the relation between these laboratory period averages and the longer-term normals, as set forth in paragraph 2-04.26.

2-04.35 In order to compare the basin hydrologic character with respect to the time delay to runoff, empirically derived recession curves are presented on plate 2-11 for Skyland Creek, UCSL; Mann Creek and Blue River, WBSL; and Castle Creek, CSSL. Recession curves are shown both in terms of cfs and cfs per square mile. There are also shown the recession coefficients, $C_{\bar{r}}$, and their corresponding values of $t_{\bar{s}}$, for the various ranges in flow for each laboratory. Inspection of the curves shows that Castle Creek, CSSL, has the fastest recession and accordingly the least time of storage delay. Mann Creek and Blue River at WBSL have similar recession characteristics in terms of unit rates of runoff and show a relatively slow recession, particularly for low flows. The recession for Skyland Creek is intermediate between Castle Creek and Blue River, for flows expressed in cfs per square mile. For flows higher than 5 cfs per square mile (equivalent to about 0.2 inch per day), the recession for Mann Creek is nearly identical to that of Skyland Creek.

2-05. REGULAR OBSERVATIONS AND INSTRUMENTATION AT LABORATORIES

2-05.01 General. - The observations of hydrometeorological elements at the snow laboratories were taken with two basic considerations: (1) adequate sampling of the elements with respect to time variations, and (2) adequate sampling of elements with regard to areal variation over the laboratory basins. Determining the sampling requirements of each of the elements with respect to those considerations is a complex problem, and considerable subjective judgment is required in order to arrive at a proper balance between feasibility of measurements with available resources, the variability of the elements with respect to time and space, and the relative importance of the elements in the hydrologic cycle. The amount and type of instrumentation were also chosen to meet the specific problems of snow hydrology as set forth in the initial objective of the program, which are problems of snow accumulation and its relation to the water balance, snowmelt, and the storage and transmission of liquid water in the snowpack. A network of regular observations at predetermined frequency of attendance was established for each laboratory and included both recording and non-recording types of instrumentation. Many of the observations were made in the vicinity of the headquarters living area to insure continuity of record during adverse weather conditions, and in some cases, concentration of observations were made in an area of particular environment, away from the headquarters area. The major portion of the time spent in servicing gages, however, was for the outlying stations which were established for measurement of precipitation, snow depth, water equivalent, wind, air temperature and humidity over the laboratory areas.

2-05.02 Table 2-6 summarizes the number of regular observation stations or points maintained at each laboratory during each of the years of operation. This summary provides a synopsis of the scope of the regular observational program and how it varied with time. The elements are listed in the categories of general weather, radiation, air pressure and wind, temperature and humidity, precipitation, snow, soil, and streamflow. Instrumentation is listed as recording or non-recording. A synopsis of hydrometeorological elements, showing in graphical form the daily progress of each of the elements as measured at key stations, is presented for each laboratory-year of record in its appropriate hydrometeorological log. In addition to the regular observations, there were many special observations taken for a specific purpose of analysis of conditions at a point, but not necessarily continuously with respect to time. These are listed in section 2-07.

2-05.03 Each of the hydrometeorological logs published for each laboratory (see Appendix I) contains an inventory of meteorologic and hydrologic data, which consists of a graphical day-to-day plotting of the actual period of record of all elements observed at the various stations throughout the laboratory. Instrumental characteristics, such as type of instrument, height above ground, number of snow course points, etc., are listed in each log under tabulations entitled, "Status of laboratory observations." A summary of station site characteristics is also published with each log and contains data on the physical conditions of local environment at the site of each gage. Site maps were prepared for the stations and show the topographic features of the sites within a radius of about 200 feet from the gage. Copies of these site maps are included in the hydrometeorological log for 1948 at UCSL, 1950-51 at CSSL, and 1947-48 and 1948-49 at WBSL.

2-05.04 Methods of observation. - In general, the measurements of meteorologic and hydrologic elements were performed in accordance with normal procedures used in data gathering by the U. S. Weather Bureau, U. S. Geological Survey, and other governmental agencies. Some instrumentation, however, was developed for special purposes, as for example the non-selective radiometers for measuring the transfer of radiant energy to and from the snowpack. Modifications of commonly used equipment were made in some cases to adapt them to conditions at the laboratory. There was little precedent for performing routine measurements of the characteristics of the snowpack and the underlying soils; accordingly, many of the methods for measuring them were developed in connection with the laboratory program of the Cooperative Snow Investigations. The quality of data obtained under the regular observational program is discussed in section 2-06.

2-05.05 A discussion of the methods used in obtaining each of the measured elements is contained in the prefacing remark for each of the hydrometeorologic logs, under the title of "Discussion of Tabulated Values." Reference is made to the logs for this information.

2-06. QUALITY OF DATA

2-06.01 General. - A major portion of the effort of the entire Cooperative Snow Investigations program was that of collecting and processing snow laboratory data. Considerable time and expense went into providing adequate instrumentation, but in any measurement involving hydrologic application, variability of the measured element in time and space precludes an exact determination of the quantity on an areal basis. Even at a single point, accurate measurements are often difficult to obtain in the field, and when considering local variability of the element, they may be meaningless. Some of the observations of snow processes at a point were made by precise methods from which quantitative physical relationships could be derived. The majority of the observations, however, were made to evaluate relative variability of the elements and to provide a network of observations much denser than ordinarily available on project basins, thereby leading to a more adequate understanding of hydrologic processes in areas of snow accumulation. These observations cannot be considered to be precise in the sense of laboratory controlled scientific measurements. The emphasis of the observational program was to minimize the controllable errors caused by mechanical deficiencies of the instruments, inadequate frequency of servicing, substandard methods of observation, or untrained personnel.

2-06.02 At the snow laboratories, most of the observational stations were located too far from the headquarters to be serviced at less than weekly intervals. At times, extended storm periods caused delays in the weekly visits. These delays, if prolonged, could cause serious loss in record. Fortunately, they were at a minimum, and in most cases the record could be kept intact through carefully processing the data.

2-06.03 In general, the types of errors introduced into the data were similar at all three laboratories; i.e., observer and instrument errors. Others that were peculiar to individual laboratories, were such things as the effect on recording instruments of the extremely cold temperatures at UCSL or the impounding of water at the interface of the soil and snow which affected the ground water level in one well at CSSL. As much as possible, errors were corrected or compensated for before the data were published. Some of the most common sources of error were those encountered in the recording instruments. They were such items as pen running dry or not being set on the chart; pen clogging; ink blurring or smearing; clock gaining or losing time; clock stopping; time checks missing or illegible; chart being on crooked; trace overlapping due to delay in changing chart; wind vibration or other interference with trace; ice and snow on working mechanism; instrument or ink frozen; chart distorted due to changes in temperature or humidity. Many of these deficiencies were overcome by careful reduction of the data through comparison with nearby instruments.

2-06.04 In the introductory statements in each of the logs, there are references to quality of data, some for that particular water year, and others that obviously apply more broadly. These statements are too numerous to index in this brief section, but, for example, cover such items as the following:

1. Position and accuracy of temperature-sensing elements in the snowpack.

2. Apparent inconsistencies between ground-water stage and nearby streamflow.

3. Limitations in exposure of the reflected pyrheliometer at UCSL for estimating albedo.

4. Apparent inconsistencies in dewpoint data, with respect to air temperature.

5. Below-freezing water temperatures.

This brief section on quality of data supplements the earlier statements which are in the logs rather than including them in a comprehensive manner.

2-06.05 Radiation. - The incident radiation pyrheliometer bulbs were given a good exposure at both CSSL and UCSL, but the reflected radiation bulb at UCSL was poorly exposed, both because the area at which it was aimed was in the shade during early and late portions of the day and because road dust occasionally fell in the area, giving a low bias to the reflected radiation readings. Occasional errors in measurement could be attributed to the collection of frost on the bulb. This resulted in readings being at times too high and at others too low. Generally, this could easily be adjusted for in the tabulations. At times there would be gaps in the record due to power failure; the Micromax recorder being out of balance or running slow; recorder pens being out of ink or clogged; a Micromax chart not changed and running out of paper. It should be pointed out that some instrumental error has been attributed to effects of ambient temperature. 6/ On one or two occasions the pyrheliometers were checked against a standard instrument with "good correspondence."

2-06.06 Air temperature. - Between 10 to 20 percent of the hourly temperature data was lost in the coldest months when the clock mechanism would freeze or fine snow would accumulate on the instruments. The over-all loss for the year was only about 2 or 3 percent. The hygrothermograph record was adjusted to agree with the maximum and minimum readings and the dry bulb temperature checks at the time the charts were changed. Some difficulty was experienced because of non-agreement of the hygrothermograph readings with maximum and minimum thermometer temperature readings, presumably due to separation of the alcohol column in the minimum thermometer. It can be assumed that the temperature values are correct within 2°F. Some possible error might be attributed to reflection from snow up through the louvers or slits in the bottom of the instrument shelter. On rare occasions some fine snow would blow into the shelter and settle on the thermometers and thermograph. The quality of temperature data from thermograph charts is generally not as good as that from direct reading thermometers. Most of the error can be attributed to lack of attendance, a condition that would be impossible to overcome at remote stations but could be controlled at the headquarters site. Temperature measured by Thermohms and liquid thermometers showed a quicker response, indicating a 10 to 20 minute temperature lag in the thermograph reading during rapid changes in temperature. Thermohms and liquid thermometers showed also a greater range (2-4 degrees in maximum and/or minimum readings) than the thermograph.

2-06.07 Humidity. - Humidity was measured by means of the hygrothermograph and the psychrometer. Values of wet and dry bulb temperatures were taken daily at the laboratory headquarters but, at field stations, only at times of changing charts (generally at weekly intervals, except when extreme storm conditions prevented attendance). From the available psychrometric data, the dewpoint was calculated by either conversion tables or psychrometric slide rule.

2-06.08 Errors inherent in the hygrothermograph and the wet bulb thermometer are similar to those already discussed under air temperature (para. 2-06.06). Some error in dewpoint could be attributed to the observer not reading the wet-bulb temperature at its coldest point. Difficulty in checking psychrometric data with hygrograph data was due largely to the different time response between the two instruments as well as to the fact that they were read several feet apart at the field stations. Most comparative data were taken during the morning hours when hygrothermograph temperature and humidity readings lagged as much as one-half to one hour behind readings from the wet and dry bulb thermometers. These factors should be taken into consideration when relating the data to other parameters.

2-06.09 Under certain conditions, the humidity element of the hygrothermograph was not very satisfactory. The hairs would collect moisture and then freeze. Fine snow would drift into the shelter and clog the hairs, later to melt and re-freeze. The slow reaction time made it difficult to calibrate. In reducing the charts it was often necessary, due to the poor quality of the data, to compare the trace from one station with those from other nearby stations with similar elevations and exposures. It was found that, for the most part, the character of the traces was the same and that they agreed for humidities below 70 percent. The greatest disagreement was found in the range from 80 to 100 percent, where corrections up to 20 percent occasionally had to be applied. A tolerance of $\pm$ 5 percent was used for humidity. This would amount to 4 to 6 degrees in corresponding dewpoints when the air temperature was about 20°F and 1 to 2°F where the air temperature exceeded 60°F. This would account, in part, for some of the occasions when the dewpoint at the time of minimum temperature was recorded as being higher than the minimum temperature.

2-06.10 Precipitation. - Perhaps the most intensive observation of any of the meteorological elements, aside from snow surveys, was that of precipitation. Several types of gages were installed and tested under field conditions. A wide variety of exposures, both good or bad, were used in an attempt to determine the variability attributable to differences in exposure. As a consequence, some of the records are quite reliable as indexes whereas others are almost worthless. Much has been written on the subject of precipitation gage exposure (see chapter 3). In this section, the types of gages used in the laboratories will be treated as to their individual deficiencies and sources of error. The problem is twofold: (1) to determine the comparative performance of various types of gages under optimum observational conditions (at headquarters sites where they could be in constant attendance), and (2) to devise and compare the best methods of obtaining snowfall records in the field at long unattended sites and under adverse weather conditions.

2-06.11 Performance of recording gages. - At headquarters sites, the caliber of record of the recording gages was generally high. Operational differences between the standard Friez and Stevens gages were slight, the catch amounts agreeing with a correlation coefficient of 0.98. Some observer preference has been expressed for the greater ease of servicing of the Friez gage during weighout and weigh-in operations. Unfortunately, the small capacity of the standard Friez gage precluded its use at the field stations, where single storms could exceed its effective capacity of approximately 6 inches (or 12 inches using the enlarged Friez gage). For experimental use at outlying field stations, a number of large-capacity Stevens gages were built to CSI specifications, capable of holding 96 inches of liquid (of which approximately half would be anti-freeze charge). Several years observation indicates that, in general, the recording difficulties experienced even in small-capacity, frequently-attended gages, were multiplied considerably in these special gages. In every case where these large-capacity gages were exposed to wind for considerable periods of time, the recorder trace tended to become obscured by the pen vibrations. At times when the wind speed was relatively light, it was not too difficult to follow the pen trace during times of precipitation. But high winds are frequently associated with precipitation, and as the wind speed increased, the trace would become at times as much as $1\frac{1}{2}$ inches wide. In such cases, only a rough estimate of the actual precipitation accumulation could be made. Occasionally, the system of weights and balances became fouled. There was some loss due to leaky valves. Trace variations of 1 inch occasionally occurred without precipitation--probably due to expansion and contraction from temperature and humidity changes. There were some instances when as much as one-half inch of precipitation fell without being recorded. Errors such as these are difficult to account for. Perhaps friction or fine blowing snow accumulating on the interior mechanism retarded the movement. The capacity of the gage made it difficult to read the chart in increments of less than 6 hours or 0.10" precipitation. Between 5 and 10 percent of the record was interpolated. The short-duration records from these gages should not be considered as being better than 20 percent accurate, though the seasonal totals have been corrected for weighout, and have been pro-rated.

2-06.12 Sacramento storage gage. - A comparative study made of the weigh-in of the initial charge versus the amount indicated by the initial stick reading revealed that there was often a discrepancy (up to 25 percent) between the two values. In comparing the weighout versus the stick reading at the end of the season, this discrepancy averaged only 4 percent, with the weighout volume generally greater than that indicated by the stick measurement. This obvious error may be attributable either to poor calibration, deformation of the bottom of the gage from the increased pressure of accumulated precipitation, improper stick measurement or incorrect values of weigh-in or weighout. The seasonal increments, as determined by the stick, were adjusted to agree with the value obtained from the final weighout. This final value can be considered to be of good quality even though there might be some error in the incremental readings. Other sources of measurement error were: failure to read an average depth in the gage; taking readings with ice or slush in the gage. The largest error encountered was from leaky valves and faulty weighouts. There was no way to account for these losses except to indicate their occurrence, which fortunately was very rare.

2-06.13 Precipitation gage catch deficiencies due to turbulence, wind, and capping are known to affect the quality of data of any of the gages but are not considered instrumental or observer errors. The magnitude of these deficiencies are dependent upon gage exposure, frequency of servicing, and meteorological conditions, and they are discussed in detail in chapter 3.

2-06.14 Snow depth and water equivalent. - This element was measured by standard procedures set forth in Snow Surveyors Manual. 5/ Considerable improvement was made from year to year as experience in the techniques of snow surveying was gained. Also, the snow courses were cleared of rocks and brush over the period of years, which, in itself, aided immeasurably in improving the quality of the measurements. As was the case with the precipitation gages, the snow survey courses were placed in a variety of exposures, mainly to test the variability over the area as well as to determine the influences of the various terrain parameters on the course. For this reason, some of the courses are of much better quality than others. This is discussed in a paper by Wilson on snow measurements at the laboratories. 8/ Some errors occur when the measurement is taken. Certain temperatures of the snow and sampling tube cause the snow to freeze to the tube, creating a plug which does not allow a clean cut through the snow. Ice planes in the snow cause similar errors. Ice at the ground surface may not be completely penetrated, giving an incomplete core. Striking a rock or brush buried beneath the snow also results in an erroneous reading. Strong wind puts enough pressure on the tube to cause the weighing scale to indicate too high a value of water equivalent. Many of these errors were detectable by trained observers, but in the earlier years of the program or with new observers they were allowed to enter into the record. It would be difficult to determine exactly the magnitude of error in water equivalent. The error in the mean value obtained for a given snow course most likely does not exceed 5 percent.

2-06.15 Wind speed. - For the most part, the anemometers performed adequately. The most serious source of error was icing due to freezing rain or from wind-driven snow. It was impossible to determine if this condition existed at field stations unless it occurred at the time of observations. It was difficult also to ascertain how long such icing had prevailed. For that reason it was customary to clear the mechanism of ice and accept the reading, at the same time making a note of the possible error. At the headquarters stations where a closer watch could be kept of the instruments, icing conditions could be cleared up very quickly with little loss of record.

2-06.16 Wind direction. - The values reduced from the Esterline-Angus strip-charts are fairly reliable. Hourly wind directions were obtained by summing the total directions for each minute and determining the prevailing direction by standard Weather Bureau procedure.

2-06.17 Air pressure. - The barometric pressure is considered to be one of the most reliable observations taken at the laboratories.

2-06.18 Snow cover. - Visual observations of snow cover were made at intermittent intervals and were subject to considerable error. It was impossible to observe an entire laboratory basin from a single point on the ground. At times, observers in two different parts of the basin would combine observations to get an over-all estimate. These observations are probably good only to the nearest 20 percent of the value given. More extensive observations of snow cover were made by aerial photography. These estimates were generally better at UCSL where the photos were taken at a higher elevation and perpendicular to the ground rather than at an angle as were many at CSSL. Most of the aerial estimates of snow cover are probably within 5 percent of the correct amount.

2-06.19 Snow thermal quality. - The liquid water content of the snow, computed from calorimetric measurements of thermal quality, was sampled unsystematically with respect to time and space, and the data do not indicate basin averages or depth profiles. Sources of error include drifting of the calorimeter constant; errors in the temperatures of the snow, water, and mixture; and heat loss through the calorimeter stoppers. Average error is estimated at 5 percent of the values given for thermal quality, which, of course, corresponds to an error in the neighborhood of 100 percent in values for liquid water content.

2-06.20 Soil moisture. - The calibration of the Bouyoucos blocks, and their care after installation, was not adequate. The blocks frequently disintegrated after a few months in the ground. The rating curves are poorly defined, particularly under wet conditions. Soil temperature data were not taken at the immediate soil-moisture points. The logs contain inconsistencies, and, in general, the soil-moisture data have little or no quantitative value. A more complete discussion of various types of soil moisture measurements at the laboratories is contained in paragraph 2-07.13.

2-06.21 Ground water. - During the winter, between 50 and 80 percent of the data was either questionable or missing. This was generally due to the water freezing in the well, or to ice or snow dams in the nearby stream. During the spring melt period the record was good, for the most part, with loss of record rarely exceeding 5 percent. No areal sampling or ground water profiles were made.

2-06.22 Soil temperature. - The soil temperatures, indicated by telethermoscope readings, are reliable to the nearest whole degree with a tolerance of $\pm$ 1°F. This applies also to those values reduced from the Micromax recorder charts. Occasionally, the surface Thermohm was exposed to the direct rays of the sun at the time of observation. This gave a reading above what would normally be expected. No attempt was made to correct these values and they were included as reported.

2-06.23 Water temperature. - On the whole, these data are reliable to the nearest whole degree.

2-06.24 Snow temperature. - The quality of these data was dependent upon the calibration of the Thermohms and the security of the support on which they were mounted. The weight of the snow accumulation caused the support wires to sag as the season progressed, making it difficult to determine where the Thermohms were at a given time in relation to the ground surface. Therefore, the temperatures reported in the logs give a good representation of the temperature profile through the snow but the accuracy of the height of the Thermohms is not reliable. Some error can be ascribed to the absorption of radiation through the snow by the Thermohms. There were many instances when temperatures of 33° or 34° were reported below the snow surface due to the heating effect of radiation penetrating to the Thermohm. Readings of snow temperature by the telethermoscope were made only to the nearest whole degree. Readings to any finer degree would not be realistic. In general, the values are regarded as reliable.

2-06.25 Streamflow. - The record obtained from CSSL was excellent throughout the year, mainly due to the daily attention paid to the gage in keeping it cleared of ice and snow. The use of the Parshall flume and V-notch weir along with careful gagings by competent personnel makes this record acceptable for the entire period of record. During the winter periods at UCSL, ice formed in the stream beds, making it difficult to maintain an accurate record. However, this occurred during periods of low flow and only a small portion of the record was erroneous. The over-all error probably did not exceed 5 percent, which is considered "excellent" by USGS standards. Skyland Creek record in the winter of 1946-47 was obviously too low, in comparison with runoff in adjacent streams and in other years. At WBSL, inadequate control and river gaging practices put this record in the "fair" to "poor" category. In addition to the deficient rating curves, there was a poor tie-in between the elevation of the outside gage and the chart. Rating curves for Mann Creek and Blue River were better than those for Wolf Creek. The float froze to the walls and to ice planes in the stilling wells, causing loss in record. Logs and debris in stream beds influenced the control to a great extent, making it difficult to apply rating curves with much degree of confidence. The stilling well intakes were rather sluggish, failing to respond to quick changes in streamflow. As in the case of UCSL, the data at the Willamette laboratory were poor during periods of low flow and ice effect, but were acceptable during the spring melt season.

2-06.26 Lysimeter. - Lysimeter data were obtained at CSSL for the Headquarters and Lower Meadow sites. In the early years from 1949 through 1952, definition of the lysimeter boundaries was indefinite during some periods because of ice planes which formed within the snowpack and thereby caused water loss or gain from adjacent areas. By trenching the perimeter during 1953 and 1954, good definition of daily volumes has been obtained. Pertinent information on the two lysimeters constructed at CSSL is given in paragraph 2-07.02.

2-06.27 Site maps. - In general, the topography and the location of structures, well defined obstructions, and observation points, are precise. There are unidentified instances of changes in snow-sampling points, from year to year. The portrayal of vegetative cover was based partly on aerial photographs and partly on ground observations, and possible errors arise from interpretation and generalizing. The actual height and density of surrounding vegetation is for the most part only a rough estimate rather than a true measurement. The effective height of vegetation varied greatly because of the great range of snow depths. There is a wide range of quality and precision of site maps among laboratories.

2-07. SPECIAL OBSERVATIONS

2-07.01 General. - Several categories of measurements in connection with the laboratory program are classified as special observations, in that either (a) records were not continuous, (b) the measurements were designed for a specific analytical project rather than for studies of the basin as a whole, or, (c) special observational techniques were being developed or tested. The following paragraphs summarize the purpose and extent of these observations. Since many of the special observations were not published in the hydrometeorological logs, reference is made specifically as to availability of data.

2-07.02 Lysimeters. - Two impervious snow lysimeters were constructed at CSSL, one having 1300 square feet of area and located near the headquarters, and the other having 600 square feet and located at the Lower Meadow (Station 3). The methods of construction and physical characteristics were described in the appropriate logs for CSSL, and also in Research Notes 17, 18, and 25. The purpose of the lysimeters was to provide data for (1) the travel and storage of liquid water in the snowpack, and (2) accurate determinations of daily snowmelt, unaffected by soil and ground storage, which could be related to meteorological parameters causing melt. The headquarters lysimeter had provision for artificial sprinkling to simulate the effects of rain on snow. Records for the headquarters lysimeter began in the 1949 water year, and for the Lower Meadow lysimeter in the 1952 water year. Much of the data for the years through 1952 is contained in the CSSL logs. Both lysimeters were operated on a part-time basis through 1954, and tabulations and plotting of the data for the Lower Meadow lysimeter during the 1954 melt season are contained in Research Note 25. Data from the lysimeters have been used extensively in determining the effect of the snowpack on rainfall runoff, and in providing means of estimating snowmelt at a point in the open. With the improvements made in observational techniques during the later years of operation, the lysimeters have proved to be invaluable for determining factors affecting snowmelt runoff at a point and form much of the basis for application of methods to basin areas.

2-07.03 A small (2 square-foot) portable lysimeter was constructed at WBSL in 1950 for the purpose of determining condensation or evaporation from the snowpack, as well as melt. Difficulties in separation of natural and artificial effects preclude its use for quantitative evaluation of amounts.

2-07.04 Deep and shallow pit data. - Observations of the character of the snowpack were obtained at CSSL by SIPRE during the water years 1951 through 1953. The observations were obtained by digging pits in the snow at the Lower Meadow site, at time intervals varying from one to two weeks. "Deep pits" were dug to the snow-ground interface, and the snow structure classified throughout the pack depth. Density and temperature profiles were also obtained, showing the vertical variation of these amounts. "Shallow pits" (usually about one to two feet deep) were dug in connection with vehicle traction tests. Data for 1951 and 1952 are published in the CSSL logs for those years, and data for 1953 are presented in graphical form in chapter 8. Miscellaneous observations of the vertical structure of the snowpack were made at other times and at the other laboratories, but they are not sufficiently complete to present the time variation of the snowpack character through the winter season.

2-07.05 Settling meter data. - In connection with the deep pit data at CSSL, SIPRE installed for the 1953 water year, a slide-wire settling meter, for obtaining undisturbed profiles with respect to time of each snow horizon for individual layers of the snowpack. This device was patterned after one constructed by Bader as reported in "Der Schnee und Seine Metamorphose" 1/ and consists of a vertically mounted electrical resistance wire, to which are attached sliding markers which define the position of each layer of snow. Observations of the positions of snow layers were made once a week through the entire period of snow accumulation and melt, but tabulations of these data have not been published. A graphical plotting of the data is shown on plate 8-1.

2-07.06 Micrometeorological data, CSSL. - Late in 1950, SIPRE instrumented four micrometeorological masts at the Lower Meadow, CSSL, for the purpose of obtaining vertical and horizontal gradients of wind, temperature and humidity over the snow, in an open meadow and adjacent forest areas. Each mast extended approximately 50 feet above ground, and there were various fixed and adjustable levels of measurement. The site is described in the CSSL log for 1951-52, and levels of instrumentation are shown in Miscellaneous Report 5 (see Appendix I). Air temperatures were measured by Thermohms equipped with polished metal radiation shields, dewpoints were measured by Foxboro Dewcels, wind speed was measured by three-cup anemometers, and wind direction by a wind vane. All were recorded by continuous strip-type recorders housed in a shelter with thermostatically controlled heat. Only portions of the basic data from this installation have been published. CSSL logs for 1951 and 1952 and Miscellaneous Reports 4, 5, and 6 contain some of the data for selected periods. Comparisons of air temperature measured on the hygrothermograph at station 3 with the temperature data from the Thermohms indicated that the shielding of the Thermohms was probably inadequate for periods of clear weather.

2-07.07 Special observations of air temperature and humidity near the snow surface (up to 4 feet above the snow) were obtained for short periods during active melt periods at WBSL and CSSL, in connection with special observations of point melt and moisture transfer. This was accomplished late in 1952 at WBSL and reported in Research Note 11. The observations at CSSL were taken in 1952, 1953, and 1954; the data for 1954 are summarized in Research Note 25.

2-07.08 Snow-cover determinations. - Special observations of snow cover were performed by aerial photography or ground surveys at each laboratory. Aerial photographs were taken at UCSL for the years 1946 through 1950, and for CSSL for the years 1947, 1948, 1950, and 1952. From 2 to 6 flights were made each year, during the course of the spring melt season. Data from these flights are presented in the appropriate logs. Ground surveys of snow cover were performed at CSSL in 1946, 1947, and 1948, and at WBSL in 1950.

2-07.09 Radioisotope snow gage. - The radioisotope radiotelemetering snow gage was developed under Civil Works Project 170, assigned to the South Pacific Division Office of the Corps of Engineers, for the purpose of providing an unattended measurement of water equivalent from an undistrubed sample of the snowpack, and telemetering the information by radio to a receiving station. A comprehensive report prepared by the South Pacific Division office describes the development of the gage, its technical aspects, and its application to the Kings River Basin, California. 2/ Data collected with this type of gage during its development at CSSL, for the years 1950 through 1952, are published in the appropriate CSSL logs.

2-07.10 Snow crust thickness and temperatures. - Daily observations of snow crust characteristics at an open site near station 3, CSSL, were obtained during the 1954 melt season in connection with the operation of the lysimeters. They were made near sunrise each morning, before melt had begun, as a means of evaluating total nighttime loss of heat from the snowpack. Measurements were made of the total thickness of the refrozen layer and the temperature profile within that layer. Also, the time variation of temperature of the crust during each night was obtained from Thermohms buried in the snow, as a means of providing continuous estimates of the snow surface temperatures through the night. These estimates are contained in Research Note 25.

2-07.11 Atmospheric moisture transfer. - Direct measurements of the transfer of moisture between the snowpack and the atmosphere by periodic weighings of blocks of snow in pans set on the snow surface were made in order to define the amount of evaporation from or condensation on the snowpack in relation to meteorologic variables. Observations of a few hours to several days duration were made at both CSSL and WBSL. Data have been published in Research Notes 11 and 25, for use in connection with evaluation of heat transfer to the snowpack.

2-07.12 Radiation and snowmelt observations in the forest. - Special measurements of radiation in the forest were made at all three laboratories for relatively short periods. Those at UCSL consisted of incident shortwave radiation through varying densities of forest canopy, and cover the period August to November, 1947. These are reported in Research Note 5. Observations of radiation in the forest at CSSL were made by Gier-Dunkle non-selective radiometers of net allwave radiation exchange, as well as total incoming and outgoing radiation, for the period 27 April 1950 through 9 June 1950, at both forested and open sites. Results of these observations are presented in Technical Bulletin 12. At WBSL in July 1952, observations were made of shortwave and allwave radiation exchange over the snow beneath the forest canopy late in the season. Data from these observations are presented in Research Note 12. In connection with the WBSL observations, snowmelt in the forest was measured by snowpack ablation for the purpose of relating melt in the forest to parameters of heat exchange. Results are presented in Research Note 11.

2-07.13 Special soil moisture observations. - Special observations of soil moisture at the snow laboratories were of two types: (1) testing electrical soil-moisture meters, using experimental units installed adjacent to the regular soil-moisture measurement installation at CSSL and at UCSL; and (2) measuring areal variation in soil moisture, using standard Colman soil-moisture meters installed at field stations at UCSL. The principal purpose of the instrument testing was to find a porous material which combined satisfactory dielectric qualities with sufficient durability under field conditions. In addition to several different types of soil-moisture meters, more than a half a dozen different porous materials were tested. (See "Status of Laboratory Observations" in the hydrometeorological logs.) The results of the tests were summarized by Gerdel in Miscellaneous Report 2 and are abstracted in section 4-06. The second group of special observations consisted of readings from Colman soil-moisture units at nine field stations at UCSL in 1949-50 and 1950-51, in cooperation with the U. S. Forest Service. Sites of the field stations were chosen to sample vegetation cover and, so far as operational schedules permitted, areal variation. Each station consisted of Colman moisture units and temperature units at each of five to seven depths below the soil surface: 1, 3, 6, and 12 inches, plus two or three additional depths usually at 12-inch intervals. The observations were made by laboratory personnel; calibration of the units and reduction of the data were performed by Forest Service personnel. Reduced data were not received by Cooperative Snow Investigations in time to be included in the logs. Analysis of the data is expected to form part of the watershed research program of the Intermountain Forest and Range Experiment Station, Ogden, Utah.

2-07.14 Liquid water in snow. - Observations of the thermal quality of snow, which may be used as a measure of the liquid water held in the snowpack, were obtained by calorimetric methods and are listed under regular observations. A special device for determining changes of liquid water in the snowpack in situ was developed at the CSSL, based on the principle of the differences in dielectric constants of ice and water. The instrument is known as a snow-probe capacitor, and a report on its use is contained in the Transactions of the American Geophysical Union. 4/ In general, the instrument has greatest use in detecting time changes with free-water content of snow, but difficulties in calibration preclude its use for quantitative measures.

2-07.15 In addition to measurements of liquid water in the snow, experiments involving the use of fuchsine dye were performed for the purpose of tracing the movement of liquid water through the pack. Miscellaneous experiments of this type were performed at all three laboratories, and a report of early measurements at CSSL is contained in Technical Report 15, Interim Report No. 1. Reports on other experiments of this type have not been published.

2-07.16 Precipitation gage battery. - A battery of precipitation gages of various types was installed at station 1, CSSL, and operated for 2 to 3 years, for the purpose of comparing the efficiencies of the gages and wind shields in areas where precipitation is predominantly in the form of snow. All were placed in an open clearing approximately 100 feet in diameter, and centered about 100 feet northeast of the laboratory headquarters building. The following tabulation describes the gages in this battery:

STATION	TYPE OF GAGE*	WIND SHIELD	MOUNT AND HEIGHT OF ORIFICES	ANTI-FREEZE CHARGE	PERIOD OF OPERATION (WATER YEARS)
1-A	Std. W.B., manual, 24" capacity	no	on snow surface orifice: 2'	no	1946-1952
1-B	Friez recorder 12" capacity	yes	on tower, orifice: 23'	yes	1947-1952
1-C	Stevens recorder 24" capacity	yes	on tower, orifice: 19'	yes	1946-1951
1-D	Enlarged Friez recorder, 24" capacity	yes	on tower, orifice: 24'	yes	1947-1949
1-E	Std. W.B., manual, 24" capacity	no	on tree stump, orifice: 24'	yes	1947-1949
1-F	Std. W.B., manual 24" capacity	yes	on tree stump, orifice: 24'	yes	1947-1949
1-G	Sacramento, Manual, 200" capacity	yes	on tower, orifice: 24'	yes	1947-1949 1952-1954

*All gage orifices are of standard 8" diameter.
Capacities listed are without anti-freeze charge.
Effective operational capacities, with charge, are approx $\frac{1}{2}$ those above.

The location of these gages is shown on the site map for station 1, contained in the 1950-51 log for CSSL. Data have been published in the appropriate logs for only stations 1-B and 1-C. Comparative data for seven gages for selected periods in the 1946-47 water year were published by Wilson in the Monthly Weather Review. 8/

2-07.17 Supplementary snow-course data, CSSL. - In the spring of 1951, 15 supplemental snow courses at CSSL were measured for more adequate sampling of the snowpack with respect to topographic features. Results of these measurements are contained in the 1950-51 log for CSSL and in Research Note 13.

2-07.18 Penetration of solar radiation into the snowpack. - An experiment on the penetration of solar radiation into the snowpack was devised in 1947, utilizing specially constructed pyrheliometers. A report of the instrument and measurements obtained by it is found in Technical Report No. 8, Interim Report No. 1.

2-08. DATA PUBLICATION

2-08.01 General. - Systematic publication of snow laboratory data was accomplished for the bulk of the observations. Reference is made to the individual hydrometeorological logs for listing of elements and periods for which values have been published (see Appendix I). Each log contains an inventory of meteorologic and hydrologic data, which indicates in chart form the status of the processing and publishing of each of the elements observed at the field stations and at headquarters. No further summary is presented herein.

2-08.02 Method of publication. - All published data are presented in the logs in tabular form from previously reduced and verified recorder charts or field notes. The time interval for tabulated values is not the same for all elements nor is it consistent throughout the year; the selection of the time intervals was dependent upon the time variability of the element, the relative importance of the element, the prospective use of the observations in hydrologic analysis, and the availability and quality of record. In some cases, short periods of hydrologic significance (unusual winter rain storms or periods of melt, etc.) were treated more fully than normal periods when there was little change in conditions.

2-08.03 Unpublished data. - All original field notes, recorder charts and data tabulations are preserved in the files of the North Pacific Division Office, U. S. Corps of Engineers. Any inquiries or requests for transcripts of unpublished data should be directed to that office. These data include records for periods which have not been encompassed by the logs. In addition, there were observations for the 1951-52 water year at WBSL and the 1952-53 water year at CSSL, for which records have not been published. Reduction of data for those years is about 50 percent complete, and information on availability of records may be obtained upon request. Original records for the entire laboratory program of the Cooperative Snow Investigations have been microfilmed, so that copies of the original field notes and recorder charts are available on microfilm. Data for approximately 30 percent of the observations have been placed on IBM punch cards for machine mass data analysis. Reference is made to the inventory of meteorologic and hydrologic data contained in each log, for listing of periods for which IBM punch cards are available.

2-09. REFERENCES

1/ BADER, H. and others, "Snow and its metamorphism," SIPRE Translation 14, (transl. by J. C. Van Tienhoven from Der Schnee und Seine Metamorphose, Beitrage zur Geologie der Schweiz, Geotechnische Serie, Hydrologie, Lieferung 3, Bern, 1939, Snow, Ice and Perm. Res. Estab., Corps of Engrs., Wilmette, Ill., May 1954.

2/ CORPS OF ENGINEERS, South Pacific Division, "Development and test performance of radioisotope-radiotelemetering snow-gage equipment," Civil Works Investigation Project CWI-170, May 1955, 74 pp. and appendix.

3/ FLINT, Richard F., "Snow, ice and permafrost in military operations," SIPRE Report 15, Corps of Engineers, September 1953.

4/ GERDEL, R. W., "The transmission of water through snow," Trans. Amer. Geophys. Union, Vol. 35, No. 3, June 1954, pp 475-485.

5/ MARR, J.C., "Snow surveying," Misc. Pub. No. 380, U. S. Dept. of Agriculture, Washington, D. C., June 1940, 45 pp.

6/ McDONALD, T. H., "Some characteristics of the Eppley pyrheliometer," Mon. Wea. Rev., Vol. 79, No. 8, August 1951, pp. 153-159.

7/ U. S. WEATHER BUREAU, "Normal weather charts for the Northern Hemisphere," Technical Paper No. 21, October 1952.

8/ WILSON, W. T., "Analysis of winter precipitation observations in the Cooperative Snow Investigations," Mon. Wea. Rev., Vol. 82 No. 7, July 1954, pp. 183-195.

TABLE 2-1

SUMMARY OF SNOW LABORATORY CHARACTERISTICS

PART I - TOPOGRAPHIC AND ENVIRONMENTAL FEATURES

LABORATORY AREA	DRAINAGE AREA	MEAN LATITUDE	MEAN LONGITUDE	ELEVATION			SLOPE			BASIN ORIENTATION				FOREST COVER			
				Mean	Max.	Min.	Mean	Steepest Quartile	Flattest Quartile	NE	SE	SW	NW	Predominant Type	Forested Area	Mean Canopy Density	Basin Area under Canopy
	(sq.mi.)	°N.	°W.	(ft. msl)			(%)	(%)	(%)	(%)	(%)	(%)	(%)		(%)	(%)	(%)
UCSL (Entire Area)	20.7	48°18'	113 °20'	5700	8605	4480	-	-	-	-	-	-	-		-	-	-
Skyland Creek	8.1	48°17'	113 °20'	5920	7610	4800	32	40	22	27	11	37	25	Lodgepole Pine, Red Fir	90	80	72
WBSL (Entire Area)	11.7	44°18'	122 °10'	3430	5364	1960	40	50	27	15	27	32	26	Douglas Fir, Hemlock, Noble Fir	93	90	84
Mann Creek	5.12	44°19'	122 °10'	3750	5235	2490	36	45	26	13	32	43	12	Douglas Fir, Hemlock, Noble Fir	93	90	84
Wolf Creek	2.07	44°18'	122 °08'	3590	5364	2490	45	59	26	19	16	27	39	Douglas Fir, Hemlock, Noble Fir	88	90	79
CSSL (Entire Area)	3.96	39°22'	120 °22'	7500	9106	6892	21	23	10	11	36	33	20	Lodgepole Pine	40	50	20

PART II - CLIMATIC CHARACTERISTICS

LABORATORY AREA	MEAN BASIN TEMPERATURE*					MEAN BASIN PRECIPITATION *			MEAN W.E. OF TOTAL NEW-FALLEN SNOW, OCTOBER THROUGH MARCH		MEAN BASIN WATER EQUIVALENT	MEAN SOLAR INSOLATION			MEAN WIND AT 700 mb LEVEL							
	Annual	Jan.	Apr.	Jul.	Oct.	Annual	Oct. through March	Apr. through Sept.	Before Interception	After Interception	1 Apr.	Apr.	May	June	JAN. Speed	JAN. Dir.	APRIL Speed	APRIL Dir.	JULY Speed	JULY Dir.	OCT Speed	OCT Dir.
	(°F)	(°F)	(°F)	(°F)	(°F)	(in.)	(in.)	(in.)	(in.)	(in.)	(in.)	(ly/day)			(knots)		(knots)		(knots)		(knots)	
UCSL (Skyland Creek)	34	14	32	56	37	50.5	32.0	18.5	29	23	19	428	489	499	28	290°	15	270°	14	250°	15	270°
WBSL	45	32	41	62	46	122.0	97.0	25.0	46	41	18	-	-	-	24	260°	15	260°	12	240°	14	260°
CSSL	38	23	34	56	41	70.0	58.2	11.8	52	48	40	563	646	706	20	280°	12	270°	7	240°	7	260°

* Temperature and precipitation data adjusted to estimated long term average.

PART III - STREAMFLOW CHARACTERISTICS

LABORATORY AREA	PERIOD OF RECORD	MEAN RUNOFF				EXTREME DISCHARGES				RECESSION DISCHARGES AT VARIOUS FLOWS							
		Annual		Seasonal		Maximum		Minimum		2 cfs/sq.mi.		5 cfs/sq.mi.		10 cfs/sq.mi.		20 cfs/sq.mi.	
		Observed Period of Record	Adjusted to 1921-50 Normal	Oct. through Mar. Period of Record	Apr. through Sep. Period of Record					t_s	C_r	t_s	C_r	t_s	C_r	t_s	C_r
		(in.)	(in.)	(in.)	(in.)	(cfs)	(cfs/sq.mi.)	(cfs)	(cfs/sq.mi.)	(days)	(per day)	(days)	(per day)	(days)	(per day)	(days)	(per day)
UCSL (Entire Area)	10/46-9/51	31.2	26.5	4.68	26.50	696	33.6	5.5	0.26	-	-	-	-	-	-	-	-
Skyland Creek	10/46-9/51	33.3	28.5	4.31	29.02	284	35.0	0.1	0.01	12.0	0.92	4.6	0.80	3.0	0.72	2.3	0.65
WBSL (Entire Area)	11/47-9/51	92.1	77.0	54.95	37.18	1410	122.6	2.1	0.18	40.0	0.975	7.2	0.87	4.5	0.80	2.7	0.69
Mann Creek	12/48-9/52	105.0	93.3	55.43	49.61	585	114.3	1.0	0.20	7.8	0.88	4.2	0.79	2.7	0.69	2.0	0.61
Wolf Creek	12/48-9/52	95.7	84.9	56.00	40.10	124	59.9	0.5	0.24	-	-	-	-	-	-	-	-
CSSL (Entire Area)	2/46-8/51	46.0	42.9	10.11	35.85	1200**	300.0**	0.0	0.00	6.5	0.86	3.2	0.73	1.5	0.52	0.9	0.35

** [illegible]

TABLE 2-2

SNOW LABORATORY PERSONNEL

WATER YEAR	CSSL Laboratory Director	CSSL Number of Employees CE	CSSL Number of Employees WB	UCSL Laboratory Director	UCSL Number of Employees CE	UCSL Number of Employees WB	WBSL Laboratory Director	WBSL Number of Employees CE	WBSL Number of Employees WB
1945-46	A. R. Codd	1	3*	F. L. Rhodes	3*FS	1	-	-	-
1946-47	A. R. Codd R. W. Gerdel B. L. Hansen	3	3*	F. L. Rhodes W. R. Demme	5*	2	-	-	-
1947-48	B. L. Hansen	3	2*	W. R. Demme	5*	2	W.G.Somerville	5*	0
1948-49	B. L. Hansen	3	2*	W. R. Demme	5*	2	W.G.Somerville	5*	0
1949-50	C. W. Mansfield	2	2*	W. R. Demme	5*	2	W.G.Somerville J. Summersett**	6*	0
1950-51	P. Merrill	3*	0	R. K. Brown	2*	0	W.G.Somerville	6*	0
1951-52	W. H. Parrott	5*#	0	-	-	-	W.G.Somerville	5*	0
1952-53	W. H. Parrott	6*#	0	-	-	-	-	-	-
1953-54	P. B. Boyer	3*+	0	-	-	-	-	-	-

* Includes Laboratory Director
** Technical Director
\# Includes approximately 8 man-years SIPRE participation, combined 1951-52 and 1952-53.
\+ For approximately 6 weeks during melt season.
FS Forest Service personnel participated at UCSL during 1945-46.

NOTES: Personnel from Analyses Unit also participated in observational program for orientation and special studies, particularly at CSSL, after the year 1950. The number of employees listed are for the major portion of the water year.

TABLE 2-3

COMPARATIVE HYDROCLIMATIC DATA - UCSL

ITEM	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	TOTAL	MEAN
TEMPERATURE, °F 1/														
Long-term														
Means, 1938-54	14.9	18.5	23.2	34.3	43.3	49.0	56.5	54.6	47.7	39.5	25.0	19.3		35.5
Means, 1947-50*	8.0	17.4	21.7	35.0	43.6	50.0	55.8	55.0	47.2	36.8	26.9	17.9		34.6
Deviation from Long-term Means	-6.9	-1.1	-1.5	0.7	0.3	1.0	-0.7	0.4	-0.5	-2.7	1.9	-1.4		-0.9
PRECIPITATION, IN. 1/														
Long-term														
Means, 1937-54	4.85	3.48	3.31	2.60	2.98	3.93	1.16	1.54	2.52	2.97	3.52	3.97	36.83	
Means, 1947-50*	5.03	4.47	3.69	2.43	2.74	4.30	1.45	1.54	1.97	3.99	5.08	4.67	41.36	
Deviation from Long-term Means	0.18	0.99	0.38	-0.17	-0.24	0.37	0.29	0.00	-0.55	1.02	1.56	0.70	4.53	
SNOWPACK W.E., IN. 2/	1 JAN	1 FEB	1 MAR	1 APR	1 MAY									
Long-term														
Means, 1936-55	7.3	11.7	15.7	18.2	11.6									
Means, 1947-50	10.0	15.4	20.4	23.1	16.3									
Deviation from Long-term Means	2.7	3.7	4.7	4.9	4.7									
RUNOFF, IN. 3/														
Long-term														
Means, 1940-52	0.47	0.43	0.52	3.16	9.53	6.87	2.30	0.71	0.50	0.75	0.74	0.76	26.74	
Means, 1947-50*	0.50	0.43	0.57	2.83	11.67	9.42	2.72	0.80	0.47	0.79	0.77	0.69	31.66	
Deviation from Long-term Means	0.03	0.00	0.05	-0.33	2.14	2.55	0.42	0.09	-0.03	0.04	0.03	-0.07	4.92	

* For period beginning Sep 1946 and ending Aug 1950.
1/ Summit, Montana - Elev. 5213 ft.
2/ Marias Pass " - Elev. 5250 ft.
3/ Middle Fork, Flathead River at Essex, Montana - Drainage Area: 510 sq. mi.

TABLE 2-4

COMPARATIVE HYDROCLIMATIC DATA - CSSL

ITEM	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	TOTAL	MEAN
TEMPERATURE, °F 1/														
Long-term														
Means, 1930-54	24.8	25.8	29.5	36.8	42.9	50.4	58.6	57.0	53.5	44.6	34.8	28.0		40.6
Means, 1947-51*	23.9	26.1	29.3	37.3	43.0	51.5	57.6	56.0	53.2	43.5	35.1	28.0		40.4
Deviation from Long-term Means	-0.9	0.3	-0.2	0.5	0.1	1.1	-1.0	-1.0	-0.3	-1.1	0.3	0.0		-0.2
PRECIPITATION, IN. 1/														
Long-term														
Means, 1930-54	9.24	7.31	7.46	4.07	2.61	1.26	0.30	0.22	0.54	3.49	6.68	8.66	51.84	
Means, 1947-51*	8.84	5.78	8.43	4.83	3.20	0.98	0.04	0.28	0.49	3.99	8.83	7.01	52.70	
Deviation from Long-term Means	-0.40	-1.53	0.97	0.76	0.59	-0.28	-0.26	0.06	-0.05	0.50	2.15	-1.65	0.86	
SNOWPACK, W.E., IN. 1/	1 JAN	1 FEB	1 MAR	1 APR	1 MAY									
Long-term														
Means, 1930-55	-	22.2	31.3	37.1	-									
Means, 1947-51	-	14.8	21.3	30.1	13.5									
Deviation from Long-term Means	-	-7.4	-9.8	-7.0	-									
RUNOFF, IN. 2/														
Long-term														
Means, 1943-52	2.14	20.08	3.21	10.90	17.55	8.00	1.97	0.91	0.48	0.50	2.64	2.95	53.33	
Means, 1947-51*	2.31	1.95	2.73	10.31	15.01	6.75	1.32	0.74	0.60	0.63	4.24	3.53	50.12	
Deviation from Long-term Means	0.17	-0.13	-0.48	-0.59	-2.54	-1.25	-0.65	-0.17	0.12	0.13	1.60	0.58	-3.21	

* For period beginning Sep 1946 and ending Aug 1951.
1/ Soda Springs, California - Elev. 6750 ft.
2/ South Yuba River near Cisco, California - Drainage Area: 50.2 sq. mi.

TABLE 2-5

COMPARATIVE HYDROCLIMATIC DATA - WBSL

ITEM	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	TOTAL	MEAN
TEMPERATURE, °F 1/														
Long-term														
Means, 1934-54	39.6	43.1	46.3	52.0	57.1	61.3	66.9	66.3	62.7	54.1	45.7	41.5		53.0
Means, 1948-51*	35.4	40.8	43.7	50.8	56.2	63.0	66.4	66.0	62.1	52.3	46.4	41.2		52.0
Deviation from Long-term Mean	-4.2	-2.3	-2.6	-1.2	-0.9	1.7	-0.5	-0.3	-0.6	-1.8	0.7	-0.3		-1.0
PRECIPITATION, IN. 1/														
Long-term														
Means, 1934-54	9.40	7.73	6.90	4.43	3.61	2.70	0.64	0.89	2.09	7.00	9.11	9.81	64.31	
Means, 1948-51*	10.45	8.89	7.46	3.58	4.01	1.56	0.50	1.16	2.41	9.13	9.50	9.08	67.73	
Deviation from Long-term Means	1.05	1.16	0.56	-0.85	0.40	-1.14	-0.14	0.27	0.32	2.13	0.39	-0.73	3.42	
SNOWPACK W.E., IN. 2/	1 JAN	1 FEB	1 MAR	1 APR	1 MAY									
Long-term														
Means, 1941-55	9.7	18.8	23.3	26.4	-									
Means, 1948-51	12.0	24.0	31.5	35.7	-									
Deviation from Long-term Means	2.3	5.2	8.2	9.3	-									
RUNOFF, IN. 3/														
Long-term														
Means, 1936-52	9.37	9.14	8.75	9.16	6.85	3.23	1.05	0.50	0.49	2.58	7.51	10.20	68.83	
Means, 1948-51*	10.47	12.46	10.03	10.51	10.27	4.05	1.13	0.54	0.55	5.40	8.92	9.76	84.09	
Deviation from Long-term Means	1.10	3.32	1.28	1.35	3.42	0.82	0.08	0.04	0.06	2.82	1.41	-0.44	15.26	

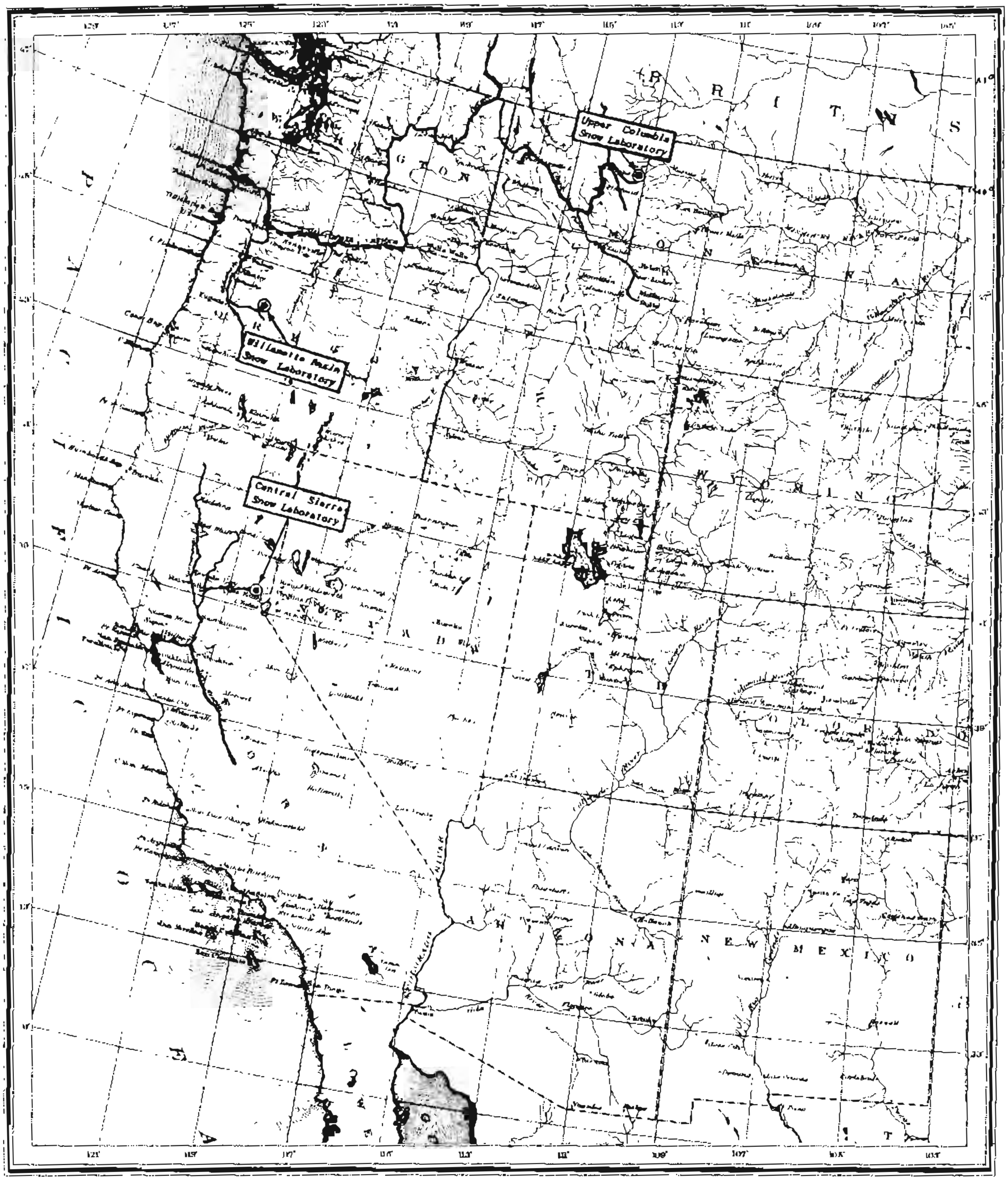
* For period beginning Sep 1947 and ending Aug 1951.
1/ Leaburg, Oregon - Elev. 675 ft.
2/ Santiam Junction, Oregon - Elev. 3780 ft.
3/ Blue River near Blue River, Oregon - Drainage Area: 75 sq. mi.

TABLE 2-6

SUMMARY OF REGULAR OBSERVATIONS AT SNOW LABORATORIES

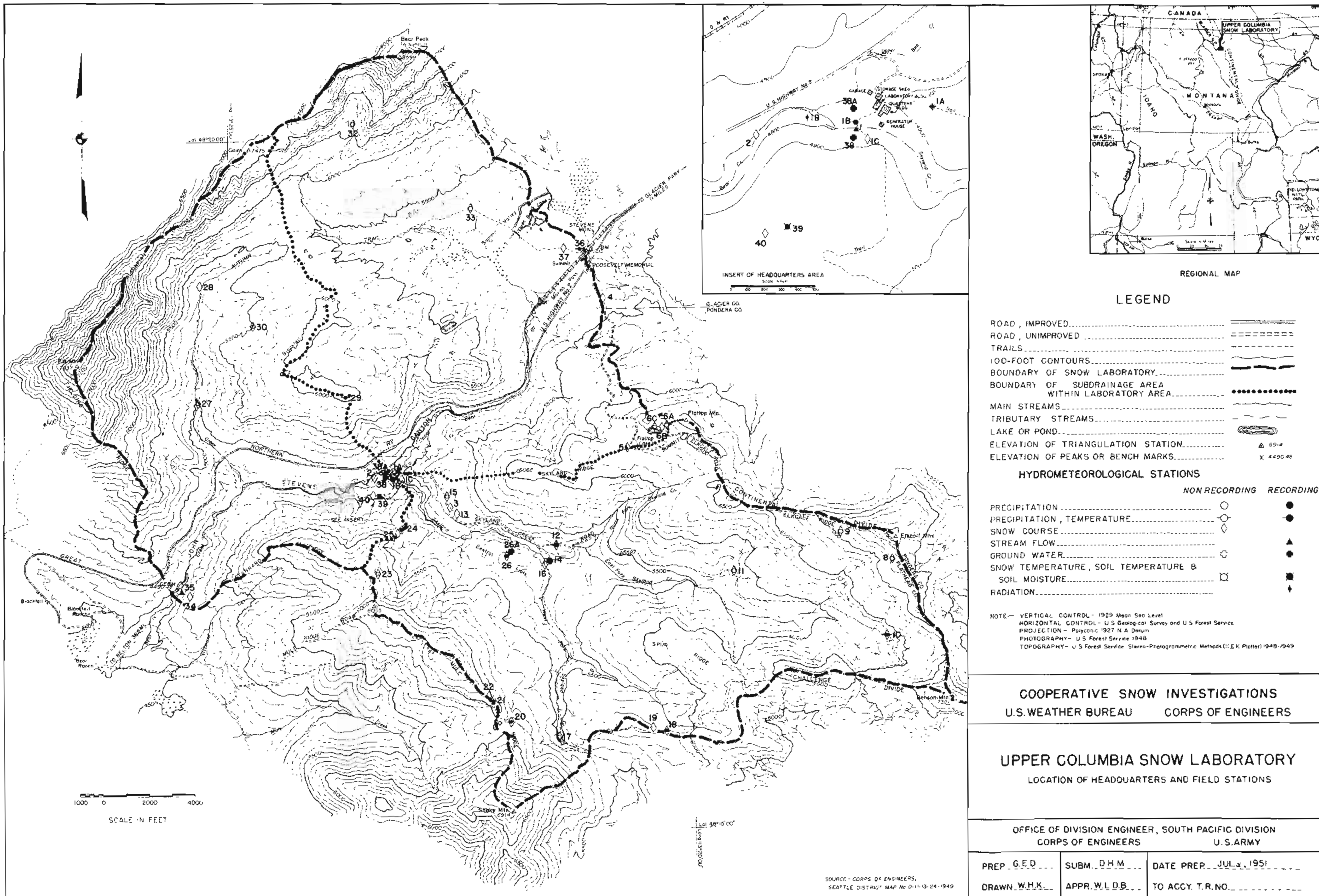
HYDROMETEOROLOGICAL ELEMENT	UNITS	NUMBER OF UNITS OPERATED EACH YEAR																					
		CENTRAL SIERRA SNOW LABORATORY									UPPER COLUMBIA SNOW LABORATORY						WILLAMETTE BASIN SNOW LABORATORY				
		1945-46	46-47	47-48	48-49	49-50	50-51	51-52	52-53	53-54	1945-46	46-47	47-48	48-49	49-50	50-51	1947-48	48-49	49-50	50-51	51-5
GENERAL WEATHER																					
Cloudiness, state of weather, wind direction (visual)	Station	1	1	1	1	1	1	1	1	2P	1	1	1	1	1	1	1P	1	1	1&1P	2
RADIATION																					
Sunshine duration (mercury switch w/recorder)	Station	1	1	1P	-	-	-	-	-	-	1	1	1	-	-	-	-	-	-	-	-
Shortwave incident in open (pyrheliometer w/recorder)	Station	1	1	1	1	1	1	1	1	1P	1	1	1	1	1	1	-	-	-	-	-
Shortwave reflected from snow (pyrheliometer w/recorder)	Station	1	1	1	1	1	1	1	1	1P	1	1	1	1	1	-	-	-	-	-	8d
Longwave net (radiometer w/recorder)	Station days	-	-	-	-	36d*	-	92d	300d	40d	-	-	-	-	-	-	-	-	-	-	8d
Longwave total hemispheric (radiometer w/recorder)	Station days	-	-	-	-	36d*	-	92d	300d	40d	-	-	-	-	-	-	-	-	-	-	8d
AIR PRESSURE AND WIND																					
Pressure (barograph without barometer)	Station	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	1P	1	1	1	-
Pressure (barograph checked by barometer)	Station	1	1	1	1	1	1	1	1	-	1	1	1	1	1	1P	-	-	-	-	-
Wind direction (vane w/recorder)	Station	1	1	1&1P	1P	-	1	1	1	1P	1	1	1	1	1	1	-	-	-	-	-
Wind movement (anemometer totalizer)	Station	-	6	7	6	7	5	2	2	1P	5	5	5	5	5	3	-	-	-	2	2&2P
Wind speed (anemometer w/recorder)	Station	3	3	3	2	-	14P	6&6P	11	5P	1	2	2	2	2	2	-	-	-	-	-
TEMPERATURE AND HUMIDITY																					
Air temperature (thermograph w/checks)	Station	6P	6&3P	9	8&2P	7	5	2	2	2P	7P	7&2P	9	9&1P	9	4	-	1&4P	5	5	5
Air temperature profile (Thermohms w/recorder unless marked-N)	Station	1P	1	1	-	-	4	2&2P	2&1P	1P&1PN	-	1P	1	1	1	1N	-	-	-	-	-
Humidity (hydrograph w/checks)	Station	6P	6&3P	9	8&2P	7	5	1&5P	2	2P	6P	6&2P	8	7&2P	8	2&1P	-	4P	4&1P	5	5
PRECIPITATION																					
Precipitation (standard 8" U.S.W.B. gage)	Station	1P	1	1	1	1	1	1	1	2P	2	2	2	2	2	1	-	1	1	1	1
Precipitation (storage gage-N)	Station	-	14	14	16	12	12	1	4	1	12	12	12	12	12	5	7	7	11	10&1P	13
Precipitation (recording weighing-type gage)	Station	1	4	4	4&4P	7	6	2	2	1&1P	8	8	8	8	8	4	1	1&1P	5	5	4
SNOW																					
Snowfall (snowboard or snowboard battery)	Station	1	1	1	1	1	1	1	1	1P	1	1	1	1	1	1	-	-	1P	1&1P	2
Water equivalent of new snow	Station	-	-	-	-	1	3	3P	1	1P	-	-	-	-	-	-	-	-	-	-	-
Snow depth(snow stake read daily)	Station	1	1	1	1	1	1	1	2	2P	2	2	2	2	2	2	1P	1	1	1	1
Snow cover(aerial photograph set)	Flights	-	11	4	8	7	2	6	-	-	3	5	5	5	5	-	-	-	-	-	-
Snow depth, water equivalent, density (snow course)	Courses	18	22	22	21	23	41	3&15P	6	2&2P	21	28	28	28	29	6	10	11	19	18	13
Temperature profile (Thermohm bridge-N)	Sets of determinations	-	88	22	26	28	-	5	-	-	-	-	-	-	-	1	-	-	14	64	106
Temperature profile (Thermohm w/recorder)	Station	1P	-	1	1	-	-	-	S	1CP	-	-	1	1	1	-	-	-	-	-	-
SOIL																					
Temperature profile (Thermohm bridge-N)	Sets of determinations	19	191&47	54	52	111&48	-	-	-	-	-	-	-	-	-	191	-	-	14	64	106
Temperature profile (Thermohm w/recorder)	Station	-	-	1	1	-	1	1	1	-	-	1P	1	1	-	-	-	-	-	-	-
Moisture profile (porous blocks w/resistance bridge-N)	Sets of determinations	19	216&48	183&53	128&51	120&49	-	-	-	-	-	5	69	83	70	8	-	-	-	-	-
Ground water (well w/float and tape-N)	Wells	-	-	-	1	-	-	-	-	-	-	-	-	-	-	-	-	-	-	2	4
Ground water (float activating a recorder)	Wells	-	1	2	1&3P	2&1P	1&1P	-	-	-	-	-	1	3	4	2	-	-	2	2	1
STREAMFLOW																					
Water stage (staff only)	Station	-	-	-	-	-	-	1P	1P	-	1	-	-	-	-	-	-	-	-	-	2P
Water stage (recorder w/staff for checks)	Station	1	1&2P	1&1P	1	1	1	1	1	2P	1	2	2	2	2	2	1	3	3	3	3
Water temperature (Thermohm w/recorder)	Station	-	1	1	1	1	-	-	-	-	-	1	1	1	1	1N	-	-	-	-	-

N - Non-recorder
P - Partial year's record
m - 1954 melt season only (12 April - 22 May)
d - Days of record
S - From snow pits only
* - Instantaneous readings - not continuous record
C - In crust zone



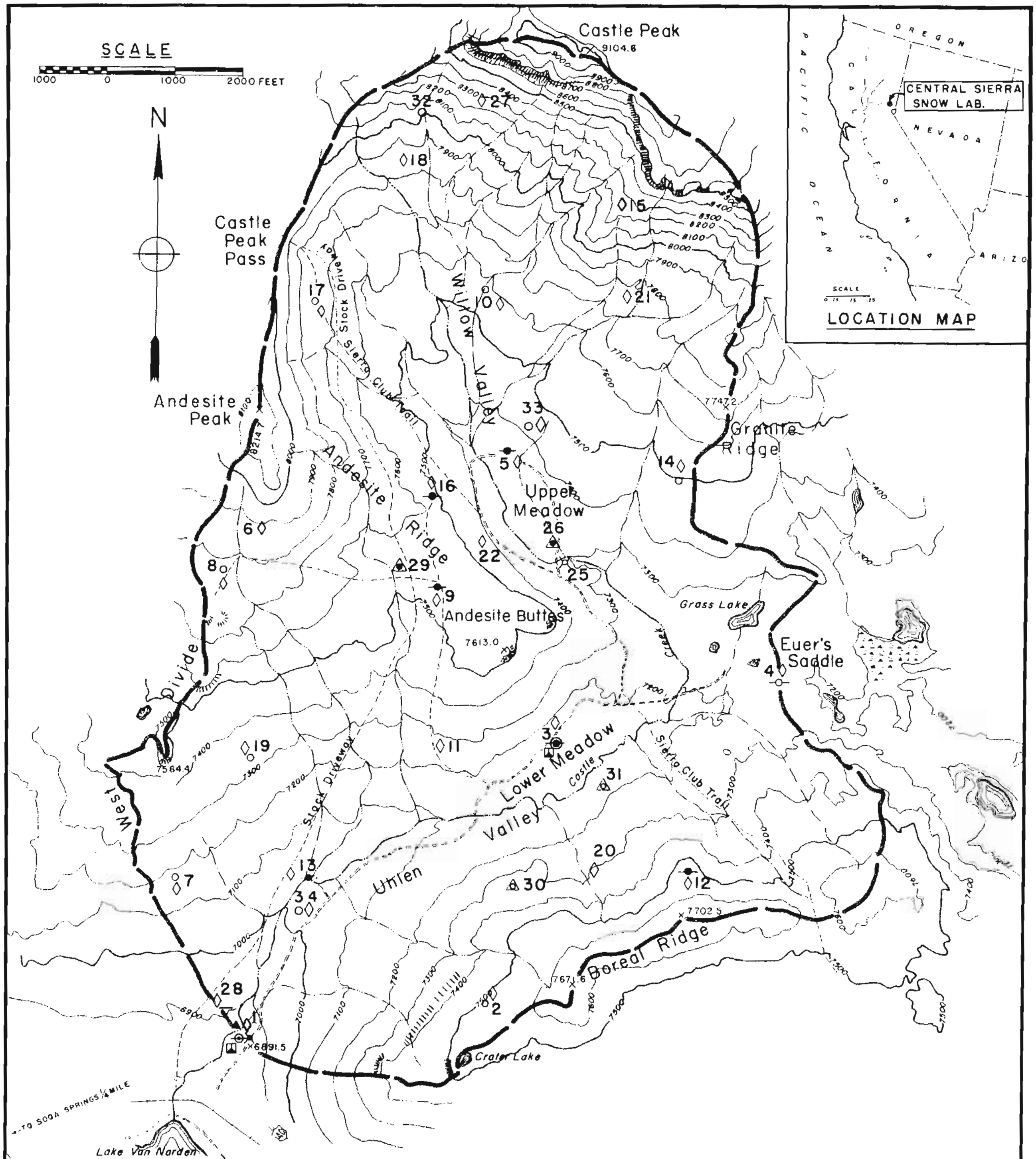
MAP OF WESTERN UNITED STATES
SHOWING LOCATION OF SNOW LABORATORIES
OF THE
WEATHER BUREAU — CORPS OF ENGINEERS COOPERATIVE SNOW INVESTIGATIONS

PLATE 2-1



INSERT OF HEADQUARTERS AREA
REGIONAL MAP
CANADA
UPPER COLUMBIA SNOW LABORATORY
MONTANA
IDAHO
WASH.
OREGON
WYO.
LEGEND
ROAD, IMPROVED
ROAD, UNIMPROVED
TRAILS
100-FOOT CONTOURS
BOUNDARY OF SNOW LABORATORY
BOUNDARY OF SUBDRAINAGE AREA WITHIN LABORATORY AREA
MAIN STREAMS
TRIBUTARY STREAMS
LAKE OR POND
ELEVATION OF TRIANGULATION STATION
ELEVATION OF PEAKS OR BENCH MARKS
HYDROMETEOROLOGICAL STATIONS
NON RECORDING
RECORDING
PRECIPITATION
PRECIPITATION, TEMPERATURE
SNOW COURSE
STREAM FLOW
GROUND WATER
SNOW TEMPERATURE, SOIL TEMPERATURE & SOIL MOISTURE
RADIATION
NOTE— VERTICAL CONTROL - 1929 Mean Sea Level
HORIZONTAL CONTROL - U.S. Geological Survey and U.S. Forest Service
PROJECTION - Polyconic 1927 N.A. Datum
PHOTOGRAPHY - U.S. Forest Service 1948
TOPOGRAPHY - U.S. Forest Service Stereo-Photogrammetric Methods (K.E.K. Plotter) 1948-1949
COOPERATIVE SNOW INVESTIGATIONS
U.S. WEATHER BUREAU
CORPS OF ENGINEERS
UPPER COLUMBIA SNOW LABORATORY
LOCATION OF HEADQUARTERS AND FIELD STATIONS
OFFICE OF DIVISION ENGINEER, SOUTH PACIFIC DIVISION
CORPS OF ENGINEERS
U.S. ARMY
PREP. G.E.D.
SUBM. D.H.M.
DATE PREP. JUL. x, 1951
DRAWN W.H.K.
APPR. W.L.D.B.
TO ACCY. T.R. NO.
SOURCE - CORPS OF ENGINEERS, SEATTLE DISTRICT MAP No. D-11-13-24-1949
SCALE IN FEET
1000
0
2000
4000
U.S. HIGHWAY No. 2
GREAT NORTHERN RY.
ROOSEVELT MEMORIAL
SKYLAND RIDGE
CONTINENTAL DIVIDE
CHALLENGE
SPUR RIDGE
GLACIER CO.
PONDERA CO.
Bear Peak
Snow Mtn.
Elkcalf Mtn.
Flattop Mtn.
Skyland
Summit
Lat. 48°20'00"
Lat. 48°15'00"

PLATE 2-2



LEGEND

MEASUREMENTS TAKEN	NON RECORDING	RECORDING
PRECIPITATION	○	●
TEMPERATURE		—
PRECIPITATION AND TEMPERATURE	-○-	-●-
PRECIPITATION, TEMPERATURE & RADIATION		-⊙-
SNOW COURSE	◊	
STREAMFLOW	△	▲
GROUND WATER WELLS		
SNOW TEMPERATURES, SOIL TEMPERATURES, AND SOIL MOISTURE		
LYSIMETER, TEMPERATURE AND WIND PROFILES		

RIMROCK AND KNOBS
U.S. HIGHWAY NO. 40
ROAD, UNPAVED
TRAILS
100 FOOT CONTOURS
BOUNDARY OF BASIN
STREAMS
LAKES
MARSH
SOUTHERN PACIFIC RAILROAD
ELEVATION OF TRIANGULATION STATION × 6891.5

LOCATION OF STATIONS FROM JAN. 1946 TO JAN. 1952

CENTRAL SIERRA SNOW LABORATORY
SHOWING LOCATION OF HEADQUARTERS AND FIELD STATIONS
COOPERATIVE SNOW INVESTIGATIONS
JANUARY 2, 1952

PLATE 2 - 3

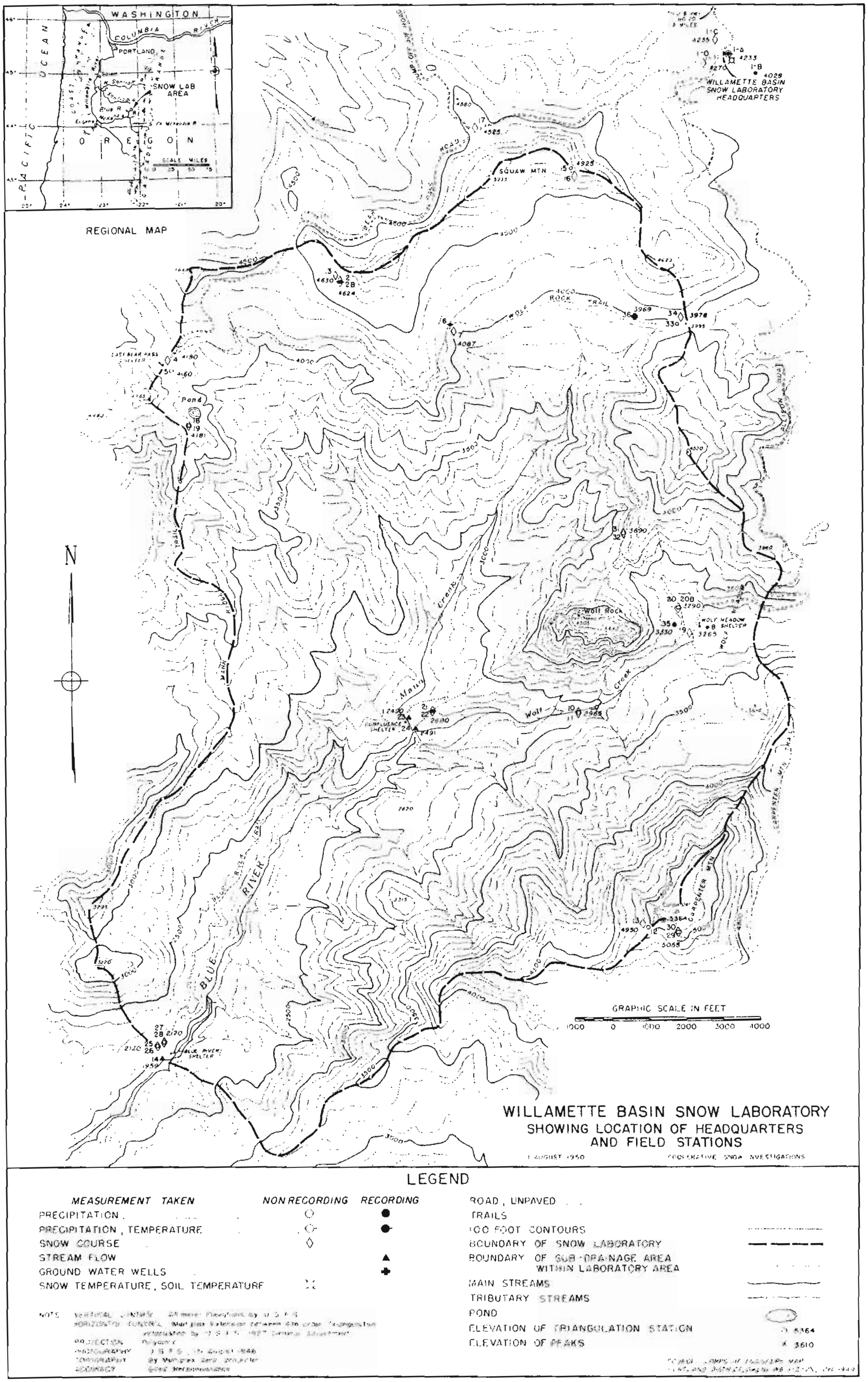
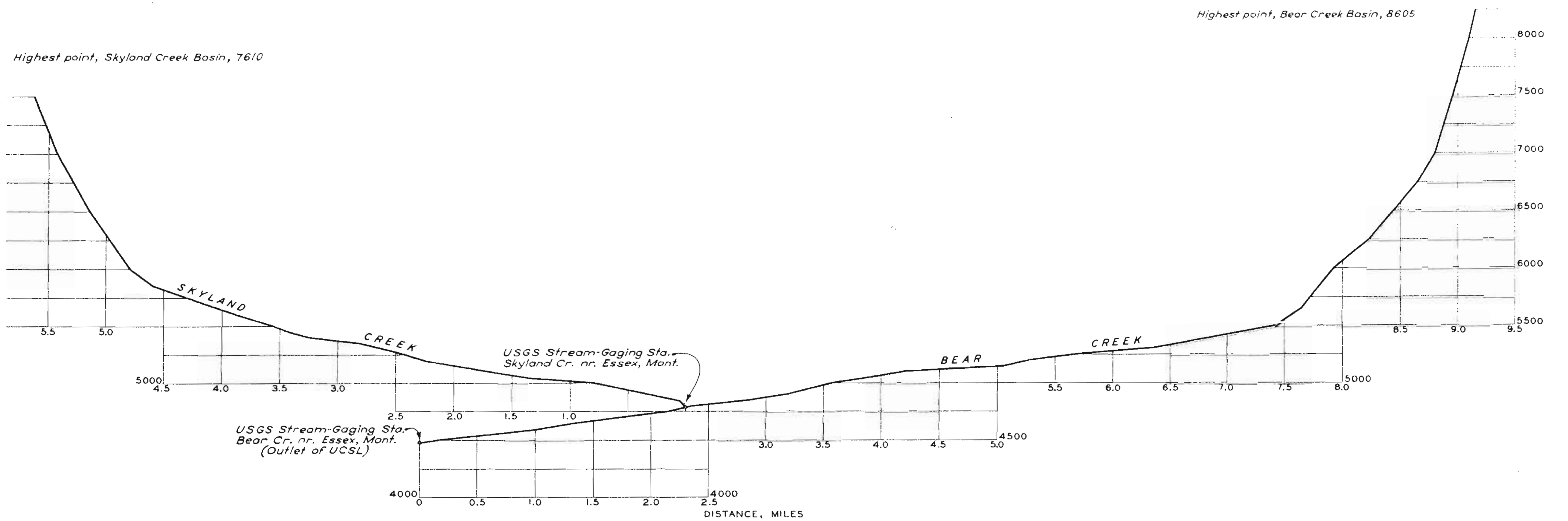
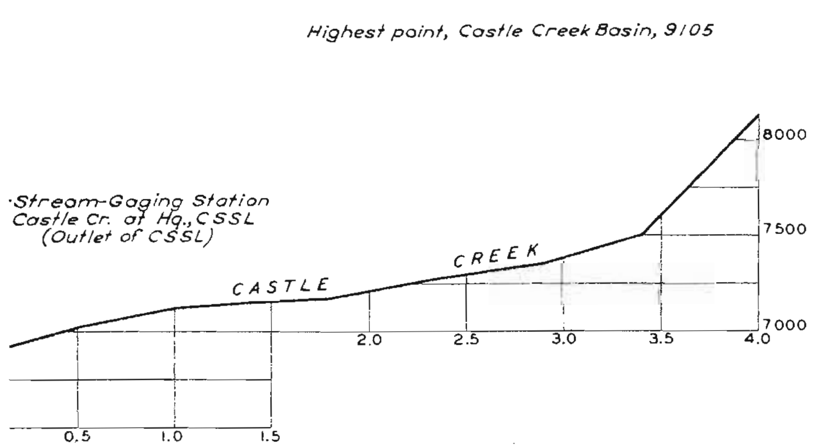

WASHINGTON
COLUMBIA RIVER
PORTLAND
SNOW LAB AREA
OREGON
PACIFIC OCEAN
SCALE MILES
REGIONAL MAP
WILLAMETTE BASIN SNOW LABORATORY HEADQUARTERS
SQUAW MTN
ROCK TRAIL
Pond
Wolf Rock
WOLF MEADOW SHELTER
CONFLUENCE SHELTER
Wolf Creek
Mann Creek
BLUE RIVER
BLUE RIVER SHELTER
CARPENTER MTN
N
GRAPHIC SCALE IN FEET
1000 0 1000 2000 3000 4000
WILLAMETTE BASIN SNOW LABORATORY
SHOWING LOCATION OF HEADQUARTERS
AND FIELD STATIONS
1 AUGUST 1950
COOPERATIVE SNOW INVESTIGATIONS
LEGEND
MEASUREMENT TAKEN
NONRECORDING
RECORDING
PRECIPITATION
PRECIPITATION, TEMPERATURE
SNOW COURSE
STREAM FLOW
GROUND WATER WELLS
SNOW TEMPERATURE, SOIL TEMPERATURE
ROAD, UNPAVED
TRAILS
100 FOOT CONTOURS
BOUNDARY OF SNOW LABORATORY
BOUNDARY OF SUB-DRAINAGE AREA WITHIN LABORATORY AREA
MAIN STREAMS
TRIBUTARY STREAMS
POND
ELEVATION OF TRIANGULATION STATION
ELEVATION OF PEAKS

PLATE 2-4



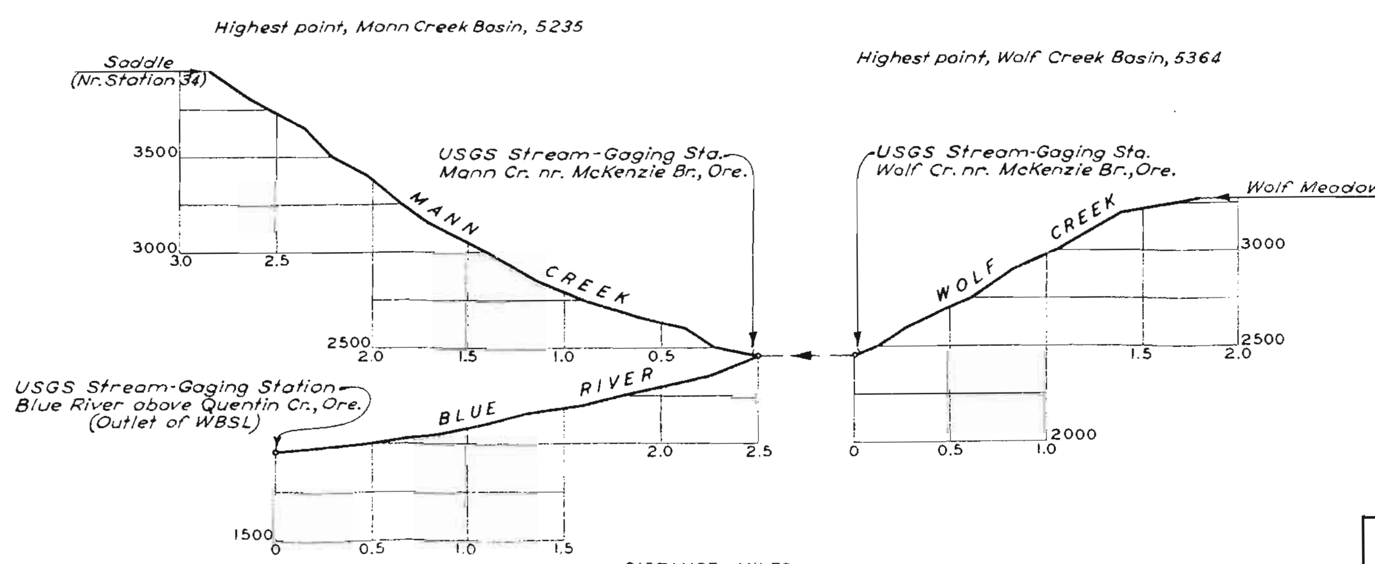
UPPER COLUMBIA SNOW LABORATORY

FIGURE 1



CENTRAL SIERRA SNOW LABORATORY

FIGURE 2



WILLAMETTE BASIN SNOW LABORATORY

FIGURE 3

SNOW INVESTIGATIONS
SUMMARY REPORT
SNOW HYDROLOGY

SNOW LABORATORY
STREAM PROFILES

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

PREPARED: M.W. SUBMITTED: P.B.B. TO ACCOMPANY REPORT DATED 30 JUNE 1956
DRAWN: D.V. APPROVED: D.M.R.

PD-20-25/6

PLATE 2-5

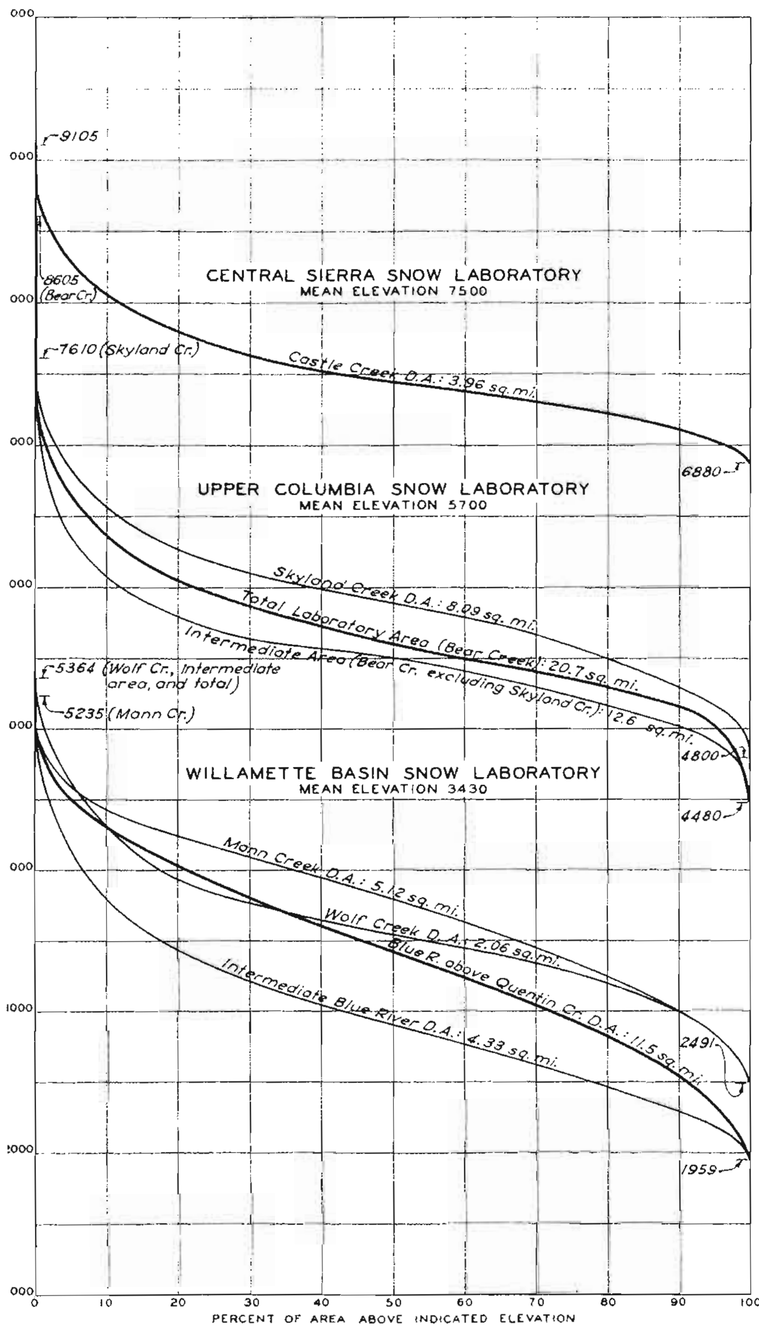


FIGURE 1 — AREA-ELEVATION CURVES

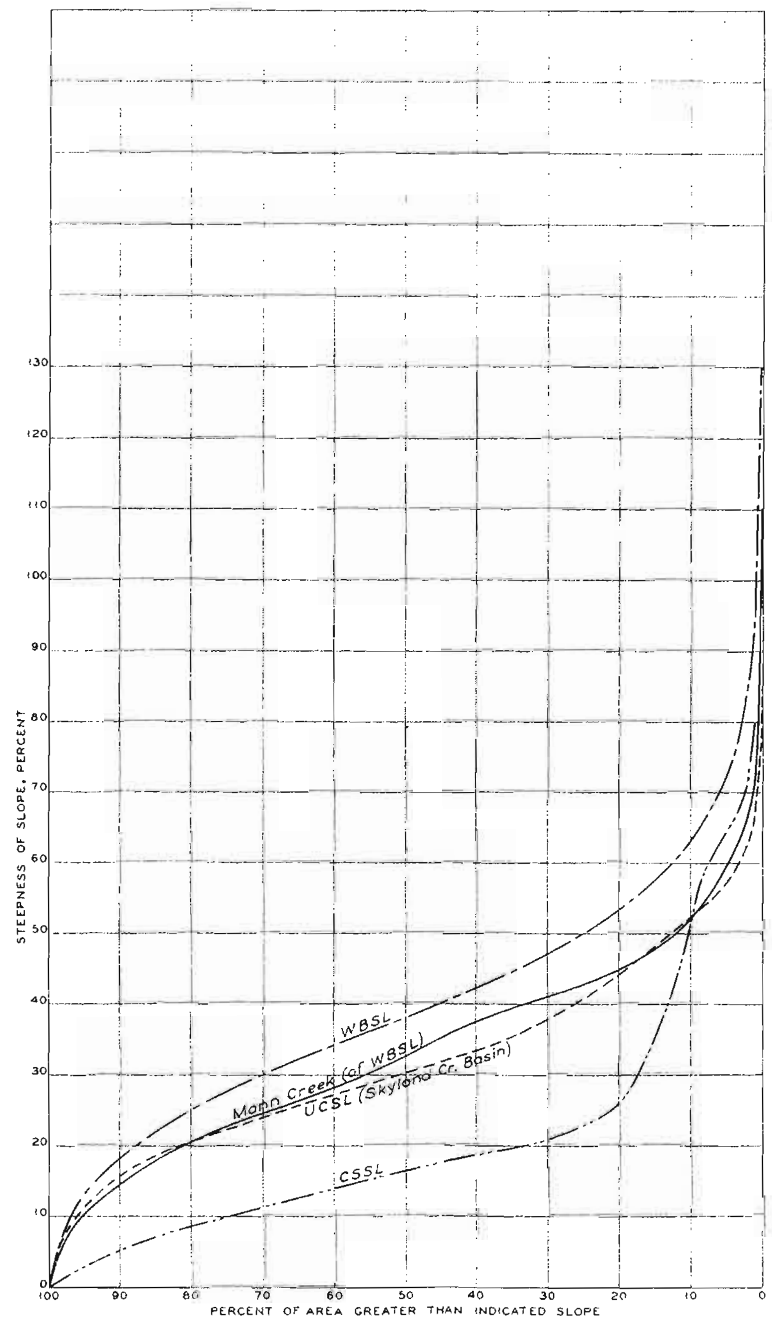
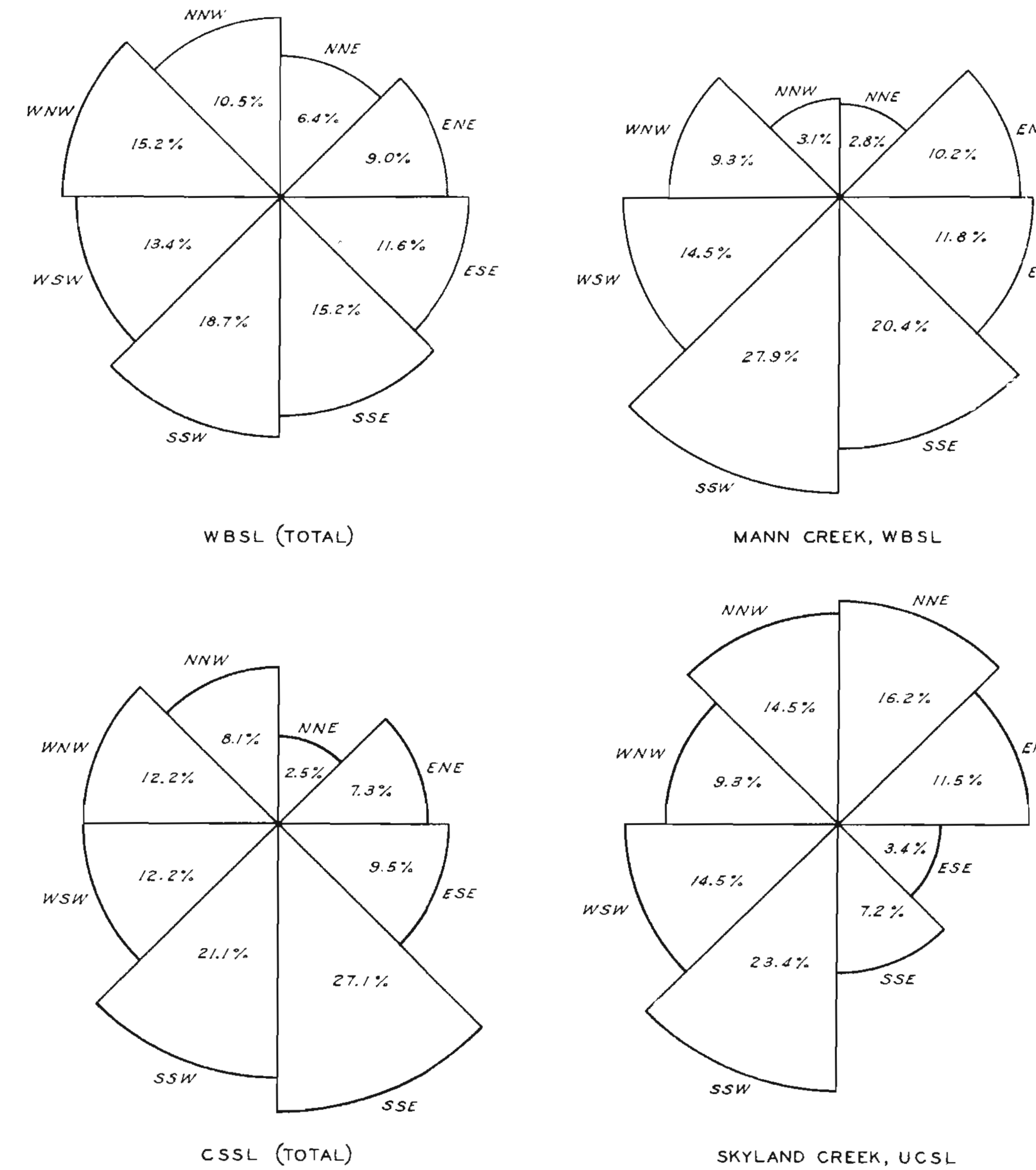


FIGURE 2 — AREA-SLOPE CURVES



Notes:
1. Values shown for each octant represent percent of basin area facing in that direction.
2. Area of each octant is directly proportional to percent of basin area facing in that direction.

FIGURE 3 — BASIN ORIENTATION

SNOW INVESTIGATIONS
SUMMARY REPORT
SNOW, HYDROLOGY

SNOW LABORATORY
BASIN CHARACTERISTICS

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U.S. ARMY

TO ACCOMPANY REPORT DATED 30 JUNE 1956

PD-20-25/7

PLATE 2-6

APPROXIMATE SCALE IN FEET
1000 0 1000 2000 3000 4000 5000 6000

ORIGINAL MOSAIC FROM SEATTLE DISTRICT,
CORPS OF ENGINEERS NO. D-6-1-61
DATE OF PHOTOGRAPH—12 JUNE 1946

SNOW INVESTIGATIONS
SUMMARY REPORT

SNOW HYDROLOGY

AERIAL MOSAIC
UCSL

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U.S. ARMY

PREPARED T.H.P.	SUBMITTED P.B.B.	TO ACCOMPANY REPORT DATED 30 JUNE 1956
DRAWN W.V.K.	APPROVED G.M.R.	PD-20-25/8

PLATE 2-7

CORPS OF ENGINEERS U.S. ARMY

APPROXIMATE SCALE IN FEET

1000 0 1000 2000 3000

SNOW INVESTIGATIONS
SUMMARY REPORT

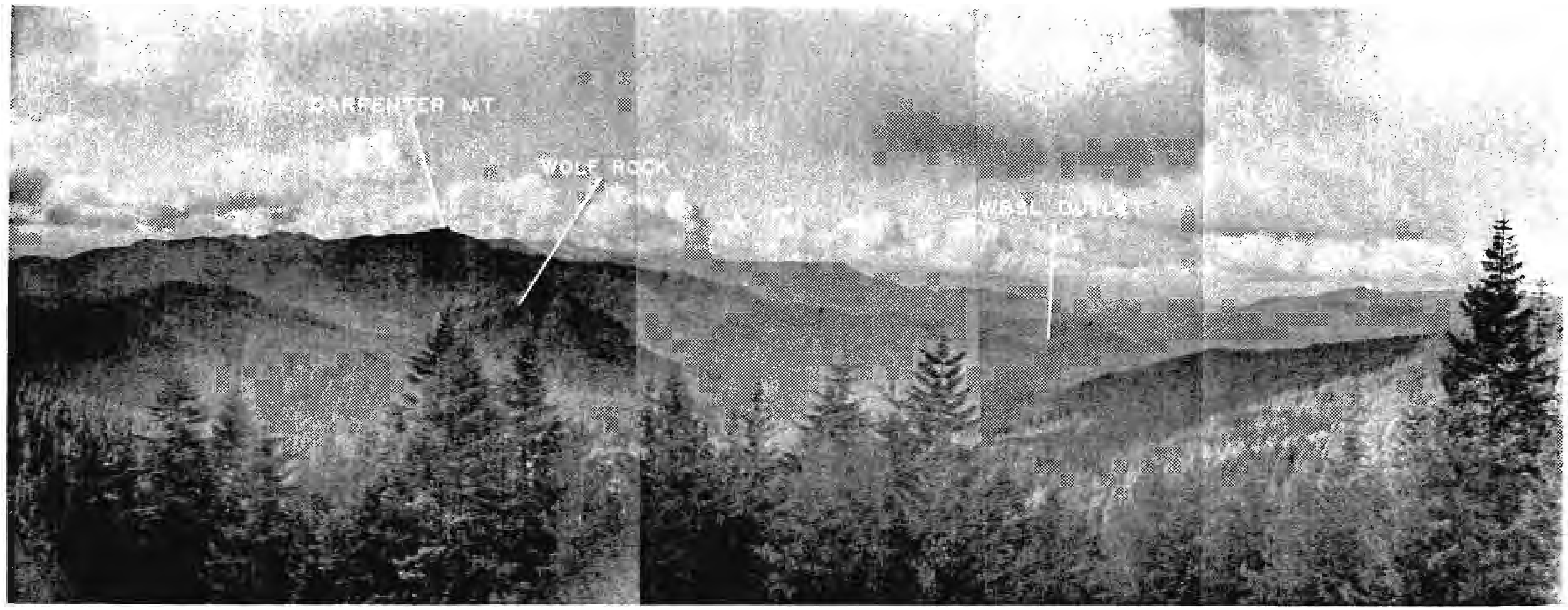
SNOW HYDROLOGY

AERIAL MOSAIC
CSSL

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

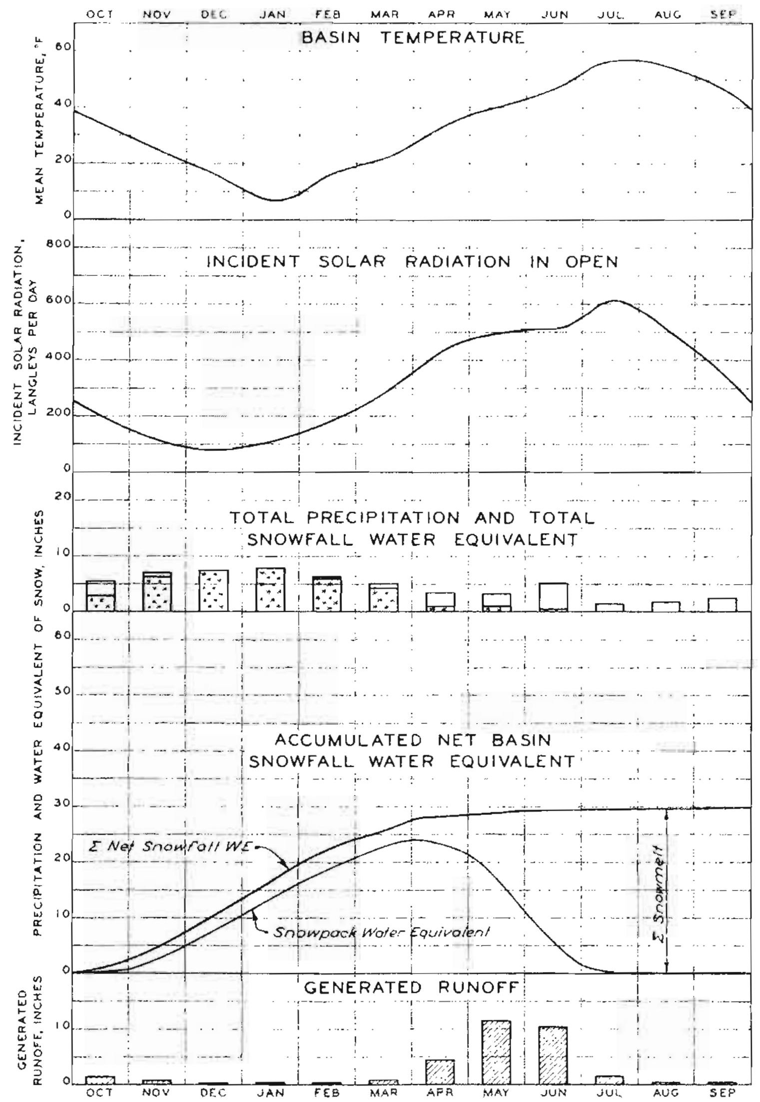
PREPARED CEJ.	SUBMITTED FDM	TO ACCOMPANY REPORT DATED 30 JUNE 1956
DRAWN MY	APPROVED JMR	PD-20-25/9

PLATE 2-8

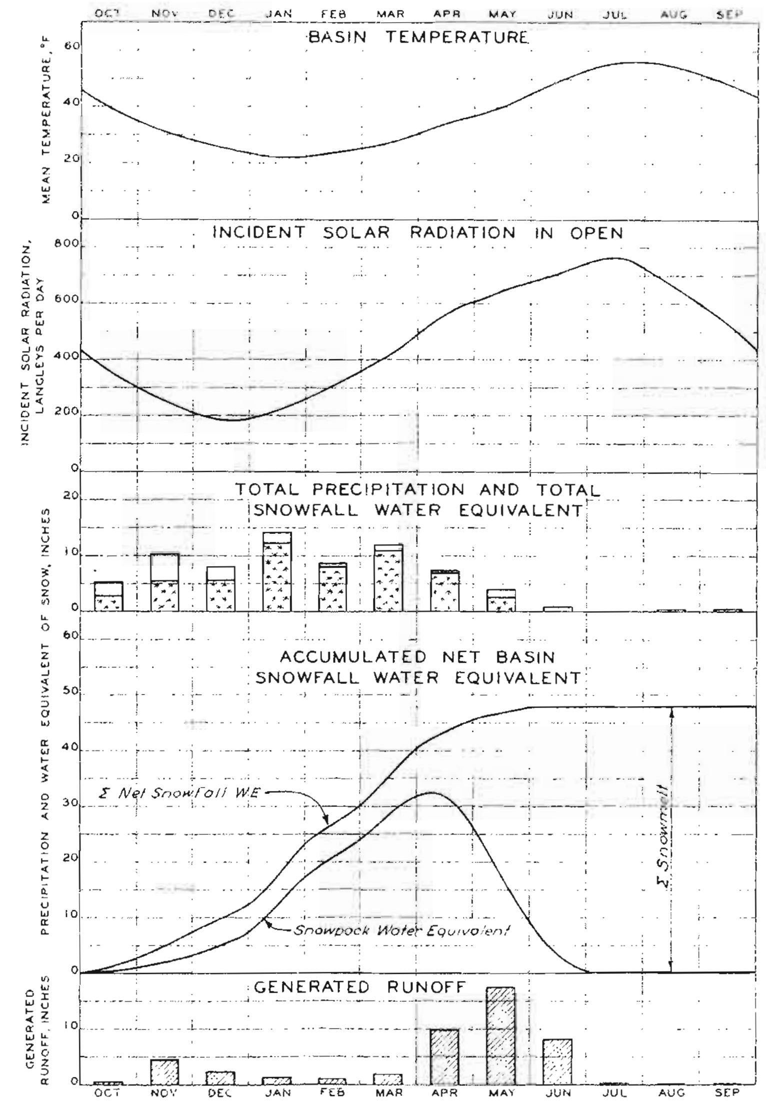


Note:
View taken from Squaw Mountain, looking south over Mann Creek and intermediate Blue River drainage areas. Wolf Creek drainage area lies between Wolf Rock and Carpenter Mountain. Photographs by R. W. Gerdel.

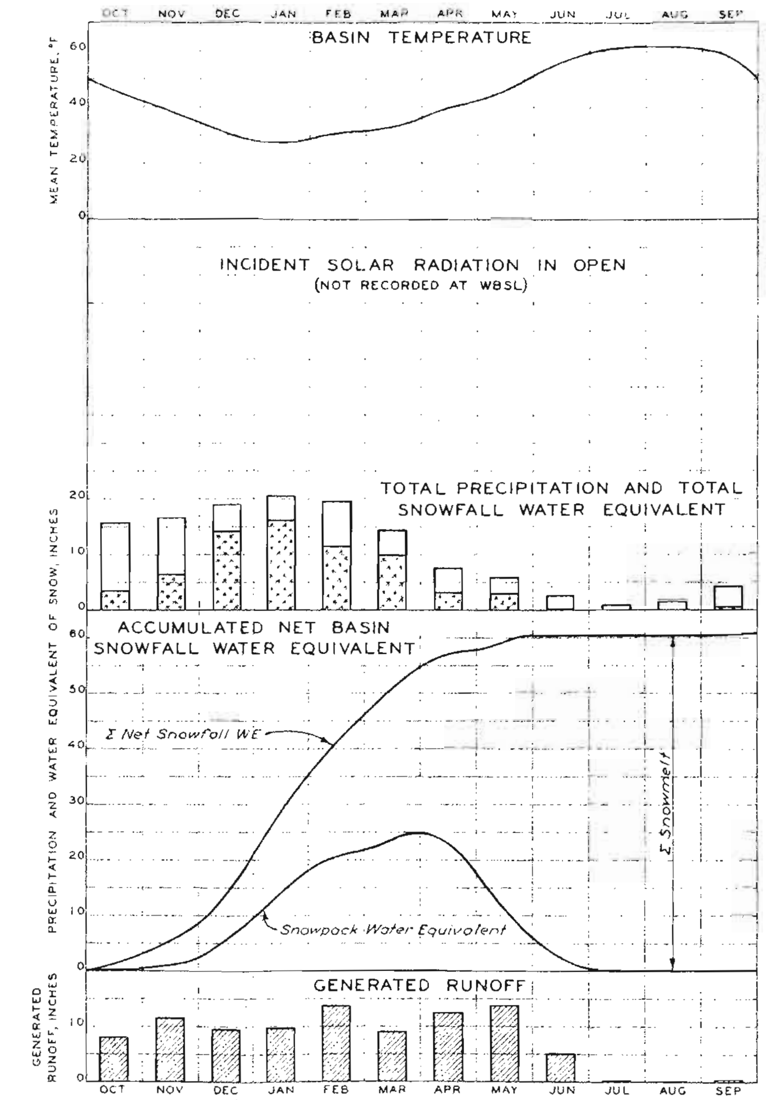
SNOW INVESTIGATIONS SUMMARY REPORT		
SNOW HYDROLOGY		
PANORAMA, WBSL		
OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION CORPS OF ENGINEERS U.S. ARMY		
PREP: T.H.P.	SUBM: P.B.B.	TO ACCOMPANY REPORT DATED 30 JUNE 1956
DRAWN: R.Y.	APPR: D.M.R.	PD-20-25/10



SKYLAND CREEK, UCSL
DRAINAGE AREA 8.09 SQ MI
OCT 1946 — SEP 1950



CASTLE CREEK, CSSL
DRAINAGE AREA 3.96 SQ MI
OCT 1946 — SEP 1951



BLUE RIVER, WBSL
DRAINAGE AREA 11.5 SQ MI
OCT 1947 — SEP 1951

Notes:

1. *Data shown in these graphs represent basin mean values for the period of record shown.*
2. *Temperature data represent basin mean values used in computing losses by evapotranspiration in water-balance computations.*
3. *Incident solar radiation was measured by pyrheliometers at UCSL and CSSL. No measurements of solar radiation were made at WBSL.*
4. *Total precipitation and total snowfall were derived from water-balance computations described in chapter 4. Values shown represent total mean basin precipitation and snowfall before interception loss.*
5. *Accumulated net basin snowfall water equivalent represents mean water equivalent of snow accumulation after interception by forest. Difference between ordinates of net snowfall water equivalent and snowpack water equivalent at a particular date represents the accumulated melt to that date from the beginning of the water year. All values were derived from water-balance computations.*
6. *Generated runoff represents the volume of water made available for runoff each month, expressed in inches of runoff over the basin area. Corrections for ground-water storage were made by streamflow recession analysis.*

LEGEND

TOTAL PRECIPITATION
RAINFALL IN INCHES OVER BASIN
SNOWFALL IN INCHES OVER BASIN
GENERATED RUNOFF IN INCHES OVER BASIN

SNOW INVESTIGATIONS
SUMMARY REPORT

SNOW HYDROLOGY

HYDROLOGIC SUMMARY SNOW LABORATORIES

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

PREPARED: M.W.	SUBMITTED: B.M.S.H.H.M.	TO ACCOMPANY REPORT DATED 30 JUNE 1956
DRAWN: A.K.	APPROVED: D.M.R.	PD-20-25/11

CHAPTER 3 - PRECIPITATION AND SNOW ACCUMULATION

3-01. INTRODUCTION

3-01.01 A fundamental problem confronting the snow hydrologist is the determination of the amount and distribution of precipitation and snowpack water equivalent in a given area at a given time. Inasmuch as the amount of snow in the accumulation phase of the hydrologic cycle is largely a function of amount of precipitation, it is desirable to consider snow accumulation and precipitation jointly in this chapter. Factors determining the amount and distribution of precipitation and of the snowpack are classified broadly as being meteorologic or topographic. Under the former are grouped such factors as temperature of the air, precipitable water therein, the circulation pattern, frontal activity, and stability of the airmass. Topographic factors include elevation, slope, aspect, exposure, forest, and vertical curvature. The effect of these factors upon amount and form of precipitation and snow accumulation is discussed herein. The problem of basic measurement is also considered, including characteristics of precipitation gages, problems of gage catch, both incremental and cumulative, and the use of snow courses and snow surveying techniques. Methods and reliability of measurements are discussed as well as techniques for translating individual values at a single point into an integrated basin total.

3-02. EFFECT OF METEOROLOGIC FACTORS ON PRECIPITATION

3-02.01 General. - Effects of meteorologic factors upon amounts and distribution of precipitation are given in various meteorological texts; therefore, comprehensive examination of these effects is not within the scope of this report. However, it is appropriate to consider briefly the more important factors and their effects upon precipitation. Basically, the requisites for production of precipitation are a supply of moisture from inflowing air and mechanisms for release of the moisture. Moisture is acquired by the air through evaporation, principally over ocean surfaces, and it is carried, largely in the form of water vapor, to the land by large-scale air movements. There is an upper limit to the amount of water vapor that can exist at a given temperature. Air that has reached this limit is said to be saturated. In saturated air, water vapor may undergo a change of state to water droplets or ice crystals. The mechanisms by which water droplets or ice crystals are precipitated from the atmosphere are complex and require that the multitude of minute droplets of condensed water which form clouds be combined into a smaller number of much larger drops. The two principal mechanisms theoretically proposed for producing moderate to heavy precipitation are (1) the colloidal instability of a mixed water-ice cloud at temperatures below $0^{\circ}C$ and (2) the coalescence of drops of un-uniform size in the gravitational field. It is believed that the two processes act together in the middle latitudes, the ice crystal process being dominant in the initiation of precipitation droplets and further growth being dependent upon the coalescing process. It is observed that, during winter storm conditions in the Pacific Northwest, all precipitation is initially in the form of snowflakes which later melt into raindrops at lower elevations. The level to which the snowflakes fall before melting is dependent upon the temperature distribution of the atmosphere. Since the moisture capacity of the atmosphere is a function of temperature, any action that will result in cooling of the air will tend to aid condensation, which in turn may produce precipitation. The principal means of cooling air in the atmosphere is through expansion, which is largely accomplished by lifting the airmass to levels of lower pressure. Frontal activity, orographic lifting, instability of the airmass, and convergence all tend to cause lifting and consequent precipitation.

3-02.02 Form of precipitation. - One aspect of precipitation is particularly important in the practical consideration of snow hydrology, that being the form in which the precipitation occurs. Of the many possible forms of precipitation, snow and rain are of prime importance, since they comprise the principal source of water deposited on the earth's surface from the atmosphere. Direct observation of form of precipitation is limited to first-order and airways meteorological stations, generally necessitating use of indirect methods for estimating areal distribution of rainfall and snowfall where required for hydrologic problems. Of particular significance, moreover, are mountainous areas where slight changes in airmass characteristics will cause relatively wide fluctuations of the boundaries between areas on which rain or snow is falling.

3-02.03 Estimation of form of precipitation. - No relationship has been found to define exactly, by use of meteorological parameters of airmass conditions, whether precipitation at a given level will occur in the form of rain or snow. Murray [15] attempted to relate the form of precipitation to such variables as the thickness of the 1000-700mb or 1000-500mb layers, the height of the freezing level in the atmosphere, and surface air temperatures. All of the variables tested showed a range of values for which either rain or snow could be expected. Each variable was analyzed on the basis of the statistical distribution of occurrences of rain or snow or both to determine the most probable frequencies of occurrence. The study showed that surface air temperature (approx. 4 feet) is as reliable as any other of the variables tested for differentiating between rain and snow. Accordingly, and since surface air temperatures are generally more available than upper air soundings, hourly observations of precipitation and surface air temperature at Donner Summit, California (elev. 7200 ft.) were used in a study of form of precipitation. Data for the period October through April for the 1946 through 1951 water years were used. Some 2400 occurrences of precipitation at air temperatures ranging from 29°F to 40°F were analyzed to obtain the distribution of occurrences of rain, snow, or mixed rain and snow. The results of this study are summarized in the following table which gives the percentage occurrences of both forms of precipitation for the various surface air temperatures:

FORM OF PRECIPITATION	SURFACE AIR TEMPERATURE, °F											
	29	30	31	32	33	34	35	36	37	38	39	40
SNOW (%)	99	99	97	93	74	44	32	29	8	8		3
RAIN (%)		1	2	3	12	31	51	57	81	90	100	97
MIXED RAIN AND SNOW (%)	1		1	4	14	25	17	14	11	2		
Number of occurrences, (all forms)	281	419	304	459	229	191	154	94	74	79	56	79

The above data are shown graphically in figure 1, plate 3-1. These data conform closely to those quoted by Murray for a markedly different condition of elevation and topography. Accordingly, they are believed to be applicable generally. On the basis of this analysis, the differentiation between rain and snow can be estimated from surface temperatures by assuming rain to occur whenever the air temperature is 35°F or greater, and assuming snow to occur whenever the air temperature is less than 35°F. Within the range of temperatures given in the above table, which includes all questionable precipitation forms, about 90 percent of the cases would be correctly designated by this 34-35°F division between rain and snow. When considering the various inaccuracies in measurements and application to specific problems, further refinements in estimating form of precipitation would be seldom if ever warranted.

3-02.04 Additional evidence supporting the selection of 34-35°F as the dividing line between rain and snow appears when the rate of fall of snowflakes and the relation between the freezing level and the average level of melting of snowflakes, are considered. Murray found that on the average there was an equal probability of rain or snow when the height of the freezing level was approximately 1000 feet above the ground. During times of significant precipitation, the atmospheric lapse rate is usually nearly wet adiabatic. With the freezing level at 1000 feet, the corresponding surface air temperature would be about 3°F warmer than freezing, or 35°F. The average rate of fall for snowflakes, determined by Nakaya [16] for several of the predominant crystalline forms of snow, is about 50 cm/sec, which is equivalent to about 1.6 feet per second. Accordingly, it would take slightly in excess of 10 minutes for snowflakes to fall 1000 feet, which thus represents the average time of melting of snowflakes in the atmosphere.

3-02.05 An additional study shows the relative frequencies of forms of precipitation at varying temperatures for three elevations: 1000, 4000, and 7000 feet. Figure 2, plate 3-1, presents the results of this study. The frequency distribution of precipitation with surface air temperatures for the 7000-foot level was derived from the Donner

Summit data used in the form-of-precipitation study. For temperatures below 29°F, the frequency curve was extended on the basis of data presented by D'Ooge 5/ for the White Mountain Research Station (37°30'N, 118°10'W) located at an elevation of 10,600 feet. These data were corrected to elevation 7000 feet by assuming a wet adiabatic lapse rate, and frequencies were adjusted to give values comparable to the Donner Summit data. The combined curve shows schematically the average distribution of precipitation occurrences with varying temperatures. The equivalent distribution curves for 4000 and 1000 feet were determined simply by applying the wet adiabatic lapse rate to the surface air temperatures. This may be done with assurance since, during times of significant precipitation, the polar maritime airmasses usually dominant in this area are conditionally unstable, and the uplift along the mountain ranges causes the airmasses to release their instability, resulting in atmospheric mixing and the establishment of a wet adiabatic lapse rate of temperature. Also, during storm conditions, the horizontal variation of air temperature is relatively small because of the turbulent mixing which occurs, and the surface air temperatures at a given elevation are representative of the airmass temperature at that level. The relative frequencies of rain and snow forms of precipitation from the air temperature function may be applied to each elevation for which the average distribution of total precipitation with temperature is known, to determine the average snow and rain frequency distribution for that elevation. The frequency distribution of precipitation in the vicinity of CSSL is shown schematically in the diagrams of figure 2, plate 3-1. Inspection of these diagrams reveals that on the average approximately 95 percent of occurrences of winter precipitation at 7000 feet are in the form of snow, while at 4000 feet about 50 percent are snow. At 1000 feet, only 2 percent of the occurrences are snow. The diagrams in figure 2 represent average conditions on the west coast of the United States at latitude 39°N. They are, however, believed to be generally applicable along the windward side of major mountain ranges in western United States and Canada, with the exception that with increasing latitude there would be a slight lowering of the elevation levels. For example, it is estimated that at latitude 50°N, the elevations of the levels shown for the CSSL would be 6000 feet, 3000 feet, and sea level, an approximate decrease of 1000 feet per 10 degrees of latitude. It is emphasized that these conditions are based on averages over several years of record. Since for a given airmass the form of precipitation is particularly sensitive to changes in elevation, the distribution of rain and snow over a project basin for a particular occurrence or season must be evaluated individually on the basis of prevailing meteorological conditions.

3-03. EFFECT OF TERRAIN ON PRECIPITATION

3-03.01 General. - Over uniform level terrain, as, for example, open ocean or plains, the areal distribution of precipitation is a function entirely of meteorologic variables and their relation in time and space. In mountainous regions, on the other hand, precipitation distribution is largely a function of the character of terrain, upon which is superimposed the distribution due to meteorological conditions of airmasses, fronts, and general atmospheric circulation. The distribution patterns of precipitation over rugged terrain are, therefore, complex and in extreme conditions may well be considered to be as irregular as the terrain itself. To evaluate precipitation satisfactorily in mountainous regions, consideration must be given to the relationship of terrain to meteorologic conditions producing precipitation. An additional consideration of importance is the large scale atmospheric circulation, its seasonal variation, and its change with varying synoptic situations, during times of significant precipitation.

3-03.02 The effects of terrain on precipitation are broadly classifiable as being either large scale or local. Many of the factors are interrelated and there is no simple relationship which is applicable to all areas and conditions of terrain. In addition, errors in precipitation measurements used to represent conditions at a point impose further difficulty in establishing relationships between terrain and precipitation. Measurements must be corrected for known deficiencies (for example, those resulting from poor exposure of precipitation gages or poor conditions of snow sampling) before analyses are made. The relationships developed between uncorrected measurements and terrain features are meaningless.

3-03.03 Small-scale terrain effects. - Studies undertaken by the Snow Investigations to determine relationships between terrain and precipitation in the form of snow were concerned primarily with small scale or local influences. The laboratory basins provided data for such analyses; consequently generalized conclusions drawn therefrom were restricted to ranges of conditions experienced in the small laboratory areas. Since each laboratory has unique terrain features (e.g., at the CSSL the majority of steep slopes are south facing), the correlations are not valid for direct application to other areas. They do, however, show relative variation of measured amounts with respect to principal topographic features at the laboratories, and the method is illustrative of techniques which can be applied to other basins.

3-03.04 Snow laboratory analysis. - Snowpack water equivalent data from CSSL and UCSL were used for analysis of the influence of terrain characteristics upon precipitation accumulated in the form of snow. The water equivalent at each of the laboratory snow courses was related to parameters of aspect, slope, curvature, elevation, exposure, and vegetation, by use of multiple linear regression analysis. Research Notes 2 and 14 report the results of these investigations. The latter study incorporates data from 36 snow courses in the CSSL (including 16 special snow courses for which observations were made on 19 April, 9 May, and 22 May 1951). The observations for 19 April are most representative of snow accumulation since by the latter dates considerable melt had taken place. The early period also reflects to some degree the varying effect of melt over the basin, which began about 1 April. A multiple regression analysis for the 19 April data, using water equivalent of the snow courses as the dependent variable and parameters of aspect, slope, elevation, and exposure sector as independent variables, gave a correlation coefficient of 0.76 and showed the parameters of aspect, elevation, and exposure sector to be statistically significant. A graphical solution of the derived regression equation is given on plate 3-2. A summary of the studies showing the numerical effects of topographic characteristics on water equivalent at the CSSL for the period studied is as follows:

Elevation - Approximately 1- to 2.5-inch increase in water equivalent for each 100-foot increase in elevation.
Slope - Approximately 0.2- to 0.5-inch decrease in water equivalent for each 1 percent increase in slope.
Aspect - Approximately 0.25- to 1-inch increase in water equivalent for each 10° deviation of aspect from the south.
Exposure sector - About 0.5- to 0.75-inch decrease in water equivalent for each 10° increase in exposure sector (sum of the sectors of a circle of half-mile radius, centered at the snow course, within which there is no land higher than the points on the snow course).
Curvature - No definite indication found.
Vegetation - No definite indication found.

More detailed quantitative interpretation of the results is not believed to be warranted; however, qualitative interpretations may be made on the basis of these observations. The main deterrent to application of the results to other basins is the small range of conditions experienced at the laboratories.

3-03.05 Forest effects. - The effect of forest cover on the accumulation of snowpack water equivalent was not evident from the analysis of the CSSL data. This is due primarily to the fact that the snow courses did not adequately sample the range of conditions of forest effect, most snow courses being located in open areas. Later measurements made in connection with research work done by the Northern Rocky Mountain Forest and Range Experiment Station of the Forest Service clearly define the effect of forest in the accumulation of snow. Since the forest effect is considered primarily to be one of interception of precipitation, the results of the study are discussed in the paragraphs on interception loss in section 4-03.

3-03.06 Large-scale terrain effects. - Spreen's [18] analysis of the effect of topography on winter precipitation for western Colorado is an attempt to correlate larger-scale terrain features with 11-year average winter-season precipitation, as measured by standard U. S. Weather Bureau precipitation gages. His parameters include elevation, slope, orientation, and exposure. Instead of using a multiple linear regression analysis, he chose to relate the variables by means of a graphical correlation technique. The derivation of some of the parameter curves appears to be somewhat arbitrary, particularly that of exposure, and the derived curves do not represent continuous functions of the variables. The method does, however, indicate the magnitude of terrain effects on precipitation in a general sense, and the results of his analysis show that the parameters used account for a high percentage of the variability of precipitation in that area. The reader is referred to the above cited report for the results of this analysis.

3-03.07 Summary. - Because of the lack of comprehensive data on the effect of terrain upon precipitation, it does not seem valid to apply the quantitative relationships for the snow laboratories to project basins. The values cited from analysis of laboratory data show only the effect of the integration of the existing meteorological conditions over the basin for a particular season, as they are related to the particular basin's terrain. For any other area, different relationships would be expected. In order to generalize on the effect of terrain, average seasonal amounts based on a period of several years would provide more usable information. Also, it would appear more reasonable to use the values of point precipitation or point water equivalent in terms of percentage of average basin amounts rather than using actual values, when considering the variation for individual years or occurrences.

3-03.08 In the final analysis, either basin precipitation or basin water equivalent must be estimated from measured values by sampling techniques. The gaged values, corrected for deficiencies in measurements, must be related to basin amounts on the basis of normals. Each measurement of precipitation, then, becomes an index of the basin amount in actual basin application. If the precipitation distribution pattern over a basin, for a season as a whole, is uniform from year to year, any well-located station will be a satisfactory index of basin amounts regardless of the terrain type being sampled. While there is variability in precipitation distribution from year to year, due primarily to atmospheric circulation and airmass characteristics at times of precipitation, experience in the western United States indicates that nearly all winter precipitation occurs at times of strong westerly circulation. The synoptic pattern during such times is generally characterized by the passage of mature occluded frontal systems entering from the Pacific Ocean, and the region is dominated by maritime polar airmasses. Therefore, in this region during conditions of significant winter precipitation, terrain variables such as aspect, slope, and curvature (but excluding elevation effects on snow accumulation) do not materially affect the reliability of any one well-located station as an index of basin precipitation.

3-03.09 When using snow course water equivalent measurements to represent either basin precipitation or the residual water stored on a basin as of a particular date, the same general considerations are involved as for total seasonal precipitation. There is, however, one important variable of terrain which is not proportionally represented on the basis of average meteorological conditions, that being elevation. The effects of elevation on the accumulated water equivalent vary considerably from season to season and year to year, due primarily to the variability of temperature, both during and between the storm periods. The airmass temperature determines the level of change between rain and snow forms of precipitation. Also, during the accumulation period there is differential melting which is primarily a temperature function and in turn varies with elevation. Both of these processes result in varying rates of accumulations of snow with elevation, depending on the temperature regime during the accumulation period, as will be discussed in the following paragraphs.

3-04. ELEVATION EFFECTS ON SNOW ACCUMULATION

3-04.01 General. - It has been pointed out in the previous section that, of all terrain parameters, elevation is the principal one which must be taken into account in determinations of basin water equivalent from point snow course measurements because of the variability, from season to season, of snowpack accumulation with elevation. In order to demonstrate the range of conditions that may be expected, an analysis of data for WBSL is given from which generalizations on the variation of the snowpack with elevation can be made.

3-04.02 The analysis for WBSL was chosen because of (1) the relatively large range of elevation sampled by the snow courses (2100 to 5000 feet msl), (2) the relative consistency of the effect of meteorologic and terrain influences, and (3) meteorological conditions which normally permit occurrence of either snow or rain forms of precipitation at any elevation. Data from CSSL are not amenable to such analysis owing to the limited range of elevation of snow courses. UCSL snow courses sample an area whose winter precipitation is nearly 100 percent snowfall, so that elevation change does not adequately reflect the result of varying percentages of rain and snow. The ordinary snow survey network established for seasonal runoff forecasting in non-laboratory basins is not adequate for this type of analysis because of insufficient density of sampling by snow courses. Data from Technical Bulletin 17, which evaluates snow conditions from 2000 to 7000 feet on the Yuba River basin, California during the heavy snow of December 1951 and January 1952, were used to provide an independent check on results obtained from the WBSL analysis.

3-04.03 WBSL study. - For each of the three water years, 1949 through 1951, snowpack water equivalent data from WBSL snow courses were adjusted to 1 January, 1 February, 1 March, 1 April, 1 May and 1 June, and plotted as a function of elevation. Straight lines were fitted to these points, giving due consideration to the relative exposures and reliability of measurements at each snow course; each line so derived shows the average depth of snow at various elevations of the basin. This line together with the zero base line forms a wedge hereinafter referred to as the snow wedge. Figure 1, plate 3-3, shows the plot of the snow wedge for each month in each of the three years at WBSL, as well as for the months of December, January, and February of the 1952 water year in Yuba River basin. Inspection of these diagrams reveals the relative changes in slope of the snow wedge within seasons as well as from year to year. The slope of the snow wedge plotted against time, to illustrate the seasonal change in slope, is shown in figure 2, plate 3-3. Adjacent to each of the plotted points is the water equivalent at the 5000-foot elevation, which indicates the relationship of the slope of the snow wedge to the magnitude of the water equivalent near the upper limit of the basin during the accumulation period. Finally, the relationship between the slope of the snow wedge and the water equivalent near the upper limit of the basin (5000 feet for WBSL and 7000 feet at Yuba River) during the accumulation period is shown in figure 3.

3-04.04 Accumulation period. - Inspection of figure 2, plate 3-3, shows that the slope of the snow wedge becomes progressively steeper through the accumulation period, but tends to remain more or less constant during the melt period. During the accumulation period, the variation in melt with elevation is a minor factor, the principal factor affecting the slope of the snow wedge being the change from rain to snow form of precipitation with elevation. Another factor affecting the slope of the snow wedge is the normal increase of precipitation with elevation. The plotted points for WBSL in figure 3, plate 3-3, show the relationship between depth of water equivalent at the 5000-foot elevation and the slope of the snow wedge during the accumulation period. A straight line fitted to these points shows an average increase in slope of the snow wedge of 0.3 inch per 100 feet for each increase of 10 inches of water equivalent at 5000 feet. The difference in the relationship which may occur between individual years is shown by the difference of accumulation characteristics for the 1949 and 1950 water years. The 1949 water year, which had near maximum of record snow accumulation at high elevations, was characterized by a relatively steep snow wedge; there was not an abnormal snowpack accumulation for the basin as a whole. By contrast, the 1950 water year with a similar amount of snow in the high portions of the basin, had relatively less change of water equivalent with elevation and as a result had a proportionally greater snow accumulation over the basin as a whole. This difference in snow wedges indicates the necessity for considering elevation effects in computing basin water equivalent. The three points for Yuba River basin for 1 December, 1 January, and 1 February, related to the water equivalent at 7000 feet, show an increase in slope of the snow wedge similar in magnitude to that for WBSL.

3-04.05 Melt period. - Data for the melt period, as represented by 1 May and 1 June observations, are not as reliable as data for the accumulation period, because of the varying effect of local exposures of snow courses on melt rates. Analysis of data from the heavily forested WBSL indicates that the slope of the snow wedge remains nearly constant during the melt period, with only a slight tendency to increase as the melt season progresses. The increase with respect to time is considerably less than that for the accumulation period. In order to evaluate rationally the changes in melt rates with elevation, it is necessary to consider all methods of heat exchange for specific conditions of environment. For the heavily forested WBSL, however, snowmelt runoff can be estimated fairly well by an air temperature function which indicates 0.041 inch of melt per degree-day of mean daily air temperature above 29° base (see chapter 6). During active melt periods in May or June, the mean daily temperature at the mid-point of the basin (elevation 3500 feet msl) is about 60°F, which represents an average daily ablation of the snowpack water equivalent of about 1.27 inches. Assuming an average temperature lapse rate of 3°F per 1000 feet over the snow-covered area, the melt at 5000 feet would be 1.11 inch per day, or 88 percent of the melt at the mid-point. Since the forest cover tends to decrease in density with elevation, average melt rates per degree-day tend to increase, so that actually the melt at the top of the basin is even closer to the mean melt rate for the basin than is indicated by the simplified illustration above.

3-04.06 Summary. - In general, the increase in the snowpack water equivalent with elevation varies widely from year to year and from season to season. During the accumulation period, the largest factor affecting the variation in snowpack accumulation with elevation is the form of precipitation which in turn is dependent upon meteorological conditions during times of precipitation. Secondary effects are the elevation differences in melt, both during and between storms, and the normal increase of precipitation with elevation. During the accumulation period there is a fair relationship between slope of the snow wedge and the total water equivalent in the upper portion of the basin. The relationship varies from year to year, however, so that in individual cases, the basin snowpack water equivalent must be adequately sampled throughout its range in elevation. After the time of maximum snow accumulation, the slope of the snow wedge becomes more or less constant. There is only a slight further increase in slope during the active melt period. During the active melt period, the slope of the snow wedge bears little relationship to the water equivalent of the snowpack at high elevations. Special care should be taken to adequately represent the actual basin condition from late-season snow surveys in evaluating basin snowpack water equivalent. It is pointed out that the above generalizations apply to the zones of elevation where straight lines adequately represent the variation of water equivalent with elevation. For areas lying above elevations where more than 90 percent of the precipitation during the accumulation season is in the form of snow, the slope of the snow wedge would tend to flatten, and curvature would be introduced into the lines.

3-05. POINT PRECIPITATION MEASUREMENTS IN AREAS OF SNOWFALL

3-05.01 General. - Proper appraisal of precipitation data necessitates a knowledge of the characteristics and limitations of the instruments used to gage precipitation catch and the observational procedures used in determining point precipitation values. Much has been written on the general problem of measuring precipitation, and reference is here made to a comprehensive summary of rain and snow gaging methods by Kurtyka. 11/

3-05.02 Gaging of snow. - The accuracy of snow measurement by standard rain gages is subject to sharp limitations due to the physical characteristics of the snow itself. These limitations have long been recognized through many experiments conducted both in this country and abroad. Experience gained at the snow laboratories in measuring winter precipitation has provided factual data on reliability of precipitation measurements in areas of snowfall and a measure of the gage-catch deficiencies involved. Wilson's report analyzing winter precipitation observations in the Cooperative Snow Investigations 23/ summarizes the problem of measurement of point precipitation, both by precipitation gages and snow courses. The study generalizes the effects of weather degree of accuracy, basin precipitation from station precipitation amounts. That is to say, the ratio of station precipitation to basin precipitation is approximately constant for any given storm or group of storms, and hence the normal precipitation for any station bears a constant relation to the normal basin precipitation. Thus,

$$\frac{P_b}{P_a} = \frac{N_b}{N_a} \text{ or,} \qquad P_b = P_a \frac{N_b}{N_a}$$

where P_b is the desired basin precipitation, P_a is the gage catch at a particular station (or group of stations), N_b is the basin NAP, and N_a is the NAP for the station(s) used to determine P_a. Such an equation yields good results provided that the number and distribution of the stations used in the relationship adequately represent the basin. As a further refinement, if stations are not evenly distributed, they may be weighted in accordance with the fraction of the basin represented by each station. These weights are treated as coefficients by which the corresponding station precipitation and station normals are multiplied before substituting in the formula.

3-06.04 The assumption of a fixed relationship between the precipitation of any station and the basin precipitation can be considered at best an approximation. A more rational method is that of the isopercentual map. Here, storm (or annual) precipitation at the various stations is expressed in percent of NAP, and isopercentual lines are then drawn for these values. This isopercentual map may then be superimposed on the NAP map and isohyets shown for the particular storm by noting intersections. The NAP map reflects the normal topographic effects while the isopercentual map indicates deviations from normal. The fewer and more widely spaced the isopercentual lines, the more nearly does the individual storm conform to the normal pattern. The chief value of this approach over the more conventional direct drawing of an isohyetal map, lies in the ability to utilize quickly and with considerable confidence a detailed pattern of precipitation and still take into account much of the obvious variation of individual storms from the mean.

3-07. POINT MEASUREMENTS OF SNOW

3-07.01 General. - Measurements of snow include the water equivalent both of newly fallen snow and the net accumulation of snow. The latter is particularly useful for basin water supply forecasts, since it integrates in a single set of measurements the basin snowfall and melt. It thereby represents water remaining in storage in the snowpack. As in the case of total precipitation, snowfall or water equivalent must be measured by sampling at a point and extrapolating point values to represent basin amounts. The following paragraphs describe commonly used methods of measurement of snow, sources of error, limitations of measurements, and general requirements for locating snow courses.

3-07.02 Measurement of snowfall. - A simple method of measuring snowfall is by use of the snowboard. The board is placed on the ground or old snow surface to permit accumulation of new snow. An inverted rain gage cylinder is utilized to isolate a core of the new snow, which is then melted and measured in the same manner as rainfall. By measuring each fall of snow in this manner and replacing the clean board again to receive fresh snowfalls, accumulated total snowfall throughout the season may be known at any time. Such measurements are fairly reliable, provided that they are taken soon after each snowfall and that the snow on the board has not been subject to drifting, melting, or evaporation. Other devices based on the same principle as the snow board are Angot's snow basket, snow tables, and snow bins.[11/]

3-07.03 Another method of determining water equivalent of new snow is by multiplying the depth of new snow by an appropriate density factor. Studies based on data at CSSL indicate that a fair relationship exists between air temperature and density of new fallen snow. [4/] A graphical plot and discussion of this relationship is contained in chapter 8. Densities of new snow based on air temperatures are acceptable for determination of water equivalent if the period involved is sufficiently long to permit compensation of errors in density values.

3-07.04 Snow density. - The specific gravity of snow, (a dimensionless ratio) is commonly called the snow density (which properly would be mass per unit volume). Following usage, this terminology is used in this report. Snow density is generally expressed numerically as a percentage. The condition of snow density in the snowpack varies widely within the vertical structure of the snowpack as well as with time. Chapter 8 describes the changes of density that occur within the snowpack and the processes that affect the metamorphism of snow.

3-07.05 Snow stakes. - Snow stakes are often used to measure depths of snow accumulation. Water equivalent of the snowpack can be determined from depth measurements by evaluating the time-duration of processes affecting the settlement of the snow or by using known densities of snow under similar conditions of environment. The use of snow stakes does allow the evaluation of relatively short time increases or decreases in water equivalent without disturbing the snowpack sampling area. Also, snow stakes may be used where it is impractical to obtain water equivalent by direct sampling, particularly by reconnaissance from the air. This method has been used successfully by California Electric Power Company, as reported by Henderson. [9/]

3-07.06 Sampling of accumulated water equivalent. - Direct sampling of water equivalent is usually accomplished by use of snow sampling equipment, the most common of which are the Mount Rose and the Utah samplers. These two types are basically similar, being comprised of a tube fitted with a cutter on one end and a handle on the other end. The tube has an inside diameter of 1.485 inches (which makes one inch water equivalent weigh one ounce) and is made up of sections to facilitate transportation. Sampling consists of pushing the tube vertically into the snow to full snow depth, withdrawing the tube with the snow content, and weighing. Weighing scales are designed to give readings in inches of water equivalent; tare weight, of course, must be subtracted to obtain the weight of the snow. A record is made of the depth, length of core, weight, and any other useful information such as estimated moisture content of the underlying soil.

3-07.07 Sources of errors. - Water equivalent measurements are subject to a variety of errors. As in any type of work where measurements are made, snow surveying data are subject to such usual observer errors as misreading the sampler weighing scale. Probably the most common error is that resulting from an incomplete core of snow in the sampler tube. This may be caused by clogging of the cutter by corky snow, obstructions such as cones or sticks, or sticking of the snow to the tube. Such occurrences can generally be detected by comparing the length of core with depth of snow. Another source of error is sampling of ponded water in the lower portion of the core, resulting from poor snowpack drainage. In such cases the water equivalent may be computed by multiplying the depth of snow by densities obtained at reliable sample points. Where frequent observations are made, care must be exercised to avoid holes left by prior sampling. Any dirt or other foreign matter must be removed from the cutter end of the sample before weighing. The many details pertaining to snow surveying for the purpose of obtaining the water equivalent of the snowpack at a given point are beyond the scope of this report; for details the reader is referred to a comprehensive report on snow surveying by Marr. [13]

3-07.08 Snow courses. - The common practice in taking snow measurements is to sample the water equivalent at a number of points along an established line called a snow course. Snow courses are located with the objective of obtaining data representative of a given area, the number depending largely upon the terrain and meteorological characteristics of the area. Other factors such as accessibility, availability of funds, and purpose for which the data are to be used, must, of course, be considered in the establishment of the network of stations. Site selection should be based on the same general requirements as for precipitation gages, considering: (1) meteorological conditions with respect to storm experience; (2) position with respect to large-scale topographic features; (3) position with regard to local environmental features, such as exposure, aspect, orientation, and ground slope; and (4) conditions on the site itself, such as local drainage and the occurrence of brush and rocks. In addition, snow courses should be located to adequately sample ranges in elevation, and they also should be so located that they are representative of average basin melt conditions as well as basin snow accumulation. As is the case for precipitation gages, snow courses should be located in areas well protected from wind, since wind erosion and drifting snow cause unrepresentative snow accumulations. An ideal location consists of an opening in the forest surrounded by hills for protection from high winds, and sloped sufficiently to permit runoff of water beneath the snowpack. The number of sample points is variable, depending largely upon the consistency of the distribution of snow. Sample points are located with the objective of avoiding variations in snow depth due to causes such as drifting, interception by trees, and presence of boulders or other obstructions. If protection from wind is altogether lacking the sampling points must be spread over a wide area to average out variations due to drifting. In general, five snow-course sample points are probably adequate for well-located snow courses on which there is a minimum amount of irregularities caused by drifting or wind erosion, if the ground surface is smooth and clear of all obstructions, and if the snow course is not too close to the forest or other obstructions of a local nature to be influenced by local irregularities in deposition. When conditions are less than ideal, however, additional snow course points are required to adequately sample the water equivalent. Basic data from snow courses are obtained under cooperative arrangements between various Federal, state and private organizations and are coordinated and published for most of the western United States by the Soil Conservation Service. In California, the supervision of collecting and publishing snow course data is by the Division of Water Resources for the State of California.

3-07.09 As in the case of precipitation gages, snow courses also have certain limitations as measures of precipitation. Wilson discusses the reliability of snow course measurements in connection with measurement of winter precipitation at the snow laboratories. Reference is made to his article [23] for comparative data showing reliability of individual point measurements of snow accumulation, as well as relative reliability of snow courses. Studies for CSSL have shown that a single, well-located snow course (such as station 5) will more adequately represent the basin water equivalent for this relatively small area than a group of poorly-located snow courses. Although care is exercised in selecting locations having stable physical features, changes affecting the deposition of snow at sampling points may occur. A common change in physical features is the removal of all or a portion of surrounding timber by causes such as fire, cutting, bug infestation, or severe wind storms. On the other hand an opposite effect can be produced by growth of brush or timber in the vicinity of the sampling points. In the latter case annual changes may not be detectable; nevertheless, the change over a period of years may be significant. Another important effect of physical changes is improper drainage of free water as a result of obstructions such as beaver dams or accumulation of debris in drainage channels in the snow course area. Occasionally physical features may change sufficiently to necessitate abandonment of the snow course. Often, however, the location is acceptable despite some changes in physical features. In such cases adjustment in records must be made, using the double-mass curve method as for precipitation.

3-07.10 Radioisotope snow gage. - A radioisotope-radiotelemetering snow gage has been developed by the Corps of Engineers (see para. 2-07.09) to make measurements of snowpack water equivalent at remote, unattended sites, and to transmit these data to a central receiving station. A general description of this gage and its operation is given in the following excerpt from the final report on its development and test performance: 3/

> "As presently developed, the equipment includes a gamma ray source, cobalt 60 (Co 60) enclosed in a lead collimating cylinder that is installed in a block of concrete set flush with the ground surface. The beam of gamma radiation leaving the lead collimating cylinder strikes a Geiger-Muller radiation-detector type (G-M tube) suspended 15 feet overhead. The G-M tube is designed to measure the residual radiation after attenuation by the snowpack. The pulses of electricity caused to flow in the G-M tube by the gamma rays are first amplified, then are divided by eight(for transmission), and then are fed into a frequency-modulated radio transmitter with a $\frac{1}{4}$-watt power output. The pulses are broadcast in the VHF range using a high-gain directionalized output antenna. The installation, which is operated on battery power (including high voltage for the G-M tube), is capable of sustained, unattended operation for eight months. The transmitter is operated for only short periods each day by means of an electrically wound, spring-driven clock with an automatic switching device."

Since the measurements made by the gage are non-destructive, the same point may be sampled over and over again, thus doing away with the need for numerous samples to average out errors which result from the unevenness of the underlying ground and other errors inherent in individual samples of the snowpack obtained with a snow tube. Moreover, the non-destructive aspect of the gage, in combination with its telemetering feature, makes it possible to take daily (or even more frequent) readings of the water equivalent in contrast to the monthly or bi-monthly data obtained by conventional snow-tube measurements. Thus it has the particular advantage of providing continuous information on increments of snowpack accumulation from storm to storm throughout the accumulation period, and also of providing a means of evaluating increments of melt during the ablation period, from an undisturbed snow sample point at a remote site. As presently developed, the gage can obtain an accurate measure of water equivalent up to about 40 inches, for an unattenuated counting rate of 20,000 counts per minute. The accuracy of measurement for a ten-minute counting period under this condition is illustrated by the relative magnitude of errors as follows:

Water equivalent (in.)	10	20	30	40	50
Std. error of est. (in.)	0.1	0.1	0.2	0.7	3.4

Using present equipment, this error of measurement may be reduced and an accurate measure of water equivalent in excess of 40 inches may be obtained by (1) increasing the strength of the radioactive source, (2) by decreasing the distance between the source and the sensing unit, or (3) by increasing the length of the counting period. In addition, to provide more accurate measurements at greater depths, such changes in the equipment as shielding the sensing unit from cosmic radiation or using multiple sources of radioactive material at varying heights above the ground have been considered. The figures given in the tabulation above are representative of the accuracy of the gages currently in use by the Sacramento District in the Kings River drainage above Pine Flat Dam.

3-08. RELATIONSHIP OF POINT VALUES TO BASIN WATER EQUIVALENT

3-08.01 General. - Having determined point values of all snow courses representative of an area, the problem is to utilize these values to determine the water equivalent on the area. The basin value may be expressed either as an index or as a quantitative measure such as inches depth over the basin. Volume of the snowpack is an important factor in forecasting seasonal volume of runoff, as a determinant of peak flow, and for short-term rate-of-flow forecasting based simply on depth of water equivalent. Index relationships are most commonly used for forecasting seasonal runoff volumes and to some extent for forecasting peak flows. Regardless of how a basin index is derived from point values, its usefulness is dependent upon how well it represents basin conditions, rather than upon its representativeness of the point values.

3-08.02 The relationship of basin water equivalent to point values is dependent upon the location of the snow courses chosen to represent the area. If the snow courses are distributed equally throughout all elevations of the basin, an arithmetic average of the point values will often provide a satisfactory index of the basin water equivalent. Refinements can often be made by weighting the snow course data in accordance with the percentage of the basin area represented by each. Averages of point values, whether weighted or unweighted, are generally acceptable only as indexes; i.e., unless properly adjusted, the average is not a reliable measure of the actual volume of water equivalent on a basin.

3-08.03 Elevation effects. - Of particular significance is the fact that on most basins, snow courses are not distributed evenly throughout all elevations, but are generally concentrated at the higher levels. As previously discussed, the distribution of snow over a given range of elevations can vary widely and, accordingly, measurements at high-level stations are not representative of the basin as a whole, particularly if the high elevations comprise only a small percentage of the basin area.

3-08.04 Snow charts. - A logical step toward the solution of the problem of unrepresentative snow-course data is the development of a method whereby each snow course is given weight corresponding to the percent of basin area that it represents. Because of the importance of variation of water equivalent with elevation, a chart on which water equivalent depth is plotted against elevation has been designed to facilitate computation of basin water equivalent. On the chart, as shown by the example on plate 4-2, it will be noted that water equivalent at a given snow course is plotted against its corresponding elevation.

A line connecting the points determines an area on the chart which is proportional to the water equivalent over the basin, provided that proper consideration is given to the area-elevation relationship of the basin. Using area-elevation data, an auxiliary scale representing percent of area is marked along the abscissa with each percentage located at its corresponding elevation. For convenience the chart is divided into zones, each of which represents 10 percent of the area, a dashed vertical line being drawn through the mean elevation of each zone. The intersection of the water-equivalent line at each of the dashed lines represents the average depth for the corresponding zone. The sum of the depths for each zone divided by 10, therefore, is an index of the mean water equivalent depth over the basin.

3-08.05 In addition to elevation, factors such as slope, exposure, and aspect have an effect upon distribution of snow on a basin, as pointed out in the discussion on terrain parameters. Accordingly, a snow course at a given elevation does not necessarily represent the water equivalent at that elevation throughout the basin. Therefore, plotted points representing water equivalent depths at snow courses do not necessarily fall along a well-defined line and generally show considerable dispersion. If all factors affecting distribution of snow were properly considered and an average basin depth for each elevation were established, then a line drawn through these average depths would indicate the true basin water equivalent. Since it is impractical to properly evaluate the effects of all factors on a basin, a line of best fit through plotted snow-course values may be drawn and the basin water equivalent derived from it may be used as an index. Such an index is more reliable than an average of snow course readings, inasmuch as each course is given weight in accordance with the percentage of basin area that it represents.

3-08.06 Index values vs. actual values. - It must be remembered that the discussion above concerns the establishment of an index which, though considered to represent with fair accuracy the changing conditions of the basin snowpack through the season, is not to be mistaken for an actual quantitative evaluation. However, if the basin water equivalent can be quantitatively evaluated by other means such as by subtracting runoff and loss from total precipitation, a ratio between the basin snow index and computed actual volume of snow can be determined. This empirical conversion factor can then be used for computing actual basin water equivalents from the snow chart values.

3-09. SUMMARY

3-09.01 This chapter has dealt with the measurement of precipitation and the accumulation of snow, and methods of estimating amounts over basin areas. The form of the precipitation, as rain or snow, was shown to be a function of meteorologic variables, and frequency diagrams illustrated the climatic averages of the relative proportion of winter rain and snow in the western United States, for varying elevation and temperature conditions. Both meteorologic and terrain conditions affect the amount and distribution of precipitation. In mountainous regions, the terrain effects may be classified as large or small scale, and it was shown that terrain differences cause wide variability of precipitation amounts, depending upon such factors as aspect, slope, curvature, and elevation, as well as location with respect to major mountain barriers and the opportunity for airmass modification. It was indicated that in the western United States, the major portion of winter precipitation occurs at times of strong westerly circulation and of uniform airmass characteristic, so that terrain effect, averaged over a winter season, causes a more or less constant relationship to exist between point precipitation and basin amounts for reasonably small basin areas. Elevation effects on snow accumulation, however, should be considered in the evaluation of the specific snowpack condition, because of the variability of form of precipitation and melt with elevation during the winter season.

3-09.02 Inasmuch as the primary hydrologic application of the evaluation of basin precipitation and snow is the forecasting of residual runoff, it is important to know (1) the accuracy and limitations of point measurements by existing techniques (precipitation gages, measurments of newly-fallen snow, and snow course measurements), (2) the factors affecting the reliability of individual measurements, and (3) the best methods of determining basin amounts from point measurements. The determination of basin precipitation or snow accumulation is usually the most important factor in the water balance, and the accuracy with which it is determined is dependent not only on the accuracy of the point measurement, but also on how well it may be used in estimating the total hydrologic balance of a basin. It should be recognized that new techniques, such as evaluation of precipitation by radar or atmospheric moisture balance may some day cause obsolescence of existing techniques of sampling precipitation. Chapter 4, which presents the hydrologic balances of each of the snow laboratories, presents further data from which the relative importance of the various terms of the water balance may be shown.

3-09.03 Deficiency of measurements by precipitation gages has been shown to result primarily from turbulence, and a knowledge of its effects is important not only in obtaining an estimate of the true precipitation at a point, but also in assessing the reliability of a point measurement as an index of precipitation over a large area. Precipitation gages are much more efficient in catching rainfall than snowfall, but gage deficiencies should be recognized as occurring with precipitation in either form. In general, the primary consideration in locating gages is to eliminate wind turbulence, by selecting locations where natural shelter will reduce mean wind speed to less than about 2 miles per hour at the gage orifice. The use of wind shields is necessary in areas where natural shelter is not available. Capping of the gage by snow may be serious in some instances, but experience at the snow laboratories shows this to be of relatively minor consequence, when gages are serviced weekly or oftener.

3-09.04 Evaluation of basin precipitation from point precipitation can be done for specific periods isopercentually by use of normal annual isohyetal maps. Point values of precipitation should be corrected for known deficiencies of gage catch. Where station coverage is adequate, a weighted average based on ratios of normal annual basin to station precipitation may be used. When storm types exhibit a seasonal trend, varying ratios by months may be applied to account for the affects of changing conditions on the basin-to-point precipitation ratio.

3-09.05 Snow courses exhibit variability in measured values because of the variety of conditions affecting deposition and measurement of snow. The primary factors influencing the accuracy of measurement and representativeness of snow courses are (1) drifting and turbulence, (2) early season melt or non-representative melt, (3) freezing of the snowpack during and after deposition, (4) ground surface conditions within the snow course area, (5) inadequate free-water drainage, and (6) a variety of errors which may be caused by faulty observer techniques. As in the case of precipitation gages, wind is a primary factor in the selection of snow-course sites. An adequate sampling of the range of elevation must be made in order that the varying slope of the snow wedge may be evaluated.

3-09.06 Basin evaluation of the snowpack water equivalent can be made through use of the snow charts described previously. The density of sampling required is dependent upon the relative homogeneity of the basin. Atmospheric circulation during significant winter precipitation in the mountains of western United States is such that the effect of terrain on the distribution of snow accumulation is believed to be fairly uniform from year-to-year within small to moderate-sized areas. Sampling of snow accumulation for conditions of terrain other than elevation may not be warranted. In locating snow courses, the main consideration is to obtain sites which are representative of basin conditions as a whole with regard to both snow accumulation and melt, and which adequately sample the range of elevation within the basin. The atmospheric circulation patterns during storm conditions should be taken into account, so that the combined effect of weather and large-scale terrain features on the site may be representative of the basin. Finally, the characteristics of the local environment should be carefully studied in order to obtain a site which possesses: (1) adequate sheltering from wind, (2) freedom from drifting to or from the site,(3) relative assurances of stability of characteristics over a period of time, (4) gentle, well-drained slopes, (5) a smooth ground surface, free of brush and rocks, and (6) representativeness of melt conditions for the basin as a whole.

3-09.07 Snow courses and precipitation gages both have their place in the evaluation of runoff potential for basin application. Limitations in the accuracy of point measurements are inherent in both. For a complete hydrologic balance, both the total precipitation and the water equivalent of the snowpack must be evaluated. When precipitation alone is used as a measure of runoff potential remaining at a given time, prior runoff may be subtracted from the total water-year runoff to obtain residual runoff. The snowpack represents a direct evaluation of residual runoff remaining in storage, but soil moisture and ground-water deficiencies must be evaluated. Precipitation amounts are particularly suited to early-season runoff evaluation. The water equivalent of the snowpack is suited to late-season evaluation of runoff, because residual errors are proportionally a smaller part of residual runoff for the snowpack than for total precipitation. In many basins, the sparsity of data requires the use of indexes of runoff potential (from either snow courses or precipitation gages), and the data must be handled statistically in order to define relationships. In such cases, the available period of record is an important consideration in the selection of indexes.

3-10. REFERENCES

1/ ALTER, J.C., "Shielded storage precipitation gages," Mon. Wea. Rev., Vol. 65, July 1937, pp. 262-265.

2/ BLACK, R.F., "Precipitation at Barrow Alaska, greater than recorded," Trans. Amer. Geophys. Union, Vol. 35, No. 2, April 1954, pp. 203-206.

3/ CORPS OF ENGINEERS, South Pacific Division, "Development and test performance of radioisotope-radiotelemetering snow-gage equipment," Civil Works Investigation Project CWI-170, May 1955, 74 pp. and appendix.

4/ DIAMOND, Marvin and W. P. Lowry, "Correlation of density of new snow with 700 mb temperature," Res. Paper 1, Snow, Ice and Permafrost Research Establishment, August 1953.

5/ D'OOGE, C.L., "Snowstorm initiation and air temperature," Bull. Amer. Met. Soc., Vol. 36, No. 3, March 1955, pp. 127, 137.

6/ GERDEL, R.W., "Snow studies at the Cooperative Snow Research Project, Soda Springs, California," Annual Report 1943-44 and 1944-45, U.S. Weather Bureau - University of Nevada.

7/ HAMILTON, E.L., "Rainfall sampling on rugged terrain," Tech. Bull. 1096, U.S. Department of Agriculture, Washington, D. C., December 1954, 41 pp.

8/ HELMERS, A.E., "Precipitation measurements on wind-swept slopes," Trans. Amer. Geophys. Union, Vol. 35, No. 3, June 1954, pp. 471-474.

9/ HENDERSON, T.J., "The use of aerial photographs of snow depth markers in water supply forecasting," Proc. West. Snow Conf., 21st Ann. Meet., Boise, Idaho, April 1953.

10/ KOHLER, M.A., "On the use of double-mass analysis for testing the consistency of meteorological records and for making required adjustment," Bull. Amer. Met. Soc., Vol. 30, 1949, pp. 188-189.

11/ KURTYKA, J.C., "Precipitation measurements study," Report of Investigation No. 20, Illinois State Water Survey Division, 1953, 178 pp.

12/ LEE, C.H., "Precipitation and altitude in the Sierra," Mon. Wea. Rev., Vol. 39, 1911, pp. 1092-1099.

13/ MARR, J.C., "Snow surveying," Misc. Pub. No. 380, U.S. Dept. Agric., Washington, D.C., June 1940, 45 pp.

14/ MERRIAM, C.F., "Progress report on the analysis of rainfall data," Trans. Amer. Geophys. Union, Vol. 19, Part I, 1938, pp. 529-532.

15/ MURRAY, R., "Rain and snow in relation to the 1000-700mb and 1000-500mb thicknesses and the freezing level," The Meteorological Magazine, No. 955, January 1952, pp. 5-8.

16/ NAKAYA, Ukichiro, Snow Crystals, Harvard Univ. Press, 1954, pp. 108-116.

17/ SNOW, ICE AND PERMAFROST RESEARCH ESTABLISHMENT, "Bibliography on snow, ice and permaforst," SIPRE Report No. 12, (continuing project; Vols. 1-8 publ. as of July 1955), Corps of Engineers, U.S. Army, Wilmette, Illinois.

18/ SPREEN, W.C., "A determination of the effect of topography upon precipitation," Trans. Amer. Geophys. Union, Vol. 28, No. 2, April 1947, pp. 285-290.

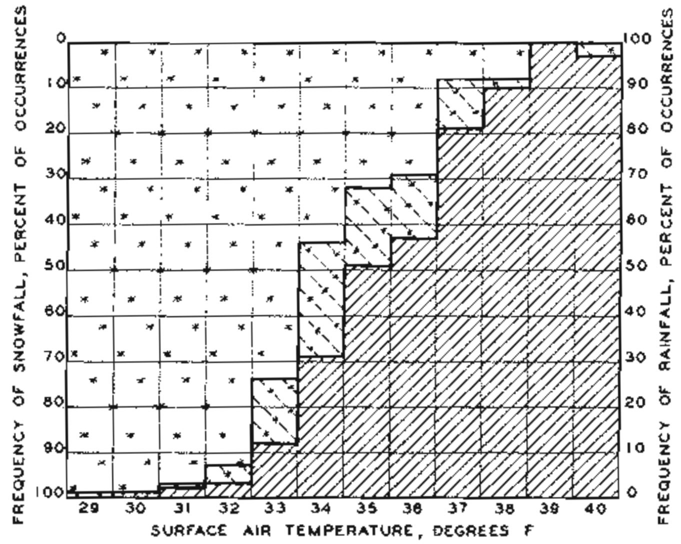
19/ U.S. WEATHER BUREAU, "Operation and Maintenance of storage precipitation gages," U.S. Weather Bureau (Washington, D.C.), 1951.

20/ WARNICK, C.C., "Experiments with windshields for precipitation gages," Trans. Amer. Geophys. Union, Vol. 34, No. 3, June 1953, pp. 379-388.

21/ WEISS, L.L. and W.T. Wilson, "Evaluation of significance of slope changes in double-mass curves," Trans. Amer. Geophys. Union, Vol. 34, No. 6, December 1953, pp. 893-896.

22/ WILSON, W.T., "Discussion of 'Precipitation at Barrow, Alaska, greater than recorded,' by R.F. Black," Trans. Amer. Geophys. Union, Vol. 35, No. 2, April 1954, pp. 206-207.

23/ WILSON, W.T., "Analysis of Winter precipitation observations in the Cooperative Snow Investigations," Mon. Wea. Rev., Vol. 82, No. 7, July 1954, pp. 183-195.



FREQUENCY DISTRIBUTION — RAIN AND SNOW FORMS OF PRECIPITATION

FIGURE 1

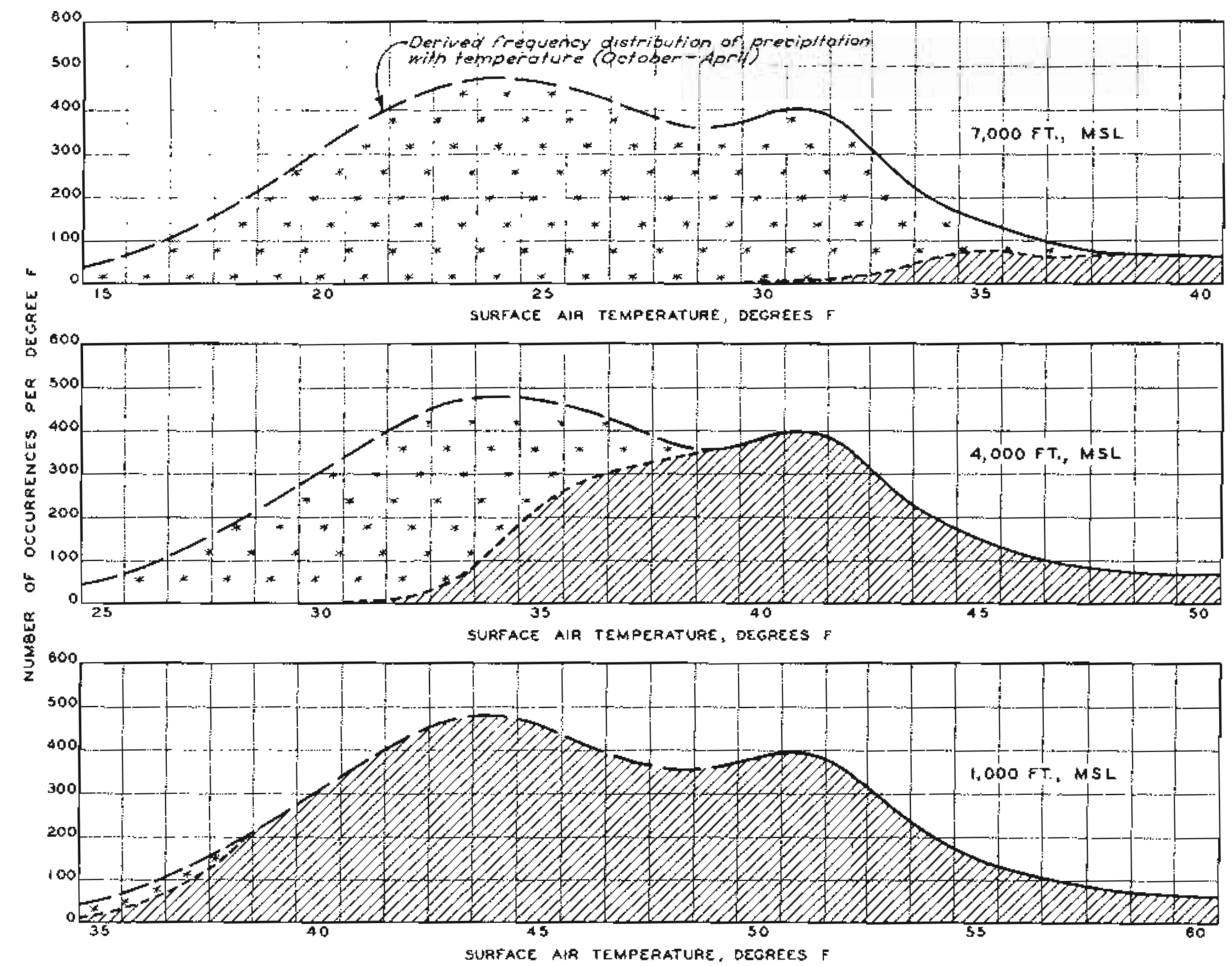
NOTES:

FIGURE 1

1. *Data for this diagram derived from analysis of hourly observations of precipitation and temperature at Donner Summit, California (El. 7200).*
2. *Approximately 2,400 observations of precipitation during the periods October through April, 1946 through 1951 water years, used in obtaining this frequency distribution.*

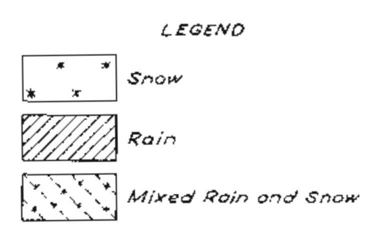
FIGURE 2

1. *Frequency distribution of precipitation for temperature range between 29° and 40°F, for El. 7000, derived from study of rain and snow forms of precipitation at Donner Summit, California. Distribution for temperature range between 15° and 28° estimated by extension of data for White Mountain Research Station, (El. 10,600), reported by Charles L. D'Ooge in Bulletin, American Meteorological Society, March, 1955, and corrected to El. 7000.*
2. *Distribution for El. 4000 and El. 1000 estimated by applying wet adiabatic lapse rate to conditions at El. 7000.*
3. *Separation between rain and snow forms determined by applying average distribution shown in Figure 1.*



SCHEMATIC FREQUENCY DISTRIBUTION OF WINTER PRECIPITATION IN VICINITY OF CSSL

FIGURE 2



SNOW INVESTIGATIONS SUMMARY REPORT

SNOW HYDROLOGY

FORM OF PRECIPITATION DIAGRAMS

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U.S. ARMY

PREPARED: D.M.R.	SUBMITTED: D.M.R.	TO ACCOMPANY REPORT DATED 30 JUNE 1956
DRAWN: B.W.	APPROVED: D.M.R.	PD-20-25/13

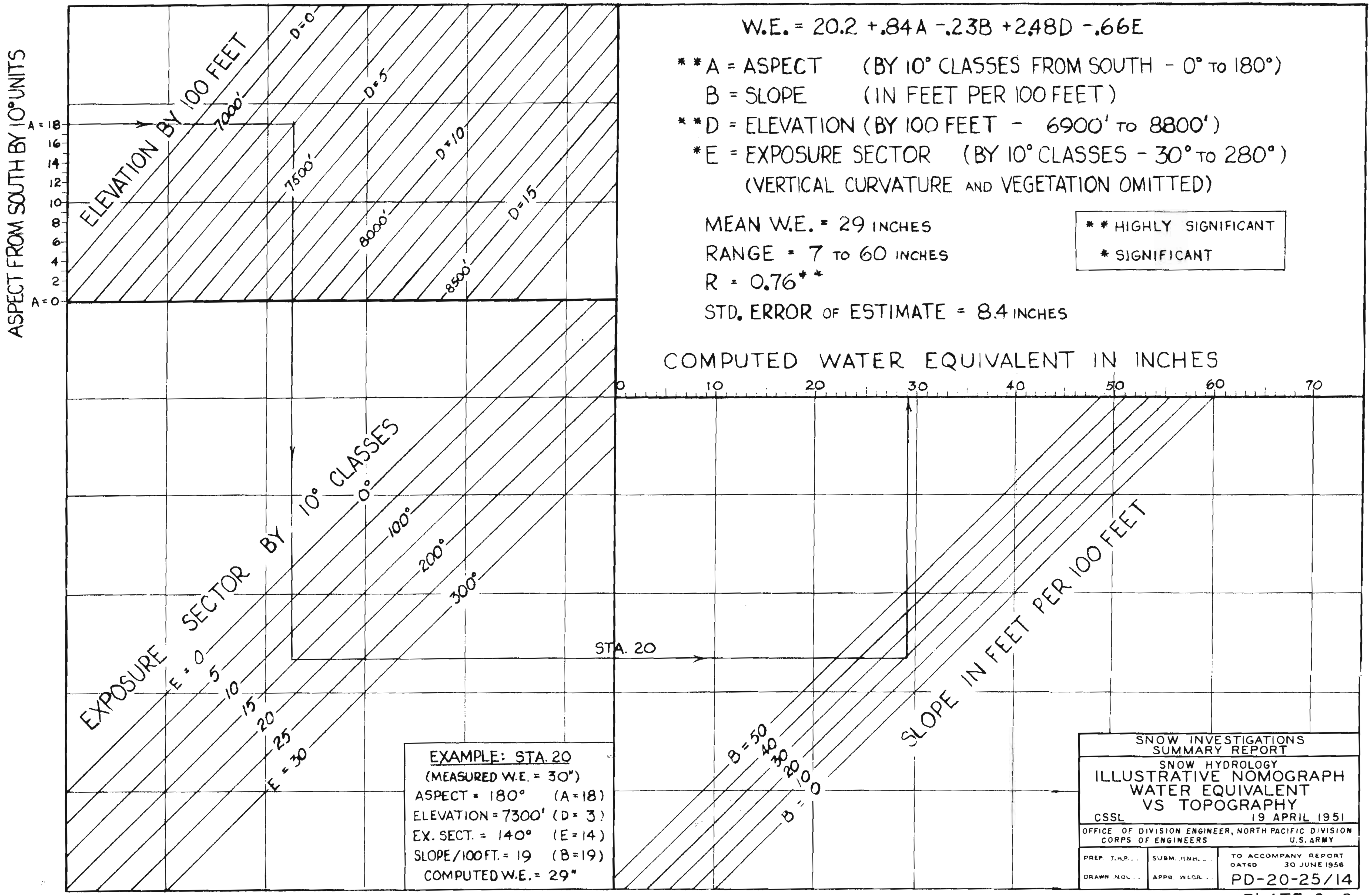
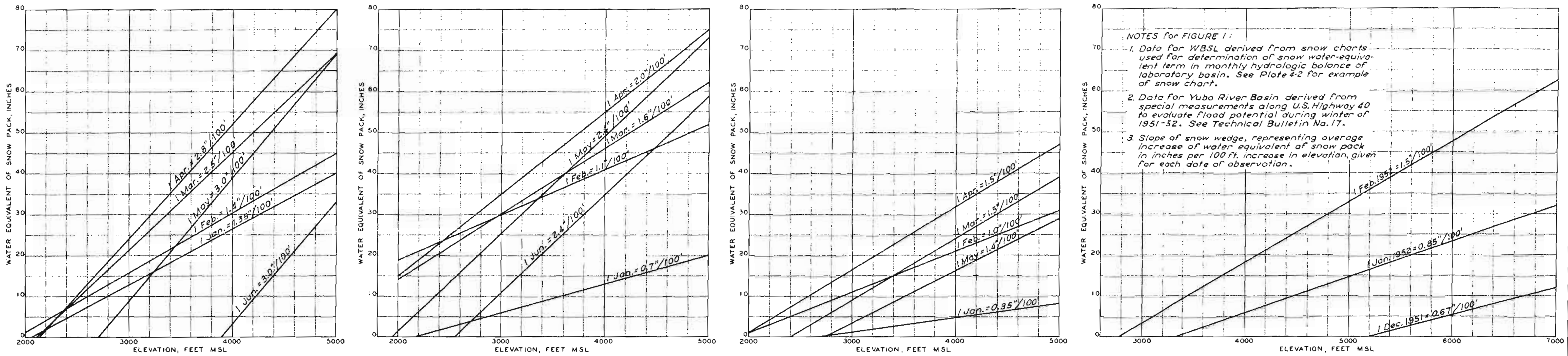

W.E. = 20.2 +.84A -.23B +2.48D -.66E
* * A = ASPECT (BY 10° CLASSES FROM SOUTH - 0° TO 180°)
B = SLOPE (IN FEET PER 100 FEET)
* * D = ELEVATION (BY 100 FEET - 6900' TO 8800')
* E = EXPOSURE SECTOR (BY 10° CLASSES - 30° TO 280°)
(VERTICAL CURVATURE AND VEGETATION OMITTED)
MEAN W.E. = 29 INCHES
RANGE = 7 TO 60 INCHES
R = 0.76* *
STD. ERROR OF ESTIMATE = 8.4 INCHES
* * HIGHLY SIGNIFICANT
* SIGNIFICANT
COMPUTED WATER EQUIVALENT IN INCHES
0
10
20
30
40
50
60
70
ASPECT FROM SOUTH BY 10° UNITS
A = 18
16
14
12
10
8
6
4
2
A = 0
ELEVATION BY 100 FEET
D=0
D=5
D=10
D=15
7000'
7500'
8000'
8500'
EXPOSURE SECTOR BY 10° CLASSES
0°
100°
200°
300°
E = 0
5
10
15
20
25
E = 30
STA. 20
SLOPE IN FEET PER 100 FEET
B = 50
40
30
20
10
B = 0
EXAMPLE: STA. 20
(MEASURED W.E. = 30")
ASPECT = 180° (A=18)
ELEVATION = 7300' (D = 3)
EX. SECT. = 140° (E=14)
SLOPE/100 FT. = 19 (B=19)
COMPUTED W.E. = 29"
SNOW INVESTIGATIONS
SUMMARY REPORT
SNOW HYDROLOGY
ILLUSTRATIVE NOMOGRAPH
WATER EQUIVALENT
VS TOPOGRAPHY
CSSL
19 APRIL 1951
OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U.S. ARMY
TO ACCOMPANY REPORT
DATED 30 JUNE 1956
PD-20-25/14

PLATE 3-2



WBSL — 1949 WBSL — 1950 WBSL — 1951 YUBA RIVER BASIN, CALIFORNIA, 1952

FIGURE 1 — SNOW WATER EQUIVALENT VS ELEVATION

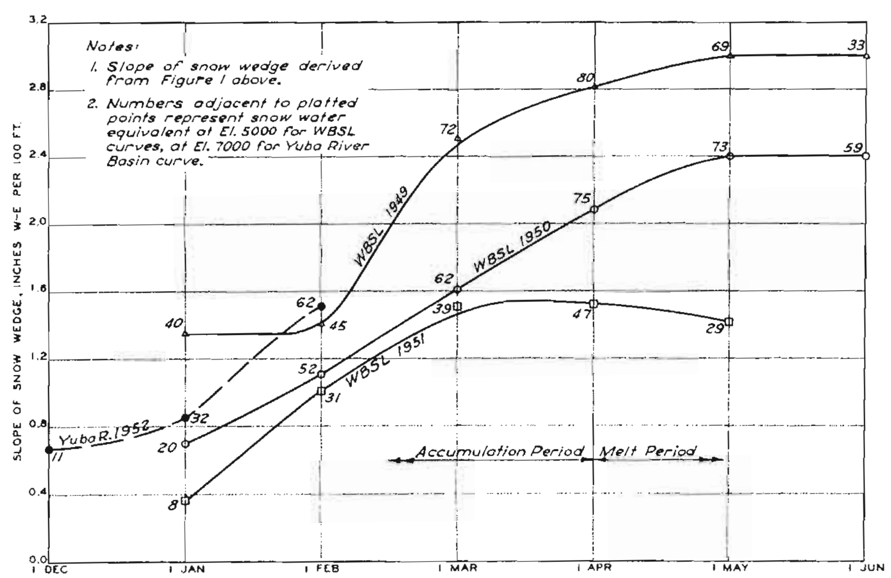


FIGURE 2 — SEASONAL VARIATION IN SLOPE OF SNOW WEDGE

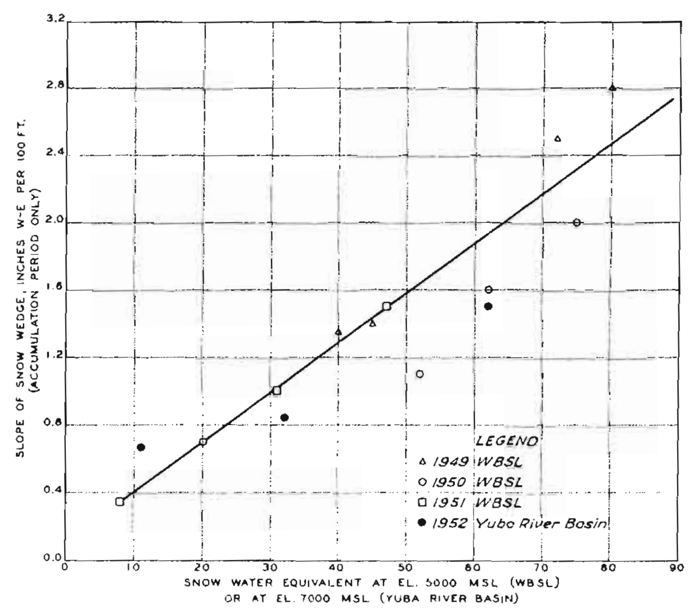


FIGURE 3 — SNOW WATER EQUIVALENT VS SLOPE OF SNOW WEDGE DURING ACCUMULATION PERIOD

SNOW INVESTIGATIONS
SUMMARY REPORT
SNOW HYDROLOGY

ELEVATION EFFECTS ON SNOW ACCUMULATION

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

PREPARED: D.M.R. SUBMITTED: D.M.R. TO ACCOMPANY REPORT DATED 30 JUNE 1956
DRAWN: BV APPROVED: D.M.R.

PD-20-25/15

PLATE 3-3

CHAPTER 4 - WATER BALANCE IN AREAS OF SNOW ACCUMULATION

4-01. INTRODUCTION

4-01.01 General. - In areas of snow accumulation, water stored in the snowpack provides a time delay between precipitation and runoff. Hence, by proper evaluation of the variables affecting runoff, volumetric forecasts of snowmelt runoff may be made several months in advance of the runoff event. Runoff forecasting methods have developed largely on the basis of statistical correlations of runoff with either prior precipitation amounts or snowpack water equivalent. Additional indexes of soil moisture, winter temperature, and winter runoff are also sometimes used. (A discussion of volume-of-runoff forecasting procedures is contained in chapter 11.) In most cases, the period of record for the necessary basic data is too short to ensure an adequate statistical sample of the range in variation and probable long-term mean values. Consequently, there may be serious bias in forecasting procedures derived by statistical methods from data unrepresentative of long-term normals. Such a bias results in unrealistic weighting of the independent variables in the forecasting equation. A knowledge of the water balance for a given area is essential in order to select appropriate forecasting parameters and interpret their reliability. Moreover, the water-balance technique may be used as a forecasting procedure by quantitatively balancing runoff against precipitation, change in snowpack water equivalent, and losses. This procedure is particularly useful for areas where hydrometeorological records are short. In addition, knowledge of the water balance is necessary for rate-of-flow forecasting and for design flood computations.

4-01.02 The principal deterrent to computing water balances for project basins has been the lack of adequate basic information. Scanty areal sampling and unrepresentative measurements of precipitation and accumulated snow, as well as inadequate information on losses, have made water-balance determinations very indefinite. In contrast, the snow laboratories provide relatively dense areal sampling of the prime hydrometeorologic variables. These areas are also relatively free from hydrologic uncertainties involved in computation of losses. One of the primary considerations in the implementation of the snow laboratory program was to provide adequate data for obtaining hydrologic balances in areas of snow accumulation. Such balances for each of the snow laboratories, for its period of complete record (4 to 5 years each), are given in this chapter. This chapter also discusses the individual components of the water balance and the methods of derivation of monthly values for each laboratory.

4-01.03 Even with the dense sampling and better-than-average quality of data for the laboratory areas, errors in measurement of the larger and more important components of the water balance (e.g., precipitation and runoff) may obscure the effect of lesser components. Since the computed areal mean value of any given component cannot be considered to be precisely the "true" value, these computed values must be adjusted to provide the most reasonable balance, using the best information available.

4-01.04 Definition of components. - Because of the varied interpretations of the components of the water balance, each will be specifically defined as used in this report. Observed runoff is defined as the gaged volume of water passing a gaging station on a river or stream. Generated runoff is observed runoff corrected by recession analysis for transitory storage in the soil, ground, and stream channels. Total basin precipitation, in the form of either rain or snow, is defined as that hypothetical precipitation which falls above tree-crown level. Net precipitation is that portion of the total precipitation which reaches the ground or snow surface, after partial interception by the forest canopy (see next paragraph). Since the amount intercepted by vegetation is affected by the form of precipitation, the total precipitation is divided into total snowfall and total rainfall. Interception amounts are computed separately, and values of net snowfall and net rainfall are derived. The water equivalent of the snowpack is exactly what the term implies: the volume of water stored in the snowpack (both solid and liquid forms). Melt is defined as the net decrease in water equivalent of the snowpack after allowance is made for increases as a result of precipitation; thus it excludes water which re-freezes or is retained as liquid water within the snowpack.

4-01.05 Loss is defined as that part of total precipitation which is permanently lost to runoff by evaporation, sublimation, transpiration, and retention as stored soil moisture. (Soil-moisture storage differs from ground or channel storage in that water stored as soil moisture can be removed from the soil only by evaporation and transpiration, whereas water in ground or channel storage is temporarily stored while in transit and will ultimately appear as runoff.) Soil moisture is further subdivided into available and unavailable soil moisture; the latter is not subject to transpiration or evaporation under normal field conditions. In addition to the losses occurring after precipitation has reached the ground or snow surface, other losses result from the interception of precipitation by the forest cover. Interception loss consists of the evaporation and sublimation of intercepted precipitation from vegetation surfaces. It therefore represents the net difference in precipitation within a large forest opening and that received beneath the tree crowns. Since precipitation measurements are generally made in the open, areal interpretation of precipitation or snowpack measurements must consider forest effects. To facilitate computations, losses are grouped in three categories: (1) evapotranspiration losses, which include evaporation from the ground or snow surface, transpiration from all types of vegetation, and sublimation of water from snow surfaces to the atmosphere; (2) soil-moisture changes, which are computed on the basis of assumed fixed capacities of the soil to act as a storage reservoir which must be filled before runoff occurs and upon which evapotranspiration may draw; (3) interception loss, which is measured as the net difference between precipitation in the open and precipitation reaching the ground beneath the forest canopy. Delay to flow through ground and channel storage is accounted for by recession analysis. Deep percolation of water in underground channels, not measured as runoff at the gaging stations, is assumed to be negligible for snow laboratories.

4-01.06 Water-balance equations. - The water year selected for the purpose of this report ends on 31 August, at which time the snowpack water equivalent is zero or negligible for the laboratory areas.* Using the terms as defined in the preceding paragraphs, the water balance for a complete water year may be expressed as follows:

$$Q_{gen} = P - L \qquad (4\text{-}1a)$$

where Q_{gen} is the generated runoff, P is the basin precipitation and L is the loss, all expressed in inches over the basin. For basins in which there is some carryover of snowpack water

* This special period was selected to better define the annual precipitation regime for the snow laboratories since, hydrologically, this is the most quiescent time of the year. The conventional October-through-September water year would serve almost as well, however.

equivalent from year to year, or at the beginning or end of any other shorter period during the snow season, the above equation must be expanded to include a snowpack water equivalent term. A water-balance equation for any such area or for any portion of the year would thus be expressed as follows:

$$Q_{gen} = P - (W_2 - W_1) - L \tag{4-2}$$

where $\underline{W_1}$ and $\underline{W_2}$ are the initial and final water equivalents respectively, expressed in inches. Losses, $\underline{L}$, may be subdivided as follows:

$$L = L_i + Q_{sm} + L_{et} \tag{4-3}$$

where $\underline{L_i}$ is the interception loss, $\underline{Q_{sm}}$ is the change in available soil moisture, and $\underline{L_{et}}$ is the loss by evapotranspiration. The basin precipitation term may be subdivided into the following equations:

$$P = P_r + P_s = P_n + L_i = P_{rn} + L_{ri} + P_{sn} + L_{si} \tag{4-4}$$

where the subscripts $\underline{r}$, $\underline{s}$, $\underline{n}$, and $\underline{i}$ refer to rain, snow, net, and interception, respectively. The interception loss may be excluded from total loss provided that total precipitation is replaced by net precipitation. Equation (4-2) may then be written as follows:

$$Q_{gen} = P_{rn} + P_{sn} - (W_2 - W_1) - Q_{sm} - L_{et} \tag{4-5}$$

Since $\underline{P_{sn} - (W_2-W_1)}$ represents melt, the water-balance equation may be written in the following form:

$$Q_{gen} = P_{rn} + M - Q_{sm} - L_{et} \tag{4-6}$$

where $\underline{M}$ is the snowmelt in inches over the basin. Ground and channel storage is not explicitly included as a separate term in the above water-balance equation, since generated rather than observed runoff is used. In water-balance equations where observed runoff is involved, the equation would be changed by the following relationship:

$$Q_{gen} = Q + Q_g \tag{4-7}$$

where $\underline{Q}$ is observed runoff and $\underline{Q_g}$ is change in ground and channel storage. Substituting in equation (4-1a) and transposing terms,

$$Q = P - L - Q_g \tag{4-8}$$

4-01.07 Water-balance components. - The preceding paragraphs have listed and defined the components to be considered and given the equalities to be used in water balances for areas of snow accumulation. The following sections deal with the technical aspects of the measurement of each of the variables, the reliability of measurements, and the general methods used in determining basin amounts from the point measurements. The elements so discussed are precipitation, interception loss, snowpack water equivalent, evapotranspiration, soil moisture, and ground-water storage and runoff.

4-02. PRECIPITATION

4-02.01 General. - The precipitation term in the water balance presents problems not ordinarily encountered in its use in the various index methods of determining basin precipitation amounts. The value of index methods in the determination of precipitation is dependent more upon their reliability as indexes than upon their representation of the true quantity. That is, an index of precipitation is acceptable if its deviation from the true value is a constant proportion of the true value. On the other hand, precipitation values used in a water balance must represent true quantities. A logical advantage claimed for precipitation indexes is that less rigorous treatment is required than if the true quantity were being determined. However, the claim of such an advantage can be justified only if it can be shown that the index is proportional to the actual amount. Unless the true value is known, however, the reliability of an index can be judged only by the results of its use. Runoff values obtained by statistical procedures often show considerable deviations from actual values; therefore, the reliability of the independent variables, including the precipitation index, remains questionable.

4-02.02 Problems of basin evaluation. - The discussion of precipitation in chapter 3 described the two basic problems involved in obtaining estimates of basin precipitation: (1) to obtain accurate values of point precipitation at the gage sites and (2) to estimate basin precipitation from these corrected station values. Deficiencies in gage catch are due primarily to turbulence and snow capping. Corrections for turbulence may be made on the basis of wind measurements, while loss of gage catch due to snow capping may be corrected by double-mass-curve analysis. Derivation of basin precipitation for mountainous regions from corrected station values is usually accomplished by the isopercentual method using a normal annual isohyetal map.

4-02.03 In the absence of a direct means of obtaining the annual basin precipitation, indirect methods may be employed. Using the basic water-balance equation,

$$P = Q_{gen} + L \tag{4-1b}$$

the annual basin precipitation may be computed from observed runoff and estimated losses. Loss estimates may be made from accepted evapotranspiration equations (see section 4-05), and generated runoff is derived from direct observations of runoff by means of recession analysis (see section 4-07). In this method, there is no independent check on the water balance as a whole. On basins where precipitation is large in comparison with losses, the method provides reasonably accurate values of annual precipitation from which normal relationships between basin and station precipitation may be computed.

4-02.04 For basins where the quality of precipitation data is poor because of gage-catch deficiencies or poor areal coverage, true mean basin precipitation may be estimated by the use of adjustment factors derived from other components in the water balance. For example, during periods when precipitation is known to be in the form of snow over the entire basin and there is little or no loss or snowmelt, the basin precipitation must be the same as the change in the basin snowpack water equivalent. Computations based on monthly values will show seasonal variations in adjustment factors for converting observed station values to basin means. This would represent a refinement of the use of a constant adjustment based on normal annual values. The use of average relationships to relate station to basin precipitation, based on the average seasonal or annual point-to-basin relationship, was discussed in section 3-06. The use of average relationships is valid in areas where precipitation distribution is normally characterized by a relatively fixed pattern, as in the mountains of western United States.

4-02.05 Form of precipitation. - In the computation of monthly water balances, the basin precipitation must be separated into rainfall and snowfall. In areas where precipitation is predominantly in the form of snow throughout the winter period (as in the case of UCSL), this is a relatively minor problem. Occurrences of both rain and snow during the same month are usually confined to the fall and spring months. In the case of WBSL, however, the separation between rain and snow is a critical problem throughout the year except during summer, when precipitation amounts are so small that they are inconsequential in the annual total. Paragraph 3-02.03 described relationships which can be used in separating the rain and snow forms of precipitation.

4-03. INTERCEPTION LOSS

4-03.01 General. - The snow courses and precipitation gages, by which the hydrologist aims to sample basin precipitation, are usually located in the open and thus give no measure of the precipitation reaching the ground in that area of the basin which has a vegetation canopy. Part of the precipitation reaching tree-top level is intercepted by the vegetation surfaces and returned to the atmosphere by evaporation. This part of the basin precipitation is called interception loss, and refers to a permanent loss of precipitation to runoff; thus, it does not include temporarily stored water or snow which later reaches the ground by falling from overloaded branches or by melting and dripping. Although studies of interception loss usually concern trees or shrubs, even low-growing herbaceous plants may reduce the quantity of precipitation reaching the ground surface.

4-03.02 Interception loss warrants careful consideration because of its importance in the water balance. For example: for a moderately dense coniferous forest in an area with annual precipitation of thirty to fifty inches, conservative values of snowfall interception loss usually range between 15 and 30 percent of the total winter precipitation. Loss rates for summer rain usually range between 20 to 40 percent of the summer precipitation. Closer study of amounts of interception loss for different areas shows a wide range due to variation in the type and density of vegetation cover and in the type, magnitude, intensity, and frequency of storms. Consequently, interception-loss percentages should be chosen from studies done for areas of similar vegetation and climate. In water balances prepared without considering interception loss, the resulting error has been obscured by the use of unrepresentatively low precipitation values uncorrected for gage deficiencies. There is, however, increasing recognition of the advantages to be derived from using corrected precipitation data (i.e., data not affected by gage-catch deficiencies and unrepresentative gage or snow-course location). Consequently, there is an increased need to take account of interception loss in the water balance. Since studies of interception loss are now an integral part of present-day research in watershed management, it may be expected that more information on interception loss will be available in the near future.

4-03.03 Interception terminology. - In terms of the effect of vegetation on precipitation reaching the ground surface, total precipitation is divided into throughfall (precipitation which reaches the ground either by falling through spaces between branches and leaves or by dripping from vegetation surfaces), stemflow (precipitation which reaches the ground by flowing down stems after having been temporarily intercepted), and interception loss (intercepted precipitation which is returned to the atmosphere and does not reach the ground). Temporarily intercepted precipitation is referred to as drip and is part of the throughfall.

4-03.04 Interception loss is commonly expressed in terms of the percentage of loss to total precipitation for the selected time unit. The use of loss-percentage for relating measured interception loss to precipitation for a given season or water year has the disadvantage of being limited in application. A loss-percentage can be accurately applied only to the particular study area or to areas with closely similar climatic regimes as well as similar vegetation cover, as was previously mentioned. Presently available studies show an approximately linear relationship between loss and storm size for storms above a minimum size of about one-half inch. Below this minimum storm size, interception loss increases proportionally as storm size decreases. Virtually all of a light shower falling on vegetation is intercepted and evaporated. More useful forms of expressing findings on interception loss, such as the average amount of loss for storms in selected size ranges, are subsequently given herein.

4-03.05 In theory, interception loss is considered to be the difference between precipitation reaching tree-top level and that reaching the ground surface; in practice, however, interception loss is usually the measured difference between precipitation catch in an open area and catch beneath the vegetation canopy. These two differences are not necessarily synonymous. Differences in deposition are added to the difference in catch due to interception loss. A major factor in deposition difference is the difference in wind velocities in the open and in the forest. Environmental differences in gage-catch deficiency may aggravate the deposition difference and thus further obscure the true interception loss. (See chapter 3 for other factors in the micro-environment of the precipitation gage which influence precipitation-gage catch.) In measuring interception loss as the difference between gage or snowboard catch in the open and catch under canopy, the resulting values may be too high if stemflow is not also measured. Stemflow amounts to only about one to three percent of precipitation in coniferous trees, but may amount to more than 30 percent for deciduous trees or shrubs. (Values of stemflow are given in the table in par. 4-03.11.) If values of snowpack water equivalent measured at snow courses located in the open and in the forest are used to determine interception loss, there is less initial difference in catch due to environmental differences than there is if precipitation gages are used. However, here the differences in the environment result in differences in melt. Because of the many differences between the environment in the open and under canopy, many of the measured interception data are actually measures of what is more appropriately called "catch difference" than interception loss.

4-03.06 Measuring interception. - The basic instrumentation for measuring interception loss or "catch difference" consists of precipitation gages or snowboards installed beneath the canopy with similar control gages or snowboards located in an adjoining clearing. Where winter rain is negligible, snowpack measurements for sampling points under canopy and in the open may provide all or part of the data. The large possible variations in catch beneath even a single tree crown (due to random concentration of drip or complete shelter by overhanging canopy) necessitates use of many gages to assure representative data. More complete experiments may include: devices to measure stemflow (such as a water-tight collar around the tree trunk with connection to covered collection can); devices to measure throughfall (such as an impervious surface installed on ground beneath vegetation canopy with covered collection can for resulting runoff); and precipitation gages installed at tree-top level.29/ Correlative data for analyzing interception loss should include measurements of canopy density as well as a description of the type of vegetation. Other useful data include: water-surface evaporation, humidity, wind velocity, soil-moisture, runoff, and temperatures of the air, soil, and vegetation surfaces. Comparative studies may use vegetation as a variable by making observations on both untreated plots and treated plots altered by clearing out the forest understory, keeping the ground surface bare around the trees, or cutting trees according to selected densities.

4-03.07 Interception storage. - The major determinants of interception loss are the interception storage capacity of the vegetation and the evaporation opportunity. Interception storage capacity is the maximum quantity of water (or snow) which can be stored on the leaves and branches of a specific type and density of vegetation. It is usually expressed in the same units as precipitation, that is, inches depth of water or water equivalent over the area. Determinations of interception storage are made by analysis of precipitation and interception-loss data. Amounts thus determined for coniferous forests range from 0.01-0.12 inch for rainfall and 0.01 to 0.34 inch (water equivalent) for snowfall. (Interception storage of snowfall is not necessarily three times greater than rainfall—0.34 vs. 0.12—since these maximum values are for different areas.) Few studies give data on interception storage capacity for both snow and rain for the same area. In a study on Sierra Nevada ponderosa pine, 29/ interception storage capacity was determined as 0.09 inch for snow and 0.12 inch for rain. These data are not a conclusive measure of the comparative storage of rain and snow, because the data for snow are for storms designated as snowstorms if only 50 percent of the total precipitation was snow. Interception storage capacity appears to be primarily a function of canopy density. Other important determinants are: branching type (whether branches are essentially horizontal, as in many coniferous trees, or slanting); foliage type (shape and plane of leaves, and whether foliage is evergreen or deciduous); and vegetative type (tree, shrub, or herb) and height.

4-03.08 Canopy density. - Canopy density is probably the single most important parameter in the determination of interception loss, when considering interception loss in the same climatic region. In this report, canopy density refers to the percentage of the forested area which is covered by a horizontal projection of the vegetation canopy. It does not refer to all the area beneath the periphery of a tree canopy unless all the area is sheltered. Until recently, estimating canopy density has been tedious and comparatively subjective. Recently, however, several instruments have been developed which make possible an objective numerical measure of canopy density.28/ 14/ 18/ These canopy-density meters consist basically of a convex silvered glass surface on which a grid may be placed or cast. The instrument is located at the sampling point and levelled, and a reading is made of the number of points on the grid which are shaded by the canopy. Readings may be satisfactorily duplicated by different observers. A meter of this type, called a "ceptometer," has been used in detailed measurements of forest influences on snow accumulation and melt reported on by Ingebo 14/ (see plate 4-11, fig. 6). A present drawback to the use of a canopy-density meter, as well as other methods of estimating canopy density, is the lack of standardization on the size of the solid angle to be included in the measurement.

4-03.09 Basin canopy cover. - Estimates of basin mean canopy cover may be made as follows. First, the proportion of forested area to total basin area is computed, using aerial photographs or large-scale vegetation maps such as the forest-type maps of the U. S. Forest Service. From point measurements or estimates of forest canopy density, a mean canopy density for the forested part of the basin is determined. Basin canopy cover is the product of the percentage of forested area and the canopy density within the forested area (see table 2-1 for examples).

4-03.10 Evaporation opportunity. - Evaporation opportunity, which determines how much of the intercepted precipitation can be evaporated, varies greatly between regions and at different seasons in the same region. Furthermore, evaporation of intercepted precipitation differs in several respects from evaporation from more continuous water or snow surfaces. In the first place, the evaporation of intercepted precipitation can take place only as long as there is intercepted precipitation remaining on the vegetation. Secondly, the environment of water or snow stored on vegetation surfaces differs from that of water or snow resting on the ground. For example, vegetation surfaces have lower albedoes and warm more rapidly than water or snow surfaces. In addition, the intercepted precipitation is freely exposed to air circulation. Estimates of the evaporation of snow intercepted by conifers range from less than 5 to more than 20 times as high as snowpack evaporation (cf. Kittridge17/).

4-03.11 Interception related to storm-type. - Studies of interception loss for single storms show that the total amount of interception loss is closely related to the frequency of occurrence of precipitation-free intervals during the storm periods. For a storm with continuous precipitation, interception loss is limited to little more than the amount of precipitation stored on the vegetation at the end of the storm. During a storm including precipitation-free intervals, the interception loss may be several times as large as the interception storage capacity. This is graphically illustrated by studies by Rowe and Hendrix29/

and Hamilton and Rowe13/, among others. Summarized below are the results of studies which show the disposition of precipitation in areas having forests and shrub growth:

Experimental Area	Bass Lake,* Sierra, Nevada	San Dimas, ** San Gabriel Mts.		North Fork,** Sierra Nevada
Vegetation	Ponderosa pine	Chaparral Evergreen shrubs		Chaparral Deciduous shrubs
Interc. storage capacity (inches)	0.12	0.08		0.03 (winter & fall only)
Length of storm (hours)	70	73	27	23
Number of pcpt.-free intervals	5	12	1	1
Length of pcpt.-free intervals (hours)	28	36	4	1½
Total pcpt. (inches)	3.19 (100%)	3.35 (100%)	2.61 (100%)	3.14 (100%)
Throughfall (inches)	2.76 (86%)	2.69 (89%)	2.20 (84%)	1.87 (60%)
Stemflow (inches)	0.12 (4%)	0.28 (9%)	0.25 (10%)	1.17 (37%)
Interception loss (inches)	0.32 (10%)	0.38 (11%)	0.16 (6%)	0.10 (3%)

* Rowe and Hendrix29/

** Hamilton and Rowe13/

4-03.12 Interception-loss analyses. - There are two basic methods used in determining snow interception, both of which involve differences between measurements made under the forest canopy and at an open site. They are as follows: (1) measuring increments of snowfall (new snow) for individual storms; (2) measuring snowpack (accumulated) water equivalent. The snowfall data may be readily analyzed in terms of volume of snowfall for individual storms, an important variable in interception loss determination. The snowpack data used in interception analysis are usually for the maximum annual snowpack water equivalent. Interception-loss amounts computed by these two methods are not comparable unless it can be assumed that no melt has occurred since the beginning of snow accumulation, or that melt rates during the entire accumulation period are identical in the open and under canopy. In the analysis of interception loss for a series of individual storms, the results are often expressed in terms of the linear regression equation:

$$L_i = \frac{bP}{100} + a \qquad (4\text{-}9)$$

where L_i is the interception loss in inches per storm, P is the precipitation in inches for the storm, a is the interception storage capacity of the vegetation cover, and b is the percentage loss of precipitation during the storm. The effect of canopy cover is inherent in the data; hence, the constants should not be used for areas with canopy cover which is much different from that of the study area. Simple linear equations giving the relationship between canopy density and snowpack accumulation are also used to express interception loss. In this case, the dependent variable gives the snowpack water equivalent under canopy in percent of the snowpack water equivalent in the open, and the independent variable is the canopy density in percent of complete cover. The effect of storm size is inherent in such relationships; hence, the constants should not be used for areas which are climatically different from the study area.

4-03.13 Snowfall and rainfall interception measurements. - A summary is given in table 4-1 of snowfall or rainfall interception loss for different storm amounts and canopy densities. This table summarizes an unpublished study by Munns 25/ of forest influences in the San Bernardino mountains of southern California. These data are for a stand of Jeffrey pine at an elevation of 6000 feet; the density of the stand as a whole is 0.8. Total amounts of precipitation are given in the table for each of the storm-intensity classes to indicate the size of the sample for each class. Interception-loss percentages are expressed by the formula

$$L_i = \frac{P_{open} - P_{canopy}}{P_{open}} \qquad (4\text{-}10)$$

where L_i is interception-loss percentage, P_{open} is precipitation catch in the open, and P_{canopy} is the precipitation catch under canopy. Reference is also made to work of, Johnson 15/ and Wilm and Neiderhof 38/ for evaluation of interception by storm sizes.

4-03.14 Snowpack interception-loss measurements. - A graphical summary of selected data on interception loss as measured by snowpack data is shown in figure 5 of plate 4-11. Interception loss here refers to the difference between snowpack water equivalent under canopy and that in the open, expressed as a percentage of the water equivalent in the open. Most data are for maximum seasonal values of snowpack water equivalent. These percentage-losses are plotted against canopy data. Part of the scatter in the plotted points is due to the difference in the methods used by the various authors in measuring and expressing canopy density. Qualitative expressions of canopy density are shown as a line extending over the probable range of the qualitative term.

4-03.15 The most conclusive information yet available on the influence of canopy cover on snowpack accumulation is from data collected in the upper Columbia River basin by Ingebo for hundreds of snow sampling points intentionally located to sample various conditions of forest cover. A unique feature of the study is that canopy cover data were obtained for each sampling point by the ceptometer, an instrument which gives a numerical measure rather than an estimate of the cover directly above the point (see par. 4-03.08). Additional analyses were made of these data by the Snow Investigations.* Preliminary results of the correlation between canopy density and water equivalent are given below. (X is the canopy density in percent of complete cover and Y is the snowpack water equivalent for the various canopy densities, expressed in percent of the snowpack water equivalent in the open.) Graphical plots of the relationships are shown on figure 6, plate 4-11.

* Basic data and preliminary analyses made available to Snow Investigations Unit through courtesy of the Missoula Research Center, Intermountain Forest and Range Experiment Station.

Year and no. of sample points	Regression equation	S_{yx}	D	r
1951 only (383)				
By individual points	Y = 95.5 - 0.387 X	19.1	0.257	0.507**
By means of 10% canopy-density classes	Y = 96.8 - 0.401 X	2.6	0.996	0.998**
1949, 1950 (340)				
By individual points	Y = 99.9 - 0.359 X	19.1	0.237	0.487**
By means of 10% canopy-density classes	Y = 99.9 - 0.366 X	5.9	0.805	0.897**

** = significant at the 99% level (highly significant)

4-04. SNOWPACK WATER EQUIVALENT

4-04.01 General. - Quantitative values for basin snowpack water equivalent must be used in a water balance. Indexes of basin snowpack water equivalent cannot be used for the same reasons that indexes of basin precipitation have no place in a water balance (par. 4-02.02). The geographical variation in snowpack accumulation over a given area has the same general pattern as the areal variation in total precipitation, since deposition of snowfall and of rainfall are similarly affected by the terrain of the area. In addition, the distribution of the snowpack is affected by factors which have no effect upon basin precipitation. These factors are the difference in forms of precipitation and the variation in melt rates. During the accumulation period, only the deposition effects upon distribution are appreciable, and usually an elevation parameter will adequately express differences in snowpack water equivalent. Data from snow courses which adequately sample a drainage basin with respect to elevation may then be related to basin amounts, providing no consistent bias results from other terrain factors. During the melt period, on the other hand, areal variation in melt rates tends to make snow course data unrepresentative of the basin water equivalent.

4-04.02 Snow chart. - The difficulty of evaluating basin water equivalent favors the use of an index as a means of evaluating the accumulation of snow. However, the validity of the water-equivalent index is questionable, not only for the reasons mentioned in connection with precipitation indexes (par. 4-02.02), but also for the additional reasons in the preceding paragraph. Accordingly, the index methods are generally inadequate to derive a measure of water equivalent which can be checked against independently derived values for the other terms in the water balance. Variation of the snowpack with elevation is a primary consideration when evaluating the snowpack water equivalent during the accumulation season (see chapter 3). The variation due to other terrain factors (e.g., orientation, slope and exposure) is usually less important and tends to be relatively constant from year to year. The snow chart, therefore, which has elevation as one of its ordinates, is an effective means of integrating snow-course measurements into basin mean snowpack water equivalent (see par. 3-08.04). For basins having relatively few snow courses, the difficulty of determining the water equivalent of the various elevation zones reduces the reliability of the results obtained by use of the snow chart. Also, its use is generally confined to areas within which there is a relatively consistent pattern of climatic conditions from year to year.

4-04.03 When using the snow chart to determine the mean snowpack water equivalent of a basin area, the volume represented by a line of best fit with respect to the plotted points can be considered to be a fixed percentage of the true value. The percentage correction factor may be derived from the water balance as a whole. This is most readily done by analyzing periods when precipitation is entirely in the form of snow and when snowmelt and losses are negligible, since basin precipitation can then be compared directly with water-equivalent change. If net precipitation values are used, as distinguished from total precipitation, the correction factor implicitly includes the effect of interception loss on the snow accumulation. (Snow courses are generally located in the open.) Also, since snow courses are usually situated in areas where local terrain favors above-average snow accumulations, values for the correction factor are generally less than unity, ranging from 0.75 to 0.90.

4-05. EVAPOTRANSPIRATION

4-05.01 General. - Part of the water which enters the soil is removed and returned to the atmosphere by evapotranspiration. This loss occurs not only while water is being supplied to the soil, but also as long as stored soil moisture is available (see section 4-06). In addition, evaporation may take place from the snow surface itself as well as from water surfaces and water intercepted by vegetation (condensation may also occur, it being considered negative evaporation). Evapotranspiration, like interception, represents a permanent loss to runoff. Average annual evapotranspiration losses for humid mid-latitude regions range between 15 and 30 inches, with smaller amounts for arid or alpine areas and larger amounts for areas with long growing seasons and an ample water supply during the growing season.

4-05.02 Knowledge of the amount of evapotranspiration loss is important to the water balance of an area in several ways. First, as one of the components of the water balance, it provides a partial check on the other components. This is particularly useful in the evaluation of precipitation. Since net precipitation is the sum of runoff and evapotranspiration loss, an estimated value of net precipitation can be determined in this manner for comparison with the computed value of net precipitation. Such a check is especially useful for a basin with heavy precipitation or with a significant part of the precipitation falling as snow. Here, gage-catch deficiencies or errors in determining basin precipitation from point measurements may go unsuspected if there is no such check. The value of the check results from the relative magnitude of evapotranspiration and precipitation. Because evapotranspiration is usually one of the smaller components in the water balance for areas of significant snowmelt runoff, the errors in computing evapotranspiration are relatively small in comparison with the errors in computing precipitation, the largest item in the water balance. A second way in which a knowledge of the amount of evapotranspiration is important to the water balance is in computing the soil-moisture deficit, as discussed in paragraphs 4-05.11 and 4-06.19.

4-05.03 Evapotranspiration terminology. - Definitions of evapotranspiration differ as to which of the component parts of total evaporation are included. In this report, evapotranspiration is considered to include transpiration by plants, evaporation from soil particles, and evaporation from the snow surface. The other components of total evaporation, not included as evapotranspiration in this report, are interception loss and evaporation from lakes or other water bodies. Transpiration and soil evaporation are included in one term since most experimental data combine them because of the difficulty of measuring transpiration separately from soil evaporation.

4-05.04 Potential vs. actual evapotranspiration. - Potential evapotranspiration is the amount of water which would be lost by transpiration and evaporation if sufficient water were available in the soil at all times to meet the demand. Potential loss is determined by the energy supply, without reference to the water supply. Actual evapotranspiration refers to the actual loss resulting from the combined effects of the demand and the available water supply. On an annual basis, actual loss is almost invariably less than potential loss, since even in areas with high annual precipitation, the summer water supply (precipitation plus stored soil moisture) is usually not large enough to meet the demand throughout the entire summer. In general, there is much less areal variation in heat supply than there is in water supply, particularly in the mountain watersheds where, due to orographic effects, the areal distribution of precipitation is characterized by large variations. As a result, areas with large ranges in normal annual precipitation usually have much smaller ranges in actual evapotranspiration loss. This is especially true of areas where much of the precipitation falls in winter, the time when the potential evapotranspiration loss is at a minimum.

4-05.05 Transpiration. - Transpiration refers to the loss of water in vapor form from living plants. This loss is not to be confused with the evaporation of water from the outer surfaces of the plant (which is termed interception loss); transpiration loss occurs from within the leaves of the plant. Most of the transpiration loss occurs through stomata (very small openings in the lower surfaces of leaves). Water-vapor loss ordinarily occurs only during the daylight hours while the stomata are open. The vapor-pressure gradient is almost always directed outward from the leaves, resulting in loss of water molecules from the leaf. (Because the leaf temperature is usually warmer than the surrounding air during the day, its saturated vapor pressure is greater than that of the air, even for air with 100 percent relative humidity.) Because of the arrangement of cells within the leaf, the internal surface of the leaf is many times larger than the external surface.10/ The diffusion of water vapor through stomata can take place at a high rate. In general, transpiration is a very efficient means of water loss. Botanists have recorded annual transpiration losses of more than 100 inches of water.16/

4-05.06 Soil evaporation. - Unlike transpiration, soil evaporation is limited by the difficulty of moving the water stored in the soil up to the evaporating surface. In transpiration, water is withdrawn by the roots and transported inside the plant up to the evaporating surface in the leaf. In soil evaporation, water must be transported up through the soil to the evaporating surface. Since the permeability of the soil decreases sharply as the water content of the soil decreases, even though a steep vapor-pressure gradient may exist at the soil surface, soil evaporation may be restricted because capillary rise of water in the soil is slow. As a result of this retarding effect of permeability upon capillary rise, evaporation becomes decreasingly effective with increasing distance of the water from the soil surface. Consequently, plant roots usually remove stored soil moisture to a considerably greater depth than soil evaporation alone (see par. 4-06.08).

4-05.07 Evapotranspiration formulas. - Since it is not practicable to install and service the instrumentation necessary to measure evapotranspiration directly in all areas where such data are needed, it must be estimated by means of an appropriate formula. Many formulas have been developed to express the relation between observed evapotranspiration data and the concurrent hydrometeorological conditions. A formula used to compute evapotranspiration amounts in the water balance should meet the following requirements: good agreement with measured quantities; applicability to climate and vegetation of basin area; basic data ordinarily available as to variety and detail; basic time period of one month or less; and, if possible, quick computation. A formula which meets each of these requirements at least moderately well is that of Thornthwaite.33/34/ Whereas other formulas have been shown to reproduce measured loss more accurately for specific sites, these formulas require more data than is ordinarily available. Such formulas include those of Penman26/ (data required: duration of bright sunshine, air temperature, air humidity and wind speed) and Halstead12/ (data required: maximum and minimum air temperature). A promising method of computing loss for large regions, using radiosonde data, is based on the net increase in water-vapor content of the air in passing over a given region. 2/ 3/ This mass transfer method appears practicable only for large regions; it has been applied satisfactorily to regions as small as the Ohio River basin. 5/

4-05.08 Thornthwaite's evapotranspiration method. - From an analysis of the use of water by many kinds of vegetation, Thornthwaite concluded that climate was the principal determinant of evapotranspiration loss and that the type of vegetation and the character of the soil made relatively little difference. Limiting himself to a consideration of the climatic elements for which data are generally available, he found that the potential loss in any area could be evaluated satisfactorily by an empirical formula using only air temperature as a variable. In addition, an adjustment must be made for length of day or number of hours of possible sunshine (which vary with latitude and season). Thornthwaite's method also includes a monthly bookkeeping method by which monthly actual evapotranspiration is obtained by balancing potential loss against supply (precipitation and available soil moisture). The required basic data are latitude of station, monthly mean air temperature, and monthly precipitation. In addition, information on the average storage capacity for available soil moisture within reach of plant roots is required. Thornthwaite suggests that an average value of four inches of water may be used in default of specific local information.

4-05.09 In Thornthwaite's evapotranspiration method, the effect of latitude and season are standardized to a standard month of 30 days with 12 hours of possible sunshine each for convenience of computation. Using the formulas presented in the following paragraph, values of "unadjusted" potential evapotranspiration are computed on the basis of the standardized month. These values are then adjusted for the number of hours of possible sunshine for the given latitude and month.

4-05.10 Thornthwaite's specific formula for computing potential evapotranspiration postulates that evaporation and transpiration vary with temperature as expressed in the general formula*

$$e = ct^{a} \qquad (4-11)$$

where $\underline{e}$ is the monthly potential evapotranspiration in cm, $\underline{t}$ is the monthly mean air temperature in °C, and $\underline{c}$ and $\underline{a}$ are coefficients which relate evapotranspiration to monthly mean air temperature. The coefficients $\underline{c}$ and $\underline{a}$ are both functions of an annual heat index, $\underline{I}$, which is the summation of monthly indexes $\underline{i}$ for the twelve months of the year. The monthly heat indexes are computed by the formula

$$i = (t/5)^{1.514} \qquad (4-12)$$

* Thornthwaite's usage of symbols is followed here; the reader is cautioned that the symbol $\underline{e}$ has been used elsewhere in this report to denote vapor pressure.

where i is the monthly heat index (dimensionless), and t is the monthly mean air temperature in °C. For the range of I from 0 to 160, the exponent a ranges from 0 to 4.25; c varies inversely with I. From the above relations, and incorporating results from his earlier work on temperature-evaporation relations, Thornthwaite derived the following specific formula for potential evapotranspiration:

$$e = 1.6\ (10t/I)^a \qquad (4\text{-}13)$$

The author supplies tables and a graph which make it easy to compute potential evapotranspiration, as given by the above equation, and also to determine from this, the "adjusted potential evapotranspiration." (See Research Note 20.)

4-05.11 To compute so-called actual evapotranspiration, values of adjusted potential evapotranspiration are used in the month-to-month bookkeeping method presented by Thornthwaite. Demand (adjusted potential evapotranspiration) is balanced against supply (precipitation and available soil moisture). An average value of four inches (10 cm) of water is given for the storage capacity for available soil moisture within reach of most plant roots. As long as demand is met by precipitation or precipitation-plus-soil-moisture, actual evapotranspiration is equal to potential evapotranspiration. When precipitation and the water stored in the soil are insufficient to meet demand, actual evapotranspiration is less than potential. When the water stored in the soil has been depleted, it must be recharged (at a rate not exceeding the infiltration-capacity of the soil).

4-05.12 The quality of actual-evapotranspiration values computed on a monthly basis is affected by the within-month variation in temperature or precipitation. This variation is particularly important in the case of precipitation. Even though the total precipitation recorded for the month may be more than adequate to meet the total demand, it may not be available for use throughout the month. For example, most of the monthly total may fall in the last few days of the month, thus satisfying demand for only these days; or it may fall in a few high-intensity storms with little infiltration. The quality of monthly actual evapotranspiration values may be considerably improved by computing the loss on a daily basis. The probable improvement is most significant for months when stored soil moisture at the beginning of the month is not adequate to fill the demand during the month. An inspection of daily precipitation records for such months will usually show a few periods when supply and demand do not overlap. Computation on a daily basis is advisable for such periods.

4-05.13 Several minor modifications in the Thornthwaite method have been suggested for application to snow-covered areas. During the snowmelt period, values computed for potential evapotranspiration may be somewhat low. This is because the air-temperature data on which the formula was based were generally measured over snow-free surfaces; the air temperature in a snow-covered basin is usually measured above a snow surface. Reference is made to Research Note 17 for a discussion of the relation between temperature data measured above a ground surface and above a snowpack surface. In brief, while there is an energy-absorbing snowpack on the ground, air temperature does not represent the incoming energy supply as it does during other times. The potential evapotranspiration loss for snow-covered areas is thus probably greater than that computed by Thornthwaite's method since, for the same insolation, the measured air temperature is not as high as it would be over snow-free ground. Another minor modification which should be made when computing actual evapotranspiration for snow-covered areas is to include snowmelt as a supplementary source of water supply. Thus, precipitation and snowmelt should both be used to satisfy the potential demand before drawing on stored soil moisture. Equal consideration should be given to both the amount of snowmelt during the month (setting the limits to the quantity which could be used) and the mean area of snow cover (limiting the area where the supply is available).

4-05.14 In computing evapotranspiration loss by Thornthwaite's method, certain periods are critical, namely, times when appreciable demand (warm air temperatures) and water supply (precipitation and available soil moisture) overlap. (Hence soil moisture storage capacity is also critical in the determination of the available moisture supply). Soil moisture storage capacity is particularly important for areas with marked summer drought. On the other hand, quality of winter precipitation data is not significant in computing actual evapotranspiration for most areas where snowmelt is important, because winter precipitation usually exceeds the low potential evapotranspiration amounts of winter. The Thornthwaite formula presupposes measurable water use to begin at monthly mean air temperatures above 32°F, so no precipitation data are required for months with mean air temperatures at or below 32°F.

4-05.15 Thornthwaite has pointed out that when the albedo of the vegetation surface is higher or lower than average, the potential loss rate will be lower or higher, respectively.34/ Also, the potential loss rate applies only to closely-growing vegetation. The effects of decreased density and increased exposure cannot be assessed in general terms. In view of the empirical basis of his formula, Thornthwaite advises for any area where it is used that the formula be tested against reliable measured evapotranspiration data. Such comparisons have shown the formula to be satisfactory in many areas.22/

4-05.16 The results of the use of the Thornthwaite formula with data from snow laboratories are not conclusive. Comparing computed evapotranspiration with the net precipitation minus runoff, results were good for UCSL. Since at WBSL the value for precipitation was not independent of loss, agreement between computed and residual values was not significant. At CSSL, while computed potential loss agreed well with the difference between net precipitation and runoff, actual loss (computed using the assumed soil-moisture storage capacity) was considerably lower than this residual. This lack of agreement may be due to incomplete measurement of outflow, as a result of unmeasured deep-percolation (see par. 4-10.08). Good results were obtained using the Thornthwaite method to compute evapotranspiration for the 438-square mile basin of the North Santiam River above Detroit (see Research Note 22).

4-05.17 Thermodynamics of transpiration at WBSL. - An independent evaluation of transpiration in the heavily forested WBSL during active snowmelt was made on the basis of energy-balance computations for periods of local climate (when advection of energy by the airmasses was known to be negligible). For this case, the only external source of heat energy is solar radiation. The measured quantity of insolation may be balanced against the energy used for snowmelt, transpiration, and loss through the atmosphere by longwave radiation. (The energy required for photosynthesis is negligible, having been estimated to be less than 3 percent of the energy absorbed by the tree crowns.24/ Because the amount of snowmelt is a measured quantity and the amount of longwave loss can be estimated from theoretical considerations, the energy required for transpiration may be treated as the residual in the basinwide energy balance. Transfer of heat by convection from the needle surfaces to the adjacent air need not be considered. Considering the area as a whole, it represents merely an intermediate process in the transfer of heat to the snow surface. When dealing with a snow-covered area, the energy balance as outlined above does not involve the negligible changes in the storage of heat in the ground. The energy-balance computations for the WBSL were made for a five-day clear-weather period in May of 1949. The residual energy, expressed in terms of transpired water, represents the potential transpiration rate for that time of year. Details of the study are presented in Supplement to Research Note 19 and summarized below.

4-05.18 The diagram shown on plate 4-12 illustrates schematically the daily mean balance of energy exchange for the snow-covered area of Mann Creek basin, WBSL, for the period 9-13 May 1949. The net allwave energy input of 490 lys per day was divided almost equally between transpiration and melt: 263 lys per day were used directly for transpiration; 227 lys per day were transferred to the snowpack by longwave radiation and convection, resulting in snowmelt. In addition, about 45 lys per day were transferred to the snowpack as a result of the condensation of water vapor transpired from the forest. This heat of condensation constituted a secondary heat supply for melting the snow. The net generated runoff for the period was 1.24 inches per day, representing an energy equivalent of 244 lys per day (heat of fusion of snow approximately equal to 198 cal per inch of resultant melt). The net transpiration loss to the atmosphere was 0.14 inches per day, representing an energy equivalent of 246 lys per day (heat of vaporization of water approximately equal to 1520 cal per inch of water evaporated). The gross transpiration rate, including water vapor condensed on the snow surface, was 0.17 inches per day—the maximum potential transpiration rate for the specified conditions. The potential evapotranspiration rate as computed by Thornthwaite's method for this condition is 0.165 inches per day. From the results of an energy-balance analysis therefore, the potential transpiration rate for this time of year as computed by Thornthwaite's method, appears to be reasonable.

4-06. SOIL MOISTURE

4-06.01 General. - The soil functions as a reservoir, storing water when available to be used during periods when potential evapotranspiration exceeds current supply. Under average conditions the depth of water stored as soil moisture available for use is about four inches.35/ In extreme cases, however, it may be less than one inch or more than 20 inches. Such a wide range in possible amounts makes accurate evaluation of the soil-moisture capacity of individual basins difficult. From the standpoint of computing basin soil-moisture storage capacity, data on soil-moisture storage and movement are inadequate, and empirical values of soil-moisture storage capacity which may be used in the actual evaluation of soil moisture are generally lacking. A brief review of soil and soil moisture is included here in order to assist in the interpretation of available information on soils.

4-06.02 Only a part of all the moisture in the soil is involved in the water balance: the stored soil moisture which can be removed by plant roots and natural evaporation. Since this available soil moisture is not measured directly, even in point measurements, it must be indirectly estimated. Methods by which this can be done are reviewed briefly in the following paragraphs. The terminology used is not that of the soil scientist, but that in most common use in hydrological studies. The discussion is a simplification of the complex interaction of forces controlling soil-moisture movement and content. Reference is made to the text, Applied Hydrology 19/, which contains a survey of the field of soil-moisture theories and a review of soil physics. Reference is made to other texts which discuss more specific aspects of soil moisture with respect to forest soils 21/ 16/ and with respect to the hydrologic cycle. 6/

4-06.03 Soil-moisture terminology. - The term soil is used here in its agricultural or soil-science usage: the surface layer of the earth, adapted by soil-forming processes to support plant life. Soil as thus defined is only the weathered top layer of the total mass of earth materials of concern in soil mechanics. This top layer is the zone from which stored water may be removed by transpiration and evaporation. Soil is made up of (1) a relatively inert "skeleton" of larger unweathered mineral particles, primarily sands and silts; (2) a physically and chemically active part consisting of tiny, plate-like clays, super-clays, and colloids, plus particles of humus; (3) water; (4) gases. In forest soils, the surface layer, consisting of partly decomposed vegetation (litter or duff), is usually at least several inches thick. The soil profile is the vertical section from the surface down to the unaltered parent material. A systematic vertical variation in texture and composition is typical for soils which have been subjected to seasonal variations in heat and water supply. Soils are commonly grouped into texture classes on the basis of the proportion of particles within specified size ranges (for example, sandy loam). The water storage capacity of soil is principally determined by its texture. This storage capacity is, however, affected by other factors such as the chemical activity of the soil particles, the shapes and arrangement of the particles, the proportion of admixed humus (decomposed vegetation), and the stoniness of the soil. Consequently, considerable variation is possible even in the storage capacity of soils of the same texture group.

4-06.04 The part of the soil moisture which is in permanent storage, and which cannot be removed from the soil by plant roots or evaporation under natural conditions, is the water content that exists at the permanent wilting point (commonly abbreviated as PWP). Although terminology varies, the terms wilting percentage and wilting coefficient may be assumed to refer to water content of the soil at which plants wilt beyond recovery. Although there is some variation among different species and for different stages of growth, the PWP is approximately the same for all plants in a given soil. Both plant roots and evaporation processes in the soil exert about the same maximum force to remove water films from soil particles; consequently, the PWP itself for a given soil is not affected by the presence or absence of plant cover.* The water content left in the soil at the PWP is appreciable. It ranges from less than one-half inch to more than two inches of water per foot depth of soil, increasing with increasing fineness in soil texture. The soil moisture in the soil at PWP is held tightly in the soil. In a laboratory, for example, in order to remove this remaining moisture, soil must be heated to a temperature above the boiling point of water for 24 hours. The PWP for a given soil is determined by growing plants under specified conditions. As a rough approximation, it is equal to about half the field capacity or moisture equivalent, discussed in the next paragraph. An approximate measure of PWP, used when laboratory data are available, is the water content when the tension in the soil sample is at 15 atmospheres. 7/21/

4-06.05 The field capacity (or field moisture capacity) is the upper limit to the amount of water which can be stored in the soil. It is the amount of water left in an initially saturated soil with unobstructed drainage after the downward movement of soil moisture has "materially decreased." Field capacity thus includes the soil moisture below the PWP as well as the available soil moisture. Field capacity is hypothetically equivalent to the capillary-moisture-holding capacity of the soil, or to the total amount of water which can be held against the force of gravity under natural conditions. Actually, gravity is only one of the directional forces acting on water in the soil. The total water content of the soil is the net result of all the directional forces or tensions affecting soil-moisture movement at a given time. Field capacity is an arbitrary measure which, like PWP, is widely used because it represents a useful quantity,

* Since plant roots commonly penetrate deeper than does effective soil evaporation, the total quantity of soil moisture removed seasonally is usually considerably larger when a plant cover is present. See 4-06.08.

notwithstanding the fact that it is not a true equilibrium point on the curve of moisture depletion versus tension. For the purposes of the water balance, it may be assumed that, after field capacity has been reached, there is little additional downward movement of soil moisture to the ground-water table. The time required for soil to drain to field capacity is about one to five days or more, being shorter for sandy soils and longer for fine-textured soils. Two to three days is the period commonly accepted for medium-textured soils.21/ The amount of soil moisture at field capacity ranges from about one inch to over four inches of water per foot depth of soil, for very coarse and fine-textured soils, respectively. A laboratory measure of water content, which is approximately equal to field capacity, is moisture equivalent (M.E.). It is the water content remaining when an initially saturated soil sample is centrifuged under standardized conditions. For most fine-textured soils, M.E. is nearly the same as field capacity; for sandy soils, it is lower. An approximate measure of field capacity is the amount of water in a soil sample when the tension in the soil is at one-third atmosphere; for moisture equivalent, one-half atmosphere.

4-06.06 Values of soil moisture. - Empirical values of storage capacity for different soil texture classes are given below. (Storage capacity for available soil moisture is equal to the difference between field capacity and capacity at the permanent wilting point.) The values given provide only rough approximations for any given soil. In using these values to compute basin soil-moisture storage capacity, it should be kept in mind that in many soils the texture of the surface is often very different from that of underlying horizons. All values are in units of inches of water per foot depth of soil.

Texture Class	Field Capacity			Permanent Wilting Point			Available Soil Moisture		
	a/	b/	c/	a/	b/	c/	a/	b/	c/
Sand	1.2	-	-	0.3	-	-	0.9	-	0.5
Fine Sand	1.4	1.6	-	0.4	0.8	-	1.0	0.8	-
Sandy Loam	1.9	-	-	0.6	-	-	1.3	-	-
Fine Sandy Loam	2.6	-	-	0.8	-	-	1.6	-	1-1.5
Loam	3.2	-	-	1.2	-	-	2.0	-	-
Silt Loam	3.4	3.2	-	1.4	1.6	-	2.0	1.6	-
Light Clay Loam	3.6	-	-	1.6	-	-	2.0	-	-
Clay Loam	3.8	-	-	1.8	-	-	2.0	-	1.5-2
Heavy Clay Loam	3.9	-	-	2.1	-	-	1.8	-	-
Clay	3.9	4.4	-	2.5	2.3	-	1.4	2.1	-

a/ Mean values for texture classes from graph "Typical water-holding characteristics of different-textured soils," USDA Yearbook of Agriculture, 1955, p. 120.37/

b/ Computed from percentage weight of moisture data for field capacity or moisture equivalent, Lutz and Chandler, p. 294;21/ specific gravity of dry soils from ASCE Hydrology Handbook, p. 137.1/

c/ ASCE Hydrology Handbook, p. 134.1/

4-06.07 Soil moisture and the water balance. - The measure of soil moisture needed for the basin water balance is the usable storage capacity for available soil moisture. Available soil moisture is usually expressed in inches of water per foot depth of soil. To estimate the total storage capacity for available soil moisture, the amount per unit depth is multiplied by total depth of soil. The essential item for the water balance, however, is not the total depth of stored water but only the part which is within reach of plant roots and soil evaporation, as discussed in the next paragraph. A preliminary step is the estimation of total capacity. Since information on depth of soil is often given in qualitative terms, some commonly used quantitative equivalents are given below. Two groups of figures are given because the same qualitative terms are commonly used in spite of the fact that there is considerable difference in the mean maximum depth of the soil for different areas. In general, forest soils (that is, the soils in present forest areas) are usually not as deep as agricultural soils. The depths given below for forest soils are in use among foresters;21/ those for agricultural soils are average values for agricultural areas in general. Local usage may differ considerably.

Descriptive terminology	Depth of soil in inches	
	Forest soils	Agricultural soils
Very shallow	<6	<12
Shallow	6 - 12	12 - 20
Moderately deep	12 - 24	20 - 36
Deep	24 - 48	36 - 60
Very deep	>48	>60

4-06.08 How much of the total storage capacity will be used is largely determined by the depth to which plant roots penetrate the basin soil. Evaporation from the soil is usually insignificant below the top foot or two, and rarely penetrates further than plant roots. Available information on root depths is incomplete and in part contradictory. In general, water-absorbing roots are concentrated in the top two to three feet of the soil. Pending better information, the following values indicate the range in average normal root depth for major vegetation types:

Vegetation type	Root depth (feet)
Coniferous trees	2 - 5
Deciduous trees and evergreen broad-leafed trees	3 - 6 or more
Evergreen shrubs (chaparral)	2 - 6 or more
Deciduous shrubs	2 - 6
Tall herbaceous vegetation (principally grasses)	2 - 5 or more
Low-growing herbaceous vegetation	1 - 2

Root depth may be restricted by impervious hardpans, high ground-water tables, or the shallow soils often characteristic of steep slopes in mountainous areas. Within the ranges suggested, plants in summer-dry areas commonly have deeper roots than plants in summer-rain areas. The range of average maximum depth for specific types of vegetation is, of course, larger than the above values.

4-06.09 The light, porous forest litter on most forest soils intercepts and retains for later evaporation some of the precipitation which reaches the forest floor. On the other hand, it retards soil evaporation. Although the moisture-holding capacity of the litter is high in terms of its weight, the amount of water involved is usually small. Under average conditions of depth and porosity, the water-holding capacity of the litter is only about 0.1 inch of water. An extremely deep accumulation of very porous litter can hold as much as two inches of water.16/ Litter is not considered to be a source of moisture supply to plant roots; where it is deep, it may be considered an additional component of interception loss.

4-06.10 Measuring soil moisture. - There are two basic methods of measuring the quantity of water in soil: (1) sampling methods which require soil samples for each laboratory analysis and (2) non-destructive methods which measure soil moisture at a given point by means of meters which remain in the soil. If a quantitative measure of the water content is needed, soil-moisture meters must be calibrated by laboratory analysis of the soil at the location and soil depth where the meter unit is to be installed. Meters may, however, serve as indexes of soil moisture without prior establishment of the relation between meter readings and the actual quantity of water in the soil.

4-06.11 Laboratory measurements. - The basic method of laboratory analysis of soil moisture is known as the gravimetric method. It determines the percentage weight of the water relative to the weight of the dried soil. Soil samples are weighed, oven-dried, and re-weighed. The soil-moisture percentage is the ratio of the decrease in weight to the weight of the dried soil. Because the water-holding capacity of soil is usually different for the various horizons or layers within the soil, it is necessary to take samples at intervals throughout the depth of the soil. Soil-moisture percentages may be assumed to be in percentage by weight unless specifically stated otherwise.

4-06.12 The percentage of soil moisture by volume (and consequently, the derived value for inches depth of water per unit depth of soil) is rarely measured directly. Instead, data for percentage soil moisture by weight are used to compute volumetric measures of soil moisture. Because of the considerable range in density of soil, the adjustment for weight per unit volume adds further sources of error. The difficulty in removing and moving samples without compressing the soil is the principal reason for this error. Since natural, uncultivated soils usually have a more open structure than agricultural soils of the same texture class (not having been subjected to compaction), a considerable error in true volume may result from compression of soil samples during removal and transportation to the laboratory.

4-06.13 Field measurements. - Many different types of instruments make measurements of soil-moisture variation without removing samples for each observation. 20/11/ Two types currently in common use are tensiometers and electrical resistance-type soil-moisture meters. Electrical meters in common use measure the range of available soil moisture. Tensiometers are not relevant here because they operate in the limited range of moisture between field capacity and saturation and do not include the range of available soil moisture. The operating principle of the electrical resistance-type meters makes use of the variation in the electrical resistance of a porous non-conductor with its moisture content. The greater the water content, the smaller the electrical resistance. The meters consist of two parts: (1) the sensing unit (or soil unit) which remains in place in the soil and consists of a porous non-conducting block with embedded electrodes and lead wires extending up through the soil surface; (2) the metering unit which measures the electrical resistance of the buried block.

4-06.14 The electrical soil-moisture meters in most common use at present are those developed by Bouyoucos and Mick 4/ and by Colman and Hendrix 6/. In the Bouyoucos meter, electrical resistance is observed by the null balance of a modified wheatstone bridge. The porous non-conductor in the basic Bouyoucos block is plaster of paris (gypsum); later modifications of the basic unit use nylon and plaster of paris. The Colman meter measures resistance by means of a battery-operated alternating-current ohmmeter; dial readings are in resistance. The Colman meter also includes a resistance-type thermometer in the soil unit. Using a separate circuit, readings of resistance for the thermistor element are made by means of the same metering unit. The temperature data are used to convert measured resistance in the porous block to a common temperature base. The porous material used in the Colman meter soil unit is fiberglass; it is sandwiched between monel-metal screen electrodes.

4-06.15 The sensing elements of the electrical soil-moisture meters appears to be the source of most of the disadvantages of the instruments themselves as far as hydrologic application is concerned. The size of the element must not be so large that there is an excessive time lag between the moisture changes in the surrounding soil and in the porous material between the electrodes. On the other hand, it must not be too small to maintain satisfactory contact with the soil throughout the random slight movements due to shrinkage from drying, etc., which occur seasonally. The porous material itself must be durable and should have an adequate range in resistance (or other electrical property) for the range in moisture content encountered in the soil. Although the material must not be so chemically active that it disintegrates in the soil, it must have sufficient "buffer effect" to insure that an accumulation of dissolved salts from percolating soil water will not significantly affect the electrical resistance between the electrodes.

4-06.16 A dozen different soil units were tested under field conditions at CSSL. Descriptions of the soil units used and results of the testing are presented in "A review of soil moisture measuring methods and apparatus" by Gerdel in Miscellaneous Report 2. The testing program was undertaken after it was found that some of the plaster-of-paris Bouyoucos blocks in use at CSSL had disintegrated after less than one year in the soil. Resistance readings from the Colman meter (1947 model) were occasionally erratic, when compared with both the Bouyoucos-block readings and the hydrological conditions. These errors were attributed to the lack of buffer effect in the chemically inert materials of the Colman units and to unsatisfactory contact between the soil and porous material of the block when soil shrinkage occurred. The Bouyoucos-block units gave readings which appeared more satisfactory; however, some of the blocks disintegrated in less than six months. The other experimental units, none of which are commercially available, had various disqualifying disadvantages. Subsequently alterations and improvements have been made in both the Bouyoucos blocks and Colman meters. In 1949, Colman meters were installed at nine stations at UCSL by the U. S. Forest Service as part of a forest-effects research program. On the basis of preliminary analysis, the data from these meters appeared to agree with concurrent hydrological conditions. The less satisfactory performance of the Colman meters at CSSL may be due to the more alkaline soils and to the use of the earlier model of the meter. In summary, the experience at the snow laboratories with electrical soil-moisture meters indicated that none of the meters used could be recommended for field use with reservations.

4-06.17 Calibration of soil-moisture meters. - Electrical soil-moisture meters must be field calibrated to relate meter readings to concurrent soil-moisture content for each soil-sampling site and each level of measurement at each site. Calibration consists of removing separate soil samples from the area adjacent to the meter and making laboratory determinations of the percentage moisture content by weight (gravimetric analysis). Samples of soil and concurrent resistance readings are taken throughout the cycle of saturation and drying. To define the relation between the electrical resistance of the porous block and the soil-moisture content of the soil, soil scientists recommend taking at least 20 sets of measurements.[20] The field moisture cycle cannot be satisfactorily duplicated by saturating and drying a soil sample in the laboratory. The magnitude of errors resulting from improper calibration procedures is discussed by Remson and Fox.[27]

4-06.18 Point-to-point variation in soil moisture makes it difficult to determine true basin soil moisture from actual measurements. Though point measurements may be accurately made, too few samples or a poor sampling plan may result in biased results. The significance of this source of error is proportional to the importance of soil moisture in the water balance as a whole. The areal variation referred to is for the same soil type, at the same depth in the profile. Even larger variation is possible between different soil types in the same area. In general, areal variation in moisture content is greater in uncultivated soils, particularly in mountainous areas, than in cultivated soils. Recent studies of areal-sampling methods and results are summarized in "Soil moisture measurements."[20] The topography and environment of soil-moisture measurement sites, as those of snow courses and ground-water wells, have a strong influence on the measured soil-moisture amounts. To mention a single determinant, the effect of slope on soil moisture is discussed by van't Woudt.[36]

4-06.19 Computation of soil moisture. - For areas where measured values are not available, the amount of available soil moisture at a given time may be computed from hydrometeorologic data. In a method presented by Mather, computations are made on a daily basis using the procedures from Thornthwaite's method of computing monthly amounts of "actual" evapotranspiration by the water-balance bookkeeping method discussed in 4-05.11.22/ The data required are daily precipitation and air temperature and the estimated quantity of usable storage capacity for available soil moisture. Beginning at a time when the amount of stored water in the soil is known—conveniently, either a maximum (field capacity) or a minimum (wilting point) value is used—, the quantity of soil moisture in the soil is computed by maintaining a daily budget of all additions (precipitation and snowmelt) and withdrawals (evapotranspiration). Published data show good agreement between computed and measured quantities of soil moisture.22/35/ In a method presented by Snyder,31/ the changes in storage of available soil moisture are computed as part of a procedure for computing daily runoff and its component surface and ground-water flow. Procedures are presented for analyzing streamflow and precipitation data in order to establish empirically the relationship between the ground-water component of flow and the initial loss to runoff attributable to soil-moisture deficiency. Together with an empirical formula for estimating evapotranspiration, this relationship is used in a method for computing daily runoff. Good agreement was shown between computed and observed amounts of runoff.

4-07. GROUND-WATER STORAGE AND RUNOFF

4-07.01 General. - From the hydrologic viewpoint, runoff may be considered the last phase of the hydrologic cycle and the end product of all that precedes it. Similarly, it is considered the dependent variable in mathematical expressions of the water balance. Runoff measurements are usually regarded as the most accurate of any variable in the water balance. This is because the measurement of runoff, unlike measurements of other variables which sample only points within an area, effectively integrates the entire area from which the measured flow originates. Even so, the measurement of runoff entails uncertainties in the water balance. These result from errors in the measurement of the runoff itself particularly during periods of ice effects, and from corrections for recession flow.

4-07.02 Deep percolation. - Knowledge of inflow into a basin or outflow from a basin through underground channels is, of course, vital to the determination of a water balance. The possibility of ground-water loss or gain should be given early consideration. Although there is no presently available means of directly evaluating deep percolation, water-balance studies for some areas indicate that a considerable quantity of water may pass from the basin through underground channels and emerge at some distance from the basin. On the other hand, in many mountainous regions, the basins appear to be relatively impervious, which is the case at UCSL and WBSL. There is some question as to ground-water outflow at CSSL. In general, it is believed that loss by deep percolation is small for most areas in the mountains of western United States. However, in some areas exceptions occur which completely invalidate water-balance computations for these areas.

4-07.03 Streamflow measurement. - Factors affecting streamflow measurements are well known and reference is made to books on hydrography for description of techniques used in determining streamflow in open channels. Specific reference is made to publications of the U. S. Geological Survey 9/ for details pertinent to establishment and operation of stream gages, and for compilation of basic streamflow data. In areas of snow accumulation, the quality of the discharge record may be adversely affected by the effects of ice in the channel and the gage installation. Special precautions must be taken in order to insure record of acceptable accuracy for these areas. All regularly established gaging stations operated by the Geological Survey are rated as to probable accuracy of measurement, for periods of both high and low flows.

4-07.04 Storage effect on streamflow. - Delay to runoff due to ground and channel storage is a basic hydrologic phenomenon. For the purpose of this report, ground-water storage is defined as the temporary storage of water in the ground, consisting of both the water under hydrostatic pressure and the water in transit through the soil under natural drainage. Direct evaluation of ground-water storage through the use of well records is impractical in mountainous areas because of the wide variability of conditions on a drainage basin. Streamflow-recession analysis provides an indirect means of evaluating both channel and ground-water storage. As previously mentioned, generated runoff is computed by adding the change in ground and channel storage to the observed runoff. Assuming that all inflow to a basin is suddenly stopped, all outflow subsequently passing the gaging station would result from depletion of ground and channel storage. A measure of this recession flow is therefore a measure of ground and channel storage. Thus, for any given period, the generated runoff may be obtained by adding to observed runoff the terminal recession flow volume and subtracting the antecedent recession flow volume.

4-07.05 Recession analysis. - Each of the several components of ground-water and channel storage has a recession flow that is essentially a decay-type curve (that is, a curve that recedes in a manner such that the incremental change in rate of flow is directly proportional to the rate of flow). Such curves may be defined by an equation of the form,

$$q = q_o \, C_r^{\,t} \qquad (4\text{-}14)$$

where q is the flow at time t after the initial flow q_o, and C_r is the recession constant (ratio of the flow on any day to the previous day's flow). The recession constant must be evaluated in the same time units used for t. Decay curves for the recession flow components can also be expressed by an equation of the form.

$$q = q_o e^{-t/t_s} \qquad (4\text{-}15)$$

where e is the base of Naperian logarithms, and t_s is the recession constant known as the "time of storage." In equation 4-15, when t equals t_s, $q/q_1 = 1/e$ (=0.368); hence t_s may be defined as the time required for the flow component to recede to 0.368 of its initial value. Moreover, the slope of the recession curve, dq/dt, at time zero equals $-q_o/t_s$; hence t_s may also be defined as the time required for a tangent to the decay curve at any point to reach zero flow. The relationship between t_s and C_r (of equation 4-14) is given by the equation,

$$t_s = \frac{-1}{\log_e C_r} \qquad (4\text{-}16)$$

The time-of-storage concept is a very useful one in several aspects of hydrology, notably storage routing. It is further considered in a discussion by Snyder.[32] Integration of equations 4-14 and 4-15 gives,

$$S = -q_o/(\log_e C_r) \qquad (4\text{-}17)$$

and,

$$S = q_o \, t_s \qquad (4\text{-}18)$$

where S is the volume of the recession flow component.

4-07.06 The total recession flow for any drainage area can be represented as the sum of two or more decay-type curves as given above. Nominally, one such flow component may represent ground-water discharge, another interflow, and a third, surface or channel runoff. These components can be derived by plotting, on semi-logarithmic paper, the observed recession flow of a basin. Since the equations of the recession components plot as straight lines on semi-logarithmic paper, a tangent to the tail of the recession curve can be drawn, extending back under the observed curve. This tangent represents the ground-water (or most sluggish) flow component. Differences between the observed curve and the ground-water curve can then be read off and plotted on the same sheet of paper and another tangent drawn to the tail of this recession. This process is continued until the residual may be fitted by a single straight line. Usually two or three such lines are sufficient to define the observed recession curve. An analysis of this kind is especially useful in the analytical determination of the volume of water discharged by the recession flow. The volumes of the individual recession components can be determined using equations 4-17 and 4-18 and the total recession volume as the combined sum.

4-07.07 An alternative method of describing a recession curve is by using variable recession constants in equations 4-14 and 4-15. The derivation of the values may be done empirically for each stream during periods of no inflow. Recession curves and curves showing the variation of t_s with discharge for each laboratory basin are shown in chapter 2. These curves were determined by plotting flows on semi-logarithmic graph paper, during times of no inflow, for all available ranges of flows. From these plottings, a single recession curve was derived, utilizing near-maximum slopes for each range in flow. Variations due to unusual storm conditions were ignored. The recession curves so derived represent average conditions over the basin. While some seasonal differences occur, these differences are small.

4-07.08 The use of a single recession curve for all conditions of flow appears to be adequate for the laboratory areas. Since it combines the components of surface flow, interflow, and ground-water flow, it is basically assumed that each component contributes its proportional part to a given flow. When attaching recession curves to hydrographs, care must be exercised to assure that the point of attachment represents the true streamflow recession. For large areas, it may be necessary to separate ground flow recession from surface flow, depending upon the character and relative magnitude of ground-water flow.

4-07.09 The following equation was used go obtain generated monthly runoff (Q_{gen}) from observed monthly runoff volumes:

$$Q_{gen} = Q + Q_{rt} - Q_{ri} \qquad (4\text{-}19)$$

where $\underline{Q}$ is the observed monthly runoff, $\underline{Q_{rt}}$ is the terminal recession volume, and $\underline{Q_{ri}}$ is the initial recession volume, all expressed in inches over the basin area. In order to facilitate these computations, volume-vs-flow curves were derived for relating the remaining runoff volume beneath the recession curve to the flow at the beginning time. This was done by incrementally summing the areas beneath the empirically-derived recession curves to the lowest value of the recession encountered in the analysis. Thus, the curves do not represent the total volume to zero flow.

4-08. WATER BALANCES FOR SNOW LABORATORIES

4-08.01 The three snow laboratories were situated in areas which represent three different climatic types found in the mountainous areas of the western United States. (These climates are described in detail in chapter 2.) In this chapter, water balances are derived for the years of record of each of the three laboratories to further illustrate the climatic differences between areas and to give firm examples of the different amounts and disposition of the precipitation that occurs in each area. Those differences are summarized in the following table which gives the annual values of the water-balance components for each of the laboratories for its period of operation, in inches depth over the drainage basin.

Laboratory (water year)	Total precipitation			Net precipitation			Loss	Runoff
	Rain	Snow	Total	Rain	Snow	Total		
UCSL								
1946-47	21.1	39.0	60.1	17.8	30.8	48.6	13.6	35.0
1947-48	22.2	35.2	57.4	19.1	28.1	47.2	13.7	33.5
1948-49	10.8	32.6	43.4	7.9	25.5	33.4	11.7	21.7
1949-50	22.2	45.4	67.6	18.9	36.2	55.1	14.0	41.1
Mean	19.1	38.1	57.2	15.9	30.2	46.1	13.2	32.9
CSSL								
1946-47	12.9	41.8	54.7	12.3	37.1	49.4	17.2	32.2
1947-48	14.9	57.4	72.3	14.1	50.1	64.2	18.9	45.3
1948-49	10.2	47.5	57.7	9.1	42.5	51.6	18.1	33.5
1949-50	9.4	68.5	77.9	9.0	61.2	70.2	15.0	55.2
1950-51	37.2	55.4	92.6	35.2	48.9	84.1	14.8	69.3
Mean	16.9	54.1	71.0	15.9	47.9	63.8	16.8	47.0
WBSL								
1947-48	69.5	58.2	127.7	58.7	51.0	109.7	18.1*	92.6
1948-49	43.4	72.2	115.6	34.6	63.4	98.0	15.7	82.3
1949-50	56.4	77.5	133.9	46.7	68.2	114.9	17.7	97.2
1950-51	66.4	68.4	134.8	56.9	60.1	117.0	14.3	102.7
Mean	58.9	69.1	128.0	49.2	60.7	109.9	16.4	93.7

* Includes one inch from soil-moisture storage carried over from previous year.

The detailed monthly data for the several years from which this summary was made are given in tables 4-2, 4-3, and 4-4 for UCSL, CSSL, and WBSL, respectively. Graphical presentation of these data is made on plates 4-3 and 4-4 for UCSL, 4-6 through 4-8 for CSSL, and 4-9 and 4-10 for WBSL. Mean monthly precipitation, snowpack, and runoff data for the period of laboratory record are summarized in plate 2-10, together with temperature and radiation data for the three laboratories.

4-08.02 The water balances are not all for the same period of record and for this reason they are not strictly comparable. There were also some differences in the methods employed in computing the balances for the three laboratories.

In general, each of the components of the water balance was computed separately. Adjustments to these computed values were then made, considering the water balance as a whole, to arrive at the adopted values of the components. Details of the methods employed are presented in the sections which follow.

4-09. WATER BALANCE FOR UCSL

4-09.01 General. - The records for Skyland Creek, UCSL, are used for deriving a water balance representing conditions in headwater areas in the upper Columbia River basin. Skyland Creek basin alone was used, rather than the combined Bear and Skyland Creek area, because of the generally better instrumentation and, consequently, the better definition of hydrologic variables in the Skyland Creek area. Monthly mean values of each component in the water balance were computed for the four water years 1946-47 through 1949-50. Since each component has inherent errors in measurement as well as errors resulting from computation of basin amounts from point measurements, adjustments must be made in the computed values in order to arrive at the most logical balance of all components, considering the water balance as a whole. Computations of individual water-balance components were performed insofar as practical by the procedures outlined in the previous sections of this chapter. There were, however, some problems peculiar to Skyland Creek basin which made some modifications necessary. The following paragraphs describe the methods and specific details of computation of each component of the water balance at Skyland Creek, UCSL. The results are given in table 4-2.

4-09.02 Basin precipitation. - Basin precipitation was computed by the isopercentual method described in chapter 3. Normally, all precipitation stations in and adjacent to a basin would be used for computing basin values. However, because wind records were used for making adjustments for gage-catch deficiency, only those stations having anemometers were used in the computation. Stations having wind records selected for use are 1-B, 10, 12, 18, 20, and 24; the locations of these stations are shown in figure 1 of plate 4-1.

4-09.03 Double-mass-curve analysis. - Double-mass curves of precipitation at station 1-B versus precipitation at each outlying station were plotted to check the reliability of the records at the outlying stations. Records at station 1-B are considered highly reliable because the station was regularly attended. On the other hand, outlying stations were attended at infrequent intervals and therefore their records were more subject to errors such as those resulting from gage malfunction or capping of the orifice. It was found that the month-to-month relation between station 1-B and the outlying stations was generally consistent and that there were no significant gage deficiencies which might be attributed to capping or gage malfunction.

4-09.04 Gage-catch-deficiency corrections. - The adjustments made for wind effect on gage catch were based largely on the degree of exposure of the precipitation gage to wind. Wind records at the outlying stations showed only the total miles of wind travel during the intervals between observations; consequently, it was necessary to use the daily records at headquarters to obtain mean monthly speeds at the outlying stations. From the studies on gage-catch deficiencies mentioned in section 3-05, a chart was prepared (fig. 4, pl. 4-2) showing turbulence correction factors for gage-catch deficiencies at various wind speeds and for various mean monthly temperatures at UCSL. Studies for UCSL have indicated that precipitation is almost entirely in the form of snow if the mean monthly temperature is 25°F or less, and that precipitation is largely in the form of rain if the mean monthly temperature is 40°F or greater. Therefore, the gage-catch deficiency for snowfall is indicated by the line for a mean monthly temperature of 25°F on figure 4. This relationship between windspeed and gage-catch deficiency for snow was established from observations reported on in Research Note 21. The line labeled 40°F represents deficiencies for precipitation in the form of rain; it was derived from a study by Wilson.40/ Lines representing gage-catch deficiencies for temperatures between 25°F and 40°F (that is, for various proportions of rain and snow) were drawn by linear interpolation.

4-09.05 Precipitation distribution. - An isohyetal map of mean annual precipitation for the four-year record was drawn for the basin, using the procedure described in chapter 3. The mean annual isohyetal pattern, together with station values, is shown in figure 4, plate 4-1. The isopercentual method was used to obtain annual basin precipitation for each of the years of study. The isopercentual maps which illustrate the year-to-year variations in precipitation pattern, are shown in figures 2, 3, 5, and 6 of plate 4-1. Having derived annual amounts of basin precipitation for each year of study, the monthly amounts of basin precipitation used in the water balance were computed by multiplying the 6-station average precipitation for each month by the ratio of the basin annual to the 6-station average annual amount. Basin snowfall and rainfall are also computed by the above relationship, using the monthly station amounts previously determined.

4-09.06 Snowpack water equivalent. - Basin snowpack water equivalent was computed by using the snow chart described in paragraph 3-08.04 and illustrated in figure 1 of plate 4-2. Where actual measurements were lacking, it was necessary to make estimates of end-of-month water equivalent. Daily snow stake readings and temperatures at station 1-B were used as aids in determining end-of-month values. A chart was prepared for each month, December through June, using stations 1-C, 10, 12, 18, and 20. These snow-course stations were selected on the basis of adequate records, general reliability of measurements, and location of adjacent precipitation gages. A preliminary line of best fit through the points was drawn on each chart. After careful study of the relationship of the individual points to the line, a fixed average relationship was established and the lines were redrawn accordingly. Since the snow courses are located in open areas, the average amount of snow on the courses exceeds the basin snowpack by the amount of the interception loss. In accordance with the snowfall interception loss of 20 percent established for this basin (see next paragraph), the preliminary basin snowpack water equivalent determined from the chart is multiplied by a factor of 0.80 to obtain corrected basin snowpack water equivalent. A sample computation of the basin snowpack water equivalent is shown in figure 1 plate 4-2.

4-09.07 Interception loss. - Approximately 90 percent of the Skyland Creek drainage area is forested; within this forested area, the canopy density is about 80 percent; the basin mean canopy cover is thus about 72 percent. For amounts of snow normally occurring in the Skyland Creek basin, accumulation of snow is approximately 30 percent less under the tree crowns than in the open (par. 4-03.15 and fig. 6, plate 4-11). The net snowfall interception loss over the basin is thus computed to be 21.6 percent (72 percent x 30 percent). A rounded value of 20 percent was adopted as the interception loss for snowfall. Interception of rainfall was determined on the basis of data quoted by Kittredge,16/ (from Munns' "Studies of Forest Influences in California") which are summarized in the following tabulation:

Rain per shower, inches	Percentage interception			
	At base of tree	Under heavy crown	Under light crown	Under edge of crown
0.01	100	100	100	81
0.06-0.10	94	84	68	48
0.11-0.30	74	48	27	5
0.51-1.00	53	33	16	4

Figures showing interception "under light crown" were used to compute interception losses. To facilitate computation, a number of months representing the complete range of monthly rainfall amounts were analyzed (by individual storms) to determine the expectable interception loss for given monthly rainfall totals. Results are shown in graphical form in figure 2, plate 4-2. Relationships of monthly rainfall and interception loss as shown by the graph were then used to obtain monthly amounts of interception loss of rainfall for the 4-year period of study.

4-09.08 Evapotranspiration. - Evapotranspiration losses were computed by the method developed by Thornthwaite (described in section 4-05). Although this method appeared to be the best of the various methods tested, one of the basic assumptions adopted in the method for computing actual evapotranspiration does not appear applicable to this area, namely the assumption of even distribution throughout the month for both precipitation and potential evapotranspiration demand. Accordingly, the computed values were modified to reduce the loss during summer months such that loss would not exceed the difference between available water and measured runoff.

4-09.09 Soil moisture. - As indicated in section 4-06, the quantitative evaluation of soil moisture is difficult. Observations of soil moisture under the snowpack were made at UCSL by use of Bouyoucos blocks and the Colman meter, both of which are electric resistance-type soil-moisture sensing devices. However, the data were not considered reliable enough to be used in the water balance. Therefore, indirect determinations of change in soil moisture were made on the basis of assumed capacity of the soil to hold moisture. A maximum value of four inches was adopted in accordance with that used by Thornthwaite, after checking its applicability to UCSL by computations based on data from other sources. Having established a maximum value for available water, the amount for any given month is calculated in the process of computing actual evapotranspiration losses by Thornthwaite's method.

4-09.10 Computed runoff. - From independent computations of snowpack water equivalent, precipitation, and losses, the water balance equation (eq. 4-2) was used to obtain computed values of generated runoff. Final values of computed generated runoff are entered in table 4-2.

4-09.11 Observed runoff. - Since stream-gage records are considered to be one of the most reliable quantitative measures in the water balance, they may be used as a check on the evaluations of the other components as integrated in the computed generated runoff. As previously pointed out, observed runoff measurements must be corrected for initial and terminal recessions for use in water balance. Figure 3 of plate 4-2 includes a curve for Skyland Creek basin relating the volume of recession flow to the observed discharge in cfs, based on the average recession curve for this area. Volumes are given for flows above an arbitrary base of 2.0 cfs. Generated runoff values computed from observed runoff values (i.e., observed runoff corrected for recession flow) are shown in table 4-2 for comparison with those computed by the water balance method. It will be noted that computed values are not entirely in agreement with observed generated runoff, particularly during the winter months of 1946-47 and spring months of 1950. A comparison of the ratios of monthly runoff values for Skyland Creek to those for other streams in the vicinity indicates that the observed Skyland Creek flow was too low during the winter months of 1946-47. Further substantiation of the low flow is obtained by comparing the 1946-47 flow with that of the following winter, when more runoff occurred even though meteorologic conditions were less conducive to high winter runoff. Similarly, a comparison of runoff from Skyland Creek with runoff from adjacent drainages for the spring months of 1950 shows a marked dissimilarity in runoff distribution, suggesting either an abnormal distribution of runoff for Skyland Creek in that period or the possibility of error in the observed runoff values.

4-09.12 Adopted values of water-balance components. - Because of the lack of agreement between observed and computed runoff values, other values were adopted where necessary, considering the water balance as a whole, to give more logical values of the various components. Changes were confined to the months when computed runoff failed to agree with observed runoff. In such cases a study was made of the hydrometeorological conditions during the month in question, in order to determine which components were incorrect. The computed values are based on procedures that will produce the best over-all results; these procedures will not necessarily give correct values for periods with unusual conditions. In a few instances a study of existing conditions failed to identify the incorrectly evaluated component, and in such cases the figures were arbitrarily changed to effect a proper balance. The outstanding examples of arbitrarily adopted values are those for the spring months of 1950. Although there were indications that the values for observed runoff were incorrect, an examination of the original hydrograph revealed no reason for changing the observed runoff values. Similarly, a recomputation of the snowpack water equivalent values during the melt season gave no indication that they were grossly in error. Accordingly, the adopted values of snowpack water equivalent and runoff were a compromise between the computed snowpack water equivalent and the observed runoff. Adopted values of all components of the water balance are shown beside the computed values in table 4-2, to permit comparison of the computed and adopted values. The adopted monthly water-balance components are shown in graphical form for each water year, 1946-47 through 1949-50, on plates 4-3 and 4-4.

4-10. WATER BALANCE FOR CSSL

4-10.01 General. - Although the methods used in computing the components of the water balance for CSSL are generally the same as those previously discussed for UCSL, the details of the methods differ somewhat due to the different nature of the area and of the data. Some of the components of the water balance were evaluated by methods suited to the hydrometeorological conditions occurring at the time rather than by the more general methods described in sections 4-01 through 4-07. The evaluations of the various components of the water balance are considered reliable, being based on an exhaustive study of the basic data and pertinent field notes. The water years 1945-46 through 1950-51 were selected for study, covering the entire period for which adequate data were available. The procedures used in evaluating the water-balance components are discussed in subsequent paragraphs under appropriate headings. Some of the columns corresponding to the columns in table 4-2 for UCSL and table 4-4 for WBSL are omitted in table 4-3 for CSSL.

4-10.02 Basin precipitation. - Basin precipitation was computed by the method used for UCSL, as described in section 4-02, with minor deviations resulting from differences in basic data. Turbulence correction factors, based on monthly mean temperature and wind speed, were applied from relationships shown in figure 9 of plate 4-5 and summarized here. For periods without melt or rainfall, the amount of precipitation at each station was compared with the increase in snowpack water equivalent at the adjacent snow course. If the two quantities were found to be in agreement, the precipitation data were considered correct and no further adjustment in station precipitation was made. Using the isopercentual method, as in the UCSL study, basin precipitation was computed for each month. The mean annual isohyetal pattern and isopercentual patterns for each year of the study are shown on plate 4-5, together with the turbulence correction-factor chart and the basin map.

4-10.03 Basin snowfall. - The separation of total precipitation into rain and snow was accomplished on a day-to-day basis, using snowboard and precipitation data for the headquarters station. In most cases the separation was clear. In marginal cases where observations of form of precipitation were lacking, the evaluation was made on the basis of air temperature, as previously discussed in section 3-02.

4-10.04 Snowpack water equivalent. - Water equivalent values used in the monthly water-balance computations were based on snow-survey data at 22 snow courses. The following data were used as aids in determining the end-of-month values of snowpack water equivalent at each course: precipitation, temperature, daily snow-stake readings at headquarters, and daily readings of the radioisotope snow gage. The snow chart was not used in computing the basin water equivalent for CSSL. Since about 80 percent of the basin area is within an elevation range of 800 feet, it is apparent that the effects of elevation upon the distribution of snow are minor and are overshadowed by the effects of other terrain features. Because the elevation range was too small to show an unequivocal increase in water equivalent with elevation, and because the areal density of snow courses was high, the basin snowpack water equivalent was based on the mean of the water equivalent depths at all the snow courses. This basic value was adjusted as follows. A comparison of snow-course sites with the basin's average topography and vegetation cover indicated that a 10-percent reduction in the value of mean snow-course water equivalent would approximate the basin snowpack water equivalent during the accumulation season (see following paragraph). During the depletion season, the rate of melt on the snow courses is about 10 percent greater than on the basin as a whole because of the predominance of southerly exposures and open sites at the snow courses. Adjustments to the end-of-month values during the depletion season consisted of determining the monthly ablation of water equivalent from the average of the snow courses, reducing the ablation by 10 percent and recomputing the end-of-month values on the basis of the adjusted ablation amounts.

4-10.05 Interception loss. - Interception losses were computed separately for rainfall and snowfall. The method used for the UCSL water balance, described in paragraph 4-09.07, was also used for this laboratory. The forested area of CSSL covers 40 percent of the total basin area. Since the mean canopy density within the forested area is about 50 percent, the basin canopy cover is about 20 percent. Rainfall interception was computed in a manner similar to that for UCSL and is illustrated in figure 8 of plate 4-5. Snowfall interception was determined largely from snow-course data from CSSL. A comparison was made between water-equivalent data from snow-course sample points located under forest canopy and data from points located in the open. The results of the comparison indicated that the basin forest cover intercepts about 10 percent of the snowfall.

4-10.06 Evapotranspiration. - Computations of evapotranspiration loss for CSSL by Thornthwaite's method (section 4-05) resulted in values which were considerably smaller than the difference between net precipitation and runoff. For the five-year period as a whole, the mean annual evapotranspiration, computed as the difference between net basin precipitation and runoff, was approximately 17 inches. Computed by Thornthwaite's method, the mean annual potential evapotranspiration was 18 inches; but the mean "actual" evapotranspiration was only 10 inches. The climatological regime at CSSL probably results in actual losses greater than those computed by Thornthwaite's method because of the carryover of water stored in the snowpack to the spring and early summer when rainfall is less than the potential demand, and because of the opportunity for loss by evaporation from the snow surface in winter. Estimated monthly values of adopted evapotranspiration are shown in table 4-3 along with values of potential evapotranspiration according to Thornthwaite's method.

4-10.07 Soil moisture. - On the basis of the other components of the water balance, it is estimated that the storage capacity for available soil moisture amounts to about six inches over the CSSL basin. This value seems high by comparison to the four-inch value normally adopted for areas having deeper soil mantles. A part of the assigned value of six inches may be due to ground-water recharge. Castle Creek normally becomes dry early in the summer and ground-water levels continue to drop after than time. As a result, streamflow recession analysis does not properly account for the resulting ground-water deficit early in the fall. Inspection of ground-water and precipitation data shows that about two inches of rainfall are required to raise ground-water levels sufficiently to produce runoff.

Although one of the criteria in the selection of the laboratory area was the absence of losses by deep percolation, the possibility of such losses cannot be ignored.

4-10.08 Analyses of ground-water well data have shown that a considerable amount of water is depleted from the meadows of the laboratory area after the cessation of surface flow at the stream-gaging station below the meadows. Furthermore, as noted by field observations, a number of springs above the meadows furnish an additional supply of water, which, together with that contained in the water table, is lost by either evapotranspiration or deep percolation after the cessation of flow at the stream gage. Although the total volume of the supply is unknown, it may be greater than that which could potentially be lost by evapotranspiration, in which case the excess loss could be accounted for only by deep percolation. With presently available data, the losses of water on the CSSL area cannot be fully accounted for, and the losses attributed to evapotranspiration in the adopted water balance may be assumed to include possible loss by deep percolation. Under such circumstances the six-inch value assigned for available soil-moisture supply would be designated as available soil-moisture supply plus loss by deep percolation.

4-10.09 Observed runoff. - As for most basins, runoff from the CSSL is considered to be one of the most accurately measured components of the water balance. Except for the possibility of deep percolation, the only source of cumulative error in streamflow measurement on this area is leakage from the flume, a structure established to provide proper channel control. Periodic field checks of the structure indicated that about two percent of the total flow may have been unmeasured as a result of flume leakage. Generated runoff was computed from the observed runoff, using the method described in section 4-07, then multiplying the resulting values by a factor of 1.02 to compensate for the estimated flume leakage.

4-10.10 The water balance. - The water balance derived for CSSL comprises both computed and adopted values of the several components, as for UCSL. However, except for the values of precipitation, runoff, and snowpack water equivalent, the computed and adopted values are the same. The above cited components incorporated all the adjustments needed in order to arrive at a proper balance between all components for the basin as a whole. The interception losses determined from preliminary evaluations of rainfall and snowfall were accepted as the adopted amounts without further corrections, even though the proportions of rainfall and snowfall were revised slightly in making the final water-balance adjustments. The tabulated values of interception loss of snowfall thus vary somewhat from the 10 percent of total snowfall value previously established. The total annual loss by evapotranspiration was obtained by subtracting the computed generated runoff from net precipitation. The annual evapotranspiration amount is prorated by months, on the basis of monthly amounts of potential evapotranspiration, precipitation, and available soil-moisture. In addition, some evapotranspiration was assigned to the winter period, the amount depending upon meteorological conditions.

4-10.11 In order to complete the basin water balance, monthly values of net rainfall, net snowfall, melt, and change in water equivalent, must be determined. An over-all balance with the other components of the water balance must be made. Melt was computed by subtracting the net rainfall from the sum of the generated runoff, change in soil-moisture, and evapotranspiration loss. The melt may also be determined independently by algebraically subtracting the change in water equivalent from the net snowfall. The evaluation of melt, then, is dependent mainly upon the accuracy of separation of net precipitation into rainfall and snowfall. Final adjustments were made in the amounts of melt, net rainfall, and net snowfall, to achieve a balance between all the factors involved. The resulting adopted values are given in table 4-3. Adopted values of generated runoff are shown in column 16a of the same table. Graphical plots of adopted monthly values of each component of the water balance are shown on plates 4-6 through 4-8.

4-11. WATER BALANCE FOR WBSL

4-11.01 General. - The water balance for the WBSL was made for the basin as a whole; that is, for the entire drainage of the Blue River above station 14. It thus includes both the Mann and Wolf Creek drainages as well as the contributing area below the confluence of these creeks. Hydrometeorologically, this area differs in several important respects from UCSL and CSSL: a considerable part of the winter precipitation occurs as rain; appreciable snowmelt occurs in most winter months. As a result, heavy winter runoff is typical. The contrast between the WBSL climate and those of UCSL and CSSL is graphically shown on plate 2-10. The opportunity to sample occurrences of rain on snow was a principal reason for selecting WBSL as a snow laboratory. Compared to UCSL and CSSL, WBSL is situated at a relatively low elevation (between 1960 and 5364 feet msl). The frequent occurrence of rain during winter is a result of the relatively low elevation and the dominance of maritime airmasses, resulting from WBSL's nearness to the ocean and location on the windward side of the Cascade Range. In addition to the variation in form of precipitation during winter, there is also a marked variation in amount of precipitation within the basin. An orographic precipitation-distribution pattern characterizes WBSL, an area of extremely rugged terrain, heavy winter precipitation, and considerable elevation range. Precipitation stations at WBSL sample an elevation range of more than 3000 feet. An extremely dense coniferous forest covers WBSL except for a few small areas of meadow or bare rock.

4-11.02 Unfortunately, the same factors that differentiate WBSL from both CSSL and UCSL and made its data unique and desirable (rugged terrain, considerable elevation range, and dense forest), also made the data harder to collect and of poorer quality. Mixed rain and snow further complicate the making of precipitation (and other) measurements. Compared to UCSL and CSSL, the quality of the snowpack water-equivalent data was poor, due to errors in basic measurements ("short cores", especially) and less frequent snow surveys. Furthermore, it was difficult to evaluate the measurements in terms of end-of-month values, both because appreciable melt occurred during the accumulation season and because rain constituted an unknown part of the precipitation occurring between the end of the month and the last preceding snow survey. Hydrometeorological data for WBSL improved considerably in quality and coverage after the first two years of laboratory operation. Water balances were made for the water years 1947-48 through 1950-51. Procedures used in computing the various components are discussed in the following paragraphs. Reference is made to table 4-4 for monthly values of the various water-balance components for the above years.

4-11.03 Basin precipitation. - The isopercentual technique used in the other laboratories for computing basin precipitation was not used for WBSL. The more empirical methods used here were made necessary by the following aspects of WBSL precipitation data: first, precipitation-gage records for the first two years of the four-year period showed obvious irregularities such as evaporation or freezing of gage contents, capping of the orifice, and errors in servicing or records. Furthermore, there were twice as many gages during the last two years of record; consequently, the use of the isopercentual method would not take advantage of the extra gages, since no four-year means would be available. (There were eight gages in the basin during 1947-48 and 1948-49, and 16 gages during the subsequent period). A second reason for not using the isopercentual technique was that most of the adjustments of precipitation records were made by comparison with the record of a single gage, the Friez at station 1B. As a result, the year-to-year variation in precipitation distribution within the basin would be partly masked, thus negating one of the principal purposes of the isopercentual method. A third reason for not using the isopercentual technique was that no adjustment could be made for gage-catch deficiency due to turbulence at individual gages, because of the lack of wind data.

4-11.04 The method used in computing basin precipitation for WBSL is as follows. First, a careful examination was made of the records of all stations with adequate records throughout all four years. (These stations were 1B, 2, 5, 6, 8, and 10.) Double-mass-curve adjustments were made for periods of missing or erratic records and monthly amounts were tabulated. The six stations were fairly well distributed throughout the basin, although there was a bias toward locations at above-average elevation. However, the slight bias toward higher elevation and consequently toward higher precipitation was probably more than compensated by gage-catch deficiencies. The mean of the six stations was used as an approximation of basin mean net precipitation, after a trial balance showed that for three of the four years, the six-station mean closely approximated the sum of basin runoff plus estimated evapotranspiration loss. For the 1950-51 water year, the six-station mean was adjusted (increased by less than five percent) to equal net basin precipitation computed as the sum of runoff plus estimated evapotranspiration loss. The basis for the adjustment was a comparison of annual totals of precipitation for all WBSL stations for their entire period of record. The comparison indicated that in 1950-51, the precipitation catch in the upper part of the basin, in terms of the basin as a whole, was relatively low as compared to the other years. Since the six-station mean was biased toward stations at higher elevations, the adjustment necessary for the 1950-51 water year was considered reasonable. Total basin precipitation for each of the four years was computed by working backward from the six-station mean, assumed to equal basin net precipitation. The difference in both gage-catch deficiencies and interception loss for snowfall and for rainfall made it necessary to evaluate total snowfall and rainfall separately in the evaluation of losses. Lacking specific data to determine gage-catch deficiencies due to wind at individual stations, an arbitrary average correction was used: a 10-percent increase in the observed quantity for snowfall and a 5-percent increase for rainfall. Monthly values of interception loss were computed, using different percentages of loss for snowfall, winter rainfall, and summer rainfall (discussed later in par. 4-11.08).

4-11.05 Form of precipitation. - The procedure used to estimate the proportions of rain and snow in the basin mean net precipitation was as follows. Using the 34° - 35° F surface air-temperature dividing line between snow and rain found for Donner Summit, California (see 3-02.03) and air-temperature data from station 1A (headquarters, WBSL), a curve was derived which relates base-station temperature to the proportion of basin snowfall in basin precipitation. A lapse rate of 3° F per 1,000 feet was assumed in deriving the curve. This curve includes the effect of both the proportion of the basin area within given elevation zones and the normal increase of precipitation with elevation. Thus it gives the proportion of snowfall in basin precipitation rather than the area over which snow is falling relative to the total basin area (see figure 2, plate 4-11).

4-11.06 Basin snowpack water equivalent. - As was done for UCSL, basin mean values of snowpack water equivalent were computed using the snow chart (described in chap. 3). The actual basin snowpack water equivalent was determined by multiplying the basin mean value from the snow chart (referred to hereinafter as the index value) by an adjustment factor representing the ratio of actual to index values. The actual value used in the adjustment factor was determined for WBSL by a preliminary water balance. Unlike UCSL, for WBSL it was not possible to make a direct comparison of basin snowpack accumulation and basin precipitation for periods of 100 percent basinwide snowfall, since few such periods occurred. Interpolation of snow-survey data to determine end-of-month values was made difficult because of the frequent occurrence of rain interspersed with snowfalls. This problem was especially acute because the changing elevation of the snowfall line (that is, the dividing line between rainfall and snowfall) usually fluctuated within the elevation range of the basin. In comparison with UCSL and CSSL, the basic snow-survey data were of generally poor quality; many "short cores" were noted during the first years of operation. Both the selection of stations used in plotting the snow chart and the weighting of the stations varied somewhat from year-to-year because of the greater number of snow courses in the later years. On the whole, the determination of snowpack water equivalent for WBSL was not rigorous; however, in spite of the above-cited weaknesses, it is probably considerably more accurate than most such basinwide snowpack water-equivalent determinations.

4-11.07 Basin snowpack values based on snow courses are measures of total-snowpack rather than net-snowpack values, since snow-survey courses are generally located in the open. However, the snowpack adjustment factor (which relates the water-equivalent index value to the actual basin snowpack water-equivalent value, as discussed in the preceding paragraph) may be derived so that allowance is made for interception loss. This was done for WBSL.

4-11.08 Interception loss. - Interception loss for WBSL was determined separately for snowfall, for winter rainfall, and for summer rainfall. Interception loss of winter rainfall and of snowfall at WBSL was computed as shown on figures 3 and 4 of plate 4-11. For summer rainfall, a constant percentage loss of 35 percent was used, modified from studies of summer-rainfall interception loss in Douglas fir in Washington.[30] Interception loss for WBSL is not as great as might be expected from a consideration of the denseness of the forest cover alone. The frequent occurrence of storms in this area reduces the evaporation opportunity and thereby the interception loss, despite the large interception-storage capacity that exists. Storm frequency is illustrated in the following table which shows the monthly average number of days with precipitation for the four-year period on which the WBSL water balance is based.

Month	Mean no. days with precipitation	Range of days with precipitation
Sept.	9	5-17
Oct.	19	13-25
Nov.	22	15-25
Dec.	24	23-25
Jan.	24	18-31
Feb.	24	22-26
Mar.	25	23-28
Apr.	16	4-28
May	16	10-24
Jun.	11	4-18
July	5	3- 7
Aug.	5	3-11

Interception losses did not enter into the water balance itself for this laboratory since net precipitation was calculated first, based on annual runoff and evapotranspiration. Total precipitation is given only to illustrate the approximate magnitude of the interception loss.

4-11.09 Evapotranspiration loss. - Evapotranspiration losses were computed by Thornthwaite's method 33/, as for UCSL. During the spring and summer months, the contribution of snowmelt as well as rainfall was considered in the month-to-month accounting of available water. A storage capacity for available soil moisture of 5 inches was considered more representative of this area than the 4 inches used at UCSL (and recommended by Thornthwaite for areas where no local information is available).

4-11.10 Computed and observed runoff. - Runoff values were computed from the foregoing water-balance components by means of equation 4-2. For comparison, the observed values of monthly runoff were corrected by means of recession curves to represent monthly generated runoff. These data are presented in columns 16 and 17 of table 4-4.

4-11.11 Adopted values of water-balance components. - The values of computed and observed runoff of table 4-4 differ by the errors in the water-balance computations. Since observed runoff is probably the most accurate of the water-balance components, it seems unlikely that much of the error results from this source. It appears more likely that the errors result from the other components which, unlike runoff, must be estimated from point measurements. Accordingly, these other values of the water balance were adjusted to make computed runoff agree with the observed values of generated runoff. The values so adjusted are designated the "adopted" values and are identified in table 4-4 by the letter a following the column number. As for the other laboratories, the adopted values were based on the most reasonable values of the various elements, considering the water balance and the water year as a whole. These adopted values of the water-balance components are shown graphically in plates 4-9 and 4-10, for each of the four years of record.

4-12. REFERENCES

1/ ASCE, Hydrology Handbook, Manuals of Engineering Practice, No. 28, New York, 1949, 184 p.

2/ BENTON, G. S., and Jack Dominitz, "The use of atmospheric data in the evaluation of evapotranspiration," The Johns Hopkins University, Department of Civil Engineering, Water Vapor Transport Project, Scientific Report No. 3, Baltimore, Maryland, September 1954, 39 p.

3/ BENTON, G. S., and Mariano A. Estoque, "Water-vapor transfer over the North American continent," Journal of Meteorology, Vol. 11, No. 6, December 1954, pp. 462-477.

4/ BOUYOUCOS, G. J., and A. H. Mick, "An electrical resistance method for the continuous measurement of soil moisture under field conditions," Michigan Agricultural Experiment Station Technical Bulletin 172, 1947.

5/ CARNAHAN, R. L., and G. S. Benton, "The water balance of the Ohio River basin for 1949," The Johns Hopkins University, Department of Civil Engineering, Water Vapor Transport Project, Technical Report No. 1, Baltimore, Maryland, March 15, 1951, pp. 9-41.

6/ COLMAN, E. A., and T. M. Hendrix, "The fiberglass electrical soil-moisture instrument," Soil Science, Vol. 67, 1949, pp. 425-438.

7/ COLMAN, E. A., Vegetation and Watershed Management, The Ronald Press, New York, 1953, 391 p.

8/ CONNAUGHTON, C. A., "The accumulation and rate of melt of snow as influenced by vegetation," Journal of Forestry, Vol. 33, pp. 564-569, 1935.

9/ CORBETT, D. M., and others, "Stream gaging procedure," U. S. Geological Survey Water-Supply Paper 888, 1943.

10/ CURTIS, O. T., and D. G. Clark, An Introduction to Plant Physiology, McGraw-Hill, New York, 1950, 752 p.

11/ HAISE, H. R., "How to measure the moisture in soil," in Water, The Yearbook of Agriculture, 1955, U. S. Department of Agriculture, U. S. Gov't. Printing Office, Washington, D. C., 1955, pp. 362-371.

12/ HALSTEAD, M. H., "Theoretical derivation of an equation for potential evaporation," The Johns Hopkins University, Laboratory of Climatology, Micrometeorology of the Surface Layer of the Atmosphere, Interim Report No. 16, 1951, pp. 10-12.

13/ HAMILTON, E. L., and P. B. Rowe, "Rainfall interception by chaparral in California," U. S. Forest Service, California Forest and Range Experiment Station, in cooperation with Division of Forestry, Sacramento, 1949, 43 p.

14/ INGEBO, P. A., "An instrument for measurement of density of plant cover over snow course points," Proceedings, Western Snow Conference, 23rd Annual Meeting, 1955, pp. 26-28.

15/ JOHNSON, W. M., "The interception of rain and snow by a forest of young ponderosa pine," Trans. Amer. Geophys. Union, 1942 Part II, pp. 566-569.

16/ KITTREDGE, Joseph, Forest Influences, McGraw-Hill, New York, 1948, 386 p.

17/ KITTREDGE, Joseph, "Influences of forests on snow in the ponderosa-sugarpine-fir zone of the Central Sierra Nevada," Hilgardia, Vol. 22, No. 1, March 1953, 96 p.

18/ LEMMON, P. E., "The spherical densiometer for estimating forest overstory density," manuscr., Soil Conservation Service, Portland, Oregon (submitted in 1956 for publication in Journal of Forestry).

19/ LINSLEY, R. K., Jr., Max A. Kohler, and Joseph L. H. Paulhus, Applied Hydrology, McGraw-Hill, New York, 1949.

20/ LULL, H. W., and K. G. Reinhart, "Soil-moisture measurement," Occasional Paper 140, Southern Forest Experiment Station, U. S. Forest Service, New Orleans, La., 1955, 56 p.

21/ LUTZ, H. J., and R. F. Chandler, Jr., Forest Soils, John Wiley and Sons, New York, 1946, 487 p.

22/ MATHER, J. R., "Summary and conclusions," in "The measurement of potential evapotranspiration," edited by J. R. Mather, The Johns Hopkins University, Laboratory of Climatology, Publications in Climatology, Vol. VII, No. 1, Seabrook, New Jersey, 1954. pp. 210-217.

23/ MATHER, J. R., "The determination of soil moisture from climatic data," Bull., Amer. Meteorological Soc., Vol. 35, No. 2, February 1954, pp. 63-68.

24/ MILLER, David H., "Snow cover and climate in the Sierra Nevada, California," Univ. of Calif. Publications in Geography, Vol. 11, Univ. of Calif. Press, Berkeley, 1955.

25/ MUNNS, E. N., "Studies in forest influences in California," 1921, 101 p. unpubl. manuscr. on file at California Forest and Range Experiment Station, U. S. Forest Service, Berkeley, California.

26/ PENMAN, H. L., "Estimating evaporation," Trans. Amer. Geophys. Union, Vol. 37, No. 1, February 1956, pp. 43-46. Discussion, pp. 46-50.

27/ REMSON, Irwin, and G. F. Fox, "The displacement of calibration curves for electrical soil-moisture units," Trans. Amer. Geophys. Union, Vol. 36, No. 5, Oct. 1955, pp. 821-826.

28/ ROBINSON, M. W., "An instrument to measure forest crown cover," Forestry Chronicle, Vol. 23, 1947, pp. 222-225.

29/ ROWE, P. B., and T. M. Hendrix, "Interception of rain and snow by second growth ponderosa pine," Trans. Amer. Geophys. Union, Vol. 32, No. 6, December 1951, pp. 903-908.

30/ SIMSON, A. G., "The interception of summer rains by forest cover," U. S. Forest Service, Pacific Northwest Forest Experiment Station, Forest Research Notes No. 5, January 1931, p. 5.

31/ SNYDER, F. F., "A conception of runoff phenomena," Trans. Amer. Geophys. Union, 1939, Part IV, pp. 725-736.

32/ SNYDER, F. F., "Discussion of 'Storage and the unit hydrograph' by C. O. Clark," Amer. Soc. Civil Eng., Vol. 110, 1945, pp. 1419-1488.

33/ THORNTHWAITE, C. W., "An approach toward a rational classification of climate," Geographical Review, Vol. 38, No. 1, 1948, pp. 55-94.

34/ THORNTHWAITE, C. W., "A re-examination of the concept of potential evapotranspiration," edited by J. R. Mather, The Johns Hopkins University, Laboratory of Climatology, Publications in Climatology, Vol. VII, No. 1, Seabrook, New Jersey, September 1954, pp. 200-209.

35/ THORNTHWAITE, C. W., and J. R. Mather, "The water budget and its use in irrigation," in Water, The Yearbook of Agriculture, 1955, U. S. Department of Agriculture, U. S. Gov't. Printing Office, Washington, D. C., 1955, pp. 346-358.

36/ VAN'T WOUDT, B. D., "On a hillside moisture gradient in volcanic ash soil, New Zealand," Trans. Amer. Geophys. Union, Vol. 36, No. 3, June 1955, pp. 419-424.

37/ WATER, The Yearbook of Agriculture, 1955, U. S. Department of Agriculture, U. S. Gov't. Printing Office, Washington, D. C., 1955.

38/ WILM. H. G., and C. H. Niederhof, "Interception of rainfall by mature lodgepole pine," Trans. Amer. Geophys. Union, Part III, pp. 660-666, 1941.

39/ WILM, H. G. and E. G. Dunford, "Effect of timber cutting on water available for stream flow from a lodgepole pine forest," U. S. Dep't. of Agric. Technical Bulletin No. 968, Washington, D. C., Nov. 1948, 43 p.

40/ WILSON, W. T., "Discussion of 'Precipitation at Barrow, Alaska, greater than recorded,' by R. F. Black," Trans. Amer. Geophys. Union, Vol. 35, No. 2, April 1954, pp. 206-207.

TABLE 4-1

PERCENTAGE INTERCEPTION LOSS BY STORM INTENSITY*

	Storm intensity classes in inches/day								
	.01	.02-.05	.06-.10	.11-.30	.31-.50	.51-1.00	1.01-2.00	2.01+	Total
	Percent of total precipitation								
Edge of crown	81	23	48	5	12	4	5	2	4.8
Light crown	100	67	68	27	32	16	14	14	18.6
Heavy crown	100	93	84	48	57	33	29	25	33.8
Base of tree	100	99	94	74	67	53	34	36	47.7
Mean	94	70	74	38	42	26	20	19	26.2
	Percent of rainfall only								
Mean	93	70	74	38	44	28	15	4	26.6
	Percent of mixed rainfall and snowfall								
Mean	100	73	76	39	42	26	22	24	25.7
Total precipitation in open, in inches	0.16	0.84	2.00	11.43	13.80	36.29	29.99	48.13	142.34

*Data are from Munns 25/

TABLE 4-2

UPPER COLUMBIA SNOW LABORATORY

WATER BALANCE BY MONTHS
(Inches depth over basin)

MONTH	TOTAL PRECIPITATION Total (1)	(1a)	Snowfall W.E. (2)	(2a)	Rainfall (3)	(3a)	INTERCEPTION LOSS Snowfall W.E. (4)	(4a)	Rainfall (5)	(5a)	Total (6)	(6a)	NET PRECIPITATION Snowfall W.E. (7)	(7a)	Rainfall (8)	(8a)	Total (9)	(9a)	SNOWPACK WATER EQUIVALENT Cumulative (10)	(10a)	Change (11)	(11a)	MELT (12)	(12a)	EVAPO-TRANSPIRATION 1/ (13)	(13a)	AVAILABLE SOIL MOISTURE Initial (14)	(14a)	Change (15)	(15a)	RUNOFF COMPUTED Generated (16)	(16a)	ACTUAL Gen. (17)	Obs. (18)	MONTH
																			1946 - 47																
SEP	2.56	2.6	0.00	0.2	2.56	2.4	0.00	0.0	0.43	0.4	0.43	0.4	0.00	0.2	2.13	2.0	2.13	2.2	0.00*	0.0	0.00	0.0	0.00	0.2	1.53	1.5	0.00	0.0	0.00	0.0	0.60	0.7	0.28	0.74	SEP
OCT	9.19	9.2	4.55	4.6	4.64	4.6	0.91	0.9	0.52	0.5	1.43	1.4	3.64	3.7	4.12	4.1	7.76	7.8	1.34*	1.3	1.34	1.3	2.30	2.4	0.00	0.0	0.00	0.0	4.00	4.0	2.42	2.5	1.00	0.92	OCT
NOV	11.21	11.2	11.21	11.0	0.00	0.2	2.24	2.2	0.00	0.0	2.24	2.2	8.97	8.8	0.00	0.2	8.97	9.0	9.33*	9.4	7.99	8.1	0.98	0.7	0.00	0.0	4.00	4.0	0.00	0.0	0.98	0.9	0.83	0.67	NOV
DEC	9.21	9.2	9.21	9.1	0.00	0.1	1.84	1.8	0.00	0.0	1.84	1.8	7.37	7.3	0.00	0.1	7.37	7.4	16.48	16.2	9.15	6.8	0.22	0.5	0.00	0.0	4.00	4.0	0.00	0.0	0.22	0.6	0.07	0.65	DEC
JAN	6.67	6.7	6.67	6.6	0.00	0.1	1.33	1.8	0.00	0.0	1.33	1.8	5.34	4.8	0.00	0.1	5.34	4.9	20.64	20.6	4.16	4.4	1.18	0.4	0.00	0.0	4.00	4.0	0.00	0.0	1.18	0.5	0.22	0.55	JAN
FEB	4.05	4.0	4.05	3.8	0.00	0.2	0.81	0.8	0.00	0.0	0.81	0.8	3.24	3.0	0.00	0.2	3.24	3.2	23.50	23.5	2.86	2.9	0.38	0.1	0.00	0.0	4.00	4.0	0.00	0.0	0.38	0.3	0.00	0.44	FEB
MAR	4.20	4.2	3.61	3.6	0.59	0.6	0.72	0.7	0.08	0.1	0.80	0.8	2.89	2.9	0.51	0.5	3.40	3.4	23.78	24.7	0.28	1.2	2.61	1.7	0.00	0.0	4.00	4.0	0.00	0.0	3.12	2.2	1.07	0.45	MAR
APR	2.20	2.2	0.09	0.1	2.11	2.1	0.02	0.0	0.38	0.4	0.40	0.4	0.07	0.1	1.73	1.7	1.80	1.8	21.50	20.9	-2.28	-3.8	2.35	3.9	0.68	0.7	4.00	4.0	0.00	0.0	3.40	4.9	4.99	1.75	APR
MAY	1.35	1.3	0.00	0.0	1.35	1.3	0.00	0.0	0.44	0.4	0.44	0.4	0.00	0.0	0.91	0.9	0.91	0.9	7.20*	6.5	-14.30	-14.4	14.30	14.4	2.63	2.6	4.00	4.0	-1.72	-0.7	14.30	13.4	13.44	12.91	MAY
JUN	4.60	4.6	0.00	0.0	4.60	4.6	0.00	0.0	0.59	0.6	0.59	0.6	0.00	0.0	4.01	4.0	4.01	4.0	0.00*	0.0	-7.20	-6.5	7.20	6.5	3.30	3.3	2.28	3.3	0.71	-1.0	7.20	8.2	6.05	7.30	JUN
JUL	0.69	0.7	0.00	0.0	0.69	0.7	0.00	0.0	0.31	0.3	0.31	0.3	0.00	0.0	0.38	0.4	0.38	0.4	0.00*	0.0	0.00	0.0	0.00	0.0	3.27	2.6	2.99	2.3	-2.99	-2.3	0.10	0.1	1.13	1.86	JUL
AUG	4.18	4.2	0.00	0.0	4.18	4.2	0.00	0.0	0.55	0.6	0.55	0.6	0.00	0.0	3.63	3.6	3.63	3.6	0.00*	0.0	0.00	0.0	0.00	0.0	2.93	2.9	0.00	0.0	0.00	0.0	0.70	0.7	1.44	1.81	AUG
TOTAL	60.11	60.1	39.39	39.0	20.72	21.1	7.87	8.2	3.30	3.3	11.17	11.5	31.52	30.8	17.42	17.8	48.94	48.6	-	-	0.00	0.0	31.52	30.8	14.34	13.6	-	-	0.00	0.0	34.60	35.0	30.52	30.05	TOTAL
																			1947 - 48																
SEP	3.74	3.7	0.00	2.5	3.74	1.2	0.00	0.5	0.50	0.2	0.50	0.7	0.00	2.0	3.24	1.0	3.24	3.0	0.00*	0.0	0.00	0.0	0.00	2.0	2.25	1.3	0.00	0.0	0.00	0.5	0.99	1.2	1.14	1.04	SEP
OCT	6.23	6.2	0.00	3.5	6.23	2.7	0.00	0.7	0.73	0.3	0.73	1.0	0.00	2.8	5.50	2.4	5.50	5.2	0.00*	0.0	0.00	0.0	0.00	2.8	1.39	1.4	0.00	0.5	1.78	0.5	2.33	2.3	1.98	1.57	OCT
NOV	3.53	3.5	3.53	3.4	0.00	0.1	0.71	0.7	0.00	0.0	0.71	0.7	2.82	2.7	0.00	0.1	2.82	2.8	2.50*	2.5	2.50	2.5	0.32	0.2	0.00	0.0	1.78	1.0	0.22	0.0	0.10	0.3	0.25	1.06	NOV
DEC	3.45	3.4	3.45	3.4	0.00	0.0	0.69	0.7	0.00	0.0	0.69	0.7	2.76	2.7	0.00	0.0	2.76	2.7	4.68	5.0	2.18	2.5	0.58	0.2	0.00	0.0	2.00	1.0	0.00	0.0	0.58	0.2	0.16	0.76	DEC
JAN	7.88	7.9	7.88	7.8	0.00	0.1	1.58	1.6	0.00	0.0	1.58	1.6	6.30	6.2	0.00	0.1	6.30	6.3	10.03	10.3	5.35	5.3	0.95	0.9	0.00	0.0	2.00	1.0	0.00	0.9	0.95	0.1	0.24	0.69	JAN
FEB	7.96	8.0	7.96	7.9	0.00	0.1	1.59	1.6	0.00	0.0	1.59	1.6	6.37	6.3	0.00	0.1	6.37	6.4	15.28	16.2	5.25	5.9	1.12	0.4	0.00	0.0	2.00	1.9	0.00	0.1	1.12	0.4	0.40	0.50	FEB
MAR	6.01	6.0	6.01	6.0	0.00	0.0	1.20	1.2	0.00	0.0	1.20	1.2	4.81	4.8	0.00	0.0	4.81	4.8	19.92	21.0	4.64	4.8	0.17	0.0	0.00	0.0	2.00	2.0	0.00	0.0	0.17	0.0	0.00	0.48	MAR
APR	4.64	4.6	2.30	0.5	2.34	4.1	0.46	0.1	0.30	0.5	0.76	0.6	1.84	0.4	2.04	3.6	3.88	4.0	19.94	20.2	0.02	-0.8	1.82	1.2	0.00	0.0	2.00	2.0	0.00	0.0	3.86	4.8	3.66	0.90	APR
MAY	5.22	5.2	0.00	0.2	5.22	5.0	0.00	0.0	0.64	0.6	0.64	0.6	0.00	0.2	4.58	4.4	4.58	4.6	9.80*	9.8	-10.14	-10.4	10.14	10.6	1.78	1.8	2.00	2.0	2.00	2.0	10.94	12.2	12.26	10.19	MAY
JUN	6.90	6.9	0.00	0.0	6.90	6.9	0.00	0.0	0.79	0.8	0.79	0.8	0.00	0.0	6.11	6.1	6.11	6.1	0.00*	0.0	-9.80	-9.8	9.80	9.8	4.24	4.2	4.00	4.0	0.00	0.0	11.67	11.7	10.25	11.72	JUN
JUL	1.50	1.5	0.00	0.0	1.50	1.5	0.00	0.0	0.45	0.4	0.45	0.4	0.00	0.0	1.05	1.0	1.05	1.0	0.00*	0.0	0.00	0.0	0.00	0.0	4.50	4.5	4.00	4.0	-3.79	-3.8	0.34	0.3	1.64	2.47	JUL
AUG	0.54	0.5	0.00	0.0	0.54	0.5	0.00	0.0	0.26	0.3	0.26	0.3	0.00	0.0	0.28	0.3	0.28	0.3	0.00*	0.0	0.00	0.0	0.00	0.0	0.49	0.5	0.21	0.2	-0.21	-0.2	0.00	0.0	0.88	1.41	AUG
TOTAL	57.60	57.4	31.13	35.2	26.47	22.2	6.23	7.1	3.67	3.1	9.90	10.2	24.90	28.1	22.80	19.1	47.70	47.2	-	-	0.00	0.0	24.90	28.1	14.65	13.7	-	-	0.00	0.0	33.05	33.5	32.86	32.79	TOTAL
																			1948 - 49																
SEP	0.85	0.8	0.00	0.0	0.85	0.8	0.00	0.0	0.36	0.4	0.36	0.4	0.00	0.0	0.49	0.4	0.49	0.4	0.00*	0.0	0.00	0.0	0.00	0.0	0.27	0.3	0.00	0.0	0.00	0.0	0.22	0.1	0.68	0.98	SEP
OCT	1.98	2.0	0.00	1.0	1.98	1.0	0.00	0.2	0.43	0.4	0.43	0.6	0.00	0.8	1.55	0.6	1.55	1.4	0.00*	0.0	0.00	0.0	0.00	0.8	1.20	1.0	0.00	0.0	0.00	0.0	0.35	0.4	0.00	0.84	OCT
NOV	7.57	7.6	6.80	6.8	0.77	0.8	1.36	1.4	0.09	0.0	1.45	1.4	5.44	5.4	0.68	0.8	6.12	6.2	4.80*	4.8	4.80	4.8	0.64	0.6	0.00	0.0	0.00	0.0	1.00	1.0	0.32	0.4	0.31	0.59	NOV
DEC	6.26	6.3	6.26	6.3	0.00	0.0	1.25	1.3	0.00	0.0	1.25	1.3	5.01	5.0	0.00	0.0	5.01	5.0	9.80	9.8	5.00	5.0	0.01	0.0	0.00	0.0	1.00	1.0	0.00	0.0	0.01	0.0	0.41	0.59	DEC
JAN	3.37	3.4	3.37	3.4	0.00	0.0	0.67	0.7	0.00	0.0	0.67	0.7	2.70	2.7	0.00	0.0	2.70	2.7	12.59	12.5	2.79	2.7	-0.09	0.0	0.00	0.0	1.00	1.0	0.00	0.0	-0.09	0.0	0.36	0.51	JAN
FEB	8.57	8.6	8.57	8.6	0.00	0.0	1.71	2.2	0.00	0.0	1.71	2.2	6.86	6.4	0.00	0.0	6.86	6.4	18.89	18.9	6.30	6.4	0.56	0.0	0.00	0.0	1.00	1.0	0.00	0.0	0.56	0.0	0.35	0.45	FEB
MAR	3.51	3.5	3.51	2.0	0.00	1.5	0.70	0.4	0.00	0.2	0.70	0.6	2.81	1.6	0.00	1.3	2.81	2.9	20.02	20.2	1.13	1.3	1.68	0.3	0.00	0.0	1.00	1.0	1.00	0.5	0.68	1.1	0.46	0.46	MAR
APR	2.72	2.7	0.26	1.0	2.46	1.7	0.05	0.2	0.43	0.3	0.48	0.5	0.21	0.8	2.03	1.4	2.24	2.2	14.14	14.4	-5.88	-5.8	6.09	6.6	0.79	0.8	2.00	1.5	2.00	2.5	5.33	4.7	4.11	1.26	APR
MAY	4.57	4.6	0.00	3.0	4.57	1.6	0.00	0.6	0.58	0.5	0.58	1.1	0.00	2.4	3.99	1.1	3.99	3.5	4.60*	4.2	-9.54	-10.2	9.54	12.6	2.66	2.7	4.00	4.0	0.00	0.0	10.87	11.0	11.66	9.65	MAY
JUN	1.66	1.7	0.00	0.5	1.66	1.2	0.00	0.1	0.45	0.3	0.45	0.4	0.00	0.4	1.21	0.9	1.21	1.3	0.00*	0.0	-4.60	-4.2	4.60	4.6	3.56	3.6	4.00	4.0	-2.35	-1.8	4.60	3.7	2.89	4.63	JUN
JUL	1.64	1.6	0.00	0.0	1.64	1.6	0.00	0.0	0.46	0.5	0.46	0.5	0.00	0.0	1.18	1.1	1.18	1.1	0.00*	0.0	0.00	0.0	0.00	0.0	2.52	3.0	1.65	2.2	-1.65	-2.2	0.31	0.3	0.45	1.30	JUL
AUG	0.58	0.6	0.00	0.0	0.58	0.6	0.00	0.0	0.28	0.3	0.28	0.3	0.00	0.0	0.30	0.3	0.30	0.3	0.00*	0.0	0.00	0.0	0.00	0.0	0.25	0.3	0.00	0.0	0.00	0.0	0.05	0.0	0.32	0.87	AUG
TOTAL	43.28	43.4	28.77	32.6	14.51	10.8	5.74	7.1	3.08	2.9	8.82	10.0	23.03	25.5	11.43	7.9	34.46	33.4	-	-	0.00	0.0	23.03	25.5	11.25	11.7	-	-	0.00	0.0	23.21	21.7	22.08	22.13	TOTAL
																			1949 - 50																
SEP	3.97	4.0	0.00	0.5	3.97	3.5	0.00	0.1	0.53	0.5	0.53	0.6	0.00	0.4	3.44	3.0	3.44	3.4	0.00*	0.0	0.00	0.0	0.00	0.4	2.67	2.4	0.00	0.0	0.00	0.2	0.77	0.8	0.30	0.74	SEP
OCT	4.01	4.0	1.69	2.5	2.32	1.5	0.34	0.5	0.30	0.2	0.64	0.7	1.35	2.0	2.02	1.3	3.37	3.3	0.00*	1.0	0.00	1.0	1.35	1.0	0.00	0.0	0.00	0.2	2.43	1.4	0.94	0.9	0.94	0.67	OCT
NOV	6.05	6.0	6.05	4.0	0.00	2.0	1.21	0.8	0.00	0.2	1.21	1.0	4.84	3.2	0.00	1.8	4.84	5.0	1.50*	2.5	1.50	1.5	3.34	1.7	0.00	0.0	2.40	1.6	1.67	2.4	1.77	1.1	0.79	0.74	NOV
DEC	11.33	11.3	11.33	11.3	0.00	0.0	2.27	2.3	0.00	0.0	2.27	2.3	9.06	9.0	0.00	0.0	9.06	9.0	10.40	11.4	8.90	8.9	0.16	0.1	0.00	0.0	4.00	4.0	0.00	0.0	0.16	0.1	0.18	0.68	DEC
JAN	13.58	13.6	13.58	13.6	0.00	0.0	2.72	2.9	0.00	0.0	2.72	2.9	10.86	10.7	0.00	0.0	10.86	10.7	21.12	22.1	10.72	10.7	0.14	0.0	0.00	0.0	4.00	4.0	0.00	0.0	0.14	0.0	0.00	0.42	JAN
FEB	4.10	4.1	4.10	3.5	0.00	0.6	0.82	0.6	0.00	0.1	0.82	0.7	3.28	2.9	0.00	0.5	3.28	3.4	24.01	25.0	2.89	2.9	0.39	0.0	0.00	0.0	4.00	4.0	0.00	0.0	0.39	0.5	0.34	0.39	FEB
MAR	6.03	6.0	5.40	5.5	0.63	0.5	1.08	1.1	0.08	0.1	1.16	1.2	4.32	4.4	0.55	0.4	4.87	4.8	28.32	29.4	4.31	4.4	0.01	0.0	0.00	0.0	4.00	4.0	0.00	0.0	0.56	0.4	0.44	0.43	MAR
APR	4.95	5.0	2.65	2.5	2.30	2.5	0.53	0.4	0.29	0.5	0.82	0.9	2.12	2.1	2.01	2.0	4.13	4.1	27.78	30.4	-0.54	1.0	2.66	1.1	0.00	0.0	4.00	4.0	0.00	0.0	4.67	3.1	1.86	0.66	APR
MAY	2.50	2.5	0.93	1.0	1.57	1.5	0.19	0.3	0.28	0.2	0.47	0.5	0.74	0.7	1.29	1.3	2.03	2.0	14.38	21.1	-13.40	-9.3	14.14	10.0	1.80	1.8	4.00	4.0	0.00	0.0	13.63	9.5	9.48	6.39	MAY
JUN	7.34	7.4	0.00	1.0	7.34	6.4	0.00	0.2	0.84	0.7	0.84	0.9	0.00	0.8	6.50	5.7	6.50	6.5	3.00*	5.9	-11.38	-5.2	11.38	16.0	3.24	3.2	4.00	4.0	0.00	0.0	14.64	18.5	22.17	21.53	JUN
JUL	1.73	1.7	0.00	0.0	1.73	1.7	0.00	0.0	0.41	0.4	0.41	0.4	0.00	0.0	1.32	1.3	1.32	1.3	0.00*	0.5	-3.00	-5.4	3.00	5.4	5.32	5.6	4.00	4.0	-4.00	-4.0	3.00	5.1	5.15	7.30	JUL
AUG	2.00	2.0	0.00	0.0	2.00	2.0	0.00	0.0	0.42	0.4	0.42	0.4	0.00	0.0	0.58	1.6	0.58	1.6	0.00*	0.0	0.00	-0.5	0.00	0.5	0.97	1.0	0.00	0.0	0.00	0.0	0.61	1.1	1.14	1.79	AUG
TOTAL	67.59	67.6	45.73	45.4	21.86	22.2	9.16	9.2	3.15	3.3	12.31	12.5	36.57	36.2	18.71	18.9	55.28	55.1	-	-	0.00	0.0	36.57	36.2	14.00	14.0	-	-	0.00	0.0	41.28	41.1	42.79	41.74	TOTAL

1/ Evapotranspiration as computed by the Thornthwaite method.

(a) Columns whose numbers include "a" give adopted values which bring all components into balance.

* Estimated values.

CENTRAL SIERRA SNOW LABORATORY

WATER BALANCE BY MONTHS
(Inches depth over basin)

MONTH	TOTAL PRECIPITATION Total (1)	(1a)	Snowfall W.E. (2a)	Rainfall (3a)	INTERCEPTION Snowfall W.E. (4a)	Rainfall (5a)	Total (6a)	NET PRECIPITATION Snowfall W.E. (7a)	Rainfall (8a)	Total (9a)	SNOWPACK WATER EQUIVALENT Cumulative (10)	(10a)	Change (11)	(11a)	MELT (12a)	EVAPO-TRANSPIRATION 1/ (13)	2/ (13a)	AVAILABLE SOIL MOISTURE Initial (14a)	Change (15a)	RUNOFF COMPUTED Gen. (16a)	RUNOFF ACTUAL Gen. (17)	Obs. (18)	MONTH
1946 - 47																							
SEP	1.90	1.9	0.0	1.9	0.0	0.1	0.1	0.0	1.8	1.8	0.0	0.0	0.0	0.0	0.0	(3.5)	1.8	0.0	0.0	0.0	0.00	0.00	SEP
OCT	2.98	3.0	1.7	1.3	0.1	0.1	0.2	1.6	1.2	2.8	1.1*	0.7	1.1*	0.7	0.9	(0.7)	1.1	1.0	1.0	0.0	0.00	0.00	OCT
NOV	12.36	12.4	10.4	2.0	1.2	0.0	1.2	9.2	2.0	11.2	13.2*	8.2	12.1*	7.5	1.7	(0.0)	0.3	3.7	2.7	0.7	0.35	0.09	NOV
DEC	5.81	5.8	4.5	1.3	0.6	0.0	0.6	3.9	1.3	5.2	16.7*	10.7	3.5*	2.5	1.4	(0.0)	0.5	5.7	2.0	0.2	0.11	0.16	DEC
JAN	5.33	5.3	5.1	0.2	0.5	0.0	0.5	4.6	0.2	4.8	20.3	14.1	3.6*	3.4	1.2	(0.0)	0.5	6.0	0.3	0.6	0.32	0.23	JAN
FEB	5.59	5.6	4.9	0.7	0.6	0.0	0.6	4.3	0.7	5.0	23.4	17.2	3.1	3.1	1.2	(0.0)	0.4	6.0	0.0	1.5	1.29	0.87	FEB
MAR	13.81	13.8	12.3	1.5	1.4	0.0	1.4	10.9	1.5	12.4	33.1*	26.2	9.7*	9.0	1.9	(0.0)	0.4	6.0	0.0	3.0	2.83	2.35	MAR
APR	3.09	3.1	2.9	0.2	0.3	0.0	0.3	2.6	0.2	2.8	21.2	17.0	-11.9	-9.2	11.8	(0.8)	1.5	5.9	-0.1	10.6	10.08	8.94	APR
MAY	1.64	1.6	0.0	1.6	0.0	0.2	0.2	0.0	1.4	1.4	0.0*	0.7	-21.2	-16.3	16.3	(2.6)	3.9	5.7	-0.2	14.0	13.67	14.11	MAY
JUN	2.21	2.2	0.0	2.2	0.0	0.2	0.2	0.0	2.0	2.0	0.0	0.0	0.0	-0.7	0.7	(2.7)	3.8	3.0	-2.7	1.6	1.57	2.83	JUN
JUL	0.00	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	(3.8)	2.5	0.5	-2.5	0.0	0.00	0.12	JUL
AUG	0.03	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	(3.5)	0.5	0.0	-0.5	0.0	0.00	0.00	AUG
TOTAL	54.75	54.7	41.8	12.9	4.7	0.6	5.3	37.1	12.3	49.4	-	-	0.0	0.0	37.1	(17.6)	17.2	-	0.0	32.2	30.22	29.70	TOTAL
1947 - 48																							
SEP	0.05	0.1	0.0	0.1	0.0	0.0	0.0	0.0	0.1	0.1	0.0	0.0	0.0	0.0	0.0	(3.2)	0.1	0.0	0.0	0.0	0.00	0.00	SEP
OCT	10.58	10.6	5.2	5.4	0.6	0.3	0.9	4.6	5.1	9.7	2.0*	1.9	2.0*	1.9	2.7	(1.5)	1.8	4.5	4.5	1.4	1.38	0.58	OCT
NOV	2.55	2.6	2.1	0.5	0.3	0.0	0.3	1.8	0.5	2.3	2.7*	2.7	0.7*	0.8	1.0	(0.0)	0.3	5.4	0.9	0.3	0.18	0.49	NOV
DEC	1.64	1.6	1.6	0.0	0.2	0.0	0.2	1.4	0.0	1.4	3.6	3.8	0.9	1.1	0.3	(0.0)	0.2	5.5	0.1	0.0	0.04	0.32	DEC
JAN	10.56	10.6	7.8	2.8	0.8	0.2	1.0	7.0	2.6	9.6	10.4	9.9	6.8	6.1	0.9	(0.0)	0.4	6.0	0.5	2.6	2.39	1.94	JAN
FEB	8.13	8.1	7.7	0.4	0.8	0.0	0.8	6.9	0.4	7.3	17.0	16.2	6.6	6.3	0.6	(0.0)	0.3	6.0	0.0	0.7	0.66	0.75	FEB
MAR	13.78	13.8	13.3	0.5	1.4	0.0	1.4	11.9	0.5	12.4	29.5	27.4	12.5	11.2	0.7	(0.0)	0.2	6.0	0.0	1.0	0.92	0.73	MAR
APR	20.00	20.0	16.3	3.7	2.8*	0.2	3.0	13.5	3.5	17.0	42.6	38.6	13.1	11.2	2.3	(0.0)	0.8	6.0	0.0	5.0	4.64	4.22	APR
MAY	4.13	4.1	3.4	0.7	0.4	0.0	0.4	3.0	0.7	3.7	24.6	22.8	-18.0	-15.8	18.8	(1.3)	3.5	5.9	-0.1	16.1	15.89	14.95	MAY
JUN	0.83	0.8	0.0	0.8	0.0	0.1	0.1	0.0	0.7	0.7	1.0	1.8	-23.6	-21.0	21.0	(3.1)	4.0	5.4	-0.5	18.2	18.03	18.59	JUN
JUL	0.00	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	-1.0	-1.8	1.8	(4.3)	5.3	1.9	-3.5	0.0	0.00	1.40	JUL
AUG	0.01	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	(3.4)	1.9	0.0	-1.9	0.0	0.00	0.07	AUG
TOTAL	72.26	72.3	57.4	14.9	7.3	0.8	8.1	50.1	14.1	64.2	-	-	0.0	0.0	50.1	(16.8)	18.9	-	0.0	45.3	44.13	44.05	TOTAL
1948 - 49																							
SEP	0.27	0.3	0.0	0.3	0.0	0.1	0.1	0.0	0.2	0.2	0.0	0.0	0.0	0.0	0.0	(2.8)	0.2	0.0	0.0	0.0	0.00	0.00	SEP
OCT	1.23	1.2	0.8	0.4	0.0	0.1	0.1	0.8	0.3	1.1	0.2*	0.2	0.2*	0.2	0.6	(1.6)	0.4	0.4	0.4	0.1	0.00	0.00	OCT
NOV	6.67	6.7	2.9	3.8	0.2	0.3	0.5	2.7	3.5	6.2	3.3*	2.0	3.1*	1.8	0.9	(0.2)	0.4	3.9	3.5	0.5	0.46	0.31	NOV
DEC	10.61	10.0	10.4	0.2	1.1	0.0	1.1	9.3	0.2	9.5	12.8	10.5	9.5*	8.5	0.8	(0.0)	0.2	4.5	0.6	0.2	0.16	0.11	DEC
JAN	6.07	6.1	6.1	0.0	0.6	0.0	0.6	5.5	0.0	5.5	18.4	15.2	5.6	4.7	0.8	(0.0)	0.2	4.9	0.4	0.2	0.12	0.14	JAN
FEB	12.89	12.9	12.7	0.2	1.3	0.0	1.3	11.4	0.2	11.6	31.3	25.9	12.9	10.7	0.7	(0.0)	0.4	5.3	0.4	0.1	0.14	0.12	FEB
MAR	13.64	13.6	12.9	0.7	1.3	0.2	1.5	11.6	0.5	12.1	42.3	36.9	11.0	11.0	0.6	(0.0)	0.5	5.7	0.4	0.2	0.37	0.22	MAR
APR	0.40	0.4	0.2	0.2	0.2	0.0	0.2	0.0	0.2	0.2	23.9	21.6	-18.4	-15.3	15.3	(1.3)	3.5	6.0	0.3	11.7	11.41	9.65	APR
MAY	4.74	4.7	1.5	3.2	0.3	0.1	0.4	1.2	3.1	4.3	1.3	3.6	-22.6	-18.0	19.2	(1.8)	4.5	5.8	-0.2	18.0	17.69	18.04	MAY
JUN	0.08	0.1	0.0	0.1	0.0	0.1	0.1	0.0	0.0	0.0	0.0	0.0	-1.3	-3.6	3.6	(3.5)	5.1	1.8	-4.0	2.5	2.39	3.69	JUN
JUL	0.07	0.1	0.0	0.1	0.0	0.1	0.1	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	(4.0)	1.8	0.0	-1.8	0.0	0.00	0.14	JUL
AUG	0.98	1.0	0.0	1.0	0.0	0.1	0.1	0.0	0.9	0.9	0.0	0.0	0.0	0.0	0.0	(3.5)	0.9	0.0	0.0	0.0	0.00	0.00	AUG
TOTAL	57.65	57.7	47.5	10.2	5.0	1.1	6.1	42.5	9.1	51.6	-	-	0.0	0.0	42.5	(18.7)	18.1	-	0.0	33.5	32.74	32.42	TOTAL
1949 - 50																							
SEP	0.11	0.1	0.0	0.1	0.0	0.1	0.1	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	(3.0)	0.0	0.0	0.0	0.0	0.00	0.00	SEP
OCT	0.76	0.8	0.1	0.7	0.1	0.0	0.1	0.0	0.7	0.7	0.1*	0.0	0.1*	0.0	0.0	(1.4)	0.7	0.0	0.0	0.0	0.00	0.00	OCT
NOV	6.13	6.1	5.5	0.6	0.6	0.0	0.6	4.9	0.6	5.5	3.9	2.5	3.8*	2.5	2.4	(0.9)	1.5	1.0	1.0	0.5	0.38	0.11	NOV
DEC	5.71	5.7	5.7	0.0	0.6	0.0	0.6	5.1	0.0	5.1	9.4	7.1	5.5	4.6	0.5	(0.0)	0.4	1.1	0.1	0.0	0.00	0.09	DEC
JAN	31.74	31.7	26.6	5.1	2.7	0.3	3.0	23.9	4.8	28.7	34.3	29.5	24.9	22.4	1.5	(0.0)	0.2	6.0	4.9	1.2	1.04	0.64	JAN
FEB	7.43	7.4	6.8	0.6	0.7	0.0	0.7	6.1	0.6	6.7	42.2	34.7	7.9	5.2	0.9	(0.0)	0.3	6.0	0.0	1.2	1.07	0.72	FEB
MAR	12.28	12.3	10.8	1.5	1.2	0.0	1.2	9.6	1.5	11.1	54.6	43.1	12.4	8.4	1.2	(0.0)	0.4	6.0	0.0	2.3	2.16	1.68	MAR
APR	8.12	8.1	7.8	0.3	0.8	0.0	0.8	7.0	0.3	7.3	51.9	38.0	- 2.7	-5.1	12.1	(0.7)	1.5	6.0	0.0	10.9	10.63	9.75	APR
MAY	4.65	4.7	4.7	0.0	0.5	0.0	0.5	4.2	0.0	4.2	20.4	16.0	-31.5	-22.0	26.3	(1.3)	3.0	5.7	-0.3	23.6	23.44	23.19	MAY
JUN	0.90	0.9	0.5	0.4	0.1	0.0	0.1	0.4	0.4	0.8	0.7	0.5	-19.7	-15.5	15.8	(2.6)	3.5	3.2	-2.5	15.2	14.97	15.70	JUN
JUL	0.00	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	-0.7	-0.5	0.5	(4.0)	3.4	0.0	-3.2	0.3	0.19	1.52	JUL
AUG	0.14	0.1	0.0	0.1	0.0	0.0	0.0	0.0	0.1	0.1	0.0	0.0	0.0	0.0	0.0	(3.7)	0.1	0.0	0.0	0.0	0.00	0.00	AUG
TOTAL	77.97	77.9	68.5	9.4	7.3	0.4	7.7	61.2	9.0	70.2	-	-	0.0	0.0	61.2	(17.6)	15.0	-	0.0	55.2	53.88	53.40	TOTAL
1950 - 51																							
SEP	0.80	0.8	0.0	0.8	0.0	0.1	0.1	0.0	0.7	0.7	0.0	0.0	0.0	0.0	0.0	(3.6)	0.7	0.0	0.0	0.0	0.00	0.00	SEP
OCT	10.19	10.2	6.0	4.2	0.7	0.2	0.9	5.3	4.0	9.3	3.0*	2.7	3.0*	2.7	2.6	(1.4)	1.5	4.5	4.5	0.6	0.55	0.13	OCT
NOV	23.09	23.1	6.1	17.0	0.8	1.0	1.8	5.3	16.0	21.3	1.0*	0.4	-2.0*	-2.3	7.6	(0.8)	1.5	6.0	1.5	20.6	20.35	19.55	NOV
DEC	16.90	16.9	6.0	10.9	0.9	0.5	1.4	5.1	10.4	15.5	9.3*	4.5	8.3*	4.1	1.1	(0.3)	0.3	6.0	0.0	11.2	11.14	12.13	DEC
JAN	16.89	16.9	15.9	1.0	1.7	0.0	1.7	14.2	1.0	15.2	26.4*	17.9	17.1*	13.4	0.8	(0.0)	0.2	6.0	0.0	1.6	1.47	1.62	JAN
FEB	9.23	9.2	8.5	0.7	0.9	0.0	0.9	7.6	0.7	8.3	34.0*	24.5	7.6*	6.6	1.0	(0.0)	0.1	6.0	0.0	1.6	1.45	1.52	FEB
MAR	6.57	6.6	6.1	0.5	0.7	0.0	0.7	5.4	0.5	5.9	35.0	26.9	1.0*	2.4	3.0	(0.0)	0.6	6.0	0.0	2.9	2.83	2.38	MAR
APR	4.42	4.4	3.5	0.9	0.4	0.1	0.5	3.1	0.8	3.9	21.8	17.6	-13.2	-9.3	12.4	(1.0)	1.5	6.0	0.0	11.7	11.63	11.34	APR
MAY	4.19	4.2	3.3	0.9	0.4	0.0	0.4	2.9	0.9	3.8	5.0	3.2	-16.8	-14.4	17.2	(1.7)	2.5	5.6	-0.4	16.0	16.07	15.64	MAY
JUN	0.21	0.2	0.0	0.2	0.0	0.1	0.1	0.0	0.1	0.1	0.0	0.0	-5.0	-3.2	3.2	(3.4)	4.0	1.8	-3.8	3.1	3.15	4.41	JUN
JUL	0.00	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	(3.9)	1.8	0.0	-1.8	0.0	0.00	0.26	JUL
AUG	0.12	0.1	0.0	0.1	0.0	0.0	0.0	0.0	0.1	0.1	0.0	0.0	0.0	0.0	0.0	(3.6)	0.1	0.0	0.0	0.0	0.00	0.01	AUG
TOTAL	92.61	92.6	55.4	37.2	6.5	2.0	8.5	48.9	35.2	84.1	-	-	0.0	0.0	48.9	(19.7)	14.8	-	0.0	69.3	68.64	68.99	TOTAL

1/ Potential evapotranspiration as computed by the Thornthwaite method.

2/ Adopted evapotranspiration, modified to CSSL conditions.

(a) All columns whose numbers include "a" give adopted values which bring all components into balance.

* Estimated values.

TABLE 4-4

WILLAMETTE BASIN SNOW LABORATORY

WATER BALANCE BY MONTHS
(Inches depth over basin)

MONTH	TOTAL PRECIPITATION						INTERCEPTION LOSS						NET PRECIPITATION						SNOWPACK WATER EQUIVALENT				MELT		EVAPO-TRANSPIRATION			AVAILABLE SOIL MOISTURE				RUNOFF				MONTH
	Total		Snowfall W.E.		Rainfall		Snowfall W.E.		Rainfall		Total		Snowfall W.E.		Rainfall		Total		Cumulative		Change				Thornthwaite 1/	Thornthwaite 2/		Initial		Change		COMPUTED Generated		ACTUAL Gen.	ACTUAL Obs.	
	(1)	(1a)	(2)	(2a)	(3)	(3a)	(4)	(4a)	(5)	(5a)	(6)	(6a)	(7)	(7a)	(8)	(8a)	(9)	(9a)	(10)	(10a)	(11)	(11a)	(12)	(12a)	(13)	(13')	(13a)	(14)	(14a)	(15)	(15a)	(16)	(16a)	(17)	(18)	
1947 - 48																																				
SEP*	3.5	3.5	0.0	0.0	3.5	3.5	0.0	0.0	1.2	1.2	1.2	1.2	0.0	0.0	2.3	2.3	2.3	2.3	0.0	0.0	0.0	0.0	0.0	0.0	2.9	2.9	2.8	1.0	1.0 3/	-0.6	-1.0	0.0	0.5	0.53	0.56	SEP*
OCT*	22.5	23.0	0.7	0.7	21.8	22.3	0.2	0.2	2.2	2.3	2.4	2.5	0.5	0.5	19.6	20.0	20.1	20.5	0.3	0.3	0.3	0.3	0.2	0.2	1.6	1.6	1.6	0.4	0.0	4.6	5.0	13.6	13.6	13.57	9.56	OCT*
NOV	12.0	13.7	1.9	1.9	10.1	11.8	0.4	0.4	1.2	1.3	1.6	1.7	1.5	1.5	8.9	10.5	10.4	12.0	0.0	0.0	-0.3	-0.3	1.8	1.8	0.2	0.2	0.2	5.0	5.0	0.0	0.0	10.5	12.1	12.09	12.58	NOV
DEC	11.6	11.6	9.8	9.8	1.8	1.8	1.1	1.1	0.4	0.4	1.5	1.5	8.7	8.7	1.4	1.4	10.1	10.1	2.4	1.9	2.4	1.9	6.3	6.8	0.0	0.0	0.0	5.0	5.0	0.0	0.0	7.7	8.2	8.18	8.04	DEC
JAN	17.5	17.5	12.3	12.3	5.2	5.2	1.4	1.4	0.5	0.5	1.9	1.9	10.9	10.9	4.7	4.7	15.6	15.6	5.2	2.9	2.8	1.0	8.1	9.9	0.3	0.3	0.3	5.0	5.0	0.0	0.0	12.5	14.3	14.33	16.04	JAN
FEB	20.3	20.3	11.4	11.4	8.9	8.9	1.2	1.2	0.9	0.9	2.1	2.1	10.2	10.2	8.0	8.0	18.2	18.2	10.9	11.0	5.7	8.1	4.5	2.1	0.0	0.0	0.0	5.0	5.0	0.0	0.0	12.5	10.1	10.06	7.90	FEB
MAR	13.0	12.1	10.9	10.0	2.1	2.1	1.2	1.2	0.5	0.5	1.7	1.7	9.7	8.8	1.6	1.6	11.3	10.4	11.2	15.2	0.3	4.2	9.4	4.6	0.1	0.1	0.1	5.0	5.0	0.0	0.0	10.9	6.1	6.08	5.78	MAR
APR	11.7	11.7	7.2	5.7	4.5	6.0	0.9	0.8	0.8	0.9	1.7	1.7	6.3	4.9	3.7	5.1	10.0	10.0	14.7	15.4	3.5	0.2	2.8	4.7	0.6	0.6	0.6	5.0	5.0	0.0	0.0	5.9	9.2	9.21	10.06	APR
MAY	7.1	7.1	6.4	6.4	0.7	0.7	0.9	0.9	0.2	0.2	1.1	1.1	5.5	5.5	0.5	0.5	6.0	6.0	6.0	5.8	-8.7	-9.6	14.2	15.1	2.0	2.0	2.0	5.0	5.0	0.0	0.0	12.7	13.6	13.64	12.91	MAY
JUN	3.6	3.6	0.0	0.0	3.6	3.6	0.0	0.0	1.3	1.3	1.3	1.3	0.0	0.0	2.3	2.3	2.3	2.3	0.0	0.0	-6.0	-5.8	6.0	5.8	4.0	4.0	4.0	5.0	5.0	-0.2	0.0	4.5	4.1	4.10	6.44	JUN
JUL	1.9	1.9	0.0	0.0	1.9	1.9	0.0	0.0	0.7	0.7	0.7	0.7	0.0	0.0	1.2	1.2	1.2	1.2	0.0	0.0	0.0	0.0	0.0	0.0	4.4	4.4	4.4	4.8	5.0	-3.2	-4.0	0.0	0.8	0.85	1.77	JUL
AUG	1.7	1.7	0.0	0.0	1.7	1.7	0.0	0.0	0.6	0.6	0.6	0.6	0.0	0.0	1.1	1.1	1.1	1.1	0.0	0.0	0.0	0.0	0.0	0.0	3.7	2.7	2.1	1.6	1.0	-1.6	-1.0	0.0	0.0	0.13	0.89	AUG
TOTAL	126.4	127.7	60.6	58.2	65.8	69.5	7.3	7.2	10.5	10.8	17.8	18.0	53.3	51.0	55.3	58.7	108.6	109.7	-	-	0.0	0.0	53.3	51.0	19.8	18.8	18.1	-	-	0.0	0.0	90.8	92.6	92.77	92.53	TOTAL
1948 - 49																																				
SEP	5.7	5.7	1.6	1.6	4.1	4.1	0.3	0.3	1.4	1.4	1.7	1.7	1.3	1.3	2.7	2.7	4.0	4.0	0.0	0.0	0.0	0.0	1.3	1.3	2.7	2.7	1.0	0.0	0.0	1.3	1.3	0.0	1.7	1.70	0.86	SEP
OCT	6.2	6.2	4.4	4.4	1.8	1.8	0.7	0.7	0.5	0.5	1.2	1.2	3.7	3.7	1.3	1.3	5.0	5.0	0.0	0.0	0.0	0.0	3.7	3.7	1.9	1.9	1.9	1.3	1.3	3.1	1.0	0.0	2.1	2.12	2.04	OCT
NOV	17.8	18.8	10.9	10.9	6.9	7.9	1.3	1.3	0.7	0.8	2.0	2.1	9.6	9.6	6.2	7.1	15.8	16.7	4.5	4.5	4.5	4.5	5.1	5.1	0.3	0.3	0.3	4.4	2.3	0.6	2.7	10.4	9.2	9.21	6.48	NOV
DEC	25.4	26.6	21.0	22.4	4.4	4.2	2.2	2.3	0.4	0.4	2.6	2.7	18.8	20.1	4.0	3.8	22.8	23.9	17.4	18.7	12.9	14.2	5.9	5.9	0.0	0.0	0.0	5.0	5.0	0.0	0.0	9.9	9.7	9.66	11.09	DEC
JAN	4.1	4.1	4.1	4.1	0.0	0.0	0.7	0.7	0.0	0.0	0.7	0.7	3.4	3.4	0.0	0.0	3.4	3.4	17.3	20.5	-0.1	1.8	3.5	1.6	0.0	0.0	0.0	5.0	5.0	0.0	0.0	3.5	1.6	1.55	2.70	JAN
FEB	28.2	28.2	22.2	22.2	6.0	6.0	2.3	2.3	0.6	0.6	2.9	2.9	19.9	19.9	5.4	5.4	25.3	25.3	27.2	33.6	9.9	13.1	10.0	6.8	0.0	0.0	0.0	5.0	5.0	0.0	0.0	15.4	12.2	12.19	8.96	FEB
MAR	8.9	10.0	5.0	5.0	3.9	5.0	0.8	0.8	0.9	1.0	1.7	1.8	4.2	4.2	3.0	4.0	7.2	8.2	25.0	31.4	-2.2	-2.2	6.4	6.4	0.5	0.5	0.5	5.0	5.0	0.0	0.0	8.9	9.9	9.86	11.17	MAR
APR	5.1	6.2	0.8	0.8	4.3	5.4	0.2	0.2	0.9	1.0	1.1	1.2	0.6	0.6	3.4	4.4	4.0	5.0	18.5	18.5	-6.5	-12.9	7.1	13.5	1.5	1.5	1.5	5.0	5.0	0.0	0.0	9.0	16.4	16.18	14.40	APR
MAY	7.2	7.2	0.8	0.8	6.4	6.4	0.2	0.2	2.2	2.2	2.4	2.4	0.6	0.6	4.2	4.2	4.8	4.8	3.4	3.8	-15.1	-14.7	15.7	15.3	2.6	2.6	2.6	5.0	5.0	0.0	0.0	17.3	16.9	16.87	18.20	MAY
JUN	1.9	1.9	0.0	0.0	1.9	1.9	0.0	0.0	0.7	0.7	0.7	0.7	0.0	0.0	1.2	1.2	1.2	1.2	0.0	0.0	-3.4	-3.8	3.4	3.8	3.6	3.6	3.6	5.0	5.0	-1.4	-1.2	2.4	2.6	2.56	4.66	JUN
JUL	0.4	0.4	0.0	0.0	0.4	0.4	0.0	0.0	0.1	0.1	0.1	0.1	0.0	0.0	0.3	0.3	0.3	0.3	0.0	0.0	0.0	0.0	0.0	0.0	4.5	3.9	4.1	3.6	3.8	-3.6	-3.8	0.0	0.0	0.00	1.23	JUL
AUG	0.3	0.3	0.0	0.0	0.3	0.3	0.0	0.0	0.1	0.1	0.1	0.1	0.0	0.0	0.2	0.2	0.2	0.2	0.0	0.0	0.0	0.0	0.0	0.0	3.8	0.2	0.2	0.0	0.0	0.0	0.0	0.0	0.0	0.10	0.50	AUG
TOTAL	111.2	115.6	70.8	72.2	40.4	43.4	8.7	8.8	8.5	8.8	17.2	17.6	62.1	63.4	31.9	34.6	94.0	98.0	-	-	0.0	0.0	62.1	63.4	21.4	17.2	15.7	-	-	0.0	0.0	76.8	82.3	82.00	82.29	TOTAL
1949 - 50																																				
SEP	3.0	3.0	0.0	0.0	3.0	3.0	0.0	0.0	1.0	1.0	1.0	1.0	0.0	0.0	2.0	2.0	2.0	2.0	0.0	0.0	0.0	0.0	0.0	0.0	3.2	2.0	1.6	0.0	0.0	0.0	0.0	0.0	0.4	0.37	0.37	SEP
OCT	8.2	8.2	1.6	1.6	6.6	6.6	0.4	0.4	1.1	1.1	1.5	1.5	1.2	1.2	5.5	5.5	6.7	6.7	0.0	0.0	0.0	0.0	1.2	1.2	1.3	1.3	1.3	0.0	0.0	5.0	3.9	0.4	1.5	1.52	1.04	OCT
NOV	14.1	14.1	2.3	2.3	11.8	11.8	0.5	0.5	1.3	1.3	1.8	1.8	1.8	1.8	10.5	10.5	12.3	12.3	0.0	1.3	0.0	1.3	1.8	0.5	1.5	1.5	1.5	5.0	3.9	0.0	1.1	10.8	8.4	8.43	4.40	NOV
DEC	18.8	18.8	16.8	16.8	2.0	2.0	1.8	1.8	0.2	0.2	2.0	2.0	15.0	15.0	1.8	1.8	16.8	16.8	11.7	14.2	11.7	12.9	3.3	2.1	0.0	0.0	0.0	5.0	5.0	0.0	0.0	5.1	3.9	3.86	5.28	DEC
JAN	33.8	33.8	29.0	29.0	4.8	4.8	2.9	2.9	0.5	0.5	3.4	3.4	26.1	26.1	4.3	4.3	30.4	30.4	29.1	34.8	17.4	20.6	8.7	5.5	0.0	0.0	0.0	5.0	5.0	0.0	0.0	13.0	9.8	9.81	9.77	JAN
FEB	15.4	15.4	7.3	7.3	8.1	8.1	0.9	0.9	1.1	1.1	2.0	2.0	6.4	6.4	7.0	7.0	13.4	13.4	28.4	29.1	-0.7	-5.7	7.1	12.1	0.0	0.0	0.0	5.0	5.0	0.0	0.0	14.1	19.1	19.14	16.98	FEB
MAR	20.0	20.0	12.2	12.2	7.8	7.8	1.4	1.4	0.8	0.8	2.2	2.2	10.8	10.8	7.0	7.0	17.8	17.8	30.9	34.0	2.5	4.9	8.3	5.9	0.0	0.0	0.0	5.0	5.0	0.0	0.0	15.3	12.9	12.89	15.21	MAR
APR	7.6	8.7	3.4	4.5	4.2	4.2	0.6	0.7	0.9	0.9	1.5	1.6	2.8	3.8	3.3	3.3	6.1	7.1	27.1	29.6	-3.8	-4.4	6.6	8.2	0.7	0.7	0.7	5.0	5.0	0.0	0.0	9.2	10.8	10.84	10.37	APR
MAY	3.7	3.7	3.5	3.5	0.2	0.2	0.6	0.6	0.1	0.1	0.7	0.7	2.9	2.9	0.1	0.1	3.0	3.0	14.7	14.3	-12.4	-15.3	15.3	18.2	1.9	1.9	1.9	5.0	5.0	0.0	0.0	13.5	16.4	16.45	14.86	MAY
JUN	5.1	5.1	0.3	0.3	4.8	4.8	0.1	0.1	1.7	1.7	1.8	1.8	0.2	0.2	3.1	3.1	3.3	3.3	1.1	1.3	-13.6	-13.0	13.8	13.2	3.0	3.0	3.0	5.0	5.0	0.0	0.0	13.9	13.3	13.34	14.28	JUN
JUL	0.7	0.7	0.0	0.0	0.7	0.7	0.0	0.0	0.2	0.2	0.2	0.2	0.0	0.0	0.5	0.5	0.5	0.5	0.0	0.0	-1.1	-1.3	1.1	1.3	4.3	4.3	4.3	5.0	5.0	-3.2	-3.0	0.6	0.5	0.53	3.35	JUL
AUG	2.4	2.4	0.0	0.0	2.4	2.4	0.0	0.0	0.8	0.8	0.8	0.8	0.0	0.0	1.6	1.6	1.6	1.6	0.0	0.0	0.0	0.0	0.0	0.0	4.5	3.3	3.4	1.7	2.0	-1.7	-2.0	0.0	0.2	0.22	1.08	AUG
TOTAL	132.8	133.9	76.4	77.5	56.4	56.4	9.2	9.3	9.7	9.7	18.9	19.0	67.2	68.2	46.7	46.7	113.9	114.9	-	-	0.0	0.0	67.2	68.2	20.4	18.0	17.7	-	-	0.0	0.0	95.9	97.2	97.40	96.99	TOTAL
1950 - 51																																				
SEP	2.7	2.7	0.0	0.0	2.7	2.7	0.0	0.0	0.9	0.9	0.9	0.9	0.0	0.0	1.8	1.8	1.8	1.8	0.0	0.0	0.0	0.0	0.0	0.0	3.0	1.8	1.5	0.0	0.0	0.0	0.0	0.0	0.3	0.27	0.51	SEP
OCT	25.7	25.7	7.5	7.5	18.2	18.2	1.0	1.0	1.8	1.8	2.8	2.8	6.5	6.5	16.4	16.4	22.9	22.9	1.0	1.3	1.0	1.3	5.5	5.2	1.4	1.4	1.4	0.0	0.0	5.0	5.0	15.5	15.2	15.17	9.60	OCT
NOV	20.2	20.2	10.8	10.8	9.4	9.4	1.3	1.3	0.9	0.9	2.2	2.2	9.5	9.5	8.5	8.5	18.0	18.0	2.6	3.0	1.6	1.7	7.9	7.8	0.5	0.5	0.5	5.0	5.0	0.0	0.0	15.9	15.8	15.77	16.41	NOV
DEC	17.1	19.2	7.0	8.1	10.1	11.1	0.9	1.0	1.0	1.0	1.9	2.0	6.1	7.1	9.1	10.1	15.2	17.2	2.8	3.4	0.2	0.4	5.9	6.7	0.5	0.5	0.5	5.0	5.0	0.0	0.0	14.5	16.3	16.29	16.77	DEC
JAN	25.5	26.6	18.4	19.5	7.1	7.1	2.0	2.1	0.7	0.7	2.7	2.8	16.4	17.4	6.4	6.4	22.8	23.8	12.2	14.7	9.4	11.3	7.0	6.1	0.0	0.0	0.0	5.0	5.0	0.0	0.0	13.4	12.5	12.52	13.29	JAN
FEB	14.2	14.2	5.7	5.7	8.5	8.5	0.8	0.8	1.1	1.1	1.9	1.9	4.9	4.9	7.4	7.4	12.3	12.3	13.4	13.5	1.2	-1.2	3.7	6.1	0.1	0.1	0.1	5.0	5.0	0.0	0.0	11.0	13.4	13.38	14.61	FEB
MAR	14.7	14.7	13.1	13.1	1.6	1.6	1.4	1.4	0.4	0.4	1.8	1.8	11.7	11.7	1.2	1.2	12.9	12.9	18.1	19.2	4.7	5.7	7.0	6.0	0.1	0.1	0.1	5.0	5.0	0.0	0.0	8.1	7.1	7.08	5.91	MAR
APR	3.6	3.6	2.2	2.2	1.4	1.4	0.4	0.4	0.5	0.5	0.9	0.9	1.8	1.8	0.9	0.9	2.7	2.7	7.0	6.8	-11.1	-12.4	12.9	14.2	1.8	1.8	1.8	5.0	5.0	0.0	0.0	12.0	13.3	13.33	12.66	APR
MAY	5.7	5.7	1.5	1.5	4.2	4.2	0.3	0.3	1.5	1.5	1.8	1.8	1.2	1.2	2.7	2.7	3.9	3.9	0.6	0.5	-6.4	-6.3	7.6	7.5	2.2	2.2	2.2	5.0	5.0	0.0	0.0	8.1	8.0	7.98	9.77	MAY
JUN	0.3	0.3	0.0	0.0	0.3	0.3	0.0	0.0	0.1	0.1	0.1	0.1	0.0	0.0	0.2	0.2	0.2	0.2	0.0	0.0	-0.6	-0.5	0.6	0.5	3.9	3.9	3.9	5.0	5.0	-3.5	-3.8	0.4	0.6	0.63	2.27	JUN
JUL	0.4	0.4	0.0	0.0	0.4	0.4	0.0	0.0	0.1	0.1	0.1	0.1	0.0	0.0	0.3	0.3	0.3	0.3	0.0	0.0	0.0	0.0	0.0	0.0	4.5	1.8	1.5	1.5	1.2	-1.5	-1.2	0.0	0.0	0.04	0.83	JUL
AUG	1.5	1.5	0.0	0.0	1.5	1.5	0.0	0.0	0.5	0.5	0.5	0.5	0.0	0.0	1.0	1.0	1.0	1.0	0.0	0.0	0.0	0.0	0.0	0.0	3.9	1.0	0.8	0.0	0.0	0.0	0.0	0.0	0.2	0.23	0.40	AUG
TOTAL	131.6	134.8	66.2	68.4	65.4	66.4	8.1	8.3	9.5	9.5	17.6	17.8	58.1	60.1	55.9	56.9	114.0	117.0	-	-	0.0	0.0	58.1	60.1	21.9	15.1	14.3	-	-	0.0	0.0	98.9	102.7	102.69	103.03	TOTAL

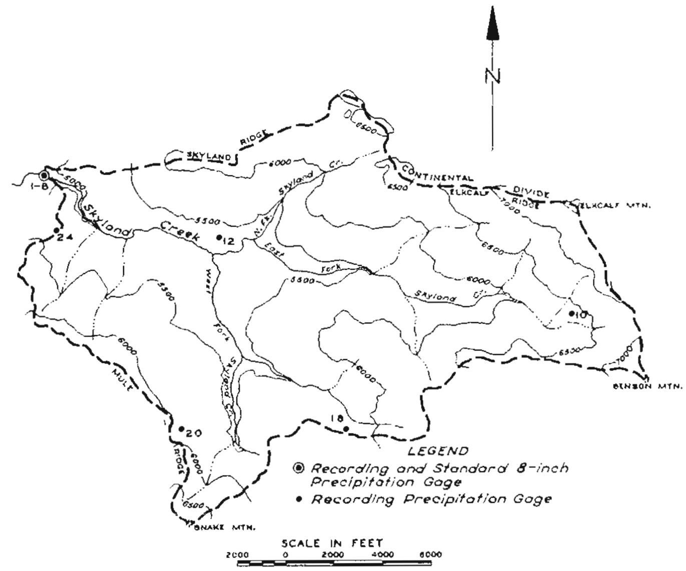
1/ Potential evapotranspiration by the Thornthwaite method.

2/ "Actual" evapotranspiration by the Thornthwaite method.

3/ In 1947-48 water year there was a carryover of an estimated 1.0 inch of soil moisture from the previous year.

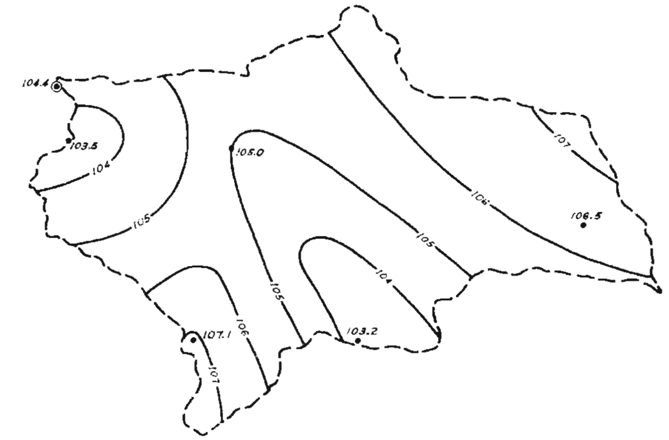
(a) All columns whose numbers include "a" give adopted values which bring all components into balance.

* Estimated values.



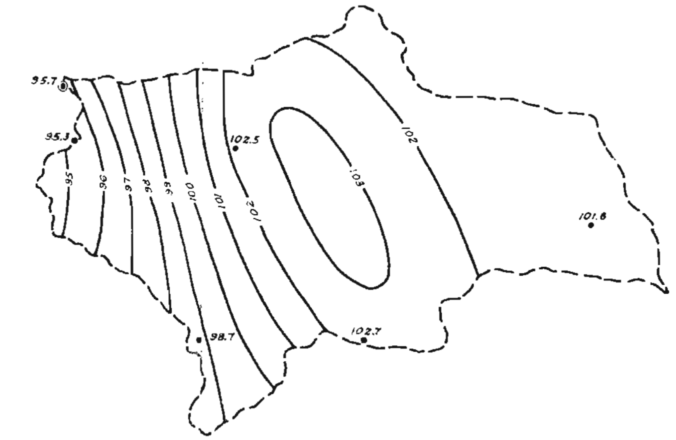
INDEX MAP

FIGURE 1



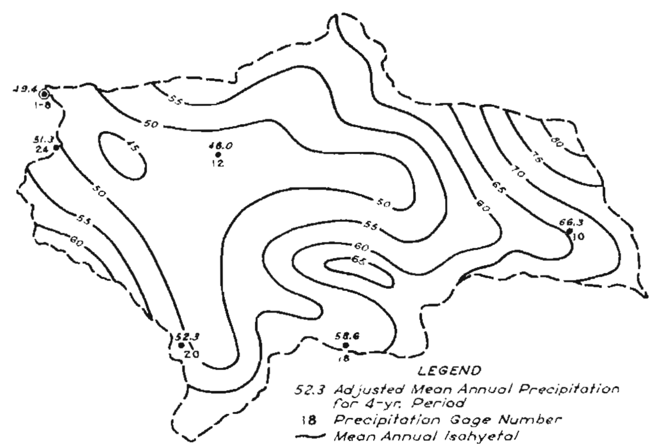
ISOPERCENTUAL MAP FOR 1946-47 WATER YEAR

FIGURE 2



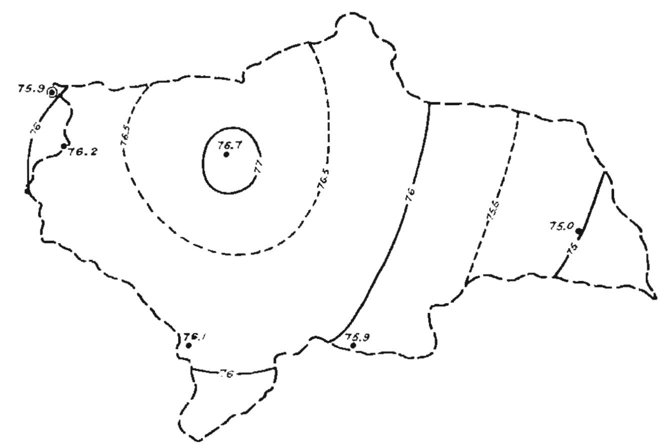
ISOPERCENTUAL MAP FOR 1947-48 WATER YEAR

FIGURE 3



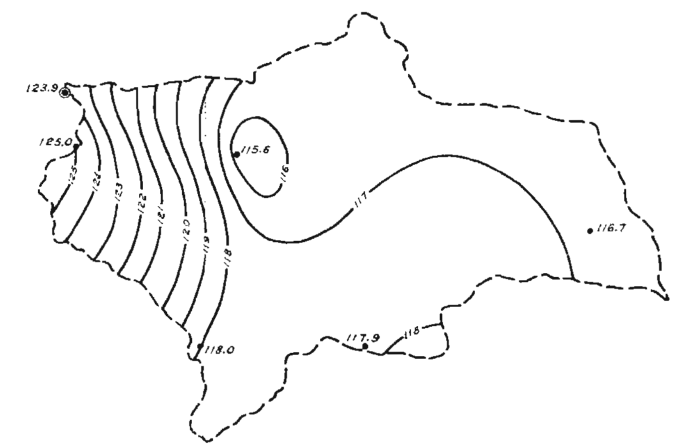
MEAN ANNUAL ISOHYETAL MAP, 1946-47 THROUGH 1949-50

FIGURE 4



ISOPERCENTUAL MAP FOR 1948-49 WATER YEAR

FIGURE 5



ISOPERCENTUAL MAP FOR 1949-50 WATER YEAR

FIGURE 6

NOTES for FIGURES 2, 3, 5, and 6:

Lines on isopercentual maps show water-year precipitation expressed as percent of the 4-year mean annual amounts shown in Figure 4.

Numbers adjacent to precipitation gage symbols indicate percent of 4-year mean annual amounts.

SNOW INVESTIGATIONS
SUMMARY REPORT

SNOW HYDROLOGY

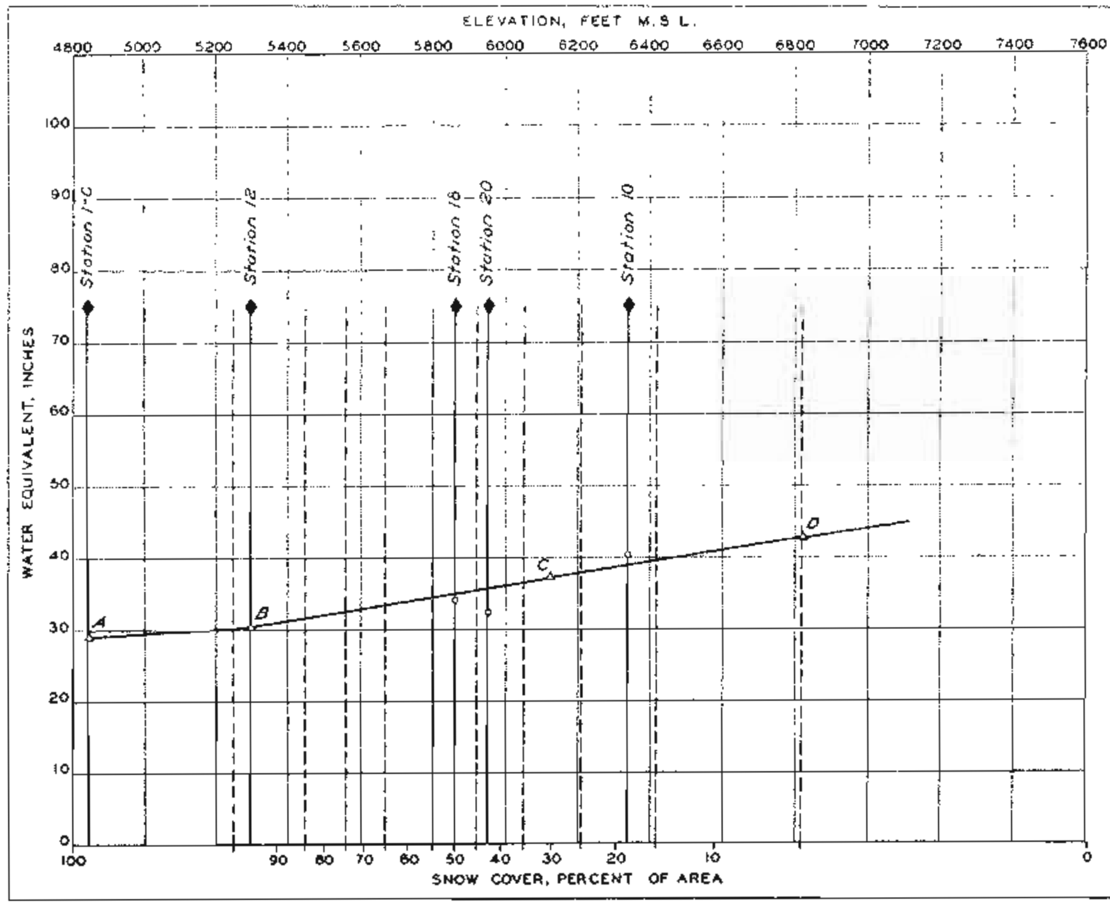
PRECIPITATION DISTRIBUTION, UCSL

SKYLAND CREEK - DRAINAGE AREA 8.09 SQ. MI.

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

PREPARED: [illegible]	SUBMITTED: [illegible]	TO ACCOMPANY REPORT DATED 10 JUNE 1956
DRAWN: [illegible]	APPROVED: [illegible]	PD-20-25/16

PLATE 4-1



Notes:
1. Plotted values show depth of snow water equivalent at numbered snow courses.
2. Line ABCD, representing unadjusted snowpack water-equivalent depth over basin, is derived in accordance with procedure described in text.
3. Dashed vertical lines indicate mid-point of each zone of 10% of basin area.
4. Intersections of line ABCD with dashed lines indicate mean depth of snow water equivalent for each zone; summation of zonal depths divided by 10 gives unadjusted basin snow water-equivalent depth.
5. Corrected basin snow water-equivalent depth is obtained by multiplying unadjusted depth by 0.80, a factor derived from water balance studies.

EXAMPLE

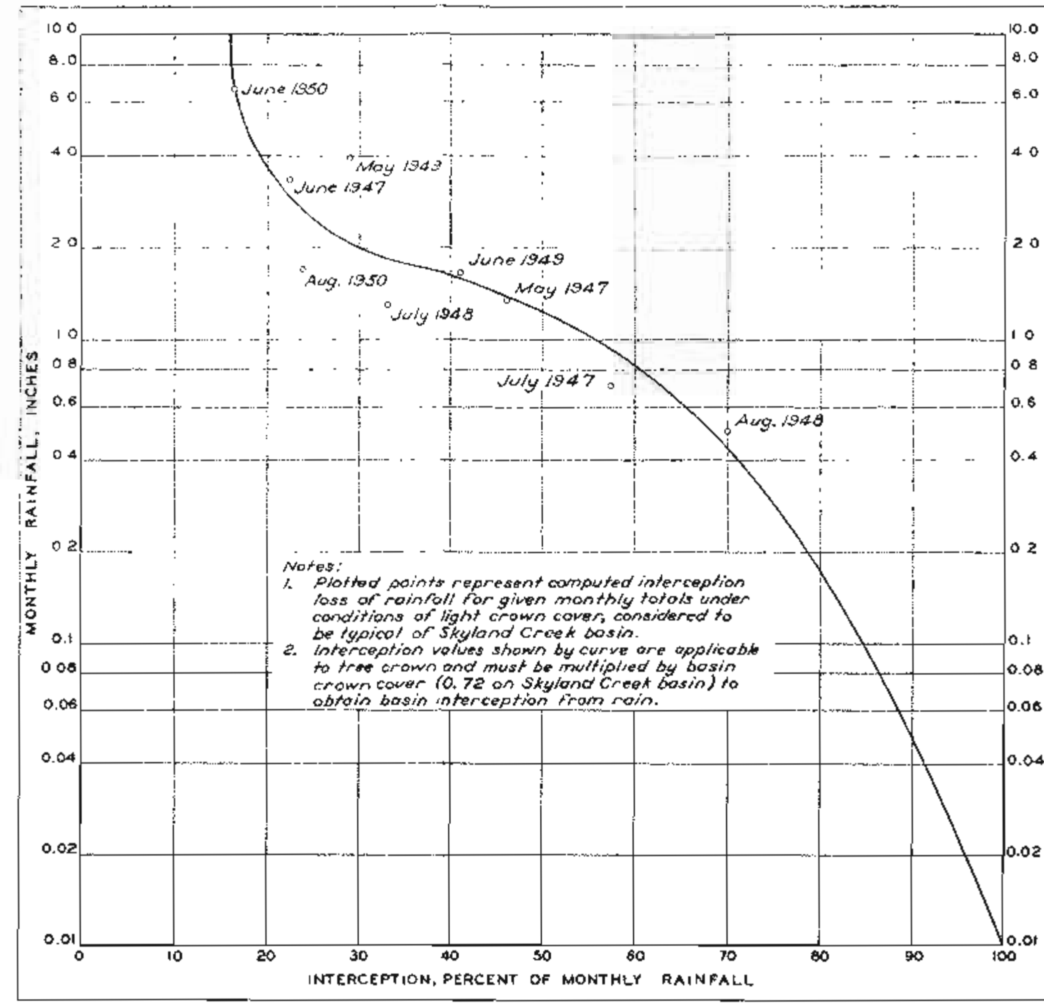
SNOW WATER EQUIVALENT

Skyland Creek Basin
UCSL - Drainage Area 8.09 Sq.Mi.

ZONE	INCHES
90 - 100	30.5
80 - 90	31.6
70 - 80	32.5
60 - 70	33.5
50 - 60	34.5
40 - 50	35.5
30 - 40	36.5
20 - 30	37.7
10 - 20	39.4
0 - 10	42.5
Wtd. Total	35.4
Date	1 April 1950

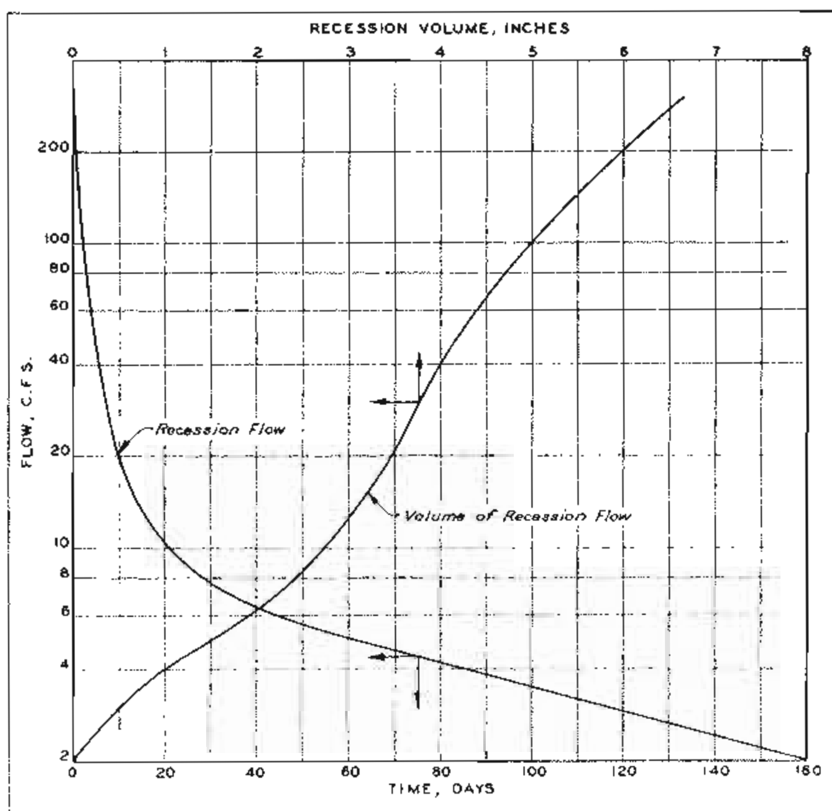
Unadjusted basin water equivalent = 35.4
Corrected basin water equivalent = 28.3

FIGURE 1 - SNOW WATER EQUIVALENT



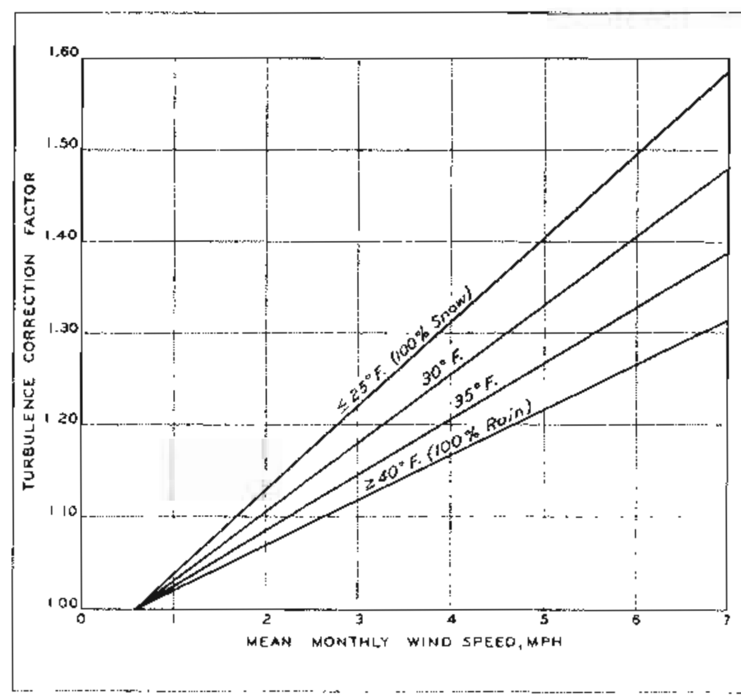
Notes:
1. Plotted points represent computed interception loss of rainfall for given monthly totals under conditions of light crown cover, considered to be typical of Skyland Creek basin.
2. Interception values shown by curve are applicable to tree crown and must be multiplied by basin crown cover (0.72 on Skyland Creek basin) to obtain basin interception from rain.

FIGURE 2 - INTERCEPTION LOSS FROM RAIN



Note:
Volume of recession flow curve is based on arbitrarily assigned zero recession volume at flow of 2.0 c.f.s. Curve is designed for computing generated runoff on flows of 2.0 c.f.s. or more.

FIGURE 3 - STREAMFLOW RECESSION CHARACTERISTICS



Notes:
1. Temperature parameters are for monthly mean temperature.
2. Lines represent turbulence correction factors for rain (≥ 40°F) and for snow (≤ 25°F) with interpolated values for mixed rain and snow (30°F and 35°F).
3. The relation between form of precipitation and monthly mean temperature is discussed in paragraph 4-04.04.

FIGURE 4 - TURBULENCE CORRECTION FACTOR

SNOW INVESTIGATIONS
SUMMARY REPORT

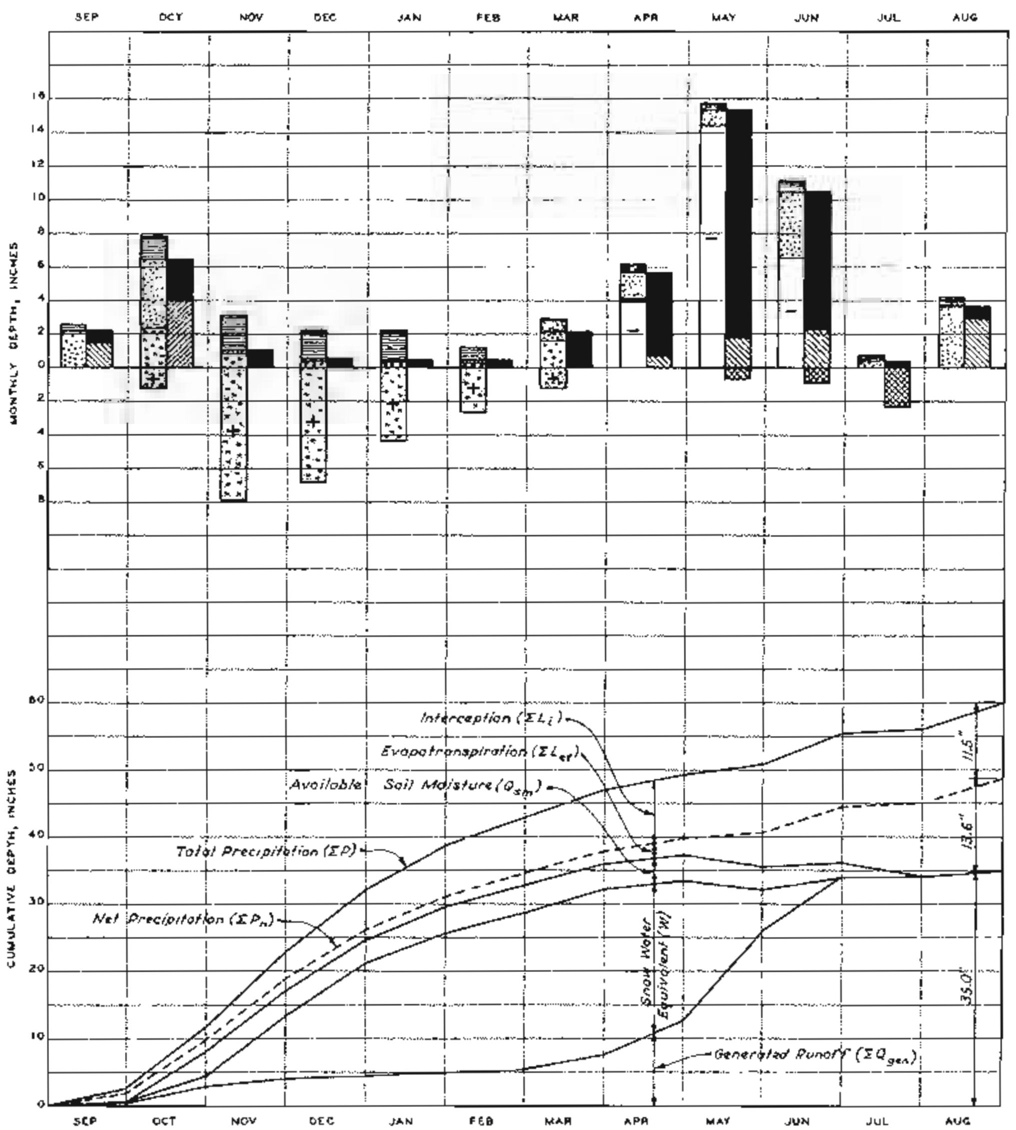
SNOW HYDROLOGY

SNOW CHART AND MISCELLANEOUS RELATIONSHIPS, UCSL

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U S ARMY

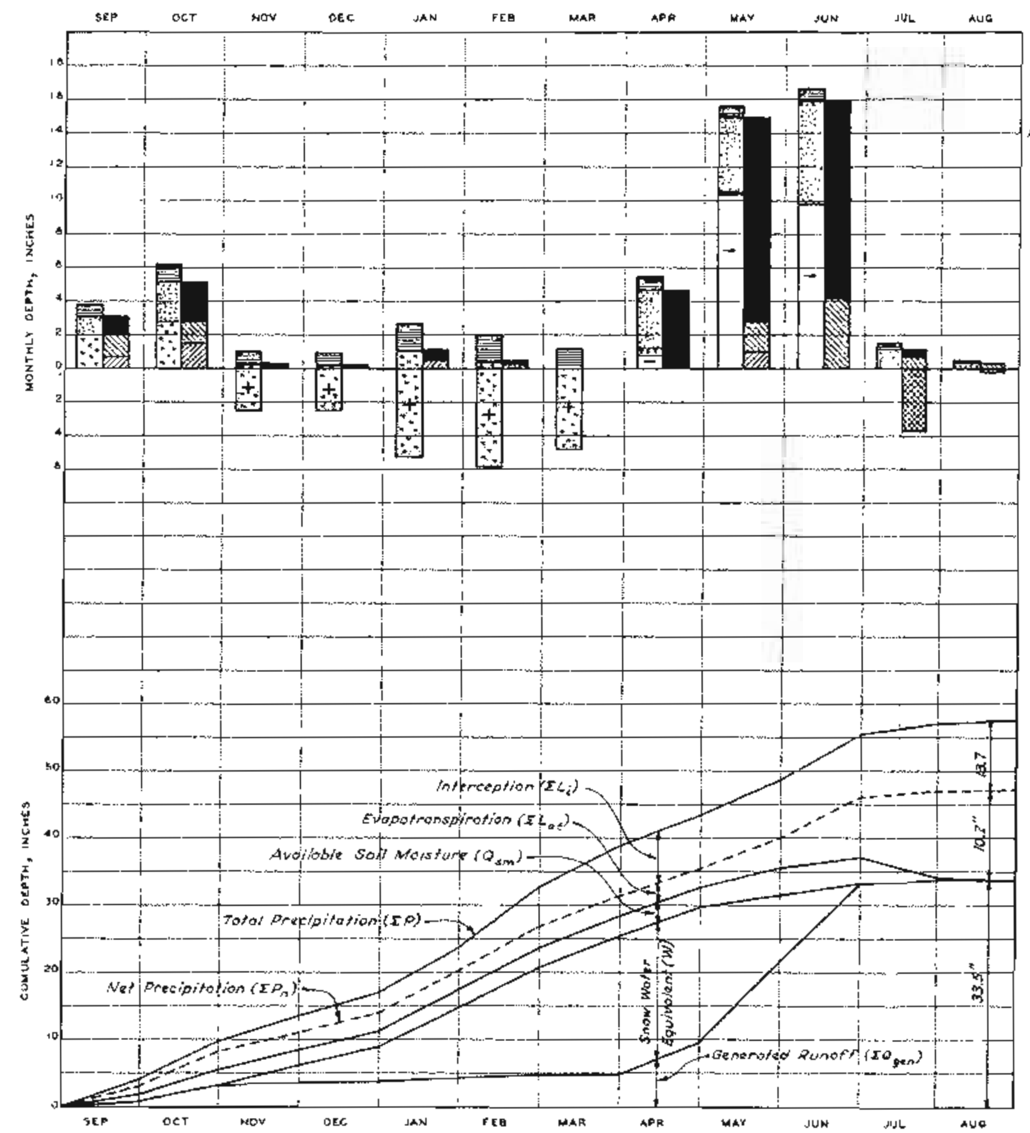
PREPARED SM, MW	SUBMITTED SM	TO ACCOMPANY REPORT DATED
DRAWN HEB	APPROVED DMR	PD-20-25/17

PLATE 4-2



HYDROLOGIC BALANCE, 1946-47 WATER YEAR

FIGURE 1



HYDROLOGIC BALANCE, 1947-48 WATER YEAR

FIGURE 2

LEGEND

- Total interception loss, L_i
- Net rainfall, P_{rn} = rainfall minus rain interception
- P_{sn}: Net snowfall, melted; Net snowfall, accumulated = positive water equivalent change, ΔW
- Ablation of snow on ground = negative water equivalent change, ΔW
- Soil-moisture change, ΔQ_{sm}
- Evapotranspiration, L_{et}
- Generated runoff, Q_{gen}

ΣP is the accumulated total precipitation above tree crown level.
ΣL_i is the accumulated interception loss.
ΣL_{et} is the accumulated evapotranspiration loss.
Q_{sm} is the available soil moisture (Assigned max. Q_{sm} = 4.00")
ΣQ_{gen} is the accumulated generated runoff.
W is the water equivalent of snow on the basin

Water-balance Equations (for monthly values)

$$P - L_i - \Delta W = \Delta Q_{sm} + L_{et} + Q_{gen}$$
$$P - L_i = P_{sn} + P_{rn}$$

Notes:

1. All values shown are average basin amounts.
2. Snow water equivalent changes (ΔW) are shown as "+" or "−" in accordance with the net change during the month.
3. Soil-moisture net change (ΔQ_{sm}) is "+" above and "−" below the "0" reference line.

SNOW INVESTIGATIONS SUMMARY REPORT

SNOW HYDROLOGY

WATER BALANCE, UCSL 1946-47 AND 1947-48

SKYLAND CREEK - DRAINAGE AREA 8.09 SQ. MI.

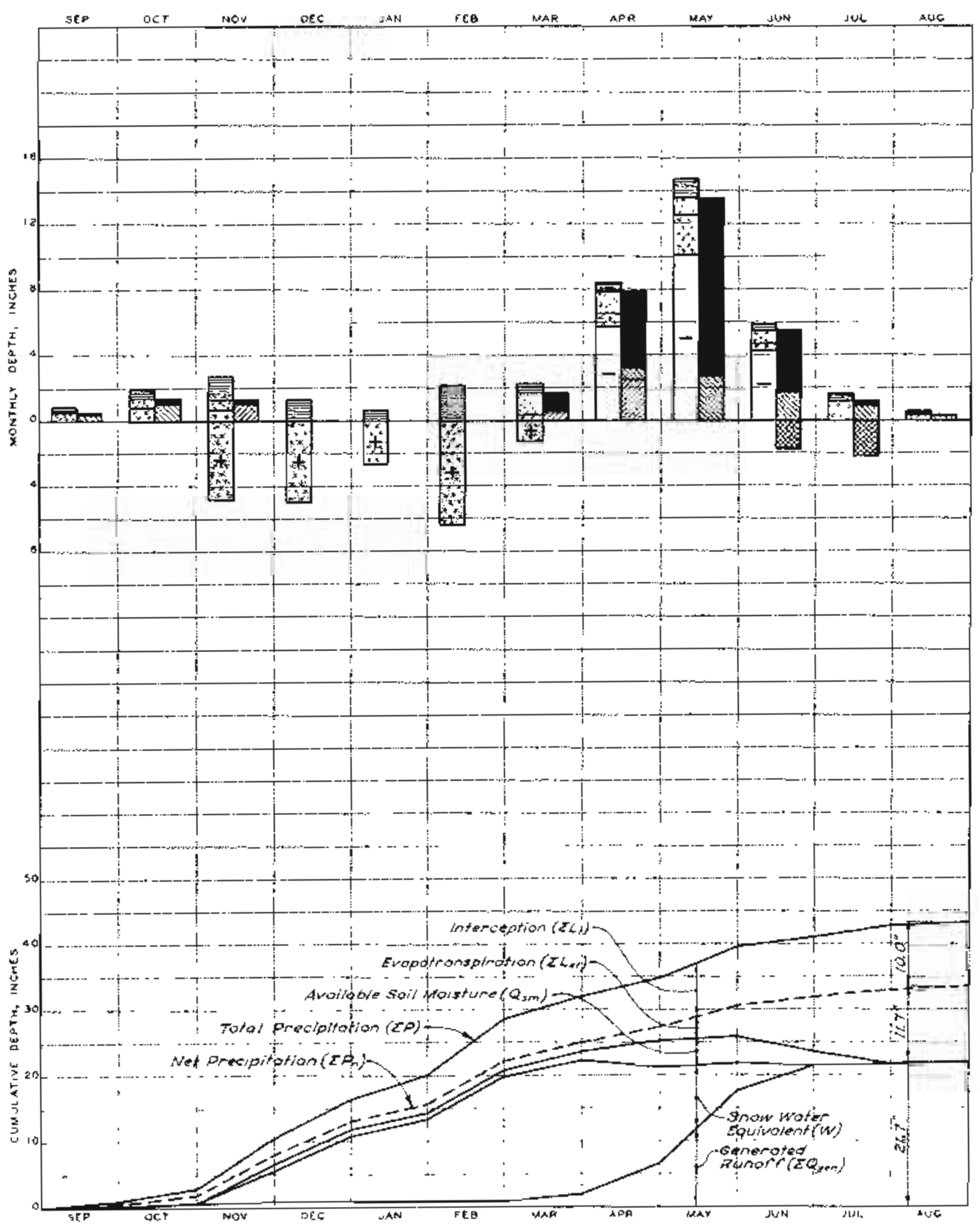
OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U.S. ARMY

PREPARED SM MW. SUBMITTED SM. DRAWN APPROVED DMR

TO ACCOMPANY REPORT DATED 30 JUNE 1956

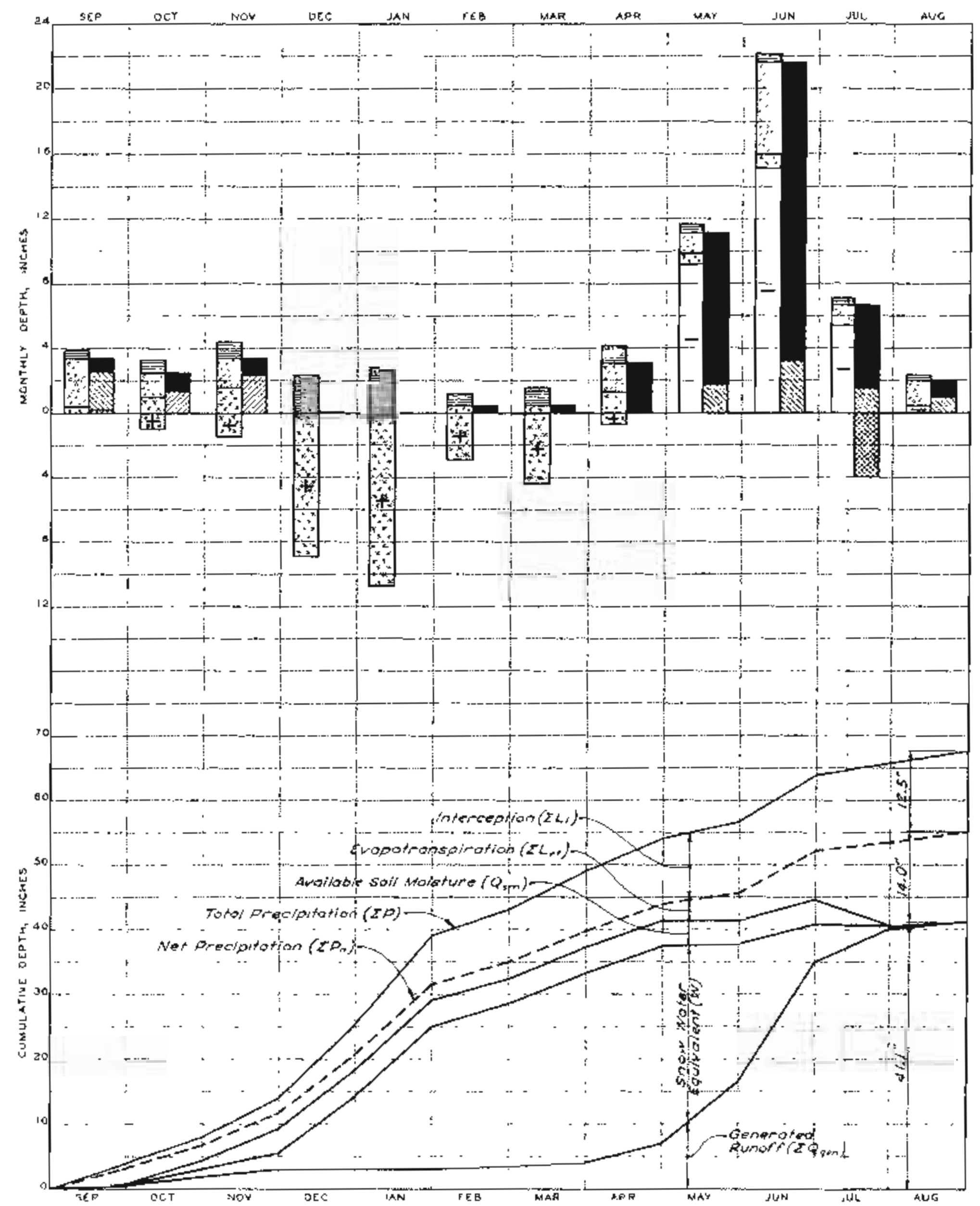
PD-20-25/18

PLATE 4-3



HYDROLOGIC BALANCE, 1948-49 WATER YEAR

FIGURE 1



HYDROLOGIC BALANCE, 1949-50 WATER YEAR

FIGURE 2

LEGEND

- *Total interception loss, L_i*
- *Net rainfall, P_{rn} = rainfall minus rain interception*
- P_{sn} *Net snowfall, melted*
- P_{sn} *Net snowfall, accumulated = positive water equivalent change, ΔW*
- *Ablation of snow on ground = negative water equivalent change, ΔW*
- *Soil-moisture change, ΔQ_{sm}*
- *Evapotranspiration, L_{et}*
- *Generated runoff, Q_{gen}*

ΣP is the accumulated total precipitation above tree crown level.
ΣL_i is the accumulated interception loss.
ΣL_{et} is the accumulated evapotranspiration loss.
Q_{sm} is the available soil moisture (Assigned max Q_{sm} = 4.00")
ΣQ_{gen} is the accumulated generated runoff.
W is the water equivalent of snow on the basin.

Water-balance Equations (for monthly values)

$$P - L_i - \Delta W = \Delta Q_{sm} + L_{et} + Q_{gen}$$
$$P - L_i = P_{sn} + P_{rn}$$

Notes:

1. *All values shown are average basin amounts.*
2. *Snow water equivalent changes (ΔW) are shown as "+" or "–" in accordance with the net change during the month.*
3. *Soil-moisture net change (ΔQ_{sm}) is "+" above and "–" below the "0" reference line.*

SNOW INVESTIGATIONS
SUMMARY REPORT

SNOW HYDROLOGY

WATER BALANCE, UCSL
1948-49 AND 1949-50

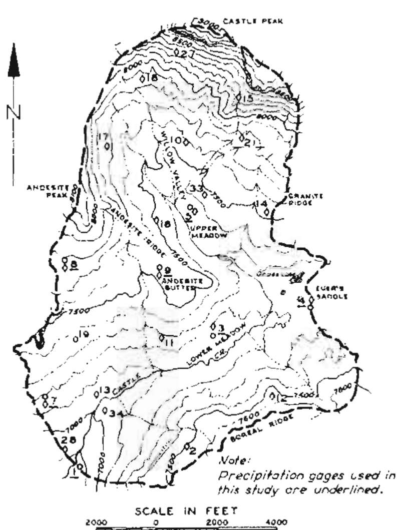
SKYLAND CREEK - DRAINAGE AREA 8.09 SQ. MI.

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

PREPARED [illegible] SUBMITTED [illegible] TO ACCOMPANY REPORT DATED 30 JUNE 1956
DRAWN [illegible] APPROVED [illegible]

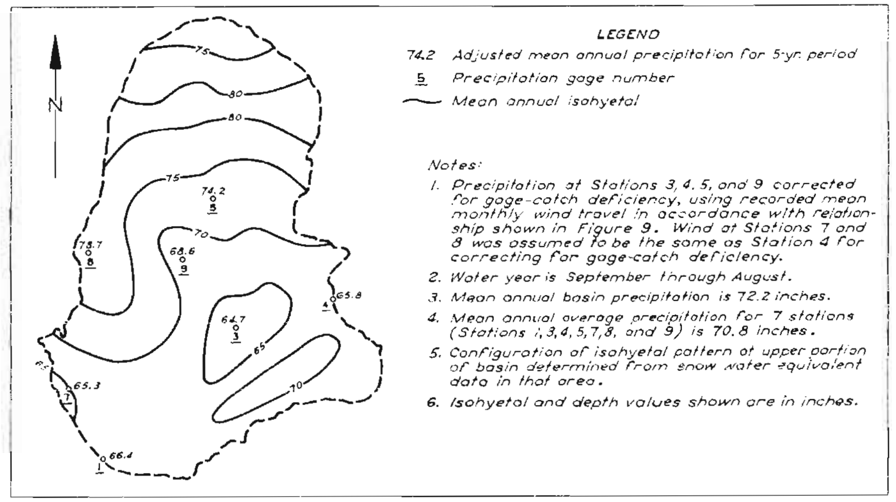
PD-20-25/19

PLATE 4-4



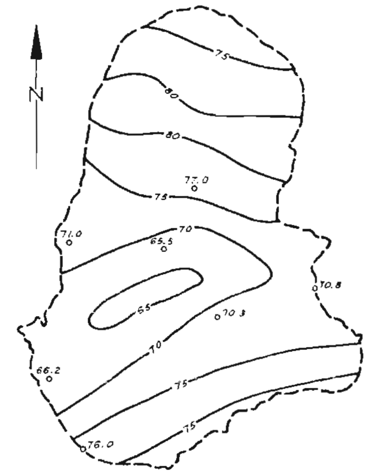
PRECIPITATION GAGES AND SNOW COURSES

FIGURE 1



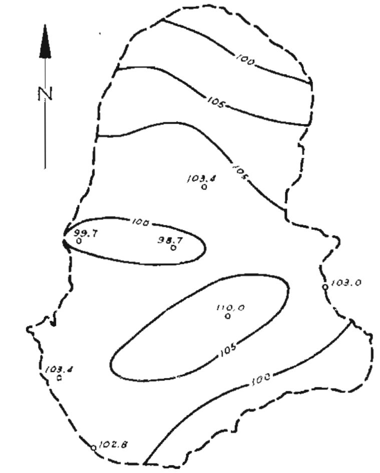
MEAN ANNUAL ISOHYETAL MAP, 1946-47 THROUGH 1950-51

FIGURE 2



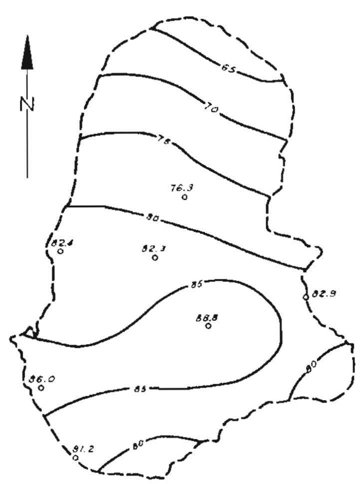
ISOPERCENTUAL MAP
1946-47 WATER YEAR

FIGURE 3



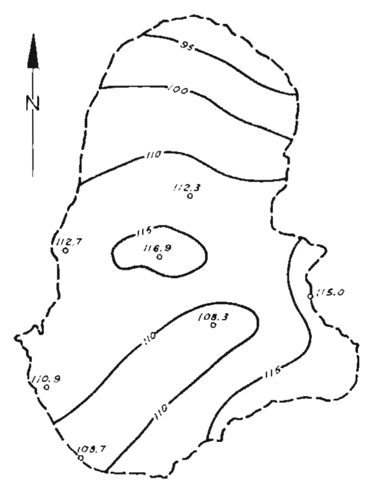
ISOPERCENTUAL MAP
1947-48 WATER YEAR

FIGURE 4



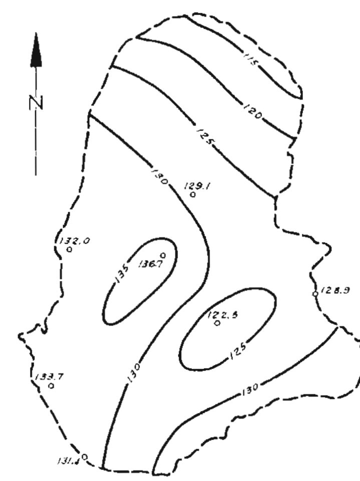
ISOPERCENTUAL MAP
1948-49 WATER YEAR

FIGURE 5



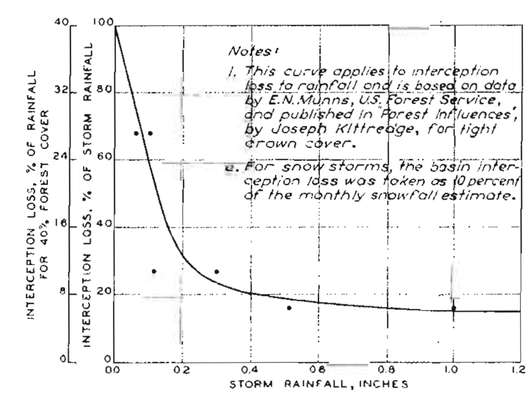
ISOPERCENTUAL MAP
1949-50 WATER YEAR

FIGURE 6



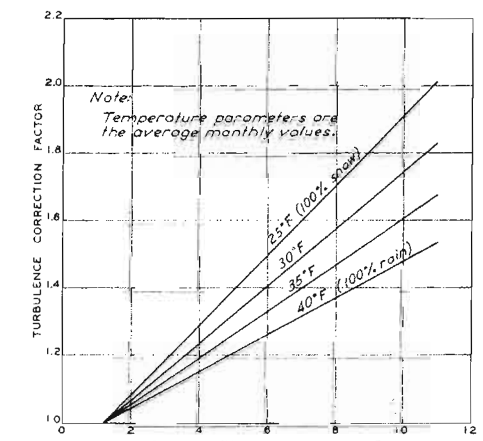
ISOPERCENTUAL MAP
1950-51 WATER YEAR

FIGURE 7



INTERCEPTION LOSS FROM RAIN

FIGURE 8

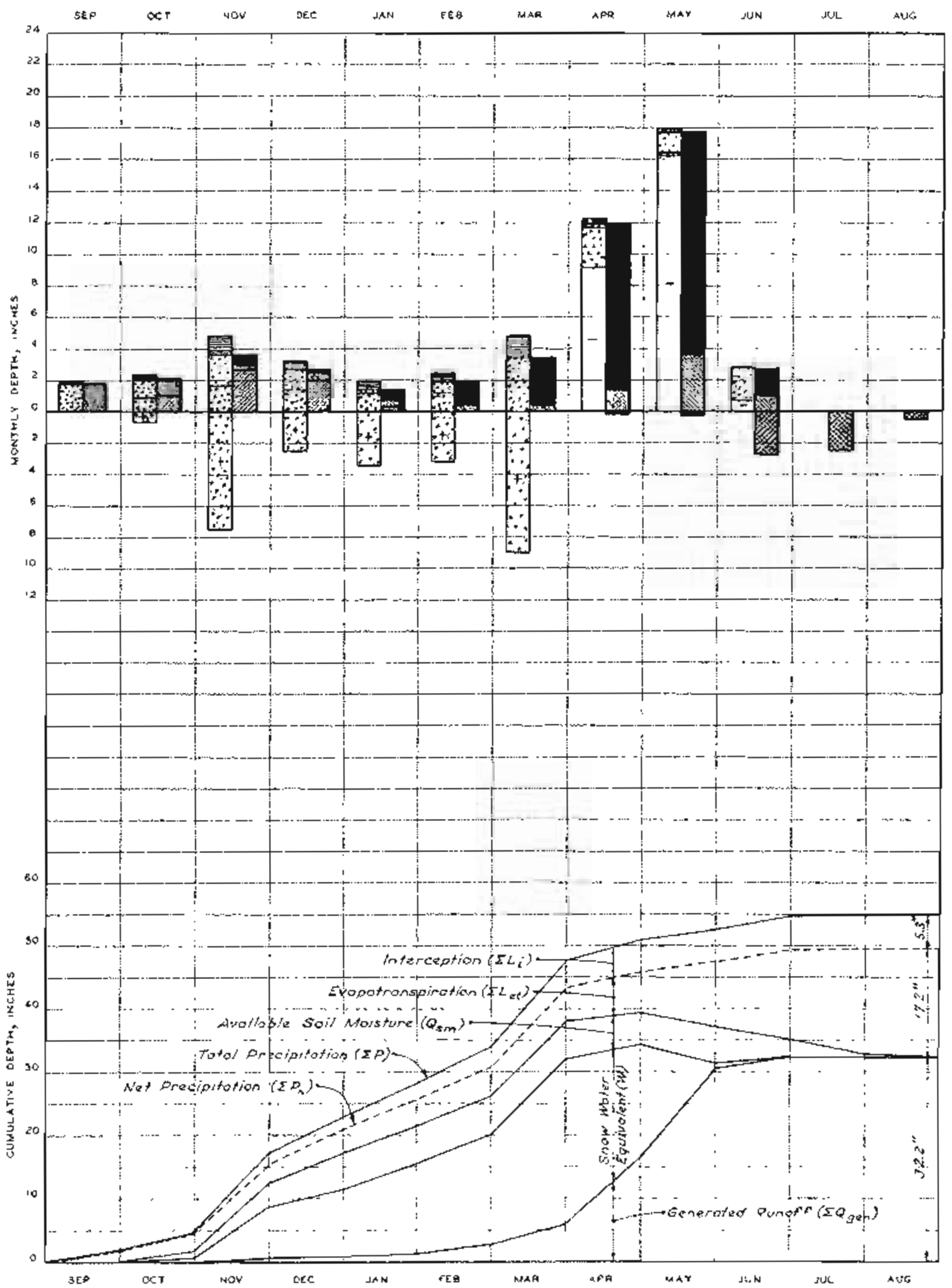


TURBULENCE CORRECTION FACTOR
FOR PRECIPITATION GAGES

FIGURE 9

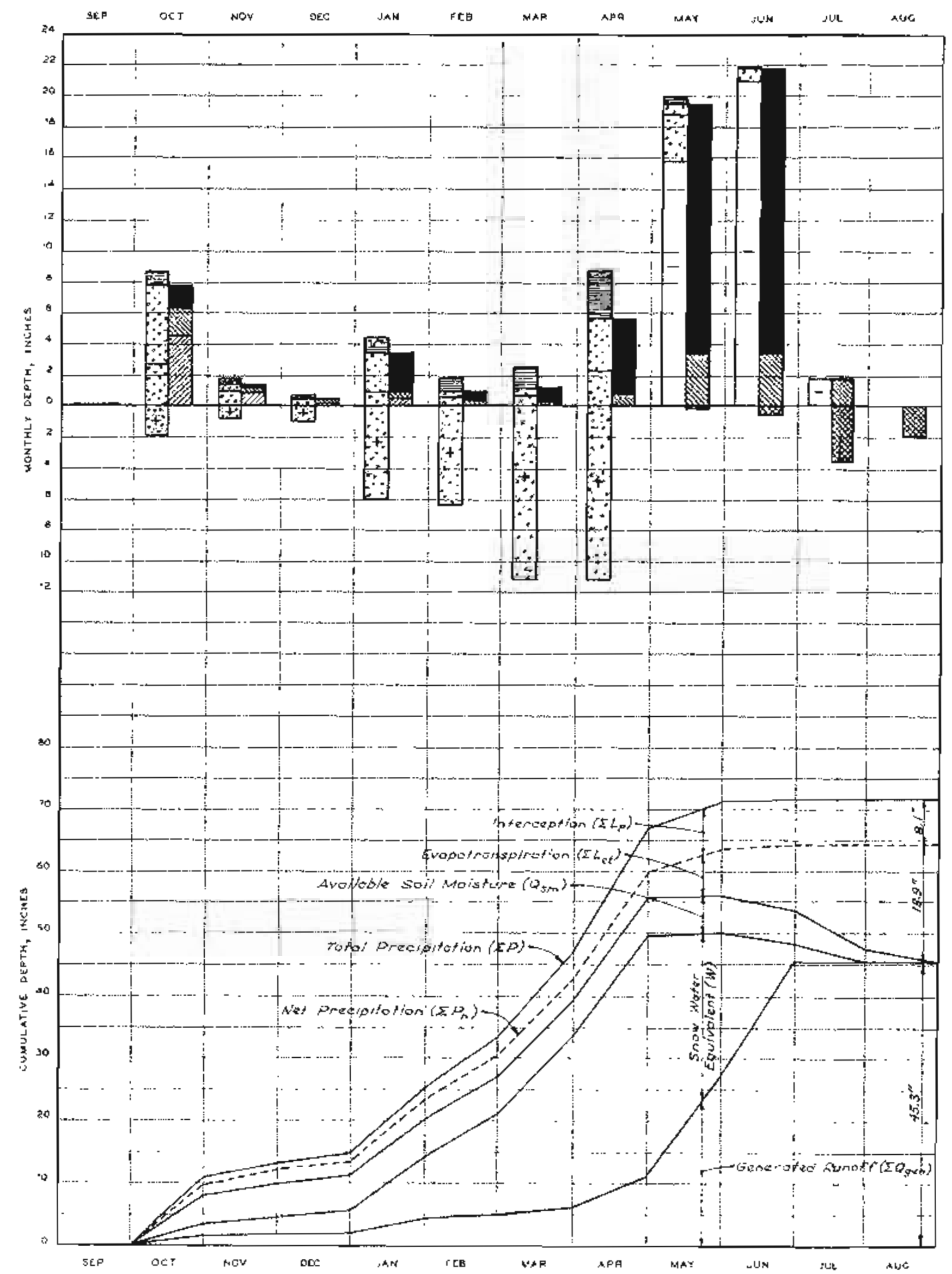
Note:
Isopercentual lines shown in Figures 3 through 7 show water year precipitation in percent of the 5-year average, 1946-47 through 1950-51.

SNOW INVESTIGATIONS SUMMARY REPORT		
SNOW HYDROLOGY PRECIPITATION CHARACTERISTICS CSSL CASTLE CREEK - DRAINAGE AREA 3.96 SQ. MI.		
OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION CORPS OF ENGINEERS U. S. ARMY		
PREPARED: F.K.B. DRAWN: B.V.	SUBMITTED: F.K.B. APPROVED: S.M.K.	TO ACCOMPANY REPORT DATED 30 JUNE 1956 PD-20-25/20



HYDROLOGIC BALANCE, 1946-1947 WATER YEAR

FIGURE 1



HYDROLOGIC BALANCE, 1947-1948 WATER YEAR

FIGURE 2

LEGEND:

- Total interception loss, L_i
- Net rainfall, P_{rn} = rainfall minus rain interception
- P_{sn} Net snowfall, melted
- Net snowfall, accumulated = positive water equivalent change, ΔW
- Ablation of snow on ground = negative water equivalent change, ΔW
- Soil-moisture change, ΔQ_{sm}
- Evapotranspiration, L_{et}
- Generated runoff, Q_{gen}

ΣP is the accumulated total precipitation above tree crown level.

ΣL_i is the accumulated interception loss.

ΣL_{et} is the accumulated evapotranspiration loss.

Q_{sm} is the available soil moisture (Assigned max Q_{sm}=6.00")

ΣQ_{gen} is the accumulated generated runoff.

W is the water equivalent of snow on the basin.

WATER-BALANCE EQUATIONS (for monthly values)

$$P - L_i - \Delta W = \Delta Q_{sm} + L_{et} + Q_{gen}$$

$$P - L_i = P_{sn} + P_{rn}$$

NOTES:

1. All values shown are average basin amounts
2. Snow water equivalent changes (ΔW) are shown as "+" or "−" in accordance with the net change during the month
3. Soil-moisture net change (ΔQ_{sm}) is "+" above and "−" below the "0" reference line

SNOW INVESTIGATIONS
SUMMARY REPORT
SNOW HYDROLOGY

**WATER BALANCE, CSSL
1946-47 AND 1947-48**

CASTLE CREEK - DRAINAGE AREA 3.98 SQ. MI.

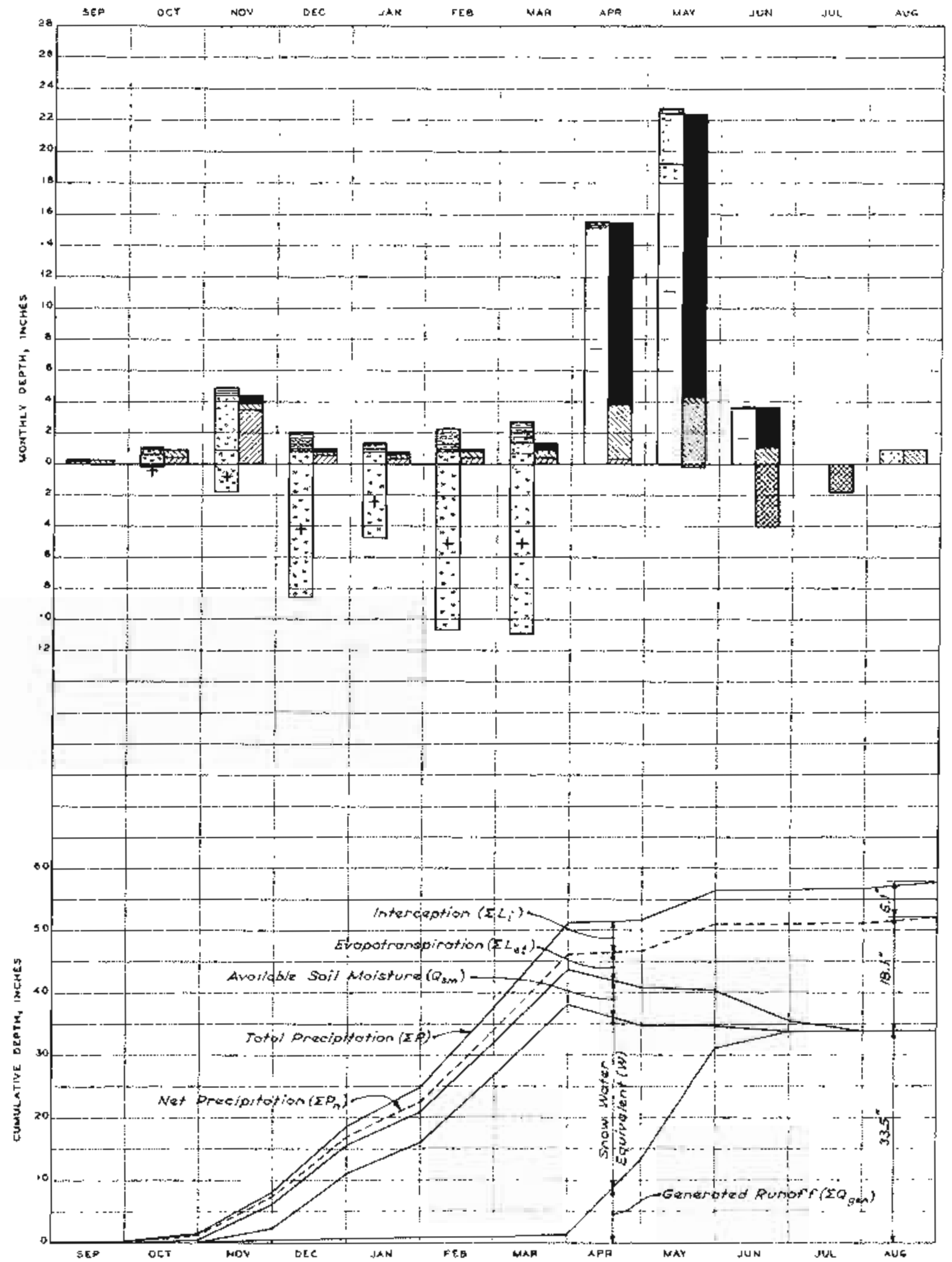
OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

PREPARED F.S.B. SUBMITTED F.S.B.
DRAWN F.W.W. APPROVED D.M.R.

TO ACCOMPANY REPORT DATED 30 JUNE 1955

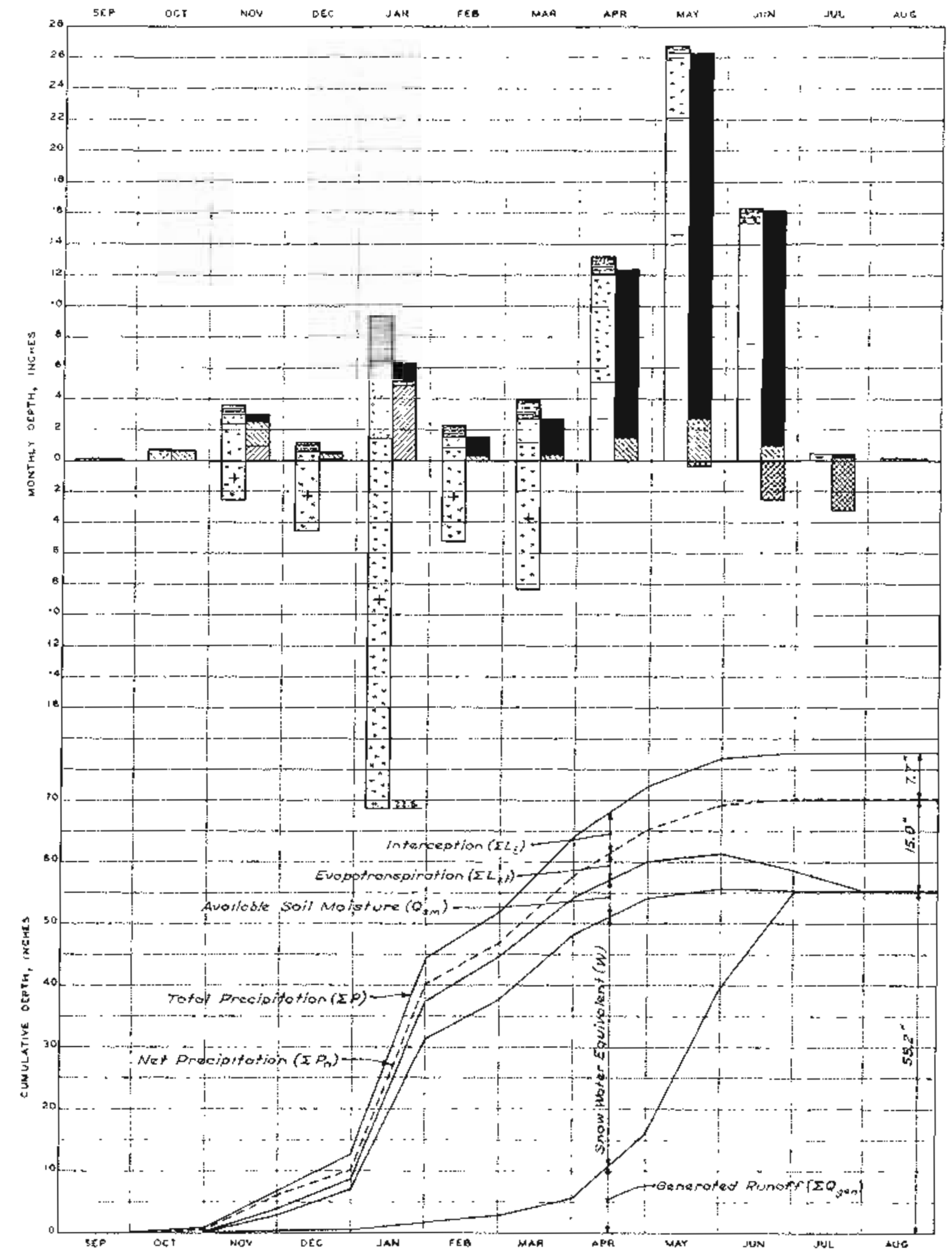
PD-20-25/21

PLATE 4-6



HYDROLOGIC BALANCE, 1948-49 WATER YEAR

FIGURE 1



HYDROLOGIC BALANCE, 1949-50 WATER YEAR

FIGURE 2

LEGEND

- Total interception loss, L_i
- Net rainfall, P_{rn} = rainfall minus rain interception
- P_{sn}: Net snowfall, melted
- P_{sn}: Net snowfall, accumulated = positive water equivalent change, ΔW
- Ablation of snow on ground = negative water equivalent change, ΔW
- Soil-moisture change, ΔQ_{sm}
- Evapotranspiration, L_{et}
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ΣL_i is the accumulated interception loss.
ΣL_{et} is the accumulated evapotranspiration loss
Q_{sm} is the available soil moisture (Assigned max Q_{sm} = 6.00")
ΣQ_{gen} is the accumulated generated runoff.
W is the water equivalent of snow on the basin.

Water-balance Equations (for monthly values)

$$P - L_i = \Delta W \mp \Delta Q_{sm} + L_{et} + Q_{gen}$$
$$P - L_p = P_{sn} + P_{rn}$$

Notes:

1. All values shown are average basin amounts.
2. Snow water equivalent changes (ΔW) are shown as "+" or "-" in accordance with the net change during the month.
3. Soil-moisture net change (ΔQ_{sm}) is "+" above and "-" below the "0" reference line.

SNOW INVESTIGATIONS
SUMMARY REPORT

SNOW HYDROLOGY

WATER BALANCE, CSSL
1948-49 AND 1949-50

CASTLE CREEK - DRAINAGE AREA 3.98 SQ. MI.

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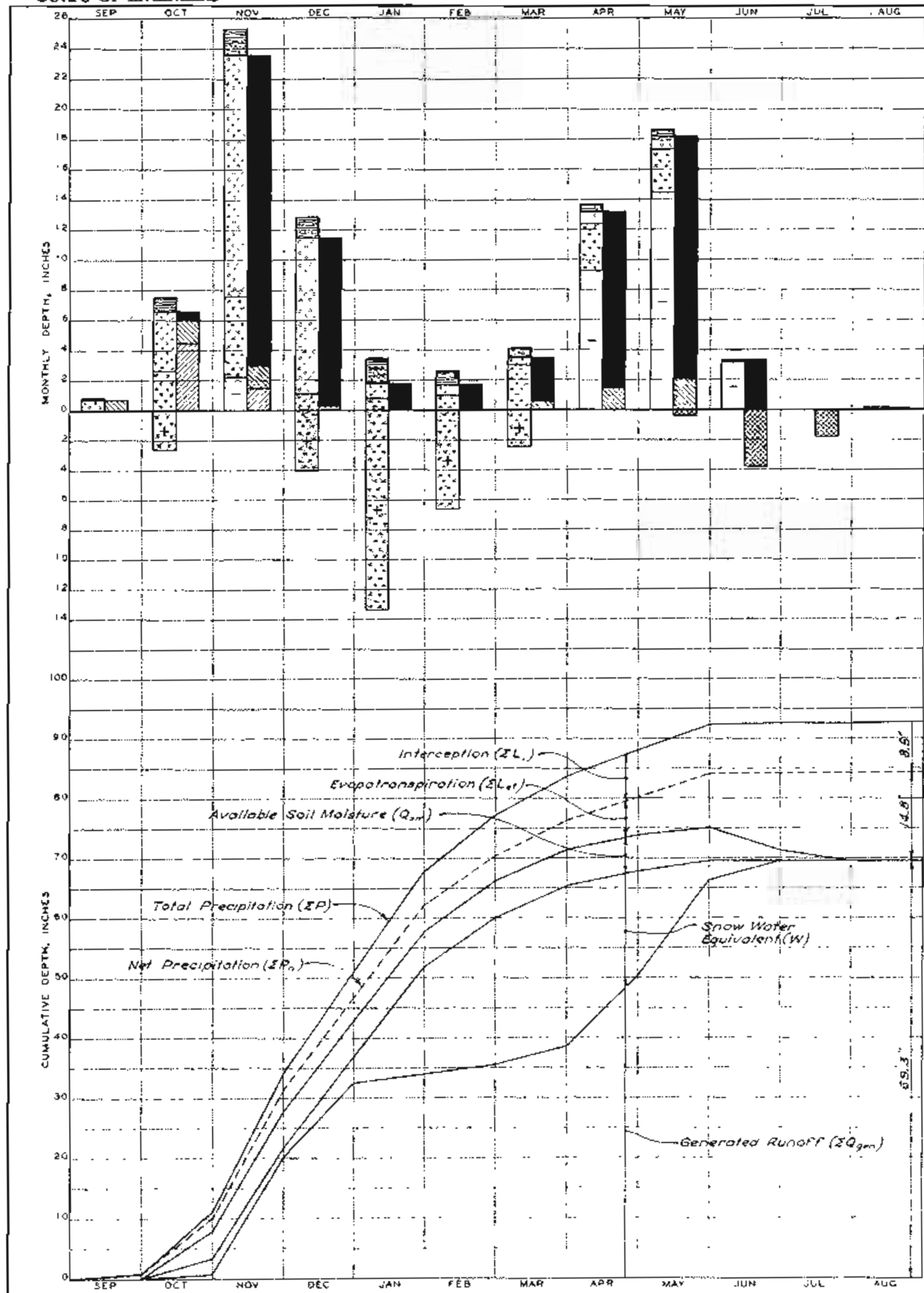
DRAWN	APPROVED DMR	PD-20-25/22

PLATE 4-7



LEGEND

- Total interception loss, L_i
- Net rainfall, P_{rn} = rainfall minus rain interception
- P_{sn}: Net snowfall, melted
- Net snowfall, accumulated = positive water equivalent change, ΔW
- Ablation of snow on ground = negative water equivalent change, ΔW
- Soil-moisture change, ΔQ_{sm}
- Evapotranspiration, L_{et}
- Generated runoff, Q_{gen}

ΣP is the accumulated total precipitation above tree crown level
ΣL_i is the accumulated interception loss
ΣL_{et} is the accumulated evapotranspiration loss
Q_{sm} is the available soil moisture (Assigned maximum Q_{sm} = 6.00")
ΣQ_{gen} is the accumulated generated runoff
W is the water equivalent of snow on the basin

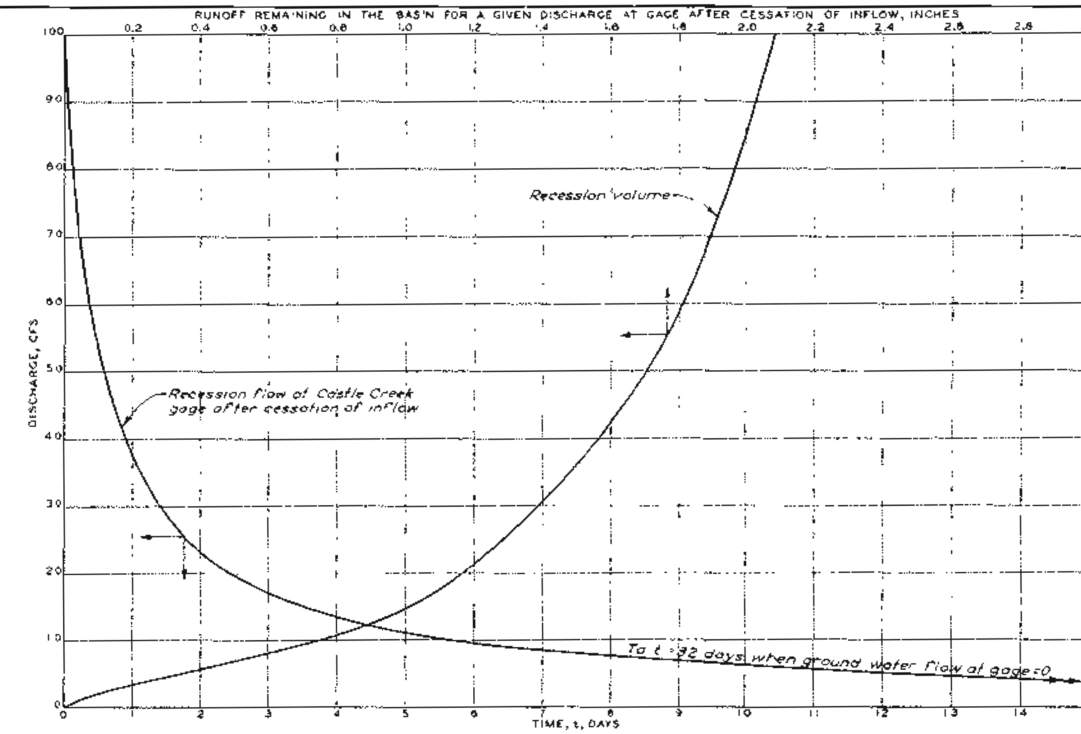
Notes:
1. All values shown are average basin amounts.
2. Snow water equivalent changes (ΔW) are shown as "+" or "-" in accordance with the net change during the month.
3. Soil-moisture net change (ΔQ_{sm}) is "+" above and "-" below the "0" reference line.

Water-balance Equations (for monthly values)

$$P - L_i - \Delta W = \Delta Q_{sm} + L_{et} + Q_{gen}$$
$$P - L_i = P_{rn} + P_{sn}$$

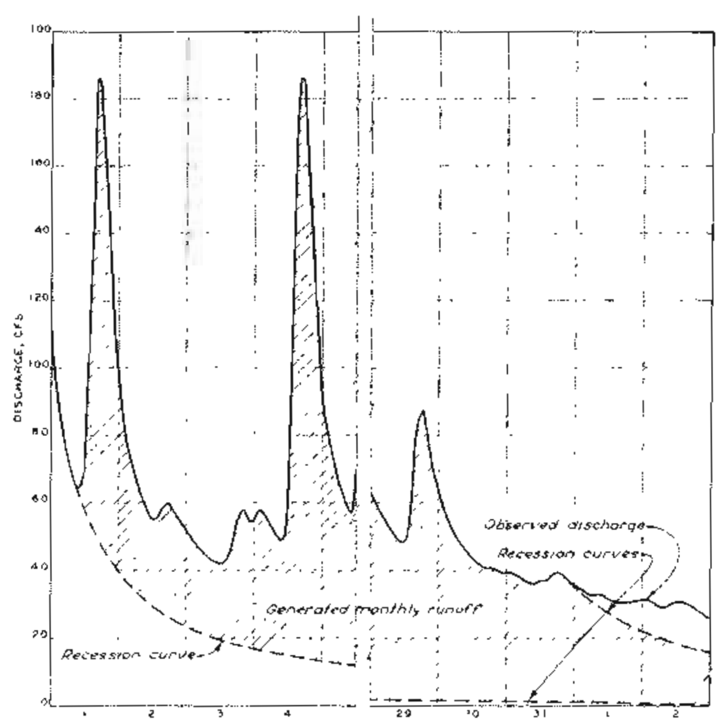
HYDROLOGIC BALANCE, 1950-51 WATER YEAR

FIGURE 1



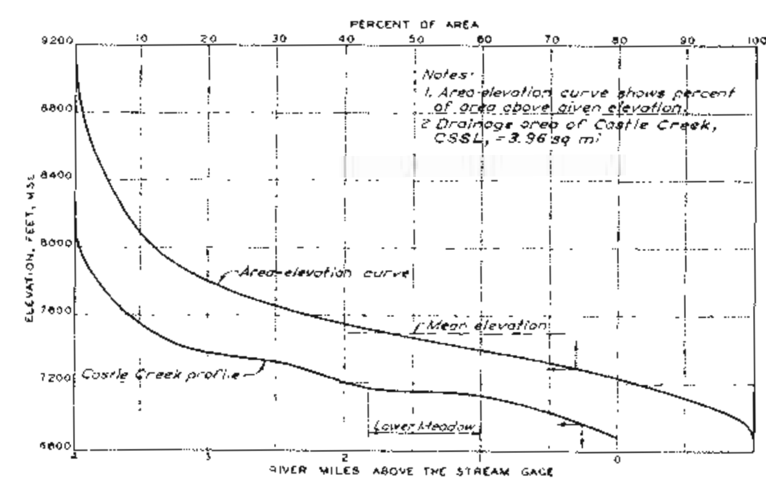
HYDROGRAPH CHARACTERISTICS

FIGURE 2



TYPICAL DAILY HYDROGRAPHS, ILLUSTRATING MONTHLY RUNOFF EVALUATION

FIGURE 3



BASIN CHARACTERISTICS

FIGURE 4

SNOW INVESTIGATIONS
SUMMARY REPORT

SNOW HYDROLOGY

WATER BALANCE, CSSL, 1950-51 AND MISCELLANEOUS RELATIONSHIPS

CASTLE CREEK - DRAINAGE AREA 3.96 SQ. MI.

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
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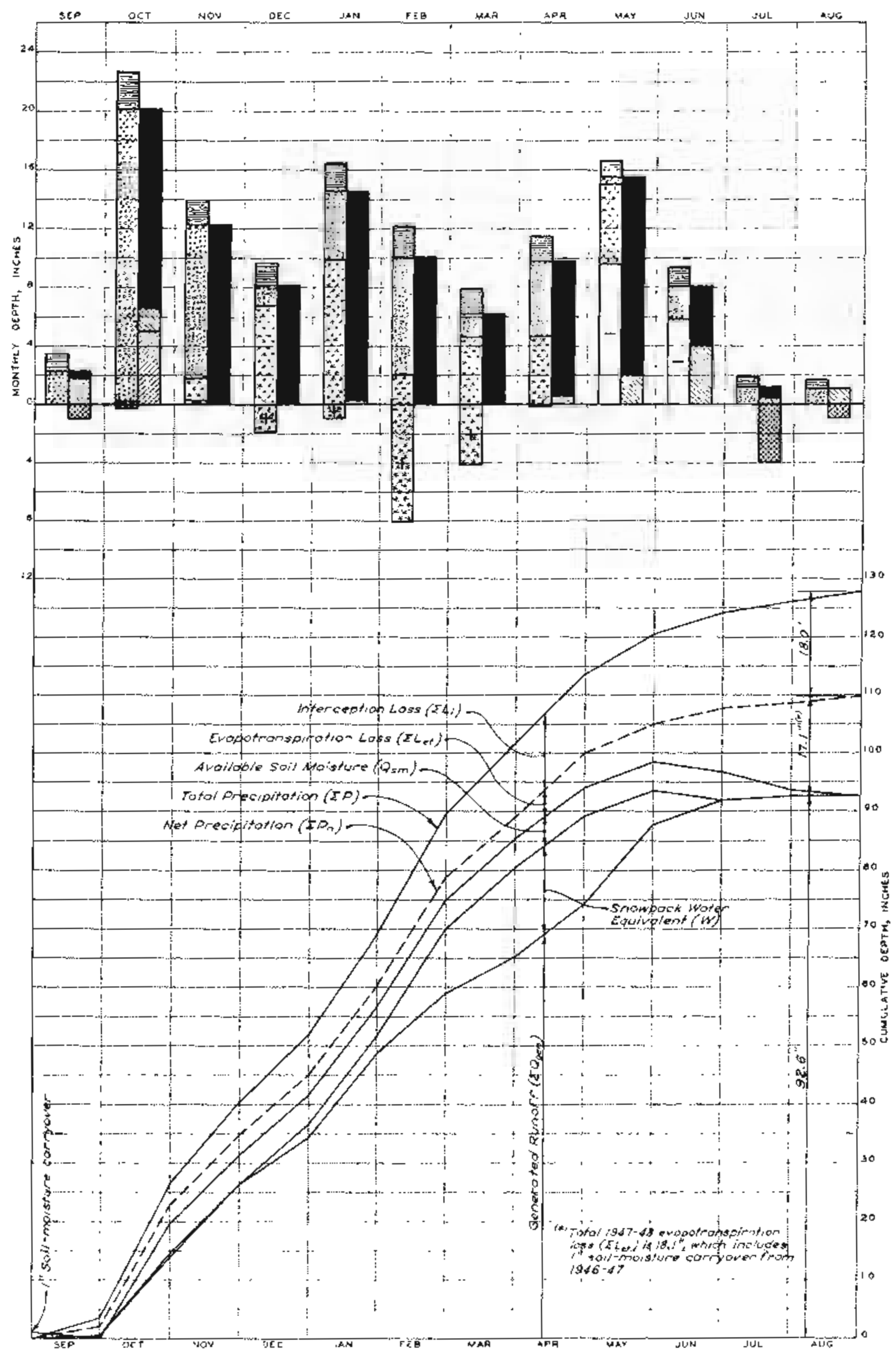
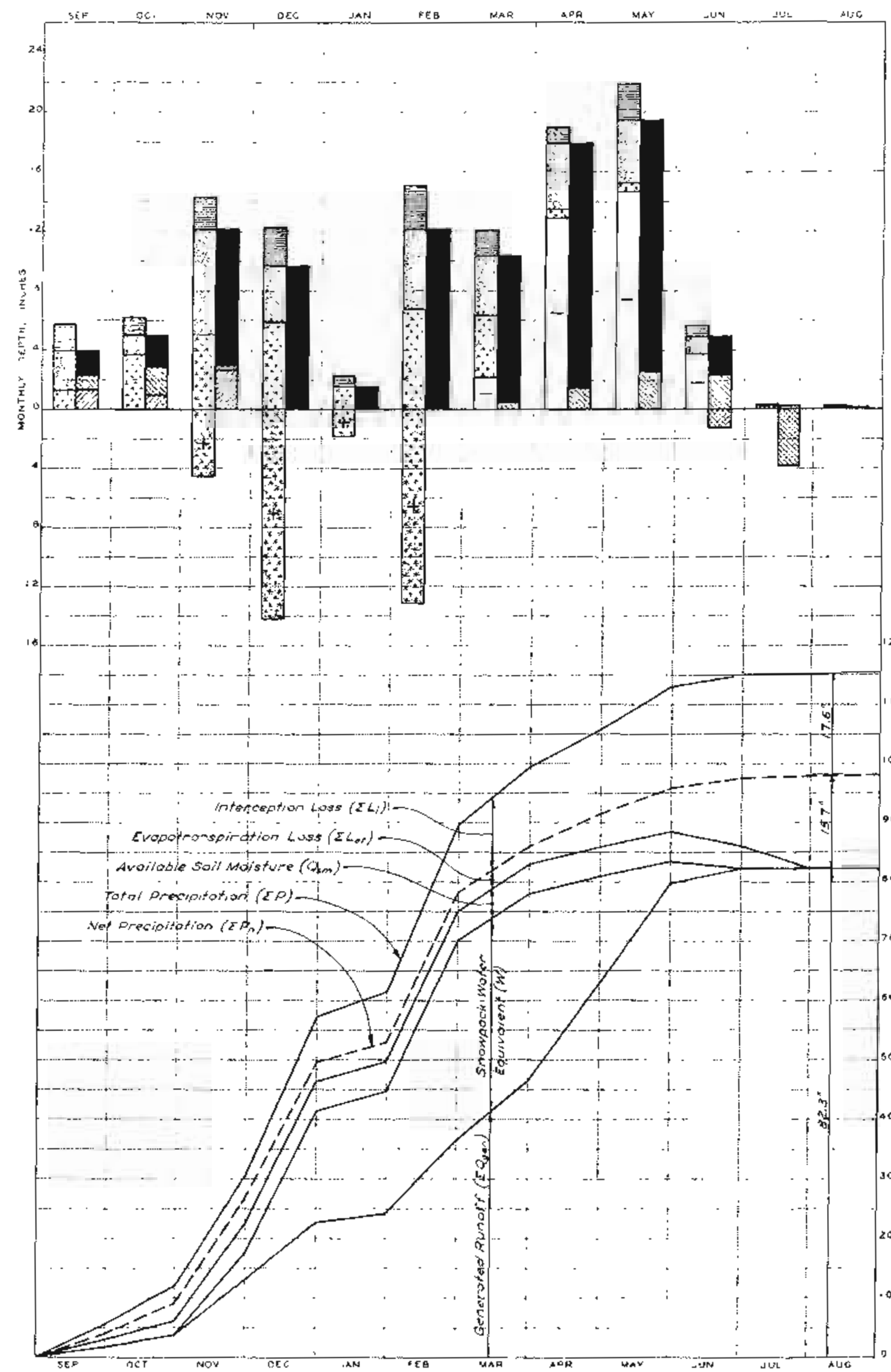
PLATE 4-8



HYDROLOGIC BALANCE, 1947-48 WATER YEAR

FIGURE 1



HYDROLOGIC BALANCE, 1948-49 WATER YEAR

FIGURE 2

LEGEND

- Total interception loss, L_i
- Net rainfall, P_{rn} = rainfall minus rain interception
- P_{sn}: Net snowfall, melted
- P_{sn}: Net snowfall accumulated = positive change in snowpack water equivalent, ΔW
- Ablation of snow on ground = negative change in snowpack water equivalent, ΔW
- Soil-moisture change, ΔQ_{sm}
- Evapotranspiration loss, L_{et}
- Generated runoff, Q_{gen}

ΣP is the accumulated total basin precipitation, above level of tree crowns.
ΣL_i is the accumulated interception loss.
ΣL_{et} is the accumulated evapotranspiration loss.
Q_{sm} is the available soil moisture (Assigned max. Q_{sm} = 5.0")
ΣQ_{gen} is the accumulated generated runoff.
W is the basin snowpack water equivalent.

Water-balance Equations (for monthly values)

$$P - L_i - \Delta W = \Delta Q_{sm} + L_{et} + Q_{gen}$$
$$P - L_i = P_{sn} + P_{rn}$$

Notes:

1. All values shown are average basin amounts.
2. Snowpack water equivalent changes (ΔW) are shown as "+" or "-" in accordance with the net change during the month.
3. Soil-moisture net change (ΔQ_{sm}) is "+" above and "-" below the "0" reference line.
4. WBSL basin elevations:
   Minimum, 1960 feet msl
   Maximum, 5364 " "
   Mean, 3430 " "

SNOW INVESTIGATIONS
SUMMARY REPORT

SNOW HYDROLOGY

WATER BALANCE, WBSL
1947-48 AND 1948-49

BLUE RIVER ABOVE QUENTIN CREEK
DRAINAGE AREA 11.5 SQ. MI

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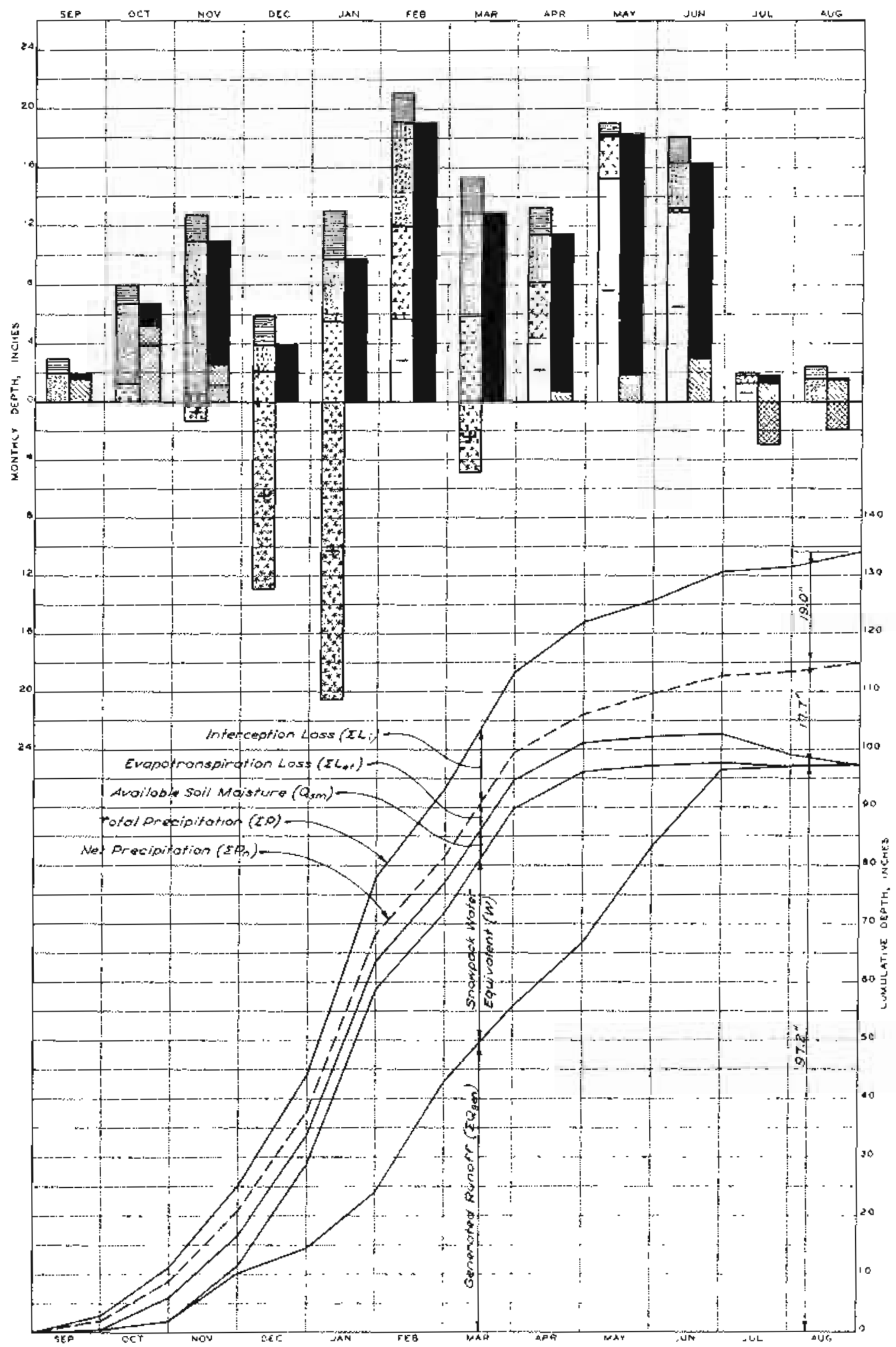
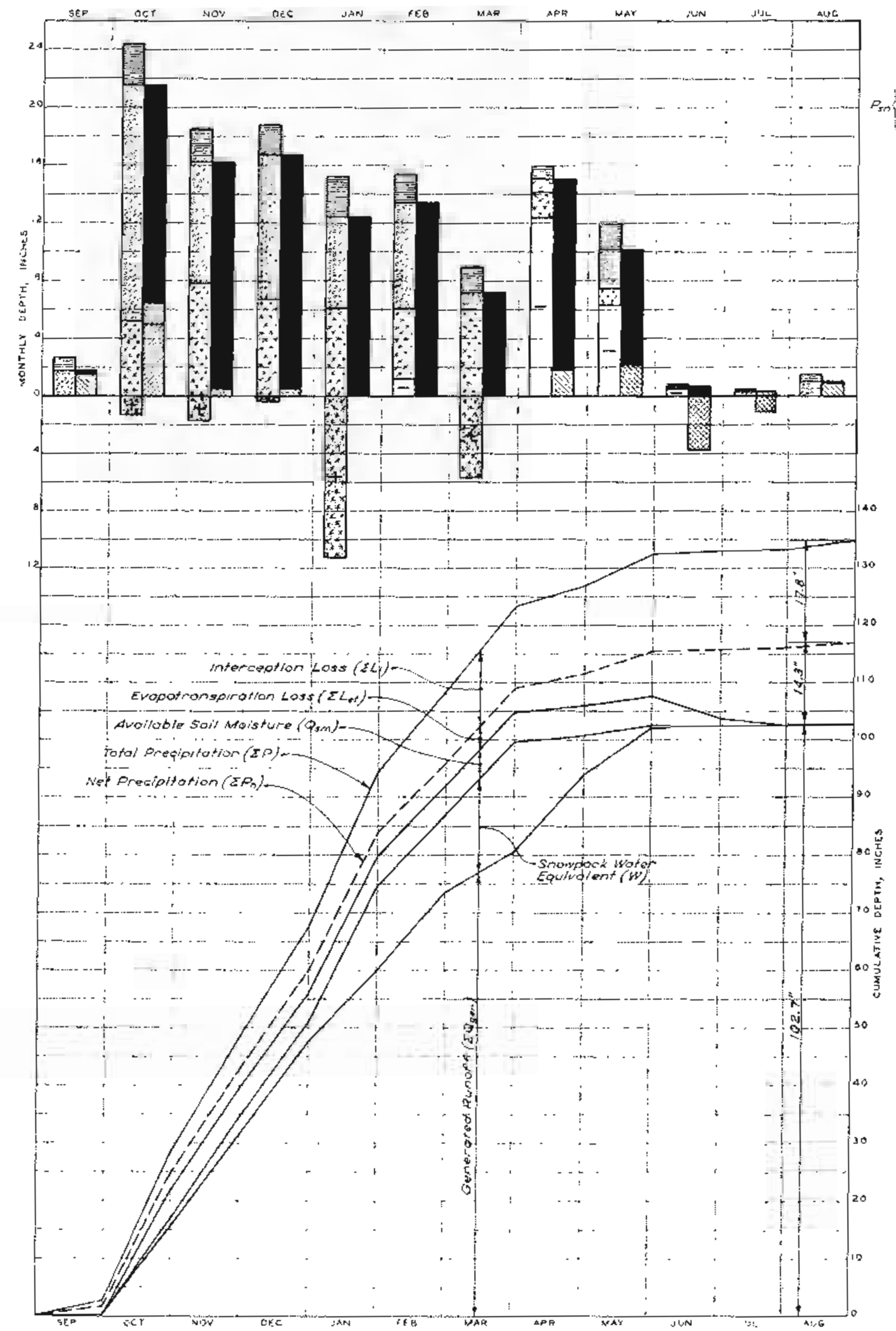


HYDROLOGIC BALANCE, 1949-50 WATER YEAR

FIGURE 1



HYDROLOGIC BALANCE, 1950-51 WATER YEAR

FIGURE 2

LEGEND

- Total interception loss, L_i
- Net rainfall, P_{rn} = rainfall minus rain interception
- P_{sn}: Net snowfall, melted
- P_{sn}: Net snowfall accumulated = positive change in snowpack water equivalent, ΔW
- Ablation of snow on ground = negative change in snowpack water equivalent, ΔW
- Soil-moisture change, ΔQ_{sm}
- Evapotranspiration loss, L_{et}
- Generated runoff, Q_{gen}

ΣP is the accumulated total basin precipitation, above level of tree crowns.
ΣL_i is the accumulated interception loss.
ΣL_{et} is the accumulated evapotranspiration loss.
Q_{sm} is the available soil moisture (Assigned max Q_{sm} = 5.0")
ΣQ_{gen} is the accumulated generated runoff.
W is the basin snowpack water equivalent.

Water-balance Equations (for monthly values)

$$P - L_i - \Delta W = \Delta Q_{sm} + L_{et} + Q_{gen}$$
$$P - L_i = P_{sn} + P_{rn}$$

Notes:

1. All values shown are average basin amounts.
2. Snowpack water equivalent changes (ΔW) are shown as "+" or "-" in accordance with the net change during the month.
3. Soil-moisture net change (ΔQ_{sm}) is "+" above and "-" below the 0" reference line.
4. WBSL basin elevations:
   Minimum, 1960 feet msl
   Maximum, 5364 " "
   Mean, 3430 " "

SNOW INVESTIGATIONS
SUMMARY REPORT

SNOW HYDROLOGY

WATER BALANCE, WBSL
1949-50 AND 1950-51

BLUE RIVER ABOVE QUENTIN CREEK
DRAINAGE AREA 11.5 SQ MI.

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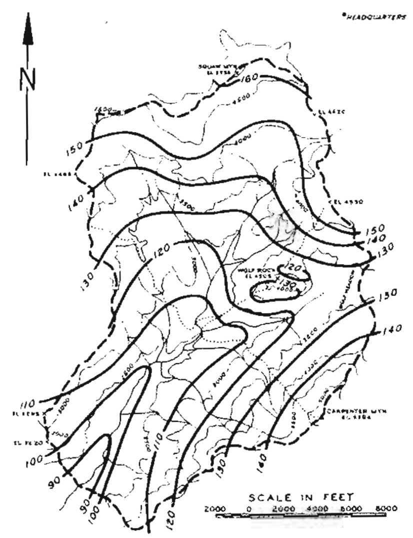
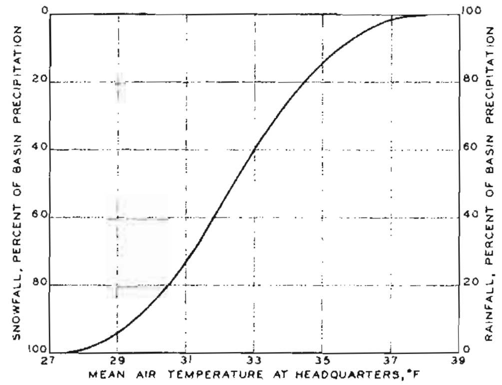
DRAWN: [illegible]	APPROVED: [illegible]	PD-20-25/25

PLATE 4-10



ISOHYETAL MAP
MEAN ANNUAL PRECIPITATION,
1947-48 THROUGH 1950-51, WBSL

FIGURE 1

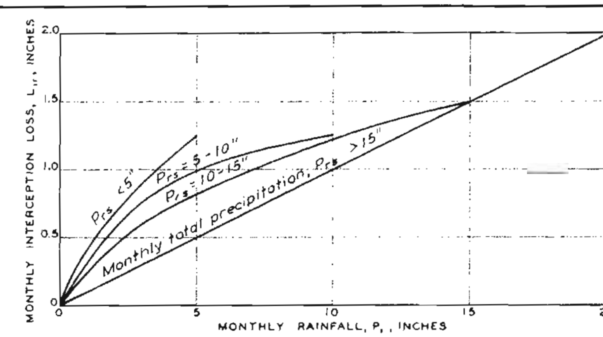


Notes:

1. Curve relates air temperature at the base station (Station 1A, WBSL) to the percentage of snowfall and of rainfall in basin precipitation for the area as a whole. It incorporates the effects of the basin area-elevation and of orographic increase in precipitation. (See Text, par. 4-11.05)
2. Division between rain and snow form of precipitation assumed to be 34.5°F (see Plate 3-1). Since this dividing temperature was determined from simultaneous observations of form of precipitation and air temperature, it should not be used with average temperatures for periods longer than about one day. A wet adiabatic lapse rate was used to extrapolate air temperatures measured at Station 1A to other elevations.

AIR TEMPERATURE VS FORM OF PRECIPITATION, WBSL

FIGURE 2

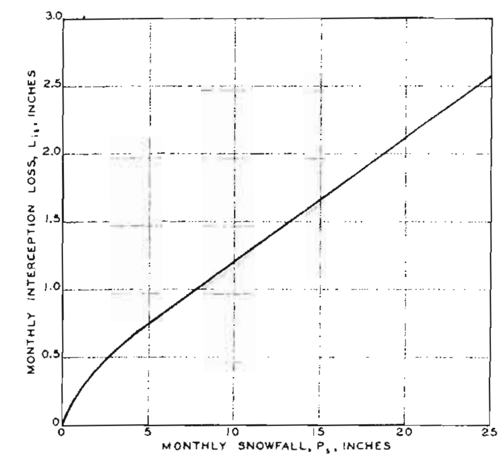


Notes:

1. Parameters are for monthly total precipitation including rain and snow, P_{rs}.
2. The above curves for WBSL were extrapolated from findings on winter rainfall interception loss for other areas [29] and are regarded as rough approximations only.

WINTER-TYPE RAINFALL INTERCEPTION LOSS, WBSL

FIGURE 3

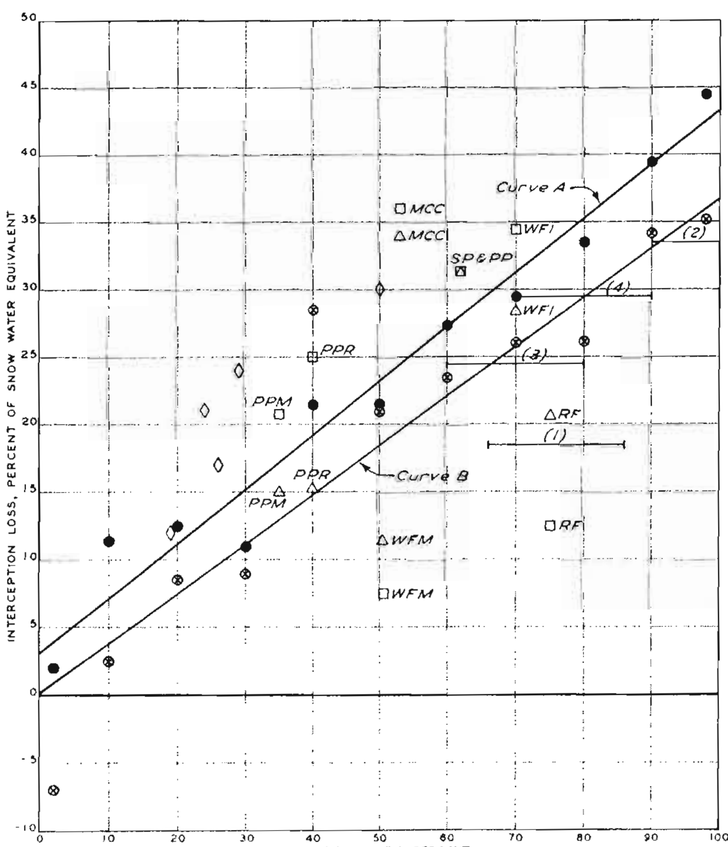


Note:

Curve extrapolated from findings on snowfall interception for individual storms for other areas, [29] on the basis of frequencies of storm occurrence in WBSL.

SNOWFALL INTERCEPTION LOSS, WBSL

FIGURE 4



SNOWFALL INTERCEPTION LOSS (BY SNOWPACK MEASUREMENTS)

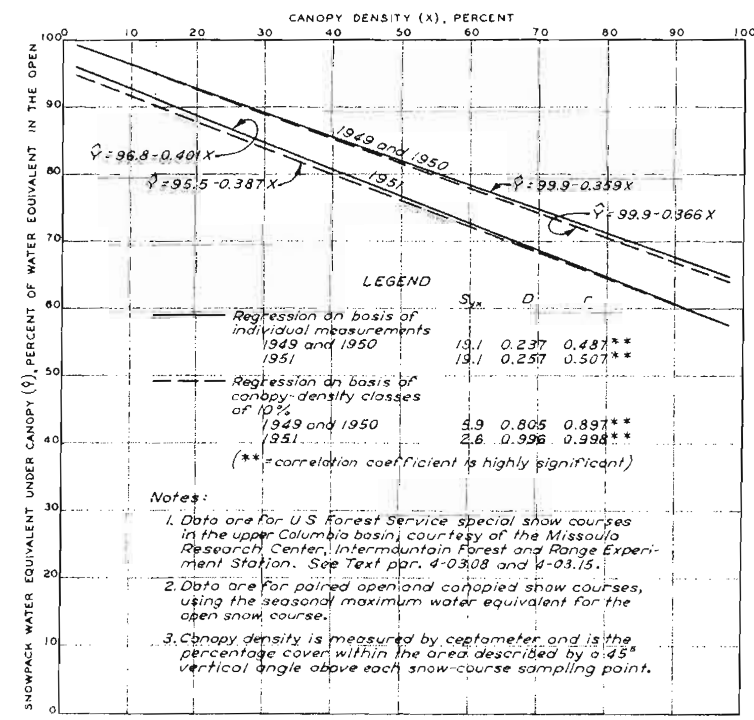
FIGURE 5

L E G E N D

SYMBOL	DATA			
	SOURCE	LOCATION	VEGETATION	REMARKS
●	Ingebo (see par. 4-03.15)	Upper Columbia Basin (4,000-5,000 ft MSL)	Pine	Data for 1951; canopy density measured by ceptometer; regression line = curve A
⊗	Ingebo (see par. 4-03.15)	Upper Columbia Basin (4,000-5,000 ft MSL)	Pine	Data for 1949 & 1950; canopy density measured by ceptometer; regression line = curve B
◊	Wilm & Dunford [38]	Fraser Experiment Forest, Colorado (9,200-9,700 ft MSL)	Lodgepole Pine	Interception loss computed from authors' data
├─┤	Connaughton [5]	Boise R Basin, Idaho	Ponderosa Pine (see remarks)	Canopy density range estimated. Numbers indicate maturity as follows: (1) under crowns <80-years old; (2) under mature dense crowns. Average density of plots having virgin timber: (3) with no young trees; (4) with young trees.
□	Kittredge [17]	Sierra Nevada, Calif (5,000-6,000 ft MSL)	(see remarks)	Letters by points indicate vegetation as follows: PPM, mature ponderosa pine; PPR, ponderosa pine reproduction (14 ft high); WFM, mature white fir; MCC, cutover mixed conifer; SP&PP, sugar pine & ponderosa pine; WFI, immature white fir; RF, red fir
△	Kittredge [17]	Sierra Nevada, Calif (5,000-6,000 ft MSL)	(see remarks)	Interception loss determined by snowboard measurements. Letters by points indicate vegetation as above.

Notes:

1. Methods of determining interception loss and canopy density differed with the different investigations. Most interception loss values were computed as the difference in water equivalents between the maximum snowpack accumulation in the open and that under the various forest canopies. Kittredge used, in addition, differences in snowboard measurements as a measure of interception loss. Ingebo made use of a canopy-density meter to determine canopy density.
2. Horizontal bars indicate probable range of canopy density (no quantitative values given).
3. Curve A fitted by least squares regression to 1949 and 1950 data and curve B to 1951 data from Ingebo's data.



SNOWPACK ACCUMULATION UNDER FOREST CANOPY

FIGURE 6

SNOW INVESTIGATIONS
SUMMARY REPORT

SNOW HYDROLOGY

MISCELLANEOUS RELATIONSHIPS, WBSL
AND INTERCEPTION LOSS

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PREPARED: MHM. SUBMITTED: MHM.
DRAWN: SV. APPROVED: DMB.
TO ACCOMPANY REPORT DATED 30 JUNE 1956

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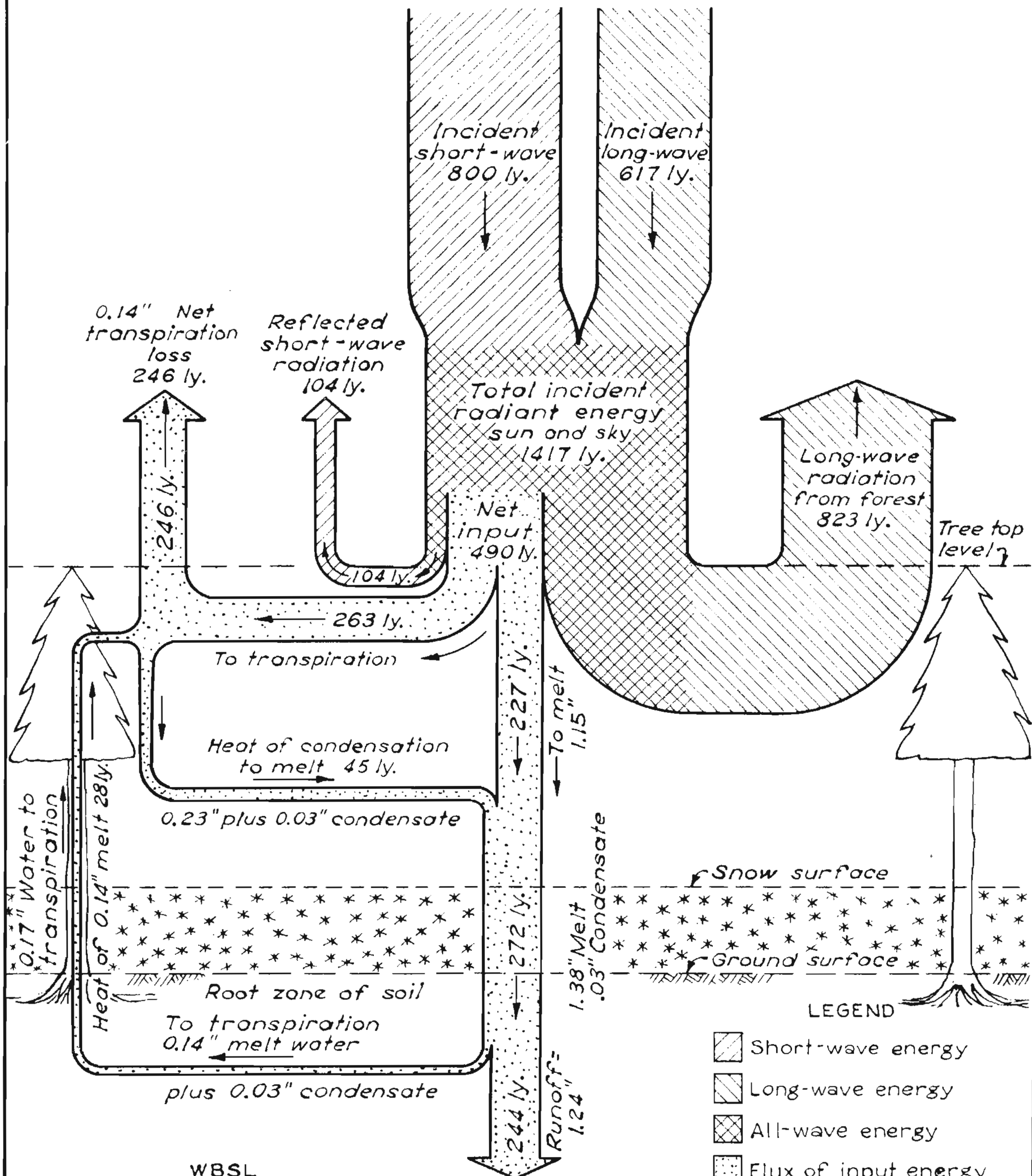

WBSL
9-13 MAY 1949
SUMMARY OF ENERGY — ly/day

		Down	Up	Net
Radiation	Short-wave	800	-104	+696
	Long-wave	617	-823	-206
	All-wave	1417	-927	+490
Transpiration		45†	-291	-246
Runoff				244

† Condensation

SNOW INVESTIGATIONS
SUMMARY REPORT

SNOW HYDROLOGY

DAILY ENERGY BALANCE IN HEAVY FOREST DURING ACTIVE SNOWMELT

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U.S. ARMY

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DRAWN B.M. APPR. D.M.R.

PD-20-25/27

CHAPTER 5 - MELTING OF THE SNOWPACK

5-01. INTRODUCTION

5-01.01 General. - Snowmelt is the over-all result of many different processes of heat transfer. The quantity of snowmelt is, moreover, dependent upon the condition of the snowpack itself. As a consequence, the rigorous determination of snowmelt amounts is quite complex and certain simplifying assumptions are used in the practical computation of snowmelt. The relative importance of the various heat-transfer processes involved in the melting of the snowpack vary with time and with locale. Considering solar radiation for example, in the plains area of the United States it is an important direct cause of snowmelt, while it is relatively unimportant except indirectly, in the heavily forested areas of the Pacific Northwest; it is of considerably less importance during the wintertime than during the spring melt season, and decreases in importance with increasing latitude. As a result of this variation in the relative importance of the several heat transfer processes involved in the melting of the snowpack, no single method or index for computing snowmelt is universally applicable to all areas and at all times of the year. In order to intelligently select the best method of computing snowmelt for a given time and area, a complete understanding of the snowmelt process is necessary. In this chapter the heat-transfer processes involved in the melting of the snowpack are first enumerated; a general snowmelt equation is next formulated which relates the net heat transfer to the resultant snowmelt, taking into consideration the thermal condition of the snowpack. Each heat transfer process is then considered separately in some detail; its variation with meteorological conditions, time of year, and place is discussed. The several processes are then summarized and interactions between them pointed out.

5-01.02 Sources of heat energy. - The principal fluxes of heat energy involved in the melting of the snowpack were enumerated by Wilson in his paper, "An outline on the thermodynamics of snowmelt," 38/ and also in Technical Bulletins 2 and 13. These fluxes, with the symbols used subsequently to identify them are:

Absorbed solar radiation (H_{rs})

Net longwave radiation exchange between the snowpack and its environment (H_{rl})

Convective heat transfer (sensible heat) from the air (H_c)

Latent heat of vaporization released by condensate (H_e)

Conduction of heat from underlying ground (H_g)

Heat content of rain water (H_p)

Each of the above is, itself, a function of several components. For example, absorbed solar radiation is the difference between the solar radiation incident on the snowpack and that reflected by it; net longwave radiation is the difference between the longwave radiation emitted by the snowpack and the portion of it radiated back from its environment (i.e., air, trees, and clouds). In the sections which follow each of the foregoing heat fluxes shall be considered separately.

5-01.03 The energy-budget equation. - If all heat fluxes directed toward the snowpack are considered positive and those away from the pack, negative (the sign being included in the term), and these fluxes are then summed, the total must be zero. Thus, considering snowmelt as but another heat transfer process and including the change in thermal quality of the snow itself (see par. 5-01.05),

$$\Sigma H = H_{rs} + H_{rl} + H_c + H_e + H_g + H_p + H_q + H_m = 0 \qquad (5\text{-}1a)$$

where H_q is the change in the energy content of the snowpack and H_m is the heat equivalent of the snowmelt (that is, the quantity of heat involved in the change of state from ice to water--the latent heat of fusion). All other terms are as previously defined in paragraph 5-01.02. From equation 5-1a it follows that

$$H_m = H_{rs} + H_{rl} + H_c + H_e + H_g + H_p + H_q \qquad (5\text{-}1b)$$

considering the heat equivalent of the melt to be positive. In the above equation, H_{rl} is ordinarily negative in the open, H_e and H_q may be either positive or negative, H_c is generally positive, and other terms in the equation are almost always positive.

5-01.04 Units. - The units used in this chapter are a combination of several systems of units. For example, the English units of degrees Fahrenheit and miles per hour are used to express temperatures and wind speeds, while the metric units of calories will ordinarily be used to express heat quanitities. Snowmelt will ordinarily be given in inches depth while unit area will be taken as one square centimeter. This heterogeneous system of units results from an attempt to express the results of this work in the familiar English units such that available data may be used directly in the resulting equations, at the same time taking advantage of some of the numerical simplicities of the metric system. Moreover, some of the basic data are commonly expressed in metric units; for example, radiation quantities are usually given in gram calories per square centimeter. In spite of the different systems of units used in this chapter, the units used to express given quantities are consistent throughout the chapter. The units will be introduced and defined as needed.

5-01.05 Thermal quality of the snowpack. - If snowmelt is defined as the liquid water which leaves the snowpack, the amount of snowmelt resulting from a given quantity of heat energy is dependent upon the thermal quality of the snowpack. While the latent heat of ice is a well established quantity (80 cal/g or 144 Btu/lb), only rarely is snow encountered which consists of pure ice at 32°F. More often, especially during the winter months, its mean temperature is less than 32°F, and some additional heat is required to first raise its temperature to the melting point before melt can begin. On the other hand, during the melt season the snowpack is not only isothermal at 32°F but also contains some free water. That is to say, instead of pure ice to be melted, there is a mixture of ice and water. When the ice matrix is melted, this free water is also released, resulting in a total quantity of water in excess of that required to melt the ice particles themselves as indicated by the latent heat of fusion for water. The actual condition of the snowpack with regard to the amount of water resulting from a given quantity of heat energy is designated as the "thermal quality" of the snowpack. Thermal quality of snow is defined as the ratio of the heat necessary to produce a given amount of water from snow to the amount of heat required to produce the same quantity of melt from pure ice at 32°F. It is usually expressed as a percentage. It may thus be seen that snow at sub-freezing temperatures will have a thermal quality greater than 100 percent, while snow containing free water will have a thermal quality less than 100 percent. This topic is considered in more detail in chapter 8. Methods whereby the thermal quality of the snowpack can be determined are discussed there and typical values of thermal quality given.

5-01.06 Resultant melt. - Since 80 langleys (calories per square centimeter) of heat energy are required to produce one centimeter of water from pure ice at 32°F, 203.2 langleys (2.54 x 80 ly) are required to produce one inch of runoff from a snowpack having a thermal quality of 100 percent. Thus letting H_m represent the total heat in langleys supplied to the snowpack, and M represent the resultant melt in inches,

$$M = H_m / 203.2 \tag{5-2a}$$

for pure ice at 32°F. Letting B represent the thermal quality of the snow, 203.2 x B langleys are required to produce one inch of melt from any snowpack having a thermal quality B. Hence,

$$M = H_m / 203.2B \tag{5-2b}$$

for any snowpack. Figure 1 of plate 5-1 illustrates this relationship between heat supply and resultant snowmelt for snowpacks of various thermal qualities.

5-01.07 In connection with the foregoing it should be pointed out that the amount of heat required to ripen* the snowpack is relatively small compared to the amount of heat required to melt the snowpack. An example will serve to illustrate this fact. Consider one gram of snow at an initial temperature of -10°C (14°F). Since the specific heat of ice is approximately 0.5, only 5 calories of heat energy are required to bring its temperature to the melting point. Assuming the free-water-holding capacity of a ripe snowpack to be 3 percent (see chap. 8 for a discussion of free water), an additional 2 calories of heat energy are required to produce sufficient melt water to satisfy this free-water-holding capacity. Thus a total of about 7 calories are required to ripen the gram of snow, while about 78 calories are subsequently required to melt one gram of the resulting water-ice mixture.

5-01.08 Nocturnal snow crusts. - During the snowmelt season, it is usual for net flux of heat from all sources to be positive (toward the snowpack) during the day and negative (away from the snowpack) during the night. This diurnal change is especially noticeable in areas of little or no forest cover and during periods of clear weather. It results from the net longwave radiation loss during the nighttime exceeding the gains of heat resulting from convection and condensation. (During the daytime, the solar radiation combined with the greater convective transfer resulting from higher air temperatures usually far exceeds the net longwave loss.) This nocturnal loss of heat energy from the snowpack results in the formation of a crust on the surface of the snowpack. The free water within the pack is refrozen and the snow cooled to some temperature below freezing. The effect is usually confined to the top layers of the pack. In spite of the fact that this deficit may amount to some 80 langleys during a single night, such crusts seldom exceed 10 inches. They are generally much thinner, being around 6 inches in thickness and representing heat deficits of 20 - 40 calories. The nocturnal snow crust represents a heat-energy deficit that must be subtracted from the subsequent day's net gain in determining daily snowmelt amounts. It is sometimes referred to and expressed as a "negative melt" quantity, in which case it is expressed in units of inches of melt rather than langleys.

5-01.09 Data. - Much of what follows is based on snow lysimeter studies of snowmelt made at CSSL (see Res. Notes 17 and 25); however, what is presented herein is generally applicable to any snow-covered area. Reference is made to several publications which contain general data pertinent to the energy-budget approach to snowmelt, and

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*For the purposes of this chapter a ripe snowpack is defined as one which is isothermal at 32°F and has all of its free-water-holding capacity satisfied. (Free water includes only that water permanently held within the snowpack; that is, water held by adsorption and capillarity. It does not include water in the process of percolating through the pack or water impounded in the pack as a result of poor drainage conditions.)

which were consulted in the preparation of this report. Sverdrup's, "The eddy conductivity of the air over a smooth snow field," 34/ gives much valuable data from carefully made meteorological observations over a barren snow field. It also presents the general energy-budget approach to snowmelt along with detailed theoretical background. John Hopkins University, Publications in Climatology, 18/ 36/ likewise present much detailed data on vertical air temperature, humidity, and wind-speed gradients near the ground, and on radiative heat transfer near the ground. These data are generally for snowfree conditions although some data are included for measurements made over snow. These publications also include some of the most recent developments in the theory of heat transfer near the ground. Geological Survey Circular 229, "Water-loss Investigations: Volume 1 - Lake Hefner Studies, Technical Report," 2/ and Sverdrup's Oceanography for Meteorologists 35/ both contain a general reivew of the theory of the energy-budget approach and also include considerable data obtained over water surfaces. Brunt's Physical and Dynamical Meteorology, 6/ Sutton's Micrometeorology, 33/ and Geiger's The Climate Near the Ground 13/ are three excellent texts dealing with the questions pertinent to this report which were referred to frequently in its preparation.

5-02. RADIATION THEORY

5-02.01 Planck's law. - All bodies radiate energy, the intensity of the radiation being a function of the temperature of the radiating body; moreover, the spectral distribution of the radiation is also a function of the temperature of the radiating body. Generally speaking, the higher the temperature, the greater the intensity of the total radiation emitted and the shorter the wave length of the maximum intensity. The spectral distribution of the energy of a radiating black body is given by Planck's Law,

$$E_\lambda = \frac{C_1}{\lambda^5 \left(e^{C_2/\lambda T} - 1 \right)} \qquad (5\text{-}3)$$

where E_λ is the intensity of the emitted radiation of wave length λ, T is the temperature of the radiating body and C_1 and C_2 are constants. Figure 2 of plate 5-1 and figure 1 of plate 5-3 show the theoretical distribution of radiation intensities in the spectrum of a black body in accordance with equation 5-3. Figure 2 is for a body at a temperature of 6000 degrees K (approximate sun temperature) and figure 1 for a body at 273 degrees K (temperature of a melting snowpack). This is the general expression for radiant energy. Other expressions may be derived from it as given in the following two paragraphs.

5-02.02 Wien's law. - For any given temperature, E_λ is zero for $\lambda = 0$ and for $\lambda = \infty$; for some intermediate value of λ, E_λ has its maximum value. This wave length of maximum intensity of radiation can be determined from equation 5-3. Differentiating and equating to zero, gives,

$$\lambda_m T = \text{constant} \tag{5-4}$$

where λ_m is the wave length at which E_λ is a maximum. Equation 5-4 is know as Wien's Law. The value of the constant is usually taken as 2940 for wave lengths expressed in microns (equals 10^{-6}m) and temperatures in degrees K. Thus the wave lengths of maximum intensity are 0.49μ and 10.8μ for the temperatures of 6000 and 273 degrees K, respectively.

5-02.03 Stefan's law. - The total energy emitted in all wave lengths (per unit time and area) by a black body may be determined by integrating equation 5-3. Thus the total radiation in all wave lengths, E, from a black body at a temperature T is,

$$E = \sigma T^4 \tag{5-5}$$

where,

$$\sigma = \int_0^\infty \frac{C_1}{(\lambda T)^5 \left(e^{C_2/\lambda T} - 1\right)} d(\lambda T)$$

Equation 5-5 is known as Stefan's law and σ as Stefan's constant. The value of σ is 0.826×10^{-10} (ly/min) / $(^{\circ}K)^4$. For a radiating body other than a black body, its radiation relative to the radiation of a black body is expressed by a ratio known as its emissivity. Figure 2 of plate 5-3 is a graphical presentation of Stefan's law, and it also shows radiation intensities corresponding to the various temperatures for emissivities less than unity (black-body emissivity).

5-02-04 Solar and terrestrial radiation. - Only a very small portion of the entire electromagnetic spectrum (which ranges from cosmic rays and the emissions of radioactive substances with wave lengths of the order $10^{-6}\mu$ to low-frequency radio waves having wave lengths of the order 10^4 m) is involved in the radiation melt of the snowpack: the radiation between about 0.15μ to 80μ. This radiation is divided into two general categories: solar and terrestrial. Solar (or shortwave) radiation is included in the range from about 0.15 to 4μ which encompasses the visible spectrum (0.4 to 0.7μ). It has its maximum intensity in the visible spectrum at about 0.5μ. It also extends into the ultraviolet and the infrared. Figure 2, plate 5-1 shows the theoretical distribution of intensity of solar radiation at the different wave lengths, and also shows the limits of the visible spectrum. It may be seen that roughly half of the solar radiation lies within the range 0.4 to 0.7μ, that is, half the total solar radiation is in the form of visible radiation or light. Terrestrial (or longwave) radiation is generally included in the range 3μ to 80μ: it has its maximum intensity in the infrared at around 11μ. Figure 1, plate 5-3 shows the spectral distribution of radiation intensity from a black body at 273°K (32°F), as given by Planck's law (equation 5-3). From Stefan's law, the total energy radiated per unit time and area is found to be 0.459 ly/min or 27.5 ly/hour. In the following two sections, both of the foregoing types of radiation involved in the melting of the snowpack will be considered separately; their variations with cloud and tree cover will be discussed and methods presented by which they may be estimated in the absence of measurements.

5-03. SOLAR RADIATION

5-03.01 The solar constant. - Of the tremendous quantity of radiant energy emitted by the sun, only an infinitesimally small portion is intercepted by the earth and its atmosphere. Yet this small portion is the ultimate source of all the earth's energy. The amount of solar energy intercepted by the earth varies slightly with seasons due to the varying distance between the earth and sun, and is thought to have small day-to-day variations due to changes in the solar output; however, these variations are quite small. The intensity of incident radiation on the earth is given by the solar constant which is defined thus: the intensity of solar radiation received on a unit area of a plane normal to the incident radiation at the outer limit of the earth's atmosphere with the earth at its mean distance from the sun. The value of the solar constant is generally taken to be 1.94 langleys per minute which is based on the 1913 Smithsonian standard scale. 29/ Until recently, it was thought this value was too high; a solar constant of 1.90 ly/min being the best available figure. 11/ More recently there has been evidence that these values are too low, a value of 2.00 ly/min being offered. 24/ These differences arise from the fact that solar radiation, measured at or near the earth's surface, is less than that incident at the outer limit of the earth's atmosphere even during clear weather. Estimates must be made of the portion absorbed and reflected by the atmosphere to arrive at the solar constant. This is particularly true in the ultraviolet where the radiation is largely absorbed. Recently observations of the solar spectrum have been made at high altitudes from rockets, and estimates of the solar constant have been made using these data. 24/ However, even those observations leave portions of the spectrum to be estimated. In this report the solar constant will be taken as 1.94 ly/min.

5-03.02 Insolation. - Of direct concern to the study of snowmelt is the amount of solar radiation incident on a horizontal surface. This is termed insolation. The daily amount of insolation received at the outer limit of the earth's atmosphere (or at the earth's surface in the absence of an atmosphere) may be calculated from the solar constant for any given latitude and time of year by taking into consideration: (1) the distance between the earth and sun, (2) the angle of incidence of the sun's rays, and (3) the duration of sunlight. Figure 3, plate 5-1 shows daily insolation amounts as a function of latitude and time of year, between latitudes 20° and 70° N.

5-03.03 Transparency of atmosphere. - The portion of the insolation given by figure 3 of plate 5-1 which actually reaches the earth's surface depends upon the transparency of the atmosphere and the optical airmass through which it must pass.* Some of the incident solar radiation is reflected, some scattered, and some absorbed by the atmosphere. In the absence of clouds these amounts are relatively small and quite constant barring unusual atmospheric conditions such as dust storms. The variations that occur are chiefly a result of variations in the amount of water vapor and dust in the air. The effects of water vapor and dust on the absorption and scattering of solar radiation has been extensively investigated; however, a detailed consideration of these effects is beyond the scope of this report. Reference is made to Technical Bulletin 5 and Research Note 3 and an article by Klein 27/ for summaries of work done on this subject. For purposes of this report it is sufficient to point out that the average daily insolation received at the earth's surface with clear skies may be determined for any given locality by plotting daily totals of measured insolation (see par. 5-03.09) as a function of time of year, and drawing a curve which envelopes these values, being guided by the curve giving the values of daily insolation received at the outer limits of the earth's atmosphere for the latitude of the particular site. This is done for CSSL (lat. 39° 22' N) as shown by figure 4 of plate 5-1. It is to be pointed out that, generally speaking, the higher the elevation of the station the greater is the atmospheric transmission, other things being equal. Also, the less the zenith distance, the greater is the atmospheric transmission. These follow from the fact that the smaller the optical airmass, the greater is the transmission for any given condition of the atmosphere. (Atmospheric transmission coefficient is defined as the ratio of the insolation received at the earth's surface with a cloudless sky, I_c, to the insolation received at the outer limit of the earth's atmosphere, $\overline{I_o}$. Actually, for the same airmass, the transmission is somewhat greater in the winter than in the summer because of the usually clearer air that prevails during the winter. It may be noted in figure 4 of plate 5-1 that the atmospheric transmission is greatest in the summer and least in the winter in consequence of the differences in optical airmasses. The atmospheric transmission coefficient varies from about 80 percent at time of the winter solstice to about 85 percent at the time of the summer solstice.

5-03.04 Atmospheric transmission coefficients are based on the total insolation received at the earth's surface and, as such, include both the direct solar beam and the diffuse sky radiation (scattered light reaching the earth's surface). They include an amount of diffuse sky radiation based on normal conditions. One factor that has a pronounced effect on the amount of diffuse sky radiation reaching the earth's surface is the albedo or reflectivity of the surface itself. Since a

*Optical airmass is very closely given by the secant of the zenith (distance) angle for sea-level locations. For elevated stations, the secant must be multiplied by the ratio of the average station pressure to standard sea level (p/p_o).

portion of the reflected bea is also scattered back to the earth's surface, the greater the reflectivity of the surface, the greater is the diffuse sky radiation, other things being equal. Since for ordinary conditions of bare ground, the albedo is quite small and constant, this is a minor effect. However, over snow surfaces, the generally high albedo and its large range (40 to 80 percent) makes this effect of considerable importance. Thus the amount of insolation received at the earth's surface increases with the albedo of the snowpack, other things being equal. Figure 4 of plate 5-1 includes the average result of this effect at CSSL. During the winter, the higher average albedoes resulting from new-fallen snow tend to increase the amount of insolation reaching the earth's surface. During the spring, the lower albedoes of the older snow decrease the amount of diffuse sky radiation, while during the summer months, the bare ground makes this effect practically nil. Thus the decrease in the atmospheric transmission coefficient during the winter months due to the greater optical airmass is somewhat offset, for snow covered areas, by the increase in diffuse sky radiation (in addition to the clearer air of winter which also tends to increase the atmospheric transmission coefficient.)

5-03.05 Effect of clouds. - By far the largest variations in the portion of solar radiation transmitted by the atmosphere are caused by clouds. Yet this variation is also one of the most difficult to evaluate. The transmitted radiation varies with type, height, density, and amount of clouds. Several investigations have been made which relate the ratio of the insolation actually received at the earth's surface (I) to the average insolation received at the earth's surface with cloudless skies (I_c), to the amount of cloud cover (N). Thus,

$$I/I_c = 1 - k'N \tag{5-6a}$$

For N expressed in tenths of sky cover an average value for k' of 0.71 is given by Sverdrup (p. 51). 35/ However, there must necessarily be a considerable variation in this value from place to place, season to season and for different types and densities of clouds. It may be seen that for an overcast sky ($N = 1$), this value of k' makes, $I = 0.29I_c$. Recently a value of 0.54 has been suggested by Newmann 32/ which results in $I = 0.46I_c$ for an overcast sky. Haurwitz 21/ gives the I/I_c ratio in the form,

$$I/I_c = 1 - (1 - k)\ N \tag{5-6b}$$

where k is equal to $1 - k'$ and has a certain physical significance: it is the ratio of the insolation received with overcast skies to the insolation received with cloudless skies. This is the more usual form of the equation used to relate the ratio I/I_c to the amount of cloud cover, N. Values of k have been determined by Haurwitz 21/ for different types of clouds. Figure 5 of plate 5-1 shows variations in the value of k with cloud type and height (after Haurwitz) and also the resultant variation in the percentage of insolation transmitted with various cloud heights and amounts using the values of k when substituted in equation 5-6b.

5-03.06 Another approach which has been used to estimate the depletion of insolation by clouds is to relate the ratio I/I_c to the percentage of possible sunshine (S), as determined by a sunshine recorder, by an equation of the form,

$$I/I_c = a + b\ S \tag{5-7a}$$

The coefficients a and b may be evaluated statistically. Fritz and MacDonald 12/ determined values of 0.35 and 0.61 using monthly data for the United States. A more usual form of the equation used to relate the ratio to sunshine amounts is,

$$I/I_c = k'' + (1-k'')\ S \tag{5-7b}$$

Here k" is the value of the ratio (fraction of cloudless sky insolation) on a day with zero recorded sunshine. Values of k" in equation 5-7b for the United States are given by Kimball. 26/ The most generally used value of k" is 0.22; however, it too is subject to variation with all of the aforementioned factors plus possible differences in the settings of the sunshine recorders. A general summary of this approach to the estimation of insolation--i.e. from sunshine data--is to be found in a paper by Hamon and others. 19/ This paper gives a graphical means of computing insolation from percent possible sunshine, with the additional parameters of latitude and time of year. (See pl. 6-1 for a reproduction of this graph.) The linear relationship between the insolation ratio, I/I_c, and the percent sunshine, S, implicit in equations 5-7 is not used in this paper; rather an empirical curvilinear relationship is derived. Equations 5-6 and 5-7 are for total insolation received at the earth's surface; hence they include diffuse sky radiation. As was previously mentioned, the quantity of diffuse sky radiation relative to the direct solar beam, is affected by the albedo of the surface. This effect is even more pronounced with cloudy skies than it is with clear skies; not only is the ratio of diffuse sky radiation to direct radiation increased by the presence of clouds, but the light reflected from the earth's surface is strongly re-reflected by the clouds. Thus, over snow-covered areas, greater values of the ratio I/I_c are to be expected than are found over non-snow-covered areas for a given type and amount of cloud cover. The actual values of the constants are also dependent, for snow-covered areas, upon the albedo of the pack and forest cover, in addition to all of the aforementioned variables. In view of these complexities, the previously given values of the constants should be used as general guides only in the determination of insolation amounts. Moreover, they apply, strictly, only to long-term averages of data and may be considerably in error for a given day. Only by actual measurement at the site can the true amount of daily insolation be accurately determined.

5-03.07 Effect of slope. - It is obvious that outside of the tropics (northern hemisphere), the radiation incident on south-facing slopes exceeds that on north-facing slopes. For moderate slopes during the springtime, as a result of the high solar altitude, the effect of slope is slight. During the winter the effect is more pronounced. At any given instant, the radiation on a sloping surface relative to the radiation received on a horizontal surface may be determined from the geometry of the situation (the slope and its aspect in conjunction with the solar altitude and azimuth). However, if daily totals are to be determined, the problem is more complex. Such a determination involves the integration of the solar path relative to the sloping surface. It, of course, varies with the time of year as a result of the changing solar path and is different for every slope and slope aspect. In addition, there is the variability in the transmission of radiation with solar altitude in consequence of the differences in optical airmass through which radiation must pass. Then too since, diffuse sky radiation is the same for all slopes and aspects, this constant factor must be included in all computations. All these effects have been included in some analyses made by Hoeck 22/ of daily totals of radiation on slopes of 25 degrees for a latitude of 46° 30' N. and an elevation of 1577 meters. Data are given for both south- and north-facing slopes (it is shown that radiation on east- and west-facing slopes is the same as that on a horizontal surface). These data are given in figure 6 of plate 1. This figure also presents values of total solar radiation (direct plus diffuse sky radiation). The above cited paper by Hoeck is an excellent and detailed treatise on the role of radiation (both longwave and short-wave) in the melting of the snowpack. Reference is also made to Geiger 13/ (p.224) and a paper by Thelkeld and Jordan 37/ for more detailed information on evaluation of incident radiation on sloping surfaces.

5-03.08 Effect of forest cover. - As is the case with cloud cover, the determination of the effect of forest cover on the amount of insolation reaching the ground is somewhat inexact. The transmission percentage varies with the density, type and condition of the trees. Deciduous trees, of course, show marked variations in the transmission ratio with the season; the analysis of transmission variation for this type of forest is most difficult. Fortunately, however, the principal type of forest in the snow-covered areas of the western United States is coniferous. Figure 1 of plate 5-2 gives an average curve for the transmission percentage of insolation through coniferous forest canopies of various densities. It is presented only as a general guide; actual quantities for a particular area may vary considerably from the curve. For the data in the figure, the canopy cover is defined as the horizontal projection of tree crown area. This relationship is based on snow laboratory data, and includes cloudy- as well as clear-weather data. There is some variation in the ratio with the relative amounts of direct and diffuse sky radiation, hence the ratio would also vary with degree of cloudiness and solar altitude. This subject is considered in some detail by Miller 31/ who examined the data from the snow laboratories along with other pertinent data. It will not be dwelt on further here. Research Notes 5, 9, and 12 also deal with this question in more detail.

5-03.09 Measurement. - The preceding paragraphs describe the causes of variations in insolation incident on the snow surface and give relationships by which insolation amounts may be estimated in the absence of actual measurements. They were derived from actual measurements at pyrheliometer sites for application to areas where no such instruments are located. Of course, the most accurate and sometimes the only practical method of determination of insolation is to be had from actual measurements. In this country, measurements of insolation are made almost exclusively by means of Eppley pyrheliometers. 20/ This instrument consists of an evacuated bulb in the center of which is a disk having a white center and concentric bands of black and white. A thermopile, having its alternate junctions in the black and in the white, produces an emf in proportion to the temperature difference between the rings, and hence to the radiation incident upon them. This emf may then be recorded on a recording potentiometer, suitably calibrated to give radiation intensity. Because of the glass envelope, longwave radiation is excluded (as is also a portion of the solar radiation, both due to reflection at the air-glass interface, and to absorption of certain wave lengths by the glass). Errors result from the degree of heating of the bulb in different ambient temperatures; however, they are quite small and are generally ignored. There are currently, in the United States, some seventy pyrheliometer stations at which insolation is continuously recorded. These data are published monthly in the U. S. Weather Bureau's National Summary of Climatological Data. (Prior to January, 1950 they were published in the Monthly Weather Review.) Continuous measurements of insolation were made at UCSL and CSSL throughout the period of operation of these laboratories (see chap. 2). These data are published in the Hydrometeorological Logs for these laboratories.

5-03.10 Albedo of the snowpack. - The albedo, or reflectivity, of the snowpack varies over a considerable range: new-fallen snow may reflect 80 percent or more of the incident insolation, while a ripe, granular snowpack may reflect as little as 40 percent. Consequently, the albedo of the snow has an important role in the melting of the snowpack. The albedo is primarily a function of the condition of the surface layers of the snowpack. Before taking up methods by which the albedo of the snow surface may be estimated, it is well first to consider briefly the manner in which albedo of the snow is commonly measured and also to consider the manner in which this reflection takes place and its effects upon the measurement of reflected radiation. Albedo of the snowpack is commonly measured (in this country) by means of two Eppley pyrheliometers as was done at the snow laboratories. The two pyrheliometer bulbs are first compared in direct sunlight to assure identical calibration. One of these is mounted in the normal position (see par. 5-03.09) and measures the insolation received. The other is inverted and measures the shortwave radiation reflected by the snowpack. Albedo of the snow is generally considered to be the inverse ratio of these two quantities. Two considerations that are of importance to the measurement of the albedo of snow are: (1) the diffuse and (2) the spectral character of the albedo. If the albedo of the snow varies with the angle of incidence of the radiation, then, even for identical snow conditions, different albedoes will result from annual and diurnal changes in the sun's altitude. If the albedo of the snow is different for different wave lengths of solar radiation, then the measurements of reflected and incident radiation intensities may not be directly comparable. Each of these two considerations will be examined separately in the following paragraphs.

5-03.11 Snow is generally considered to be a good diffuse reflector; that is to say, the intensity of reflected light is independent of the angle of the incident beam. The intensity of reflected light should thus conform to the cosine of the angle of the reflected beam. While this is nearly true for small angles of incidence, the cosine law does not hold strictly for large angles of incidence, higher albedoes being associated with larger angles of incidence. This variation may also be partially the result of changes in the structure of the snow itself. In the morning and in the evening a crust may occur on the snow surface. During midday, when the melt rate is at a maximum, the higher concentration of liquid water in the top layers of the snowpack undoubtedly decreases the albedo. However, even an immature, non-melting snowpack exhibits the same characteristic, although, it is generally believed to be to a lesser degree. This effect has also been investigated by Hubley. 23/ In practice, this diurnal variation in albedo is obviated by use of a mean daily albedo. This specular quality of the snow also has a seasonal effect on the albedo of the snowpack as a result of the annual change in the sun's altitude. Thus slightly higher albedoes would be expected to result during the winter than during the spring for the same snow conditions in consequence of the smaller solar altitudes.

5-03.12 As previously mentioned, the albedo of the snow is usually determined from the readings of a pair of Eppley pyrheliometers, one in the normal position measuring incident radiation on a horizontal surface and the other inverted to measure reflected radiation. Since Eppley bulbs are calibrated in direct sunlight and since varying amounts of incident radiation are reflected and absorbed by the glass envelope of the bulb, depending upon the wave length of the radiation, the resulting calibration is for the spectral distribution of the radiant energy of the incident solar radiation only. Should the spectral distribution of the reflected rays differ significantly from the incident, the calibration would no longer strictly hold. Spectral measurements of the albedo of snow have shown it to be quite constant through the visible range, as is obvious from the dazzling whiteness of the snow surface. Progressing into the near infrared the reflecting power decreases rapidly. This is illustrated in figure 2 of plate 5-2 which shows the spectral reflectivity for a ripe, melting snowpack. (Data for this figure are from SIPRE Report 4. 30/) However, since the transmission of the glass envelope of the Eppley bulb and the intensity of the radiation itself both decrease with increasing wave lengths in the near infrared, the longer wave lengths of solar radiation have relatively little effect upon the calibration of the pyrheliometer and consequently, the lesser albedoes in this portion of the solar spectrum are quite unimportant.

5-03.13 Continuous measurements of incident and reflected radiation over a snowpack were made at CSSL during the period of snow cover for the years 1946 through 1954. These measurements were made using the two-Eppley pyrheliometer method previously described. Several studies have been made, using these data, to determine the variation of albedo with time, and with accumulated heat supply (as determined by radiation and temperature indexes). (Reference is made to Tech. Bull. 6 and to Res. Note 1.) The results of these studies are summarized in figures 3 and 4 of plate 5-2 which show the variation of albedo with time and with accumulated heat supply index. Different curves are given for different times of the year in figure 4. It may be noted that during the melt season, the high albedoes of new-fallen snow quickly decrease to the albedo of the older snow. There is a lower limit of about 40 percent. These curves are for an uncontaminated snow surface; snow naturally contaminated by forest litter, dust, etc., or artifically contaminated would, of course, exhibit lower albedoes. The curves are general ones derived from the study of many separate occurrences. Individual situations may vary significantly from the average values represented by the curves. For example, when an old, melting snowpack having an albedo of, say, 50 percent is covered by a light snow fall, its albedo may increase to 80 percent and then return to 50 percent in a day or two when the thin cover of newly-fallen snow is melted. The curves presented, however, are good general guides. They may be entered at any point corresponding to the current albedo of the snow, and estimates of the future albedo of the snowpack made therefrom. A more complete discussion of the albedo of the snowpack can be found in the previously referenced report by Miller.31/

5-03.14 Absorption of radiation by the snowpack. - The difference between the solar radiation incident on the snow surface and that reflected by it is the solar radiation absorbed by the snowpack. This absorption occurs not only at the surface as with more opaque materials, but, because of the translucent nature of snow, extends to some depth within the pack. For deep, ripe snowpacks, this penetration is of little practical concern; the heat absorbed by the snow would result in the same quantity of melt were it all absorbed in the top surface or should it penetrate to a depth of a foot or so. However, for shallow snowpacks the penetration of radiation may result in a measurable quantity being transmitted by the snowpack to the underlying surface. As a result of the usual low albedo of this surface (rock or soil), most of this transmitted radiation is absorbed. Most of the heat energy may be returned to the snowpack by conduction and/or longwave radiation, producing about the same melt as would have occurred in a deeper snowpack. On the other hand, in the case of frozen ground, some of this heat may be absorbed by the ground without any melt resulting at the ground level.

5-03.15 For snowpacks having a homogeneous structure, the penetration of solar radiation into the snowpack may be expressed by a logarithmic law,

$$I_d = I_a e^{-kd} \qquad (5\text{-}8)$$

where I_a is the intensity of radiation transmitted through the snow surface, I_d is the intensity at depth d beneath the surface, and k is the extinction* coefficient for snow. In experiments made of the absorption of solar radiation by snow, the value of k has been found to vary, its value being chiefly dependent upon the density of the snowpack; the higher the density the greater is the penetration. For depths expressed in centimeters, in equation 5-8, the following values of k have been experimentally determined (Tech. Rpt. 8, Int. Rpt. 1):

Snow Density (percent)	Extinction Coefficient (k)
26.1	0.280
32.2	0.184
39.7 } 44.8 }	0.106

Figure 5 of plate 5-2 illustrates the penetration of radiation in the snowpack for different densities of snow using the coefficients given above. From this figure it may be seen that, for ripe, high density snow, about 4 percent of the radiation absorbed by the snow penetrates as far as one foot; for lesser densities of snow the radiation penetration is less. A further discussion of this topic may be found in Technical Report No. 8 (Interim Report No. 1) and in a paper by Gerdel.14/

5-03.16 The albedo of snow is mainly determined by the character of the snow surface; sub-surface conditions have little effect on the albedo. When an old surface of low-albedo snow is buried by an appreciable layer of new-fallen snow, the snowpack exhibits the albedo of the new surface. When the old surface is reexposed by the melting of the new layers, its albedo becomes as it was previously. Similarly the albedo of shallow snow (as low as 6 inches) is thought to be little affected by the type and condition of ground beneath the snow. Measurements of albedo, however, usually show a marked decline in albedo as the snow cover becomes thin. This results mostly from the patchiness of the shallow snow, the reflected radiation being the result of reflection both from bare ground and snow. Under these conditions, melt is usually accelerated because of the greater portion of solar radiation absorbed by the combined ground and snow surfaces, even though little if any more radiation is directly absorbed by the snow itself. The warming of the ground by solar radiation accelerates melt both by conduction of heat from the ground, and indirectly, by warming the air passing over it. This results in a rapid edge melting of snow patches.

*The term "extinction coefficient" is used rather than "absorption coefficient" inasmuch as some of the decrease in the intensity of the incident beam with depth results from internal reflections; not all the radiation which penetrated the snow surface is absorbed.

5-04. TERRESTRIAL RADIATION

5-04.01 Radiation emitted by the snowpack. - Snow is very nearly a perfect black body with respect to longwave radiation, 10/ absorbing all such radiation incident upon it and emitting the maximum possible radiation in accordance with Stefan's law (equation 5-5). While this may seem somewhat strange, considering the high albedo of snow with respect to shortwave radiation, particularly in the visible spectrum, a consideration of the character of the snow surface indicates the reason for this phenomenon. Since the snow surface is composed of small grains of ice having many facets, when considered microscopically, it is extremely rough. The multi-faceted crystals and the interstices between the crystals quite effectively trap incoming longwave radiation, and, conversely, are an efficient emitting surface. Since snow radiates as a black body in accordance with Stefan's law, the longwave radiation emitted by the snowpack may be readily calculated. The intensity of black-body radiation for different temperatures is given in figure 2, plate 5-3. Since the temperature of snow is limited to a maximum of 32°F, the maximum intensity of radiation that may be emitted by it is 0.459 ly/min (equals 27.5 ly/hr). Radiation intensities corresponding to emissivities other than unity are also given by the figure (these will be discussed subsequently).

5-04.02 Back radiation to the snowpack. - Back radiation to the snowpack is the integrated result of radiation from: (1) the earth's atmosphere, (2) clouds, and (3) forest cover. Each of these radiative fluxes will be considered separately in the paragraphs which follow; the combined effect will then be discussed. Considering first the back radiation to the snowpack in an unforested area under conditions of clear skies, it is pointed out that the earth's atmosphere, unlike the snowpack, is not a black or even a gray body*. Rather it absorbs and emits radiation to varying degrees dependent upon the wave length. At certain wave lengths the air absorbs and radiates almost as a black body; at others it is practically transparent to radiation. Only two of the gases of the atmosphere, carbon dioxide and water vapor, have any appreciable effect on absorption in the longwave portion of the spectrum. Since the proportion of carbon dioxide in the atmosphere is practically constant, its effect in absorbing and radiating longwave radiation may be considered as fixed. The over-all effect of carbon dioxide is also less important to radiation exchange, both by virture of its lesser quantity and also by virtue of its lesser absorption bands. Carbon dioxide has a narrow, intense absorption band centered at 14.7μ and extending from 12μ to 16.3μ. The amount of water vapor in the atmosphere, however, exhibits wide variations. As a result it is the controlling variable in the amount of back radiation from the atmosphere with clear skies. The absorption

*A gray body is one which at a given temperature, emits a fixed proportion of the black body radiation at that temperature <u>in all wave lengths</u>.

spectrum for water vapor is quite complex and extensive. For certain wave lengths in the longwave spectrum almost no absorption takes place, while other wave lengths are almost totally absorbed by the existing water vapor in the atmosphere. Other wave lengths are absorbed to varying degrees. Thus there are "windows" through which some of longwave radiation emitted by the pack (and other terrestrial surfaces) may escape to space.

5-04.03 The back radiation to the snow surface from the earth's atmosphere is the result of radiation from all levels of the atmosphere. It is dependent upon the moisture content and temperature distribution of the entire atmosphere. A method has been advanced whereby the temperature and moisture distribution throughout the troposphere may be utilized in determining downward longwave radiation at the earth's surface;9/ however, the method is quite complex and requires that upper air sounding be made. Since the layers of the atmosphere nearest the earth's surface ordinarily have the greatest moisture content and the highest temperatures, they have the greatest influence on the downward longwave radiation. The temperature and moisture content of the upper atmosphere has comparatively little variation, and as a result, its contribution to downward longwave radiation is, like that of carbon dioxide, fairly constant. As a consequence, several investigators have found that estimates of downward longwave radiation from the earth's atmosphere can be had from surface air temperature and vapor pressure alone. For example, Brunt6/ has proposed an equation wherein the ratio of the back radiation from the earth's atmosphere (R_d), to the theoretical black-body radiation computed using surface air temperature (σT_a^4) was correlated with the square root of the surface vapor pressure (e_a). that is,

$$R_d/\sigma T_a^4 = a + b\sqrt{e_a} \qquad (5\text{-}9)$$

Other investigators have determined values of a and b in Brunt's equation for a variety of locations, some of which are given in the table below (as listed by Goss and Brooks17/):

Investigator	Place	a	b	Corr. Coeff.	Range of e_a (mb)
Goss & Brooks	California	0.66	0.039	0.89	4 - 22
Angstrom	California	0.50	0.032	0.30	------
Ramanathan & Desai	India	0.47	0.061	0.92	8 - 18
Eckel	Austria	0.47	0.063	0.89	------
Dines	England	0.52	0.065	0.97	7 - 14
Asklof	Sweden	0.43	0.082	0.83	2 - 8
Angstrom	Algeria	0.48	0.058	0.73	5 - 15
Boutaric	France	0.60	0.042	----	3 - 11
Anderson	Oklahoma	0.68	0.036	0.92	3 - 30

Another form of empirical equation, advanced by Angstrom, 4/ relates the ratio to the vapor pressure of the air, as follows:

$$R_d/\sigma T_a^4 = a - be^{-k\,e_a} \qquad (5\text{-}10)$$

where values of e is the base for Naperian logarithms and e_a is the vapor pressure in millibars. Values of the constants a, b, and k, are given in the following table (from Anderson 2/):

Investigator	Place	a	b	k
Angstrom	Sweden	0.806	0.236	0.115
Kimball	Virginia	0.80	0.326	0.154
Eckel	Austria	0.71	0.24	0.163
Raman	India	0.79	0.273	0.112
Anderson	Oklahoma	1.107	0.405	0.022

A straight line relationship between the ratio and the vapor pressure of the air has also been derived from measurements made at Lake Hefner.2/ Thus,

$$R_d/\sigma T_a^4 = a + b\,e_a \qquad (5\text{-}11)$$

For e_a expressed in millibars in the above equation, values of 0.740 and 0.0049 are given for a and b. The differences that exist between the foregoing equations are small, especially when one considers the large scatter that exists in the actual observations. Figure 3 of plate 5-3 shows the variation of the ratio, $R_d/\sigma T_a^4$, with vapor pressure, e_a, as given by equations 5-9, 5-10, and 5-11, using values of the coefficients and exponents given for the Lake Hefner study (Anderson, in the above tables). The values of the ratio, as determined by the Lake Hefner coefficients, are generally greater than those found by the other investigators yet they agree most closely with the constant ratio found for the CSSL, within its limited range (see following paragraph). It will be noted that all three of the equations give quite similar results.

5-04.04 Over extensive snowfields, wide variations of vapor pressure of the air are not ordinarily encountered. The vapor pressure has a strong tendency to remain close to that of the snow surface since the snowpack is both a sink and a source for vapor pressures greater or less than that of the snow. For air over a melting snowpack, the tendency is thus toward a vapor pressure of 6.11 millibars (the saturated vapor pressure at 32°F); the range of observed vapor pressures is usually between 3.0 and 9.0 millibars. Measurements of back radiation made over snow at CSSL, with vapor pressures within this range, indicate the ratio, $R_d/\sigma T_a^4$, to be quite constant and independent of the vapor pressure of the air for this limited range. The value of this constant ratio of

0.757 is also shown in figure 3 of plate 5-3 for the limited range of vapor pressures for which it holds. It will be noted that within this range, the values given by equations 5-9, 5-10, and 5-11 also show little change. The use of a constant ratio is tantamount to making the coefficient b in equations 5-9, 5-10, and 5-11 equal to zero, the constant ratio then being the value of the constant, a.

5-04.05 Net radiation with clear skies. - Using the constant ratio of the preceding paragraph for $R_d/\sigma T_a^4$ (= 0.757), the net longwave radiation exchange over a melting snow surface (R_u = 0.459 ly/min) for clear-weather conditions is given in figure 4, plate 5-3 as a function of air temperature. (Values are also given for ratios equal to 0.80, 0.85 and unity.) From this figure it may be seen that with clear skies the air temperature must exceed 69°F in order for a net gain of longwave radiation by the snowpack to result. This is, of course, strictly true only for conditions of a melting snowpack and for vapor pressures near the saturated vapor pressure for melting snow (6.11 mb). Yet it is generally true for clear weather radiation exchange over the snowpack during the snowmelt season since these conditions usually prevail. Snow surface temperatures are usually not too different from 32°F, and the differences that do occur result in a relatively small change in the emitted longwave radiation. For example, a snow surface temperature of 20°F results in an emitted radiation of 24.9 ly/hr in contrast to the 27.5 ly/hr emitted by the melting snowpack (see fig. 2 plate 5-3); this is 90.5 percent of the black-body radiation at 32°F. Also, over a melting snow surface, the vapor pressure of the air usually remains fairly close to that of the snow surface (i.e., 6.11 mb) as was previously discussed.

5-04.06 Radiation from clouds. - So far the discussion has been restricted to longwave radiation exchange between the snowpack and the atmosphere during clear weather only. In the presence of clouds the foregoing relationships do not hold, since clouds have a dominant effect on longwave radiation. The absorption spectrum for liquid water is quite similar in pattern to that for water vapor; however, the magnitude of the absorption is much greater for liquid water. For example at a wave length of 10 microns, which is in the middle of the transparent band for water vapor and also in the portion of the spectrum where the absorption is least for liquid water, 0.1 mm of liquid water transmitts only 1 percent of the incident radiation. Since even relatively thin clouds contain more precipitable water than this, all clouds are considered to be black bodies with respect to longwave radiation. Under overcast conditions, the net longwave radiation exchange between the snowpack and the atmosphere may be considered to be the net longwave radiation exchange between two black bodies having temperatures corresponding to the snow surface temperature and the cloud base temperature. That is,

$$R = \sigma(T_c^4 - T_s^4) \qquad (5\text{-}12)$$

where R is the net longwave radiation and T_c denotes cloud base temperature. The net radiation exchange for overcast conditions over a melting snow surface is given by the curve labeled "(black-body)" in figure 4 of plate 5-3. It may thus be seen that in situations where the cloud base temperature is greater than the snow surface temperature, there will be no loss of heat energy from the snowpack by longwave radiation, but rather a net gain will result.

5-04.07 Net longwave radiation loss from the snowpack under conditions of partial cloud cover may be estimated as follows. Angstrom 4/ found that since the net longwave radiation loss from the snowpack is inversely proportional to the amount of cloud cover, the net longwave radiation loss with cloudy skies (*R*) may be roughly approximated by an equation of the form

$$R = R_c\ (1 - kN) \qquad (5\text{-}13)$$

where R_c is the net longwave radiation loss with clear skies and *N* is the portion of sky covered by clouds. An average value of *k* of 0.9 has been suggested by Angstrom; however, the value of the coefficient, *k*, has been shown to vary with type and height of the clouds among other things. Meinander (as quoted by Geiger 13/) found the following *k* values for the different cloud types:

Cloud type	k
Low thick clouds (Ac, Sc, Ns, St)	0.76
High thinner clouds (Ac, As, Cs)	0.52
Thin cirrus veils	0.26

Phillips (as quoted by Geiger 13/) found the following values of *k* as a function of cloud height:

Ceiling		k
km	1000 ft	
1.5	4.92	0.87
2	6.56	0.83
3	9.84	0.74
5	16.40	0.62
8	26.24	0.45

From this it would seem the average *k* of 0.9 must thus be best applied to low-level clouds. From the above table, the approximate relationship, $k = 1 - 0.024z$, where *z* is the height of the cloud base in thousands of feet, may be derived (see fig. 5, plate 5-3). Substituting this relationship in equation 5-13, gives the equation,

$$R = R_c \left[1-(1-0.024z)N\right] \qquad (5\text{-}14)$$

which is illustrated in figure 5 of plate 5-3. Using Meinander's k values, the relationship is illustrated in figure 3 of plate 5-6. Also shown in this figure is an average relationship between the degree of cloudiness and the longwave loss ratio as found by Lauscher (as quoted by Hoeck 22/). This latter relationship reflects average conditions wherein the tendency is for lower, thicker clouds to be associated with higher degrees of cloudiness. Obviously these relationships are only approximations of net longwave radiation under cloudy conditions. The value of k in equation 5-13 is dependent upon other factors such as thickness and density, as well as cloud type and height. Moreover, the value of the ratio, R/R_c varies with the distribution of the partial cloud covers, N. The assumed linear relationship with each of these variables (z and N) is not absolutely valid. Then too, the equation represents only average conditions. A normal decrease in cloud base temperature with elevation is implied; however, in any given instance, the actual temperature may vary from the normal. For these reasons, the relationship of figure 5, plate 5-3 (equation 5-14) is of little practical value in estimating back radiation from partially cloudy skies. It is given, rather, to illustrate the relative effects of clouds on the back radiation from the sky. Because of the many complexities involved in its computation, it is necessary that back radiation from the sky be measured if accurate results are to be obtained (see par. 5-04.11).

5-04.08 To summarize the effect of cloud height and amount in the net exchange of longwave radiation over the snowpack, and give a general expression which also holds approximately for clear skies, the relationships of the foregoing paragraphs may be combined into one general equation. Since R_c may be approximated by the equation,

$R_c = 0.757\sigma T_a^4 - \sigma T_s^4$ it follows that,

$$R = \left[0.757\sigma T_a^4 - \sigma T_s^4\right]\left[1-(1-0.024z)N\right] \qquad (5\text{-}15)$$

Equation 5-15 is a general expression for the net longwave radiation exchange over a melting snowpack for an unforested area. The sign convention followed here is as elsewhere in this chapter: Fluxes of heat energy directed toward the snowpack are considered positive and those directed away, negative. Again the reader is cautioned to remember this relationship is only an illustrative one; practical determination of longwave radiation under conditions of partial cloud cover must be based upon actual measurements. Quite often several layers of clouds are involved, in varying amounts, such that computation of their net effect is practically impossible.

5-04.09 Radiation from forest canopy. - Somewhat analogous to the radiation from clouds is the radiation from the forest canopy. A solid canopy also approximates a black body in the longwave portion of the spectrum, absorbing and emitting all possible radiation. The effective leaf temperature can conveniently be considered to be the same as the ambient air temperature.* Thus the net longwave radiation heat exchange between a solid forest canopy and the snowpack (R_f) may be expressed as

$$R_f = \sigma(T_a^4 - T_s^4) \tag{5-16}$$

since both the snowpack and the tree canopy are, effectively, black bodies, and the effective temperature of the tree leaves is taken as air temperature. For forest conditions other than 100 percent cover, the situation is more complex. Assuming a melting snowpack (σT_s^4 is constant), and a constant value for the ratio $R_d/\sigma T_a^4$, it is possible to illustrate the effect of a varying forest cover in longwave radiation exchange. Since net longwave radiation exchange for the forested areas is given by equation 5-16, and radiation in the open may be expressed as $R_c = (0.757\sigma T_a^4 - \sigma T_s^4)$ these two terms can be weighted in accordance with the amount of forest cover and combined to arrive at the over-all net exchange in the forest. Letting F represent the degree of forest cover (solid canopy equals unity),

$$\begin{aligned} R &= FR_f + (1-F)\ R_c \\ &= \sigma T_a^4 \left[F + (1-F)\ 0.757\right] - \sigma T_s^4 \end{aligned} \tag{5-17}$$

This variation in R with forest cover is illustrated in figure 6 of plate 5-3 for net exchange over a melting snow surface. The foregoing (eq. 5-17) is, of course, an oversimplification of the problem; ramifications are considered in Research Note 12 and in a paper by Miller.31/

5-04.10 The relationship given by figure 6 of plate 5-3 (eq. 5-17) is, as is the case with the relationship for cloud cover, only a relative one used to illustrate the role of forest cover in the flux of longwave radiation. The type of trees, degree of maturity, spacing, etc., all affect the relationship. In addition, the sun's altitude, the wind speed, and degree of cloud cover, affect the approximation of using air temperature as the effective canopy temperature. Still another important source of variation in the above equation is the method by which the degree of forest cover is estimated. This is a quite subjective quantity, and estimates for the same site may vary considerably

*Actually the leaves heated by direct solar radiation may be somewhat higher than the ambient air temperature; they transfer heat to the air by longwave radiation and by convection to the air passing over them. However, the leaves which face the snowpack and hence have the dominant role in radiative heat transfer between the trees and the snowpack, are generally shaded by the tree crown and hence are very nearly at air temperature. There is a radiative heat transfer that occurs within the foliage itself.

with the method used and even for a given method, with the observer. In view of the complexities, the accurate determination of instantaneous or daily amounts of net longwave radiation exchange is dependent upon measurements. This will be discussed in the paragraph which follows.

5-04.11 Measurement. - In this country most current measurements of longwave radiation are made using Gier-Dunkle Radiometers. These instruments are non-selective absorbers of radiation, that is, they are sensitive to both shortwave and longwave radiation. Two varieties are made: one, a total hemispherical radiometer, measures the total hemispherical irradiation upon a plane surface; the other, a net-exchange radiometer, measures the net radiative heat transfer across the plane of the meter surface. Reference is made to papers by Gier and Dunkle 16/ and by Dunkle and others 8/ for a description of the construction and operation of these instruments. Basically they consist of a 4-inch-square flat plate which serves as a heat-flow meter. Both surfaces of the plate are blackened to absorb, non-selectively, a high percentage of the radiation incident upon them. A silver-constantan thermopile inside the plate has its alternate junctions in the upper and lower surfaces and thus produces an emf portional to the temperature difference between the surfaces and hence to the difference in irradiation falling upon the surfaces. The flat plate is mounted in an air blast from a blower which keeps the convective losses from the two surfaces approximately equal. Thus the net radiometer measures the difference in radiation falling on the two surfaces. The total hemispherical radiometer has one of its plates shielded by a highly reflecting plate which is also located in the air blast and thus remains at about the same temperature as the shielded surface. Since these meters measure both shortwave and longwave radiation, they are ideal for the measurement of total radiative heat transfer between the snowpack and its environment. Thus the net-exchange radiometer integrates into a single measurement all radiative fluxes to and from the snowpack. For analytical purposes, however, it is sometimes desirable to determine the longwave component separately. One method of doing this is to make the measurements at night when shortwave radiation is non-existent. This method has commonly been employed in the past, which accounts for the fact that longwave radiation is often referred to as "nocturnal radiation." Nighttime measurements, however, usually restrict the range of air temperatures to the lower values; during the daytime, air temperatures, and hence the back radiation, are usually higher than at night. In order to determine the back radiation during daylight hours it is necessary that measurements be made of the incident shortwave radiation (see para. 5-03.09) simultaneously with the allwave radiation measurements; this former quantity can then be subtracted from the latter to arrive at the longwave component. This method, however, gives rise to considerable error. Since the longwave component is but a small difference between two much larger quantities, any small errors in either of them is magnified many-fold in the final result. Because of this source of error, other investigations, for example the Lake Hefner study, 2/ have led to the conclusion that the measurement of longwave radiation during the day is impractical.

Nevertheless, the nocturnal measurements still offer a reliable means of determining longwave radiation, and the radiometers are well suited to the determination of the total radiative heat transfer, day or night. Measurements of longwave radiation exchange by Gier-Dunkle radiometers were made at CSSL during several years of its operation. Several studies have been made of the data collected. Reference is made to Research Notes 6, 7, and 11 and Technical Bulletin 12. Another study of net longwave radiation exchange is given in Technical Bulletin 7.

5-05. RADIATION SUMMARY

5-05.01 General. - So far the shortwave and longwave components or the total radiative flux have been considered separately. The effects of atmospheric water vapor, clouds, forest cover, etc., on each were individually discussed. In this section the effects of those factors on the combined allwave radiative flux will be considered. Illustrative examples are given for conditions as they exist at CSSL (approx. 40° N and 7200 ft msl) and for the north slopes of the Alps (approx. 46° N and 5200 ft msl). For the Alps, the variation of radiation throughout the year is shown. For CSSL example two situations will be examined: one for the winter and the other for the spring melt season. The winter situation is included to illustrate the importance of the seasonal change in daily insolation amounts in the melting of the snowpack and the dominant effect of radiative heat transfer in controlling snowmelt. February 15 was selected to typify winter conditions and May 20 to typify spring melt conditions. On those dates the daily amounts of insolation received with clear skies at CSSL are 400 langleys and 800 langleys, respectively. This marked change in insolation coupled with the usual change in albedo of the snow and the air temperatures from winter to spring results in a great increase in radiation melt from winter to spring as will be shown. In the discussions which follow, it is assumed that 200 langleys result in one inch of snowmelt (thermal quality of 98.4 percent). During the winter the thermal quality would usually be greater than this, resulting in even less melt than indicated, while during the spring lesser thermal qualities and greater melts would be the rule. Reference is made to SIPRE Research Paper 8 15/ for a nomograph which expresses total radiation heat supply (both shortwave and longwave) to the snowpack. Parameters of time of year, latitude, sky condition, and albedo of the snow are involved.

5-05.02 Clear-weather melt. - During clear weather, the important variables in radiation melt are: (1) the insolation, (2) the albedo of the snow, and (3) the temperature of the air. Humidity of the air also affects the radiation melt; however, its effect is relatively minor. Figures 1(a) and 1(b) of plate 5-4 are illustrative of daily clear-weather radiation melt amounts for the winter and spring conditions. Back radiation to the snowpack is estimated using the constant ratio, 0.757, for $R_d/\sigma T_a^4$, which is generally applicable for vapor pressures in the vicinity of 6 millibars. In computing longwave radiation emitted by the snowpack, the snowpack was assumed to remain at 32°F. These conditions hold quite well for the spring melt situation, but during the winter lower vapor pressures and snow surface temperatures would be expected. Hence, both the back radiation to the snowpack and the longwave radiation emitted by the snowpack would be less than assumed. Consequently, these assumptions tend to cancel one another, making the net longwave radiation loss from the snowpack during the winter reasonably correct. In figure 1(b), "negative melts" are shown as dashed lines, since the values given in this portion of the figure are dependent upon the snow surface being at a temperature of 32°F, and the indicated loss of heat would indicate this not to be the situation. This section of the figure is strictly applicable only if convection-condensation amounts of heat transfer are sufficient to make up the indicated deficit.

5-05.03 Typical albedoes of the snow during the spring may be taken as 50 percent, and a typical mean daily temperature of 50°F. During the winter, a higher albedo of, say 75 percent, and a mean daily temperature of 30°F are more representative of actual conditions. The points fulfilling these conditions are indicated on the figures. Thus, for clear weather, a typical springtime daily radiation melt of about 1.6 inches is indicated, while during the winter a heat deficit is indicated. (This heat deficit may be partially or wholly made up by convection-condensation.) These conditions of albedo and temperature for winter and spring are assumed in the following paragraphs where the effects of clouds and trees on radiative melt are considered.

5-05.04 Effect of clouds. - The dominant role of clouds in the fluxes of both shortwave and longwave radiation over snow has already been discussed. Since clouds are such a powerful controlling factor in radiative heat exchange, other minor factors such as humidity of the air are often ignored in the estimation of radiative heat exchange on cloudy days. Figures 2(a) and 2(b) of plate 5-4 illustrate the effect of clouds on daily radiation melt during the spring and winter. These figures are simply a combination of shortwave and longwave radiation exchange corresponding to given cloud heights and amounts given by figure 5 of plate 5-1 and by figure 5 of plate 5-3; the amounts of shortwave and longwave radiation with clear skies are as previously mentioned. It will be noted that during the winter the effect of clouds on radiative heat exchange is relatively less than during the spring, due to the lesser radiation melts during the winter time. Also, during the winter, radiation melt tends to increase with increasing cloud cover and lower cloud heights in consequence of the more important role played by longwave radiation during this time of the year.

5-05.05 Effect of forest canopy. - Similar to the effect of clouds, the forest canopy exerts a powerful controlling influence on net allwave radiation exchange between the snowpack and its environment. However, its effect is different from that of clouds, particularly with respect to shortwave radiation. While both the clouds and trees restrict the transmission of insolation, clouds are highly reflective, while the forect canopy absorbs much of the insolation. As a result, the forest canopy tends to be warmed and in turn gives up a portion of the incident energy to the snow. (It radiates, in the longwave portion of the spectrum, directly to the snowpack and also warms the air by convection and radiation which in turn gives up some of its heat to the snowpack.) Clouds, on the other hand, reflect back to space a large portion of the incident radiation, which is thus lost. The role of the forest canopy in radiative heat exchange between the snowpack and its environment is illustrated in figures 3(a) and 3(b) of plate 5-4. These figures represent the typical spring and winter snowmelt conditions in the middle latitudes, as previously specified, under a coniferous forest cover. It shows the variation of the several radiative components and the net allwave radiation with degree of forest cover during clear weather. The curves of these figures are based on the transmission coefficients for insolation given by figure 1 of plate 5-2 and on the net longwave radiation exchange in the forest given by figure 6 of plate 5-3. It may be seen that for the conditions specified, during the spring the maximum radiation melt occurs in the open and the minimum with a canopy density of about 50 percent. During the winter the maximum melt (for the conditions specified) is with 100 percent canopy cover and the minimum with about 20 percent cover. These curves, of course, are merely for average conditions of canopy cover and do not reflect variations in the spacing of trees or take into account clearings in the forest canopy. These effects are discussed in section 5-12 of this chapter.

5-05.06 Effect of slopes. - The effect of slope on allwave radiation exchange over a barren snowfield can be illustrated by an analysis made by Hoeck 22/ for the Alps. This analysis carries on the investigation of the same 25-degree slope gradients previously considered in paragraph 5-03.07 for shortwave radiation alone. It applies to a latitude of 46° 30' N and to an elevation of about 5200 feet msl. Conditions of air temperature and humidity assumed are mean values for the north slopes of the Alps. The effects of variable cloudiness and albedo, as well as the effect of slope, are included in the relationships. These relationships are illustrated in figure 1, of plate 5-6. Figure 1(a) gives the relationship for horizontal surfaces and for east- and west-facing slopes which are the same as a horizontal surface. Figure 1(b) shows the relationship for a south-facing slope of 25-degree gradient and figure 1(c) for a north-facing slope of 25-degree gradient. The two dates used in the foregoing examples for CSSL (plate 5-4) are indicated on these figures for comparison.

5-06. THEORY OF TURBULENT EXCHANGE

5-06.01 General. - Of secondary importance to radiation in the transmission of heat to the snowpack is the process of turbulent exchange in the overlying air. With a downward temperature gradient there is a direct transfer of heat from the air to the snow, and with a downward vapor pressure gradient there is a direct transfer of moisture from the air onto the snow surface, releasing, in addition, its latent heat of vaporization. The reverse processes occur as well; it is the net effect that is of concern. During periods of melting, the temperature of the snow surface remains at 32°F and the corresponding vapor pressure is 6.11 mb. If the air temperature and the vapor pressure immediately above the surface (within the laminar layer--a fraction of an inch in thickness) were known, the flow of heat and moisture to the snow and the resulting melt could be easily computed, using the thermal conduction equation. Such refined measurements, however, are unobtainable with available instruments, and the practical alternative has been to establish the relationship of observed melts to air temperatures and vapor pressures measured at some higher level. These relationships have been determined experimentally for measurements at several sites and for the particular measurement conditions prevailing at the time of the experiment. In order to generalize the equations thus derived to make them hydrologically applicable to other sites and other conditions of measurement, it is of value to consider what is known of the processes of turbulent exchange, since the usual heights of measurement of air temperature and air moisture content are within the region of turbulence, where the vertical distribution of temperature, water vapor and wind speed is governed by the action of eddies.

5-06.02 Basic equation. - The basic equation for turbulent exchange is,

$$Q = A\, dq/dz \qquad (5\text{-}18a)$$

where Q is the flow of some property of the air through a unit horizontal area per unit time, dq/dz is the vertical gradient of this property (q = the property, such as temperature or water vapor; z = height), and A is the exchange or "Austausch" coefficient which will be discussed later. The property q may be any property which is itself unaffected by vertical transport. Since the interest here is in the exchange of heat and moisture to the snow, only the properties of air temperature* and moisture content (often expressed in various ways, as specific humidity, vapor pressure, etc.) and wind speed shall be considered.

5-06.03 Derivation of practical equation. - To put the basic equation in a form usable for hydrologists, it is necessary to know the variation of air temperature, humidity and wind speed with height in the zone of turbulent mixing. The matter of vertical gradients under a variety of conditions of stability of the air has been extensively investigated (Sverdrup 34/, Johns Hopkins Publications in Climatology 36/) and the profiles of these properties during conditions of atmospheric stability, such as prevail over snowfields, have been shown to follow a

*Strictly speaking, potential temperatures should be used, as the equation is applicable only to properties of the air which do not change with vertical motion. If, however, temperatures are measured within ten feet of the snowpack, as they usually are, the error in using measured air temperatures is negligible for practical purposes, amounting to less than 0.05°F for the movement of the air between the level of measurement and the snow surface.

power law distribution. (A logarithmic profile has been found to more adequately represent the distribution under neutral or unstable conditions, but it is disregarded in this presentation as being uncharacteristic of conditions over snowfields.) According to a power law, the ratio of values of some property of the air, $\underline{q}$, at two heights above the snow surface is equal to some power of the ratio of the heights, $\underline{z}$, themselves. It is expressed by the relationship

$$q_2 / q_1 = (z_2 / z_1)^{1/n} \tag{5-19a}$$

If $\underline{z_1}$ is unity, then

$$q = q_1 z^{1/n} \tag{5-19b}$$

where $\underline{q}$ is the value of the property at level $\underline{z}$ and $\underline{q_1}$ the value at unity level. (The values of the property $\underline{q}$ in the above equations represent the differences between the values measured at their respective heights and the values of the properties at the snow surface. For wind speed and for temperatures measured in degrees C, the snow surface value is ordinarily zero, and the measured values of these elements may be used directly in the equations. For temperatures measured in degrees F and for vapor pressures, the snow surface values must be subtracted from the measured values to make the relationship valid.) Differentiating equation 5-19b,

$$dq/dz = (q_1 / n) z^{(1-n)/n} \tag{5-20}$$

and substituting in equation 5-18a,

$$Q = A (q_1/n) z^{(1-n)/n} \tag{5-18b}$$

which expresses the eddy exchange of the property at any level $\underline{z}$ from measurements of the property made at unity level. The exchange coefficient, $\underline{A}$, also applies to the level $\underline{z}$. The value of the exchange coefficient varies with height as is discussed in the following paragraph.

5-06.04 Under equilibrium conditions, the gradients of temperature and moisture assumed by the air above the snow are such that the eddy transfer of heat and moisture are constant with height up to normal heights of measurement. Consequently, in equation 5-18a, the exchange coefficient must vary inversely with the gradient, dq/dz. That is,

$$A / A_1 = (dq/dz)_1 / (dq/dz) \tag{5-21a}$$

and, from equation 5-20,

$$A = A_1 z^{(n-1)/n} \tag{5-21b}$$

Where A_1 is the exchange coefficient for the 1-foot level. Substituting this value of $\underline{A}$ in equation 5-18b,

$$Q = A_1 \, q_1 \, / \, n \qquad (5\text{-}18c)$$

Sverdrup 34/ has shown that the exchange coefficient at a given level is directly proportional to the wind speed at that level. Thus,

$$A_1 = k \, v_1 \qquad (5\text{-}22)$$

where v is the wind speed and the subscripts indicate unity level; k is a proportionality constant. Substituting this value of A_1 in equation 5-18c,

$$Q = (k/n) \; q_1 \; v_1 \qquad (5\text{-}18d)$$

Using the power law (equation 5-19b) to express the variation of the property q and the wind speed v with height, equation 5-18d may be given as,

$$Q = (k/n) \quad (z_a z_b)^{-1/n} \; q_a v_b \qquad (5\text{-}18e)$$

where the subscripts a and b are used to identify the levels of measurement of the property and the wind speed respectively, since the two may be different.

5-06.05 Condensation melt. - Considering now the specific case of the exchange of moisture by eddy diffusion. Since the exchange coefficient expresses the mass (of air) exchanged per unit time and area, it is necessary to know how much of the property being considered --here water vapor--is contained in unit mass of air. Thus specific humidity (mass of water vapor per unit mass of air) must be used to express the humidity gradient. Since the specific humidity is approximately given by the expression, $(0.622/p)e$, where p is the atmospheric pressure and e is the vapor pressure of the air, values of vapor pressure may be substituted for specific humidity. Thus the moisture transfer, Q_e, is given by the equation,

$$Q_e = (k/n) \; (z_a z_b)^{-1/n} (0.622/p) \; e_a v_b \qquad (5\text{-}23a)$$

Equation 5-23a thus gives the moisture transfer and hence the amount of condensate given up by the air to the snow surface. (In the above equation, e_a represents the difference in vapor pressure between the air at level z_a and the vapor pressure of the snow surface. If the vapor pressure of the air is greater than that of the snow surface, e_a is positive and condensation results. If the vapor pressure of the air is less than that of the snow surface, e_a is negative and evaporation occurs. In what follows, where condensation is discussed, evaporation is considered simply as negative condensation.) In addition to the condensate itself, for every gram of water condensed on the snow surface, approximately 600 calories of heat energy are released (latent heat of vaporization). This is sufficient to melt 7.5 grams of snow having a thermal quality of 100 percent (ratio latent heat of vaporization to latent heat of fusion of water equals 600/80). Adding this melt to the condensate gives a total of 8.5 grams of melt-plus-condensate. Equation 5-23a thus gives,

$$M_e = 8.5\ (k/n)\ (z_a z_b)^{-1/n}(0.622/p)\ e_a v_b \qquad (5\text{-}23b)$$

where $\underline{M_e}$ represents the total melt plus condensate.

5-06.06 Convection melt. - In the case of heat transfer by eddy exchange, $\underline{H_c}$, the measured property of the air used in equation 5-18e becomes air temperature; the specific heat of the air, $\underline{c_p}$, must also be included in the equation to properly convert the temperature measurements into heat units. That is,

$$H_c = (k/n)\ (z_a z_b)^{-1/n}\ c_p\ T_a\ v_b \qquad (5\text{-}24a)$$

In cgs units, the equivalent snowmelt is equal to the net heat exchange divided by 80 (for pure ice at 32°F).

$$M_c = (1/80)\ (k/n)\ (z_a z_b)^{-1/n}\ c_p\ T_a\ v_b \qquad (5\text{-}24b)$$

where $\underline{M_c}$ is the resultant melt in grams.

5-06.07 Elevation effect. - The coefficient $\underline{k}$ in the foregoing equations is a complex function itself. Among other things, its value is dependent upon the density of the air. Since density of the air varies with elevation, so does the value of $\underline{k}$, and hence it is not a constant for all locations. In order to make this coefficient independent of air density, and hence elevation, another term giving the variation of density with elevation, may be included in the equation, the sea-level density of the air being implicitly included in the value of the coefficient. Thus the term, p/p_o, where $\underline{p}$ is the atmospheric pressure at the elevation of the site and p_o is sea-level pressure, may be included in the equation to represent the variation in density of the air with elevation. Including this term in equations 5-23b and 5-24b,

$$M_e = 8.5\ (k'/n)(z_a z_b)^{-1/n}(0.622/p_o)\ e_a v_b \qquad (5\text{-}23c)$$

$$M_c = 1/80\ (k'/n)\ (z_a z_b)^{-1/n}(p/p_o)\ c_p\ T_a\ v_b \qquad (5\text{-}24c)$$

where $\underline{k'}$ is now a constant applicable to sea level pressure ($p = p_o$).

5-06.08 Combined equation. - Since many of the terms of equations 5-23c and 5-24c are common to both, the two equations may conveniently be combined into a single expression of convection-condensation melt, $\underline{M_{ce}}$. Thus

$$M_{ce} = \frac{k'}{n} (z_a z_b)^{-1/n} \left(\frac{1}{80} \frac{p}{p_o} c_p T_a + 8.5 \frac{0.622}{p_o} e_a \right) v_b \qquad (5\text{-}25)$$

This is the theoretical expression for the snowmelt resulting from eddy transfer of heat and moisture to the snowpack. It is based on the concept that the exchange coefficients for both the exchange of heat and moisture are the same. In the sections which follow, the values of the coefficients for convection and condensation melt will be arrived at separately by experimental means.

5-07. CONDENSATION AND EVAPORATION

5-07.01 General. - Numerous investigations have been made of actual amounts of water vapor condensed upon or evaporated from exposed water surfaces. Most studies have been primarily interested in water losses due to evaporation. 2/ 35/ In snow hydrology, in addition to the amounts of water evaporated or condensed, the latent heat of vaporization involved in the change of state from gas to liquid or vice versa is also of concern. In the case of water surfaces, such heat may be absorbed or given up without immediate hydrologic effect; on snow surfaces this heat is of considerable hydrologic significance, for it is capable of producing a more than sevenfold increase (or decrease in the case of evaporation) in the amount of water available for runoff over that actually condensed (or evaporated). In what follows, the discussion shall be concerned with condensation, it being understood that evaporation is merely the negative case. The experimental methods are similar in either case. The condensation or evaporation amounts are determined volumetrically or by weighings. Simultaneously, measurements of vapor pressure and wind speeds are made. The equation which is used to relate vapor pressure and wind speed to the amounts of condensate is of the form,

$$q_e = k_e (e_a - e_s) v_b \qquad (5\text{-}26)$$

where q_e is amount of condensate, e_a and e_s are the vapor pressure of the air and snow surface respectively, v_b is the wind speed, and k_e is a coefficient relating the two.

5-07.02 Condensation over snow. - Detailed experiments of condensation and evaporation over snow have been made by several investigations. Among these are studies made at CSSL (see Res. Note 25) and by deQuervain at the Weissfluhjoch Institute in the Swiss Alps. 7/ The method used has been to place pans filled with snow into the snowpack so that the surface of the snow in the pans is flush with the surrounding snow surface, and to note the moisture gain or loss in the pans by periodic weighings. It is to be remembered, however, that where snow is involved, for every unit of water vapor condensed, additional heat of vaporation is released, capable of melting 7.5 times this amount of snow. Thus to represent the condensate plus its accompanying melt, assuming this latent heat is completely effective, the constant k_e above must be multiplied by 8.5 (1 + 7.5), and the condensation melt equation for any particular site becomes

$$M_e = 8.5\, k_e\ (e_a - e_s)\, v_b \tag{5-27a}$$

5-07.3 Generalization of equation. - Since the vapor pressure of the air and the wind speed vary with height above the snow surface, the value of the coefficient $\underline{k_e}$ is dependent upon the height of measurement of these meteorological elements. By assuming the power law variation of vapor pressure and wind speed with height, discussed previously (par. 5-06.03), it is possible to generalize equation 5-27a so that the coefficient $\underline{k_e}$ is constant regardless of heights of measurement and is applicable to any site over open snow. The value of the exponent $\underline{n}$ in the power law (equations 5-19) has been the subject of considerable investigation. A good summary is found in Sutton's Micrometeorology 33/ where a spread of values is reported on from a variety of conditions of stability of air. Over snow, however, an inversion generally exists and values of $\underline{n}$ are fairly constant. Power law exponents derived by Walsh* as a best fit to wind speed gradients at CSSL range from 4.8 to 7.1, with an average value of 5.8. Sverdrup found by observations on Isachsen's Plateau 34/, that an $\underline{n}$ of 5.6 applied to the vertical distributions of wind speed, air temperature and vapor pressure. Recent experimentation by the Snow Investigations tends to confirm a value of $\underline{n} = 6$ as an adequate representation of the vertical distribution of these meteorological parameters above a melting snow field in an open, unforested site. Thus

$$e_a - e_s = (e_1 - e_s)\, z_a^{1/6}$$

and

$$v_b = v_1 z_b^{1/6}$$

where the subscript $\underline{1}$ refers to the various properties measured at a standard reference level of one foot (or other unit). Substituting these relationships in equation 5-27a,

$$M_e = 8.5 k_e\, (z_a z_b)^{1/6}\, (e_1 - e_s)\, v_1 \tag{5-27b}$$

Letting $k_e' = 8.5\, k_e\, (z_a z_b)^{1/6}$, where $\underline{k_e'}$ is now a constant (the value of the coefficient when wind speed and vapor pressure are measured one foot above the snow surface),

$$M_e = k_e'\ (e_1 - e_s)\, v_1 \tag{5-27c}$$

Moreover, since

$$e_1 - e_s = (e_a - e_s)\, z_a^{-1/6}$$

* See Miscellaneous Report 6.

and

$$v_1 = v_b z_b^{-1/6}$$

vapor pressure and wind speed measured at any levels (z_a and z_b) can be reduced to their equivalent values at the one-foot level, and

$$M_e = k_e' \ (z_a z_b)^{-1/6} \ (e_a - e_s) \ v_b \qquad (5\text{-}27d)$$

where k_e' is a constant relating condensation melt (plus condensate) to the condensation parameter for values of wind speed and vapor pressure measured at the one-foot level.

5-07.04 Evaluation of constant. - The value of k_e in equation 5-26 has been evaluated by relating measurements of moisture exchange, q_e, to simultaneous products of vapor-pressure gradient and the first power of the wind speed, $(e_a - e_s) \ v_b$, by a linear least-squares regression. From this, the coefficient k_e' of equation (5-27d) was then evaluated by substituting actual heights of measurements of wind speed and vapor pressure. Values of the constant, k_e' so determined (for daily melts in inches, for vapor pressure measured in millibars and wind speed in miles per hour, heights of measurement in feet) are as follows:

Central Sierra Snow Laboratory (Res. Note 25)	0.0540
Weissfluhjoch Institute (deQuervain)	0.0770

Figure 2, plate 5-5 illustrates this relationship, giving daily melts in terms of mean daily wind speed and vapor pressure at one-foot level, using the CSSL coefficient. (Melts for other heights of measurement may be determined from figure 2 by use of figure 5, plate 5-5.)

5-08. CONVECTIVE HEAT TRANSFER FROM THE AIR

5-08.01 Unlike radiative heat transfer and the transfer of moisture between the air and snow surface, convective heat transfer from the air to the snow surface cannot be measured and has to be evaluated indirectly. This has been done by considering convection melt as a residual in the general snowmelt equation. Since the total melt from experimental areas is measurable, and the melts from the processes of radiation and condensation can be computed as previously explained, the difference between the total melt and these two computed melts is considered due to convective heat transfer. By relating quantities so determined to the meteorological parameters of air temperature and wind speed by the equation,

$$M_c = k_c \ (T_a - T_s) \ v_b \qquad (5\text{-}28a)$$

where $\underline{M_c}$ is the convective melt quantity, $\underline{T_a}$ and $\underline{T_s}$ are the temperatures of the air and snow surface, and v_b the wind speed, the convective melt coefficient, k_c, can be determined. This was done at CSSL using data from a special snow lysimeter constructed expressly for this purpose (see Res. Note 25). The convective melt coefficient obtained reflects the conditions peculiar to that site and the particular instrumentation involved. By a process similar to that described in paragraph 5-07.03, the foregoing equation can be generalized for application to any heights of measurement and to any site over open snow. Thus

$$M_c = k_c' \ (p/p_o)(z_a z_b)^{-1/6} \ (T_a - T_s) \ v_b \qquad (5\text{-}28b)$$

where $k_c' \ (p/p_o)(z_a z_b)^{-1/6}$ equals $\underline{k_c}$ in equation 5-28a. The term $\underline{p/p_o}$ (where $\underline{p}$ is atmospheric pressure at the observation level, and $\underline{p_o}$ is sea level pressure) is introduced to correct for the variation of air density with elevation, since the coefficient $\underline{k_c}$ is directly proportional to the air density. Values of $\underline{p/p_o}$ as a function of elevation are obtainable from figure 6 of plate 5-5. Values of $z^{-1/6}$ for various heights of measurement are obtainable from figure 5 of plate 5-5. The value of the convection coefficient, k_c', of equation 5-28b was determined for an open site from the CSSL experiments to be 0.00629 for daily (24 hour) melt rates expressed in inches, temperatures measured in degrees F, wind speeds in miles per hour, and heights of measurement in feet. Figure 1, plate 5-5 shows this relationship for measurements of air temperature and wind speed at one-foot level. (Melts for other heights of measurement may be determined from figure 1 by use of figure 5, plate 5-5.)

5-09. SUMMARY OF CONVECTION-CONDENSATION MELTS

5-09.01 Comparison with other investigations. - In present use by snow hydrologists in the computation of condensation or convection melts are the results of experiments and the melt equations presented by other investigators, notably Sverdrup 34/ and deQuervain 7/. Since the form of their equations is similar to those given herein, the coefficients for convection and condensation melt may be compared as follows:

Investigator	Condensation coefficient (k_e')	Convection coefficient (k_c')	Ratio: k_c'/k_e'
Snow Investigations (1954)	0.0540	0.00629	0.12
Sverdrup (1936)	0.0674	0.0215	0.32
deQuervain (1951)	0.0770		

Light, 28/ who used a theoretical equation advanced by Sverdrup, arrived at coefficients essentially the same as those of Sverdrup's given above.

All coefficients above are for wind speeds, air temperatures, and vapor pressures measured at the one-foot level; they are for temperatures measured in degrees F, vapor pressures measured in millibars, and wind speeds in miles per hour; they are for melts expressed in inches per day (24 hours).

5-09.02 As may be seen, the condensation coefficients of the other investigators given above are of the same order of magnitude as those of Snow Investigations. The convection coefficients, however, are more than three times as great as those determined by the Snow Investigations. Sverdrup's coefficients were derived from experiments over snow on Isachsen's Plateau, in which ablation measurements furnished total melt data, from which was subtracted computed radiation melt, leaving a residual melt assigned to combined condensation and convection.* Sverdrup, being unable, in the absence of measured condensation amounts, to separate condensation-convection melt except on a theoretical basis, assumed that the exchange coefficients of moisture and heat are the same (as described in section 5-06). This assumption is becoming less satisfactory with accumulating experimental evidence. 3/ 18/ A further check on the condensation constant comes from deQuervain's investigations at Weissfluhjoch Institute by experimental methods similar to those of Snow Investigations. The ratios, k_c'/k_e', given above afford a means of comparing the relative magnitudes of the convection and condensation melts as found by the several investigations. The value of this ratio, as given by Bowen 5/, ranged from 0.32 to 0.37, the smaller value of the ratio being associated with stable atmospheric conditions (such as are found over snow fields). Thus Bowen's ratio agrees with that of Sverdrup. In summary, theory accords reasonably with experiment in considerations of moisture exchange. For convective heat exchange there is notable discrepancy. The smaller exchange coefficient indicated by Snow Investigations is recommended in view of the more detailed measurements of meteorological parameters (including radiation) available at present.

5-09.03 Combined equation. - Many of the terms of equations 5-27d and 5-28b are common to both, and the two equations may conveniently be combined into one equation of convection-condensation melt. Using the coefficients derived at the CSSL, this equation is as follows:

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*Light's equation is theoretical, based on one advanced by Sverdrup which assumes a logarithmic variation of air properties with height, in contrast to the power law variation cited previously and also used in Sverdrup's experimental work. Light's equation, in application, employs a basin factor of less than unity, by which the theoretical equation must be multiplied to reproduce measured basinwide melt amounts. (Radiation melts are not considered). It would be reasonable to expect this factor to be greater than unity, since it must include the not inconsiderable melt from radiation. Its empirical value of approximately 0.65 would seem to indicate that the coefficients of the theoretical equation are too large.

$$M_{ce} = (z_a z_b)^{-1/6} \left[0.00629\ (T_a - T_s)(p/p_o) + 0.0540\ (e_a - e_s)\right] v_b \quad (5\text{-}29a)$$

or

$$M_{ce} = 0.00629\ (z_a z_b)^{-1/6} \left[(T_a - T_s)(p/p_o) + 8.59\ (e_a - e_s)\right] v_b \quad (5\text{-}29b)$$

The above equation is for daily (24-Hour) melt in inches where the temperatures are in degrees F, vapor pressures in mb, and wind speed in mph and height in feet. They are for melt from ripe snow packs and apply strictly only to unforested sites. This combined relationship is given graphically in figures 3 and 4 of plate 5-5. Figure 4 expresses the total melt in terms of temperature and relative humidity.

5-09.04 Discussion. - The foregoing coefficients for convection and condensation melt were derived from measurements made at a point in the open. To apply these results to other areas, the effects of terrain on air temperature, vapor pressure and wind speed must be considered. In forested areas wind speeds are less than in the open; hence if convection-condensation melt in the forest is to be determined, based on measurements of wind speed made in an open area, coefficients smaller than those given herein would certainly be expected. Moreover, the power law variation of wind speed with height applies to open areas only. In the forest no such simple relationship of wind speed with height applies. Since each forested area presents a different problem due to differences in type and spacing of trees, topography, etc., no general relationships can be given. Areal variations in air temperature and vapor pressure are less than those of wind speed but are also of some consequence. In the computation of basinwide melts due to convection and condensation, the variation of both air temperature and vapor pressure with elevation is an important consideration. It is to be pointed out that, unlike solar radiation, the temperature and the vapor pressure of the air are not independent of the resultant melt they produce. For a given airmass, an increase in wind speed tends to produce more convection-condensation melt but at the same time tends to reduce the temperature and vapor pressure of the air near the snow surface. Thus there is a tendency for the value of the measured gradients to vary inversely with the wind speed. In general, the relationships given in this section should not be extrapolated to values outside the range of those indicated in the several figures which illustrate them.

5-10. CONDUCTION OF HEAT FROM THE GROUND

5-10.01 General. - In the preceding section the major heat fluxes to the snowpack have been considered. Ordinarily these heat fluxes--radiation and convection-condensation--are all that need be considered in the determination of daily melt quantities. Yet there is still another heat flux which, although negligible in daily computations of melt, becomes significant when the melt season as a whole is considered. This flux is the conduction of heat upward to the snowpack from the underlying ground. This source of heat has special hydrologic significance since it can cause melting during the winter and early spring when melt at the snow surface is non-existent. Thus the melt due to this cause is capable of priming the underlying soil in advance of the actual melt season, and may also help to ripen the snowpack, readying it for melt.

5-10.02 Ground-temperature gradients. - The flux of heat upward from the underlying ground to the snowpack during the winter and spring months results from thermal energy that is stored in the ground during the summer and early fall when no snow cover exists. During the summer months, the ground surface is heated, primarily by solar radiation and as a result the thermal gradient is directed into the ground. In consequence of this thermal gradient, heat is conducted downward into the ground, the amount being dependent upon the thermal gradient and the conductivity of the ground itself. Thus the ground may be warmed to a considerable depth. During the winter, with snow on the ground, this process is reversed. The ground surface is cooled to 32°F (or below) and the thermal gradient is directed upward. Figure 5 of plate 5-6 shows the annual variation in soil temperature at several depths for an area having a shallow winter accumulation of snow. This figure was prepared from soil temperature data secured in Minnesota by Algren. 1/ In figure 4, plate 5-6, these data are plotted to show vertical temperature gradients through the soil profile. Mean monthly values are given for the months September through April. Since snow is a good insulator,* the ground is shielded from the sub-freezing air temperatures of winter in the case of deep snowpacks. Before the deep pack is established, however, the ground may become frozen to some depth. Ground that may have been frozen before the deposition of the permanent snowpack will generally be thawed by the conduction of heat from greater depths once the protective snow cover is deposited. It will be noted in figure 4 that during brief transition periods in the spring and the fall the situation exists in which there is a dual temperature gradient in the ground. During the fall, the ground surface may be suddenly cooled by early snow fall and (or) cold air temperatures, resulting in an upward gradient near the surface, while deeper in the ground the gradient is still directed downward. Heat flows in both directions from the depth of maximum temperature, further warming deeper layers and conducting heat to the surface of the snowpack. In the spring, the gradients are directed in the opposite directions. The ground surface becomes warmed as the snow cover disappears and heat flows downward; at the same time heat is being conducted upward from greater depths to some intermediate area of minimum temperature. Typical ground temperature gradients determined from data secured at CSSL (see Tech. Bull. 16) for the months January through April are given in the following table:

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* See chapter 8.

Mean Ground Temperature Gradients, CSSL
(in °F per foot)

Month	Three feet to surface	One foot to surface
January	1.6	2.1
February	1.4	1.9
March	1.3	1.6
April	0.9	1.2

These data are for silty clay loam such as is found in the meadows at CSSL where the data were obtained. They are an average of five years of record. Data for earlier and later months are not given because of the transition periods previously discussed.

5-10.03 Thermal conductivity of the ground. - The quantity of heat transported by the conduction process across unit area of a plane parallel to the ground surface during unit time is given by the equation,

$$H_g = k \, dT/dz \tag{5-30}$$

where dT/dz is the temperature gradient in a direction perpendicular to the plane, and k is a proportionality factor known as the thermal conductivity. For soils, the thermal conductivity varies with the composition, the density, and also with the moisture content; generally speaking the thermal conductivity of a soil varies directly with its density and its moisture content. The heat flux by conduction is the result of both the temperature gradient and the thermal conductivity; however, these two terms are not mutually exclusive. The better the thermal conductivity of the soil, the less the temperature gradient, other things being equal, since the better conduction tends to equalize temperature differences. On the other hand, it is possible to maintain quite steep temperature gradients in a substance having a low thermal conductivity. Some typical values of thermal conductivities are given in the table below for a silty clay loam.

Thermal Conductivity for Unfrozen Fairbanks Silty Clay Loam

(after Kersten 25/)

Mositure Content	Density in lbs per cubic ft.		
	80	90	100
2.5%	0.0005		
18.0%	0.0019	0.0025	0.0031
25.0%	0.0022	0.0028	0.0036*
30.0%	0.0025	0.0029*	
40.0%	0.0027*		

*Saturated soil.

Reference is made to the report by Kersten 25/ for an extensive tabulation of thermal conductivities of many different soils.

5-10.04 Observed melt quantities. - Data on the melt occurring at the bottom of the snowpack as a result of the conduction of heat from the ground are difficult to obtain and are not ordinarily available. Special measurements were made, however, during one year at CSSL from which this melt could be computed (see Tech. Bull. 16). Measurements of the compaction and melt of the bottom layers of the snowpack were determined by means of a slide wire settling meter. Densities of these same layers were determined from deep pit measurements. From these data the change in water equivalent of the bottom layer through the season could be computed. These data are given in figure 5 of plate 5-6, along with the computed water equivalents. The monthly ground melt amounts from this figure are as follows:

Month	Inches
January	0.11
February	0.37
March	0.69
April	0.77
May	0.96

The total seasonal loss from 1 January to 9 June amounted to 3.26 inches. It will be noted that the melt rate accelerated as the season progressed. This may be partially explained by the increase in the thermal conductivity of the soil as it became progressively more moist; however, thermal conductivities computed from these melt data and available temperature gradient data exhibit too wide a variation (by comparison with Kersten's data). It is probable that, early in the season (January and February) some of the heat conducted to the bottom of the pack is being consumed in ripening the pack, without producing melt and runoff. This is to say, some of the heat is being used to bring the bottom layer to 32°F and to saturate it to its maximum free water holding capacity. These data, while not universally applicable, are felt to be generally indicative of the magnitudes and variations in the ground melt. It may thus be seen, considerable water is available to prime the soil in advance of the main melt season. For areas having greater temperature gradients and/or thermal conductivities, even greater melts from ground heat would be expected. Rough approximations of this amount may be made for other areas from the foregoing data on thermal conductivities and temperature gradients.

5-11. HEAT CONTENT OF RAIN WATER

5-11.01 Derivation of equation. - When rain falls on the snowpack it is cooled to the temperature of the snow, the quantity of heat involved being given up to the snow. For snowpacks isothermal at

32^oF, this release of heat results in snowmelt, while for colder packs this heat tends to raise the snow temperature to 32^oF. The amount of heat given up to the snow by the rainwater is directly proportional to the quantity of rainwater and to its temperature excess (above that of the snowpack). Considering a melting snowpack, for every degree centigrade the rainwater is in excess of the snow temperature (zero °C), and for every gram of rainwater, one calorie of heat is available. Thus, each centimeter depth of rainfall releases one langley (calorie per square centimeter) for each degree centigrade above freezing. That is,

$$H_p = (T_r - T_s)\ P_r \qquad (5\text{-}31a)$$

where $\underline{H_p}$ is the heat released by the rainfall in langleys, $\underline{T_r}$ and $\underline{T_s}$ are temperatures of the rainwater and snowpack, respectively, in degrees C, and $\underline{P_r}$ is the depth of rainfall in centimeters. English units may be substituted for the temperature and rainfall measurements in the above equation. Thus,

$$\begin{aligned} H_p &= 5/9\ (T_r - T_s)\ 2.54P_r \\ &= 1.41\ (T_r - T_s)\ P_r \qquad (5\text{-}31b) \end{aligned}$$

where $\underline{T_r}$ and $\underline{T_s}$ are now in degrees F, and $\underline{P_r}$ is in inches. $\underline{H_p}$ is still in langleys. For snow having a thermal quality of 100 percent, the resultant melt, $\underline{M_r}$, in inches, is given by,

$$\begin{aligned} M_r &= \frac{1.41}{203.2}\ (T_r - 32)\ P_r \\ &= 0.00695\ (T_r - 32)\ P_r \qquad (5\text{-}32) \end{aligned}$$

This equation is shown graphically in figure 2 of plate 5-6. It may be seen from this figure that the melt from rainfall is relatively minor for normal rainfall temperatures when compared with the quantity of rainfall itself. For example, one inch of rainfall at a temperature of 46^oF produces only 0.1 inch of melt.

5-11.02 Latent heat of fusion. - When rain falls on a subfreezing snowpack, an additional quantity of heat is also given to the pack. The rainwater that is frozen within the pack releases its latent heat of fusion (80 cal/g) to the snow. Thus for each inch of rainfall, 203.2 langleys are given up to the pack. In view of this large source of heat and the small specific heat of snow, it may be seen that subfreezing snowpacks cannot prevail during rainstorms of any considerable magnitude. For example, a snowpack six-feet deep having a mean density of 40 percent and a mean temperature of -10^oC could be brought to zero degrees C by 1.8 inches of rainfall at zero degrees C. (This topic is discussed further in chapter 8.)

5-11.03 Rain temperature. - The surface wet-bulb temperature is generally considered to be a suitable estimate of the temperature of the rain. Becuase of the usual near-saturated conditions that exist during rainstorms, dewpoint, or for that matter air temperature, may be used as rain temperature with very little error. Some hydrologists make the refinement of considering the mean temperature of, say, the kilometer of air above the snow surface. This usually lower temperature assumes the raindrops to be more nearly represented by the temperatures of the region through which they pass; that is, there is a lag in the change in the temperature of the falling rain as it attempts to assume that of its environment.

5-12. INTERRELATIONSHIPS BETWEEN COMPONENT MELTS

5-12.01 Examples. - By way of summary of this chapter, the figures of plates 5-7 and 5-8 are presented. These figures give the average component heat fluxes to the snowpack for the months November through June and for three days during the spring melt season for conditions as they exist at the CSSL basin. The meteorological conditions used to compute the fluxes given in plate 5-7 are the mean values for the five years 1946-1947 through 1950-1951. All heat fluxes are expressed in inches of melt (200 ly equal one inch melt), and are for basinwide conditions; the basin is assumed to have 100 percent snow cover for the entire period. Negative melts represent heat losses from the pack, the same correspondence between heat and melt quantities mentioned above being used. The relationships between observed meteorological conditions and resultant snowmelt previously presented and discussed in this chapter are used to calculate the component melts. It will be noted in figure 5 of plate 5-7 that for the months November through February, the total heat flux to the snowpack is on the average, negative. During March the total flux becomes positive; however, the amount of heat is small. During April there is a sudden acceleration in the heat supply to the snowpack which continues through June. By considering these heat fluxes, it can thus be seen why the snowpack accumulates through March and ablates thereafter. During March and April, some of the net heat supply to the pack is consumed in ripening the pack; by the latter part of April, however, melt is generally as indicated by the figures. Individual figures on these plates are discussed in the following paragraphs.

5-12.02 Figure 1 of plate 5-7 gives a general picture of the variation of insolation with the season at this mid-latitude station. It shows how the insolation received at the outer limits of the earth's atmosphere is, on the average, depleted by the atmosphere and by clouds. Furthermore, it shows the radiation actually absorbed by the snowpack, forest cover and snowpack albedo being considered. It is interesting to compare this absorbed radiation with that originally available at the outer limit of the earth's atmosphere, noting how little of this available energy is actually directly absorbed by the snow. This is especially true during the winter when frequent new snowfalls maintain a high albedo.

5-12.03 In figure 2 of plate 5-7, the absorbed radiation of figure 1 is converted to equivalent snowmelt assuming that 200 ly produce one inch of melt. This conversion is made so that the melt quantities due to radiation can be combined with other melt components which are expressed directly in inches of melt. Actually, this is somewhat misleading since the amounts given are not the actual melt that would be realized from the snowpack but are, rather, the heat fluxes to the snow. The actual resultant melt is dependent upon the thermal quality of the snow as has been discussed. Losses of heat from the snowpack are shown as negative melt quantities. Also shown in figure 2 is the average longwave radiation loss for this area, expressed in inches of melt. It includes the effects of cloud and forest cover, of temperature and humidity of the air, and of the changes in the temperature of the snow surface itself. Shortwave and longwave radiation melt are combined into a single curve showing the variation in radiation melt through the period of interest.

5-12.04 Figure 3 of plate 5-7 shows the variation of convection and condensation melts with time and also a combined convection-condensation melt curve. These values were computed from mean observed air temperatures and vapor pressures for the years of record at CSSL. A constant wind speed was assumed throughout, since no regular seasonal trend of wind speed was discernible from the available data.

5-12.05 Figure 4 of plate 5-7 shows melts resulting from conduction of heat from the ground and from the heat content of rainwater falling on the snowpack. The former was computed from mean ground temperature gradient data for CSSL and estimated thermal conductivities for the ground as given in Technical Bulletin 16. While the temperature gradient tends to decrease throughout the period under consideration, there is an increase in the conductivity of the ground during the spring melt season as a result of the increase in moisture content. This results in the rise of ground melt during the period March-May. Rain melt was computed using the rainfall amounts from the water balance for this area (see chap. 4). The quantities of heat given (in inches of melt) are largely the result of rain falling on the sub-freezing snowpack and releasing its latent heat of fusion. The additional increment of heat which results from the rainwater being cooled to 32°F is practically negligible.

5-12.06 Figures 1, 2, and 3 of plate 5-8 give a detailed, hour-by-hour picture of the principal heat fluxes to the snowpack for three representative days during the spring melt season. A clear day, a partly-cloudy day, and an overcast day are illustrated. The data presented in those figures are from some special lysimeter studies of snowmelt made at CSSL during the 1954 spring snowmelt season (see Res. Note 25). They are for melt at an open, unforested site and for days without precipitation. Ground melt is not included, as the lysimeter introduced an artificial effect with respect to this melt component.

5-12.07 Discussion. - In the preceding section of this chapter, each of the sources of thermal energy involved in the melting of the snowpack has been examined separately. Yet these heat fluxes are not independent of one another; rather they are quite interdependent, the degree of relationship being influenced by the terrain involved. For this reason it is well that these heat fluxes also be examined collectively and with consideration of the terrain. Moreover, the discussion has thus far been concerned with heat exchange and melt as a specified point only. Areal considerations are also involved in a complete understanding of heat exchange as it affects snowmelt. In this section, these broader aspects of snowmelt will be examined.

5-12.08 Since all the thermal energy involved in melting the snowpack has its ultimate source in the solar radiation reaching the earth, the other processes serve merely as intermediate means of heat transfer. Yet it is the measurement of some of the manifestations of solar energy, such as air temperature and vapor pressure, that are most commonly used as indexes of snowmelt. The earth's atmosphere is warmed but slightly by solar radiation passing through it, as was pointed out in the section on solar radiation. Over snowfields this absorption is increased, since a portion of the reflected beam is also absorbed. Nevertheless, the degree of heating of the air by solar radiation is relatively small. Of greater consequence is the heating of air which results from air passing over lands and objects which are heated by solar radiation. These give their heat to the air by the processes of convection and longwave radiation, the air being much less transparent to the latter than it is to solar radiation. Similarly, water surfaces and ground surfaces containing water may be heated by solar radiation and by the transfer of heat from the air by the process of convection and longwave radiation and give off energy in the process of evaporation, the latent heat of vaporization being consumed in the process. Thus water vapor is added to the air and may subsequently condense upon the snow surface, releasing its latent heat.

5-12.09 In view of the foregoing, it may be seen that over barren snowfields of great areal extent, and in the absence of advection of energy, the air can neither be heated appreciably above 32°F nor can its vapor pressure exceed 6.1 millibars (saturated vapor pressure at 32°F). Since the upper limit of snow surface temperature is 32°F and its vapor pressure thus restricted to 6.1 millibars in this situation, it cannot warm the air above this temperature either by convection of heat or longwave radiation, nor can it add moisture to the air in excess of its own vapor pressure. Thus any appreciable convection and condensation melts require bare ground or water surfaces and/or forest cover which can serve as exchange mechanisms to convert radiant energy into sensible heat and moisture. (Some heat may result from subsidense; however, this is ignored in the discussion which follows.)

5-12.10 Advection of thermal energy. - Considering the snowmelt within a given drainage basin, heat and moisture may be advected either into or out of the basin depending upon the forest cover and areal extent of the snow cover in the basin relative to its environment. If the basin under consideration is but a relatively small part of a much larger homogeneous area, then very little advection would ordinarily be expected. The air leaving the basin should have about the same temperature and moisture content as the air entering. Only in the case of exceptionally warm and moist or cold and dry air masses passing over the area would advective effects be of any consequence. Even then the basin under consideration would add or subtract only its small incremental share to the modification taking place in the airmass in passing over the larger homogeneous area. Advection plays a more important role in snowmelt where the basin under consideration is adjacent to and leeward of an area having markedly different characteristics. Thus considerable heat and moisture may be advected into a basin situated to the lee of a non-snow-covered area or of an open water area. This situation is exemplified by the snowfields of the mountain ranges along the Pacific Coast. Here air coming off the ocean first passes over the valleys to the west before reaching the snowfields. Another situation favorable to the advection of heat and moisture into a given drainage basin occurs where the area to the windward is more heavily forested than is the drainage basin itself. Thus, in the windward area the air may be warmed and moisture added to a greater extent than within the basin, the net result being the advection of moisture and heat into the area. A more or less barren drainage area to the lee of a heavily forested area would result in considerable advection during clear weather.

5-12.11 On the other hand, heat and moisture may be advected from a snow-covered basin. Even air initially warmer than 32°F and having a vapor pressure in excess of 6.1 millibars, may, in passing over the basin, be warmed above 32°F by convection and longwave radiation from the trees and barren areas, and made more moist by evaporation from snow-free areas and transpiration from trees. Of course air initially having a temperature and dewpoint less than 32°F can be modified to at least those temperatures by even a barren snowfield.

5-12.12 Local effects. - Considering now snowmelt in a basin in which no advection of heat energy takes place: the condition may be approached in actual situations where weak pressure gradients exist or where a high pressure area stagnates over the basin being considered, and where the basin is but a small part of a much larger area having similar conditions of snow and forest cover. Since the air leaving the basin has, under these conditions, the same heat content as the air entering, all thermal energy, other than solar radiation, used in melting the snowpack must be generated within the basin. This situation is thus referred to as the "local climate" or "radiation climate." Since the air temperature remains constant, the heat added to the air by convection and longwave radiation from trees and snow-free ground surfaces is just balanced by the heat given up by the air to the snow by the same processes. Thus, in this situation, the total energy represented by the snowmelt and the water lost to the atmosphere by evapotranspiration is equal to the net amount of solar radiation absorbed within the area. (See supplement to Research Note 19.) One would thus expect the melt under these conditions of a local climate to vary directly with the density of the forest, the over-all albedo decreasing with increasing forest cover. Heavily forested areas, however, tend to result in advection of heat and moisture from the area, whereas more barren areas favor advection into the area. This will be considered further in the following paragraph.

5-12.13 Forest effects. - Is the melt rate greater in forested or in barren areas? This is a question that has long been debated by those concerned with snowmelt. The answer is simply this: Sometimes it is greater in the forest and sometimes it is greater in the open, depending upon the size of the clearing and other factors. With the background of this chapter it is possible to qualify this statement. But first the question of deposition of snow should be dealt with. Since relative melt rates in the forest and in the open are often judged by the disappearance of snow, the usually greater deposition in the open tends to bias the answer in favor of the forested areas where there is generally less snow to begin with. However, generalizations must be made with as much caution in the case of deposition as with melt. While small forest clearings usually collect the maximum snowpack, larger open areas may be scoured of snow which is then deposited in greater depths near and under the surrounding trees (see chap. 3 for a discussion of deposition effects). Melt rates are generally considerably greater in large clearings than they are in forested areas. These areas not only have approximately the same air temperatures and vapor pressures as do the forested areas, due to large-scale mixing of the air, (even though the heat and moisture are given to the air primarily as a result of the interception of solar radiation by the trees), but the higher wind speeds encountered over the snow in the open result in greater melts due to convection and condensation. Since the melt component due to absorbed solar radiation is considerably greater in the open than it is in forested areas, the greater convection-condensation melt component in the open, coupled with this greater radiation melt, results in greater melt in the open. (Of course evaporation might prevail, in which case heat losses in the open due to this cause would exceed those in the forest. The heat transfer due to evaporation is, however, relatively small.) Longwave radiation loss in the open is, of course, greater than in the forest; however, since a part of the solar energy absorbed by the forest canopy is re-radiated to space and part is used up in warming the air and transpiring moisture which in turn are conveyed to the snow surface, the greater loss of longwave radiation in the open is usually more than made up for by the greater shortwave radiation gain.

5-12.14 It has been observed that the last snow patches to disappear in the spring are usually found in small clearings in the forest; that is, clearings having a diameter about the same as the heights of the surrounding trees. The reason for this is two-fold: (1) the greater deposition of snow in these sheltered areas and, (2) the lesser melt rates in these areas. The first of these reasons is dealt with in chapter 3. The second is the result of the shading of the sites from direct solar radiation by the surrounding trees, while, at the same time, longwave loss is affected to a lesser degree. Then too, convection-condensation melts are less than in larger open areas by virtue of the lesser wind speeds in these areas. In summary, then, it may be said that melt rates are greatest in large open areas and least in small forest clearings, the melt rate in the forest being intermediate between these two. It is to be emphasized, however, that the greater melt rates of large open areas is still contingent upon the existence of surrounding forests or bare ground for a supply of sensible heat and moisture. Large, non-forested, snow covered areas such as are found in parts of Canada would not have as high melt rates as would a large clearing in the otherwise forested area.

5-12.15 From the foregoing the true relationship between air temperatures, vapor pressures, and snowmelt can be deduced. It may be seen that upper air temperatures and humidities should <u>not</u> be good indexes of snowmelt since they do not adequately reflect the conversion of radiant energy into sensible heat and moisture. The air may be warmed and its humidity increased in passing over trees and bare ground warmed by radiation, and it may in turn convey this thermal energy to the snowpack without any change indicated in the upper air. Only surface temperatures and humidities adequately reflect this heat transfer. Moreover, since more of the radiant energy is manifest in the air temperature and humidity in forested areas than in barren areas, temperature and humidity indexes should be expected to increase in accuracy with the degree of forest cover. These effects are considered in detail in the following chapter which deals with the practical computation of snowmelt by means of indexes.

5-13. SUMMARY

5-13.01 The relationships presented in this chapter are summarized below in outline form. The general snowmelt equation is given, followed by the equations which give the component melts. References are made to figures which illustrate these relationships where applicable. Symbols used are defined in the text.

<u>GENERAL SNOWMELT EQUATION</u> (M)

$$M = H_m/203.2 \qquad (5\text{-}2a)$$

(See fig. 1, pl. 1)

where,

$$H_m = H_{rs} + H_{rl} + H_c + H_e + H_g + H_p$$

SHORTWAVE RADIATION MELT (M_{rs})

Insolation
(See figs. 3 & 4, pl. 1)

Effect of Clouds

$$I/I_c = 1 - (1-k) N \qquad (5\text{-}6b)$$

where,

$k = 0.18 + 0.024z$
(see Fig. 5, pl. 5-1)

Effect of forest cover
(See fig. 1, pl. 5-2)

Effect of slope
(See Fig. 6, pl. 5-1)

Albedo
(See figs. 2, 3, & 4, pl. 5-2)

Absorption

$$I_d = I_a e^{-kd} \qquad (5\text{-}8)$$

(See fig. 5, pl. 5-2)

LONGWAVE RADIATION MELT (M_{rl})

Radiation from clear skies (R_d)

$$R_d = \sigma T_a^4 (a + b \sqrt{e_a}) \qquad (5\text{-}9)$$

(See figs. 2 & 3, pl. 5-3)

Radiation from snowpack (R_u)
(See fig. 2, pl. 5-3 -- Black body curve)

Net radiation exchange -- clear skies (R_c)
(See Fig. 4, pl. 5-3 -- over melting snow)

Effect of Clouds

$$R = \sigma (T_c^4 - T_s^4) \quad \text{(overcast)} \qquad (5\text{-}12)$$

(See fig. 4, pl. 5-3: black body curve)

$$R = R_c (1 - kN) \qquad (5\text{-}13)$$

where,

$k = 1 - 0.24\,z$
(See fig. 5, pl. 5-3)

Effect of forest cover

$$R_f = \sigma (T_a{}^4 - T_s{}^4) \quad \text{(Solid canopy)} \qquad (5\text{-}16)$$

$$R = FR_f + (1\text{-}F)\,R_c \quad \text{(partial canopy)} \qquad (5\text{-}17)$$

(See fig. 6, pl. 5-3)

CONVECTION MELT (M_c)

$$M_c = k_c\,(T_a - T_s)\,v_b \qquad (5\text{-}28a)$$

$$M_c = k_c'\,(p/p_o)(z_a z_b)^{-1/6}\,(T_a - T_s)\,v_b \qquad (5\text{-}28b)$$

(See fig. 1, pl. 5-5)

CONDENSATION MELT (M_e)

$$M_e = 8.5\,k_e\,(e_a - e_s)\,v_b \qquad (5\text{-}27a)$$

$$= k_e'\,(z_a z_b)^{-1/6}\,(e_a - e_s)\,v_b \qquad (5\text{-}27d)$$

(See fig. 2, pl. 5-5)

GROUND MELT (M_g)

$$H_g = k\,dT/dz \qquad (5\text{-}30)$$

(See figs 4 & 5, pl. 5-6)

RAIN MELT (M_p)

$$M_r = 0.00695\,(T_r - 32)P_r \qquad (5\text{-}32)$$

(See fig. 2, pl. 5-6)

5-14. REFERENCES

1/ ALGREN, A.B., "Ground temperatures as affected by weather conditions," Heating, Piping & Air Conditioning, June 1949, pp. 111-116.

2/ ANDERSON, Ernest R., "Energy-budget studies," Geological Survey Circular 229 (Water Loss Investigations: Volume -- Lake Hefner Studies Technical Report), 1952, pp. 71-119.

3/ ANDERSON, E.R., L.J. Anderson, and J.J. Marciano, "A review of evaporation theory and development of instrumentation," (Interim Report: Lake Mead Water Loss Investigations), Report 159, Navy Electronics Laboratory, San Diego, Calif., February 1950.

4/ ANGSTROM, A., "On the radiation and temperature of snow and the convection of the air at its surface," Arkiv. f. Math. 13, Nr. 21, 1919.

5/ BOWEN, I.S., "The ratio of heat losses by conduction and by evaporation from any water surface," Phys. Rev., Vol. 27, June 1926, pp. 779-787.

6/ BRUNT, David, Physical and Dynamical Meteorology, Cambridge Univ. Press, 1952.

7/ de QUERVAIN, M., "Evaporation from the snowpack," (translated from the German in Snow Investigations Research Note 8, October 1952).

8/ DUNKLE, R.V. and others, "Non-selective radiometers for hemispherical irradiation and net radiation interchange measurements," Univ. of Calif., Div. of Engrg. Research, Report No. 9, (Report Code NR-015-202), October 1949.

9/ ELSASSER, Walter M. "Heat transfer by infrared radiation in the atmosphere," Harvard Meteor. Studies, No. 6, 1942.

10/ FALCKENBERG, G., "The infrared absorption constants of some meteorologically important substances," (Absorptionskonstantan einiger meteorologisch wichliger, Korper fur infrarote wellen; test in German). (Meteoro. Zeitschr.) Vol. 45, September 1948, pp 334-337.

11/ FRITZ, Sigmund, "Solar radiant energy and its modification by the earth and its atmosphere," Compendium of Meteorology, American Meteor, Society, Boston, Mass., 1951, pp. 13-33.

12/ FRITZ, Sigmund and Torrence H. MacDonald, "Average solar radiation in the United States," Heating and Ventilating, Vol. 46, July 1949, pp. 61-64

13/ GEIGER, Rudolf, The Climate near the Ground, (A translation by Milroy N. Stewart and others, of the second German edition of Das Klima der Bodennahen Luftschicht with revisions and enlargements by the author.) Harvard Univ. Press, Cambridge, Mass. 1950.

14/ GERDEL, R. W., "Penetration of radiation into the snowpack," Trans. Amer. Geophys. Union, Vol. 29, No. 3, June 1948, pp. 366-374.

15/ GERDEL, R. W., M. Diamond, and K. J. Walsh, "Nomographs for computation of radiation heat supply," SIPRE Research Paper 8, Corps of Engineers, February 1954.

16/ GIER, J.T. and R.V. Dunkle, "Total hemispherical radiometers," AIEE Transactions, Vol. 70, 1951.

17/ GOSS, John R., and Frederick A. Brooks, "New constants for the empirical expression for downcoming atmospheric radiation under cloudless sky," (Paper presented at 134th Nat'l Meeting, AMS, Berkeley, California, December 1954.)

18/ HALSTEAD, Maurice H., "The relationship between wind structure and turbulence near the ground," Publications in Climatology, Vol. 4, No. 3 (The Johns Hopkins Univ., Lab. of Climatology), Seabrook, New Jersey, 1951.

19/ HAMON, Russell, W., Leonard L. Weiss, and Walter T. Wilson, "Insolation as an empirical function of daily sunshine duration," Mon. Wea. Rev., Vol. 82, No. 6, June 1954, pp. 141-146.

20/ HAND, I.F., "Pyrheliometers and phrheliometric measurements," U.S. Wea. Bureau, Washington, D. C., 1946.

21/ HAURWITZ, B., "Insolation in relation to cloud type," Jour. of Meteor., Vol. 5, No. 3, June 1948, pp. 110-113.

22/ HOECK, Erwin, "The influence of radiation and temperature on the melting process of the snow cover," (Der einfluss der strahlung und der temperatur auf den schmelzprozess der schneedeck; text in German), Bietr. Geol. Schweiz - Geotech. Serie-Hydrologie, 1952.

23/ HUBLEY, Richard C., "Measurements of diurnal variations in snow albedo on Lemon Creek Glacier, Alaska," Jour. of Glaciology, Vol. 2, No. 18, October 1955, pp. 560-563.

24/ JOHNSON, Francis, S., "The solar constant," Jour. of Meteor., Vol. 11, No. 6, December 1954, pp. 431-439.

25/ KERSTEN, Miles, S., "Laboratory research for the determination of the thermal properties of soil (final report)," Engrg. Exper. Station, Univ. of Minn., June 1949.

26/ KIMBALL, H.H., "Variations in total and luminous solar radiation with geographic position in the United States," Mon. Wea. Rev., Vol. 47, No. 11, November 1919, pp. 769, 793.

27/ KLEIN, Wm. H., "Calculation of solar radiation and the solar heat load on man," Jour. of Meteor., Vol. 5, No. 4, August 1948, pp. 119-129.

28/ LIGHT, Phillip, "Analysis of high rates of snow-melting," Trans. Amer. Geophys. Union, part I, 1941, pp. 195-205.

29/ MacDONALD, T.H., and Norman B. Foster, "Pyrheliometer calibration program of the U.S. Weather Bureau," Mon. Wea. Rev., Vol. 82, No. 8, August 1954, pp. 219-227.

30/ MANTIS, Homer T. (ed.), "Review of the properties of snow and ice," (SIPRE Report 4), Univ. of Minn., Inst. of Technology, Engrg. Exper. Station, March 1951, (Chaps. 5-7), pp. 52-85.

31/ MILLER, David H., "Snow cover and climate in the Sierra Nevada, California," Univ. of Calif. Publications in Geography, Vol. 11, Univ. of Calif. Press, Berkeley, 1955.

32/ NEWMANN, U., "Insolation in relation to cloud amount," Mon. Wea. Rev., Vol. 82, No. 11, November 1954, pp. 317-319.

33/ SUTTON, O.G., Micrometeorology, McGraw-Hill Book Co., Inc., 1953.

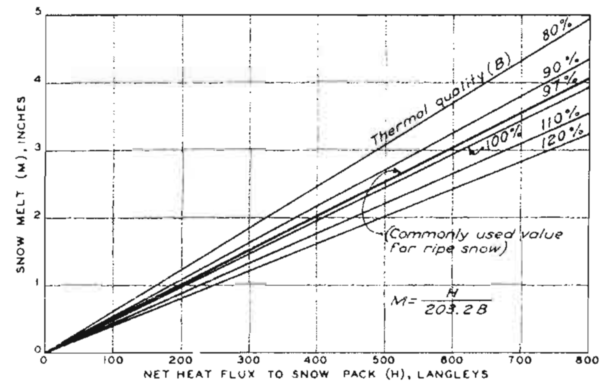
34/ SVERDRUP, H.U., "The eddy conductivity of the air over a smooth snow field," Geofysiske Publikasjoner, Vol. 11, No. 7, 1936, pp. 1-69

35/ SVERDRUP, H.U., Oceanography for Meteorologists, Prentice-Hall, Inc., New York, 1942.

36/ The Johns Hopkins University, Laboratory of Climatology, "Micrometeorology of the surface layer of the atmosphere; the flux of momentum, heat, and water vapor, final report," Publications in Climatology, Vol. 7, No. 2, Seabrook, New Jersey, 1954.

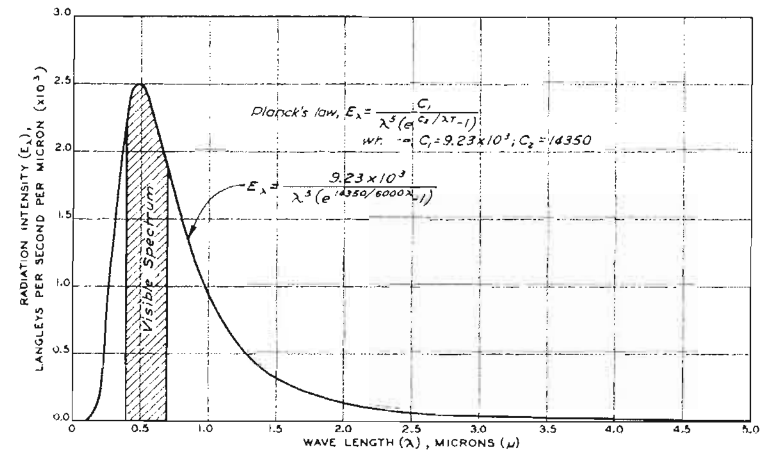
37/ THRELKELD, J.L. and R.C. Jordan, "Solar radiation during cloudless days," Heating, Piping and Air Conditioning, February 1955, pp. 117-122.

38/ WILSON, Walter T., "An outline of the thermodynamics of snow melt", Trans. Amer. Geophys. Union, Part I, 1941, pp. 182-195.



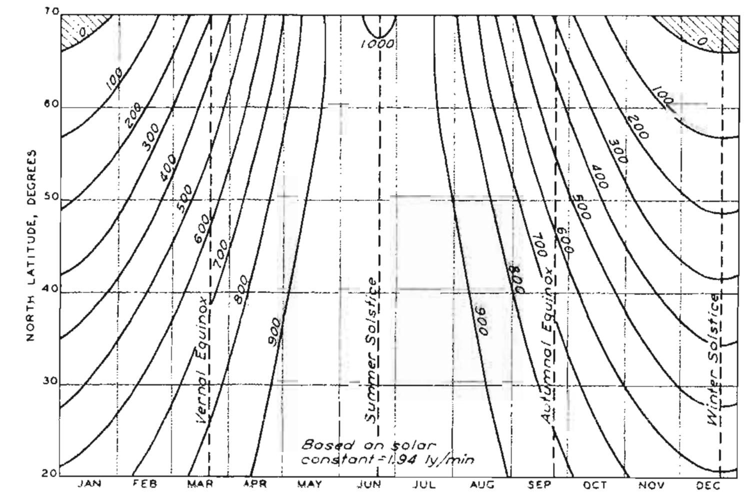
SNOW MELT RESULTING FROM THERMAL ENERGY

FIGURE 1



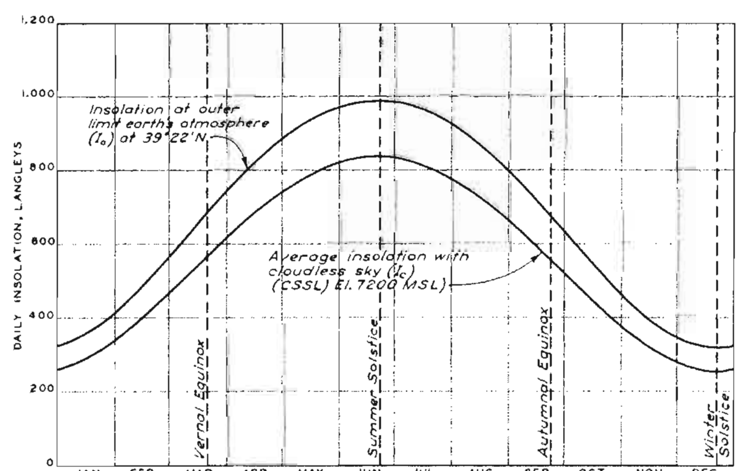
SPECTRAL DISTRIBUTION OF RADIATION INTENSITY
FOR A BLACK BODY AT A TEMPERATURE OF 6,000°K
(THEORETICAL RADIATION EMITTED BY THE SUN)

FIGURE 2



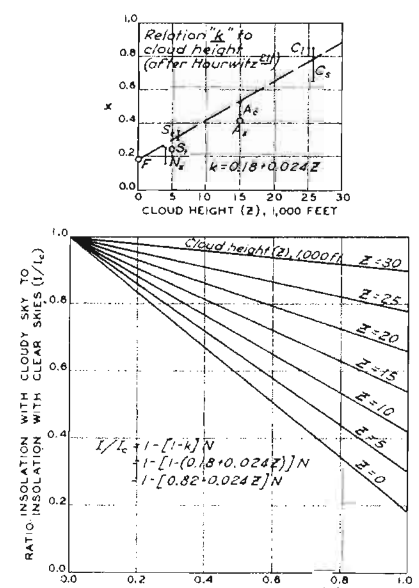
DAILY INSOLATION AMOUNTS OUTSIDE
THE EARTH'S ATMOSPHERE (LANGLEYS)

FIGURE 3



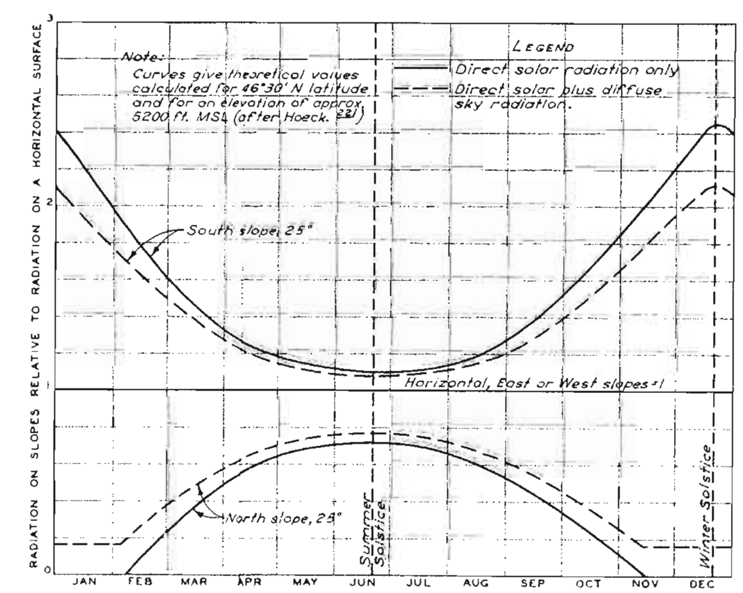
DAILY INSOLATION AMOUNTS AT CSSL

FIGURE 4



VARIATION OF INSOLATION WITH
CLOUD HEIGHT AND AMOUNT

FIGURE 5



DAILY SOLAR RADIATION ON CLEAR DAYS ON NORTH AND
SOUTH SLOPES OF 25-DEGREE GRADIENT RELATIVE TO
RADIATION ON A HORIZONTAL SURFACE

FIGURE 6

SNOW INVESTIGATIONS
SUMMARY REPORT
SNOW HYDROLOGY

SHORTWAVE RADIATION

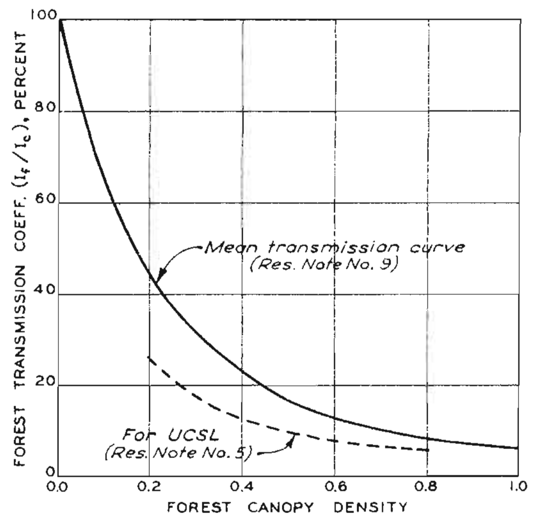
SHEET 1 OF 2

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

PREPARED C.S.H.	SUBMITTED C.S.H.	TO ACCOMPANY REPORT DATED 30 JUNE 1956
DRAWN: S.J.	APPROVED: D.M.R.	PD-20-25/28

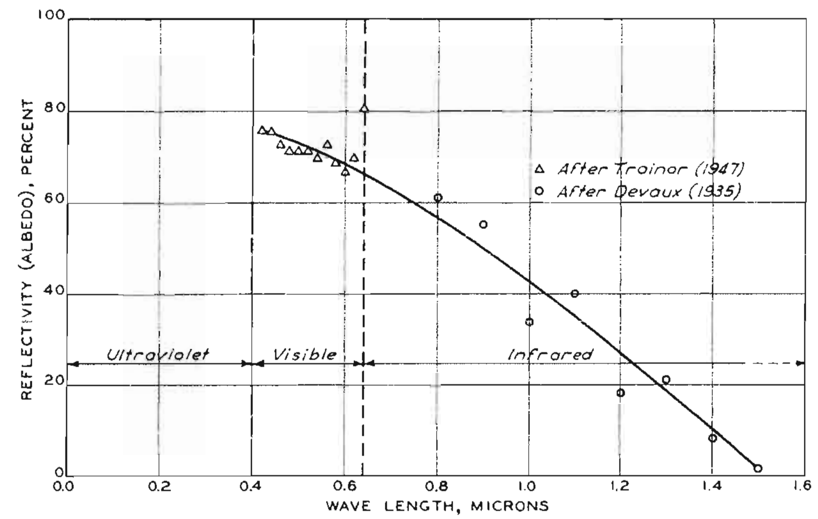
PLATE 5-1

Note:
Transmission curves are for daily insolation amounts. They represent mean conditions during the spring snow-melt season. Actual transmission of insolation may differ from that indicated due to differences in the solar altitude, height and spacing of trees, and differences in types of trees.



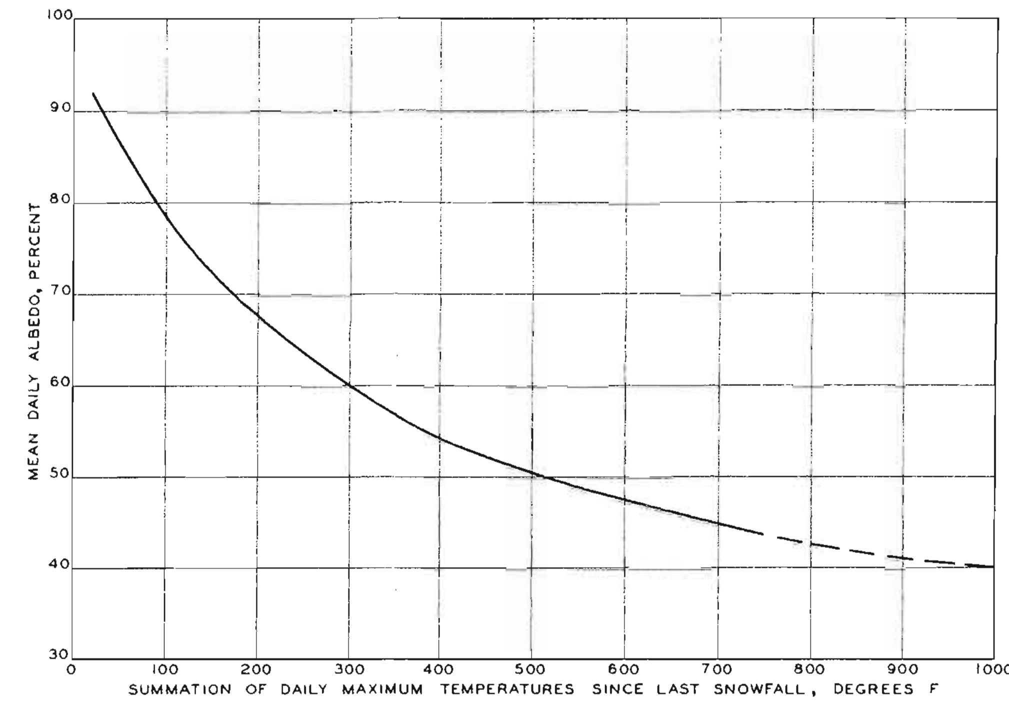
TRANSMISSION OF INSOLATION BY FOREST CANOPY
(CONIFEROUS FORESTS)

FIGURE 1



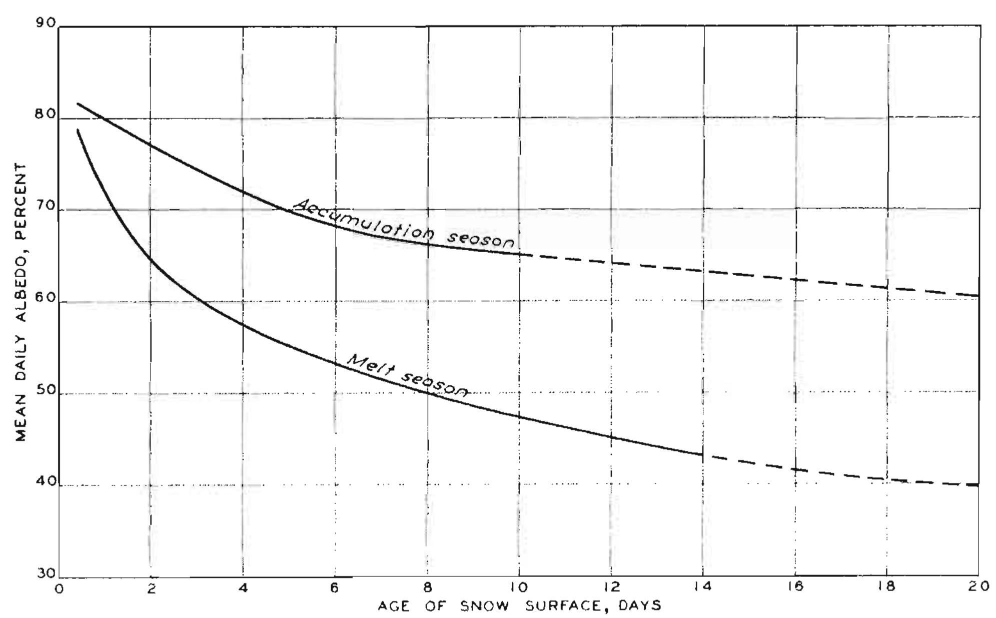
APPROXIMATE SPECTRAL REFLECTIVITY OF MELTING SNOW
(FROM DATA GIVEN IN SIPRE REPORT 4[39])

FIGURE 2



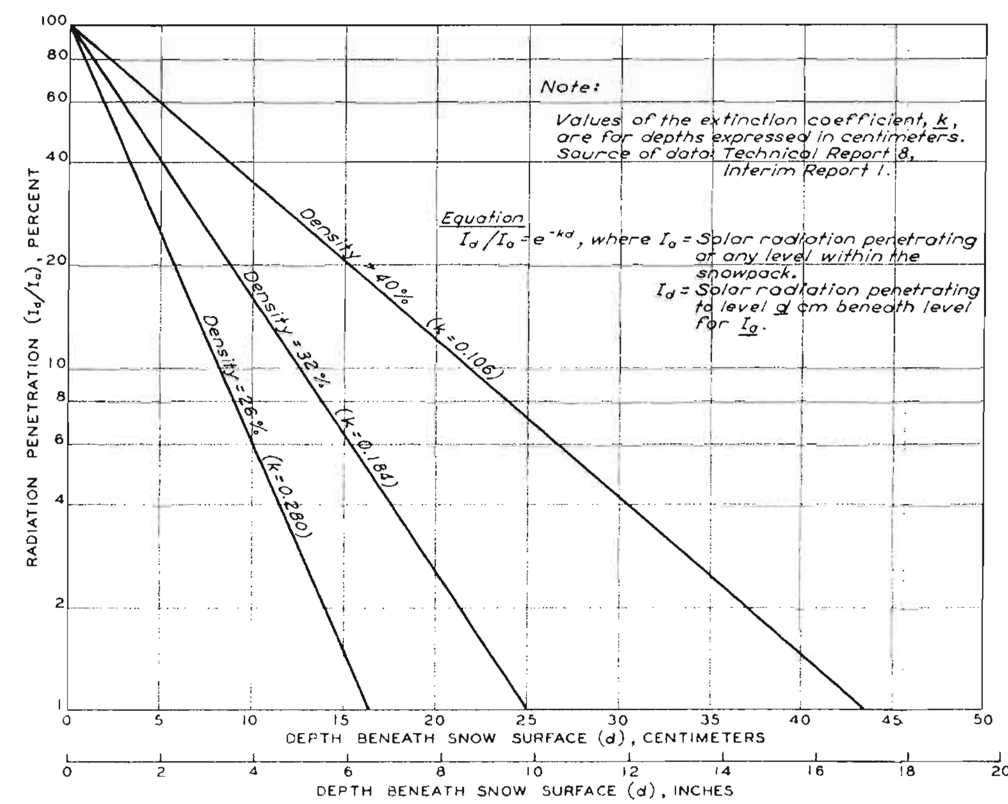
VARIATION OF ALBEDO WITH ACCUMULATED TEMPERATURE INDEX

FIGURE 3



VARIATION IN ALBEDO WITH TIME

FIGURE 4



PENETRATION OF SOLAR RADIATION INTO THE SNOW PACK

FIGURE 5

SNOW INVESTIGATIONS SUMMARY REPORT
SNOW HYDROLOGY

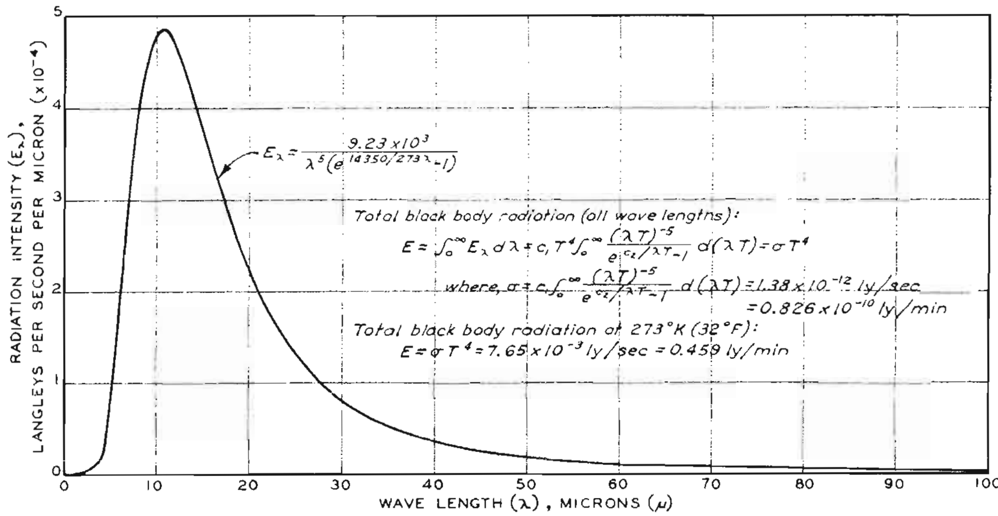
SHORTWAVE RADIATION

SHEET 2 OF 2

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U.S. ARMY

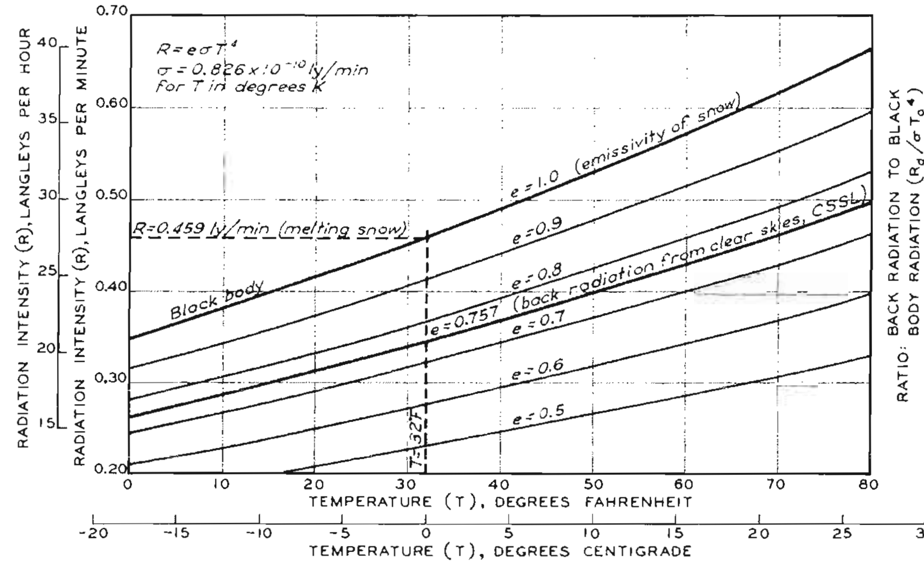
PREPARED: C.E.H. SUBMITTED: C.E.H. TO ACCOMPANY REPORT DATED 30 JUNE 1956
DRAWN: R.V. APPROVED: D.M.R.

PD-20-25/29



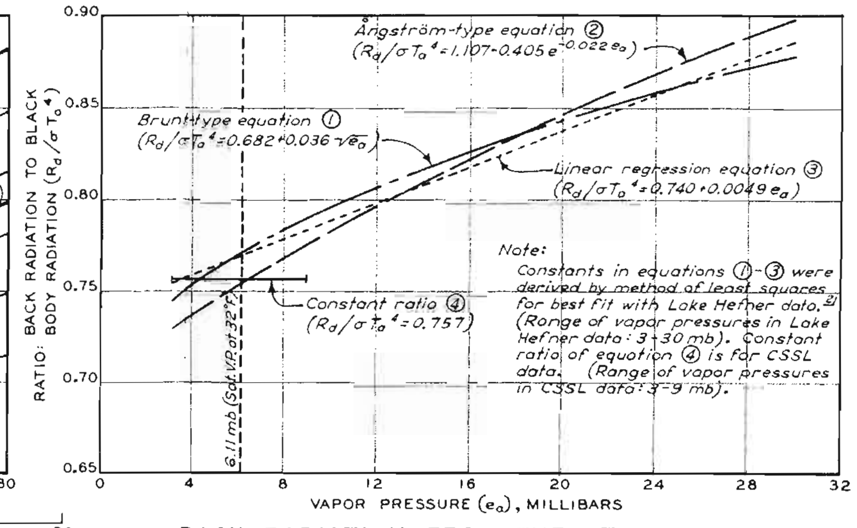
SPECTRAL DISTRIBUTION OF RADIATION INTENSITY FOR A BLACK BODY AT A TEMPERATURE OF 273°K (32°F) (THEORETICAL RADIATION EMITTED BY MELTING SNOW)

FIGURE 1



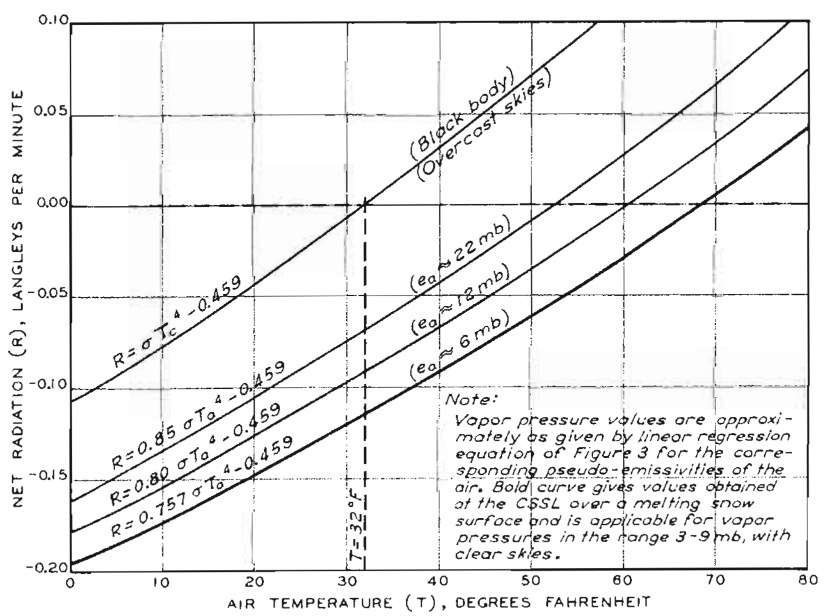
LONG-WAVE RADIATION EMITTED IN ACCORDANCE WITH STEFAN'S LAW (FOR VARIOUS EMISSIVITIES)

FIGURE 2



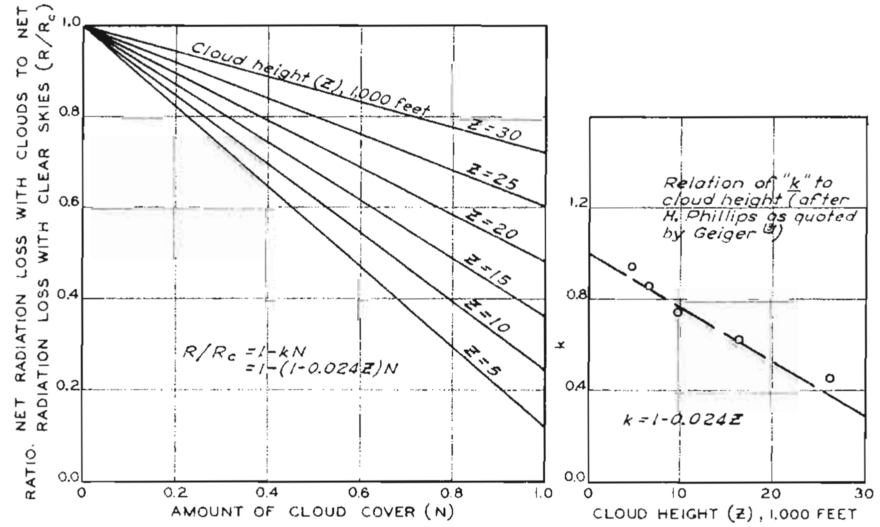
BACK RADIATION FROM THE ATMOSPHERE WITH CLEAR SKIES

FIGURE 3



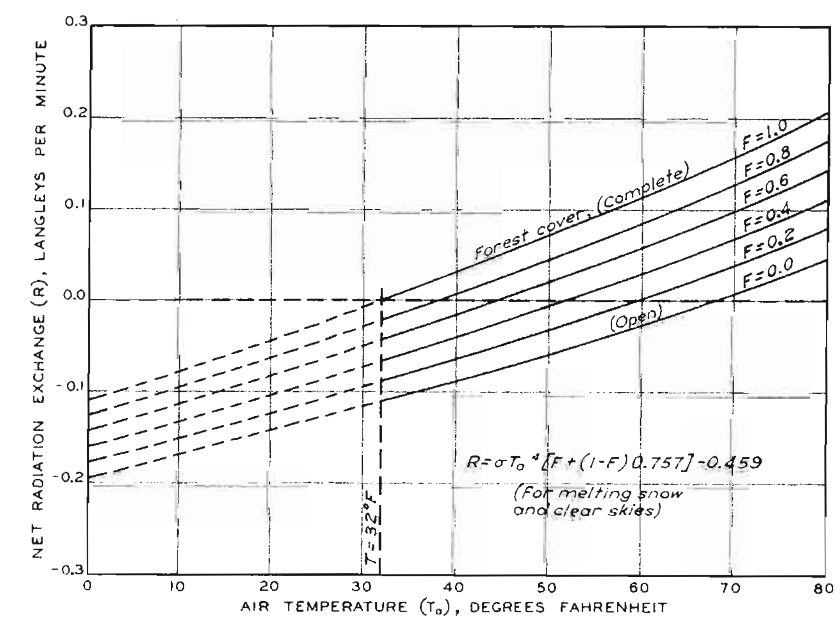
NET LONG-WAVE RADIATION EXCHANGE BETWEEN A MELTING SNOW PACK AND THE EARTH'S ATMOSPHERE

FIGURE 4



VARIATION IN NET LONG-WAVE RADIATION LOSS WITH CLOUD HEIGHT AND AMOUNT

FIGURE 5



NET LONG-WAVE RADIATION EXCHANGE IN FORESTED AREAS

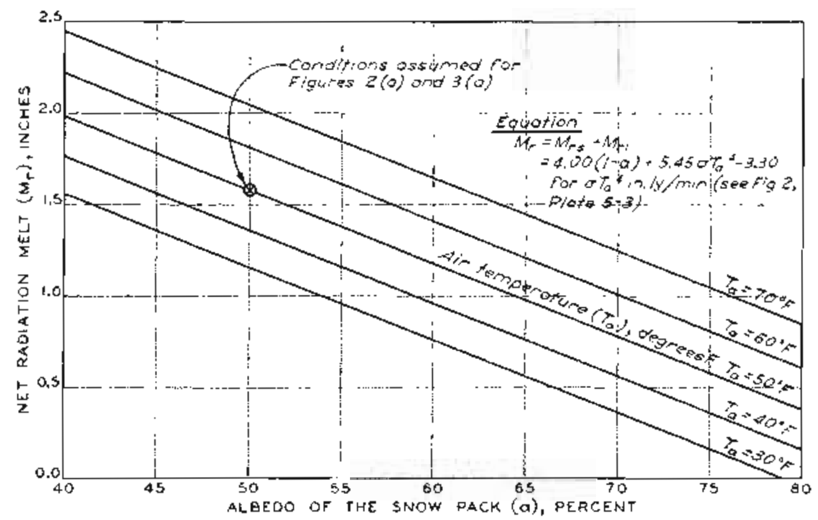
FIGURE 6

SNOW INVESTIGATIONS
SUMMARY REPORT
SNOW HYDROLOGY

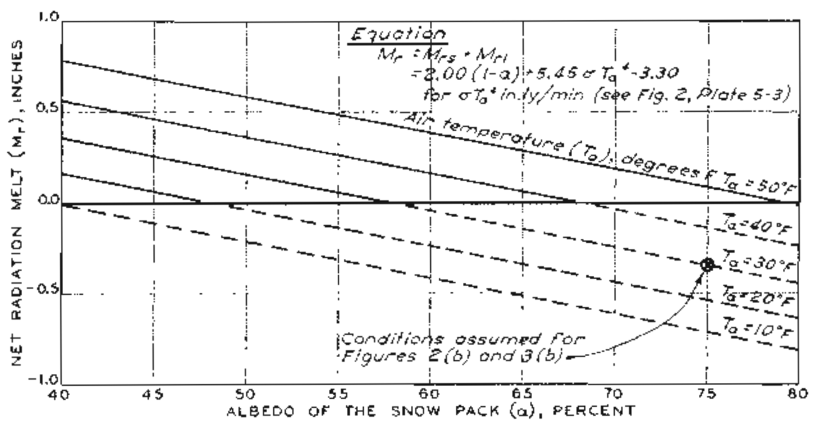
LONGWAVE RADIATION

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

PD-20-25/30



(a) DURING SPRING – 20 MAY
[INSOLATION (I_c) = 800 LY/DAY]

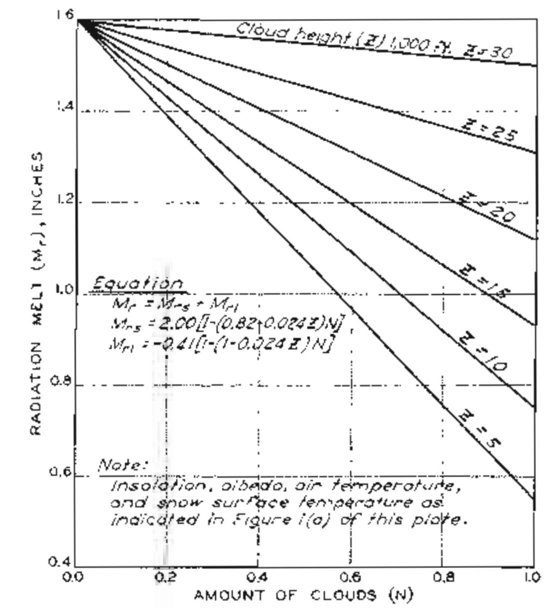


(b) DURING WINTER – 15 FEBRUARY
[INSOLATION (I_c) = 400 LY/DAY]

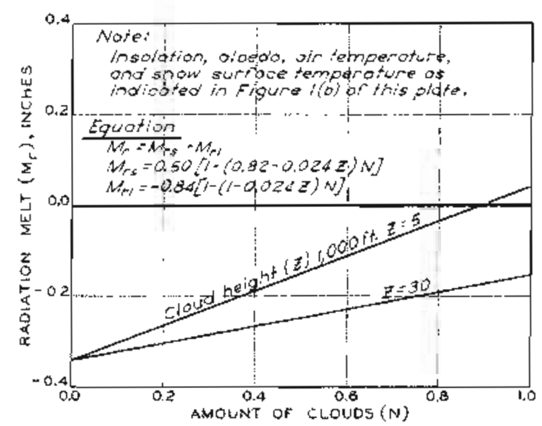
DAILY RADIATION MELT IN THE OPEN WITH CLEAR SKIES

FIGURE 1

Note:
Back radiation computed using constant ratio, $R_a/\sigma T_a^4 = 0.757$. Snow assumed to be at 32 degrees F (see Figure 4, Plate 5-3).
M_r (in) = H_r (ly)/200



(a) DURING SPRING – 20 MAY

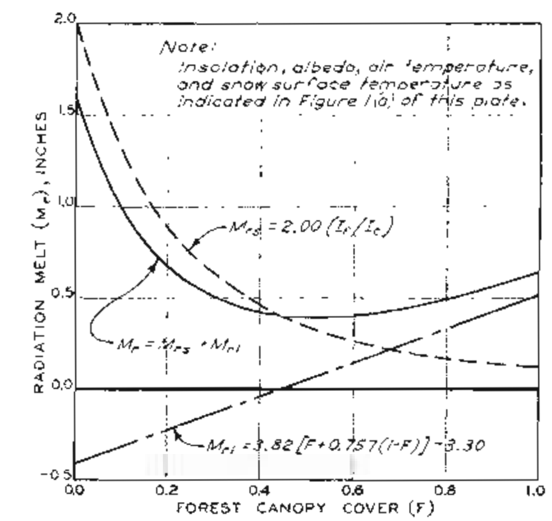


(b) DURING WINTER – 15 FEBRUARY

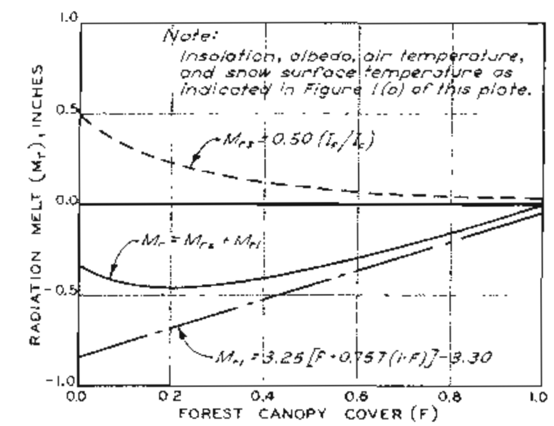
DAILY RADIATION MELT IN THE OPEN WITH CLOUDY SKIES

FIGURE 2

Note:
Insolation melt from Figure 5, Plate 5-1; long-wave radiation melt from Figure 5, Plate 5-3.
M_r (in) = H_r (ly)/200



(a) DURING SPRING – 20 MAY



(b) DURING WINTER – 15 FEBRUARY

DAILY RADIATION MELT IN THE FOREST WITH CLEAR SKIES

FIGURE 3

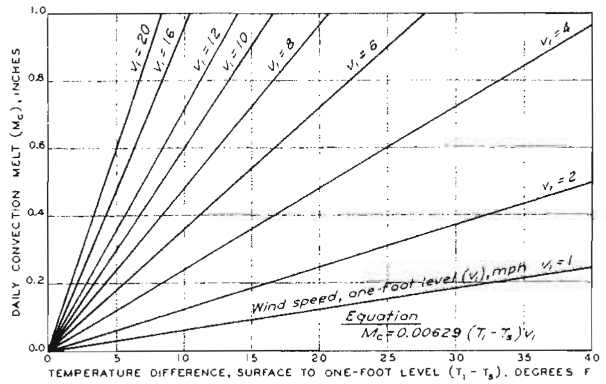
Note:
Values of I_r/I_c are from Figure 1, Plate 5-2; long-wave radiation melt from Figure 6, Plate 5-3.
M_r (in) = H_r (ly)/200

SNOW INVESTIGATIONS SUMMARY REPORT
SNOW HYDROLOGY

GENERALIZED RADIATION MELT SUMMARY FOR CENTRAL SIERRA SNOW LABORATORY

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

PD-20-25/31



Notes:

1. The general equation for daily convection melt (M_c) in inches, is as follows:

   $M_c = 0.00629\ (p/p_o)(Z_a Z_b)^{-1/6} (T_a - T_s) v_b$

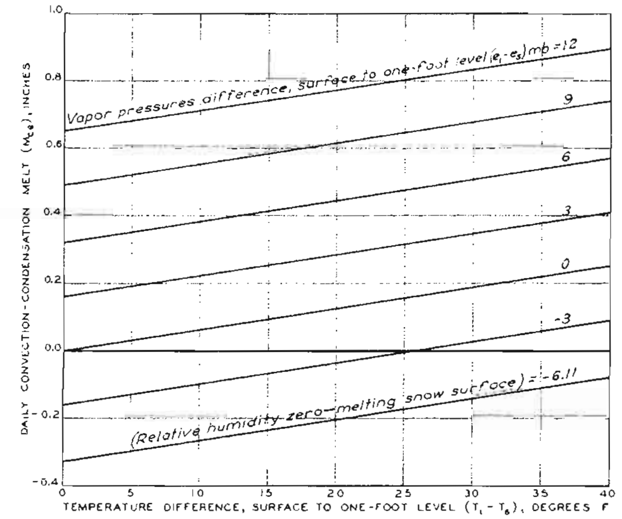
   where p/p_o is the ratio of atmospheric pressure at the site to standard sea-level pressure; Z_a is the height in feet above the snow surface of the air temperature measurement, T_a (in degrees F), and Z_b is the height in feet above the snow surface of the wind speed measurement, v_b (in mph); T_s is the snow surface temperature.

   This equation strictly applies only to unforested sites and to ripe snow packs.

2. Melt quantities given by the figure are for sea-level conditions ($p/p_o = 1$) and for T_a and v_b measured at the one-foot level [$(Z_a Z_b)^{-1/6} = 1$]. (See figure 5 for corrections for other heights of measurement and Figure 6 for adjustments for elevation.)

CONVECTION MELT

FIGURE 1

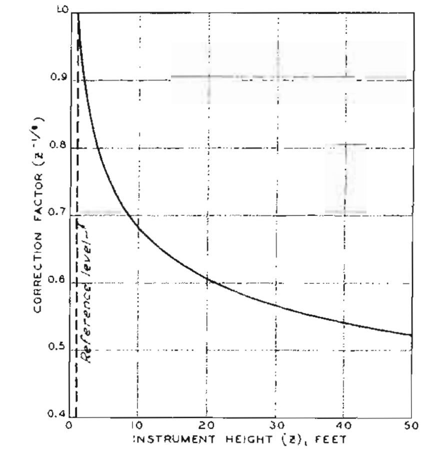


Note:

General equations for convection and condensation melts given in notes on Figures 1 and 2. Other notes given on these figures apply here also. Indicated melt is for wind speed of 1 mph measured at the one-foot level; to get melt, multiply by actual wind speed of this level.

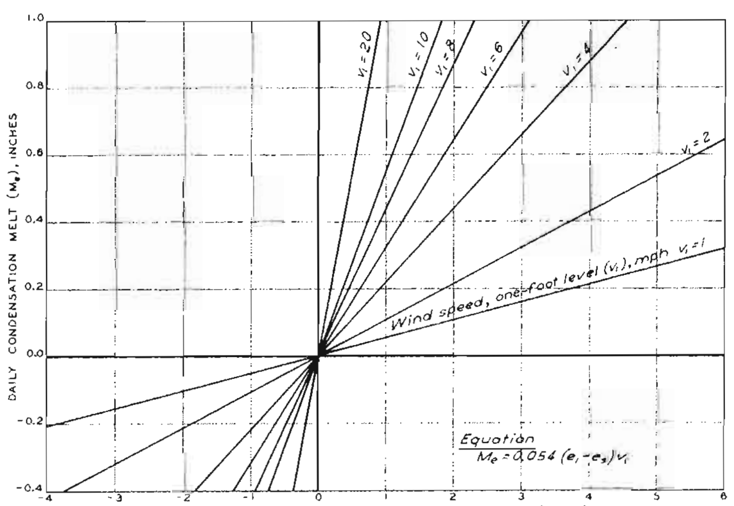
CONVECTION-CONDENSATION MELT FOR ONE MILE PER HOUR WIND SPEED

FIGURE 3



MEASUREMENT HEIGHT CORRECTION FACTOR

FIGURE 5



Notes:

1. The general equation for daily condensation melt (M_e) in inches, is as follows:

   $M_e = 0.054\ (Z_a Z_b)^{-1/6} (e_a - e_s) v_b$

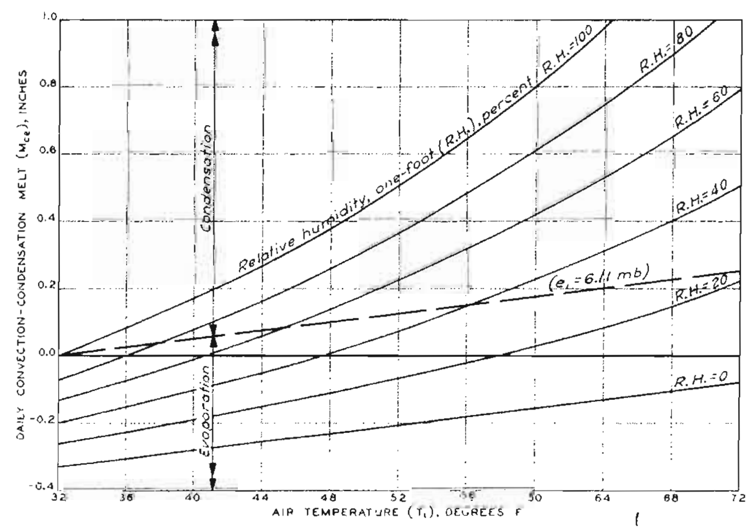
   where Z_a is the height in feet above the snow surface of the vapor pressure measurement, e_a (in mbs), and Z_b is the height in feet above the snow surface of the wind speed measurement, v_b (in mph); e_s is the saturated vapor pressure of the snow surface temperature.

   This equation strictly applies only to unforested sites and to ripe snow packs.

2. Melt quantities given by the figure are for e_a and v_b measured at the one-foot level [$(Z_a Z_b)^{-1/6} = 1$]. (See Figure 5 for corrections for other heights of measurement.)

CONDENSATION MELT

FIGURE 2

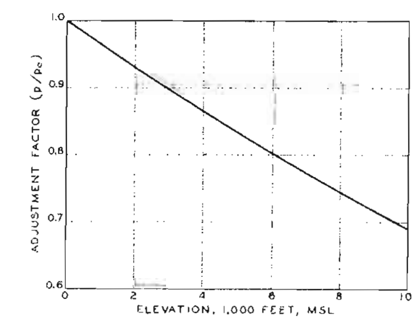


Note:

Melt quantities given are for measurements of air temperature, humidity, and wind speed made at one-foot level (see Figure 5 for corrections for other heights of measurement) and for sea-level conditions (see Figure 6 for adjustment for other elevations). Indicated melt is for 1 mph wind speed and must be multiplied by actual wind speed. A melting snow surface is assumed (snow surface temperature, 32 degrees F; vapor pressure, 6.11 mb).

CONVECTION-CONDENSATION MELT OVER A MELTING SNOW PACK FOR ONE MILE PER HOUR WIND SPEED

FIGURE 4



ELEVATION ADJUSTMENT FACTOR

FIGURE 6

Note:

The relationships shown on this plate are from "Lysimeter Studies of Snowmelt", Research Note 25, based on observations at the Lower Meadow Lysimeter, CSSL, taken during the 1954 snowmelt season.

SNOW INVESTIGATIONS SUMMARY REPORT

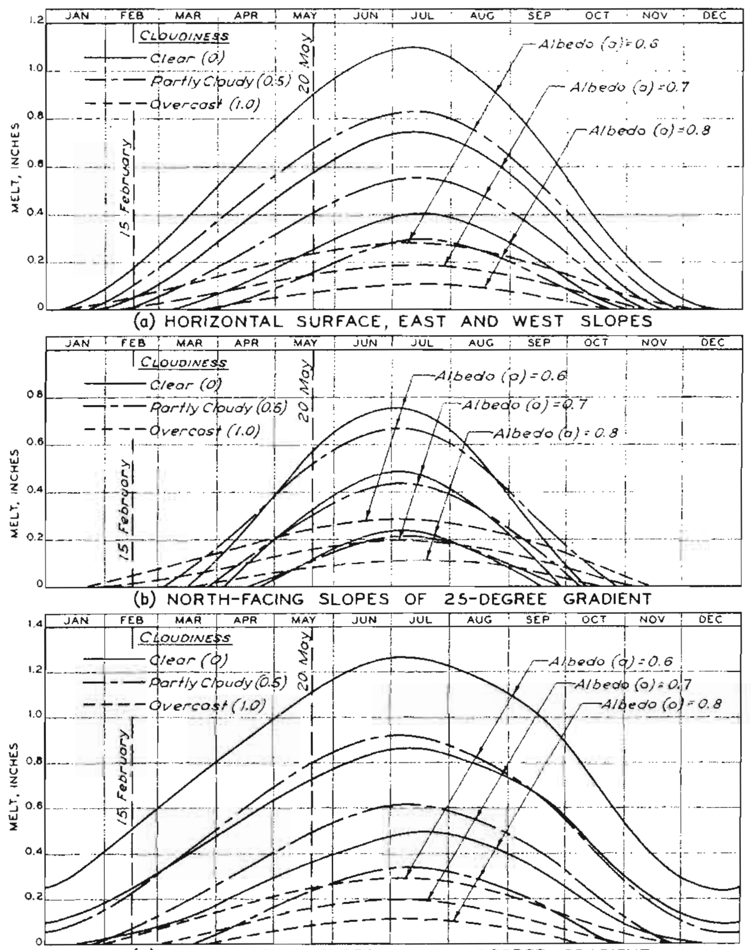
SNOW HYDROLOGY

CONVECTION-CONDENSATION MELT

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

DRAWN B.Z. APPROVED D.M.R.

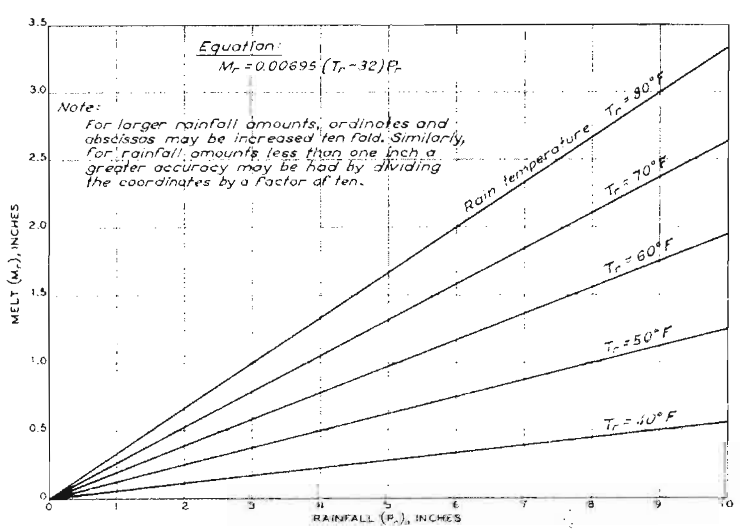
PD-20-25/32



Note: Conditions assumed are for an elevation of 5248 feet MSL and a latitude of 46°30'N. Snow temperature assumed to be 32°F throughout. Air temperature and humidity corresponds to the year-long mean values on the north slope of the Alps.

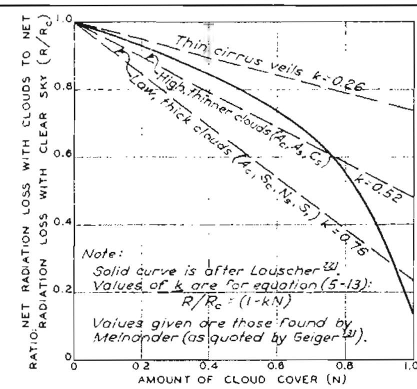
DAILY MELT WATER AMOUNTS DUE TO RADIATION (AFTER HOECK[23])

FIGURE 1



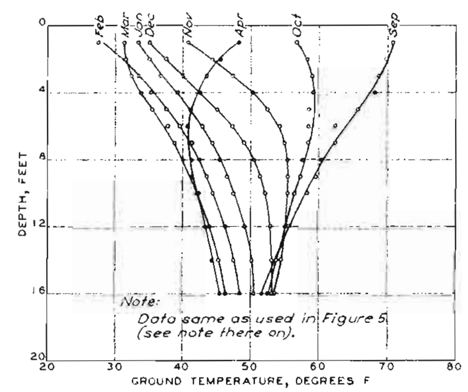
SNOWMELT RESULTING FROM RAIN FALLING ON SNOW

FIGURE 2



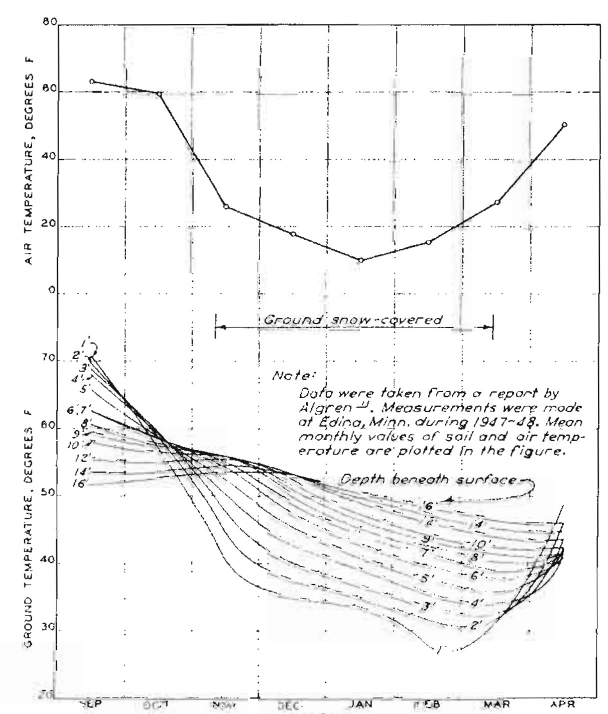
EFFECT OF CLOUD COVER ON NET LONGWAVE RADIATION EXCHANGE

FIGURE 3



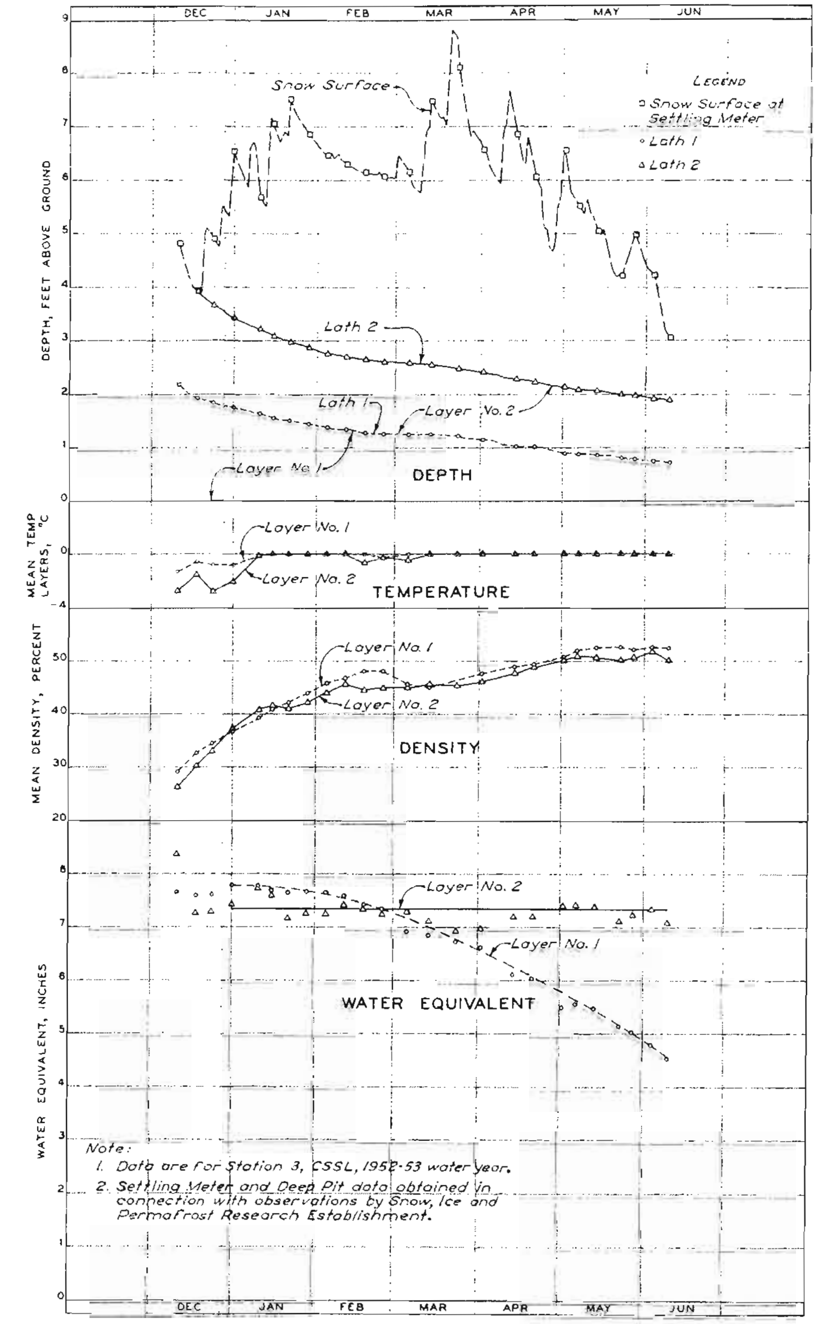
MONTHLY GROUND-TEMPERATURE PROFILES (AFTER ALGREN[11])

FIGURE 4



GROUND TEMPERATURES AT DIFFERENT DEPTHS (AFTER ALGREN[11])

FIGURE 5



CHARACTERISTICS OF LOWER LAYERS OF SNOWPACK FOR DETERMINATION OF MELT FROM GROUND HEAT

FIGURE 6

SNOW INVESTIGATIONS SUMMARY REPORT

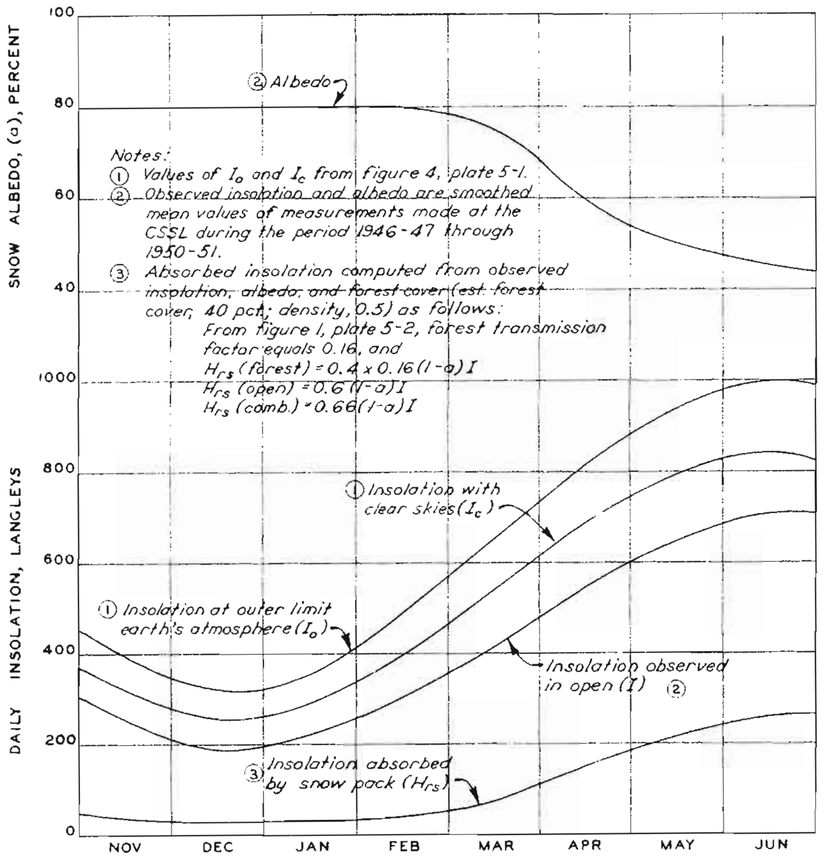
SNOW HYDROLOGY

MISCELLANEOUS SNOWMELT DATA

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION CORPS OF ENGINEERS U. S. ARMY

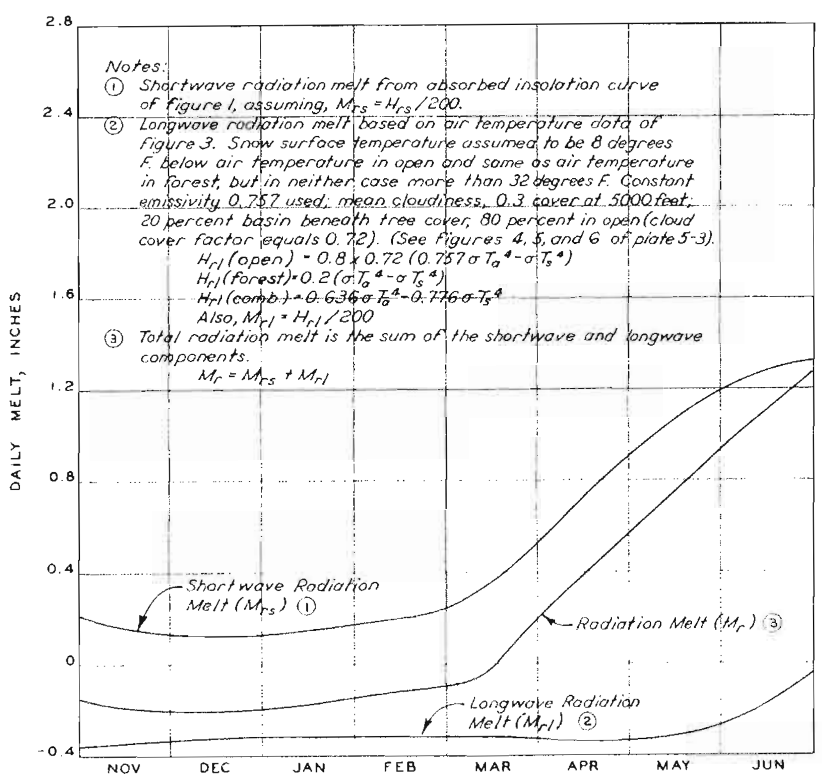
PREPARED C.E.H.	SUBMITTED C.S.H.	TO ACCOMPANY REPORT DATED 30 JUNE 1956
DRAWN S.V.	APPROVED G.M.R.	PD-20-25/33

PLATE 5-6



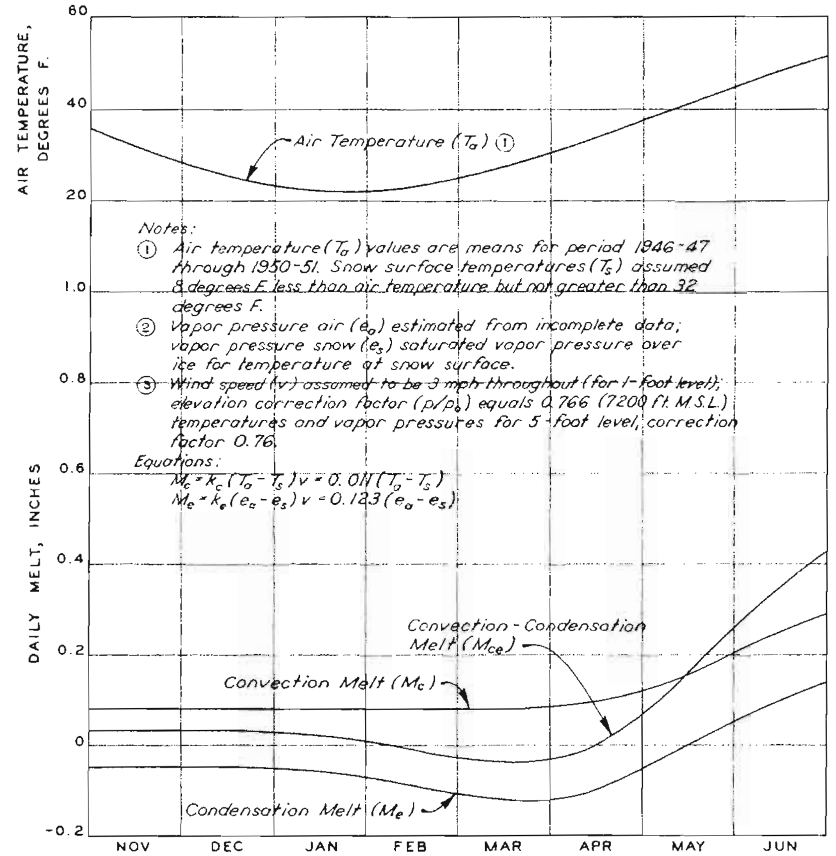
INSOLATION AND ALBEDO

FIGURE 1



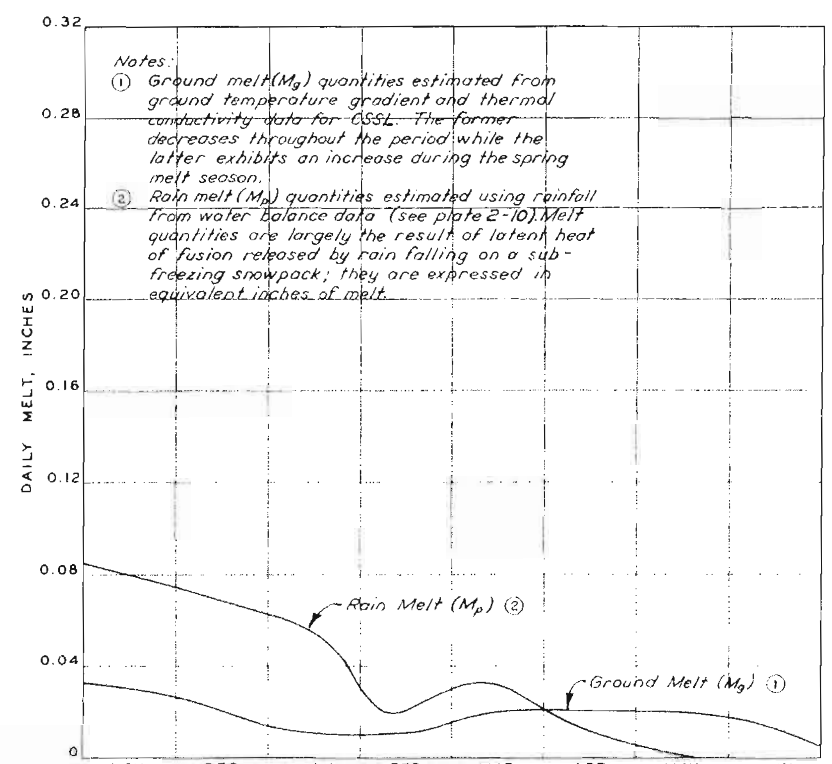
RADIATION MELT

FIGURE 2



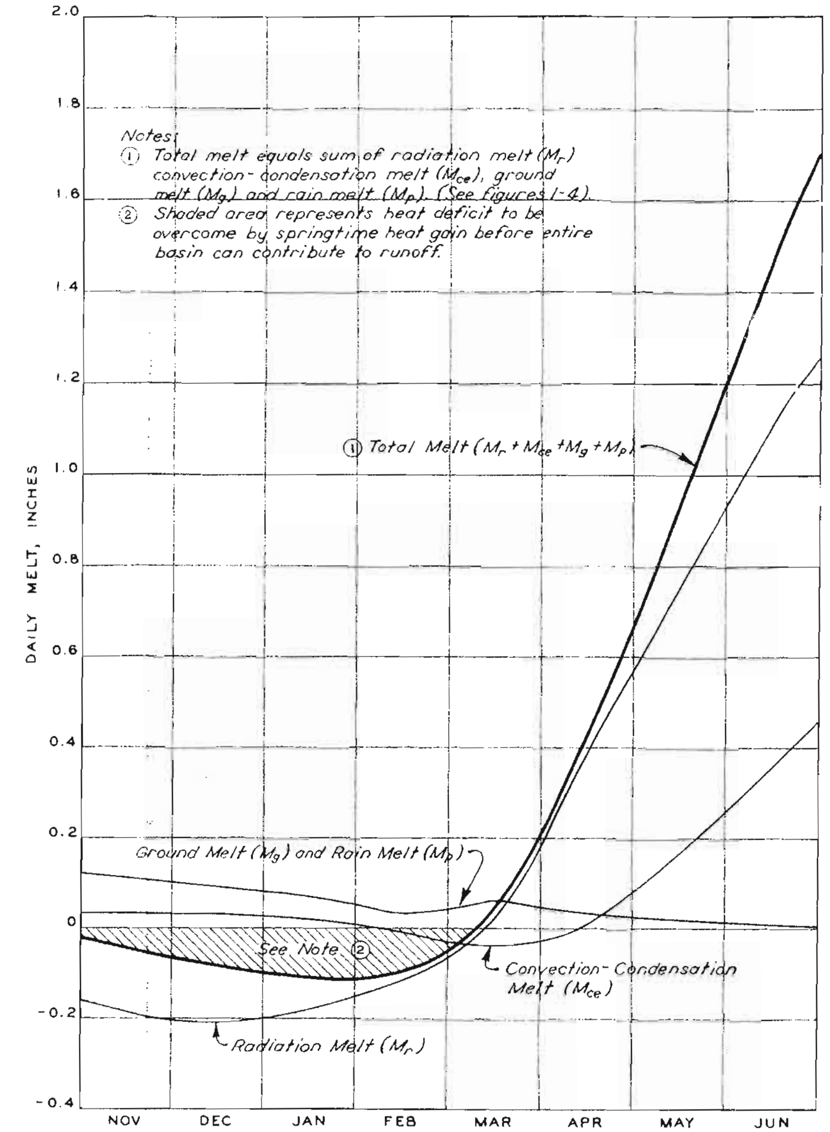
CONVECTION AND CONDENSATION MELT

FIGURE 3



GROUND MELT AND RAIN MELT

FIGURE 4



SNOWMELT SUMMARY

FIGURE 5

SNOW INVESTIGATIONS
SUMMARY REPORT

SNOW HYDROLOGY

SNOWMELT SUMMARY

CENTRAL SIERRA SNOW LABORATORY
MEAN DATA 1946-47 – 1950-51

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

PREPARED CEH	SUBMITTED CEH	TO ACCOMPANY REPORT DATED 30 JUNE 1955
DRAWN HEB	APPROVED DMR	PD-20-25/34

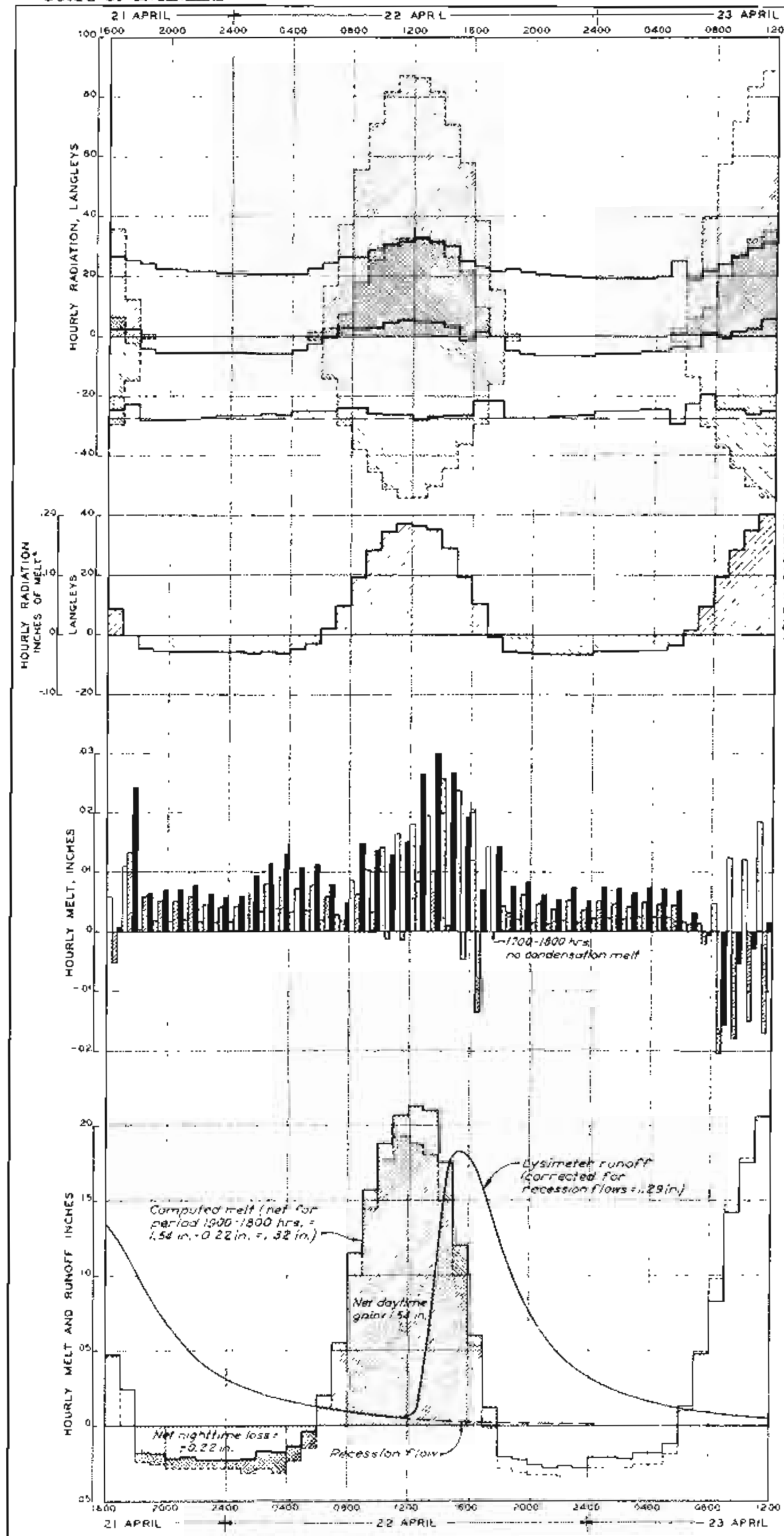


FIGURE 1 — CLEAR DAY

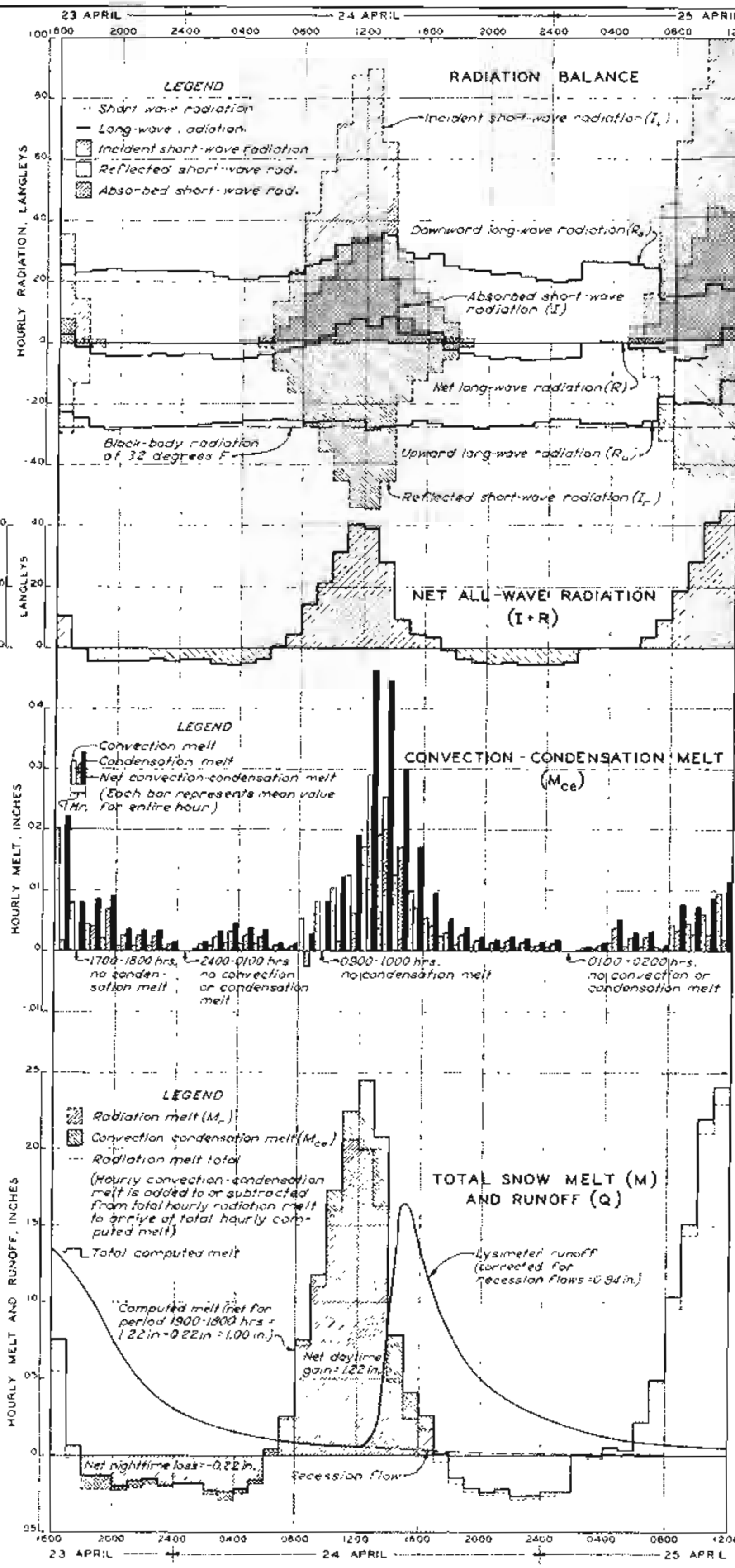


FIGURE 2 — PARTLY CLOUDY DAY

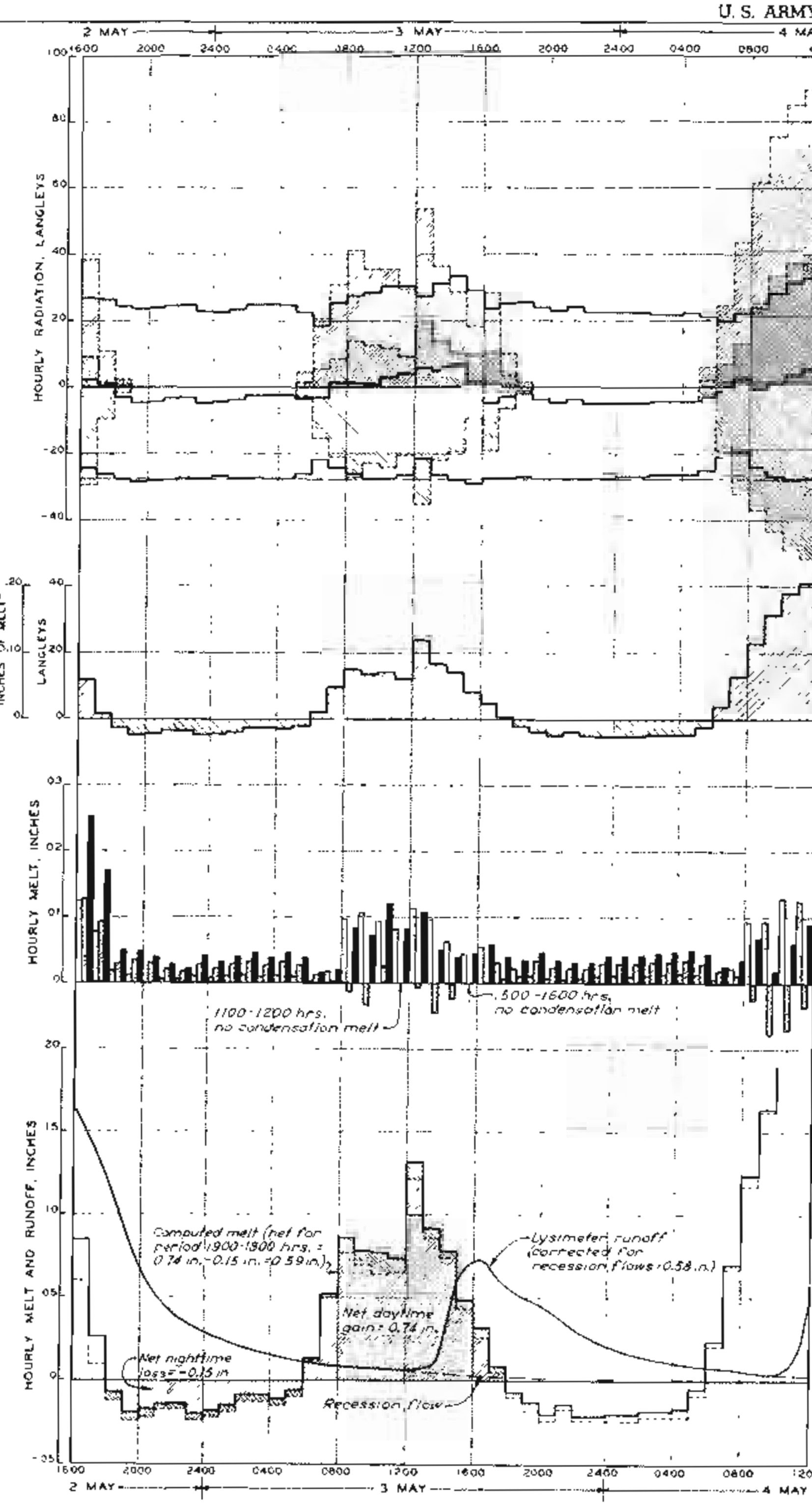


FIGURE 3 — CLOUDY DAY

*Melt computed for ripe snow pack, thermal quality of 97%.

Note:
Observations of unforested site, Lower Meadow Lysimeter, CSSL.

SNOW INVESTIGATIONS
SUMMARY REPORT

SNOW HYDROLOGY

DIURNAL VARIATION OF HEAT SUPPLY AND SNOW MELT

CENTRAL SIERRA SNOW LABORATORY
1954 MELT SEASON

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U.S. ARMY

TO ACCOMPANY REPORT DATED 30 JUNE 1956

PD-20-25/35

PLATE 5-8

CHAPTER 6 - SNOWMELT INDEXES

6-01. INTRODUCTION

6-01.01 General. - Temperature indexes of snowmelt have been widely used to estimate runoff from snowmelt for areas where its contribution to total runoff warranted consideration. Temperature was used because it was generally thought to be the best index of the heat transfer processes associated with snowmelt and because it was (and in many cases will continue to be) the only reliable and regularly available meteorological variable. At the present time, however, a general understanding of the thermal budget of a drainage area and its snowpack enables the hydrologist to select or to establish necessary instrumentation for obtaining other appropriate meteorological variables suited as indexes of snowmelt in a specific area. This can be done in the light of the broader understanding of the physical and thermodynamic aspects of snowmelt learned from work of the Snow Investigations and from other studies as discussed in the previous chapter.

6-01.02 In hydrologic practice an index is either a meteorologic or hydrologic variable whose variations are associated with those of the element it serves to estimate and which is more readily measured than the element itself. An index may be used to represent basin values from point measurements, both in space and in time, where average fixed relationships are known to exist between the measured and basin values. The purposes of indexes are (1) to allow a readily observed measurement to represent hydrologic element or physical process not ordinarily measured, and (2) to allow a single value or group of measured values to represent basin averages in time and space, in order to simplify observational and computational requirements in actual practice. The reliability of indexes depends upon (1) how well the element or physical process is described by the measured variable, (2) the random variability of the measurement, (3) the variability between point values and basin averages, and (4) the representativeness of the measured value. The index relationship may be described either by a coefficient (such as a degree-day factor) or by a formula in the case of more complex linear or curvilinear functions and may be either constant or variable depending upon the variability of associated factors.

6-01.03 Data. - The specific kinds of data required for detailed thermal-budget analysis are rarely available to the hydrologist concerned with project basins. By necessity, he must use the ordinarily available data for determining snowmelt runoff in hydrologic studies. The most generally available data are daily maximum and minimum temperatures, humidity measurements and surface wind speeds; less frequently are found stations having continuous temperature, humidity, and wind records; and few and far between are stations having solar radiation or even duration of sunshine recorders. Occasionally, hourly cloudiness data may be obtained from nearby airway stations.

6-01.04 Scope. - This chapter includes a discussion of index theory, the relative importance of the various indexes with respect to varying degrees of forest cover, and index-loss relationships during clear-weather spring melt. Preliminary analyses with laboratory data demonstrate the technique of index evaluation and degree of accuracy of the regression function that can be achieved with complete information on shortwave and longwave radiation, albedo, snow cover, and hourly data on air and snow-surface temperatures, relative humidity, and wind. Secondary analyses demonstrate the effectiveness of daily temperature, humidity, and wind data when used as convection and condensation indexes. As an illustrative example, snowmelt runoff coefficients are derived for the Boise River at Twin Springs, Idaho. A separate section deals with the general effectiveness of temperature indexes only. Finally, general basin snowmelt equations applicable to rain-on-snow and rain-free situations are presented.

6-02. INDEXES FOR THERMAL-BUDGET COMPONENTS

6-02.01 General. - Known physical laws and established empirical relationships governing the water and heat economy of the snowmelt-runoff process form the basis for the derivation of snowmelt-runoff indexes. The water balance and thermal budget have been discussed in detail in chapters 4 and 5, but will be reviewed briefly as needed for development of the theory, assumptions, derivation, and application of snowmelt indexes. The heat transfer processes which are of primary importance in producing snowmelt are radiation (shortwave and longwave) and convection of sensible heat and moisture in the atmosphere (commonly termed convection and condensation). Heat transfer from rain and underlying ground are secondary and are considered only briefly. Appropriate indexes for each of the primary heat transfer processes will be discussed in the following paragraphs.

6-02.02 There is no universal index for describing accurately the snowmelt-runoff regime for all areas. Index coefficients vary with local conditions of weather, time, snow condition, vegetation and terrain, being limited in applicability to the specific area, time of year, and weather conditions for which they are derived. Index studies carried out in the Snow Investigations have been concerned primarily with conditions of so-called "clear-weather melt." (The term clear-weather as used in this connection refers to periods of no precipitation but places no restriction on the extent or density of cloud cover.) Before proceeding to the discussion of indexes, a brief review of the water balance for clear-weather melt periods will be made.

6-02.03 The general equation for runoff is written as the sum of the terms of the water balance (see chapter 4). The water balance for rain-free periods during active spring melt is:

$$Q = M - L \tag{6-1}$$

where Q is generated runoff, M is snowmelt, and L is loss. The main assumptions which permit index-analysis of equation (6-1) are: that loss is equal to evapotranspiration, and that evapotranspiration is estimated by the same indexes that describe melt. The assumption for the active melt period that all loss is in the form of transpiration requires that all initial losses for conditioning the snowpack for runoff (see chapter 8) and satisfying soil-moisture deficits have been met over the entire basin area. All transitory storage in the soil and underlying rocks is assumed to be accounted for through use of generated runoff values derived by use of average empirical flow recession curves (see Res. Note 19). Daily values of runoff so determined become the dependent variable for the index equations. The condition of no loss to deep percolation on the three laboratory basins is predicated on the water-balance studies presented in chapter 4. The CSSL Lower Meadow lysimeter has an impervious subdeck upon which melt water is collected, thereby eliminating the possibility of loss by percolation into underlying soil or rock.

6-02.04 During periods of local climate, when advection of heat and moisture into the basin is at a minimum, there appears to be a fixed proportion of available energy used for transpiration and melt in forested areas. Reference is made to the discussion on the thermodynamics of evapotranspiration for WBSL contained in the Supplement to Research Note 19 and paragraph 4-05.17 , where it was shown that evapotranspiration water loss under this condition represents about 10 percent of the melt.

6-02.05 Shortwave radiation. - The importance of shortwave radiation in the melt budget for direct evaluation of snowmelt depends, for the most part, on areal extent and density of forest cover. In densely forested areas direct observations of solar radiation are not essential as a snowmelt parameter since 80 to 90 percent of the incident radiation energy is absorbed by the forest. In densely forested areas very little of the shortwave energy passes through the forest canopy to the snow surface, and the energy stored in the forest is released to the snow by longwave radiation, convection, and condensation. In open and partly forested areas, however, shortwave radiation is highly important, having increasing importance with decreasing forest cover. If no observations of shortwave radiation are available, estimates should be made by extrapolation from nearby stations, by means of the atmospheric depletion technique, such as described in Technical Bulletin 5 and Research Note 3, or where duration of sunshine data are available, from the chart adapted from Hammon, Weiss, and Wilson 5/ shown in figure 3, plate 6-1. The last-mentioned is the most recent and apparently most accurate technique. The authors report a coefficient of correlation of 0.97 and a standard error of 36 ly/day for observed versus estimated insolation. Independent application of the chart to Boise, Idaho, data for the months of May and June 1955 yielded the same degree of association (R = 0.97), which is equivalent in the present sample to a coefficient of determination, D, of 0.94, with a standard error of estimate, $s_{y.x}$, of 40 ly/day. The method requires knowledge only of the duration of sunshine and, with date, latitude, and a seasonal correction factor (tabulated on the chart), yields in 8 steps the estimated value of insolation for the day.

6-02.06 Diurnal temperature range is a fair index of solar radiation in heavily forested areas, accounting for 65 percent of its variance in UCSL during the spring of 1947 (see fig. 1, plate 6-1). The degree of association is less for the partly forested CSSL, as shown in fig. 2, plate 6-1, where there was only 50 percent determination for data recorded in the spring of 1954. Unfortunately, the poorer correlation with temperature exists where radiation is important and most needed, namely in partly forested or open areas. Studies of variables affecting maximum temperature at CSSL indicate that within that environment over snow, the daily maximum air temperature increases about 1.3°F per 100 ly increase in incident radiation; decreases about 1.2°F per mph wind speed increase at the one-foot level; and increases about 0.75°F per degree F increase in 10,000-foot level temperature. Maximum temperature* by itself, then is a poor index of solar radiation, its magnitude reflecting variations associated with other processes.

* As measured over snow.

The minimum temperature is also sensitive to surface wind conditions, and increasing wind tends to prevent the minimum temperature from falling as low as on calm nights. Accordingly, wind tends to decrease the spread between maximum and minimum temperature. Also, any change in airmass between times of minimum and maximum temperature would affect the spread between them. In general, diurnal temperature range is not as adequate for estimating shortwave radiation as duration of sunshine data.

6-02.07 Longwave radiation. - Unlike shortwave radiation, longwave radiation is important for all degrees of forest cover. The net effect of longwave exchange on melt does vary with the degree of forest cover and this variation must be taken into account in the design of an index for the longwave component of the thermal budget. As discussed previously, the net effect of longwave exchange in the open is usually negative, constituting the principal mode of heat loss from the snow surface. In densely forested areas, on the other hand, there is practically no longwave loss to the night sky, and the snow exchanges longwave radiation mostly with the forest canopy. Both snow and forest radiate as pure black bodies, and since the temperature of the forest canopy is generally above freezing during periods of active melt, the forest canopy radiates more longwave energy to the snow than the snow is emitting by virtue of its own temperature (which cannot exceed 32°F). Thus, under dense forest the net exchange is nearly always positive in the direction of the snow, and the snow-surface temperature is rarely below freezing during the melt period. With a fixed snow-surface temperature, the radiation exchange between the snow and forest canopy may be described by the temperature of the forest canopy only. For index purposes an approximate linear relation has been adopted. (See fig. 1, plate 6-2). With the assumption that air and forest canopy have the same average temperature, the index for longwave radiation exchange is simply the temperature of the air. The use of air temperature as a longwave index is ideally suited to densely forested areas but is inadequate as a simple index for open areas.

6-02.08 Convection. - The relative importance of convection as a heat transfer process at open sites has been a controversial subject and is difficult to evaluate from direct observation. The eddy conductivity studies appear not to have adequately separated sensible heat transfer from effects of radiation and condensation, and greater importance is consequently assigned to convective transfer than may actually be the case. In regression analyses of the 1954 CSSL lysimeter data, only five percent of the variance of the lysimeter runoff was explained by the convection parameter. In the same study, omission of the convective parameter resulted in a negligible decrease of accountable variance, demonstrating the relative weakness of convective heat transfer from air to snow at an open site. An explanation for this weakness is that the melt contribution due to sensible heat transfer is so small as to be nonsignificant with respect to contribution of the other heat-transfer processes (namely radiation). For forested areas, on the other hand, the convection parameter has greater importance, since it indexes the combined heat-transfer processes of convection and longwave exchange in the forest. Since both longwave exchange and convective transfer are temperature functions, they may be combined as shown in figure 1, plate 6-2. The index used for convective transfer (X_3) is a parameter consisting of the product of temperature difference and wind:

$$X_3 = (T_a - T_b)v \qquad (6\text{-}2a)$$

where $\underline{T_a}$ is air temperature, $\underline{T_b}$ is snow surface temperature or temperature base, and $\underline{v}$ is wind travel. Where hourly data are available, best results are obtained with a parameter for the day, computed from the sum of the hourly parameters for the twenty-four hour period from sunset to sunset. When hourly temperatures and winds are not available, as is generally the case in practice, daily maximum or daily mean temperature, and a daily index of wind may be substituted with some loss of accuracy. In this case, the base temperature, T_b is evaluated statistically by methods discussed in paragraphs 6-03.19 to 6-03.21, and the function is used in equation 6-2a as a product of wind and air temperature. Investigations have shown that the general effectiveness of maximum and mean temperatures varies from area to area and even from season to season within a single area. Therefore, either maximum temperature or mean temperature, 1/2 (max + min), may be used for the daily index. Investigations of a daily wind index have included total 24-hour wind travel (sunset to sunset) and 12-hour daytime wind travel (sunrise to sunset). For CSSL, daytime wind travel was found superior to the 24-hour wind, and thus was also used in the UCSL studies. There are other indexes of wind which may be useful, but which have not been evaluated as yet, such as 24-hour winds for some period other than sunset to sunset, (say midnight to midnight), instantaneous wind speed at a specific time of day, etc. The temperature base for the daily convection parameter may be computed from the regression analysis as will be shown later in paragraph 6-03.20.

6-02.09 Condensation. - As with convection, the condensation parameter for the open site is relatively unimportant compared with radiation. It was found in a study of the 1954 CSSL lysimeter data, that less than 3 percent of the lysimeter runoff variance was lost to the accountable variance when the condensation parameter was omitted. (The net decrease in accountable variance with omission of both the convection and condensation parameters was still less than 3 percent.) The importance of the condensation parameter, however, increases with the presence of forest cover. In CSSL, where effective forest cover is approximately 30 percent of the basin area, nearly 40 percent of the runoff variance was lost to the accountable variance when the condensation parameter was omitted. The form of the condensation parameter (X_4) is:

$$(X_4) = (e_a - e_s)\ v \tag{6-2b}$$

where e_a is vapor pressure of the air, e_s is vapor pressure of the snow surface or a derived vapor pressure base, and v is wind travel for the time interval represented by mean e_a and e_s. Hourly values of the variables constituting the condensation parameter yield better results than a daily index. However, the only daily condensation indexes tested on laboratory data have been (1) with WBSL data: maximum vapor pressure for the day, (2) with CSSL data: vapor pressure at 1800, the most consistent hour of daily CSSL maximums, and (3) with UCSL data: vapor pressure at 1400, the most consistent hour of daily UCSL maximums. As in the case of the convection parameter, bases are derived statistically when daily values are used. There are also other possible condensation indexes that might prove worthy of investigation, such as maximum or mean dewpoint temperature, etc.

6-02.10 Summary. - The following table summarizes the relative importance and proper selection of thermal-budget indexes of clear-weather melt, as determined from statistical studies of daily clear-weather snowmelt runoff at the snow laboratories, for varying degrees of effective forest cover:

Thermal Budget Component	Forest environment			
	Heavily Forested Area, WBSL	Forested Area, UCSL	Partly Forested Area, CSSL	Open Area, Lysimeter
Effective forest cover	90%	85%	30%	0%
Absorbed shortwave radiation	N - 10/	N - 10/	S - 2/	S - 2/
Longwave radiation exchange in open	N - 10/	N - 10/	S - 3/	S - 3/
Longwave radiation exchange in forest	S - 1/	S - 1/	N - 1/	N - 10/
Convective heat exchange from atmosphere	S - 4/	S - 5/	S - 5/	N - 6/
Heat of condensation or evaporation by moisture exchange between atmosphere and snow surface	S - 7/	S - 8/	S - 8/	N - 9/

S = Significant clear-weather melt index
N = Non-significant clear-weather melt index

1/ Adequately indexed by convection parameter
2/ Observe directly or estimate by index relationship of solar radiation and snow albedo
3/ Observe directly or estimate by minimum temperature and airmass-temperature index
4/ Daily or hourly convection parameter, excluding wind
5/ Daily or hourly convection parameter, including wind
6/ Daily or hourly convection parameter, including wind; may be omitted in active shortwave radiation melt periods
7/ Daily or hourly condensation parameter, excluding wind
8/ Daily or hourly condensation parameter, including wind
9/ Daily or hourly condensation parameter, including wind; may be omitted during periods of active shortwave radiation melt
10/ Non-existent or negligible heat-transfer process

The most striking deficiency in the above table is the lack of an index for longwave radiation exchange in the open. Although air temperature has been found to be a fair index of downward longwave radiation, the variability of the snow-surface temperature at an open site does not permit air temperature alone to act as an acceptable index of longwave exchange between snow and sky. It was found, however, in preliminary studies on the partly forested CSSL, that measured longwave radiation in the open was a significant variable in the estimation of Castle Creek runoff. It was further found that substitution of incident shortwave radiation for net longwave exchange in the open gave an equal level of determination from estimation of Castle Creek runoff. For runoff from an unforested site, however, incident shortwave radiation was found nonsignificant as a substitute for longwave exchange in the open. For open and partly forested areas, longwave exchange in the open is an essential part of the thermal budget. If longwave exchange in the open is not measured, as will most often be the case, it should be estimated. An empirical method for estimating longwave exchange in the open in the Boise River basin is presented in figure 1, plate 6-2.

6-02.11 Figure 3, plate 6-2 illustrates graphically the relative importance of radiation, convection, and condensation parameters in explaining the variance in daily runoff values, at Mann Creek, WBSL; Skyland Creek, UCSL; Castle Creek, CSSL; and the Lower Meadow lysimeter at CSSL, which have a relative effective forest cover of 90,85, 30, and 0 percent respectively. Effective forest cover is defined as the percent of the basin effectively shielded from direct radiation to the sky (both longwave and shortwave), based on an average in time and space for the active melt period. In the case of UCSL, the radiation term is absorbed shortwave and explains practically none of the variance in runoff, while longwave radiation exchange is adequately indexed by the convection and condensation parameters. The radiation term at CSSL for both Castle Creek and the lysimeter is allwave net exchange, as measured by the non-selective radiometer. The increasing importance of an adequate radiation index for decreasing extent of forest cover is shown by this diagram.

6-03. EVALUATION OF INDEXES FOR CLEAR-WEATHER MELT

6-03.01 General. - The preceding section defined the components of basin heat exchange in terms of thermal budget indexes for clear-weather periods. In this section there are described the analytical techniques used to evaluate the index coefficients and their relative strengths for different areas, as determined from laboratory data.

6-03.02 Analytical procedure. - The general form of the snowmelt runoff equation in terms of heat indexes is

$$Y = b_1X_1 + b_2X_2 + b_3X_3 + \ldots\ldots + b_iX_i + \ldots + c \qquad (6\text{-}3)$$

where $\underline{Y}$ is generated snowmelt runoff, the $\underline{X_i}$'s are the independent variables (indexes), and $\underline{b_i}$'s are the associated coefficients found by regression, and $\underline{c}$ is the regression constant. The symbols and units associated with the indexes used in equation 6-3 are given in the following table:

Variable	Description	Symbol	Unit	Index for
Y	Daily generated volume of snowmelt runoff, depth over snow-covered area	Q	in.	
X_1	Daily total shortwave radiation absorbed by snow in the open	I_{net}	ly	Shortwave
X_2	Longwave radiation loss in the open or daily total incident shortwave radiation in the open	R_{net} or I_i	ly	Longwave exchange in open
X_3	Convection parameter = daily maximum temperature multiplied by 12-hour wind travel, 0600-1800 daily	$T_{max}v_{12}$	°Fmi	Convection and longwave exchange in forest
X_4	Condensation parameter = vapor pressure at a specified time, $\underline{t}$, multiplied by 12-hr wind travel (0600-1800 daily)(t = 1800 for CSSL; t = 1400 for UCSL)	$T_{DA}v_{12}$	°Fmi	Condensation
X_5	12-hr wind travel, 0600-1800 daily	v_{12}	mi	Temp. & v.p. bases
X_6	Daily maximum temperature	T_{max}	°F	Convection and LW exchange in forest
X_7	Vapor pressure at time, $\underline{t}$, or max. v.p. (t=1800 for CSSL; t=1000 for UCSL; daily max. vapor pressure WBSL)	e_t or e_{max}	mb	Condensation

(Continued)

Variable	Description
X_i	General term for any of the above variables with appropriate subscript ($i = 1 - 7$)
b_i	Coefficient of any of the above variables to give runoff; determined by regression analysis
c	A constant whose value is the magnitude of runoff given by any regression equation when all $X_i = 0$. (Also: a constant whose value is the magnitude required for the regression equation to give the mean value of the dependent variable when means are substituted for all X_i's.)

By following a planned sequence of dropping, adding or substituting different hydrometeorologic variables and repeating the regression analysis with different combinations, a set of regression equations is obtained with different regression coefficients for any one variable. The change in magnitude of regression coefficients for the same hydrometeorologic variable from one regression equation to another shows how these coefficients compensate for the presence or absence of one or more of the other variables. Differences between coefficients of determination for the various regression equations measure the relative effectiveness of the variables which have been dropped, added, or substituted. Thus, the analytical schedule set up in the following table was designed to provide a systematic procedure for answering specific questions as to relative effectiveness of the hydrometeorologic variables used:

Runoff Function	X_1	X_2	X_3	X_4	X_5	X_6	X_7
(1)	*	*	*	*	*		
(2)		*	*	*	*		
(3)	*		*	*	*		
(4)			*	*	*		
(5)	*	*				*	*
(6)		*				*	*
(7)		*				*	
(8)						*	*
(9)	*	*					
(10)	*						
(11)		*					
(12)			*		*		
(13)				*	*		
(14)						*	
(15)							*

(See table on page 202 for definition of variables X_1 through X_7.)

Questions relative to each function are listed below, with question numbers corresponding to function numbers listed above:

(1) What is the accuracy of runoff estimated from the important variables: radiation, convection, and condensation?

(2)(3)(4) What is the effect on the accuracy if radiation is omitted, and runoff is estimated only from convection and condensation parameters? (for use where radiation data are unavailable.) Does incident shortwave radiation add anything to the accuracy?

(5) What is the effect on the accuracy of (1) if wind is omitted from the convection and condensation parameters? (For use where wind data are unavailable.) How important is wind speed in the relationship?

(6)(7)(8) What is the effect on (5) if radiation is omitted and runoff is estimated from temperature and vapor pressure only?

(9) How well can runoff be estimated from shortwave radiation only, omitting the convection and condensation parameters?

(10)-(15) How well does each of the variables by itself estimate runoff?

Not all functions listed in the above table were evaluated for each basin or for all years for any basin. It became readily apparent, after analysis of the 1949 data for UCSL that, as expected for a densely forested area, solar radiation was unimportant as a runoff index in this basin. Consequently, all of the functions involving absorbed shortwave radiation were omitted from the analysis of the 1950 data, since albedo information for this year was incomplete as well. Incident radiation was carried in a few of the 1950 regression functions with the same lack of effectiveness demonstrated by the 1949 data. Regression coefficients and measures of accuracy are summarized in tables 6-1 to 6-4 and discussed in subsequent paragraphs.

6-03.03 Daily indexes of snowmelt. - Analyses of the 1954 lysimeter and Castle Creek runoff data have demonstrated that best results are obtained when firsthand observations of the important melt index variables are available and when hourly data are used in the computation of the convection and condensation parameters. The main consideration in the selection of snowmelt-runoff indexes is to reduce as much as possible variations for representation of basin amounts, from measured values, both in space and time. When considering location of index stations, proximity to the melt zone and representativeness of the station must be balanced against quality of data. At the snow laboratories this was no problem because measurements of meteorologic variables were made within the drainage area of the laboratory, and index stations were selected on the basis of known quality of record. However, to evaluate the effect of time variation, convection and condensation were computed first by hourly parameters and second by daily parameters. In the case of clear-weather melt in the open, as measured by the lysimeters, errors in estimating convection and condensation by daily parameters are difficult to detect since radiation is the prime variable for an open site. For Castle Creek, on the other hand, the over-all determination is sensitive to the kind of parameter used for convection and condensation (hourly or daily) because these heat-transfer processes represent a significant portion of the total melt. The following table lists coefficients of determination obtained from analysis of 1954 data.

Convection and condensation parameter	Total determination, D	
	Lower Meadow Lysimeter, CSSL	Castle Creek, CSSL
Hourly values	.97	.94
Daily values	.95	.84

For the densely forested WBSL, where both temperature and vapor-pressure parameters (without wind) are used to represent the melt process, maximum daily, mean daily, and mean daytime values are about equally good in estimating snowmelt runoff. This reflects the relatively small variance of winds in a heavily forested area. In general, it may be stated that daily parameters may be used for estimating snowmelt runoff, but with some loss of reliability for partly forested areas, or for open areas when melt by shortwave radiation is small.

6-03.04 Regression analysis of laboratory data. - General relative effectiveness of various parameters was discussed in paragraph 6-02.10. It is not desirable herein to make comparisons of all combinations of variables and areas. Instead only those comparisons will be made which emphasize the role of a particular index in the area for which it is derived. First is presented a general discussion of the effectiveness of melt parameters for the lysimeter and for each of the laboratories, to illustrate variation of the relative importance of melt indexes with varying amounts of forest. This is followed by a more detailed discussion of the magnitude of the regression coefficients for the radiation and condensation-convection parameters. Also, the statistical derivation of temperature and vapor-pressure bases is discussed.

6-03.05 Unforested site (Lower Meadow lysimeter, CSSL, 1954). - Statements concerning the effectiveness of the various parameters for an unforested site are based on regression analyses of the 1954 lysimeter data (Research Note 25). Regression coefficients, constants, and measures of accuracy for this site are presented in table 6-1. The following information for various functions, extracted from table 6-1, provides comparative data on coefficients of determination from which it is possible to assess the relative importance of the melt parameters:

Eq. No. Table (6-1)	Independent Variables 1954 Lower Meadow Lysimeter, CSSL	Coeff. of Determination
(1)	Allwave radiation, convection, condensation	.97
(5)	Convection, condensation (omits allwave radiation)	.46
(3)	Allwave radiation (omits convection and condensation)	.94
(4)	Shortwave radiation only	.65
(9)	Maximum air temperature only	.23
(10)	Mean daily air temperature only	.08

Comparing equation (1) with equations (3) and (5), it can be readily seen that omission of allwave radiation (5) results in loss of more than half of the explained variance of equation (1), whereas omission of convection and condensation (3) results in a negligible loss. This comparison clearly demonstrates that allwave radiation is the controlling variable for the melt regime at an open site. It can be further seen by comparing (3) and (4) that approximately 30 percent of the accountable variance explained by allwave radiation is lost when the longwave component of allwave radiation is omitted. For the best estimates of snowmelt runoff from an open area, then, it is imperative to have good estimates for both shortwave and longwave radiation. When allwave radiation measurements are not available, estimates of the shortwave and longwave components must be made. Shortwave radiation may be indexed by duration-of-sunshine data as explained earlier in paragraph 6-02.05, and appropriate albedo estimates may be applied to give estimates of absorbed shortwave radiation exchange. Minimum air temperatures in conjunction with an upper-air temperature index (as will be described later in connection with the Boise River melt indexes) may be used to provide estimates of longwave radiation exchange. Convection and condensation parameters without radiation are inadequate for estimating clear-weather melt in the open, while air temperature* alone is totally inadequate, as shown by the poor determination where maximum or mean daily air temperatures are used.

6-03.06 Partly forested area (Castle Creek, CSSL). - Conclusions concerning the relative importance of thermal-budget indexes in the estimation of snowmelt runoff from a partly forested area are based on regression analyses of clear-weather data from three years of record at CSSL. Results of these analyses are presented in table 6-2. Equation 1 on that table, derived from data for 1954 utilizing hourly convection and condensation parameters and measured longwave and shortwave radiation exchange in the open, is presented as a basis of comparison for other equations. Generalizations as to the effectiveness of individual melt parameters for this type area (30% effective forest cover) can be made by comparisons of coefficients of determination for each equation. For areas of this type, adequate representation of melt processes for both the open and forested areas should be included in the melt indexes, and the problem is therefore more complex than for either open or completely forested areas. For Castle Creek, the radiation melt indexes represent mainly the melt in the open, while the convection-condensation parameters represent melt both in the forest and in the open. There are interactions between the two which would not necessarily be accounted for in a

* As measured over snow

strictly rational approach but are contained in the regression coefficients in the statistical approach. Deductions made from regression analyses are therefore mainly limited to showing the effectiveness of various parameters in estimating melt for the periods under analysis. The magnitudes of regression coefficients for the individual parameters can be rationalized only for those equations containing indexes for all thermal-budget components with hourly parameters for convection and condensation. Also, it is shown that the relative effectiveness of individual parameters varies with the melt conditions for each year. Even with these limitations, however, some interpretations concerning relative importance of the variables may be made.

6-03.07 When all indexes of thermal-budget components for Castle Creek are used, the coefficient of determination is 93 percent for two of the three years, and 84 percent for the third, (see equations 1, 3, and 5 in table 6-2). This indicates that the complete thermal-budget index accounts for all but about 10 percent of the variance of snowmelt runoff from this partly-forested area. The importance of radiation in the snowmelt runoff equation is shown by a loss of determination ranging from about 10 to 45 percent where allwave radiation in the open is omitted. Omission of the open-area longwave index represents a loss of determination of melt ranging from 3 to 10 percent. Convection and condensation parameters by themselves show a total determination of between 47 and 83 percent, while shortwave radiation by itself accounts for 22 to 71 percent of the variance of daily melt in individual years. Daily temperature and vapor-pressure parameters without wind account for between 14 and 67 percent of the variance. Maximum temperature alone is an erratic melt index for Castle Creek; it has practically no correlation with daily runoff during 1954, but it explains about 50 percent of the runoff variance in the 1952 and 1951 clear-weather periods of the present study. General effectiveness of temperature alone is discussed in section 6-06.

6-03.08 Forested area (Skyland Creek, UCSL). - The relative importance of thermal-budget indexes of melt for a forested area were tested on Skyland Creek, UCSL, for clear-weather periods in 1949 and 1950. Lodgepole pine forest with trees ranging to 50 feet in height covers about 90 percent of the area, and the average canopy density of the forested area is estimated to be 80 percent, with a resultant effective forest cover of 85 percent. Table 6-3 lists the results of the regression analyses of snowmelt runoff for Skyland Creek. The following tabulation summarizes the relative effectiveness of individual parameters for 1949, the year for which a complete analysis was performed:

Equation No. (Table 6-3)	Independent variables (Skyland Creek, UCSL, 1949)	Coefficient of Determination
2	Abs. shortwave radiation, convection, condensation	0.89
4	Convection, condensation	0.90
7	Incident and absorbed shortwave radiation	0.29
10	Convection	0.73
12	Condensation	0.48
6	Maximum temperature, 1400 vapor pressure	0.60
14	Maximum temperature	0.54

Note that convection-condensation parameters excluding shortwave radiation in equation 4 adequately indexed the melt, since they include the effect of longwave radiation exchange in the forest. Apparently, in equation 2, neither absorbed nor incident shortwave radiation contributed significantly to the determination of melt. Maximum temperature and 1400 vapor pressure without wind, in equation 6, shows the damaging effect of omitting wind from the convection and condensation parameters, which in 1949, caused the determination to drop from 90 to 60 percent. In 1950, however, the decrease in determination was less pronounced when wind was omitted from the melt equation (see table 6-3). Maximum temperature alone provided a fair index of melt in both years, the determination ranging from 54 to 62 percent. In general, it is seen that for this type of area, accurate determination of melt quantities requires parameters which include the effect of wind, to express convection and condensation, but that maximum temperature alone may be used to provide a fair estimate of daily melt if it is impractical to obtain the necessary data on wind and humidity for a more complete evaluation.

6-03.09 <u>Heavily forested area (Mann Creek, WBSL)</u>. - Analyses on four separate periods of clear-weather snowmelt runoff data for Mann Creek, WBSL, provided the basis for determining the relative effectiveness of melt parameters for a heavily forested area. Here, the forest consists of large trees (predominantly Douglas fir whose heights generally range from 150 to 200 feet) and has an estimated effective cover of 90 percent (see plate 2-9). No measurements of solar radiation were obtained at WBSL because of its expected minor importance in the direct melt process at this location, but observations of solar radiation obtained at Medford, Oregon (approximately 100 miles south of the laboratory) were tested in the melt equation and were found to be non-significant in the estimation of runoff. Longwave radiation between the snow and forest, convection, and condensation are the three primary processes of direct heat transfer to the snowpack. Parameters involving air temperature and vapor pressure thus adequately index snowmelt at WBSL. Wind was not a regularly measured variable at WBSL, but an index of wind by use of upper-air data did not provide a significant improvement in the determination of melt. The following tabulation summarizes the effectiveness of the melt parameters shown in table 6-4; the mean determination for the four periods was obtained by weighting the values for the separate periods by the number of observations in each sample.

Equation Nos. (Table 6-4)	Mann Creek WBSL Independent variables	Mean coefficients of determination
1-4	Maximum daily temperature and vapor pressure	0.79
5-8	Mean daily temperature and vapor pressure	0.81
9-12	Maximum daily temperature	0.71
13-16	Mean daily temperature	0.61

The above summary shows that either maximum or mean daily combined temperature and vapor pressure parameters are about equally effective in estimating snowmelt runoff at WBSL, but that there is a loss of 10 to 20 percent determination by omission of the vapor pressure term. It also shows that for estimating snowmelt at WBSL without vapor pressure, maximum temperature provides a better estimate than mean temperature.

6-03.10 Summary of laboratory melt indexes. - The following tabulation presents a comparative summary of the coefficients of determination showing the effectiveness of the primary melt parameters used in the study, for each of the four environments studied:

Summary of Effectiveness of Melt Indexes

Laboratory and environment	Snowmelt parameters used in multiple linear regressions								Coefficients of Determination			
	Abs. Shortwave Radiation in open	Net Longwave Radiation in open	Convection and condensation: Wind incl.: Conv.	Convection and condensation: Wind incl.: Cond.	Convection and condensation: Wind excl.: Conv.	Convection and condensation: Wind excl.: Cond.	Air Temp. only: Maximum	Air Temp. only: Mean				
									(1954)			
Open Site:	x	x	x	x					.97			
Lower	x	x							.95			
Meadow	x								.65			
lysimeter,			x	x					.46			
CSSL							x		.23			
								x	.08			
									(1954)	(1952)	(1951)	
Partly	x	x	x	x					.93			
forested	x		x	x					.90	.65	.90	
area:			x	x					.47	.59	.83	
Castle					x	x			.14	.58	.67	
Creek, CSSL	x								.51	.22	.71	
							x		.01	.54	.58	
									(1949)	(1950)		
Forested	x		x	x					.89			
area:			x	x					.90	.81		
Skyland					x	x			.60	.77		
Creek, UCSL	x								.23			
							x		.54	.62		
									(1951)	(1950)	(1949)	(1949)
Heavily					x	x			.80	.77	.93	.80
forested							x		.70	.72	.73	.70
area: Mann								x	.65	.61	.64	.49
Creek, WBSL									April	May	May	April

6-03.11 Radiation coefficients for unforested sites. - Greater confidence in statistical techniques is inspired when regression coefficients can be shown to agree favorably with "expected values" estimated from rational considerations only. Of the variables employed in these analyses, radiation lends itself most readily to the derivation of rational coefficients. For example, the rational radiation melt coefficient would be obtained from an expression involving the latent heat of fusion of ice, (80 cal/gm), and the free-water content of the snow (f_p). Thus a snowpack having a free-water content of 3% (as in Research Note 25) would require 80 x 0.97 x 2.54 = 197 langleys per inch of melt. The expected melt coefficient for inches of melt per langley is the reciprocal of 197, or 0.00508. Regression analysis of the 1954 lysimeter data yielded a radiation melt coefficient of 0.00541, which is only 7 percent larger than the expected rational value. It should be noted here also that the regression value is subject to systematic errors in the data traceable to calibration of the radiation instruments, calibration of the lysimeter orifice, and the assumed lysimeter area. The rational value depends on the assigned free-water content, which itself is subject to experimental error. To illustrate how relatively small errors may affect the results, it can be demonstrated that systematic errors of approximately two percent in all variables could have made rational and regression coefficients identical. Also, the rational value was derived on the basic assumption that the lysimeter snow surface is horizontal; approximately 4 percent increase in the rational factor would be expected due to the slope of the lysimeter. Errors of this order of magnitude are reasonable in the light of adjustments made in the original data and clearly demonstrate the remarkably close agreement between the rational and regression coefficients. Since the calibration of the instruments may vary from year to year, it should also be borne in mind that corresponding variations may be expected in derived regression coefficients.

6-03.12 The 1954 lysimeter data are again cited to demonstrate how incident shortwave radiation acts as a substitute for the longwave loss and why its coefficient is negative. The following are values of coefficients derived for two functions, one containing allwave radiation, and convection-and-condensation parameters and the other containing absorbed shortwave radiation, incident shortwave radiation, and convection-and-condensation parameters. The only difference is that in the second function incident shortwave radiation has been substituted for the longwave loss component of allwave radiation in the first.

(1) Regression coefficient, allwave radiation 0.00541

(2) Regression coefficient, absorbed shortwave radiation (I_{net}) 0.00669
Regression coefficient, incident shortwave radiation (I_i) -0.00128

At first glance, the coefficients of the second function do not appear to be compatible with the first. However, the two terms comprising allwave radiation melt (M_G) in the second function:

$$M_G = 0.00669\ I_{net} - 0.00128\ I_i \qquad (6\text{-}4a)$$

can be reduced algebraically to

$$M_G = 0.00541\ (I_{net} - 0.24\ a\ I_i) \qquad (6\text{-}4b)$$

where the quantity in parentheses is the estimate of allwave radiation from absorbed and incident shortwave, for a snow surface albedo, a. The over-all coefficient of determination for the second function (2) is less than that for the first, due to the imperfect relation between incident shortwave radiation and longwave radiation loss. However, the resulting regression coefficients obtained from either function are identical. The above computation thus explains the apparently high coefficient obtained for absorbed shortwave radiation and accounts for the minus sign associated with the coefficient for incident radiation, and further demonstrates the validity of using absorbed and incident shortwave radiation as indexes of allwave melt. However, as will be shown later, the use of incident shortwave radiation as a separate index produces undesirable effects which may offset the gain in determination.

6-03.13 Radiation coefficients for partly forested areas. - The previous discussion has been concerned with radiation melt regression coefficients for an unforested site. The next problem to be considered will be examination of the behavior of the radiation melt coefficient when the regression is performed for runoff from a partly forested area, namely Castle Creek basin, with 30% effective forest cover. Regression equation 1, table 6-2, shows that the absorbed shortwave radiation melt coefficient was 0.00312 for the entire basin. This value appears reasonable when considering the amount of bare area in the open during the melt period.

6-03.14 Selection of radiation parameter. - Referring again to the summary of regression coefficients listed in tables 6-1 and 6-2, and rewriting the radiation coefficients for the combined radiation parameter as described in 6-03.12, comparisons may be made of functions containing absorbed and incident shortwave radiation and of functions containing only absorbed shortwave radiation. Convection and condensation parameters (adjusted for temperature, humidity and wind at one foot) are included to illustrate the relative effect of the presence or absence of incident shortwave radiation on them:

CENTRAL SIERRA SNOW LABORATORY

REGRESSION COEFFICIENTS

		Radiation		Conv. $(T_a-T_s)v$	$\frac{\text{Cond.}}{\text{Conv.}}$ Ratio	Cond. $(e_a-e_s)v$	Coeff. of Det. D
		$I_{net} - kaI_i$	k				
Lysimeter	1954	0.0054	0.24	0.00044	6.82	0.0030	0.92
Castle Creek	1954	0.0036	0.28	0.00038	6.84	0.0026	0.94
	1952	0.0042	0.63	0.00023	8.26	0.0019	0.75
	1951	0.0031	0.49	0.00018	9.44	0.0017	0.93
		I_{net}					
Lysimeter	1954	0.0047		0.00040	8.00	0.0032	0.89
Castle Creek	1954	0.0028		0.00035	8.00	0.0028	0.90
	1952	0.0029		0.00029	8.28	0.0024	0.65
	1951	0.0022		0.00030	8.33	0.0025	0.90

It can be readily seen that the convection and condensation coefficients are in closer agreement among years for the function which omits incident shortwave radiation. A reason for the more extreme variation among convection and condensation coefficients, when shortwave incident is substituted for longwave loss in the open, is that incident shortwave is an index of longwave radiation in the open. By itself, shortwave incident explains only about twelve percent of the variance of the longwave loss, but with convection and condensation, the explained variance is about 54%. Thus, the convection and condensation parameters take on the additional function of contributing to the estimation of longwave exchange in addition to their normal job of indexing convection and condensation melt. The net effect of this double duty is felt in the magnitudes of the convection and condensation coefficients, which no longer bear the same relation to one another as in the case of the complete thermal budget index. Due to the relatively weak determination of longwave loss by incident shortwave, convection and condensation, there is a large sampling variation in the regression coefficients for these variables when they are used as a substitute for longwave loss. This accounts for the erratic variation of "k" and the unstable ratio of the convection and condensation coefficients in the upper portion of the above table. In view of these effects, it is doubtful that the coefficients of any one of the years for the function including incident shortwave radiation would be satisfactory for other years. Consequently, it appears better to sacrifice some of the degree of determination in individual cases for the better over-all agreement among the other coefficients by omission of shortwave incident radiation. It is emphasized that such comparisons are possible only because of the precision afforded by the use of hourly data in the convection and condensation parameters.

6-03.15 Radiation coefficients for densely forested areas. - As mentioned earlier in the discussion of the regression analysis schedule, shortwave radiation in a densely forested area is nonsignificant. It appears that the longwave component, which is essentially a temperature function in dense forest, is adequately described by the convection parameter. In the analysis of the 1949 data at UCSL (85% effective forest cover) addition of maximum temperature as a separate independent variable reduced the convection parameter by 30% but had no effect on the condensation parameter and no effect on the over-all coefficient of determination for the function. The same treatment of the 1950 UCSL data showed the addition of maximum temperature as a separate variable to be non-significant and to have no effect whatever on the coefficients for the convection and condensation parameters. Reasons for this will be discussed under relation of wind effects to the convection and condensation parameters in later paragraphs. It is mentioned here to emphasize that in densely forested areas where it is expected that longwave radiation will be important and shortwave radiation unimportant, results of regression analysis are compatible with thermodynamic considerations.

6-03.16 <u>Condensation and convection</u>. - Parameters for convection and condensation are generally discussed together since they are both jointly associated with wind travel. Although in the past it has been thought by some investigators that convection and condensation parameters are sufficient to describe melt, work of the Snow Investigations has demonstrated that this is so only for limited conditions of weather and vegetative cover. The only situations where convection and condensation adequately describe the thermal budget are those where the direct influence of solar radiation on melt is inhibited either by dense forest canopy or by dense cloud cover such as may occur during storm conditions. The importance of heat exchange by longwave radiation exchange likewise must be assessed for specific conditions, and consideration should be given as to the adequacy of a convection parameter to index longwave exchange. Snow Investigations studies have been concerned primarily with clear-weather melt to assess the relative importance of convection and condensation in areas having different degrees of forest cover. Examination of coefficients of determination for three regression functions answers the questions:

1. What is the best that can be done with all three variables?
2. How well do convection and condensation by themselves explain melt runoff?
3. How well does radiation alone explain the melt-runoff without the aid of the convection and condensation parameters?

The answers to these questions may be seen in the following table which summarizes coefficients of determination for regression analyses on three areas having different degrees of forest cover as shown:

Year	Area	Effective forest cover, percent	Function		
			(1) Rad, conv, cond.	(2) Conv. cond. (w/o rad.)	(3) Rad. (w/o conv. cond.)
1954	Lysimeter*	0	.97	.46	.94
1954	Castle Cr.*	30	.90	.55	.80
1949	Skyland Cr.**	85	.89	.90	.29

* Allwave net radiation and hourly convection-condensation parameters used.

** Daily values of convection-condensation parameters and daily absorbed shortwave radiation.

Examination of the relative magnitudes of the coefficients of determination for the open area in the above table shows that the function suffers a serious loss of determination without radiation and suffers hardly any if radiation is used without convection and condensation. This is strong evidence of the unimportance of the convection and condensation parameters in open areas where radiation is the controlling variable. It is important to bear in mind, however, that these results apply to clear-weather conditions only. In open areas the situation could be easily reversed during stormy or overcast conditions when convection and condensation, combined with longwave radiation, might well become more important than shortwave radiation. The coefficients of determination for the lysimeter and Castle Creek were obtained from functions including allwave radiation and hourly convection and condensation parameters to emphasize the effect of radiation and to give the convection and condensation parameters the best possible advantage in the open and partly forested categories. In the forested category (Skyland Creek basin, UCSL) neither allwave radiation nor hourly convection and condensation parameters were available. In this case, however, absorbed shortwave radiation and daily convection and condensation parameters adequately demonstrate the importance of convection and condensation relative to radiation. In Skyland Creek basin the presence or absence of indexes of radiation had essentially no effect on the coefficient of determination, but omission of convection and condensation parameters caused the relation to suffer a serious loss of determination. This clearly demonstrates the inadequacy of radiation alone as a melt index in a forested area. For the forested category then, convection and condensation parameters by themselves explain all of the accountable variance in clear-weather snowmelt runoff. It is to be remembered, however, that for any degree of forest cover during clear weather, solar radiation is the primary source of heat supply, and consequently it is important in thermal budget analyses for all areas. Statistical analyses, on the other hand, relegate solar radiation to a place of no importance for forested areas because there is little detectable association between the day-to-day changes in radiation absorbed by snow in the open and the day-to-day variations in generated snowmelt runoff.

6-03.17 Regression coefficients for convection and condensation. - Coefficients obtained in multiple regression analyses vary both with magnitude and kinds of indexes used for the various components of the melt budget. Different values of convection and condensation coefficients are obtained, for example, when daily parameters are substituted for the hourly parameters of the preliminary analyses. Coefficients are influenced by the presence or absence of other variables such as radiation and wind, and the particular index used for these. For example, the magnitudes of the lysimeter convection and condensation coefficients were different when the shortwave and longwave components of radiation were substituted as separate independent variables for allwave radiation. The wind index (12 hr wind, 24 hr wind, etc.) will also affect the magnitude of the regression coefficients. An effort was made to apply the same indexes of convection and condensation to each of the areas studied, but as work progressed, it became apparent that the 1400 vapor pressure would be more representative for UCSL than the 1800 vapor pressure used in the CSSL analyses. More specific differences in magnitudes of derived coefficients are attributable to height of wind measurement, (UCSL: 31 feet; CSSL: 1 foot in 1954, 50 feet in 1951 and 1952). Even if the power law for temperature and wind variation with height as used at CSSL did apply in UCSL (doubtful because of the greater density of forest cover), coefficients would still be expected to differ by virtue of the differing elevation ranges of the two areas and the different locations of the index stations in their respective areas. Regression coefficients for the lysimeter (unforested site) represent melt occurring at a point (lysimeter area = 600 sq. ft.) close to the index station (150 feet from lysimeter to Station 3) and thus are not called upon to integrate the thermal budget of an entire basin several square miles in area as is the case for the other categories of forest cover. The other categories, part forest, forest, and dense forest exhibit varying quantities of transpiration loss and this loss is taken into account by the regression coefficients, while such is not the case for the lysimeter, where loss by evaporation from the snow surface is accounted for by the condensation parameter. All these considerations are possible sources of differences among derived coefficients. Nonetheless, it is to be expected that convection and condensation melt quantities will retain the same relative proportions in a given area, from year to year for any one function. To make such comparisons, the following table summarizes convection and condensation coefficients derived for the lysimeter, Castle Creek and Mann Creek and shows, in addition, the ratio of the condensation coefficient to the convection coefficient in each case, for equations involving hourly parameters. Convection and condensation coefficients for UCSL are excluded from the table because the results were obtained with daily parameters only.

Laboratory	Year	Equation No.	Table	Coefficients Conv.	Coefficients Cond.	Ratio cond./conv.
Lysimeter, CSSL	1954	(1)	6-1	.00011	.00105	9.5
Castle Creek, CSSL	1954	(1)	6-2	0.00018	0.00148	8.2
	1954	(7)	6-2	0.00024	0.00187	7.8
	1952	(9)	6-2	0.00012	0.00097	8.1
	1951	(11)	6-2	0.00011	0.00089	8.1
Mann Creek, WBSL	1951	(5)	6-4	0.024	0.085	3.5
	1950	(6)	6-4	0.036	0.112	3.1
	1949	(7)	6-4	0.039	0.094	2.4
	1949	(8)	6-4	0.048	0.147	3.1

The consistency of the ratios of condensation to convection, for each environment, leads to confidence in proper weightings of the condensation and convection parameters by the statistical analysis. Differences in the ratio among areas are mainly due to the dual role of the convection term in indexing the effect of longwave heat exchange in the forested areas. In section 6-4, in connection with the derivation of a general expression for snowmelt during rain-on-snow, the effect of longwave radiation is separated from the convection parameter and the proportional effects of condensation and convection show remarkable consistency between laboratories. Direct comparisons of regression coefficients as shown above can be made only when hourly parameters of convection and condensation are used, and where the melt resulting from direct shortwave radiation is adequately indexed by a separate variable for open areas.

6-03.18 Wind. - Assessment of the importance of wind in the heat transfer processes involving convection and condensation has been based on the results of regression analyses in which wind travel was either used jointly in the convection and condensation parameters or omitted entirely. The following table shows coefficients of determination from analyses with daily parameters for the partly forested and forested areas (Castle Creek, CSSL; Skyland Creek, UCSL). Radiation has been excluded to emphasize the effect of the presence or absence of wind on the coefficients of determination. The following coefficients of determination are abstracted from Tables 6-2, and 6-3:

Area	Table	Eq.No.	Independent variables	Coeff of Determination
Castle Creek CSSL, 1954	6-2	(13)	Convection, condensation (including wind)	0.47
"	6-2	(16)	Convection, condensation (excluding wind)	0.14
Skyland Creek UCSL, 1949	6-3	(4)	Convection, condensation (including wind)	0.90
"	6-3	(6)	Convection, condensation (excluding wind)	0.60

From the coefficients of determination in the above table it is apparent that exclusion of wind from the regression function results in a serious loss of determination. The loss of determination by excluding winds for individual periods is a function of the variance of wind within the period. For periods other than the ones shown above, the loss of determination by omission of wind is less. Admittedly the convection and condensation parameters by themselves are weak for the partly forested category (Castle Creek, CSSL) but they are considerably weaker without wind than with it. By induction we may infer also that the joint effect of wind will be important for estimation of convection and condensation melt at open sites. Analyses made on densely forested area (Mann Creek, WBSL) did not fully assess the importance of wind for this category of forest cover, for two fundamental reasons. First, although laboratory observations of wind were not available, surface winds in a dense forest are apt to be small and have little variance due to the stilling effect of the dense forest canopy on air movement. Therefore, although it is not conclusive, it may be said that wind is relatively unimportant for the estimation of clear-weather snowmelt runoff from densely forested areas. Secondly, in preliminary correlations, upper-air winds were found non-significant.

6-03.19 Temperature and vapor pressure bases. - The question of what base to use for convection and condensation parameters has long been a matter of conjecture. Thirty-two degrees has been a commonly used base but with no more justification than that it is the temperature at which ice melts. A melt index involving maximum air temperature alone would obviously need a base greater than 32°F. by reason of the diurnal temperature variation. Where air temperature is indexing solar radiation melt, however, the temperature base tends to be well below 32°. Variation of basin to station temperature (as, for example, when the temperature station is above or below the mean basin elevation), causes a displacement of the temperature base. Rational bases may be deduced for mean temperatures and, in the absence of other criteria will give acceptable results. If, however, regression analyses are being performed to evaluate melt-runoff coefficients, it is not necessary to rationalize or choose the base in advance. The base which is appropriate not only to the index station, but to the index itself (mean temperature, maximum temperature, etc.) with or without wind data may be computed from the derived constants of the regression equation.

6-03.20 Statistically derived bases for parameters excluding wind. - The concept of statistically derived bases was originated for studies of clear-weather melt in WBSL, as presented in Research Note 19. The general regression function for that area is

$$Q_{gen} = A(T_a - T_b) + B(e_a - e_b) \qquad (6\text{-}5a)$$

where Q_{gen} is the daily generated Mann Creek snowmelt runoff in inches over the snow-covered area of the basin, and T_a and e_a are the temperature and vapor pressure respectively at Station 6 in the Mann Creek basin. The coefficients A and B, and the bases T_b and e_b were to be evaluated by regression. By expanding equation 6-5a, the following is obtained:

$$Q_{gen} = AT_a + Be_a - (AT_b + Be_b) \qquad (6\text{-}5b)$$

Now, if the regression of Q_{gen} on T_a and e_a is computed by the usual least squares technique, the equation obtained will be

$$Q_{gen} = AT_a + Be_a + c \qquad (6\text{-}5c)$$

Having derived regression constants, A, B, and c in 6-5c, it is a matter of simple algebra to find T_b and e_b in 6-4b by writing the equation:

$$c = -(AT_b + Be_b) \qquad (6\text{-}6a)$$

which may be solved (by trial and error because of the non-linearity of the relationship between air temperature and saturation vapor pressure) to find a value of T_b and the corresponding saturation vapor pressure, e_b, which will satisfy equation 6-6a.

6-03.21 Statistically derived bases for parameters including wind. - Having developed the concept of derivation of bases from regression constants, as described in the preceding paragraph, it is a relatively simple matter to amplify the technique to apply to parameters which include wind, as in the following function:

$$Q_{gen} = A(T_a - T_b)v + B(e_a - e_b)v + c \qquad (6\text{-}7a)$$

If, as before, we expand the right hand member,

$$Q_{gen} = AT_a v + Be_a v - (AT_b + Be_b)v + c \qquad (6\text{-}7b)$$

where the quantity in parenthesis may be lumped into a single constant, C, the coefficient of v. Thus, for parameters containing wind, the least squares analysis is performed in the regression of runoff on T_a, e_a, and v, to obtain the coefficients

A, B, and C. As before, T_b and e_b are obtained by trial solution of

$$C = -(AT_b + Be_b) \qquad (6\text{-}6b)$$

with the important difference that, in this case, C (big "C") is the derived coefficient of the wind term and not the regression constant, c, (little "c") as in the case of no wind. In the case with wind there will also be a regression constant, c, which is the Y-intercept and an integral part of the regression equation. It is emphasized that, in order to compute coefficients and bases correctly when convection and condensation parameters have wind as a factor, wind must be carried as a separate independent variable in the regression function. It is also important, however, to discard the wind term in writing the final regression equation, if the computed bases are written into the parameters. The coefficients for wind have been included in tables 6-1, 6-2, and 6-3, where appropriate, only to provide complete information on derivation of the bases.

6-04. ESTIMATES OF SNOWMELT DURING RAIN

6-04.01 General. - Evaluation of snowmelt during rain represents a special condition for which certain simplifying assumptions can be made in the snowmelt equation. As in the case for clear-weather snowmelt, the form of the equation is dependent somewhat upon the type of area to which it is to be applied. Heat transfer to the snowpack during rain involves the following basic considerations in application of indexes:

1. Shortwave radiation is relatively unimportant and can be evaluated as a constant amount.
2. Longwave radiation exchange between forest or low clouds and the snowpack may be adequately indexed linearly by air temperature.
3. Air is assumed to be saturated, so that air temperature (combined with wind) may be used to index both convection and condensation melt. By assuming a linear relationship between air vapor pressure and dewpoint, for the range normally experienced under these conditions, a linear expression of convection and condensation melt may be assumed as a function of air temperature and wind.
4. Wind is important and should be evaluated for open or partly forested basins. On heavily forested basins, however,

wind variation is so reduced beneath the forest canopy that an average wind condition may be assumed, thereby eliminating the wind variable.

5. Rain melt may be evaluated simply as a function of rainfall intensity and air temperature.

6. Ground melt is unimportant and may be estimated as a constant amount.

7. Water lost by evapotranspiration is negligible.

6-04.02 Melt rates during rain may be determined from clear-weather melt coefficients derived for the laboratories by applying conditions set forth in the preceding paragraph and evaluating separately the convection-condensation melt for the basin. These coefficients integrate basin characteristics and conditions of measurement for the particular laboratory to which they apply, but they may be used as a guide for general application to other basins. Coefficients of convection-condensation melt derived for the Lower Meadow lysimeter at CSSL are directly applicable to open areas. Adjusted coefficients for WBSL may be applied generally to heavily forested areas. Coefficients for partly forested areas may be estimated for the specific conditions of exposure and instrumentation.

6-04.03 In the following paragraphs a generalized snowmelt equation for computing point or basin snowmelt during rain is developed from theoretical considerations of heat transfer and derived coefficients of convection-condensation melt. Separate melt equations are presented for areas having different forest cover. For basin application, the general equation must be interpreted in terms of average conditions, considering the areal distribution of each variable. The basin may be divided into sub-areas or elevation zones in computing the snowmelt, but cognizance should be given to variability of forest cover and exposure to wind in each sub-area.

6-04.04 Project basin coefficients for heat indexes can be derived from analysis of rain-free melt data by methods set forth in section 6-03 for the snow laboratories, and extended to rain-on-snow conditions, with due consideration for particular type of environment involved. Such a procedure gives indexes of snowmelt runoff for a particular basin and condition of observation. When it is impractical to derive basin coefficients, a generalized rational equation must be used.

6-04.05 <u>Coefficients of melt during rain at laboratories.</u> - Melt coefficients during rain were derived from a combination of statistically and rationally determined melt coefficients for clear-weather melt, by separation of terms in the thermal budget. For this condition, melts by longwave shortwave radiation exchange, as well as rain melt and ground melt, are treated adequately from theoretical general equations or assumed average rates of energy transfer. Convection and condensation resulting from turbulent exchange of energy between the atmosphere and the snowpack is an important source of melt and must be evaluated for the specific condition of environment. Comparative coefficients of convection and condensation melt for varying environments are presented from the convection-condensation coefficients derived rationally for the lysimeter, and from statistically derived coefficients for the lysimeter, Castle Creek at CSSL and Mann Creek at WBSL.

6-04.06 Convection-condensation melt coefficients. - The basic equation for convection-condensation melt set forth in chapter 5 may be simplified and reduced to a function involving air and dewpoint temperatures and wind, on the assumption of linear variation of vapor pressure and dewpoint for the range between 32°F and 50°F, and for a constant snow surface temperature of 32°F. The equation may be written in the general form

$$M_{ce} = K\ (v)\ (AT_a + BT_d - 32) \qquad (6\text{-}8a)$$

where M_{ce} is convection-condensation melt, v is the wind speed, T_a and T_d are the air and dewpoint temperatures, K is the point or basin coefficient of convection-condensation melt, and A and B are the effective respective weights of air temperature and dewpoint temperature in producing convection-condensation melt. The sum of A and B is unity, and in the case of saturated air, air temperature and dewpoint are equal, so that the expression becomes

$$M_{ce} = K\ (v)\ (T_a - 32) \qquad (6\text{-}8b)$$

The coefficients K, A, and B have been evaluated for the 1954 observations at Lower Meadow lysimeter, CSSL, on the basis of the rational coefficients presented in chapter 5, as well as for statistical weightings of the data, in which net longwave, net shortwave, and hourly parameters of convection and condensation were evaluated separately in a multiple linear regression analysis. Coefficients of convection-condensation melt for Castle Creek were derived for the 1954 melt period by statistical weightings using the same parameters as for the lysimeter. For Castle Creek, the longwave radiation exchange in the forest was indexed by the convection parameter. The longwave portion of the convection parameter coefficient was separated by utilizing the ratio of longwave exchange to combined longwave and convection in forested areas of 0.72, as derived in Research Note 11, (see figure 2, plate 6-2) and applying this ratio to the area of forest canopy effective in longwave radiation exchange for CSSL, estimated to be 30%. For example, the convection parameter coefficient for CSSL in 1954 was 0.00018 (equation 1, table 6-2). With 30% of the basin forested and 72% of combined longwave and convection representing longwave exchange in the forest, the longwave-in-forest part of the convection coefficient is 0.72 x 0.30 x 0.00018 or 0.00004 and the true convection amount is (0.00018 - 0.00004) = 0.00014. With convection isolated in this way, it is possible to compare relative proportions of convection and condensation for Castle Creek with those for the lysimeter. The manner in which this is accomplished is illustrated by the following computation. Having isolated the convection coefficient as described above, an expression for convection and condensation melt may be written (equation 1, table 6-2):

$$M_{ce} = 0.00014\ (T_a - T_s)v + 0.00148\ (e_a - e_s)v \quad (6\text{-}9a)$$

Converting the condensation coefficient for use with dewpoint in $^{\circ}F$ instead of vapor pressure in millibars (multiply vapor pressure coefficient by 0.34),

$$M_{ce} = 0.00014\ (T_a - T_s)v + 0.00050\ (T_d - T_s)v \quad (6\text{-}9b)$$

By factoring the sum of the convection and condensation coefficients, the following is obtained:

$$M_{ce} = 0.00064\ (0.22T_a + 0.78T_d - T_s)v \quad (6\text{-}9c)$$

which shows that 22% of the combined convection-condensation melt is due to convection and 78% due to condensation. The coefficient 0.00064 is for hourly melt amounts for winds at 1 foot and temperatures at 10 feet above the snow surface. The coefficient must be multiplied by 24 for daily melt amounts and divided by 1.9 to correct winds to the 50-foot level. The corresponding coefficient is 0.008, which is the value shown in the table below for 1954 Castle Creek with net shortwave and net longwave as the radiation variables. It should be remembered that wherever net longwave is included as a separate variable, it represents only longwave radiation exchange in the open and should be distinguished from longwave exchange in the forest indexed by the convection parameter described above. Since there is no forest on the lysimeter, the convection coefficient derived for the open site is a pure convection coefficient and the above technique for separation of longwave in the forest does not apply in that case. The convection-condensation coefficients are presented below for CSSL, elevation 7200 feet msl, with measurements of temperature and dewpoint at the 10-foot level and wind at the 50-foot level at the Lower Meadow: (see figure 3 (b) plate 6-2)

CONVECTION-CONDENSATION MELT RATES, CSSL, IN/DAY

$$M_{ce} = K(AT_a + BT_d - T_b)v$$

Radiation variables	DATA (METHOD)	K*	A	B
Net shortwave Net longwave	1954, Lysimeter (Rational)	0.0084	0.20	0.80
	1954, Lysimeter (Statistical)	0.0058	0.24	0.76
	1954, Castle Creek (Statistical)	0.0080	0.22	0.78
Net shortwave only	1954, Castle Creek (Statistical)	0.0104	0.23	0.77
	1952, Castle Creek (Statistical)	0.0102	0.22	0.78
	1951, Castle Creek (Statistical)	0.0093	0.22	0.78

* Values of K for M_{ce} in inches per day in equation 6-8b.

It is seen that coefficients derived independently show remarkable consistency. In the upper part of the table it may be seen that the convection-condensation melt rate (K) for Castle Creek is represented closely by the rationally derived convection-condensation melt rate for the lysimeter, which seems reasonable considering the exposure to wind of the Lower Meadow in comparison with the Castle Creek basin as a whole. This concept is further supported by comparisons in the lower portion of the above table, which represents results of three clear-weather melt periods from different seasons (1954, 1952, and 1951, equations 2, 5, and 7 in table 6-2). The 1954 coefficients have been adjusted for use with winds measured at the 50-foot level, (in 1952 and 1951 recorded 50-foot winds were used in the derivations), longwave loss in the open is not a variable in the regression model for the lower portion of the table, but its average effect is felt through interaction with other variables and in the regression constant, which is remarkably uniform among the three years. Here the melt rate for convection and condensation is slightly higher than was obtained from the 1954 data when longwave loss in the open was included in the regression function. This difference may be attributed to possible calibration errors in the shortwave and allwave radiation instruments. (Net longwave radiation exchange was not separately measured but was computed as the difference between allwave and shortwave radiation observations). Thus the estimate of longwave loss in the open may be in question and the higher melt rates should be used as a guide when extrapolating to areas similar to Castle Creek.

6-04.07 Mean daily temperatures and vapor pressures were used in deriving total melt coefficients at WBSL. There, however, wind effects are greatly reduced because of the heavy forest which covers nearly the entire basin. Although no continuous wind measurements were obtained at WBSL, attempts to improve determination of melt by adding wind parameters were unsuccessful, and the degree of determination without wind was consistently high. Therefore, it was concluded that convection and condensation within heavy forest are adequately represented by an average wind amount, implicit in the total coefficient of melt. Coefficients of convection and condensation melt, shown in table 6-4, derived from mean daily ai_ temperatures and vapor pressures, provide a means for determining the basin coefficients of $\underline{K}$, $\underline{A}$, and $\underline{B}$ in equation 6-8a, modified by exclusion of the wind term, which can be compared with CSSL coefficients. The longwave exchange between the snow and forest was subtracted from the temperature coefficient. Values of $\underline{K}$, $\underline{A}$, and $\underline{B}$ for Mann Creek, WBSL, are tabulated below for each of the four periods studied, with adjustments for height of measurement of temperature and vapor pressure to 10 feet above the snow surface:

CONVECTION-CONDENSATION MELT RATES, WBSL, IN/DAY

$$M_{ce} = K\ (AT_a + BT_d - T_b)$$

Period	K*	A	B	T_b
April, 1949	0.043	0.22	0.78	31
May, 1949	0.039	0.26	0.74	33
May, 1950	0.048	0.22	0.78	30
April, 1951	0.036	0.20	0.80	29
Mean	0.042	0.22	0.78	31

* K for melt rates expressed in inches per day.

Adjusting the melt coefficients above to a theoretical 32°F base, the mean daily convection-condensation melt for Mann Creek, WBSL, is

$$M_{ce} = 0.045\ (.22T_a + .78T_d - 32), \qquad (6\text{-}9d)$$

where M_{ce} is expressed in inches per day. (see figure 3a, plate 6-2) The tabulation of coefficients above shows the stability of the relationship between the relative weights of convection and condensation melts, expressed by the coefficients A and B. The proportional weights given in the preceding paragraph for CSSL, although at a different elevation level, are in general agreement with WBSL. The magnitude of the coefficient K for the heavily forested WBSL corresponds to an average wind of approximately 5 miles per hour at the 50-foot level, or between 2 and 3 miles per hour at the 1-foot level. This is a reasonable order of magnitude for conditions in the forest.

6-04.08 Shortwave radiation melt. - Snowmelt by shortwave radiation, M_{rs}, is relatively unimportant during periods of rain-on-snow. Studies of incident radiation during these periods show that it is reasonable to assume a constant daily average of 40 ly for an open area with an average albedo of the snow surface of 65 percent. The resulting net snowmelt is 0.07 inch per day. For forested areas it may be less, depending on the areal extent and density of forest cover.

6-04.09 Longwave radiation melt. - Longwave radiation during periods of significant precipitation can be adequately estimated by the theoretical exchange of blackbody radiation between the snow surface and the forest canopy or low clouds. During storm conditions, turbulent mixing in the lower layers of the atmosphere establishes an equilibrium between air temperature measured at the normal instrument height and the temperature of the forest or low clouds. When the cloud base is less than 1000 feet above the ground, air temperature lapse rate corrections may be ignored. A linear relationship of net longwave radiation exchange closely approximates the theoretical exchange expressed by equation 5-12. The net melt by longwave radiation for rain-on-snow conditions, expressed in terms of air temperature and with a snow surface temperature of 32°F is

$$M_{rl} = 0.029\ (T_a - 32) \tag{6-10}$$

where $\underline{M_{rl}}$ is melt in inches per day resulting from net longwave radiation exchange, and $\underline{T_a}$ is the air temperature in degrees F.

6-04.10 Rain melt. - Snowmelt by the transfer of heat from rain, as discussed in chapter 5, may be expressed in terms of average daily rainfall rate and free air temperature as in the following equation:

$$M_p = 0.007\ P_r\ (T_a - 32) \tag{6-11}$$

where $\underline{M_p}$ is the daily snowmelt from rain, $\underline{P_r}$ is the daily precipitation in inches, and $\underline{T_a}$ is the air temperature in degrees F.

6-04.11 Ground melt. - Snowmelt from ground heat, $\underline{M_g}$, may be estimated at 0.02 inch per day (see section 5-10).

6-04.12 Generalized convection-condensation melt equation. - It was shown in paragraph 6-04.06 that for saturated air, convective melt is only approximately 20 percent of the total convection-condensation melt, as shown by the value of the coefficient $\underline{A}$ determined from both rational and statistical evaluations for several different conditions of environment. It was explained in chapter 5 that convective heat transfer is a function of air density, and that differences in convective heat due to elevation may be expressed as an elevation or pressure function. The magnitude of the elevation correction for convective melt, in terms of total convection-condensation melt for saturated air, is relatively small, and for an elevation difference between sea level and 7,000 feet, the correction represents only 5 percent of the total melt. Considering the accuracy of the over-all melt coefficient, the elevation correction for convective melt is not warranted. The coefficients derived for the laboratories, therefore, may be applied without regard to elevation differences. The following generalized equations represent convection-condensation melt in inches per day during rain, for varying environments:

(1) for melt at a point in the open:

$$M_{ce} = .0084\ v\ (T_a - 32) \qquad (6\text{-}12a)$$

(2) for basin melt from open or partly forested areas:

$$M_{ce} = (k)\ .0084\ v\ (T_a - 32) \qquad (6\text{-}12b)$$

(3) for heavily forested areas:

$$M_{ce} = .045\ (T_a - 32) \qquad (6\text{-}13)$$

where T_a is the temperature of saturated air at the 10-foot level in $^{\circ}F$, v is the wind speed at the 50-foot level in miles per hour, and k is a basin constant, considering the conditions of measurement with respect to average basin topographic characteristics and exposure to wind. Conversion to different observation levels of temperature and wind from those specified in the above equations may be accomplished according to the power law variation explained in chapter 5, with the reservation that the exponent n, may vary from one area to another. Increased turbulence due to rain should tend to increase M_{ce}. Experiments are needed to determine the effect of rain on both the temperature and wind profiles near the surface.

6-04.13 General equation for total basin melt during rain. - Components of melt may now be combined to form a general equation for total basin melt during rain. Since total melt, M, is expressed by the relationship,

$$M = M_{rs} + M_{rl} + M_{ce} + M_g + M_p \qquad (6\text{-}14)$$

the terms may be combined for environmental conditions as follows:

(1) for open or partly forested basin areas,

$$M = (0.029 + 0.0084kv + 0.007P_r)\ (T_a - 32) + 0.09 \qquad (6\text{-}14a)$$

(2) for heavily forested areas,

$$M = (0.074 + 0.007P_r)(T_a - 32) + 0.05 \qquad (6\text{-}14b)$$

where $\underline{M}$ is total daily snowmelt in inches per day, T_a is the temperature of saturated air at the 10-foot level in degrees F, $\underline{v}$ is the wind speed at the 50-foot level in miles per hour, P_r is the rate of precipitation in inches per day, and $\underline{k}$ is the basin constant as defined in the previous paragraph. The value of $\underline{k}$ varies from about 0.2 for densely forested areas to slightly over 1.0 for exposed ridges or mountain passes. It is emphasized that total basin melt can be computed from the above equations only by use of values of wind and temperature which are representative of average conditions over the snow-covered area of the basin or by integration of melts computed from representative zonal averages of temperature and wind.

6-05. PROJECT BASIN APPLICATION

6-05.01 General. - Previous sections of this chapter have dealt with analyses of laboratory data to demonstrate the essential elements of the thermal budget and the kinds of meteorological data required for snowmelt indexes in drainage areas having various degrees of forest cover. While coefficients derived in these analyses are strictly applicable only to the areas for which the investigations were made, the basic principles which have been demonstrated may be used to establish clear-weather snowmelt-runoff relations for other areas. A study of this kind has already been made for a densely forested area and published as Supplement to Research Note 19 (North Santiam River, Willamette Basin, Oregon). This section will deal with investigations for a project basin in the partly forested category, namely, Boise River basin above Twin Springs, Idaho. Evaluation of basin snow cover and a hydrograph reconstitution for this stream will be discussed in chapters 7 and 9.

6-05.02 Basin characteristics. - The Boise River above Twin Springs, Idaho, drains an area of 830 square miles in the upper watershed of the Boise River. Situated in the Sawtooth Mountains, the area has deep valleys, steep slopes, and narrow sharp-top ridges. It ranges in elevation from about 3000 to 9000 feet. The stream system and instrumentation are shown in figure 3, plate 9-5. Forest cover consists principally of conifers which are estimated to have a shading effect on 30% of the basin area.

Practically the entire runoff of the Boise River is contributed from headwaters areas, 3/ and spring floods on this river result primarily from melting snow. Consequently, it is of paramount importance for optimum use of storage in Boise River reservoirs to have a reliable and accurate technique of estimating the basin's snowmelt contribution to spring runoff.

6-05.03 Scope. - Exploratory analyses for selecting indexes best suited to this basin are limited to study of clear-weather days in May and June 1955. Final indexes selected on the basis of the 1955 study are used for independent derivation of a clear-weather snowmelt-runoff equation for the 1954 season and for the combined data of both seasons. Selection of the Boise River basin and the 1955 and 1954 melt seasons in particular for this application was based primarily on the availability of snow-cover information from several aerial reconnaissance flights over the area during these seasons, performed by personnel from the Walla Walla District Office of the Corps of Engineers.

6-05.04 Melt components. - It will be recalled that the primary thermal budget components affecting snowmelt in a partly forested area are shortwave radiation, longwave radiation loss in the open, convection, and condensation. The following paragraphs will discuss the data available for estimating each of these components in the Boise basin. The analysis shows that available humidity data did not satisfy the requirement of a condensation-melt index; that maximum temperature alone served as a better convection-melt index than a temperature-wind product; and that an estimate of longwave loss in the open from 700 mb and surface minimum temperatures significantly improved the overall runoff estimate.

6-05.05 Snowmelt runoff. - By application of a standard recession curve to the continuous discharge hydrograph of the Boise River above Twin Springs, separation of daily generated snowmelt-runoff amounts was accomplished after the manner described in Research Note 19. Number of available clear days and hydrograph base flow are shown for each season in the following table:

Month and year	Clear days	Base flow - cfs
May - June 1955	26	0
Apr - May 1954	21	1500

Daily runoff quantities were expressed in inches depth on the snow-covered portion of the basin. These runoff amounts were used as the dependent variable in all regression analyses of this section.

6-05.06 Area of snow cover. - The estimation of areal extent of snow cover was based on aerial reconnaissance flights in 1954 and 1955. Day-to-day variation of the snow-covered area was determined as part of the material in chapter 7 and is fully discussed therein.. In general, the estimate of snow cover appears to be too high at the start of the 1955 period of study and too low at the end. If this is actually the case, the computed values of runoff (depths over snow-covered area) are too low at the beginning and too high at the end. The net effect of this error would be to cause derived melt coefficients to be too high, resulting in melt estimates that are too high in the early part of the period and too low at the end. These errors are clearly seen in the reconstitution of the 1955 hydrograph in chapter 9. No refinements in snow cover were made, however, for derivation of the melt coefficients, since the primary purpose of this study was the demonstration of the method. The present results, even as a first approximation, show that the parameters used explain a large portion of the runoff variance and are superior to those obtained with a temperature index alone.

6-05.07 Shortwave radiation. - Pyrheliometric observations of incident shortwave radiation were available at Boise. These were used for the computation of absorbed shortwave radiation in conjunction with estimates of snow albedo based on decay curves of albedo versus age of snow surface (figure 4, plate 5-2) by methods described in Technical Bulletin 6, taking into account the occurrence of new snowfalls during the period. Albedo versus time is shown in plate 6-3, figure 1.

6-05.08 Longwave radiation. - Net longwave exchange within the forested portions of the basins is considered to be adequately described by the temperature index. Longwave loss in the open, however, was estimated (for the latter studies) from a graphical relationship as shown on figure 2, plate 6-3. It is emphasized that this relation was devised expressly for the Boise basin study and is not intended for general application. Time limitations do not permit detailed analyses for the establishment of a universally applicable relation of this kind. The parameter for net allwave radiation in the open was obtained by adding together absorbed shortwave radiation and estimated longwave loss.

6-05.09 Convection and condensation. - Required data for convection and condensation parameters are:

1. Convection - mean or maximum temperature and daytime wind.
2. Condensation - mean or some instantaneous dewpoint (or vapor pressure) and daytime wind.

In this study, temperature records were available for five stations in or near the basin (Boise, el. 2842; Idaho City, el. 3940; Lowman, el. 3794; Atlanta Summit, el. 7590; and Bald Mountain, el. 8700)*. Humidity and wind records were available only at Boise. The data selected for trial convection and condensation parameters were: Boise maximum temperature; Boise mean daytime dewpoint temperature (average of 1100 and 1700 observations); and Boise mean daytime wind (average of 1100 and 1700 observations).

6-05.10 Analytical procedure. - The following paragraphs will show the variables which were tested in regression analyses for estimation of Boise River snowmelt runoff and will present some of the intermediate results to demonstrate the process by which the final results were achieved. The following variables were tested for use as indexes of components in the thermal budget of this basin; the table shows the combinations of these variables which were studied and the coefficient of determination computed for each combination:

X_1 Absorbed shortwave radiation, ly

X_2 Incident shortwave radiation, ly

X_3 Convection parameter ($T_{max} \cdot v$), (T_{max} and v at Boise, in ^oF and in mph, respectively)

X_4 Condensation parameter ($T_d \cdot v$)

(T_d = mean Boise daytime dewpoint, average of obs. at 1130 and 1730 MST, in ^oF)

X_5 Mean Boise daytime wind, v (average of obs. at 1130 and 1730 MST, in mph)

X_6 Boise maximum temperature, (eqs. 1-10), Idaho City maximum temperature (eq. 11); ^oF

X_7 Boise 700 mb temperature (0800 MST), ^oC

* Elevations in feet above m.s.l.

X_8 Boise 700 mb dewpoint (0800 MST), °C

X_9 Longwave loss in open (estimated from Boise 700 mb temperature and Idaho City minimum temperature), ly

X_{10} Estimated allwave radiation ($X_1 + X_9$), ly

Eq. No.	X_1	X_2	X_3	X_4	X_5	X_6	X_7	X_8	X_9	X_{10}	D
(1)	*	*	*	*	*						0.82
(2)			*	*	*						0.82
(3)	*		*		*						0.83
(4)			*		*						0.82
(5)	*										0.58
(6)	*		*		*		*	*			0.87
(7)	*					*					0.86
(8)						*					0.75
(9)	*					*			*		0.89
(10)						*				*	0.89
(11)						*				*	0.86

6-05.11 Sequence of analyses. - The table of the previous paragraph shows not only the different combinations of variables tested, but also the sequence or order of progress from one function to the next. The first four equations demonstrated that convection and condensation parameters alone were as good as convection and condensation parameters with absorbed and incident radiation added. Furthermore, the convection parameter alone, in equation 4, was as good a determinant of the snowmelt runoff as any combination shown by the first three equations. Without the condensation parameter, shortwave radiation and convection were slightly better than the convection parameter alone (eqs. 3 and 4). However, the first order relation of runoff to absorbed shortwave radiation had a determination of 0.58 (eq. 5) approximately the same order of magnitude as obtained for the partly-forested laboratory area (Castle Creek, CSSL). The weakness of the radiation variable in the presence of the convection parameter indicates interaction of radiation and local climate at a valley station where the ground is bare of snow.

6-05.12 Upper air indexes. - Of the first five derived equations, equation 3 was best with a coefficient of determination of 0.83. The question of how much improvement could be gained by addition of 700 mb temperatures and dew points was answered in equation 6. (D = 0.87) Here, both absorbed shortwave radiation and the 700 mb temperature were nonsignificant, while the convection parameter continued to be highly significant, and the 700 mb dewpoint was acceptably significant. The coefficient of the 700 mb dewpoint had a negative sign, indicating that it was not performing the desired index function, namely that of representing condensation melt. For this reason, 700 mb dewpoint was dropped as a condensation index. Thus, even though this function (eq. 6), improved noticeably over equation 3, it was decided to abandon the condensation index, and to further explore modifications of equation 3 by omission of wind from the convection parameter and by addition of an estimate of longwave radiation loss in the open.

6-05.13 Wind. - Equation 7 was set up with a view to examining the relative importance of wind in the convection parameter. This function, equation 7, contained only two independent variables, absorbed shortwave radiation and Boise maximum temperature, and had a coefficient of determination of 0.86 as contrasted with equation 3 which had only 0.83. With omission of wind, absorbed shortwave radiation became highly significant, apparently due to a lesser degree of dependence between absorbed radiation and maximum temperature at Boise than between radiation and the convection parameter (X_3 = product of maximum temperature and wind). This appears reasonable for a valley station where the air temperature is not measured over snow. The importance of absorbed shortwave radiation in equation 7, is emphasized by equation 8, which shows that Boise maximum temperature without shortwave radiation is far less effective than with it. Although there is little difference in over-all effectiveness between equation 4 (which uses the Boise max temp-wind product) and equation 7 (which omits wind but includes absorbed shortwave radiation), the latter is favored because rationally it is better suited to the general theoretical thermal budget of the area. It is well to bear in mind, however, that a temperature-wind index may be better than temperature alone in situations where radiation estimates are desirable but not available.

6-05.14 Longwave radiation in the open. - Thus far, all components of the thermal budget have been accounted for, with the exception of longwave loss in the open. Examination has been made of absorbed shortwave radiation, convection, and condensation, with decisions to retain absorbed shortwave radiation, to discard Boise wind from the convection parameter, and, for lack of adequate data, to abandon completely the condensation index. The only remaining component which has not yet been examined is longwave radiation exchange in the open. Since longwave loss in the open has been demonstrated by laboratory studies to be a major component of the thermal budget for this kind of area, attention is now focussed on examination of indexes for this component.

6-05.15 A graphical relation for estimating longwave loss as a function of 700 mb and minimum surface temperatures was devised, using Boise upper-air data and Idaho City minimum temperatures. Construction of the chart is based on the concept, that, for a given airmass condition, minimum temperature is a function of the amount of nighttime cooling by longwave radiation. Therefore, the temperature condition of the airmass, as indexed by the 700 mb temperature, may be combined with the minimum nighttime surface temperature to represent the total nighttime loss by longwave radiation, which in turn may be used to represent the loss for the 24-hour period. Surface winds also affect minimum temperatures, but inclusion of a wind parameter in the chart for the Boise River basin was not required. Idaho City (rather than Boise) minimum temperatures were used in order to provide a more representative index of local climate in the basin. The adopted relation for this study is shown in figure 2, plate 6-3. It is again emphasized that this relation was specially derived as an expedient for this particular study and may not be applicable to other areas without modification.

6-05.16 The effectiveness of the longwave loss index as estimated from 700 mb and minimum surface temperatures when tested in equation 9 (D = 0.89) showed a significant improvement over equation 7. Since the coefficients for absorbed shortwave, and the longwave-loss index derived in equation 9 were nearly equal, the longwave-loss estimate was combined with absorbed shortwave to give an estimate of net allwave radiation effective in producing snowmelt. This estimate was tested in equation 10 with a coefficient of determination of 0.89 (equal to that for equation 9). Equation 10 represents the final result of this series of analyses and explains all but approximately ten percent of the variance of clear-weather snowmelt runoff from the Boise River basin above Twin Springs, Idaho. Daily runoff estimated from this equation is plotted against observed runoff in figure 3, plate 6-3.

6-05.17 Regression equations for 1955 and 1954. - To assess the universality of application of the snowmelt equation derived for this basin from 1955 data (eq. 10a, table 6-5) a similar derivation was made with data from the 1954 season and with data of both seasons combined. 1954 and 1955 snowmelt runoff coefficients are in close agreement. All coefficients, and statistical measures of accuracy are shown in table 6-5.

6-05.18 For the purpose of reconstitution of streamflow hydrograph (presented in chapter 9), a melt equation was derived, with Idaho City maximum temperatures rather than Boise maximum temperatures (eq. 11, table 6-5) because Idaho City is at a higher elevation and is nearer the basin, and because Idaho City minimum temperatures had been used for the estimation of longwave radiation loss in the open. The coefficient of determination, however, was lower than that for the case where Boise maximum temperatures were used.

6-05.19 Summary. - This section has presented an investigation of clear-weather snowmelt runoff from the Boise River basin above Twin Springs, Idaho. The basin is partly forested and suitable indexes were applied to represent the thermal-budget components of melt essential to this category of forest cover. It was found that allwave radiation was an essential item in the melt budget of the area; that a combination of temperature and wind at a valley station was less effective than temperature alone; that condensation is not adequately indexed by any presently available data; and, incidentally, that a quite satisfactory estimate of longwave radiation loss from snow in the open could be made from 700 mb temperature and surface minimum temperature. The final melt equation, (10c, in table 6-5) is

$$Q_{gen} = 0.00238\ G + 0.0245\ (T_{max} - 77) \qquad (6\text{-}15)$$

where Q_{gen} is daily generated snowmelt runoff of Boise River above Twin Springs, Idaho in inches over snow-covered area, G is estimated daily net allwave radiation exchange (snow and sky) in the open, in langleys, and T_{max} is Boise daily maximum temperature, in degrees F. The high temperature base of 77°F reflects the adjustment necessary in using data from a valley station bare of snow. The equation explains 90 percent of the runoff variance and gives runoff estimates that are correct within 0.11 inch of the observed daily values of generated runoff, approximately 67 percent of the time.

6-05.20 This example of the Boise River was primarily intended to illustrate the method of approach and the kind of problems which confront the hydrologist seeking to establish snowmelt runoff rates for project basins. The problems are as follows:

(1) Determination of physical characteristics of the basin with regard to topography, forest cover, soil, and ground-water conditions.

(2) Separation of daily increments of snowmelt runoff by hydrograph analysis.

(3) Availability of adequate meteorological data for basin representation of air temperature, humidity, wind speed, and allwave radiation.

(4) Determination of snow-surface albedo.

(5) Evaluation of snow-covered area.

6-05.22 To the hydrologist accustomed to using degree-day indexes for estimating snowmelt, the procedures presented thus far in this chapter may seem complex and cumbersome. It is important to bear in mind, however, that the best indexes of snowmelt are those which describe as closely as possible the thermal budget of the drainage area concerned. The degree of closeness sought and the availability of data both influence the choice of method. There are situations in which it is desirable to use a temperature index only. The following section deals with the applicability and effectiveness of temperature data alone as a snowmelt index.

6-06. TEMPERATURE INDEXES

6-06.01 General. - In many areas where snowmelt is an important factor in runoff, air-temperature measurements are the only data available from which snowmelt can be computed. Moreover, air temperature is a simple index of snowmelt, and, as was demonstrated in the preceding section, is the best single index of snowmelt for forested areas. For these reasons, temperature indexes are the most widely used method of computing snowmelt. Much work has been done in relating snowmelt amounts to temperature indexes; the most commonly used index is degree-days above freezing, that is, mean daily temperature in excess of 32°F. Mean temperature is usually taken as the mean of the daily maximum and minimum temperatures, those two measures of temperature being generally available. The 32°F base follows from the idea that most snowmelt results directly from the transfer of sensible heat from the air in excess of 32°F. While it is now known that air temperature is not the primary cause of snowmelt in most areas, 32°F is still a good average base value for use with mean temperatures. The use of a time period of one day results from the diurnal pattern of snowmelt; in most areas each day's melt may be identified by a drop in the discharge hydrograph associated with considerable reduction or even cessation of melt during the night. Time periods other than one day have been used under special circumstances with appropriate temperature indexes of degree-half days, degree-hours, etc. Also temperatures other than mean daily temperature are used for temperature indexes (e.g., maximum daily temperature), and bases other than 32°F are often used. These other temperatures, temperature bases and time periods will be examined later. For the moment, only the common index of degree-days above freezing will be considered.

6-06.02 Point melt rates. - Whereas snowmelt runoff is determined from hydrograph analysis, as in the preceding sections, snowmelt or snowpack ablation rates are determined from the analysis of snow survey data. This is done by noting the change in water equivalent on the snow course (or point within the snow course) over a period of several days, and relating it to the accumulated temperature index during the period. The period selected should be long enough and have a change in water equivalent large enough to make the errors of measurement small, relative to the total change in water equivalent. Where data from the entire snow course are used, care must be taken that all points remain snow covered. If any of the points become bare, a lesser melt rate than actually occurred is indicated.

6-06.03 It should be noted that snowmelt defined in this manner--the change in water equivalent of the snowpack--may differ from the melt equivalent of the heat supply to the snowpack at the snow surface, especially early in the snowmelt season. Before the pack has been thoroughly ripened, some of the melt water from the snow surface may refreeze within the pack and still more may go to satisfy the liquid-water-holding capacity of the pack (see chapter 8). Furthermore, the snow-surfaces albedo decreases as the season progresses. Assuming a temperature index to be a good indicator of heat supply, then a changing relationship between the index and snowmelt is indicated while the snowpack is being ripened and snow-surface albedo is changing.

6-06.04 Early investigations of temperature indexes of snowmelt were mostly concerned with point melt rates. In 1914, Horton 6/ performed experiments wherein cylinders of snow were cut from the snowpack and melted under laboratory-controlled temperature conditions. He found the melt rate to be about 0.04 to 0.06 inch of water per 24 hours for each degree above 32°F for the conditions of the experiment. In 1931, Clyde 2/ performed similar experiments and determined average melt rates of 0.05 to

0.07 inch per degree-day above freezing. However, these laboratory experiments are not representative of actual melt conditions, since the radiation exchange was certainly not typical of field conditions. In addition, since the snow samples stood on a table in the laboratory, the air which became cooled by contact with the snow could drain away and be replaced by warmer air, as pointed out by Horton in subsequent article. 7/ This condition does not exist over snowfields where the air cooled by the snow tends to stagnate over the snow surface. Due to this effect alone, a smaller degree-day factor than those cited above, would be indicated; however, the inclusion of allwave radiation would indicate a larger factor.

6-06.05 Clyde 2/ also determined spring snowmelt rates from a field study made at Gooseberry Creek, Utah, over a period of 17 days. Using snow-course data, he related the change in water equivalent to degree-days and arrived at a degree-day factor of 0.054 inch per degree-day using mean daily temperatures above freezing. Church 1/ compared snowmelt at Soda Springs, California with the "mean of temperatures above freezing during month" (equals one-half mean maximum temperature above 32°F for month) and arrived at a degree-day factor of 0.051 inch per degree-day. His data were for the month of April and covered the years 1936 through 1941. The foregoing factors represent melt rates per degree-day only for a given site and for a specified time of year. Obviously, there must be a considerable variation in point melt rates at different sites within a basin during the same period, and also a variation in the relationship with time of temperature indexes and point melt rates at a given site. Degree-day factors derived for just one site and for a single melt season are of limited significance in determining basin snowmelt. A more firm basis for the relationship between degree-days and point melt rates can be had from an analysis of the data from a single snow course over a number of years and from the analysis of a large number of different courses having different exposures. Such an analysis has been made using the considerable snow course data from the three snow laboratories for their years of operation. Mean degree-day factors for all the years of record of each snow course were computed. Also, a mean degree-day factor was determined for each of the snow laboratory basins for all of the years of record by averaging all snow courses within the basin. These data are summarized in the table below:

POINT MELT RATES: DEGREE-DAY FACTORS*

(Based on mean daily temperature)

Laboratory	Maximum Melt Station	Minimum Melt Station	Mean All Stations
CSSL	.128	.066	.106
UCSL	.131	.054	.090
WBSL	.108	.026	.060
Mean	.122	.049	.085

* inches of melt per degree-day above 32° F.

It is emphasized that the factors given in the table above are mean values for the several years of operation of each of the laboratories. Individual values for each year and for within-year periods exhibited a wide range of values for the individual snow courses. Deviations of the individual degree-day factors from the mean factor, computed for each snow-course, showed the more sheltered sites to have more constant factors. This bears out findings of the previous section on statistical determination of snowmelt indexes, which showed temperature indexes to be more accurate in determination of snowmelt in more heavily-forested areas. With regard to the degree-day factors for the maximum melt stations, it is pointed out that these factors are greater than those that would be observed at a horizontal, unforested site. Each of the maximum stations had a decided southerly exposure in addition to being unforested. As regards the minimum melt stations, the considerable difference in the magnitudes of these factors may be attributed to differences in forest cover. The WBSL site was especially conducive to low melt rates, being in a small clearing surrounded by high and dense trees. While solar radiation was effectively excluded from the site, longwave radiation loss from the snowpack was not reduced to the extent it would be under a solid forest canopy. The mean all-station factor for each of the laboratories represents an average in time as well as space. The mean degree-day factor gives a higher melt rate than the true melt rate over the laboratory area since they are for snow courses, which, on the whole, are in more open areas than the mean forest cover for the basin. The temperatures used in computing the degree-day factors were daily maximum and minimum values as determined by mercury-and-alcohol thermometers at the headquarters stations of each of the laboratories. A fixed lapse-rate correction of 3°F per 1000 feet was used to adjust this temperature to the elevation of the snow course being considered. Changes in water equivalent were means for the entire snow course, all points remaining snow covered throughout.

6-06.06 Temperatures. - While daily mean temperature is the most commonly used index of snowmelt, many other temperature indexes have been tried in an attempt to improve the index relationship (see Snyder, 9/ p.24). Most of these are based on daily maximum and daily minimum temperatures because of the wide availability and simplicity of use of these measures. Daily maximum temperature by itself, has been extensively used by the Snow Investigations as an index of snowmelt. When used by itself, maximum temperature has been found to be a more accurate index than daily mean temperature, and, in addition, is even simpler to use. Tabulated below are degree-day factors relating point snowmelt amounts to daily maximum temperatures in excess of 32°F. The factors are based on the same data used in computing the degree-day factors of the table of paragraph 6-06.05.

POINT MELT RATES: DEGREE-DAY FACTORS*

(Based on maximum daily temperature)

Laboratory	Maximum Melt Station	Minimum Melt Station	Mean All Stations
CSSL	.054	.029	.045
UCSL	.041	.020	.035
WBSL	.060	.015	.034
Mean	.052	.021	.038

* inches of melt per degree-day above 32° F.

The factors given here are roughly half those based on mean daily temperature. No general conclusions can be drawn regarding the relative accuracy of the two indexes even with the extensive data used in this study. The wide scatter in computed melt rates for individual snow courses obscured the relative effectiveness of the temperature parameters in computing melt.

6-06.07 It would seem that some measure of possible nocturnal snowpack heat deficit should appear in the temperature index, yet it is felt that the inclusion of minimum temperature at an equal weight with maximum temperature gives undue emphasis to this effect. On the other hand, the use of maximum temperature only, excludes this effect entirely. The use of an index such as $(2T_{max} + T_{min})/3$ would seem to offer a superior index to either maximum or mean temperature. However, in view of the slight difference in goodness of fit between maximum and mean temperatures, it appears unlikely such a refinement is warranted. Temperature indexes based on wet-bulb or dewpoint temperatures have also been suggested as being superior to dry-bulb temperatures. From the correlations presented in section 6-03, it may be seen that air temperature is a better index to snowmelt runoff than is vapor pressure. In view of the approximate linear relationship between dewpoint and vapor pressure over the limited ranges of these variables ordinarily encountered over melting snow, it follows that dewpoint temperature alone is not as good an index of snowmelt as is the dry-bulb temperature. Still another temperature index of snowmelt that has been considered is upper-air temperature. In favor of its use, it is argued that this is a temperature truly representative of the basin as a whole, being unaffected by local influences. However, studies by the Snow Investigations have shown that for small-to-moderate-sized drainage basins, upper-air temperatures by themselves are not sensitive-enough indicators of heat supply to represent melt variations adequately. Upper-air temperature variation is small compared to the daily variation of snowmelt. Much of the heat energy which goes to produce snowmelt is generated and consumed within the surface layers of the atmosphere and thus is not manifest in the upper-air temperature. It may be that for larger basins (greater than, say, 10,000 square miles), this may prove to be an adequate index, since the sensitivity required in smaller basins is not so important here.

6-06.08 <u>Temperature bases.</u> - The 32°F base used so far in this section is the most commonly used temperature base. While its use stems largely from a lack of knowledge concerning bases and their variations, it is, nevertheless, a good average base value for mean temperature indexes. Where maximum temperatures are used, higher bases and somewhat lower degree-day factors are generally indicated. The tendency is for the melt to become more independent of the temperature index as the amount of cover decreases. The temperature bases and degree-day factors for the point melt rates of this section, are given in figure 1, plate 6-4 for the extremes of forest cover. The graphs in this figure were determined by plotting the point melt data as functions of the average temperatures (both mean and maximum) for the period, for both open and forested snow courses, from data representing all three laboratories for several melt seasons. There is, of course, wide variation in melt rates between periods, because air temperature alone does not adequately represent the entire melt process, particularly for open sites. The curves do, however, represent the average relationship between air temperature and snowmelt during the active spring snowmelt period, for the extremes of forest cover. It should be noted that the temperature base decreases with decreasing forest cover, which conforms to the statistically-derived temperature bases described in section 6-03. The increase in temperature bases when maximum temperatures are used instead of mean temperatures should be noted. There is also a change in the degree-day factor. The below-freezing temperature bases for open sites do not necessarily indicate that snowmelt occurs with air temperatures below 32°F, but rather that there is included in the melt equations an average component of melt (principally solar radiation) which is independent of air temperature. Because of this, the melt equations are applicable only to average springtime melt for conditions as experienced at the snow laboratories and only for the approximate range of temperatures shown on the diagrams.

6-06.09 Basinwide snowmelt. - So far this section has been restricted to the consideration of point melt rates. Basinwide snowmelt rates present further complications. For one thing, variations in areal snow cover complicate the problem, while at the same time the progressive retreat of the snowline results in a change in the mean elevation of the snow-covered area. In addition, only a part of the snow-covered area may be contributing to snowmelt and the southerly exposed open areas become bare of snow first, leaving the more sheltered areas to produce the last of the snowmelt. As a result of these complications, basinwide snowmelt is most difficult to evaluate. While apparent point melt (ablation) rates can be determined from snow survey data and basin snowmelt runoff from hydrograph analysis, there is no single satisfactory measure of basinwide snowmelt. The mean change in water equivalent of all snow courses within the basin, in itself, is not adequate even for such densely-sampled basins as those of the snow laboratories, since snow courses are generally located in clearings and have higher than average basin melt rates. In order to evaluate basinwide snowmelt properly, it is necessary that a complete water balance be made for the area such that the snowmelt can be determined relative to the other causes of runoff. Moreover, it is necessary that the areal extent of the snowpack be known. The climatological data of chapter 2, the water balance of chapter 4, and the snow-cover depletion data of chapter 7, for each of the three laboratory basins afford the opportunity for such a determination. A study of basinwide snowmelt was made, based on these data using all of the years of record at each of the three snow laboratories. Basinwide snowmelt amounts corrected for areal snow cover, were related to temperature indexes (both mean and maximum temperatures), the temperatures being corrected to the mean height of the snow-covered area by means of a standard lapse rate of 3°F per 1000 feet. Results are given in the table below for the standard temperature base of 32°F. Monthly values for the months of April and May only are given. During June the areal extent of the snow cover was not well enough known to permit accurate computation.

BASINWIDE SNOWMELT RUNOFF: DEGREE-DAY FACTORS *

Basin	Mean temperature		Maximum temperature	
	April	May	April	May
CSSL	.089	.100	.024	.043
UCSL	.037	.072	.010	.031
WBSL	.039	.042	.021	.025

* inches of melt over the snow-covered area per degree-day above 32°F.

The values given in the above table are mean values for the several years of record at each of the laboratories. They reflect the general decrease in melt rates with increasing forest cover and also show the increase in the melt rates as the melt season progresses. Since, on the average, some of the heat transferred to the snowpack during April goes to ripen the pack, the resultant melt for this month is less than for subsequent months. Other things also influence this change in factor with time; they will be discussed in a subsequent paragraph. For comparison with the above, reference is made to a study of basinwide snowmelt by Horton. 7/ He found melt rates varying from 0.088 to 0.058 inches per degree-day (mean temperature greater than 32°F) for thinly forested and fully forested areas in western Pennsylvania during March and April.

6-06.10 Using temperature bases for snowmelt, appropriate to the degree of forest cover as previously discussed, degree-day factors for basinwide snowmelt were also computed for the three snow-laboratory areas. These were as follows:

BASINWIDE SNOWMELT: TEMPERATURE BASES AND DEGREE-DAY FACTORS *

Basin	Mean temperature			Maximum temperature		
	Base °F	Degree-day factor April	Degree-day factor May	Base °F	Degree-day factor April	Degree-day factor May
CSSL	26	.036	.062	29	.020	.038
UCSL	32	.037	.072	42	.109	.064
WBSL	32	.039	.042	42	.046	.046

* inches of melt over the snow-covered area per degree-day above indicated temperature base

The same average temperature base was used for both months for each of the laboratories since the variation with time in the base value is slight. As before, the temperatures were corrected to the mean height of the snow-covered area.

6-06.11 Degree-day factors. - In the foregoing paragraphs of this section, degree-day factors for both point and basinwide melt rates have been given. The variation in these factors with density of forest cover and with time has been shown. Here the reasons for the variations will be discussed and some general conclusions will be drawn concerning them. Regarding the increase in the degree-day factors for basinwide snowmelt from April to May, it is to be pointed out that this increase is largely fictional, the result of using monthly means in the computation. The typical April has alternate periods of snowmelting and deposition of new snow (cf. tables 4-1, 4-2, and 4-3). Even though corrections are made for additions of snow in computing the total ablation for the month, in each instance of new snowfall, this snow must first be ripened before it can be melted. Considerable heat is thus consumed which would otherwise go to produce snowmelt, and the resultant degree-day factor is accordingly lowered. During May, the occurrences of new snowfall are much less frequent; hence the degree-day factor is higher. If only periods of actual melting were considered, rather than monthly totals, the difference between the degree-day factors for the two months (April and May) would be much less. This was done in the determination of degree-day factors for point melt rates for snow known to be ripe. Only periods of active melt between two snow surveys were used in arriving at the degree-day factors. As a result, there was no apparent seasonal increase in these factors. While it would seem that there should be some increase due to the increasing quantities of solar radiation through the season, apparently this increase was adequately reflected by the temperature index. This is discussed in the next paragraph.

6-06.12 The place and method of air temperature measurement has a secondary effect upon the variation of the degree-day factor with time. If temperatures are measured at or near the basin outlet, as is quite often the case, it is usual for the ground to become bare in the vicinity of the temperature station fairly early in the melt season when considerable snow cover still remains on the basin. This change from snow-covered to snow-free ground results in a measureable increase in air temperature over what it would have been, had the ground remained snow covered throughout. This effect is further heightened if temperature measurements are made at a fixed level with respect to the ground surface. As the melt season progresses, the height of the temperature measurement with respect to the snow surface increases until the ground is bare of snow. Because of the usual inversion over the snowpack, this results in a greater increase in temperature with time than would have been the case had no snow existed or had the thermometer remained at a fixed height relative to the snow surface. In this situation, the additional temperature gain tends to offset any actual seasonal increase in the degree-day factor. For temperature index stations that remain snow-free throughout the melt season this effect is, of course, non-existent.

6-06.13 The causes of the normal increase in the degree-day factors with time are (1) increasing ripeness of the snowpack, (2) decrease of snow-surface albedo, (3) depletion of snow cover, (4) increase in insolation, (5) increase in percentage of sheltered snow-covered area, and (6) the increase in the mean elevation of the snow-covered area. These are even more important in larger basins having considerable ranges in elevation than in the examples of the snow laboratories. Failure to make adjustment for these variations results in a net decreasing degree-day factor with time. If corrections are made for the effects of snow-cover depletion and mean elevation of the snow-covered area, such minor effects as the increase in insolation and the progressive increase in shelter of the residual snow cover as the melt season progresses, tend to cancel one another, making for a more constant degree-day factor for basinwide snowmelt.

6-06.14 The usual temperature increase through the melt season suggests an empirical scheme which may be used to approximate the observed increase in degree-day factors with time as the melt season progresses. If, for example, the degree-day factor is taken as $K(T_a - T_b)^n$, then

$$M = K(T_a - T_b)^{(n + 1)} \tag{6-16}$$

The value of n may be determined from analysis of snowmelt data. This form of the melt equation has been used, with good results, in several studies of melt rates, using $n = 0.25$. Not only does it approximate the seasonal increase in degree-day factors through the seasonal temperature increase, but it allows also for such decreases in the factors as usually follow springtime frontal passages. While the decrease in degree-day factors may be, physically, the result of the high albedo and thermal quality of newly-deposited snow, they are empirically reflected in the degree-day factor in consequence of the usual colder airmass following the frontal passage. At the present writing, this method has not been extensively tested, and it is not known whether it is generally applicable to the computation of snowmelt. Consequently, it should be used only in the sense in which it is presented here, namely, as an empirical device for increasing the degree-day factor with time in situations where such an increase has been shown to exist.

6-06.15 Snowmelt runoff. - Actually, as was pointed out previously, measurements of basinwide snowmelt are not ordinarily available. Lacking a water balance, all that are available for most basins are point melt rates and basin snowmelt-runoff rates. Hence, for a measure of basinwide melts, recourse is usually made to basin snowmelt runoff. For large drainage basins, it is impossible to separate daily increments of flow for purposes of analysis as was done in the preceding section of this chapter for the small laboratory areas and the moderate sized Boise River basin above Twin Springs, Idaho, where a marked diurnal snowmelt rise and fall made identification of daily melts possible. For large basins it is possible to determine the snowmelt runoff per degree-day by plotting double mass curves of snowmelt runoff as a function of degree-days. This technique was used by Snyder 9/ in studies of the Susquehanna River Basin, by Miller 8/ in the Missouri River Basin and by Wilson 10/ for

Gardiner River, Wyoming. Inasmuch as observed mean daily runoffs rather than the actual daily generated runoff are accumulated, during periods of increasing flows, too flat a curve results and too low a degree-day factor, while during periods of receding flows the reverse is true. Periods of sub-freezing temperatures result in irregularities in the curve since the recession of the runoff values continues even though no temperature index values are accumulated. However, if the melt season is taken as a whole, such irregularities are slight and the general relationship between snowmelt runoff and the temperature index is readily apparent. Corrections must be made in the runoff hydrograph, of course, for any rainfall runoff that occurs. As an example of the double-mass curve approach to the determination of the degree-day factor for snowmelt runoff, computations of the factor were made for the 1949 snowmelt season at the UCSL. Both maximum and mean temperature indexes were used; the previously determined temperature bases were used. The results are given in figures 2 and 3 of plate 6-4. Shown on each of the figures is: (a) the snowmelt runoff in inches over the basin (not corrected for contributing area) as a function of the unadjusted temperature index; (b) the snowmelt runoff in inches depth over the contributing area as a function of the unadjusted temperature index; and (c) the snowmelt runoff corrected for contributing area as a function of the temperature index adjusted to the mean height of the contributing area.

6-06.16 The foregoing method of double-mass curve analysis was presented only as an illustrative example of the method, giving data for one year only from one of the snow laboratories. Other years and other laboratories were not analyzed because the relationships between the temperature indexes and the snowmelt runoff have already been determined by the statistical studies of the preceding sections of this chapter. The two approaches in computing degree-day factors for snowmelt runoff for this one year at the UCSL gave results which are in close agreement. In this analysis, as in the basinwide snowmelt analysis based on the monthly water balance data, it is to be emphasized that the early-season melt rates are of little significance, reflecting as they do the frequency and quantity of new snowfall. Moreover, in the case of snowmelt runoff, additional snowmelt water goes to satisfy soil moisture and depression storage (pondage) requirements early in the melt season; this fact further affects the relationship between the temperature index and the snowmelt runoff.

6-07. SUMMARY - THE GENERALIZED BASIN SNOWMELT EQUATION

6-07.01 Basin snowmelt coefficients are most accurately determined by statistical analysis of past runoff records correlated with appropriate indexes of snowmelt. Statistically derived coefficients relate snowmelt to the conditions of observation of meteorologic variables at fixed points, give proper weight to features of the physical environment of the basin (i.e., percent of forest cover, slope, orientation, loss, etc.), and allow relatively unimportant melt processes to be indexed by major melt parameters. The application of statistically derived coefficients from daily snowmelt runoff records is particularly suited to short-term streamflow forecasting, for cases where adequate past record is available for study and reconstitution. Logically, the coefficients so derived apply only to the conditions for which they were developed. For many cases, particularly in connection with design floods, it is impractical to derive basin melt coefficients, and a general snowmelt equation is required. In order to meet this requirement, the rational analysis of snowmelt presented in chapter 5 may be combined with the statistical weightings of the variables developed in this chapter to arrive at simplified general snowmelt equations which are applicable to all conditions of environment over project-sized basins. The general solution for snowmelt during rain on snow, as described in section 6-04, serves as a guide to a more general snowmelt equation applicable to rain-free conditions. The principal requirement of the general equation is to express snowmelt in terms of ordinarily available meteorologic data which most reliably represent the physical processes of melt. Simplifying assumptions which are well within the accuracy of the derived melt coefficients may be applied in order to obtain a workable snowmelt formula.

6-07.02 The following equations for rain-free periods have been developed on the basis of the above-stated requirements, for varying conditions of forest environment:

Heavily forested area:

$$M = 0.074\ (0.53T'_a + 0.47T'_d) \qquad (6\text{-}17)$$

Forested area:

$$M = k(0.0084v)\ (0.22T'_a + 0.78T'_d) + 0.029T'_a \qquad (6\text{-}18)$$

Partly forested area:

$$M = k'(1 - F)(0.0040\ I_i)\ (1 - a) + k(0.0084v)(0.22T'_a + 0.78T'_d) + F(0.029T'_a) \quad (6\text{–}19)$$

Open area:

$$M = k'(0.00508\ I_i)\ (1 - a) + (1\text{-}N)(0.0212T'_a - 0.84) + N(0.029T'_c) + k(0.0084v)(0.22T'_a + 0.78T'_d) \quad (6\text{–}20)$$

where:

M is the snowmelt rate in inches per day.

T'_a is the difference between the air temperature measured at 10 feet and the snow surface temperature, in $^{\circ}$F.

T'_d is the difference between the dewpoint temperature measured at 10 feet and the snow surface temperature, in $^{\circ}$F.

v is the wind speed at 50 feet above the snow, in miles per hour.

I_i is the observed or estimated insolation (solar radiation on horizontal surface) in langleys. (See figures 4, 5 and 6, plate 5-1.)

a is the observed or estimated average snow surface albedo. (See figures 3-4, plate 5-2 for estimating albedo of the snow.)

k' is the basin shortwave radiation melt factor. It depends upon the average exposure of the open areas to shortwave radiation in comparison with an unshielded horizontal surface. (See figure 6, plate 5-1, for seasonal variation of k' for North and South 25° slopes.)

F is an estimated average basin forest canopy cover, effective in shading the area from solar radiation, expressed as a decimal fraction.

T'_c is the difference between the cloud base temperature and snow surface temperature, in ^{o}F. It is estimated from upper air temperatures or by lapse rates from surface station, preferably on a snow-free site.

N is the estimated cloud cover, expressed as a decimal fraction.

k is the basin convection-condensation melt factor, as defined in paragraph 6-04.12. It depends on the relative exposure of the area to wind.

The melt coefficients given in the above equations represent daily melt amounts in inches. For those cases where a convection-condensation term with wind is included in the melt expression, it may be necessary to subdivide the day into smaller time increments where there is marked variation in both wind and temperature or dewpoint. The coefficients also express melt for a ripe snowpack (isothermal at $0^{o}C$ and with 3 percent initial free water content - see chap. 8). Except that they account for loss by transpiration from forested areas, the melt coefficients represent the actual melt of the snowpack averaged over a basin area, expressed as daily ablation in inches of water equivalent. The equations are based on linear approximations between saturation air-vapor pressure and dewpoint, and also on longwave radiation as a linear function of the temperature of the radiating surface. These approximations provide values which are close to the theoretical amounts for the range of temperatures and dewpoints experienced in snow hydrology, as shown in figures 1 and 5, plate 6-2. Functions involving air temperature and dewpoints are expressed as the difference between snow surface temperature and the value for a given level of observation. For a melting snowpack, subfreezing snow-surface temperatures are experienced usually only as the result of nighttime cooling in open areas (see chapter 8). Therefore, except under this condition, the temperature and dewpoint values are equal to increments above $32^{o}F$. The following paragraphs summarize the derivation of the general melt equations.

6-07.03 The snowmelt runoff equation for heavily forested areas directly evaluates melt by convection, condensation, and longwave radiation. Implicit in the relationship is an average wind (estimated to be between 2 and 3 miles per hour at the 1-foot level) which is required for the transfer of heat and water vapor to the snow by turbulent exchange in the atmosphere. The air temperature term includes the evaluation of heat transfer by longwave radiation and convection. The proportional effects of convection and condensation and their appropriate melt coefficients were derived as explained in paragraph 6-04.07. Direct melt by absorbed shortwave radiation is assumed to be equal to the basin loss of water by forest transpiration, as was explained in Research Note 19.

6-07.04 The general melt equation for less densely forested areas is derived in a similar manner, except that convection and condensation are evaluated separately from longwave radiation in order to include a wind variable. The value of k, the basin coefficient, depends on the conditions of measurement of wind with respect to the basin average, and its value is 1.0 for plains areas with no forest cover. It may be slightly greater than 1.0 for exposed mountain ridges, and for heavily forested areas it approaches a minimum value of 0.2. The 50-foot level wind value for forested areas is assumed to be the average wind in an open area resulting from the general airmass circulation prevailing at the time.

6-07.05 For partly forested areas, a term for shortwave radiation exchange in the unforested portions is included in the melt equation. The theoretical melt coefficient for absorbed shortwave radiation melt is reduced from 0.00508 to 0.0040, to account partly for longwave radiation loss in the open and partly for loss by evapotranspiration from the forest. The effective forest canopy cover, F, represents the average percent of the basin shaded by the forest from direct solar radiation, expressed as a decimal fraction. For partly forested basins, longwave loss in the open is considered to be adequately expressed by the reduction of the shortwave radiation melt coefficient for the open portions of the area.

6-07.06 The expression for snowmelt in open areas becomes more complex because of the requirement for direct evaluation of longwave radiation exchange. Incident shortwave radiation is presently observed at approximately 90 stations in the United States. For areas where observations are not available, shortwave radiation may be estimated by duration of sunshine or diurnal temperature rises. Longwave radiation, on the other hand, is not regularly measured, and although many expressions have been developed to evaluate it from both surface and upper-air meteorologic data, they are mostly too complex or cumbersome to incorporate in a generalized melt expression. From observations of longwave radiation made at CSSL, it has been found that for the condition of clear skies, the downward longwave radiation from the atmosphere over snow can be expressed simply as 0.75 of the theoretical blackbody radiation corresponding to surface air temperature measured at instrument height. The limited variation of the water-vapor content of the air usually experienced over snow, does not produce significant changes in downward longwave radiation. Therefore, in the general melt equation, with clear skies (N equal to 0), the net longwave radiation loss is expressed simply as a linear function of the surface air temperature. The net longwave radiation exchange for cloudy skies is expressed as the theoretical blackbody radiation, as approximated by a linear relationship, for the difference in temperature between the snow surface and the cloud base; cloud base temperatures may be determined either from soundings made by radiosonde, or by applying appropriate temperature lapse-rate corrections to known surface air temperatures and cloud heights. Estimates of longwave exchange under cloudy skies by this method are considered to be realistic for low or middle clouds. Convection and condensation melt for open areas is estimated by a general equation as discussed in section 6-04. It is not considered necessary to evaluate the effects of elevation in the convection term (due to decreased air density with elevation), because of the relatively small order of magnitude of this correction in comparison with the other components of melt.

6-08. REFERENCES

1/ CHURCH, J. E., "The melting of snow," Proc. Central Snow Conf., Vol. I, Dec. 1941, pp. 21-32.

2/ CLYDE, George D., "Snow-melting characteristics," Bul. 231 (Tech), Utah Agric. Exper. Station, August 1931.

3/ CORPS OF ENGINEERS, Walla Walla District, Preliminary Flood Regulation Manual, For Boise River Reservoirs, Idaho, 15 Feb 1952.

4/ FALCKENBERG, G., "The infrared absorption constants of some meteorologically important substances," (Absorption-skonstantan einiger meteorologisch wichliger, Korper fur infrarote wellen; test in German). (Meteoro. Zeitschr.) Vol. 45, September 1948, pp 334-337.

5/ HAMMON, WEISS, AND WILSON, "Insolation as an empirical function of daily sunshine duration," Mon. Wea. Rev., Vol. 82, No. 6, pp. 141-146

6/ HORTON, Robert E., "The melting of Snow," Mon. Wea. Rev., Vol. No. 43, No. 12, Dec. 1915, pp. 599-605.

7/ HORTON, Robert E., "Infiltration and runoff during the snow-melting season, with forest cover," Trans. Amer. Geophys. Union, Vol. 26, Pt. 1, Aug. 1945, pp. 59-68.

8/ MILLER, S. A., "Some snow-melt runoff characteristics," Trans. Amer. Geophys. Union, Vol. 31, No. 5, Pt. 1, Oct. 1950, pp. 741-45.

9/ SNYDER, F. F., Cooperative Hydrologic Investigations, Part II, Commonwealth of Pennsylvania, Harrisburg, 1939 (mimeo.).

10/ WILSON, W. T., "Some factors in relating the melting of snow to its causes," Proc. Central Snow Conf., Vol. I, December 1941, pp. 33-41.

TABLE 6-1

SUMMARY OF RESULTS OF REGRESSION ANALYSIS

SNOWMELT RUNOFF FROM UNFORESTED AREA

Lower Meadow Lysimeter, CSSL

1954

EQUATION NUMBER	DATA	SNOWMELT RUNOFF COEFFICIENTS											REGRESSION CONSTANT	STANDARD ERROR OF ESTIMATE	COEFFICIENT OF DETERMINATION
		Radiation		Convection and Condensation											
		I_{net} Shortwave	R_{net} Longwave	Convection parameter		Condensation parameter		Wind	Maximum temperature		Mean temperature				
		X_1	X_2	X_3	Base °F	X_4	Base mb	X_5	X_6	Base °F	X_9	Base °F	c	$s_{y.x}$	D
(1)	1954 hourly	0.00544	0.00445	0.00011	T_s 2/	0.00105	e_s 2/						-0.10	0.05	0.97
(2)	daily	0.00617	0.00758	-0.000038	78	-0.00411	33.1	0.00139					0.71	0.07	0.95
(3)	daily	0.00747	-0.00183 1/	0.00019	32	0.00056	6.1	-0.00939					0.31	0.18	0.67
(4)	daily	0.00576	0.00649										0.07	0.07	0.94
(5)	daily	0.00504											-0.03	0.18	0.65
(6)	hourly			0.00060	T_s	0.00238	e_s						0.37	0.24	0.46
(7)	hourly			0.00035	T_s								0.75	0.29	0.16
(8)	hourly					0.00116	e_s						1.22	0.30	0.09
(9)	daily								0.0334	22			-0.73	0.22	0.23
(10)	daily										0.0364	8	-0.31	0.29	0.08

Notes:
Definitions of variables and units are given in paragraph 6-03.02.

1/ Incident shortwave radiation substituted for longwave radiation loss in the open.

2/ T_s and e_s are temperature and vapor pressure of the snow surface.

TABLE 6-2

SUMMARY OF RESULTS OF REGRESSION ANALYSIS

SNOWMELT RUNOFF FROM PARTLY FORESTED AREA

Castle Creek - CSSL

1954, 1952, 1951

EQUATION NUMBER	DATA	SNOWMELT RUNOFF COEFFICIENTS											REGRESSION CONSTANT	STANDARD ERROR OF ESTIMATE	COEFFICIENT OF DETERMINATION
		Radiation		Convection and Condensation											
		I_{net} Shortwave	I_i Index of longwave loss in open	Convection parameter		Condensation parameter		Wind	Maximum temperature		1800 vapor pressure				
		X_1	X_2	X_3	Base °F	X_4	Base mb	X_5	X_6	Base °F	X_7	Base mb	c	$s_{y.x}$	D
(1)	1954 hourly	0.00312	0.00158 1/	0.00018	T_s	0.00148	e_s						-0.13	0.06	0.93
(1-a)	1954 hourly	0.00462	-0.00101	0.00026	T_s	0.00179	e_s						0.05	0.05	0.94
(2)	daily	0.00508	-0.00176	0.00038	37	0.00118	7.5	-0.02314					0.36	0.09	0.84
(3)	1952 hourly	0.00686	-0.00264	0.00009	T_s	0.00079	e_s						0.37	0.15	0.75
(4)	daily	0.00643	-0.00291	0.00013	34	0.00053	6.6	-0.00774					0.76	0.18	0.62
(5)	1951 hourly	0.00466	-0.00154	0.00006	T_s	0.00062	e_s						0.26	0.10	0.93
(6)	daily	0.00550	-0.00220	0.00008	28	0.00055	5.1	-0.00506					0.27	0.10	0.92
(7)	1954 hourly	0.00285		0.00024	T_s	0.00187	e_s						-0.22	0.07	0.90
(8)	daily	0.00242		0.00026	35	0.00113	6.9	-0.01688					-0.11	0.11	0.72
(9)	1952 hourly	0.00289		0.00012	T_s	0.00097	e_s						-0.21	0.18	0.65
(10)	daily	0.00108		0.00023	38	0.00050	7.6	-0.01247					0.32	0.17	0.64
(11)	1951 hourly	0.00224		0.00011	T_s	0.00089	e_s						-0.19	0.11	0.90
(12)	daily	0.00173		0.00020	31	0.00087	6.0	-0.01149					-0.23	0.13	0.86
(13)	1954 daily			0.00051	36	0.00141	7.2	-0.02835					-0.22	0.15	0.47
(14)	1952 daily			0.00030	41	0.00035	8.6	-0.01525					0.74	0.17	0.59
(15)	1951 daily			0.00037	33	0.00099	6.4	-0.01856					0.08	0.15	0.83
(16)	1954 daily								0.01154	8	0.04757	1.9	-0.18	0.20	0.14
(17)	1952 daily								0.02327	16	0.04467	2.9	-0.51	0.18	0.58
(18)	1951 daily								0.02239	27	0.06403	4.8	-0.91	0.20	0.67
(19)	1954 daily	0.00315											0.06	0.15	0.51
(20)	1952 daily	0.00179											0.36	0.24	0.22
(21)	1951 daily	0.00382											-0.44	0.19	0.71
(22)	1954 daily			0.00029	38			-0.01111					0.53	0.21	0.04
(23)	1952 daily			0.00032	44			-0.01422					0.74	0.18	0.54
(24)	1951 daily			0.00042	42			-0.01757					0.34	0.21	0.64
(25)	1954 daily					0.00100	1.2	-0.00120					0.52	0.19	0.20
(26)	1952 daily					0.00051	9.9	-0.00503					1.24	0.25	0.18
(27)	1951 daily					0.00133	4.0	-0.00534					0.43	0.30	0.31
(28)	1954 daily								0.01049	22			0.02	0.21	0.01
(29)	1952 daily								0.02739	17			-0.46	0.18	0.54
(30)	1951 daily								0.03045	30			-0.91	0.23	0.58

Notes: Definitions of variables and units are given in paragraph 6-03.02.

1/ Measured R_{net} (equal to $G_{net} - I_{net}$) is used rather than its index, I_i in equation 1.

TABLE 6-3

SUMMARY OF RESULTS OF REGRESSION ANALYSIS

SNOWMELT RUNOFF FROM FORESTED AREA

Skyland Creek - UCSL

1950, 1949

EQUATION NUMBER	DATA	SNOWMELT RUNOFF COEFFICIENTS											REGRESSION CONSTANT	STANDARD ERROR OF ESTIMATE	COEFFICIENT OF DETERMINATION
		Radiation		Convection and Condensation											
		I_{net} Shortwave	I_l Index of longwave loss in open	Convection parameter		Condensation parameter		Wind	Maximum temperature		1400 vapor pressure				
		X_1	X_2	X_3	Base °F	X_4	Base mb	X_5	X_6	Base °F	X_7	Base mb	c	$s_{y.x}$	D
(1)	1949 daily	0.00056	0.00041	0.00028	41	0.00155	8.8	-0.02521					0.11	0.07	0.88
(2)	1949 daily	0.00004		0.00029	41	0.00141	8.8	-0.02420					0.15	0.07	0.89
(3)	1950 daily			0.00040	39	0.00128	8.0	-0.02536					0.27	0.17	0.81
(4)	1949 daily			0.00029	42	0.00139	8.9	-0.02441					0.16	0.07	0.90
(5)	1950 daily								0.03629	36	0.07849	7.2	-1.87	0.19	0.77
(6)	1949 daily								0.01962	38	0.07225	7.6	-1.28	0.13	0.60
(7)	1949 daily	0.00380	-0.00169										0.36	0.18	0.29
(8)	1949 daily	0.00130											0.15	0.18	0.23
(9)	1950 daily			0.00048	56			-0.02647					0.52	0.25	0.59
(10)	1949 daily			0.00037	54			-0.02015					0.22	0.11	0.73
(11)	1950 daily					0.00176		-0.00469					0.28	0.31	0.36
(12)	1949 daily					0.00220	6.5	-0.01425					0.39	0.15	0.48
(13)	1950 daily								0.04254	43			-1.84	0.24	0.62
(14)	1949 daily								0.02497	44			-1.02	0.14	0.54
(15)	1950 daily										0.11641		0.03	0.32	0.33
(16)	1949 daily										0.14076	3.6	-0.50	0.17	0.35

Note:
Definitions of variables and units are given in paragraph 6-03.02.

TABLE 6-4

SUMMARY OF RESULTS OF REGRESSION ANALYSIS

SNOWMELT RUNOFF FROM HEAVILY FORESTED AREA

Mann Creek, WBSL

1951, 1950, 1949

EQUATION NUMBER	DATA	SNOWMELT RUNOFF COEFFICIENTS								REGRESSION CONSTANT	STANDARD ERROR OF ESTIMATE	COEFFICIENT OF DETERMINATION
		Maximum Temperature		Max. Vapor Pressure		Mean Temperature		Mean Vapor Pressure				
		X_6	Base °F	X_7	Base mb	$\bar{X}_6$	Base °F	$\bar{X}_7$	Base mb	c	$s_{y.x}$	D
(1)	April 1951	0.027	36	0.045	7.2					-1.30	0.14	0.77
(2)	May 1950	0.031	35	0.070	6.9					-1.57	0.12	0.80
(3)	May 1949	0.030	37	0.057	7.5					-1.54	0.14	0.87
(4)	April 1949	0.027	34	0.094	6.6					-1.53	0.16	0.74
(5)	April 1951					0.024	29	0.085	5.3	-1.15	0.13	0.80
(6)	May 1950					0.036	30	0.112	5.6	-1.70	0.18	0.77
(7)	May 1949					0.039	33	0.094	6.3	-1.88	0.10	0.93
(8)	April 1949					0.048	31	0.147	5.8	-2.34	0.14	0.80
(9)	April 1951	0.034	38							-1.29	0.16	0.70
(10)	May 1950	0.038	35							-1.33	0.19	0.72
(11)	May 1949	0.030	37							-1.11	0.19	0.73
(12)	April 1949	0.034	35							-1.19	0.17	0.70
(13)	April 1951					0.040	30			-1.20	0.18	0.65
(14)	May 1950					0.043	28			-1.20	0.23	0.61
(15)	May 1949					0.044	31			-1.36	0.22	0.64
(16)	April 1949					0.051	30			-1.53	0.22	0.49

Note: Definitions of variables and units are given in paragraph 6-03.02.

TABLE 6-5

SUMMARY OF RESULTS OF REGRESSION ANALYSIS

SNOWMELT RUNOFF, BOISE RIVER

above Twin Springs, Idaho

1955, 1954

1/ EQUATION NUMBER	SNOWMELT RUNOFF COEFFICIENTS											REGRESSION CONSTANT	STANDARD ERROR OF ESTIMATE	COEFFICIENT OF DETERMINATION
	Radiation				Convection and Condensation									
	I_{net} Shortwave	I_i Index of longwave loss in open	LW loss in open $f(T_{700}, T_{min})$	G_{net} Allwave $X_1 + X_9$	Convection parameter	Condensation parameter	Wind	Maximum Temp.	700 mb Temp.	Dewpoint	Temperature Base °F			
	X_1	X_2	X_9	X_{10}	X_3	X_4	X_5	X_6	X_7	X_8	T_b	c	$s_{y.x}$	D
(1)	0.00123	0.00024			0.00245	0.00005	-0.163				66.5	-0.31	0.16	0.82
(2)					0.00313	-0.00024	-0.195				67.0	0.11	0.16	0.82
(3)	0.00153				0.00234		-0.158				67.4	-0.17	0.15	0.83
(4)					0.00303		-0.196				64.7	0.11	0.16	0.82
(5)	0.00451										----	-0.86	0.24	0.58
(6)	0.00102				0.00268				0.0124	-0.0150	66.1	-0.30	0.14	0.87
(7)	0.00239							0.0274			83.5	-2.29	0.14	0.86
(8)								0.0375			62.4	-2.34	0.19	0.75
(9)	0.00237		0.00228					0.0200			81.4	-1.63	0.12	0.89
*(10a)				0.00233				0.0236			76.7	-1.81	0.12	0.89
*(10b)1954				0.00271				0.0267			80.2	-2.14	0.10	0.87
*(10c)'54&55				0.00238				0.0245			76.9	-1.89	0.11	0.90
(11) 2/				0.00227				0.0267 2/			75.0	-2.00	0.14	0.86

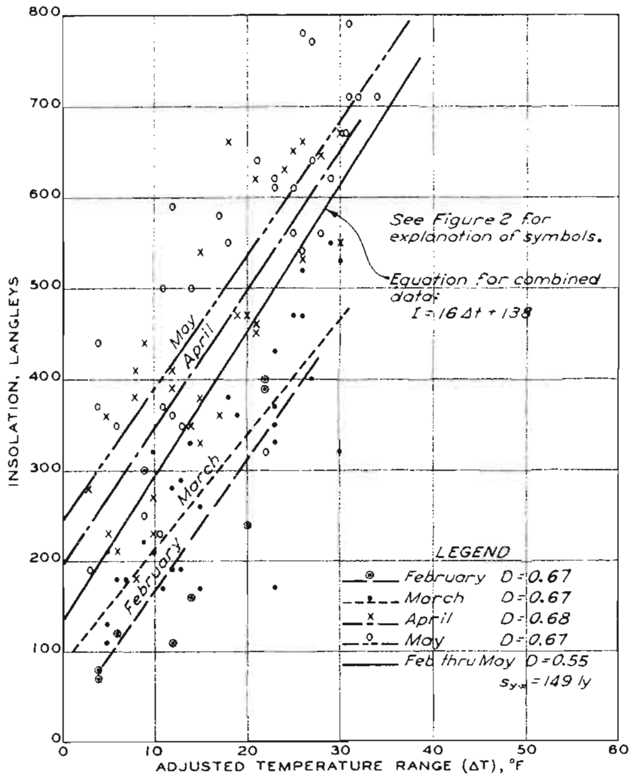
Notes:

Definitions of variables and units are given in paragraph 6-05.10

1/ All equations for 1955 data unless otherwise indicated.

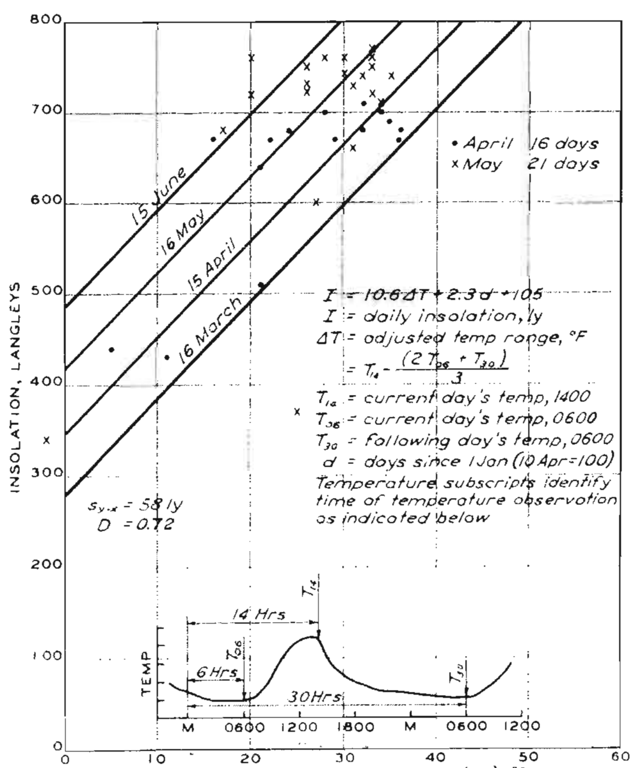
2/ For reconstitution of hydrograph in Chap. 9. (X_6 = Idaho City Max. Temp.).

* In 10a, n = 26; in 10b, n = 14; and in 10c, n = 40.



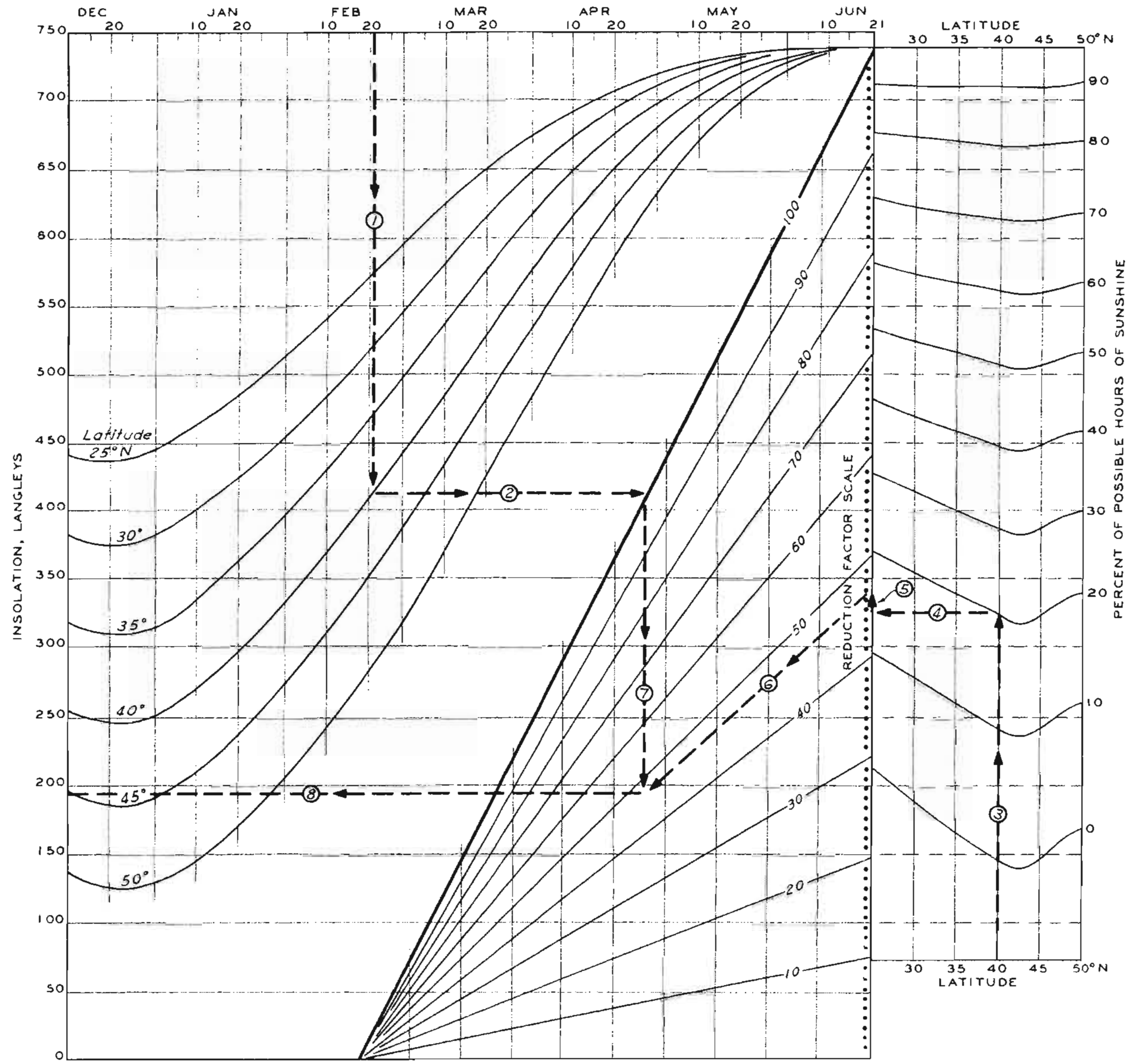
ESTIMATION OF DAILY INSOLATION FROM 6 AM AND 2 PM TEMPERATURE VALUES AT UCSL FEB-MAY 1947

FIGURE 1



ESTIMATION OF DAILY INSOLATION FROM 6 AM AND 2 PM TEMPERATURE VALUES AND DATE, CSSL 14 APRIL-23 MAY 1954

FIGURE 2



Notes:

1. The sample shown by dashed lines with arrows estimates the daily total insolation at a station at latitude 40°N, on February 21 with 20 percent possible sunshine: Step 1 — enter the graph at February 21 and move downward to the curve labelled 40°N. Step 2 — thence move horizontally to establish a point at the intersection with the heavy line. Step 3 — enter lower right of graph at 40°N, proceeding upward to 20 percent of possible hours of sunshine. Step 4 — thence move horizontally toward the left to the vertical line through 25°N. Step 5 — here the seasonal correction (+2) from the table at the right is added on the reduction factor scale. Step 6 — thence move downward toward the convergence point. Step 7 — from the reference point of step 2 move downward to an intersection with the line extended in step 6. Step 8 — from the point of intersection of these two lines move horizontally to the insolation scale to read the estimated value, 195 langleys per day.
2. For use between June 21 and December 21, the curves are symetrical about June 21.

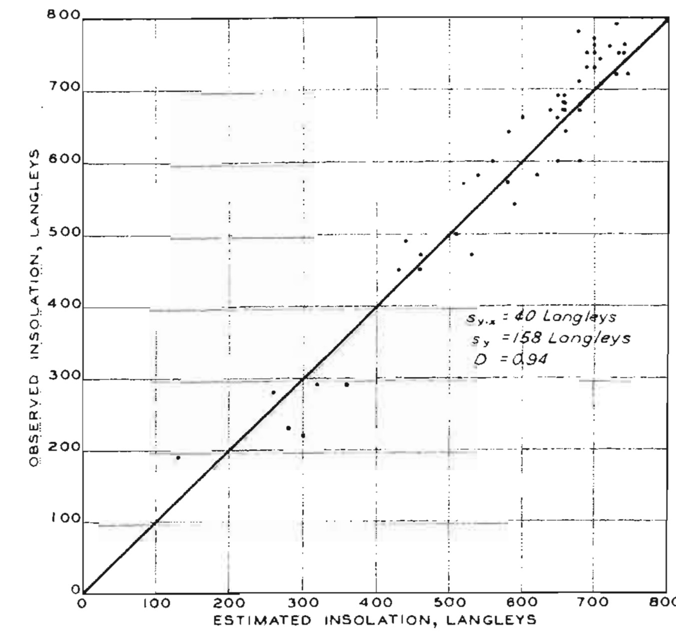
SEASONAL CORRECTION TO REDUCTION FACTOR

Month	Percent Of Possible Sunshine							
	0	10	20	30	40	50	60	70
Jan	+4	+3	+3	+2	+2	+2	+1	+1
Feb	+3	+3	+2	+2	+2	+1	+1	+1
Mar	-1	-1	-1	-1	-1	0	0	0
Apr	-2	-2	-1	-1	-1	-1	-1	0
May	-4	-3	-3	-2	-2	-2	-1	-1
Jun	-5	-4	-4	-3	-2	-2	-2	-1
Jul	-5	-4	-3	-3	-2	-2	-2	-1
Aug	-4	-3	-3	-2	-2	-2	-1	-1
Sep	-2	-2	-1	-1	-1	-1	-1	-1
Oct	0	0	0	0	0	0	0	0
Nov	+2	+2	+1	+1	+1	+1	+1	0
Dec	+4	+3	+3	+2	+2	+2	+1	+1

ESTIMATION OF DAILY INSOLATION FROM LATITUDE, DATE, AND DURATION OF SUNSHINE

(AFTER HAMON, WEISS, AND WILSON [5])

FIGURE 3



OBSERVED VS ESTIMATED DAILY INSOLATION AT BOISE, IDAHO — MAY AND JUNE 1955

(ESTIMATED FROM DURATION OF SUNSHINE AND FIGURE 3)

FIGURE 4

Notes:

1. Insolation is the quantity of solar energy received on a horizontal surface, expressed in langleys (gm-cal/cm²) for a specified time. All amounts in figures on this plate are daily totals.
2. The coefficient of determination, D, is computed from the formula

$$D = \frac{s^2_y - s^2_{y \cdot x}}{s^2_y}$$

where s^2_y is the variance of observed variable

$s^2_{y \cdot x}$ is the mean square error of estimate

SNOW INVESTIGATIONS SUMMARY REPORT

SNOW HYDROLOGY

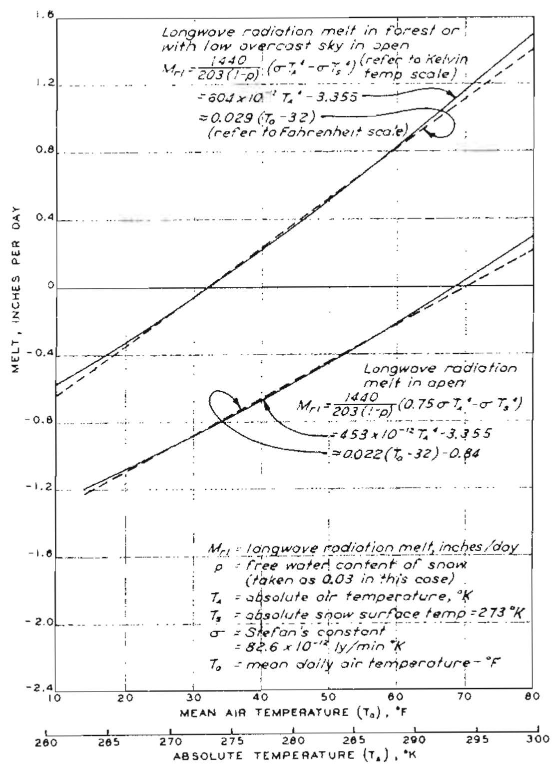
SNOWMELT INDEXES

ESTIMATION OF SOLAR RADIATION

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U.S. ARMY

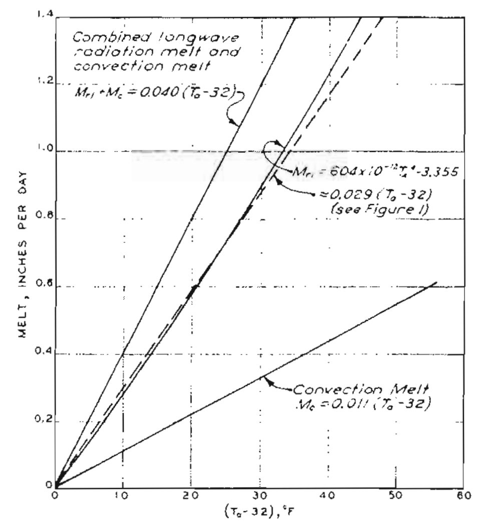
PREPARED C.D.-S.M. SUBMITTED D.M. TO ACCOMPANY REPORT DATED 30 JUNE 1956
DRAWN B.E. APPROVED D.M.R.

PD-20-25/36



LONGWAVE RADIATION MELT,
WITH LINEAR APPROXIMATION

FIGURE 1



Note:
Constants from Research Note 11.

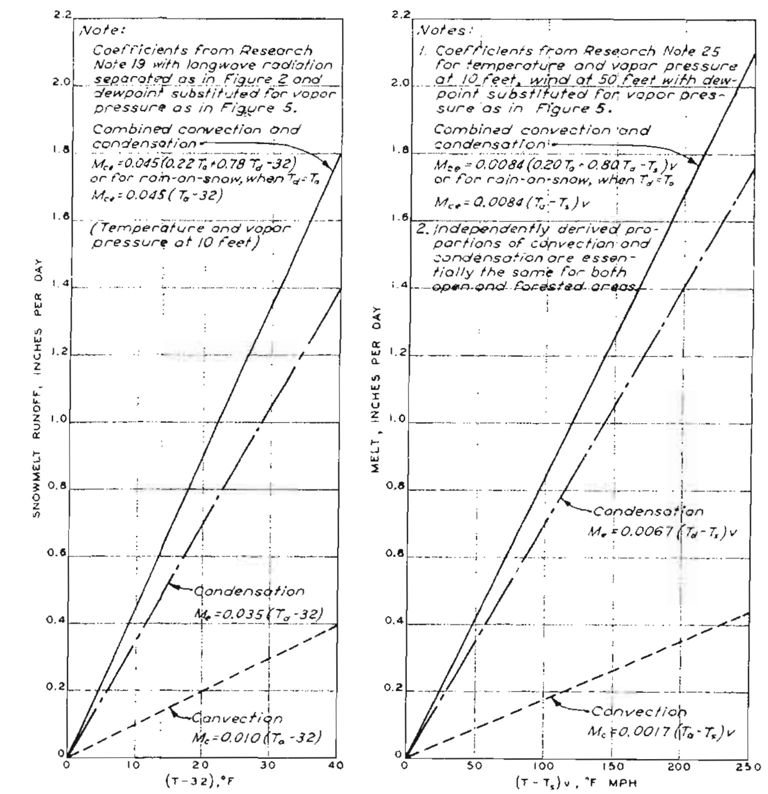
Symbol	Melt Component	Melt Rate*	Proportion**
M_{rl}	Longwave Radiation	0.029	0.72
M_c	Convection	0.011	0.28
M_{crl}	Combined Melts	0.040	1.00

*Melt rates are in inches per day per degree Fahrenheit above 32.

**Proportions used for separation of longwave and convection components of temperature coefficients in regression analyses where longwave exchange in forest and convection are indexed by a single parameter.

LONGWAVE RADIATION AND CONVECTION IN FOREST

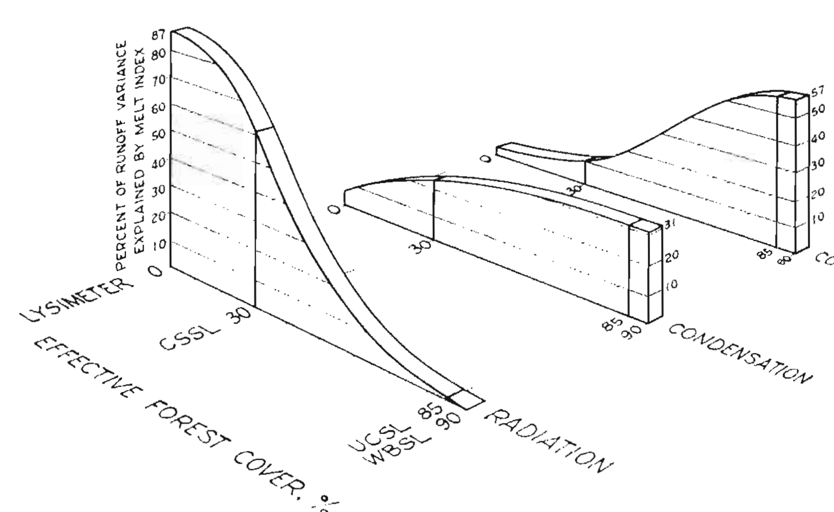
FIGURE 2



(a) FORESTED AREA — WBSL (b) OPEN AREA — LYSIMETER

CONVECTION AND CONDENSATION MELT RATES
FOR RAIN-ON-SNOW

FIGURE 3

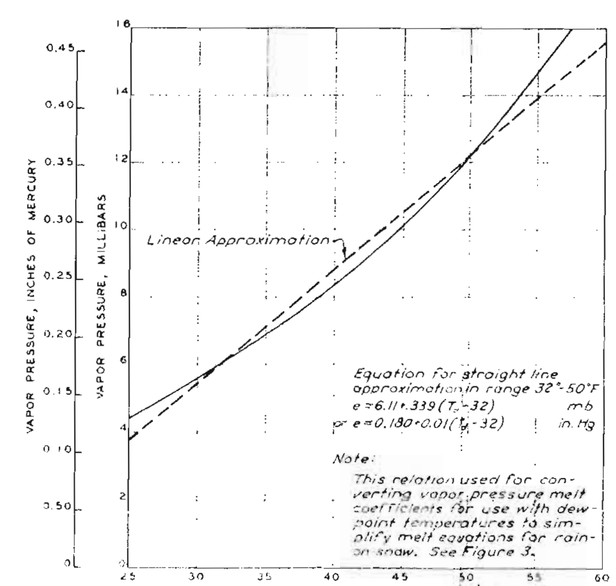


Forest Cover	Area & Stream		Partial Coefficients of Determination, δ Radiation	Condensation	Convection	Total	Period
0	Meadow	Lysimeter	0.87	0.06	0.04	0.97	1954
30	CSSL	Castle Cr	0.61	0.21	0.10	0.92	1954
85	UCSL	Skyland Cr	0.01	0.31	0.57	0.89	1949
90	WBSL	Mann Cr	—	0.36	0.55	0.91	1949

Notes:
1. Curves are drawn only to illustrate variations with forest cover and are not intended for interpolations for forest covers other than shown in table above except in a very general way.
2. Radiation = net all-wave radiation in open
3. Condensation = vapor pressure x wind (except WBSL)
4. Convection = temperature x wind (except WBSL)
5. WBSL regression function omits radiation and wind
6. Longwave radiation in forest indexed by convection parameter
7. Vertical lines are partial coefficients of determination computed in regression analyses of clear-weather snowmelt runoff on thermal budget indexes as shown above.

EFFECT OF FOREST COVER ON RELATIVE IMPORTANCE OF THERMAL BUDGET INDEXES
CLEAR WEATHER MELT

FIGURE 4



VAPOR PRESSURE VERSUS DEWPOINT,
WITH LINEAR APPROXIMATION

FIGURE 5

SNOW INVESTIGATIONS
SUMMARY REPORT

SNOW HYDROLOGY
SNOWMELT INDEXES

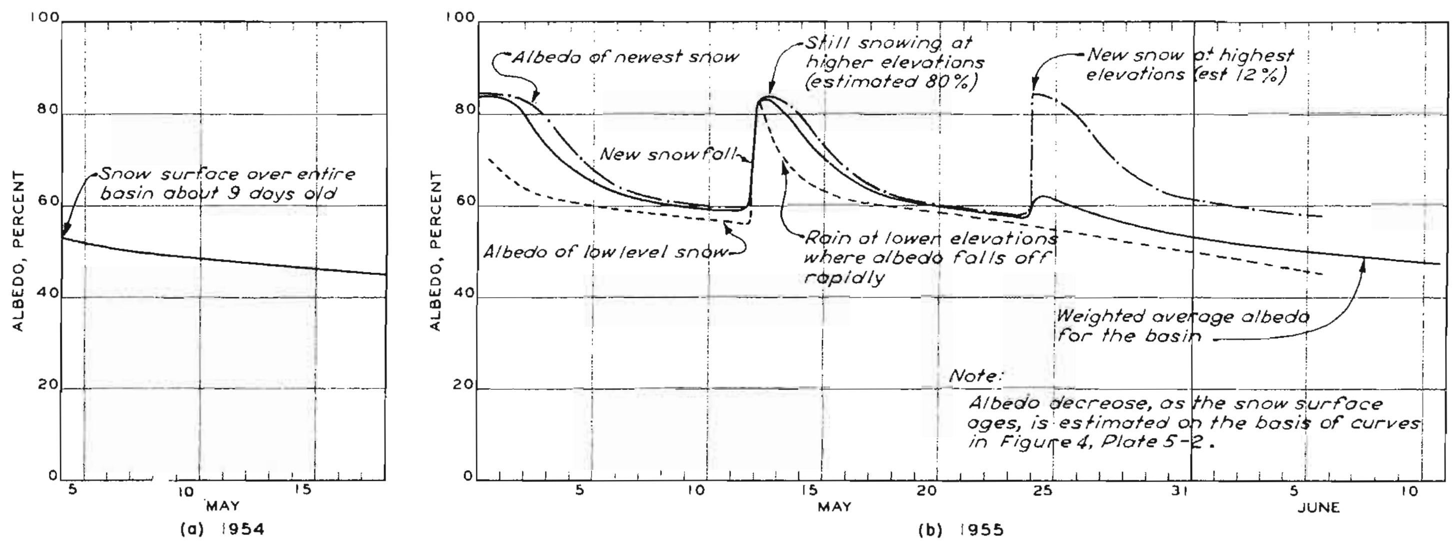
BASIC RELATIONSHIPS AND
FOREST INFLUENCES

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

PREPARED: DRAWN: SUBMITTED APPROVED TO ACCOMPANY REPORT DATED 30 JUNE 1956

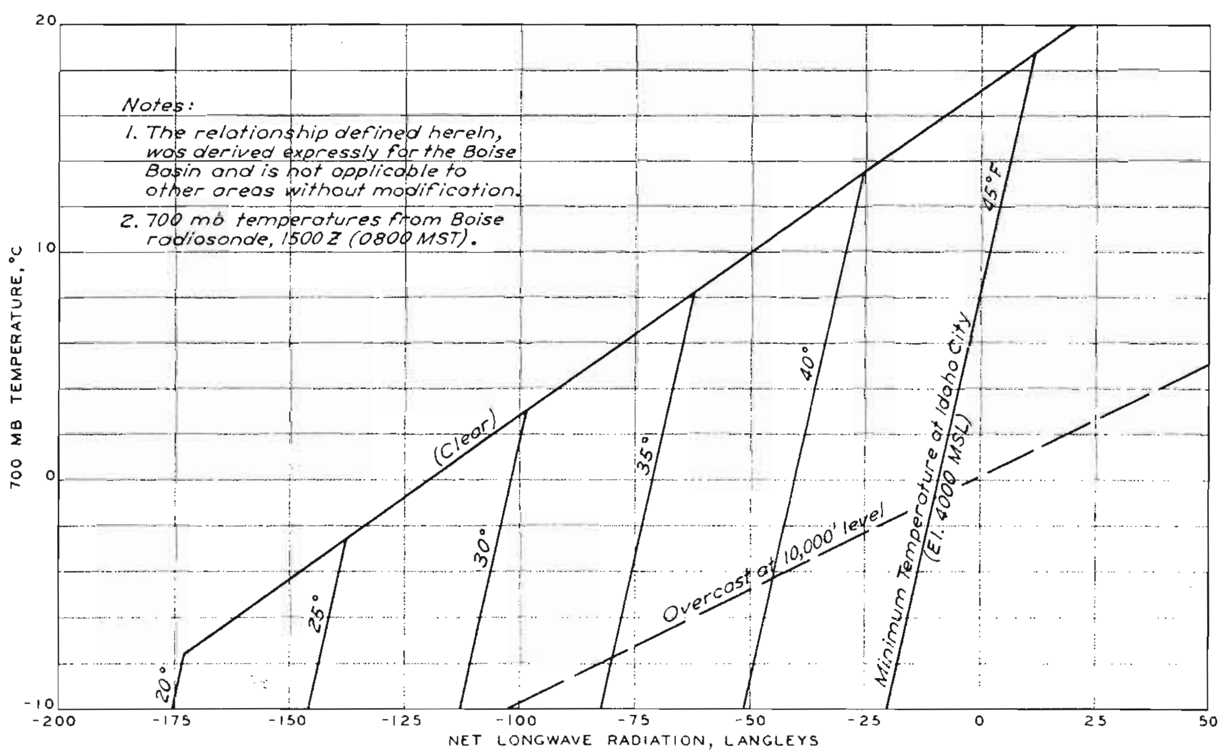
PD-20-25/37

PLATE 6-2



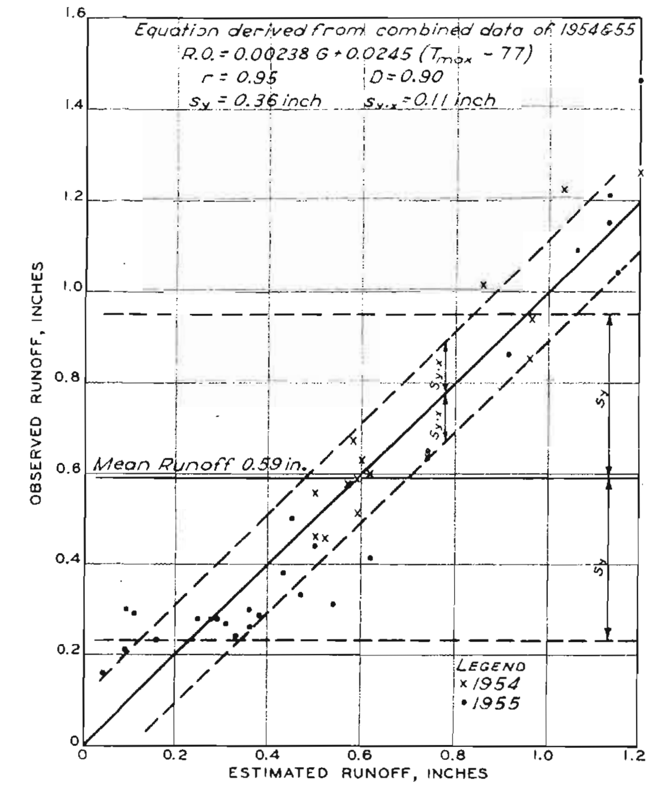
ESTIMATES OF SNOW SURFACE ALBEDO BOISE BASIN

FIGURE 1



ESTIMATION OF NET LONGWAVE RADIATION EXCHANGE BETWEEN SNOW AND SKY

FIGURE 2



R.O. = Daily generated snowmelt runoff in inches depth over snow-covered area

G = Daily net allwave radiation absorbed by snow in open, langleys

T_{max} = Daily maximum temperature, Boise, °F

r = Coefficient of correlation

D = Coefficient of determination

s_y = Standard deviation of observed runoff, inches

s_{yx} = Standard error of estimated runoff, inches

See Plate 7-6 for snow cover and other basic data pertaining to the 1954 & 1955 seasons for this basin.

OBSERVED VERSUS ESTIMATED RUNOFF

FIGURE 3

SNOW INVESTIGATIONS
SUMMARY REPORT

SNOW HYDROLOGY

SNOWMELT INDEXES

BASIN APPLICATION

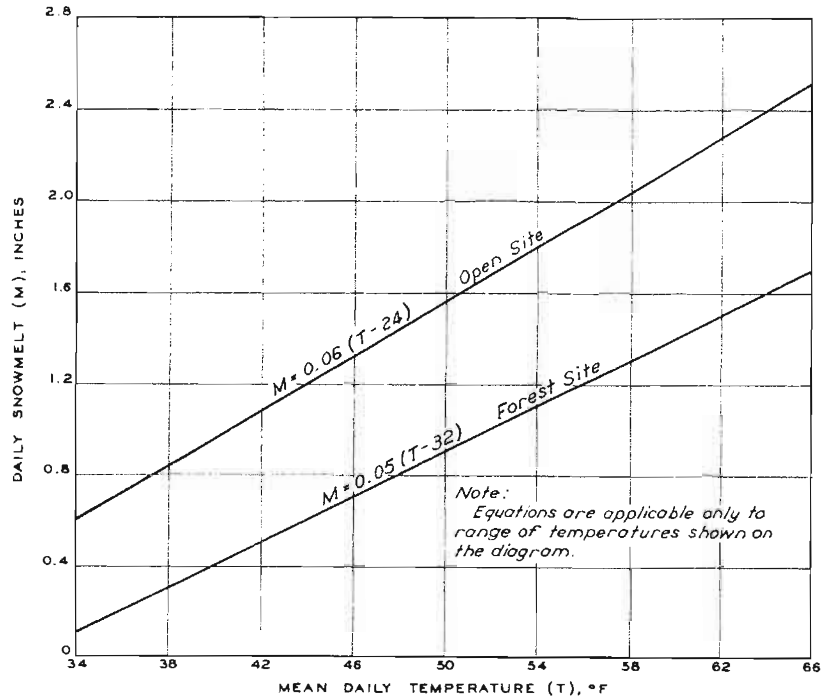
BOISE RIVER ABOVE TWIN SPRINGS, IDAHO
DRAINAGE AREA 830 SQUARE MILES

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

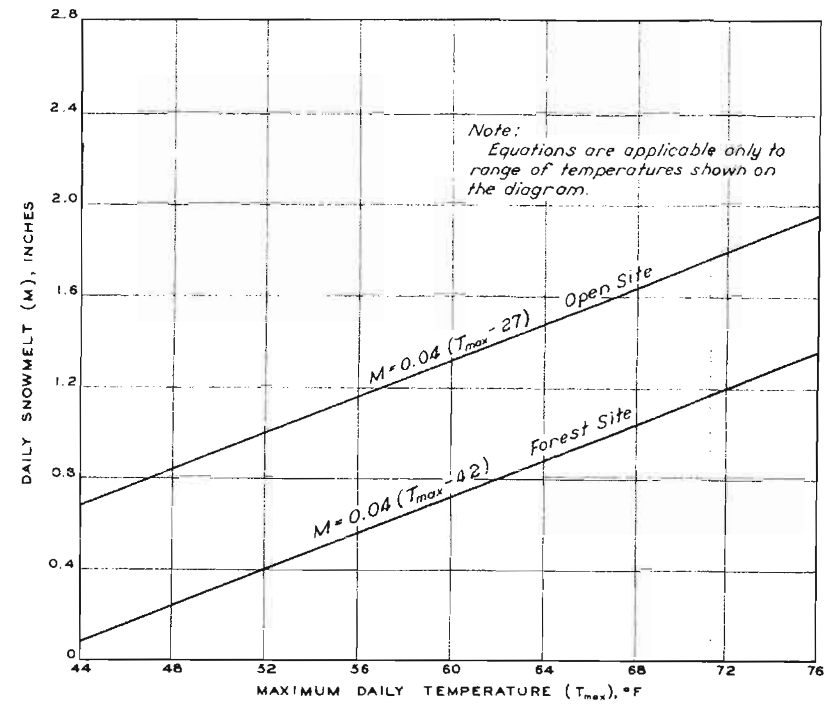
PREPARED GM·CEJ SUBMITTED GM TO ACCOMPANY REPORT DATED 30 JUNE 1956
DRAWN D.V. APPROVED DMR

PD-20-25/38

PLATE 6-3



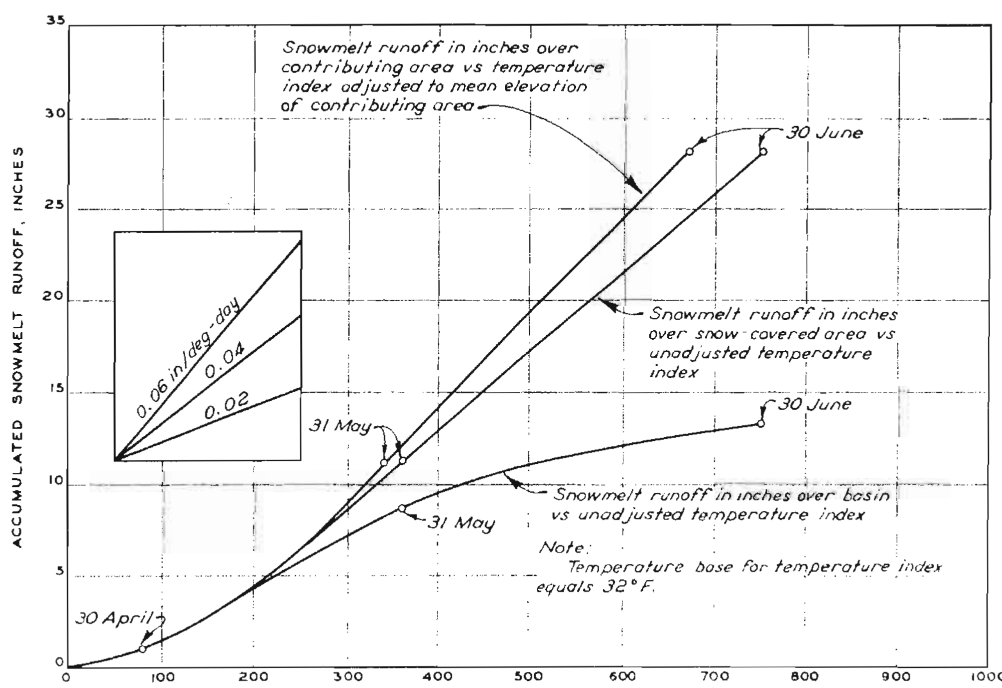
(a) MEAN TEMPERATURE INDEX



(b) MAXIMUM TEMPERATURE INDEX

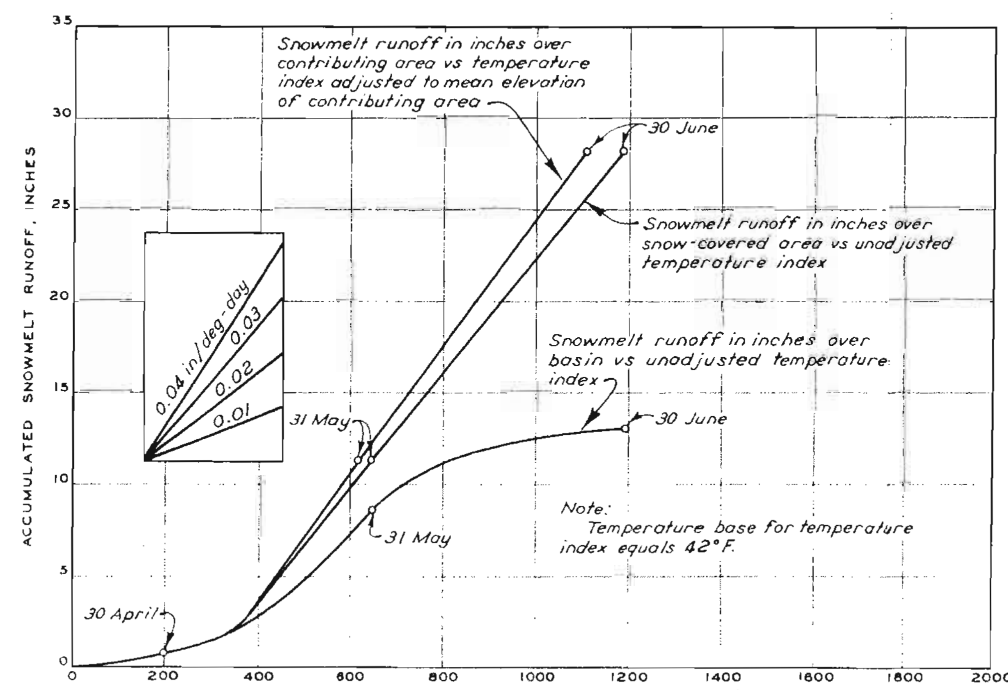
POINT MELT RATES

FIGURE 1



DOUBLE-MASS-CURVE ANALYSIS OF SNOWMELT RUNOFF
USING MEAN DAILY TEMPERATURE INDEX
SKYLAND CREEK, UCSL, 1949

FIGURE 2



DOUBLE-MASS-CURVE ANALYSIS OF SNOWMELT RUNOFF
USING MAXIMUM DAILY TEMPERATURE INDEX
SKYLAND CREEK, UCSL, 1949

FIGURE 3

Notes:

1. In figures 2 and 3 melt rates are read by referring slopes of curves to inset diagrams.

2. Temperature-index and snowmelt-runoff accumulations begin on 1 April.

SNOW INVESTIGATIONS
SUMMARY REPORT

SNOW HYDROLOGY
SNOWMELT INDEXES

TEMPERATURE INDEXES
OF SPRINGTIME SNOWMELT

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

PREPARED CEH SUBMITTED GM
DRAWN HLB APPROVED DMR
TO ACCOMPANY REPORT DATED 30 JUNE 1956

PD-20-25/39

PLATE 6-4

CHAPTER 7 - AREAL SNOW COVER

7-01. INTRODUCTION

7-01.01 General. - The area of snow cover has long been recognized as a prime variable in many applications of snow hydrology. Systematic observations of snow cover, however, have been generally lacking. Hydrologists have therefore resorted to empirically derived relationships between snow cover and runoff, assumed distributions of snowpack water equivalent by elevation zones and assumed zonal melt rates, or snow-cover indexes from ground observations of parts of a basin. Because of the lack of direct observation of snow covered area, none of these methods could be verified. As a result, the area of snow cover has been used to compensate for errors in other hydrometeorologic elements. Direct observations of snow cover are important as a forecasting tool, both for volume and rate-of-flow forecasting. In recent years there has been increasing recognition of the importance of snow cover to efficient operation of storage reservoirs, and a number of Corps of Engineers offices have begun making aerial snow-cover surveys during the spring melt season as an aid in reservoir regulation. As yet, an insufficient length of record is available to permit generalizations from snow-cover data; rather, the present use of the data has been limited to evaluating conditions at a specific time and developing observational techniques.

7-01.02 Definitions. - For the purpose of this report, the term snow cover refers to the extent of the ground area covered by snow, regardless of the depth of snow or its water equivalent. It may be expressed in units of area, such as square miles, or as a percentage of either the total basin area or an arbitrary maximum snow-covered area. The term snowpack refers to the total volume of snow on a basin. Snow-cover accretion is the increase in snow cover, while snow-cover depletion refers to a decrease in snow-covered area. Accumulation of the snowpack is the net increase in basin snowpack water equivalent, usually expressed in inches, while ablation refers to a net volumetric decrease of the snowpack water equivalent.

7-01.03 Functional use of snow-cover data. - There are two principal uses of snow-cover data in snow hydrology. One is for obtaining a measure of the areal extent of snowmelt at a given time, for the purpose of hydrograph synthesis. This may be involved in establishing procedures for streamflow reconstitutions, short-term forecasts, or design-flood computations. The second use of snow-cover data is in connection with volumetric forecasts of seasonal runoff. Snow cover may be used as a variable in establishing the volume of water stored in the snowpack, thus supplementing snowpack water-equivalent data from snow-course measurements, as was done in Research Note 22. In some mountainous regions, the area of snow cover may be used directly as an index of the water stored in the snowpack, as was done by Potts 5/ and

Croft.2/ A particularly useful application of snow-cover data is in connection with forecasting runoff for reservoir regulation after the melt season is well underway. All early-season volumetric forecasts possess a residual error, and as the melt season progresses, the magnitude of this error becomes a larger percentage of the remaining runoff. After the basin is less than, say, 50 percent covered, snow-cover data is particularly useful for verifying or adjusting earlier forecasts of runoff.

7-01.04 Requirements for hydrologic use. - In evaluating snow cover to meet the uses outlined above, there are three basic requirements to be considered: (1) the need for basic research on snow-cover accretion and depletion, and their relation to meteorologic and terrain factors which cause variation in precipitation and melt; (2) the necessity of direct observation of snow cover on project basins; and (3) the preparation of indexes or derived relationships for estimating the accretion and depletion of snow-covered area, for periods when observations are not available, or for design conditions.

7-01.05 Basic research is needed in order to improve present knowledge of the factors affecting snow cover accretion and depletion. If all basins had systematic snow-cover observations regularly taken through the accumulation and melt periods, this requirement would be much less important. Snow cover would then be simply another measured quantity. At the present time, however, observations are limited, and estimates of snow cover must be made indirectly.

7-01.06 The importance of obtaining direct observations of snow cover cannot be overemphasized. Subjective evaluations made from scanty and unrepresentative data are often misleading, because of the heterogeneity of basin areas and the complexity of the relationships between snow cover and its environment. The need for direct observations is three-fold. First, each basin has a characteristic pattern of snow-cover depletion, more or less consistent from year to year, which can be determined from direct observations of snow cover. A series of such observations over a period of years makes it possible to establish relationships for use when observations are not available. Second, such variations that do occur from year to year are so complex that they can only be determined by actual observation. Third, observations of snow-covered area, as measured quantity, are useful as a forecasting parameter for determining residual runoff volumes.

7-01.07 Primary factors affecting snow-cover accretion and depletion. - The accretion of snow cover usually begins on the higher elevations of the watershed, and continues through the accumulation period until all or a part of the watershed is covered. The depletion of snow cover begins with the exposure of the first bare ground in a completely snow-covered basin or with the date of maximum basin snowpack accumulation in a partially snow-covered basin. There are large differences between years in the length of accretion and depletion periods, depending primarily upon the meteorological regime during the accumulation and melt seasons.

7-01.08 During the accretion period, an elevation contour adequately defines the area of snow cover for small- to moderate-sized basins with relatively large ranges in elevation. This is due to the fact that the form of precipitation varies with elevation during individual storm periods, as explained in section 3-05, and the transition zone between areas of rainfall and snowfall is narrow. In addition, melt varies largely as a function of elevation during the winter period. Since solar radiation is at a minimum at this time, what little melt occurs is largely a function of air temperature, which in turn generally varies with elevation. The net result of these factors, then, is a fairly definite snowline during the accretion period.

7-01.09 Snow deposition, as related to meteorological and terrain factors, is a prime variable affecting snow-cover depletion. The variability of snow accumulation is discussed in chapter 3. During the accumulation period, the variations in the snow depths over basin areas show the combined effects of the terrain and meteorological factors. These factors include atmospheric circulation and airmass character during precipitation, opportunity for modification of airmasses, and large- and small-scale topographic influences. All of these influence the variations of snow depth from one point to another, which in turn will affect the depletion of snow cover during the melt period.

7-01.10 Snowmelt is the second prime factor affecting snow cover depletion. The meteorological and terrain factors causing variability of melt rates over a basin may act in entirely different ways from those affecting the deposition of snow. The principles involved in the variation of melt rates with respect to meteorological conditions, forest cover, and exposure to radiation are discussed in chapter 5. During winter, the melt rates over a basin are generally fairly uniform within a given elevation zone, and the amount of melt is usually small. During the spring and early summer there is wide variability to melt, due primarily to exposure; elevation effects are of lesser consequence. A definite snowline elevation does not exist during the melt period. Snow-cover depletion, therefore, reflects the variable influence of both the deposition and the melt of snow. The terrain influences on each are independent and should be considered separately.

7-01.11 In general, the seasons of heaviest snow accumulation have the longest lasting snow covers and those of lightest snowfall have the shortest. The interactions between meteorologic and topographic features determine the patterns of snow cover for individual cases. These factors are too varied and too complex to permit a practical general formula for assessment of snow cover from independent meteorological and topographic observations. Instead, a relatively simple empirical formula or chart for individual basins is needed to relate snow cover to readily observed data. Basin snow-cover depletion, snowpack ablation, and runoff are all the integrated effects of the same basic factors. Consequently, "depletion-ablation" or "depletion-runoff" curves may be constructed for a given basin if adequate and dependable data are available. The relationship may be improved by the introduction of a parameter such as the ratio of initial depth to areal snow cover, the initial basin snowpack water equivalent, or the ratio of snowpack water equivalents at low and high elevation snow courses. Using these relationships, one can determine the area of snow cover. Conversely, if a snow-cover survey is made, basin snowpack water equivalent or remaining runoff can be determined. These relationships can also provide the means for reconstructing the snow-cover depletion for years of historic floods of large magnitudes and establishing snow-cover criteria for design.

7-01.12 Organization of material and methods of approach. - The first part of this chapter deals with methods of obtaining direct observations of the snow-covered areas and summarizes generalizations on the accretion of snow cover. Then, observations of snow-cover depletion at the snow laboratories are used as a basis for discussing: (1) the relationships between snow-cover depletion, snowpack, ablation, and runoff; (2) the influences of terrain on snow cover. Finally, methods are given for using snow-cover data in forecasting residual runoff and in hydrograph synthesis.

7-02. METHODS OF OBSERVING SNOW COVER

7-02.01 General. - Systematic and complete observations of snow-covered area have been obtained only in recent years. Consequently, observational techniques for the collection of snow-cover data are not as standardized as those for many other basic hydrologic data. Also, the determination of snow-covered areas from the ground is extremely difficult, especially for rugged mountainous headwater areas, where routes of communication are generally lacking. The increasing use of aircraft for snow-cover observations has, however, made headwater areas readily accessible to both visual and photographic observations. In general, the following methods have been employed to obtain snow-cover information:

(1) Ground reconnaissance, utilizing prominent vantage points and transmountain highways to observe and map areas of snow, and also to define the elevation of the snowline when possible.

(2) Ground photography of selected sections of the drainage basin from fixed reference or vantage points.

(3) Aerial photography, either vertical or oblique.

(4) Aerial reconnaissance, from high- or low-level flight, supplemented by photographs, maps, sketches, and snowline elevation observations.

7-02.02 Ground reconnaissance. - Reliable determinations of the snow-covered area based on visual observations from the ground require considerable competence. No check or evidence exists to support the subjective opinions of the observer. The coverage cannot be as complete as coverage from aerial surveys because of obstructions, such as forest and hills, to the field of vision. In addition, the travel to satisfactory vantage points make ground observations expensive and time consuming. These limitations of ground reconnaissance surveys are more serious during the ablation than during the accumulation period. During the accumulation season, the use of ground reconnaissance is practical, since snowline elevations in mountainous areas are satisfactory indexes of snow-covered areas. In many basins, transmountain highways provide access through a range of elevations thus making possible the determination of the average snowline by automobile reconnaissance. Such observations may be made in conjunction with regular early-season snow-survey measurements. After the onset of the melt season, however, ground observations of snow cover are not recommended as a means of determining basin snow cover.

7-02.03 Simultaneous and independent observations of snow-covered areas by ground observation and aerial photographic methods were made at CSSL during the 1947 melt season. Comparison and analysis of these observations are presented in Appendix I to Research Note 16. Assuming the aerial photography analysis to give the correct snow cover, ground estimates of snow cover were found to be too high early in the season and too low later in the season. Estimates of snow cover made from high vantage points were found to be more reliable than those made from lower, inferior view points. On some of the sub-areas, the ground estimates of snow cover were as much as 50 percent less than those determined by aerial photography. For over half of the basin, however, the estimate from ground survey was within 15 percent of the value determined from the aerial photographs. From this experience, it is concluded that estimates from the ground should be made from high points whenever possible. Care should be taken not to bias the results to favor the more readily observable open areas in preference to wooded or more obscure areas. This results in overestimates early in the season and underestimates late in the season.

7-02.04 Ground photography. - Ground photographs are generally used as indexes of areal snow cover rather than measures of the total snow-covered area. Actual snow-covered area can be determined by relating these photographic indexes to actual snow-cover amounts determined by other means. Potts 5/ has utilized ground photographic methods for providing a direct index to snowmelt runoff on the middle fork of the South Platte River in Colorado, by-passing the determination of areal snow cover. Miscellaneous Report 1 describes a procedure for establishing a snowline index from horizontal ground photographs of the Sierra Nevada in the vicinity of CSSL. Panoramic photos were taken at a site at about the 5000 foot level near Emigrant Gap, California. The procedure consisted essentially of obtaining a master photograph, on which were located prominent features of the landscape with their elevations. A considerable portion of the 40 sq. mi. drainage area of the South Fork of the Yuba River is visible from this site. Between 8 March and 21 June 1948, thirteen series of panoramic photographs were taken from the site, and snowline indexes from elevation zones and exposure sectors were determined for each series. The index was used for the purpose of correlating snow cover to measurements of water equivalent of the snowpack and runoff.

7-02.05 In 1950, the U. S. Weather Bureau established a snow-cover investigations unit, for the purpose of estimating the extent of snow cover in the Columbia River basin, to be used in connection with seasonal and short-term streamflow forecasting. A report of the activities of this unit was made at the annual Cooperative Snow Investigations Conference of 1 April 1952, and at the Western Snow Conference in 1953. 1/ The principal method used by this unit to estimate snow cover has been by ground photographs from key stations. A photographic record is currently being accumulated from which indexes of snow cover may be determined. The photographs are taken periodically through the melt period by cooperative observers and using standardized photographic procedures.

7-02.06 The standardized procedure used by the snow-cover unit to photograph selected portions of a basin from fixed vantage points, for the purpose of establishing an index of snow cover, consists of:

(1) Selecting photographic stations.

(2) Preparing a master panoramic view of the watershed from each station.

(3) Dividing the master photograph into convenient exposure sectors.

(4) Identifing and determining elevation of prominent landmarks on the master photograph.

(5) Photographing the progress of snow-cover depletion from each point.

(6) Determining the average snowline in each exposure sector and assigning a snow-cover index value to each snow-line elevation.

The procedure is subject to considerable personal judgment and requires experience in taking the photographs and evaluating the snow line. A disadvantage is that some areas are obscure, and also there is great distortion with increasing distance from the camera, so that the use of a uniform grid system is not practical. An additional disadvantage of the method of ground photographs is that for routine observations a considerable time is required before the pictures are ready for analysis. Since these photographs provide only an index of snow-covered areas, if actual basin snow cover is desired, results must be correlated with simultaneous observations of the actual cover. The method does have the advantage of being inexpensive and of providing records to supplement personal judgment.

7-02.07 Aerial photography. - Aerial photography is an exact method of determining the area of snow cover. It furnishes a permanent record which can be analyzed at any time, and information which can be transferred to basin maps for evaluation of the true cover. Aerial photographs are taken both vertically or obliquely. Their use varies from a supplement to visual observations to a complete and precise delineation of the snow-covered area. The principal advantages of aerial photographs are: (1) a record is obtained which may be preserved, and from which detailed analyses may be made of true snow-covered area; (2) remote regions which are not accessible to ground surveys may be covered by air; and (3) through use of stereo-pairs, determinations of snow cover and surrounding terrain elements may be made from which the two may be correlated. Aerial photographs have been used primarily for special studies of snow cover on small basins. Their application to larger basins for operational use, however, is subject to the following limitations: (1) the large number of photographs required to cover such basins; (2) the high cost of operation of aircraft which can operate at high elevations; (3) the time required for photograph processing and evaluation; (4) the difficulty of interpretation of snow cover in forested areas; and (5) the near-ideal weather conditions required during periods when snow-cover observations are needed.

7-02.08 Aerial photographs of snow cover were obtained for UCSL and CSSL, for the years listed in table 2-6 (chapter 2). Those for UCSL were vertical photographs, taken at flight levels ranging from 13,000 to 17,500 feet and using aerial mapping cameras. For CSSL, oblique photographs were obtained primarily by light aircraft not equipped for vertical photography. Data from these flights were used for detailed studies of snow-cover relationships.

7-02.09 In the application of vertical aerial photographs for project-size basins, the most critical difficulty lies in the large number of photographs required to cover the basin. Flying at 10,000 feet above the ground surface and using a 9 x 9 inch camera with an 8.15 inch focal length, the total area covered by each picture is about 4 square miles. If only one quarter mile overlap is provided for each picture, the effective area is 2.2 square miles per picture. For a 10,000 square mile area, about 450 photographs would have to be taken, processed and interpreted. Even when photographing only the marginal zones of snow cover, a large number of pictures is required.

7-02.10 Aerial reconnaissance. - The most feasible method for obtaining timely and accurate estimates of snow cover is aerial reconnaissance. Unlike the data from aerial photographs, the reconnaissance snow-cover information is available to the hydrologist as soon as the flight is completed. The probable small increase in accuracy would not warrant taking aerial photographs, particularly when special skills for processing and interpreting them are not generally available. The cost of aerial reconnaissance using light aircraft is less than ground reconnaissance, considering the difference in time required. Its cost is far less than that for obtaining complete coverage by aerial photogrammetric methods. While suitable weather conditions are required for aerial reconnaissance, these requirements are considerably less than for aerial photos. The principal disadvantages of the method are: (1) evaluation is subjective and results cannot be verified, and (2) it is dependent to some extent upon weather conditions. Observing snow under the forest is difficult, regardless of the basic method used, whether from ground or air. Supplementing aerial reconnaissance with oblique aerial photos provides continuity between observations and aids in verifying subjective observation.

7-02.11 Various Corps of Engineers offices, including Sacramento, Portland, Seattle, Walla Walla, and Omaha districts, have made regular observations of snow cover by aerial reconnaissance for use in flood evaluation and reservoir regulation. The methods used have varied, depending upon the areas involved, aircraft available, weather, and preference of those making the flights. In general, two procedures have been used successfully. Both consist basically of observing snow-covered areas and plotting them on a topographic map. In one case, flight altitude is maintained 4000 to 5000 feet above the highest ridge lines, thus enabling the observer to obtain a fairly broad view of the basin as a whole. The other method is to fly in the canyons approximately at the elevation of the snowline. The high-level flight requires better conditions of ceiling and visibility, which often limits the time that observations may be made; however, more comprehensive definition of the snow-covered area may be made from high-level flight. Also in some cases mountain ranges may be too high to permit the use of light planes in the high-level flight procedures. Early in the season when snow generally covers a large portion of the basin, only the lower portions of the basin are exposed and there is less variability of the snowline. At such time, low-level flight may be preferred. Whenever cloud conditions prevent flying over mountain ridges, low-level flight must suffice. A hand level is sometimes used to maintain the flight level at the approximate snowline elevation so that altimeter readings can be used to measure the height of the snowline. When it is impossible to fly the entire basin, a system of sampling the elevation of snowline on various aspects and areas may be used to evaluate an average basin snowline. This in turn may be applied to an area-elevation curve to determine the snow-covered area. When viewing heavily forested areas from low-level flight, the line of sight being more or less horizontal, it is difficult to determine snow cover under trees. Viewing the area vertically gives a better opportunity to estimate snow cover under those conditions. Snow-free patches, well within the snow-covered area, often occur on steep slops. These areas appear relatively large when viewed horizontally or obliquely; actually, they may represent negligibly small areas on a horizontal projection. A detailed report on aerial recommaissance of snow cover in the Kootenai and Flathead basins by the Seattle District office is contained in Technical Bulletin 15.

7-02.12 The following general recommendations are made for conducting aerial snow-cover reconnaissance surveys:

(1) Trained personnel who know the basin hydrologic characteristics should make the survey.

(2) The observer should be familiar with landmarks throughout the basin, and ground features should be identified continually during the flight.

(3) The snowline should be identified as to elevation or location and plotted on the base topographic map.

(4) Where spotty or patchy fringe areas of snow cover exist, an average snowline should be estimated and plotted.

(5) Aerial snow-cover reconnaissance flights should be scheduled to coincide with ground snow-course surveys, insofar as possible.

(6) Aerial snow-cover surveys should not be made immediately after a new snowfall.

(7) Supplementary photographs should be taken which will show progress of snow-cover depletion in a given area from flight-to-flight. Photograph points should be established which are easily recognizable. Pictures should be taken from the same point each time, at a specific altitude. Areas photographed should show both northerly and southerly slopes.

(8) Low-level flights should be made at the snowline level and this level plotted on the topographic map.

7-02.13 The interpretation of the snow-covered area from aerial reconnaissance surveys is made by planimetering the areas of snow on the basin map. As an independent check, average snowline data may be applied to area-elevation curves, but this latter method is less reliable. Where basin coverage is not complete, area-elevation relationships must of necessity be used. For this purpose, the average snowline should be carefully weighted to reflect the various conditions of exposure throughout the basin.

7-03. SNOW-COVER ACCRETION

7-03.01 Although a simple method for quantitative evaluation of snow-cover accretion has not been developed, a few qualitative statements may be made concerning the processes involved. Broadly speaking, there are two types of areas to be considered, namely, mountainous regions and open plains. For mountainous regions, as was explained earlier, the area covered with snow during periods of accumulation varies primarily with elevation. Reference is made to section 3-04 for a discussion of elevation effects on snow accumulation. While there may be large variation in the depth of snow because of meteorological and terrain effects on snow deposition, the amount of melt during the accumulation period is small and generally is not sufficient between storms to expose the lightly covered areas. The elevation of the snow-line, therefore, is primarily a function of the form of precipitation in individual storms, and accordingly is dependent upon elevation.

7-03.02 On wind-swept open plains, snow-cover accretion may be irregular because of drifting. In such areas, the elevation range is small and the effect of elevation on accretion is usually negligible. The variation in snow cover in this case is a complex function of meteorologic conditions, primarily of wind, temperature, precipitation, and the sequence of these events, superimposed on small-scale terrain irregularities. The relative magnitudes of these effects have not been determined separately.

7-03.03 In some mountainous regions, the combined effects of steep slopes, high wind, and lack of forest may result in local patches of snow-free ground in an area which is generally snow-covered during the accumulation period. Usually, the relative magnitude of these areas is small when considered with respect to the drainage basin as a whole.

7-03.04 The effect of forest on snow accretion is to cause greater uniformity of cover than would occur in bare areas. The effect of forest on snow accumulation may be considered analogous to its effect on snowmelt, in that it tends to "level out" the variability of meteorologic processes. In the case of snowmelt, forest influence permits the use of temperature to index the radiation melt process. For snow accumulation, a uniform forest stand provides a means for distributing snow more equitably with regard to the rest of the terrain features and minimizes the variability of deposition in locations of abnormally high winds or steep slopes.

7-04. SNOW-COVER DEPLETION AND ITS RELATION TO TERRAIN

7-04.01 General. - The factors affecting snow-cover depletion are extremely complex and include the interrelationships between terrain and meteorologic conditions during snow deposition, as well as during melt periods. It is not feasible to attempt evaluation of snow cover depletion by rational procedures. Rather it is necessary to derive empirical relationships with readily observed data for the purpose of determining snow cover for individual basins. Not only do these relationships vary from basin to basin but, with regard to snow-cover depletion on a particular basin, each year has its peculiarities, resulting primarily from the meteorological differences during the accumulation and melt periods. Therefore, a precise quantitative definition of snow-cover depletion applicable to all areas and to all years is not possible from the limited observations which are available. It is, rather, the intent of this section to present a qualitative evaluation of the processes of snow-cover depletion and their relation to terrain. These are based primarily on observations at the snow laboratories. Later sections of this chapter deal with snow-cover depletion as related to ablation of the snowpack and accumulated runoff.

7-04.02 Analyses of snow-cover depletion in relation to terrain have been made on the basis of snow-cover observations at CSSL and UCSL (as reported in Research Note 16). The results of the analyses contained in Research Note 16 are summarized in the following paragraphs.

7-04.03 Analysis of the 1947 season at CSSL. - A detailed analysis of snow-cover depletion during the 1947 season at CSSL was accomplished by subdividing the basin into twenty topographic units of homogeneous character, as shown on figure 2, plate 7-1. The percent snow cover was determined for each unit from analysis of aerial photographs of the entire basin which were made that year. Several flights were made from which the progress of depletion could be determined.

7-04.04 The 1946-47 season at the CSSL was deficient in snowfall; the snowpack water equivalent was less than 70 percent of normal. The melt season was warm and free of storms from 10 April until the end of May when most of the snow was gone. Consequently, the determination of snow cover during the melt period was not complicated by new-fallen snow from spring storms. The initial streamflow rise commenced on 10 April, and the snowmelt contribution of runoff terminated early in June. The peak discharge occurred on 1 May. In general, the continuous nature of the melt season made it ideal for the study of snow-cover depletion.

7-04.05 Depletion of snow cover, 1947, at CSSL. - A generalized description of the progress of depletion of snow cover during 1947 is as follows:

a. On 31 March snow cover was substantially complete over the entire basin, with minor exception of some steep slopes on Castle Peak.

b. At the middle of April, cover was still high, averaging about 92 percent over the basin; bare areas had appeared on high parts of Castle Peak and on the south side of Andesite Ridge.

c. At the end of April, cover averaged about 80 percent. Bare spots that had appeared in the middle of April were larger, and snow cover in Uhlen Valley was also partly broken up. In the other areas little change in extent of snow cover had occurred. Most of the topographic units still had more than 60 percent snow cover, and in the upper basin there was a large block of units with more than 90 percent cover.

d. At the middle of May, snow cover averaged 37 percent over the whole basin, but the dispersion was quite large. In two topographic subdivisions, mostly in the upper part of the basin on both sides of Willow Valley, snow cover exceeded 90 percent. On the other hand, eight subdivisions, chiefly south-facing slopes in both upper and lower basins, were nearly bare. Figure 7, plate 7-1, shows the areal distribution of snow cover on 30 April and 15 May, when the average basin snow cover was 79 percent and 37 percent respectively. The similarity of patterns on the two dates may be noted. The relatively stationary status of the units with above average snow and the rapid depletion of units that were initially below average in cover result in an increase of dispersion.

7-04.06 The sequence of depletion for 1947 is illustrated in figure 6, plate 7-1. This diagram shows for each topographic unit the number of days after active melt had begun before a snow cover of 60 percent was attained. It presents, therefore, a measure of the rates of snow-cover depletion for various conditions of terrain. Seven of the units reached 60 percent snow cover within 25 days, while two of the units required in excess of 50 days to reach 60 percent cover. The shortest time was 13 days for the steep, south-facing slopes of Castle Peak, while the longest time was 60 days for the sheltered north slopes of Andesite Ridge. Figure 5, plate 7-1 illustrates schematically the sequence of snow-cover depletion for an unforested area with relatively steep north- and south-facing slopes. For the CSSL, the windward slopes face south and the leeward slopes face north. As a result of local topographic influences the accumulation of snow is greater on north than on south slopes (see chapter 3). Also, since melt rates are greater on south-facing slopes, the combined depletion effect results in south slopes going bare well in advance of other areas. North slopes, with their greater accumulation and reduced melt rates, exhibit the opposite effect.

7-04.07 Topographic influences. - Watersheds differ from one another in topography and orientation with respect to exposure to the flow of airmasses and to solar radiation and other factors affecting deposition and melt. The differences cause variation in depth of snow and in the duration of the melt season between basins. Even within a watershed, local differences in topography exist which cause variability in the accumulation and melt of the snowpack, and consequently in the snowpack ablation and snow-cover depletion. In relatively flat areas, such as open meadows and valleys or in plains regions, the snow cover tends to remain in tact for a relatively long time until it becomes quite shallow. It then exhibits a rapid change as large areas of thinned snow become bare simultaneously. This tendency is characteristic of all snowpacks of uniform depth subject to uniform melting rates. In mountainous areas, on the other hand, there is wide variability in the snow-cover depletion with area. Yet, for a given area, the depletion pattern is remarkably similar from year to year. A characteristic effect of topography is manifest in the appearance and development of bare patches, which appear at the same sites and grow in nearly identical patterns each year.

7-04.08 Orientation. - The basic considerations of the effect of slope orientation on snow-cover depletion were mentioned in paragraph 7-04.06. The following tabulation, based on CSSL data for the 1947 season, shows the progress of depletion of snow cover as a function of slope orientation:

DEPLETION OF SNOW COVER WITH RESPECT TO ORIENTATION, CSSL, 1947

Orientation	Percent of area snow covered				Percent of basin area
	31 Mar	30 Apr	13 May	16 May	
N	100	83	76	57	4
NE	100	82	72	69	5
E	100	87	76	75	8
SE	98	74	37	22	18
S	99	67	32	25	27
SW	100	74	35	27	15
W	100	79	58	38	13
NW	100	83	69	56	10

The plotting of the above data in figure 3, plate 7-1, illustrates the progressive decrease in snow-covered area for the various orientations relative to one another. In extreme, the rate of change of snow-cover depletion is from two to three times greater for south slopes than for north slopes. (Actually the depletion rate tends to be least in the northeast octant, which reflects the greater deposition of snow on the lee side of local barriers during the southwesterly atmospheric circulation accompanying storms, as well as the reduced melt rates on northerly slopes.) Strictly speaking, slope orientation should not be evaluated without also considering the steepness of slope. Very flat slopes of north and south orientation would tend to be quite similar in depletion characteristics while steep north and south slopes would be markedly different. The effect of steepness will now be examined.

7-04.09 Steepness. - In general, the accumulation of snow varies inversely with the steepness of slope, as was pointed out in chapter 3. For the CSSL, the fact that most steep slopes are for southerly orientation also results in greater melt rates. The combined effect of below average snow depths and high melt rates causes snow cover to deplete at a fast rate on these steep slopes. Separate evaluation of the relationship between depletion and steepness of slope is not practical from CSSL data, because of the interrelationship between steepness and orientation.

7-04.10 Elevation. - The data from CSSL are inadequate to relate depletion with elevation, because the entire basin is within the headwaters area of the Sierra Nevada. The range in elevation is small and other topographic influences at these high elevations obscure the effect of elevation. Data from WBSL presented in chapter 3 showing the variation in slope of the snow-wedge with time reveal the nature of depletion in that type of area. As was pointed out in paragraph 3-04.06, the slope of the snow-wedge increases through the accumulation period, but after active melt is under way, there is little variation in melt with respect to elevation, resulting in a nearly uniform decrease of the snowpack water equivalent with elevation. Under these conditions, the depletion of snow cover with respect to elevation is a function almost entirely of the variation in snow accumulation, and only slightly of the variation in melt. When considered over large ranges in elevation (sea level to, say, 10,000 feet) elevation is of course the most important single topographic variable in its effect on depletion of snow cover.

7-04.11 There are several compensating factors affecting variation of melt with elevation. What melt occurs during the accumulation season is largely a function of elevation. Solar radiation melt is small, hence air temperature mainly determines the amount of melt. During the late spring melt season, however, solar radiation is the prime source of energy for melting snow.

7-04.12 Figure 4, plate 7-1 shows the snow cover-elevation relationship for various dates of observation for CSSL during 1947. This diagram illustrates the depletion of snow cover with elevation and shows that the relative magnitude of depletion in the various elevation zones was greatest in the upper and lower portions of the basin, and least in the mid-elevation zones. The effect of topographic features other than elevation obscured any quantitative evaluation of elevation effect on depletion.

7-04.13 Forest. - It is difficult to evaluate quantitatively the effect of forest on snow-cover depletion. Studies from CSSL show little relation between forest and depletion, but the results were obscured by the effect of more significant terrain parameters. Figure 1, plate 7-1 is an aerial mosaic showing the distribution of the forest at CSSL. It was shown in chapter 4 when considering the interception of snow by the forest crown that the accumulation of snow under dense forest may be less than 80 percent of that in adjacent open areas. Factors affecting melt in various-sized forest openings have been discussed in chapter 5, and in general, melt rates are highest in large clearings and decrease to a minimum in small clearings protected from sunshine by the surrounding trees. In a broad sense, the effects of forest on accumulation and melt tend to balance each other, so that the depletion rates would be similar in magnitude. It has been observed in the heavily forested WBSL that the last remaining snow patches are in the small forest clearings, which again shows the integrated effects of above-normal accumulation and reduced melt in these locations.

7.04.14 Kittredge 4/ performed an exhaustive study on the influence of forest on snow in the central Sierra Nevada using observations made over a period of seven years. Measurements included profiles of snowpack water equivalents under various densities and species of forest, made at various times through the snow season. Those conclusions from the study directly pertinent to snow-cover depletion are quoted below:

"1. From 13 to 27 percent of the seasonal snowfall was intercepted by the forest canopies.

"2. The maximum water equivalents of the total snow on the ground or the amounts of water in storage in the snow are larger in red fir and in the cutover stand with large openings than in the clearings, and smallest in dense fir and ponderosa pine stands. The dates of maximums in the forested areas are usually later than in the open areas. Maximum water equivalents in the cutover mixed conifer and in a few other areas, for some years, vary inversely with the crown coverage within a 20-foot radius.

"3. The effect of trees on the south side of the large clearing on the water equivalents of the snow was to maintain greater storage not farther to the north than the height of the trees, as compared with the smallest amounts at greater distances where melting was more rapid.

"4. Openings between the crowns showed average maximum accumulations of 1 to 5 inches water equivalent larger than did areas under the crowns.

"5. The first exposure of bare ground varied from March 28 to May 5 between extremes in different forest types, and more than 60 days in different years in the same type.

"6. The average date of disappearance of the snow varied from April 17, in the old ponderosa pine, to June 1 in the red fir, and by about 2 months between extreme seasons.

"7. The date of disappearance of the snow varied inversely to the crown coverage within a 20-foot radius in the cutover mixed conifer area, and in some other types in certain years.

"8. The average duration of the snow cover varied from 117 days, in the ponderosa pine, to 160 days in the red fir area.

"9. The percentage of area covered by snow decreased after the first exposure of bare ground by from 4.4 percent per day, in the red fir area, to 17.2 percent in the lower meadow.

"10. The rates of melting tended to be lower under the crowns than in openings, and lower in openings than in the large clearing, per unit change in the independent variable in each case, but the influence of trees in retarding melting was quite small."

7-04.15 Snow- cover depletion, UCSL. - To illustrate the process of snow-cover depletion in an area of non-uniform deposition of snow, successive aerial photographs of the progression of depletion within the Blacktail Hills, UCSL, are shown on plate 7-3, for 1946 and 1947. The area shown in each photograph is slightly over one square mile. The growth of bare areas is apparent during successive periods of melt. The wind-swept ridge of the Blacktail Hills possesses little forest and conditions are favorable for low deposition and high melt of snow. The ridge becomes bare of snow early in the season, but on the lee side (northeast), snow remains much later in the season. These photographs illustrate the uniformity of depletion patterns between the two years (1946 and 1947) which serves to give confidence to the use of index relationships for estimating snow cover in mountainous regions. Plate 7-2 is an aerial photograph of the entire UCSL, taken on 2 May 1946, showing approximately the mid-season condition of snow-cover depletion on the basin. Also outlined on this photograph is the area of the Blacktail Hills contained in the successive photographs of plate 7-3. Notice the wide diversity of areas bare of snow for the various slopes within the basin.

7-04.16 Effect of diversity of terrain on snow-cover depletion. - The preceding paragraphs have discussed the variability of snow-cover depletion caused by each of the primary terrain factors. The integrated effect of all these factors on a basin area determines the rate of snow-cover depletion. The greater the diversity of terrain, the longer will be the time of depletion of snow cover. Areas having uniform conditions of accumulation and melt will exhibit rapid changes in snow cover from the time the first areas become bare to the condition of complete loss of snow.

7-05. SNOW-COVER DEPLETION VS. ABLATION OF THE SNOWPACK

7-05.01 General. - The preceding section described in general terms the variation of snow cover depletion with major terrain factors, in order to show the fundamental processes involved in snow-cover depletion. For the practical determination of snow-covered areas, however, it is necessary to determine average relationships between basin snow cover and commonly observed data. One such usable relationship is that with ablation of the snowpack, or as a step further, accumulated runoff.

7-05.02 Figure 1, plate 7-4 is a schematic diagram illustrating basic differences in the character of snow-cover depletion-ablation relationships of deep snowpacks. Curve A represents the conditions for heterogeneous basins, where snow accumulation and melt are affected by topographic variability. Beginning with the time the basin first begins to go bare, the area of snow cover decreases quite uniformly with ablation of the snowpack, resulting in a curve which is slightly concave downward. The reverse curvature near the bottom of the curve is caused by the few remaining deep drifts which last long after the major portion of the original snow-covered area has gone bare. Curve A is typical of mountainous areas of western United States. Curve B shows the rate of depletion on a homogeneous basin, where large amounts of snow are uniformly distributed over the area, and where melt rates are relatively uniform. Here, the snow-cover depletion with respect to ablation is slow at first and then suddenly increases. This type would be expected in the plains regions.

7-05.03 Depletion vs. ablation, CSSL. - Depletion-ablation relationships are shown in figure 2, plate 7-4 for several of the homogeneous topographic units of CSSL for the 1947 melt season. Also shown in the figure is a curve representing the basin as a whole. Curves for each of the units lie above the one for the entire basin and show that for areas of homogeneous character, there is a trend for a more pronounced "knee" in the curve as discussed in the preceding paragraph. When an area has a large variety of slope facets, as in the case of the basin as a whole, the curvature becomes less pronounced.

7-05.04 Figure 3, plate 7-4, shows the depletion-ablation relationships for four years at CSSL for the basin as a whole. Data for 1948, 1950, and 1952 are less complete than those for 1947. It is seen that the curve for 1947 lies below that of the other three years. This is accounted for by the fact that the relationship is begun on 1 April at which time the 1947 snowpack was relatively less than in the other years. Because of this fixed starting date, the curvature in the relationship is greater for years of above-normal snowpack accumulation while for years with below-normal snowpack the curvature is less than it is for normal snowpack conditions. Beginning the accumulated ablation-snowcover curves at 98 percent cover regardless of date, these curves are all similar in shape.

7-05.05 The difference between accumulated ablation of the snowpack and accumulated runoff represents the net effect of losses (evapotranspiration and soil-moisture increase) and ground-water and other basin storage. Figure 4, plate 7-4, indicates the 1947 CSSL snow-cover depletion as a function of accumulated runoff as well as accumulated ablation of the snowpack. The displacement of the runoff curve to left of the ablation curve is due to losses and storage. (In the case of CSSL, storage is relatively small in proportion to the total runoff.)

7-06. SNOW-COVER DEPLETION VS. RUNOFF

7-06.01 General. - Relating snow cover to observed runoff during the active melt period provides a convenient method for estimating snow-covered area continuously through the melt season. Snow cover may be related directly to observed data, or a mathematical function may be used to express the relationship. Data from Research Note 16, showing the relationships at the laboratories and a few miscellaneous basins, are presented to illustrate the general character of the relationships. Runoff may be accumulated commencing either from (1) the time of initial rise in streamflow, (2) the time of maximum snowpack, or (3) an arbitrary date, such as 1 April. It is also useful to accumulate historical runoff data from the end of the snowmelt runoff season, backward through the melt period, and thus relate snow cover to "future runoff." All values of accumulated runoff should be corrected for spring precipitation (either rain or snow) so that the relationships will express conditions resulting from the initial snowpack and thus will be more consistent from year to year.

7-06.02 Examples of depletion vs. runoff relationships. - Figure 5, plate 7-4 shows curves of snow-cover depletion as a function of accumulated runoff from snowmelt for the period 1 April through 31 July for the CSSL basin, Skyland Creek at UCSL, St. Louis Creek in Fraser Experimental Forest, Colorado, 3/ and for Kings River, California. These curves reflect the effects of snow-cover depletion, ground-water storage, losses, and the magnitude of the snowpack in individual years. Similar curves could be constructed on the basis of generated rather than actual runoff, and thereby eliminating the effect of storage. For the cases shown, runoff from CSSL is the least affected by storage (the curves are displaced farthest to the right). Skyland Creek at UCSL and St. Louis Creek at Fraser Experimental Forest possess longer times of storage delay to runoff, and accordingly their curves are displaced to the left.

7-06.03 Figure 6, plate 7-4 shows the snow-covered area (in percent of initial snow cover) plotted against future runoff (in inches over area initially snow covered). As would be expected, there is wide divergence in amount of future runoff associated with a given snow cover early in the season. When snow-covered area is high, the future runoff depends principally upon the water equivalent of snowpack;

as the melt season progresses, the lines converge to indicate future runoff is largely a function of remaining snow cover. Such empirical relationships suggest the possibility of forecasting from direct observation of snow cover the remaining volume of snowmelt runoff after the melt season is under way.

7-06.04 Mathematical expression for snow-cover depletion. - In the absence of observed data, snow-cover values may be obtained from a theoretical snow cover-runoff curve, (Research Note 19). The general expression used to relate snow cover to generated runoff is as follows:

$$A_s = 1.0 - (\Sigma Q_{gen})^n \tag{7-1}$$

where $\underline{A_s}$ is the fractional portion of the basin area which is snow covered, $\underline{Q_{gen}}$ is the generated runoff relative to the total seasonal runoff from the initial snow-covered area, and $\underline{n}$ is an exponent expressing the characteristic basin snow-cover depletion with runoff. For basins which are initially 100 percent snow covered, the runoff summation begins when the basin first begins to go bare. Runoff is expressed in terms of generated flows, and hence, storage effects are not pertinent. The value of $\underline{n}$ reflects the diversity of terrain effects on snow-cover depletion. In the case of WBSL, where a snow wedge adequately defines the variation in the snowpack water equivalent with elevation, a value of $\underline{n}$ = 2 gave reasonable values of snow cover. The curve for the value of $\underline{n}$ = 2 approximates closely the condition of a uniformly ablated snow wedge for a basin with a typical S-shaped area-elevation relationship, and with snowpack water equivalent proportional to elevation above the snowline. A smaller value of $\underline{n}$ would be expected in areas of greater diversity of terrain. In plains regions with uniform deposition of snow and melt, the value for $\underline{n}$ may be 3 or more. Figure 7, plate 7-4, illustrates the rate of snow-cover depletion for various values of $\underline{n}$ in equation 7-1. It is pointed out that these mathematically expressed curves do not account for the reverse curvature which appears near the end of season in some basins, as illustrated by curve A, figure 1.

7-07. METHODS OF ESTIMATING SNOW COVER FROM INDEXES OR DERIVED RELATIONSHIPS

7-07.01 General. - In the derivation of design floods and in seasonal-runoff or rate-of-flow forecasting, it is necessary to evaluate the area of snow cover for the particular melt sequence. In some such procedures, the area of snow cover is implicitly evaluated by another variable which is related to snow cover. For example, many procedures for forecasting seasonal runoff from snow-course data do not account for the area of snow cover directly, but derived relationships between snow-course water equivalent and runoff implicitly include the average relationship between area of snow cover and snowpack water equivalent. However, since there is variability in the relationship between snowpack water equivalent and snow cover, the average relationship can at best only approximate the true volume of water stored in the snowpack.

7-07.02 The most reliable estimate of snow cover is one made from direct observation, as described in section 7-02. In many cases, however, such observations are not feasible, and estimates must be indirectly made from other observed data. Also, once a sufficient period of record of snow-cover observations have been obtained and related to other data, the frequency of making snow-cover observations can be reduced, and snow-cover estimates can be made more quickly and more economically by indirect relationships than by direct observation.

7-07.03 In estimating snow cover from other observed data, derived relationships are used in two ways. One is in obtaining a single estimate of snow cover at a specific time and the second is for estimating day-to-day changes in snow cover. Methods used for determining snow-covered areas are listed below under these two categories:

A. Methods of estimating snow cover at a specific time.

1. Index relations of snow cover to fixed ground or aerial observations, or photographs of snow cover at a point or series of points.

2. Use of snowline observations and area-elevation relationships.

3. Relation of snow cover to point measurements of snow.

a. Water equivalent measurements.
b. Snow depth measurements.

B. Methods of estimating changes in snow cover.

1. Empirical relation of snow-cover depletion to accumulated runoff.

a. Curves derived graphically from observed data.
b. Mathematical equation.

2. Relation of snow-cover depletion to temperature or some index of melt.

3. Use of current precipitation and temperature data to establish areas of new snow, within the accretion or depletion period.

4. Subdivision of basin into elevation zones or homogeneous sub-areas.

7-07.04 Indexes of snow cover. - Methods of observing snow cover, from a point, either on the ground or from the air, have been described in section 7-02. The quantitative evaluation of basin snow cover using point observations as indexes, required simultaneous observations of the index and of basin snow cover until a relationship has been established between the two. Average snowline elevations may be used with area-elevation relationships to determine snow cover in mountainous regions, particularly during the accumulation period, but care should be exercised in their use during the melt period.

7-07.05 Snow course measurements of snowpack water equivalent at one or more snow courses may be related to snow cover as a simple function. Obviously, snow courses selected for use in this relationship should be those on which snow remains for the longest possible time. The principal deficiency of the method is that such a simple correlation does not account for variation in slope of the snow wedge. In Research Note 22, the area of snow cover of the North Santiam River basin above Detroit Dam was expressed as a function of the ratio between the snowpack water equivalents of two snow courses at different elevations. Only one season's snow-cover observations were available, however, so the reliability of the method on this basin cannot be assessed. Snow-depth observations are useful primarily in defining the times that areas become bare of snow for given locations and elevations. Care should be taken in selection of the point(s) to secure representativeness of basin conditions, both with regard to snow deposition and melt. In the West, snow-course measurements are made at monthly or bimonthly intervals, so that their use is limited to the times of observations. Snow-depth measurements at weather observation stations are available daily during the period of snow cover.

7-07.06 Estimates of basin snow-cover depletion. - A knowledge of the change in snow cover between times of observation during the melt period is required for many snow-hydrology problems. The most feasible method is to assume snow cover to vary with some continuous function, such as runoff, time, or a melt index. Section 7-06 described relationships between snow-cover depletion and accumulated runoff, based on snow-laboratory observations. The procedures have been used by the Seattle District on the Kootenai and Flathead River basins, as described in Technical Bulletin 15. The available observations are insufficient, however, to derive general relationships for these areas. Empirical relationships in graphical or mathematical forms may be used to relate snow-cover depletion to runoff according to methods set forth in section 7-06. The use of a time function alone to express depletion is not too reliable because of the variations in melt (and hence depletion), with time. A simple index of melt, such as degree days, may be used to evaluate depletion, but the use of accumulated generated runoff is considered to be more practical because (1) it integrates all factors affecting melt, (2) it is simpler to use than melt indexes, and (3) it is readily available.

7-07.07 During the period of snow depletion, snow-cover estimates may be improved by use of current temperature and precipitation data. The purpose is to delineate areas of shallow new snow which contribute little to runoff. Once the areas covered by new snow are evaluated the time required to melt the new snow in order to re-establish the snow-cover depletion rate of the old snow can be determined.

7-07.08 A different approach to the determination of snow-cover depletion is that of subdividing a basin area into zones of equal elevation. Beginning with an assumed or known distribution of snowpack water equivalent with elevation, values of snowpack water equivalent are determined for each zone. By maintaining an inventory of snowpack accumulation and ablation, the depletion of snow on successive elevation zones is determined. The principal difficulty of the method is in the evaluation of precipitation distribution with elevation, particularly for heterogeneous areas. A refinement of the method is to assume a non-uniform distribution of snowpack water equivalent within a given zone.

7-08. APPLICATION OF SNOW-COVER OBSERVATIONS TO BOISE RIVER BASIN

7-08.01 General. - Of recent years, the Walla Walla District of the Corps of Engineers has determined the area of snow cover on various drainage basins within their district by means of aerial reconnaissance. Some of this information has been used by the Snow Investigations in studies of daily snowmelt and streamflow for the Boise River near Twin Springs, Idaho (D.A = 830 sq. mi.). Results of those studies are presented in chapters 6 and 9. The importance of snow cover in these studies led to a detailed analysis of snow cover on the Boise River basin during the 1954 and 1955 melt seasons. These analyses are described in this section.

7-08.02 Description of 1954 and 1955 seasons. - The snowpack on 1 April was above normal in 1954 and somewhat below normal in 1955. During April, 1954, melting conditions prevailed and light precipitation fell, principally in the form of rain. April, 1955, on the other hand, was characterized by below normal temperatures and above normal precipitation, thereby resulting in a large increase in accumulation of snow. The major portion of the snowpack ablation occurred during May of both years. In 1954 the last few days of May and the first half of June were cold and wet, thereby retarding the melt of the remnants of the snowpack.

7-08.03 Progression of snow-cover depletion. - Plate 7-5 presents the results of the aerial observations of snow cover on the Boise River basin above Twin Springs, Idaho, during the 1954 and 1955 melt seasons. Principal streams and elevation contours are shown on each of the basin maps to convey a general idea of the topography. Figure 3 on plate 7-5 shows the location of hydrometeorological stations within the basin and in the surrounding area. Plate 7-6 shows the hydrometeorological events for each of the two years, including estimated basin precipitation (both rain and snow are shown separately), mean daily temperature at Atlanta (elev. 6000 ft msl), daily discharge hydrographs for Boise River near Twin Springs, Idaho, for the period March through June, and snowpack water equivalents for snow courses within the basin or adjacent areas for those dates for which records are available. Also shown are the observed snow-cover data, plotted on the same time scale. For times between observations, the 1955 values are interpolated by means of the snow cover-generated runoff relationship (corrected for subsequent precipitation), shown in figure 3, plate 7-6. For 1954, estimates of snow cover between observations were made by drawing a smooth curve drawn through the four observed points thus defining the snow-cover depletion only in a general way. Precipitation and temperature data, between dates of observation, suggest significant deviations between the actual cover and that shown by the snow cover-time curve.

7-09. SUMMARY AND CONCLUSIONS

7-09.01 Snow-cover information is, like temperature or heat supply, an important hydrometeorological element. Snow cover is a factor in all hydrologic problems which involve basin snowmelt. At present systematic snow-cover surveys are being made in a number of basins. Despite the complexity of the variables affecting the snow-cover depletion, the hydrologist is able to approximate the snow-covered area between snow-cover surveys using available hydrometeorological data.

7-09.02 The recession of snow cover is very slow early in the melt season compared to ablation of the snowpack or runoff. Since areas of homogeneous heat supply exhibit uniform melt rates, the snow cover depletes gradually to a thin layer. A sudden increase in depletion then takes place. In years with very deep snow, a large amount of snow-depth reduction or runoff takes place before the appearance or substantial enlargement of the snow-free areas in the basin. The principal factors affecting the snow-cover depletion are the variations in snow deposition and variations in snowmelt, both of which are affected by terrain features, including orientation, steepness of slope, elevation, and forest cover. The snow-cover depletion-runoff patterns vary between basins in accordance with the difference in topography and ground-water character. Variation is also expected within each watershed from year to year on account of differences in the snowpack accumulation at the onset of the snowmelt runoff season.

7-09.03 During the winter accumulation period the determination of the snow cover is relatively simple and accurate; the snowline is well defined and coincides with an elevation contour. The area-elevation curve is used in determining the snow-covered area. In the absence of snowline or snow-cover surveys, a current snow-course survey may be used to determine snowline elevations. The water equivalent-elevation curve, even though poorly defined, will indicate the average elevation below which no snow exists on the drainage basin. The snow-covered area, as of the date of snow-course curvey, can be determined from the area-elevation curve or from the snow chart, shown in figure 1, plate 4-2 of chapter 4. The snowline elevation subsequent to the most current snowline survey can be estimated by reducing the snow wedge at the time of the survey, by an amount proportional to heat supply, or by lowering the snowline elevation if subsequent precipitation, in the form of snow, caused the snowline to advance to a lower elevation.

7-09.04 During the active melt season, the determination of an average snowline is not dependable because the snowline is not as well defined as in the accumulation period. The lower portion of the snow wedge is quite ragged or patchy for 1000 feet or more in elevation. In general, this ragged zone is higher on southerly slopes than it is on northerly slopes. It is less patchy and lower in a heavily forested area than on an open slope of same exposure. During the period of active melt season, the use of an average snowline elevation for determining the snow-covered area can be considered only a rough approximation. The most dependable basin snow-cover estimates are made from aerial reconnaissance surveys. Snow cover between surveys is determined in accordance with runoff and with the meteorological events affecting new snow cover. A characteristic of snow-cover depletion is the definite pattern in which snow depletes from year to year on a given basin. As a result of this year-to-year uniformity, only a few sites, representative of the topography of the watershed, need be observed as an index to snow cover.

7-09.05 The determination of snow-covered area by means of established "cover-runoff" or "cover-ablation" curves is accomplished from analysis of historical data. If accumulated runoff is plotted in percentage of the season's total from beginning of the appearance of the effective spring melt at the stream gaging station, the curves will tend to be close together and serve as guide for the extrapolation of the snow-cover recession for the melt season considered. Undoubtedly "cover-mass" relationships can be improved if a parameter such as the initial basin snow cover is used. The relation between snow cover and "future runoff" provides a method for estimating residual runoff. Forecasts based on those relationships are particularly useful in connection with regulation of reservoirs near the end of the filling period.

7-10. REFERENCES

1/ BUTSON, K.D., "Snow cover observation program, U.S. Weather Bureau," Proc. West. Snow Conf., Boise, Idaho, April 1953, pp. 5-6.

2/ CROFT, A.R., "Some factors that influence the accuracy of water-supply forecasting in the Intermountain Region," Trans. Amer. Geophys. Union, Vol. 27, No. 11, 1947, pp. 375-388.

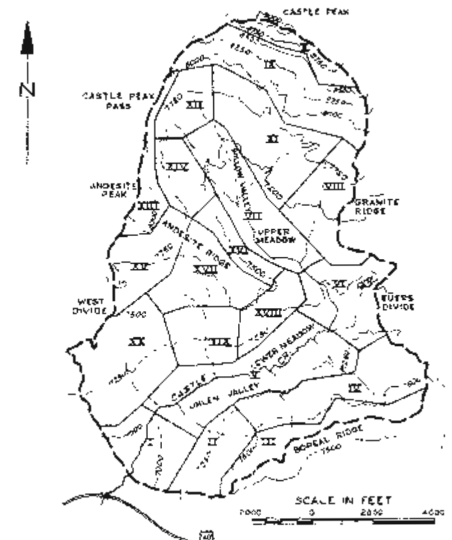
3/ DUNFORD, E.G. and L.D. LOVE, "The Fraser Experimental Forest, its work and aims," U.S.F.S., Rocky Mountain Forest and Range Exp. Sta. Paper 8, May, 1952.

4/ KITTREDGE, Joseph, "Influences of forest on snow in the ponderosa-sugar pine-fir zone of the central Sierra Nevada," Hilgardia, Vol. 22, No. 1, March 1953, Univ. Calif. Press.

5/ POTTS, H.L., "A photographic snow-survey method of forecasting runoff," Trans. Amer. Geophys. Union, Vol. 25, Part I, September 1944, pp. 194-153.

AERIAL PHOTOGRAPH, CSSL, SHOWING FOREST COVER

FIGURE 1



TOPOGRAPHIC UNITS, CSSL

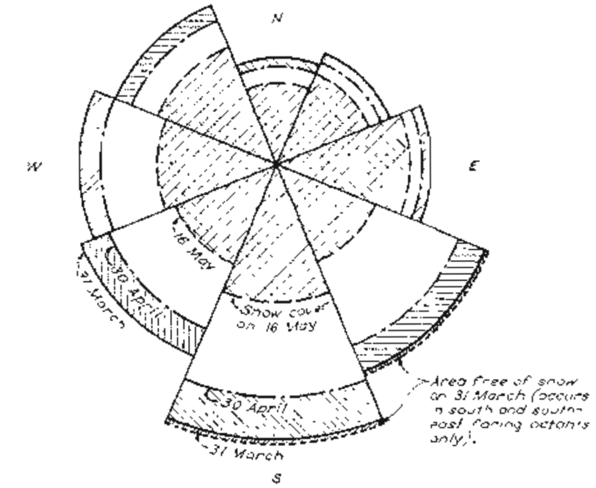
DRAINAGE AREA 3.96 SQUARE MILES

FIGURE 2

COVER vs ORIENTATION

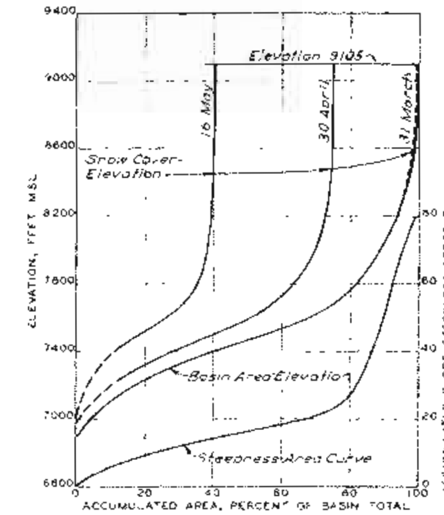
ORIENTATION	AREA IN % BASIN	SNOW COVER IN % BASIN 31 MAR	30 APR	16 MAY	MEDIAN EL.
N	4.0	4.0	3.3	2.3	7420
NE	5.0	5.0	4.1	3.5	7550
E	8.0	8.0	7.0	6.0	7620
SE	18.0	17.6	13.3	4.0	7410
S	27.0	26.1	18.2	6.7	7510
SW	15.0	15.0	11.1	4.1	7610
W	13.0	13.0	10.3	4.9	7530
NW	10.0	10.0	8.3	5.6	7360
TOTAL	100.0	98.7	75.6	37.1	

Note:
Areas shown in each octant are directly proportional to snow covered areas for that octant in basin.



PROGRESS OF SNOW COVER DEPLETION WITH RESPECT TO ORIENTATION CSSL, 1947

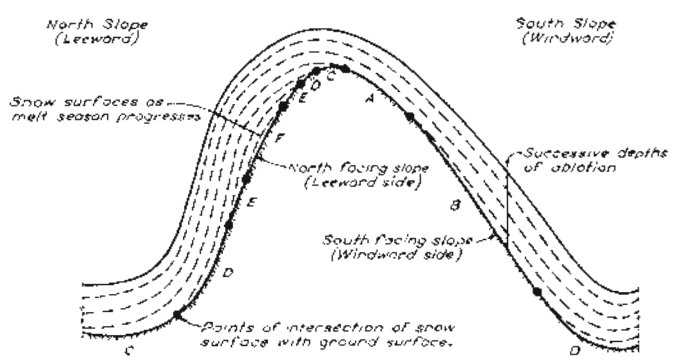
FIGURE 3



SNOW COVERED AREA-ELEVATION CURVES, CSSL, 1947

AREA-STEEPNESS CURVE, CSSL

FIGURE 4



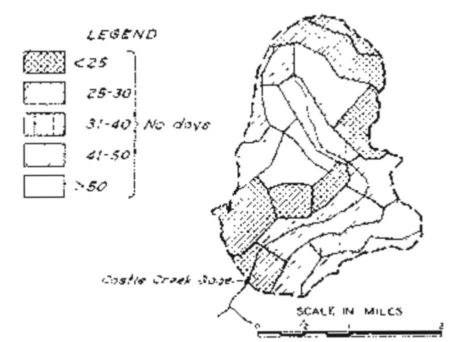
Note:
This diagram illustrates the progress of depletion of the snow cover during period of active melt. The sequence of appearance of bare ground are lettered successively beginning with the letter A. Relative melt rates are shown by the vertical distance between snow surface lines corresponding to various dates. Rate of snow cover depletion for a particular area depends chiefly on the orientation of the slope, which affects both accumulation of snow and the supply of heat for melting the snow. Note the drift effect on the accumulation of snow on windward and leeward slopes near edge of crest.

SCHEMATIC DIAGRAM OF SNOW COVER DEPLETION, UNFORESTED SLOPES, CSSL

FIGURE 5

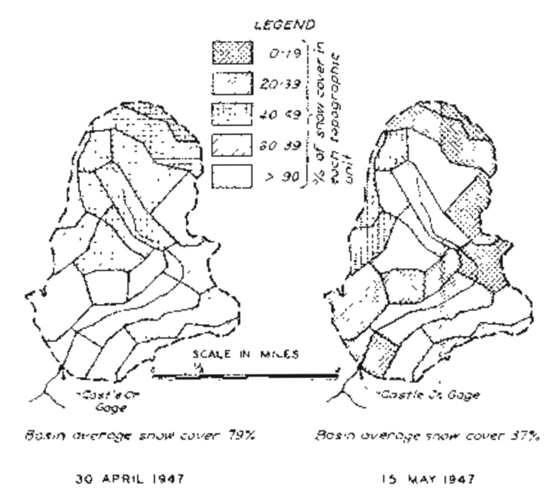
DATE WHEN SNOW COVER IS REDUCED TO 60% OF AREA OF TOPOGRAPHIC UNIT

TOPO UNIT	DATE	DAYS AFTER 10 APR	TOPO UNIT	DATE	DAYS AFTER 10 APR
I	4 May	24	XIII	13 May	29
V	7 "	27	XIV	5 June	55
VI	7 "	27	XVI	10 "	60
II	9 "	29	VII	20 May	40
III	9 "	29	XII	7 "	27
IV	20 "	40			
			VIII	2 "	22
XV	7 "	27	IX	28 Apr	18
XVII	15 "	35	X	23 "	13
XVIII	1 "	21	XI	1 June	50
XIX	3 "	23			
XX	30 Apr	18			



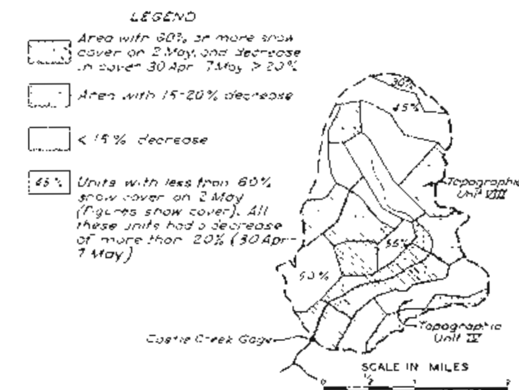
DAYS AFTER 10 APRIL 1947 WHEN 60 PERCENT OF THE AREA IN EACH UNIT WAS COVERED WITH SNOW

FIGURE 6



AREAL DISTRIBUTION OF SNOW COVER

FIGURE 7



SNOW COVER AT TIME OF PEAK FLOW ON 1 MAY 1947

FIGURE 8

Notes:

1. *On 31 March snow cover was 100 percent over all topographic units, except on a few very steep slopes of high elevations.*
2. *The basin snow water equivalent depth reached its maximum value on about 10 April when snow melt runoff began to appear at the gaging station in Castle Creek.*
3. *On 15 April approximately 92 percent of the basin was covered with snow.*

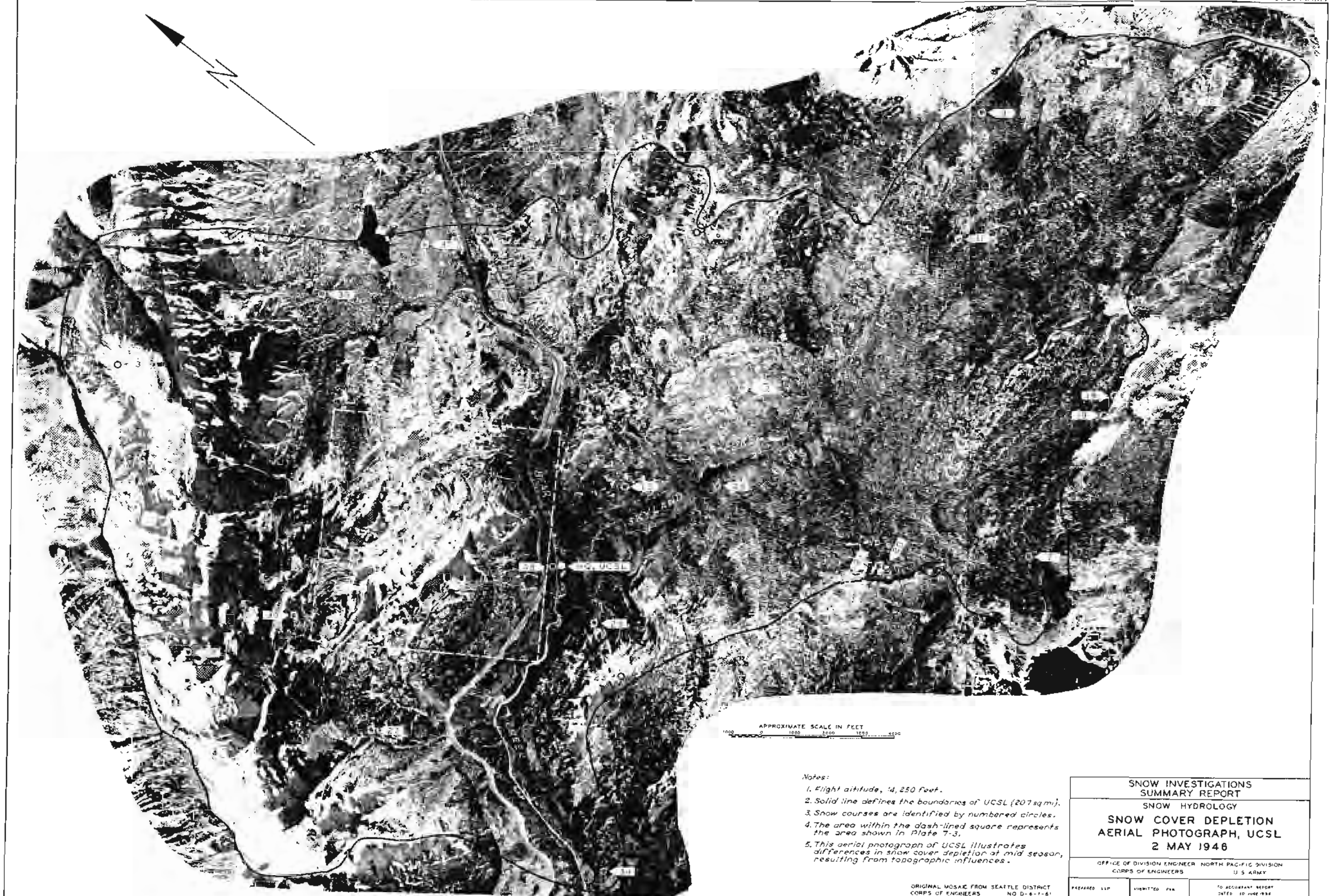
SNOW INVESTIGATIONS SUMMARY REPORT

SNOW HYDROLOGY

SNOW COVER DEPLETION CSSL, 1947

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U.S. ARMY

PREPARED T.S.S. / DRAWN B.J.	SUBMITTED J.G.B. / APPROVED T.M.M.	TO ACCOMPANY REPORT DATED 30 JUNE 1956 PD-20-25/40



APPROXIMATE SCALE IN FEET
0 1000 2000 3000 4000

Notes:

1. *Flight altitude, 14,250 feet.*
2. *Solid line defines the boundaries of UCSL (207 sq mi).*
3. *Snow courses are identified by numbered circles.*
4. *The area within the dash-lined square represents the area shown in Plate 7-3.*
5. *This aerial photograph of UCSL illustrates differences in snow cover depletion at mid season, resulting from topographic influences.*

ORIGINAL MOSAIC FROM SEATTLE DISTRICT CORPS OF ENGINEERS NO D-6-7-61

SNOW INVESTIGATIONS SUMMARY REPORT

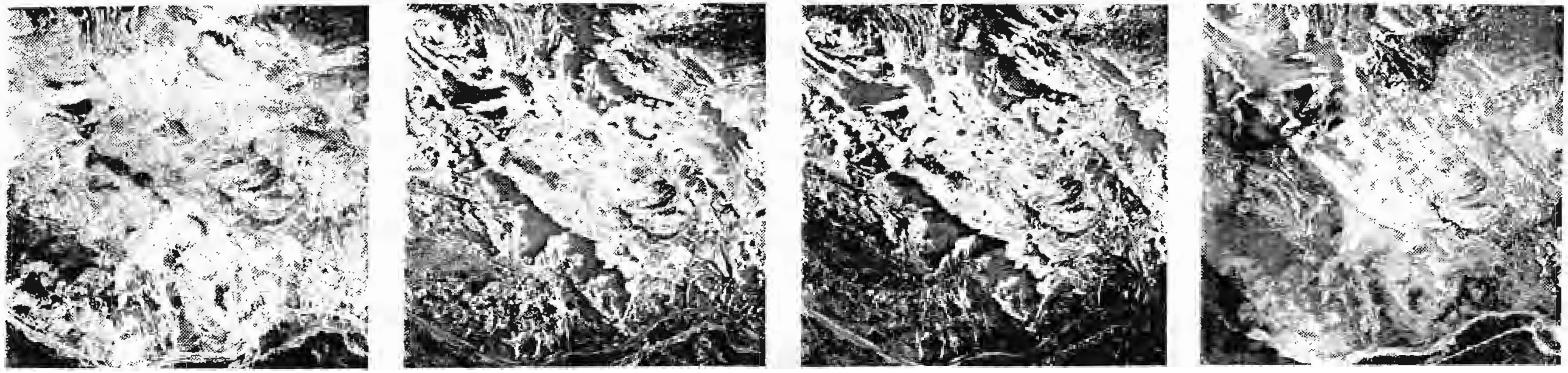
SNOW HYDROLOGY

SNOW COVER DEPLETION AERIAL PHOTOGRAPH, UCSL 2 MAY 1946

OFFICE OF DIVISION ENGINEER NORTH PACIFIC DIVISION CORPS OF ENGINEERS U. S. ARMY

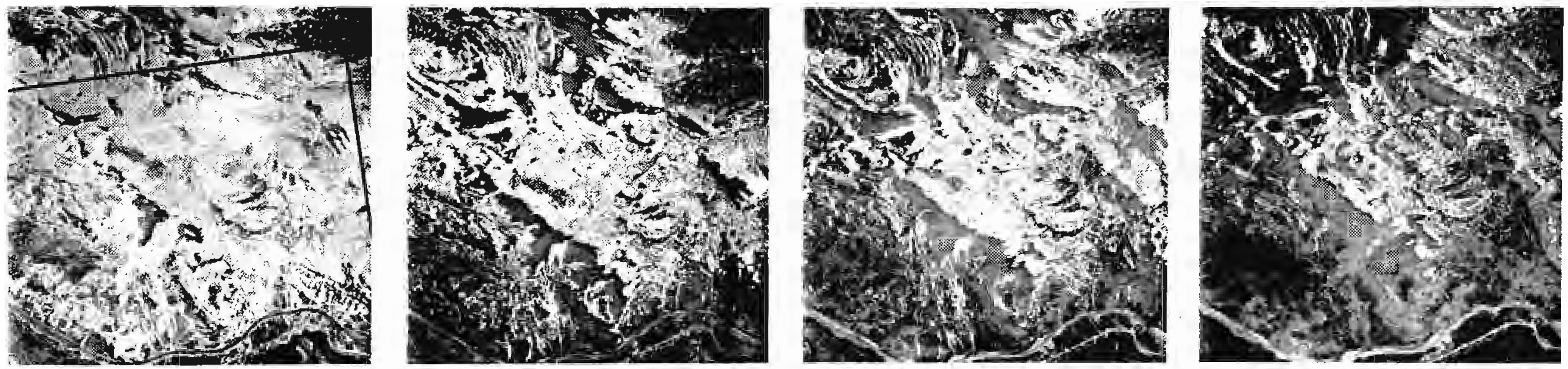
PD-20-25/41

PLATE 7-2



7 APRIL 24 APRIL 2 MAY 17 MAY

FIGURE 1 — 1946 AERIAL PHOTOGRAPHS



16 APRIL 3 MAY 8 MAY 20 MAY

FIGURE 2 — 1947 AERIAL PHOTOGRAPHS

Note:
The photographs illustrate that each year the sequence of areas going bare is the same. The pattern of growth of bare patches is similar each year if the forest cover is unchanged. The area encompassed in each photograph of this series is approximately one square mile. See Plate 7-2 for general location and orientation of area.

SNOW INVESTIGATIONS SUMMARY REPORT

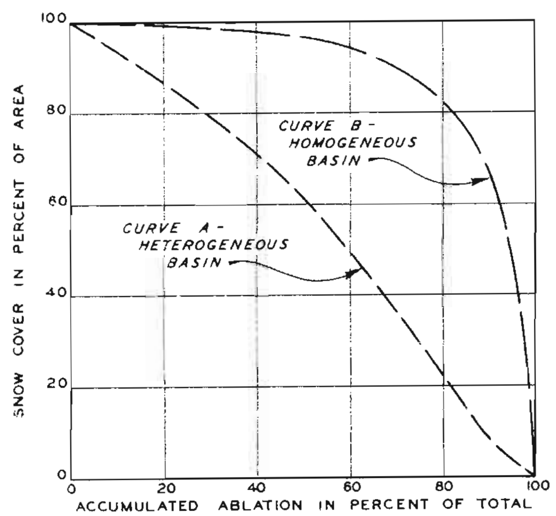
SNOW HYDROLOGY

PROGRESS OF SNOW-COVER DEPLETION UCSL, 1946 AND 1947

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION CORPS OF ENGINEERS U.S. ARMY

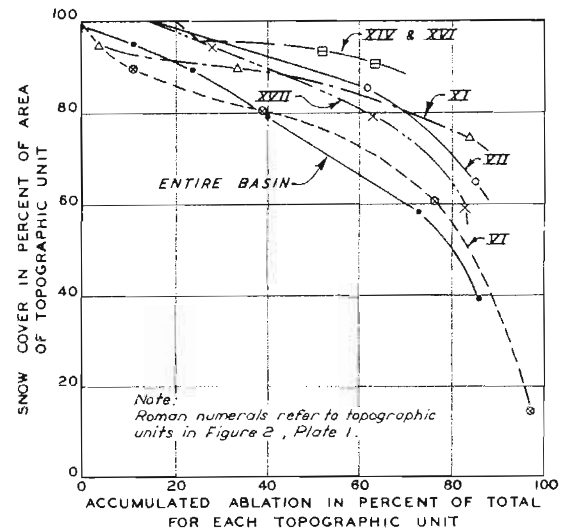
PREP: ...... SUBM: ...... TO ACCOMPANY REPORT DATED 30 JUNE 1956

DRAWN: ...... APPR: PD-20-25/42



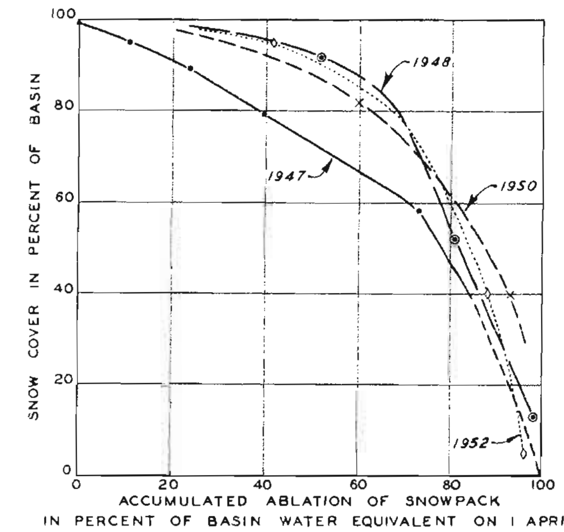
SCHEMATIC DIAGRAM OF SNOW COVER DEPLETION - ABLATION RELATIONSHIP

FIGURE 1



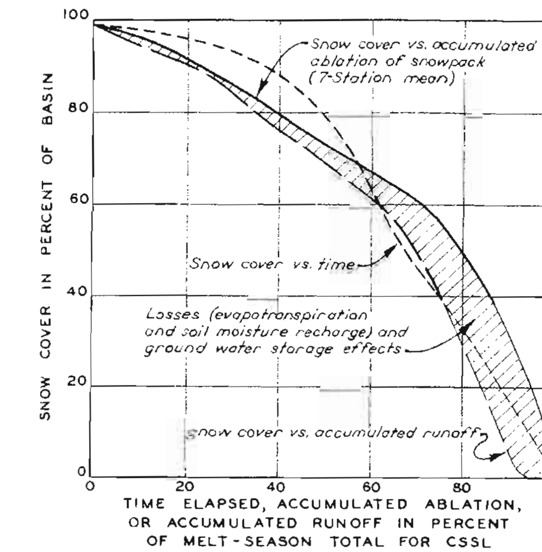
SNOW COVER DEPLETION - ABLATION RELATIONSHIP, CSSL, 1947

FIGURE 2



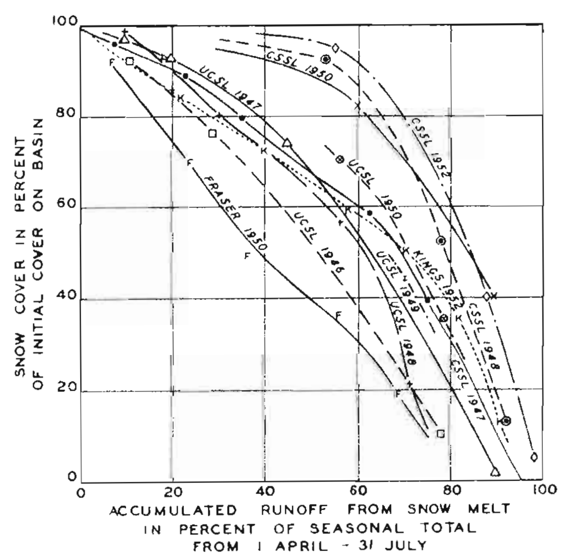
SNOW COVER DEPLETION - ABLATION RELATIONSHIP, CSSL

FIGURE 3



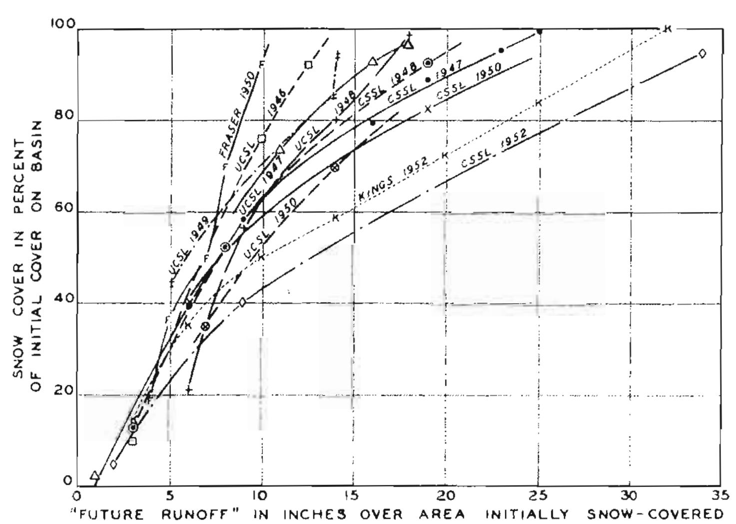
SNOW COVER DEPLETION, ABLATION AND ACCUMULATED RUNOFF, CSSL, 1947

FIGURE 4



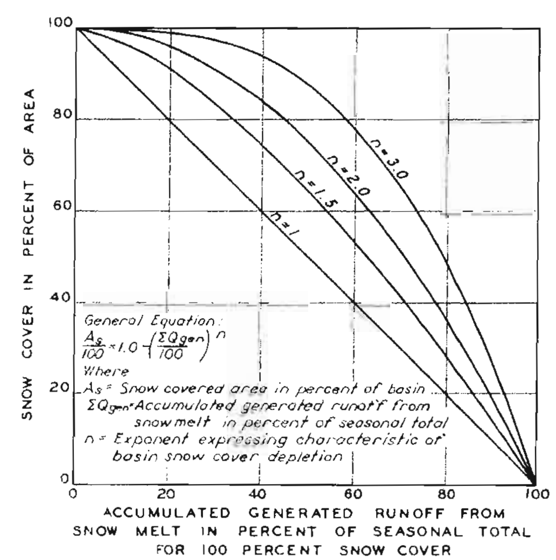
SNOW COVER DEPLETION - RUNOFF RELATIONSHIPS

FIGURE 5



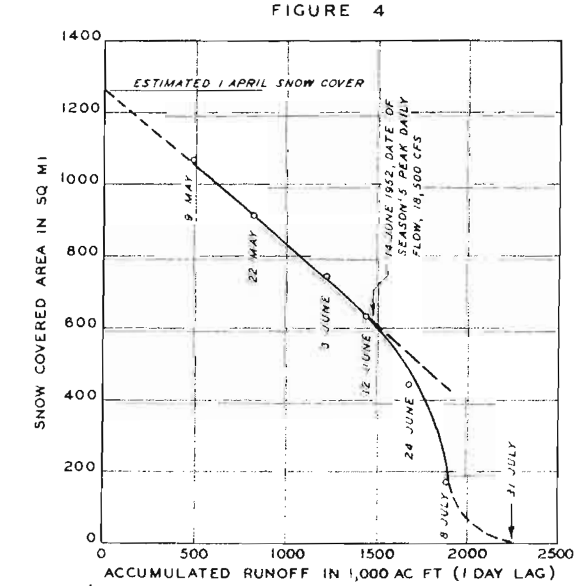
SNOW COVER - "FUTURE RUNOFF" RELATIONSHIPS

FIGURE 6



MATHEMATICAL EXPRESSIONS FOR SNOW COVER DEPLETION

FIGURE 7



(DAILY OBSERVED INFLOW TO PINE FLAT RESERVOIR, SACRAMENTO DISTRICT, CALIFORNIA, 1 APRIL-31 JULY 1952)

SNOW COVER-RUNOFF RELATIONSHIP KINGS RIVER, CALIFORNIA

FIGURE 8

Notes for Figures 2, 3, 4, 5 and 6:

1. Accumulated runoff from 1 April - 31 July is corrected for rainfall between snow cover surveys but not adjusted for ground water recession flow.
2. Initial snow cover is the snow covered area as of 1 April.
3. In Figure 3, if the accumulation for 1948, 1950 and 1952 began at 98 percent snow cover the curves would fall close to 1947 curve.

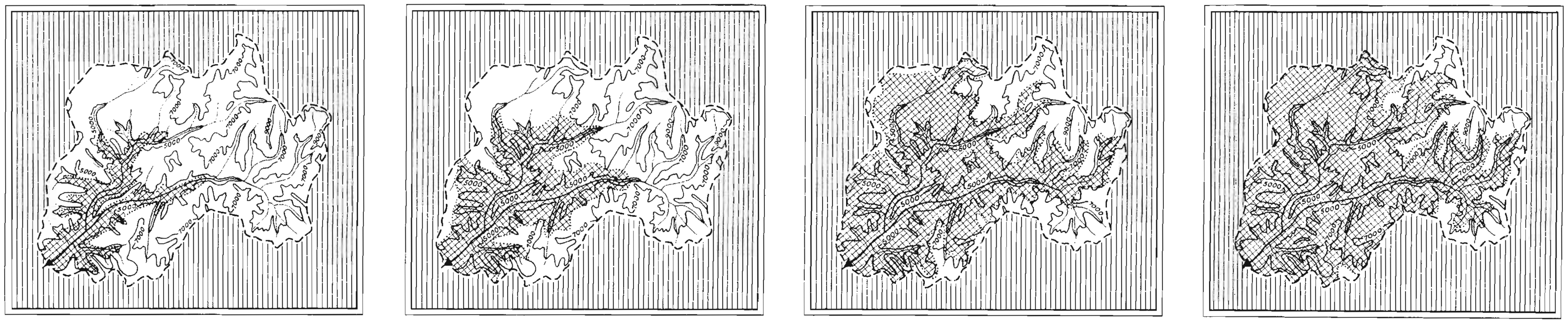
SNOW INVESTIGATIONS SUMMARY REPORT

SNOW HYDROLOGY

SNOW COVER DEPLETION, ABLATION OF THE SNOWPACK, AND RUNOFF

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION CORPS OF ENGINEERS U. S. ARMY

PREPARED P.B.B.	SUBMITTED P.B.B.	TO ACCOMPANY REPORT DATED 30 JUNE 1956
DRAWN H.L.S.	APPROVED D.M.R.	PD-20-25/43



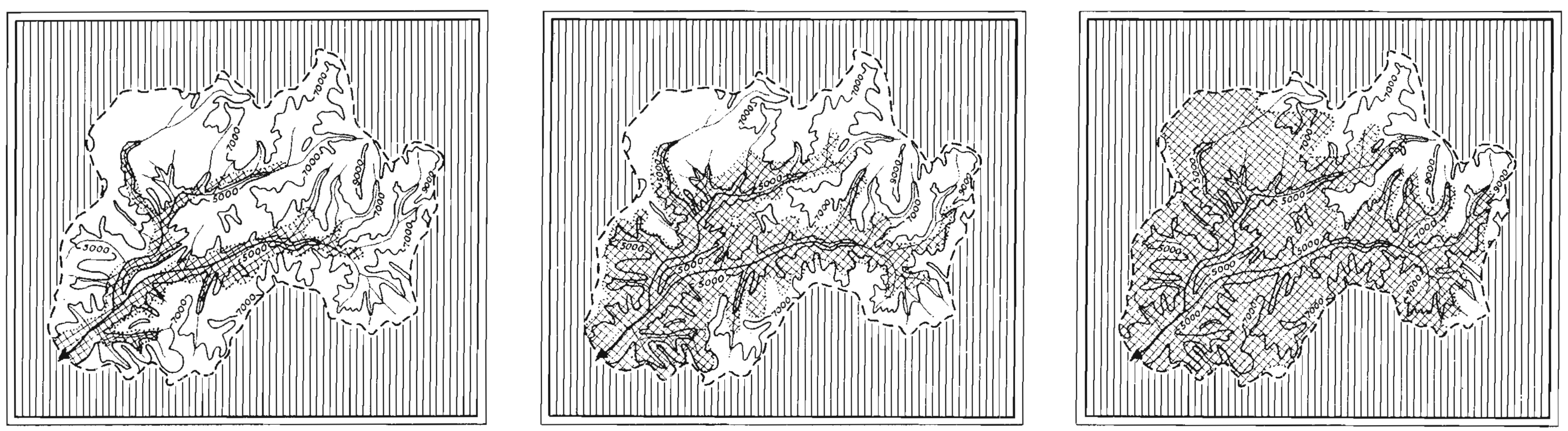
31 MARCH 1954 - 86% SNOW COVER

22 APRIL 1954 - 69% SNOW COVER

6 MAY 1954 - 38% SNOW COVER

20 MAY 1954 - 18% SNOW COVER

FIGURE 1 - 1954 AERIAL RECONNAISSANCE



5 MAY 1955 - 86% SNOW COVER

22 MAY 1955 - 59% SNOW COVER

8-9 JUNE 1955 - 22% SNOW COVER

FIGURE 2 - 1955 AERIAL RECONNAISSANCE

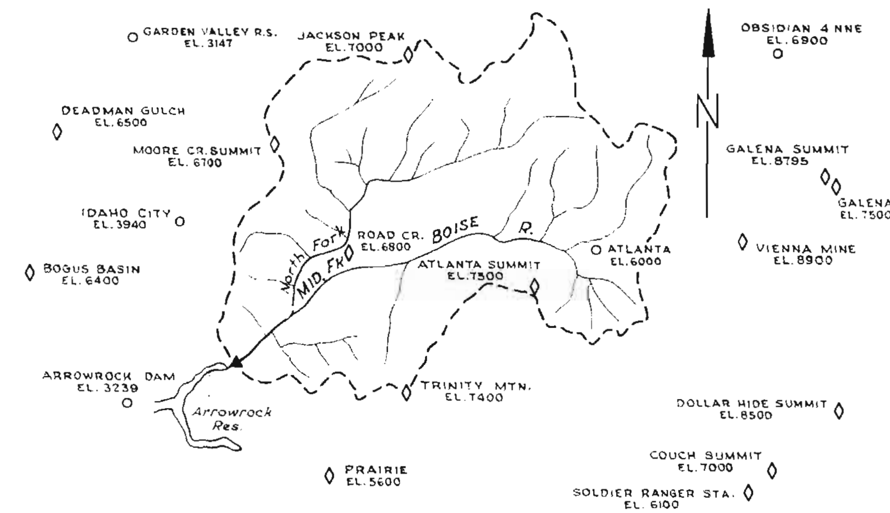


FIGURE 3 - HYDROMETEOROLOGICAL STATIONS

LEGEND

- Snow-free Area } From aerial snow line reconnaissance on dates shown.
- Snow Covered Area }
- ○ Meteorological Station
- ◊ Snow Course
- ▲ Stream Gage

SUMMARY OF WEATHER

YEAR	MARCH	APRIL	MAY	JUNE
1954	Temperature slightly below normal. Precipitation 80% of normal. Periods of major storms: 8-12, 7-21, 24-28. Snowfall at lowest elevation on 20th.	Temperature and precipitation slightly above normal. Maximum snow accumulation during first week. Storms: 3-7, 9-10, 13-15, 18-19, 20-21, 27-28, 30. Effective snow-melt runoff began on 10th.	Above normal temperatures. Peak stream flow on May 21st. Storms: 1, 16, 21-23, 25-31	Subnormal temperature. Storms: 1-2, 5-8, 8-13, 15-17, 26-29.
1955	Temperature considerably below normal. Precipitation near normal with snow at low levels. Storms: 1-5, 9-16, 22-24, 28-31.	Temperature below normal. Precipitation above normal. Storms: 1-4, 10-30. Effective snow-melt runoff began on 28th.	Temperature below normal. Precipitation above normal. Maximum snow accumulation during first week. Storms: 1-4, 14-17, 21-22, 25-28.	Temperature below normal. Precipitation near normal. Peak stream flow on 10th. Storms: 2-3, 12-15, 24, 26, 28-29.

Note: Snow cover observations by aerial reconnaissance obtained from Walla Walla District, U. S. Corps of Engineers.

SCALE IN MILES

10 0 10 20

SNOW INVESTIGATIONS SUMMARY REPORT

SNOW HYDROLOGY

SNOW COVER OBSERVATIONS 1954-55

BOISE RIVER ABOVE TWIN SPRINGS, IDAHO

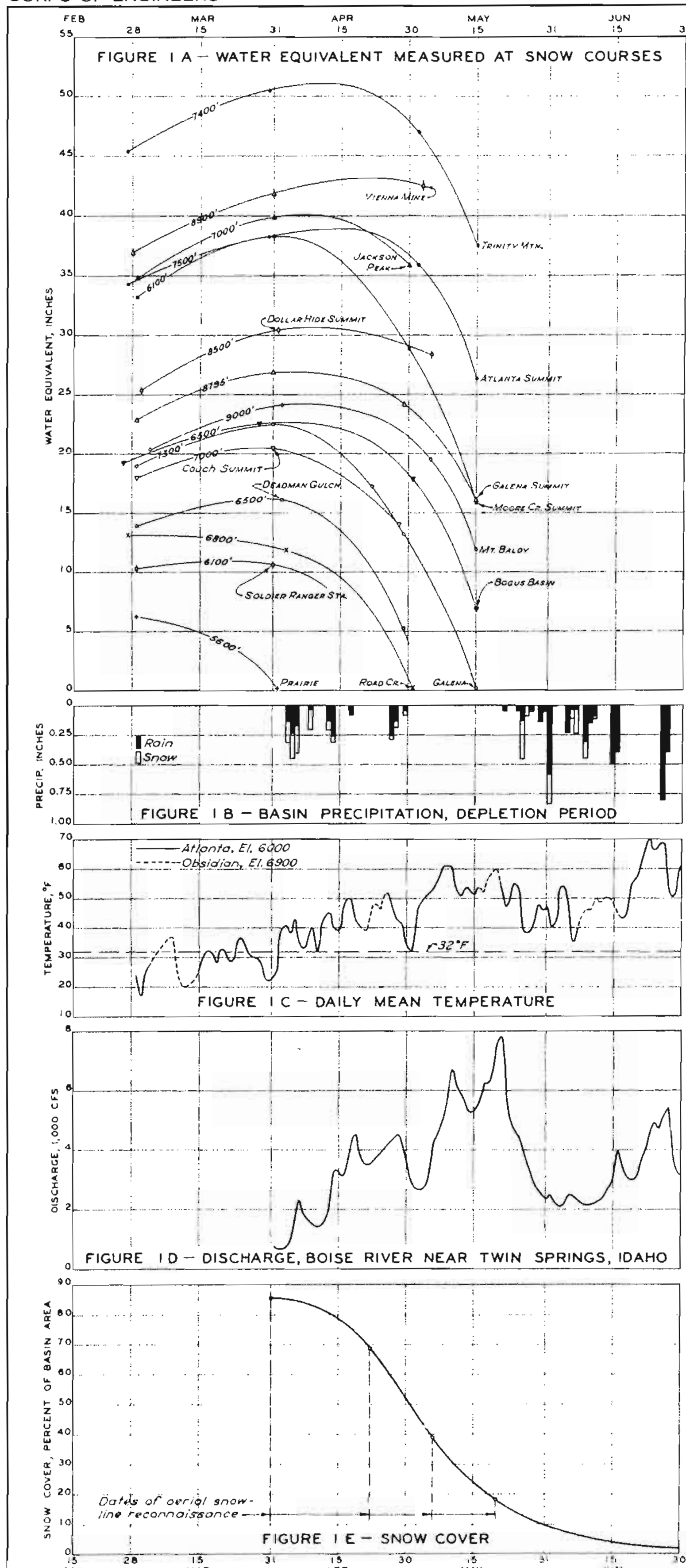
DRAINAGE AREA 830 SQ. MI.

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

PREPARED: PM SUBMITTED: PBB
DRAWN: WJM APPROVED: DMR

TO ACCOMPANY REPORT DATED 30 JUNE 1956

PD-20-25/44



NOTES for FIGURES 1A and 2A:

1. Lines between points are primarily for identification of snow courses and show only the general trend of the water equivalent variation between measurements. Actually the accumulation and depletion of the snowpack is very irregular.
2. Location of snow courses is shown in Figure 3, Plate 7-5.

NOTE for FIGURES 1B and 2B:

Basin precipitation computed from 4-Station average, adjusted to represent basin amounts on basis of normal annual precipitation.

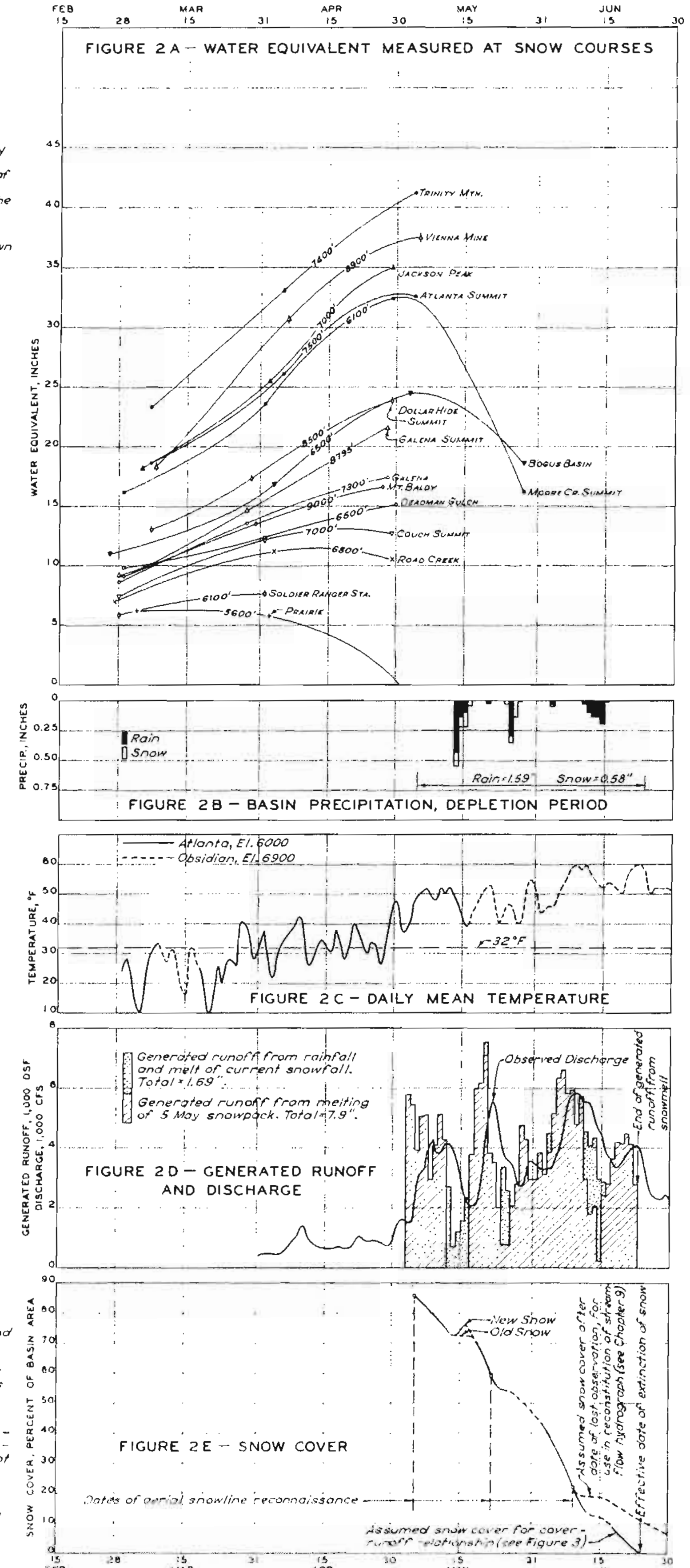
NOTES:

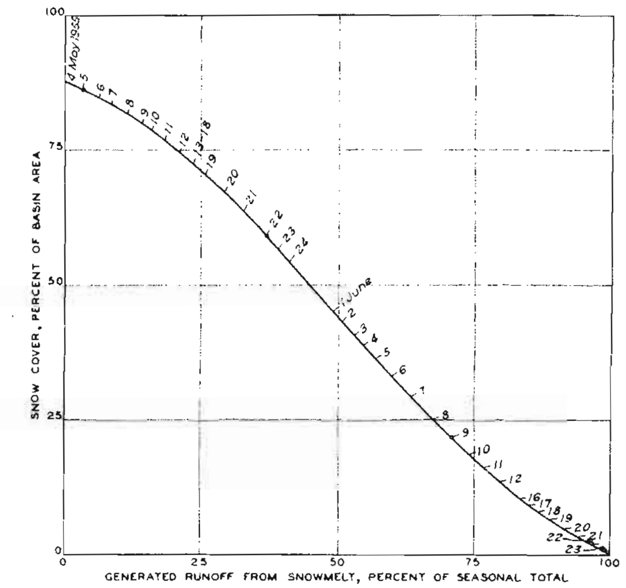
FIGURE 1E

Curve shows only the general trend between observations of snow cover. No attempt was made to estimate the probable values between observation dates as was done for 1955, Figure 2E.

FIGURE 2E

1. Snow cover during periods of precipitation (---), estimated by considering temperature and percent of basin above estimated snow isotherm.
2. Snow cover between storms and observations interpolated with aid of Figure 3, upper right.





NOTE:

This curve is derived from data found in Figures 2D and 2E. The effect of new snow and rain (coverage and runoff) was estimated and the curve adjusted accordingly to represent the cover-runoff relationship for the basin snow as observed on 5 May 1955. This is the date when appreciable snowmelt rise in the discharge hydrograph began. Season's total runoff from the melt of the 5 May snowpack (with 86% cover) was 7.9 inches.

FIGURE 3 – SNOW COVER, GENERATED RUNOFF RELATIONSHIP – 1955

SNOW INVESTIGATIONS
SUMMARY REPORT

SNOW HYDROLOGY

SNOW COVER DEPLETION 1954 AND 1955

BOISE RIVER ABOVE TWIN SPRINGS, IDAHO
DRAINAGE AREA 830 SQUARE MILES

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

PD-20-25/45

CHAPTER 8 - EFFECT OF SNOWPACK CONDITION ON RUNOFF

8-01. INTRODUCTION

8-01.01 General. - The storage effect of the snowpack is important in the evaluation of both the volume and the time distribution of runoff. For equal melt values and losses, generated runoffs may not be the same because of the snowpack condition. An initially "cold" (sub-freezing) snow will freeze a certain amount of liquid water entering it and thereby raise the temperature of the snow to the melting point. An additional amount is required to satisfy the liquid-water-holding capacity before the snow will release any water by gravity. If, however, the entire snowpack is already saturated and conditioned to yield water, all inflow will pass through the pack to the ground without depletion; the time delay depends mainly on the depth of snow, the resistance to flow, and the rate of inflow. In order to assess the snowpack condition and its effect upon runoff, it is necessary first to understand the changing character of the snowpack and the processes of heat and water-vapor transfer within the pack. Special experiments on the storage and transit of liquid water in the snowpack at CSSL, although somewhat meager and inconclusive, furnish information which serves as a guide to solution of problems involving storage and travel time of water in the snowpack.

8-01.02 The physical properties, which affect the liquid-water retention and detention capacities of the snow, continue to change from the time of deposition to melt. Even during the active melt season the proportions of ice, liquid water, and air are not constant. Likewise, the permeability and diffusivity of the snowpack to heat, air, and water are continually changing.

8-01.03 Character of the snowpack. - Snow is a precipitate. Ice crystals are formed in the atmosphere at temperatures below freezing by sublimation of water vapor on hygroscopic nuclei. Many different types of crystals form, depending on the shape of the nucleus, the rate of sublimation, and the turbulence of the air. An excellently illustrated discussion of the relation between snow crystal types, forms, mass, rate of fall, and crystal habits is found in Snow Crystals by Nakaya.11/ Because of the usual dendritic structure of these crystals, new-fallen snow is generally of low density. With time, however, the snowpack undergoes a change: the original, delicate crystals of snow become coarse grains, and the density of the pack increases. There is no definite time at which the change takes place from crystalline forms characteristic of new-fallen snow to coarse grains in the snowpack. The crystalline structure of the snow, as used in this chapter, is a general term referring to classification of either the snow crystals themselves, or the snow grains resulting from metamorphism of the snowpack. The change from a loose, dry, and subfreezing snowpack of low density to a coarse, granular, and moist snowpack of high density is spoken of as "ripening" of the snowpack. A "ripe" snowpack is said to be "primed" to produce runoff (barring temporary ponding due to resistance to flow or inadequate channel capacity). A ripe snowpack is not necessarily homogeneous; it is generally striated with ice planes and ice lenses. Density of the snowpack, which is an easily measurable element, appears to be a single variable that integrates fairly well the effect of the other physical properties of the snow. It has been used for defining the affinity of the snow for water, as well as the thermal properties of snow. Knowledge of the factors affecting the metamorphism of the snowpack will facilitate the understanding and solution of the heat transfer, the liquid-water storage, and transmission problems in snow hydrology.

8-02. METAMORPHISM OF THE SNOWPACK

8-02.01 General. - The change in the character of the snowpack has been studied at length in connection with the use of snow as an engineering material. No consideration will be given here to variation in such structural qualities as hardness, strength, or trafficability. The hydrologist's concern with the metamorphism of the snowpack is primarily limited to the function of the snowpack as a deterrent to runoff. The presentation in this section is intended to give the hydrologist a limited summary of the effects of the metamorphism of the snowpack as regards its role in the hydrologic cycle.

8-02.02 Factors affecting the metamorphism of snow. - Time is the principal factor to be considered in the metamorphism of snow. The several physical processes contributing to metamorphism of snow are: (1) heat exchange at the snow surface by radiation, convection, and condensation; (2) percolation of melt or rain water through the snowpack; (3) internal pressure due to the weight of the snow; (4) wind; (5) temperature and water-vapor variation within the snowpack; and (6) heat exchange at the ground surface. The effects of these processes which are of hydrologic importance are: (1) change in density as the result of change in crystal forms and displacement of the crystalline particles with respect to one another; (2) formation of ice planes; (3) change in air, water, and heat permeability and diffusivity; (4) change in liquid-water-holding capacity; and (5) change in temperature of the snowpack.

8-02.03 Structure of the snowpack. - As each new layer of snow is deposited, its upper surface is subjected to weathering effects of radiation, rain and wind, the under-surface to ground heat, and the interior to the action of the percolating water and water vapor. As a result, the snowpack is stratified, showing distinct layers of individual snow-storm deposits. Early in the season a granular layer is formed at the ground surface when the ground is unfrozen.

8-02.04 The change in the form of the snow crystals is believed to result from sublimation (evaporation from and condensation onto crystal surfaces) and from the action of percolating water. 1/ 4/ Due to temperature differences, air is in continuous motion within the snowpack, carrying with it heat and water vapor. This activity results in rounding off of snow crystals and growth of some at the expense of others. The circulating air (saturated with water vapor) tends to equalize the temperature and vapor pressure within the snowpack. Impermeable ice planes deflect but do not prevent the movement of the air, just as they deflect or impede the downward percolation of water. The areal extent of such impermeable planes is not great. Observations of percolation paths made at the three snow laboratories, using fuchsine dye to trace the water, indicate many weaknesses in the seemingly impermeable ice planes through which air and water can pass. As a result of the flow of air and water in the snowpack, the pack tends to become homogeneous with respect to temperature, liquid-water content, grain size, and density, as the season progresses.

8-02.05 Nocturnal snow crust. - During the melt season, on clear nights, a relatively shallow surface layer of the snowpack generally cools considerably below 0°C due to outgoing longwave radiation; the liquid water will freeze, but below this surface layer, the snow remains at 0°C and liquid water continues to drain, until the remaining liquid water in the pack equals the liquid-water-holding capacity. The combined effect of air and heat diffusion causes the surface layer to cool each night to a depth of about 10 inches. There is a change in the crystalline structure of the surface layer due to this alternately freezing and thawing effect.

8-02.06 Air permeability of the snowpack. - It was pointed out in the preceding paragraphs that the circulation of air within the pack is partially responsible for the snow metamorphism. Measurements reported by Bader and others 1/ show that air permeability varies widely within the snowpack and with time, just as the crystalline structure and porosity do. An important function of this moving air is the transportation of water vapor from high to low-temperature areas. As a result, ice lenses and ice planes will increase in size, and some crystals will grow at the expense of others.

8-02.07 Flow of moisture. - The relative magnitudes of moisture moving in snow in the form of water vapor and that moving in liquid form by capillary action have not been determined. Experiments conducted by Hadley and Eisenstadt 7/ on thermally-actuated moisture migration in granular media appear to indicate that the movement of moisture by vapor diffusion in the direction of heat flow in snow may be insignificant compared to the capillary movement of liquid water. Section 8-05 deals with the measurement, storage and movement of liquid water in snow, which is of particular hydrologic significance in evaluating the effect of the snowpack on runoff.

8-02.08 Density of new-fallen snow. - The density of new-fallen snow varies widely with the size, shape, and type of snow crystals, as well as with the temperature, humidity, and wind speed of the air through which they fall. The two primary factors are air temperature and wind. Diamond and Lowry 5/ correlated the density of new snow as measured at CSSL with surface air temperature and found an average increase in density of 0.0036 g/cc per degree F increase in surface air temperature at the time of deposition. Figure 4, plate 8-1, shows a plot of their data relating density of new snow with surface air temperatures. Rikhter12/ reported densities of new-fallen snow varying with surface wind, and ranging from 0.06 for calm conditions to 0.34 for snow deposited during gale winds. Changes in density of new snow are rapid and variable during the first few hours after deposition. Therefore, daily measurements of density reflect considerable variability due to the varying time of deposition and settlement conditions during the 24-hour period. For practical use, an assumed average density of 0.10 g/cc for converting depth of new snow to water equivalent, from once-daily measurements, will suffice in most cases for an average measure of precipitation.

8-02.09 Snowpack density characteristics. - The average density of a basin snowpack varies widely with space and time. Generally speaking, the average pack density increases with time, declines slightly with each new snow, but regains its former density soon thereafter with the settlement of the new snow. This is illustrated in figure 1, plate 8-1, by the daily density graph for an accumulating snowpack at WBSL headquarters. The depth of the snow increased almost continuously from 1 inch on 4 December 1949, to 165 inches on 15 January 1950, during which time the mean daily air temperatures were below freezing. On 12 January a deep pit was dug and densities at various levels were observed. As shown in figure 2, the densities varied from 0.07 g/cc for the new-fallen snow at the surface to 0.38 g/cc near the bottom of the pack. Wind, melt, and rain augmented the rate of settlement of new snow. The striated structure of the snowpack was caused by the succession of new snow layers on the more dense snow of greater age upon which they were laid. In general, the density of the pack varies directly with depth, but there is considerable variation as a result of individual ice lenses, ice planes, and buried snow crusts.

8-02.10 Snowpack density changes. - Rapid settling and compacting of the snowpack begins immediately following deposition, and then continues more slowly throughout the accumulation period into the ablation period. The change in form and displacement of individual particles within the snow matrix cause the settlement. The volume of voids gradually decreases from a maximum value when snow is newly deposited to a minimum amount during the melt season, and approaches zero whenever ice is formed. There are several physical processes contributing to settlement through change in crystal form and displacement, as follows: (1) percolation of melt or rain water which freezes with the pack; (2) plastic deformation of the snow matrix, from weight of overlying snow causing reduction in voids; and (3) transport of water vapor due to temperature and vapor-pressure gradients and convection of air within the snowpack. The relative importance of the processes varies with the temperature and precipitation conditions, as well as the original snow condition, and, to a considerable extent, to the length of time that each process has been operating, jointly or singly.

8-02.11 Continuous observations of snowpack conditions, CSSL, 1952-1953. - During the 1952-53 water year at CSSL, SIPRE performed observations of snowpack conditions, as outlined in paragraph 2-07.04. These observations provided a continuous record of depth and time distribution of the crystalline structure, temperature, density, and the horizons of the principal layers of the snowpack for the entire season of snow accumulation and ablation. Isopleths of temperature and density and the positions of the settling-meter markers (see paragraph 2-07.05) are plotted in figure 4, plate 8-2. Daily values of incident and reflected shortwave radiation, maximum and minimum temperature, precipitation in forms of both rain and snow, and mean daily outflow of Castle Creek, are plotted for the same time period. Selected vertical profiles of crystalline structure, temperature and density are shown for the same period on figure 1, plate 8-3. Inspection of these diagrams shows the nature of change of the snowpack condition with respect to time, and the accompanying meteorological conditions producing the change. Special attention is drawn to the rain-on-snow condition which occurred on 8-9 January 1953 and was reported on in Research Note 18. Approximately 4.8 inches of rain fell within a 40-hour period, with air temperatures averaging nearly 40°F. While the temperature condition of the snowpack changed abruptly with the onset of rain and melt, the change in density was very slight. The settlement of the individual snow layers, as marked by the settling-meter measurements, proceeded uniformly at its slow rate through the storm period, unaffected by the several inches of liquid water passing through the snowpack.

8-02.12 Application to snow hydrology. - While the study of snow metamorphism is complex and there is much to be understood about the processes of change, yet in hydrologic application, problems concerning the physical nature of the snowpack can be resolved by evaluation of its "cold content" and the retentivity and permeability of liquid water within the snow-ice matrix. The temperature of the pack may be determined by direct measurement, obviating the evaluation of heat flow into or away from the snowpack prior to the runoff period. The amount of liquid water existing in the snowpack at a specific time may also be determined by actual measurement, but where such measurements are lacking, estimates must be made on the basis of limiting conditions and prior history of the snowpack.

8-02.13 Summary. - Only the highlights of the complex phenomenon of snow metamorphism, pertinent to the storage effect of the snowpack on the amount and time distribution of the runoff, are mentioned. Reference is made to de Quervain, 4/ Hughes, 9/ and Bader 1/ for more complete information on the metamorphism of snow. It was stated that metamorphic processes begin at the point of snow origin and continue until the snow has disappeared. Even during their fall, the individual snow crystals grow together and break apart before hitting the ground. Compaction and settling begin immediately at a very rapid rate and follow a decay type of function with respect to time as shown in figure 4, plate 8-2. Besides compaction, the snowpack undergoes crystalline transformation by the processes of sublimation, melting and refreezing. To determine the storage potential of a natural snowpack and the transmission rate of water, one must appraise the stage of metamorphism of snow. At present the aggregate effect of metamorphism on runoff can best be defined by the depth, density, temperature and liquid water content of the natural snowpack at a particular date and on the march of the previous weather to which the snow was subjected. Factors neither measured nor experimentally determined must, of necessity, be assumed in order to effect a reasonable solution of water storage and transmission problems in snow hydrology.

8-02.14 As a result of the combined action of these factors, the snowpack becomes more compact, striated, and granular as the season progresses. Finally it reaches a "ripe old age" in spring, when only the upper surface undergoes appreciable changes, as manifested in the formation of the nocturnal snow crust and daytime thaw. Ripeness of snow should not be defined solely by its density. In snow hydrology, ripeness is associated with the readiness of the snowpack to transmit and discharge liquid water entering at its surface, regardless of season or density. For hydrologic purposes, therefore, snow is considered to be ripe when it contains all the water it can hold against gravity, i.e., when it is primed.

8-03. HEAT TRANSFER WITHIN THE SNOWPACK

8-03.01 General. - One of the processes involved in conditioning the snowpack to produce runoff is transmission of heat within the snowpack. During the spring melt season, the pack is normally at a temperature of 0°C, except for the nocturnal crust layer, and whatever heat is applied at the surface is converted to melt. During the winter, on the other hand, the pack is often sub-freezing, and heat must be transferred downward from the surface and upward from the ground to meet the thermal deficiency before appreciable runoff may occur. The transfer of heat may be accomplished by conduction, convection and diffusion of air and water vapor within the snowpack, or percolation and refreezing of liquid water from surface melt or rainfall. In hydrologic application the total heat deficit of the snowpack may be treated as an initial loss that must be satisfied before runoff occurs. Accordingly the processes of heat transfer are incidental to the evaluation of the total heat deficit. A general background in the theory of heat exchange within the snowpack is useful, however, in evaluating observed conditions of snowpack temperature and assessing expected changes.

8-03.02 Thermal properties of snow. - The transmission of heat by snow depends on its thermal properties, which are:

(1) the latent heat of fusion, which is the heat energy per gram of snow required for change from solid to liquid state without change in temperature. The latent heat of fusion for snow may be equal to or less than that for ice, depending upon the amount of liquid water in the snow.

(2) thermal quality, which, as defined in paragraph 5-01.05, is the ratio of the heat necessary to produce a given amount of water from snow to the amount of heat required to produce the same quantity of melt from pure ice at 32°F.

(3) specific heat (c_p), which, in c.g.s. units, is the heat in calories required to raise the temperature of one gram of snow one degree centigrade.

(4) heat conductivity (k_c) or heat permeability, which is a measure of the time rate of heat transfer. It is expressed as calories transmitted through 1 cc of snow in 1 sec when the temperature difference between two opposite faces is 1° C.

(5) diffusivity (k_d), which is related to the specific heat and thermal conductivity as follows:

$$k_d = \frac{k_c}{\rho c_p}$$

where ρ is the density of the snow. Note that ρc_p is the heat capacity or the specific heat by volume (cal/cc/deg C). Diffusivity may also be called the temperature conductivity of the snow, because, it is the temperature change in degrees centigrade that occurs in one second, when the temperature gradient is 1° C/cm for each cm depth.

8-03.03 Experimental work. - Differences in the stage of metamorphism of the layers of a natural snowpack make the determination of its thermal properties exceedingly difficult. Specifically, the factors affecting the thermal conductivity and diffusivity of snow are: (1) the structural and crystalline character of the snowpack, (2) the degree of compaction, (3) the extent of ice planes, (4) the degree of wetness, and (5) the temperature of the snow. Experimental work shows that density is a satisfactory index of the thermal properties of the snow shown in the following table:

Density, ρ	Specific heat, c_p		Conductivity, k_c			Diffusivity, k_d
	By weight	By volume	Acc. to Kondrat'eva	Acc. to Abel's	Acc. to Jansson	
g/cc	cal/g/°C	cal/cc/°C	cal/cm²/°C/cm/sec			°C/cm²/sec
1.000(Water)	1.0	1.0000	0.00130+			0.00130
0.900(Ice)	0.5	0.4500	0.00535+			0.0119
0.540	0.5	0.2700	0.00246*		0.00162+	0.00911
0.500*	0.5	0.2500	0.00205*	0.00170*	0.00095*	0.00820*
0.440*	0.5	0.220	0.00167*	0.00132*	0.00089*	0.00760*
0.365*	0.5	0.1825	0.00110*	0.00091*	0.00075*	0.00603*
0.351*	0.5	0.1755	0.00087*	0.00084*	0.00072*	0.00494*
0.340*	0.5	0.1700	0.00075*	0.00079*	0.00070*	0.00441*
0.330*	0.5	0.1650	0.00070*	0.00074*	0.00068*	0.00422*
0.250	0.5	0.1250	0.00042*		0.00053+	0.00336
0.130	0.5	0.0650	0.00011*		0.00029+	0.00169
0.050	0.5	0.0250	0.00002*		0.00010+	0.00080
0.001(Air)+	0.24				0.00006+	

+ From Beskow 2/ pp. 108, 118.
* From Kondrat'eva 10/ pp. 10, 12.

From observations at CSSL, as reported in Technical Report 3, the relation between heat conductivity and density of snow was computed by linear regression to be

$$k_c \,(10^4) = 22.7\rho - 0.46$$

Kondrat'eva 10/ suggests the use of

$$k_d = 0.0133\,\rho \quad \text{and } k_c = 0.0068\rho^2, \text{ for } \rho < 0.35 \text{ g/cc; and}$$

$$k_d = 0.0165\,\rho \quad \text{and } k_c = 0.0085\rho^2, \text{ for } \rho > 0.35$$

For $0.14 < \rho < 0.34$, Abel's gives $k_c = 0.0068\rho^2$.

For $0.08 < \rho < 0.50$, Jansson gives $k_c = 0.00005 + 0.0019\rho + 0.006\rho^4$.

For $0.10 < \rho < 0.60$, Devaux gives $k_c = 0.00007 + 0.007\rho^2$.

Attention is called to the fact that experimental work on thermal properties of the snow has been generally conducted with homogeneous dry snow, often after subjecting it to artificial compaction to attain density variation. In contrast, the natural snowpack consists of several snow layers of varying thicknesses and of different character (resulting from the seasonal snow storms), separated by ice planes.

8-03.04 Volume of air space. - The air space and the absolute porosity of the snow may be computed by considering a unit volume of snow, in which

$$P = 1 - \frac{\rho}{0.92} = 1 - 1.09\,\rho \qquad (8\text{-}1)$$

Here P is the portion of voids or air space in the unit volume considered and ρ is the density of snow. The large percentage of air (with very low heat conductivity, 0.000057) in the snowpack makes the snow a good insulating material. Even for extremely cold weather, the heat transmitted through the snowpack is small. The density of the pack reaches its maximum value in late melt season and seldom exceeds 0.55 g/cc. During the accumulation period, the density may be as low as 0.05 g/cc for a cold new snow and as high as 0.35 g/cc for the pack as a whole. Thus the snow layers may contain 95 to 62 percent air by volume during the accumulation period, or as little as 40 percent during the melt season.

8-03.05 Theory of heat flow. - In a natural snowpack the heat-transfer phenomenon is complicated by the simultaneous occurrence of many heat-exchange processes. As a result of temperature differences, air transports heat and water vapor by convection within the snowpack. Upon reaching a cold surface, some water vapor condenses and yields its heat of vaporization (approximately 600 cal/g). The transport of warm air is greatest when the temperature decreases upward. If, on the other hand, temperature decreases with depth, convection of air within the pack is suppressed. Due to the low heat conductivity of snow, the amplitude of the temperature wave diminishes rapidly with depth below the snow surface. Rain and melt water freezes within the cold (sub-freezing) layers and warms the pack by heat of fusion (80 cal/g). These two processes tend to change the conductivity and diffusivity of the snow throughout the pack and influence the heat transfer rates. The surface layer of the pack is subjected to heating and cooling effects of shortwave and longwave radiation, convection, and condensation, in amounts shown in chapter 5; ground heat flows upward, causing a reduction in the cold content of the snowpack or melting it at the bottom. Ground melt is also discussed in chapter 5. Furthermore, the absorption and transmission of heat by snow vary with the topography of the drainage basin and with the character of the individual layers of the snowpack, just as the structure, the liquid-water content, the porosity and the temperature of the snow vary. Thus, the ever-changing physical and thermal properties of the basin snowpack together with the variation in weather make the theory of heat flow in snow much more complicated than that for homogeneous solids. Variations in the composition and density between layers are of such magnitude that only average values of thermal properties can be used in the application of the fundamental heat flow principles. That is, a homogeneous snowpack (isotropic solid) of small depth is assumed, the temperature-time curve for any level or in any direction in the pack is considered to be straight or sinusoidal, and at any instant the temperature curve has the shape of a damped wave. None of these assumptions is strictly valid. The snowpack is a crystalline and anisotropic solid, in which certain directions are more favorable for conduction of heat than others. The temperature wave is complex; the character of snow varies from layer to layer, and the liquid water content varies with time and temperature. Therefore, the theoretical heat flow computations yield results which are little more than of qualitative significance. Even though the magnitude of heat flow in snow is probably smaller than the errors of some observations and assumptions used in hydrologic studies, a knowledge of the fundamental principles of heat exchange processes within the snowpack is of great value to the hydrologist. Recognition of the order of magnitudes and time lag of heat flow will enable him to make proper allowances. Some special hydrologic problems may require the application of the fundamental principles of heat flow, approximate as they may be. A brief review of these principles is presented in the following paragraphs.

8-03.06 A temperature gradient is established within the snowpack as the result of heat transfer at the surface layers. The temperature gradient is defined as the change in temperature per unit change in depth. Thus, the slopes of the temperature-depth curves, shown in figure 1, plate 8-3, represent the temperature gradients. A straight temperature-depth relationship indicates that the inflow and outflow of heat for any layer are equal. A curved relationship indicates a change in gradient or unequal inflow and outflow, with consequent changes in the temperature of the layers.

8-03.07 Temperature gradients in the snowpack are more pronounced in winter than in spring. When the snowpack reaches an isothermal condition at $0^{\circ}C$, molecular conduction of heat ceases, and heat energy is spent in melting the snow. But the cooling effect of the nocturnal radiation (particularly for open sites) still remains an effective factor in setting temperature gradients within the top 2 to 15 inches of the snowpack. It is apparent that the solution of heat transfer problems requires knowledge of (1) density of the snow, (2) temperature gradients in the snowpack, and (3) thermal properties, including specific heat, conductivity, and diffusivity for the given snow condition.

8-03.08 Inasmuch as the hydrologist is only indirectly concerned with heat transfer within the snowpack, the relative importance of the subject in applied snow hydrology does not warrant inclusion herein of detailed derivations of heat transfer equations. Reference is made to Wilson 15/ and Kondrat'eva 10/ for description of conductive heat flow equations for the snowpack.

8-03.09 Applying Fourier's expression 3/ to a snowpack of $0^{\circ}C$ whose surface is suddenly cooled to and is maintained at $-10^{\circ}C$, Wilson demonstrated how slow the diffusion of heat is through the snow and how the temperature varies with depth and time after the sudden cooling. The following tabulation lists the change of temperature that would occur at three levels of an initially isothermal snowpack whose density is 0.20 and k_c is 0.0003, as computed by Wilson for the sudden change from 0° to $-10^{\circ}C$ at the surface:

Time in hours	Temperature in °C		
	10 cm depth	25 cm depth	50 cm depth
1	-0.7	0.0	0.0
4	-2.2	-0.3	0.0
8	-3.9	-0.7	0.0
24	-6.4	-2.2	-0.4
48	-7.8	-3.8	-1.1

8-03.10 Temperature distribution in nocturnal snow crusts. - It was pointed out that nocturnal snow crusts occur on clear nights during the melt season. The depth of penetration of the sub-freezing temperatures has been computed theoretically by methods presented by Beskow 2/, to be about 13 inches. The depth should be considerably smaller than this, chiefly because of the latent heat supplied by refreezing of melt water and to a small extent the convection and condensation heat from air flowing onto the cold surface layer. In view of these facts one may assume (in this case) that the magnitude of the maximum depth of penetration of the cold wave or of the frost line is approximately 10 inches. Observations of snow temperatures in the crust layer were made in connection with the operation of the lysimeter at CSSL during May, 1954. A plot of snow temperatures against depth and time are shown in figures 2 and 3 of plate 8-3, showing the progress of cooling through a typical night with clear skies, and the subsequent warming with the onset of energy input from solar radiation.

8-04. THE SNOWPACK TEMPERATURE

8-04.01 General. - The external factors affecting the flow of heat to the snowpack have been described in chapter 5, and the processes of heat transfer within the pack were discussed in the preceding section. The resulting variations in snowpack temperature affect the water-storage potential of the snowpack and with it the runoff from snowmelt or rain. Much of the winter and early spring surface melt is stored in the snowpack, contributing little or nothing to runoff. The amount of liquid water lost to runoff because of the cold content is a function of the snow temperature below 0°C. It accounts for the gradual increase in ablation of snowpack per unit of heat absorbed or in degree-day factor. Obviously, snow temperature must be considered in hydrologic problems involving early season runoff. Unfortunately, it is not yet a regularly observed element, and no simple relationship is available for estimating it from independent data. Generally, snowpack conditions are observed by digging a pit, but an approximate temperature, moisture, and structural profile of the pack can be obtained by use of the Mt. Rose sampling tube, Weston metallic thermometers and visual observations of the core, as illustrated in plate 8-10.

8-04.02 Laboratory observations. - Continuous observations of snowpack temperatures made at CSSL and UCSL provide a basis for estimating the range in temperature to be expected and the variations that occur through the season. Data for the 1948-49 water year for these two areas were selected to show the variation of the temperature profile of the snowpack with time. The isotherms of snowpack temperature are plotted on plate 8-4, as a time function, for both CSSL and UCSL for the 1948-49 water year. Daily values of maximum, minimum and mean air temperatures are also plotted on the same time scale. The snow and ground temperatures for specified levels above and below the ground surface are shown on plate 8-5 for the same period. The previously referenced snow structure data at CSSL for the 1952-53 water year, shown on plates 8-2 and 8-3, indicate the snowpack temperature variation and gradients for that year. Inspection of these charts reveals the nature of the temperature gradients of the snowpack and how they change with time and weather.

8-04.03 The cold content of the snowpack. - The hydrologist is primarily interested in the temperature and density profiles of the snowpack and how they affect the cold content of the snow, for the evaluation of runoff potential in the winter and early spring months. The cold content is defined as the heat required per unit area to raise the temperature of the snowpack to 0°C. It is convenient to express it in inches of liquid water (produced at the surface by either rain or melt) which, upon refreezing within the pack, will warm the pack to 0°C. The relationship may be expressed as

$$W_c = \frac{\rho D T_s}{160} \qquad (8\text{-}2)$$

where W_c is the cold content in equivalent inches of liquid water, ρ is density in g/cc, D is the depth in inches, and T_s is the average snowpack (or snow layer) temperature deficit below 0°C.

8-04.04 As a means of estimating the cold content of the upper 24 inches of the snowpack from current temperature data, the empirical relationship shown on figure 1, plate 8-6 was derived on the basis of deep pit observations at CSSL. This diagram relates the cold content of the upper two feet of the snowpack to the average temperature of the preceding 3 days. It is assumed that the temperature of the snowpack below the upper two feet changes slowly, and the cold content for the lower layers can be estimated from previously obtained snow temperature data.

8-04.05 The cold content of the crust layer represents the only deficiency of heat in the snowpack during the active melt season. The penetration of cold from nocturnal cooling has been discussed in paragraph 8-02.05. A cold content of about 0.1 inch is an average value for nighttime crust formed under clear skies in the open. The liquid-water deficiency (to be discussed later) developed in the surface snow layer by virtue of the refreezing of its free water, represents an additional amount of about 0.15 inch of melt, so that the daily average water equivalent of total heat deficit in the crust layer to produce runoff is about 0.25 inch of melt. This deficit is approximately 15 percent of the average daytime energy input for clear weather spring-time melt in the open, and must be supplied each morning before there can be an appreciable contribution to runoff. Knowledge of the order of magnitude of the snow-crust deficiency is of value when results of a snowmelt study must be interpreted with respect to errors of observations, assumptions, and omission. Reference is again made to figures 2 and 3 of plate 8-3, which show the progressive cooling of the crust layer during a typical clear night at CSSL, during the active melt period. In plate 8-9, the amount of daytime energy expanded to balance this deficit is illustrated.

8-05. LIQUID WATER IN SNOW

8-05.01 Movement of water in snow. - Water moves within the snowpack in both the vapor and liquid phase. While the movement of water vapor is important to metamorphism of the snowpack (see paragraph 8-02.04), the order of magnitude is low in comparison with liquid water transport. Liquid water moves by gravitational and capillary forces in all directions. After the snow reaches its liquid-water-holding capacity, the downward movement is dependent entirely upon gravitational force.

8-05.02 Conditions of liquid water in the snowpack. The snow is said to be dry when its temperature is below 0°C. At 0°C, the degree of wetness depends on the availability of liquid water and the liquid-water-holding capacity of the snowpack. Winter rains or melt may bring the snowpack to its liquid-water-holding capacity commensurate with the stage of metamorphism of the snow. Subsequent weather will change the character of the snow and with it the liquid-water-holding capacity. Generally, however, the snow is cold and dry in winter. The forms in which liquid water exists in the snowpack are:

(1) hygroscopic water, which is adsorbed as a thin film on the surfaces of the snow crystals, and unavailable to runoff until the snow crystal has melted or changed its form.

(2) capillary water, which is held by surface tension in the capillary spaces around the snow particles. Capillary water is free to move under the influences of capillary forces, but it is not available to runoff until the snow melts or the spacing between crystals changes.

(3) gravitational water, which is in transit through the snowpack under the influence of gravity. It drains from the pack and is available for runoff.

8-05.03 The hydrologist is concerned with both liquid water content, $\underline{f_p}$, (as determined by actual measurement of liquid water in the snowpack) and the liquid-water-holding capacity, $\underline{f_p''}$, which is defined as the maximum amount it can hold against gravity at a given stage of metamorphism and density. The difference between the two quantities represents the amount of liquid-water storage capacity (in excess of the cold content of the pack) and is termed the liquid-water deficiency, $\underline{f_p'}$. Liquid-water content in excess of the liquid-water-holding capacity represents a condition where liquid water excesses are flowing through the pack. The amount of liquid water is expressed in percent by weight. The total liquid-water-holding capacity of the snowpack can be integrated over a basin area, through use of snowpack data representative of various elevation zones or areas.

8-05.04 Determination of the liquid water in the snowpack. - The temperature of the snowpack limits the requirements for determination of liquid-water content. If the temperature of the snow is below 0°C, the snow is dry, and no measurements for liquid-water content are necessary. The liquid-water-holding capacity of the snow of certain character is determined from measurements of its liquid-water content after drainage of the excess gravitational water. The most commonly used method for measuring the liquid water is the calorimetric method, by use of the thermos bottles as calorimeters, as outlined in Technical Report 1. Other methods include measurements of electrical capacitance in a parallel plate condenser having snow as its dielectric 6/, by differences in snow compaction 14/, or by use of a centrifuge.

8-05.05 While calorimetry is commonly used to measure liquid-water content, it is an indirect method. The state or degree of dampness is described by the thermal quality of the snow, which is defined in paragraph 8-03.02. Both liquid water and thermal quality are usually measured in percent by weight. The complement of the thermal quality (for thermal qualities less than 100 pct) is the percentage of water in the snow matrix. Thermal qualities in excess of 100 percent indicate no liquid water in the snow and temperatures below 0°C. The percentage above 100 percent is proportional to the cold content of the snowpack.

8-05.06 Observations of thermal quality at the snow laboratories. - The thermal quality of the snow at the snow laboratories was determined by the calorimetric method. Thermal qualities ranged from 80-110 percent. Generally low thermal quality values were obtained during times of high melt when samples of snow contained melt water in transit or in excess of the liquid-water-holding capacity of the snow. Measurements generally were taken randomly, with sampling inadequate to represent completely time, areal, and depth variations; however, in May 1948, observations were made at CSSL at four-hour intervals, over a period of two days when active melt was in progress, for six layers in the snowpack. The results of these observations are shown in figure 4, plate 8-7, along with hydrometeorological elements preceding and during the measurement period. While there is considerable variability due to errors in measurement, the diurnal fluctuation in thermal quality (and hence liquid-water content) in the various snow layers is well defined. The drainage of liquid water through the pack is shown by the time displacement of the maximum and minimum values in the lower layers. Also, it should be noted that after nighttime drainage, liquid water for the pack as a whole averaged about 3 percent, but that maximum values of 10 percent occurred during the day from the effect of water in transit. The diurnal fluctuation in density during these observations is also shown on this diagram and appears to be closely related to changes in liquid water.

8-05.07 Qualitative field tests. - The calorimetric method is not well suited for field determination of the liquid-water content of the snow. A qualitative evaluation of liquid water can be had by the "wetness test",13/ wherein the observer notes the character of the snow, cools his gloves to snow temperature, squeezes a sample and records the appearance and the degree of compaction when pressure is released. His notes will show the snow "dry" when a snowball cannot be made; "moist" when liquid water is not obvious but a snowball can be made; "wet" when liquid water is visible; and "slushy" when water drains out of the snow with slight pressure. Noting the elevation and exposure of the site, the temperature, density, date, and hour of the day may be of great value in evaluating the seasonal and areal variations in the wetness of the snowpack. Exposure is an important factor in the priming of the snowpack by surface melt. Snow surveys at CSSL show that the snowpack on southerly exposed areas reaches its liquid-water-holding capacity about 15 days earlier (and for the same date about 1500 feet higher in elevation) than northerly exposed snow.

8-05.08 Variability of liquid-water-holding capacity.- Most of the available thermal quality data for the snow laboratories was obtained during the active melt period at one site, when the snowpack density was relatively high and often included water in transit through the snowpack. A considerable number of the thermal quality observations at the snow laboratories represent results of trials for acquiring speed and consistency, and cannot be considered adequate for analysis. Furthermore, pertinent and associated information on the density and character of the snow were not always recorded. Consequently, the thermal-quality data from the laboratories are inadequate for precise analysis of the liquid-water content of the snow. The basinwide variability cannot be evaluated except in a general way, and only approximate percentages can be recommended for use in hydrology. A snowpack at $0^{\circ}C$ has a liquid-water-holding capacity of approximately 2 to 5 percent by weight, depending on the density and depth; the mass of ice layers; the size, shape and spacing of snow crystals; and the degree of channelization and honeycombing. It is difficult, if not impossible, to evaluate the individual influence of each of these factors on the liquid-water-holding capacity of a basin snowpack. Therefore, the liquid-water-holding capacity of snow may be related to density. As shown in figure 6, plate 8-7, the affinity of snow for liquid water increases with increasing snowpack density. Unfortunately, the number of thermal-quality determinations for snow of less than 40-percent density is insufficient to indicate if the decrease in liquid-water-holding capacity with decreasing snow density continues for very low snow densities. Additional measurements are required to determine the liquid-water-holding capacities in this range.

8-05.09 An approximation of the liquid-water-holding capacity of snow can be obtained from the heat-balance equation for the surface layer during clear nights in spring. This layer begins to cool at the snow surface just before sundown and continues for approximately 12 hours. The frost line reaches a maximum depth at about 0600 hours. The difference between total heat loss and the maximum cold content represents the heat gain from freezing of liquid water (not in transit) in the snowpack. The computed liquid water content for the snow crust as shown in the previously referenced observations at CSSL (plate 8-3) would be approximately 4 percent, which agrees with thermal quality determinations for liquid water in a pack after drainage.

8-05.10 Recommended liquid-water-holding capacities. - Experiments on liquid-water-holding capacity of snow are limited. Nearly all are for spring snow of densities above 35 percent, while densities of winter snowpacks usually range from 10 to 35 percent. In this range, no observations of liquid-water-holding capacities are available. From the results of observations of thermal quality shown on plate 8-7, and from Gerdel's study of transmission of water through the snow, 6/ between 2 to 5 percent by weight is recommended for the liquid-water-holding capacity of snow. Additional observations are required to establish the relationship between snow density and liquid-water-holding capacity.

8-05.11 The lack of information on the capacity of the snow to retain liquid water against gravity, as a function of some index of the stage of metamorphism, constitutes a major gap in knowledge of the storage effect of the snow on runoff. Streamflow forecasts require estimates of the basin snow temperatures and probable liquid-water content of snow at various elevations. These estimates cannot be made with confidence unless systematic observations of these snowpack conditions are made or computed from empirical relations based on adequate experimental data. The above range of values is presented as a guide for use where observations are lacking.

8-05.12 It is pointed out that the liquid-water-holding capacity of snow, as discussed in the preceding paragraphs, represents conditions where free drainage of the snowpack is assured. In flat areas, horizontal drainage through channels is impeded by the lack of sufficient slope. Thus, portions of the snowpack in foothills and flat lands may hold liquid water far in excess of that for mountainous areas where free drainage is rapid.

8-06. TRANSMISSION AND TRAVEL TIME OF WATER THROUGH THE SNOWPACK

8-06.01 General. - The condition of the snowpack determines the amount of storage and the rate of downward movement of water. The temperature, size, shape, surface area, and spacing of the snow crystals, channelization stage, and melt and rainfall intensities control retention and detention of water as it moves downward through the snowpack. Since many of these factors are continuously changing, neither the storage nor the rate of movement can remain constant. The time of travel of water in unprimed snow may be considerable, particularly when the snow is striated with ice planes which are flat or concave upward. By the time water reaches the ground surface, a water course is established in the snowpack. After this, the travel time is relatively short, being primarily a function of the snow depth. The time of travel through the established water courses in the snowpack continues to vary, because, under the action of the percolating water, the crystalline structure of the snowpack continues to change, and erosion and more extensive channelization progresses, with the consequent release of some liquid water held against gravity.

8-06.02 Experimental work. - It is impossible to estimate quantitatively the variation in travel time of water through a natural snowpack except in a general way, using laboratory experiments as guide. Electric snow-moisture meters, such as Gerdel used in CSSL, are probably the best method at present for qualitatively determining the travel time of water through the snowpack. Figure 5, plate 8-7, illustrates the results of his tests and show the rate of drainage of water through three layers of the snowpack. The following table summarizes Gerdel's results on the transmission of water through snow. 6/

Snow density	Probe spacing	Water applied	Trans-mission rate*	Liquid water content**			Duration of observation
				Initial	Peak	End	
g/cc	in	in	in/min	pct	pct	pct	min
0.35	25	2.0	1.1	4.0	16.2	5.5	19
0.35	41	2.0	1.1	1.4	9.3	1.1	19
0.	7	0.8	0.9	3.8	6.2	1.1	18
0.40	17	0.5	3.7	4.1	6.2	1.7	18
0.46	6	2.0	12.0	6.0	10.7	5.1	48
0.46	12	2.0	16.0	4.0	9.8	0.9	48
0.46	18	2.0	18.0	4.0	10.3	0.9	48
0.46	6	2.0	24.0	5.0	10.3	2.6	35
0.46	12	2.0	24.0	2.9	10.3	0.7	35
0.46	6	2.0	24.0	3.0	10.3	2.0	35

* Computed from time interval between peak flow and spacing of capicitor probes.
** Interpolated from calibration curve derived from capacitance readings and calorimetric measurements on three or more snow samples, collected during each experiment.

In general, it is seen from Gerdel's work that the transmission rate increased with increasing densities. This may reflect the effect of changing structure of the snowpack as the season progresses, rather than a direct relationship to density. Horton 8/ theorized transmission rates in snow on the basis of void spaces, as is done for other porous media, and according to his theory, transmission rates through snow would increase markedly for decreasing densities. This is in direct conflict with results obtained from Gerdel's experiments.

8-06.03 <u>The depth of penetration of water</u>. - In a snowpack of uniform texture, the depth of penetration of water varies directly with the amount of water entering the snowpack and inversely with the storage deficiency. The latter is a function of the snow temperature and the liquid-water-holding capacity. For instance, for a cold snow,

$$D = \frac{t\,(i_r + m)}{\rho\left(\frac{T_s}{160} + \frac{f''_p}{100}\right)} \qquad (8\text{-}3)$$

in which $\underline{D}$ is the depth of penetration in inches; $\underline{i_r}$ and $\underline{m}$ are the rain and melt intensities in inches per hour; $\underline{t}$ is the duration in hours; $\underline{\rho}$ is the density of snow in g/cc; $\underline{T_s}$ is the temperature of the snow in $^{\circ}$C below freezing; and $\underline{f_p''}$ is the liquid-water-holding capacity of the snow in percent. For a moist snow, $T_s = 0^{\circ}C$, and if the liquid-water deficiency is $\underline{f_p'}$, then

$$D = \frac{100\ t\ (i_r + m)}{\rho\ f_p'} \qquad (8\text{-}4)$$

For a completely primed snowpack, where $f_p' = 0$, the entire depth will be penetrated regardless of the magnitude of water entering the snowpack.

8-06.04 Method of travel. - Storage of water begins at the snow surface where melt (or rain and melt) enters the snowpack. The priming or conditioning of the snow (to pass water through it) begins with this surface layer, which is generally homogeneous. The conditioning of the snowpack progresses downward in a path of least resistance from one layer to another. Upon meeting an ice plane, the water flows over the surface until a weak point allows part or all of the water to enter and spread in the layer below. In this zigzag manner water finally reaches the ground surface. The phenomenon is illustrated in figure 3, plate 8-1, for a cold, unripe snow. Depending on the slope, curvature, and degree of impermeability of the ice planes which separate different layers of snow deposits, water may reach the ground surface before the entire snowpack is saturated or conditioned to yield runoff. That is, some cells in the snow matrix may not yet have become accessible to the conditioning action of the infiltrating water when runoff appears.

8-06.05 Examples of time of travel. - The lysimeters at CSSL provided an opportunity to study transmission of water vertically through the snowpack. In Research Note 4, rates of outflow from artificial sprinkling of the snowpack at the headquarters lysimeter were determined and storage delay within the snowpack was evaluated by means of distribution graphs. Research Note 18 describes the effect of storage and transmission of water through the snow for the natural rain-on-snow occurrence of 8 January 1953, at the same lysimeter. Clear-weather melt studies for the 1954 season at the Lower Meadow lysimeter also provide factual data on delay to runoff by the snow, through evaluation of the diurnal fluctuation of heat supply to the snow surfaces.

8-06.06 The rain-on-snow event described in Research Note 18 provided an excellent opportunity to study both storage and transmission of liquid water in the snowpack. A rain of about 4.9 inches, augmented with melt of 1.9 inches, entered a cold and dry snowpack in a period of about 2 days. Plate 8-8 shows a mass curve of the water balance of the snowpack for the storm period. The snowpack at the beginning of rain was 84 inches deep, with an average density of approximately .31 g/cc, a water equivalent of 26.7 inches, and an average temperature of $-3.0^{\circ}C$. The liquid water required to warm the pack to $0^{\circ}C$ and supply the liquid-water-holding capacity were computed to be 0.8 in. Rain and melt in excess of this amount was available for runoff. Delay to runoff of the liquid-water excess is illustrated by inflow and outflow hydrographs shown on figure 4, plate 8-9. At the headquarters lysimeter, the time delay through the snowpack was about 2 hours. A plot of outflow for the Lower Meadow lysimeter is superimposed on the same graph. The runoff deficiency at the Lower Meadow lysimeter in the early part of the storm was due to the configuration of the ice planes, causing an indeterminable part of the water to flow outside the lysimeter boundaries until channels of flow through the ice planes were developed.

8-06.07 The lysimeter outflow for daily melt contribution during rain-free periods is shown in figures 1-3, plate 8-9, for clear, partly cloudy, and cloudy days. Here, inflow at the surface is computed by energy-transfer equations as described in chapter 5. Observed outflows show the net effect of time delay to runoff caused by the snowpack. Liquid-water deficiency in the crust layer, due to nighttime heat loss, must first be satisfied before water is available for runoff the next day, and it is indicated by shading on the diagram. The time of peak flow is displaced approximately 3 hours in these cases. Here, the snowpack depth averaged about 48 inches for conditions in figures 1 and 2 and 40 inches for figure 3. Densities averaged 0.50 g/cc.

8-06.08 A study was made of time of diurnal peak discharges in Castle Creek, CSSL, to determine changes in peak lag time resulting from varying snowpack conditions. From 5 years of study, it was determined that early in the melt season, (usually about 1 May), Castle Creek peaks daily at about midnight, but the peak advances to about 1700 hours after a week or two of active melt and occurs at about this hour for the remainder of the season. While the time of peak discharge is a function of both natural basin storage and the storage effect of the snowpack, the change in peaking time is caused primarily by change in snowpack conditions.

In hydrologic studies, after the melt season has progressed for a relatively short period, changes in travel time caused by change in snowpack conditions are not significant, and the use of an average storage time is justified.

8-06.09 The February 1951 rain-on-snow analysis for Mann Creek, WBSL, described in Research Note 24, is discussed in chapter 9. While the study was complicated in the initial part of the storm by precipitation in the form of both rain and snow, by storage of snow and liquid water in the trees, and by uncertainties of melt computations, fairly definite evaluation can still be made of the effect of the snowpack on runoff. In general, the lag time between peak inflow and outflow increased about 2 hours from that which would occur for bare ground conditions. The snow depth averaged about 5 feet. Appraisal of the storage effect of the pack is shown in plates 9-1 and 9-2.

8-06.10 Horizontal drainage. - Where horizontal drainage is inadequate (as in the Great Plains, in contrast to mountainous regions), the delay to runoff caused by the snowpack may be much larger than that required for the vertical transit of water through the pack alone. Unfortunately, adequate information on horizontal flow rates and the stage of metamorphism of the snowpack is not available.

8-07. STORAGE POTENTIAL AND TIME DELAY TO RUNOFF

8-07.01 General. - The preceding sections have considered separately the processes involved in conditioning the snowpack to produce runoff and the methods for evaluating amount of storage of liquid water that results from a given snowpack condition. This section deals with the problem of combining amounts numerically, for the purpose of applying the theory to project basins, and estimating the time and storage required to prime the snowpack for given rain or melt rates and snowpack conditions. In formulating storage and time delay it is assumed that the snowpack is homogeneous; the storage is a direct function of the cold content and liquid-water deficiency; and the total time delay is a function of the rate of inflow. It is therefore assumed that the total storage potential of the snowpack must be satisfied before runoff occurs. Actually, the pack is not homogeneous. Some water will appear at the bottom of the pack before the entire pack is primed. The storage potential in the snowpack decreases gradually as the percolating water disintegrates the ice planes shielding the cold cells.

8-07.02 The storage potential of the snowpack is the sum of its cold content (expressed in inches of liquid water) and its liquid-water-holding capacity. These amounts are small relative to the total energy required to melt the snowpack, and constitute an initial loss before runoff occurs. The energy required to condition the pack, in percent of the energy required to melt it, is

$$E = \frac{T_s}{1.6} + f_p'' \qquad (8\text{-}4)$$

where $\underline{E}$ is equivalent energy to condition the snowpack in percent of melt energy, $\underline{T_s}$ is the average snow temperature below zero, in $^{\circ}C$, and $\underline{f_p''}$ is the liquid-water-holding capacity of the snowpack in percent. Thus, with an average snowpack temperature of $-5^{\circ}C$, and liquid-water-holding capacity of 3 percent, the total energy required to condition the snowpack is only 6 percent of that required to melt it. This energy is normally supplied during the transition between the accumulation and melt periods, so that its effect on flows during the active spring melt season may be ignored. In the winter, however, the magnitude of storage potential in the snowpack may be an appreciable part of the runoff quantity, depending upon snowpack condition. An example is presented in this section to show how to estimate the storage effect over a basin.

8-07.03 Basic data requirements. - The condition of the snowpack can be determined by direct observation, augmented by estimates based on day-to-day variations in the meteorologic regime. In mountainous areas, elevation differences must be taken into account, and sampling points should adequately represent all elevations within the basin. Time frequency of sampling is dependent upon known conditions of the snowpack. Information required is temperature, depth, density, liquid-water content (when applicable), and structural characteristics such as crystalline types and locations of ice planes. Plate 8-10 shows the type of information as obtained during January 1952 on the west slope of the Sierra Nevada along U. S. Highway 40. Various elevations ranging from 3000 feet msl to 6900 feet msl were sampled by means of digging snowpits, in order to assess the storage potential of the snowpack in the Yuba and American River basins. Also shown on the plate is a method for determining the temperature and structural profile of the snowpack by use of a Mt. Rose snow sampler and Weston bi-metallic thermometers. A means of estimating snow temperatures in the top 2 feet of the snowpack from air temperatures of the preceding 3 days is shown in figure 1, plate 8-6.

8-07.04 <u>Formulas for computing runoff delay</u>. - In order to evaluate the storage of liquid water and time delay to runoff in a given elevation zone, basic storage equations are presented. The cold content may be represented by the equivalent liquid water requirement,

$$W_c = \frac{W_o T_s}{160} \tag{8-5}$$

where W_c is the liquid water in inches to raise the temperature of the pack to $0^{\circ}C$, W_o is the initial water equivalent of the snowpack in inches, and T_s is the average snowpack temperature in $^{\circ}C$ below zero. Neglecting the small amount of warming by rain heat, the time in hours, t_c, required to warm the pack to $0^{\circ}C$ is

$$t_c = \frac{W_o T_s}{160\ (i_r + m)} \tag{8-6}$$

where i_r is the rainfall intensity and m is the melt rate in inches per hour. Additional storage of liquid water to satisfy the liquid-water deficiency may be expressed as

$$S_f = \frac{f'_p}{100}(W_o + W_c) \tag{8-7}$$

where S_f is the water stored and f'_p is the liquid-water deficiency in percent. The time required to store S_f may be represented by t_f as follows:

$$t_f = \frac{S_f}{i_r + m} = \frac{f'_p\ (W_o + W_c)}{100\ (i_r + m)} \tag{8-8}$$

The total storage potential of liquid water not available for runoff is

$$S_p = W_c + S_f \tag{8-9}$$

where S_p is "permanent" storage, in that it is not available for runoff until the snowpack is melted. Transitory storage constitutes an additional delay to runoff. Initially, the transitory storage up to the instant runoff began would be

$$S_t = \frac{D\,(i_r + m)}{v_t} \tag{8-10}$$

where S_t is the transitory storage in inches, D is the snowpack depth in feet and v_t is the transmission rate in ft/hr. The time (in hours) is simply

$$t_t = \frac{D}{v_t} \tag{8-11}$$

so that

$$S_t = \frac{W_o}{\rho\, v_t}(i_r + m) \tag{8-12}$$

and, since $\quad D = W_o/\rho$

$$t_t = \frac{W_o}{\rho\, v_t} \tag{8-13}$$

Total storage of liquid water before appreciable runoff occurs is

$$S = W_c + S_f + S_t \tag{8-14}$$

or, adding (8-5), (8-7), and (8-12), and assuming W_c is negligibly small in comparison with W_o,

$$S = W_o\left[\frac{T_s}{160} + \frac{f'_p}{100} + \frac{(i_r + m)}{\rho\, v_t}\right] \tag{8-15}$$

The time required to produce runoff is

$$t = t_c + t_f + t_t \tag{8-16}$$

or

$$t = W_o\left[\frac{T_s}{160\,(i_r + m)} + \frac{f'_p}{100\,(i_r + m)} + \frac{1}{\rho\, v_t}\right] \tag{8-17}$$

After runoff has begun, the delay caused by transitory storage in the snowpack is negligibly small in comparison with the usual magnitude of natural basin storage times. As the inflow continues, and the drainage channels within the pack become more efficient in transmission of water, there is an indeterminate amount of previously withheld water which is released to runoff. The actual magnitude of this effect is unknown and represents a gap in basic knowledge.

8-07.05 Example of storage potential evaluation. - The data shown on plate 8-10 provide the basis for a numerical example of storage potential of the snowpack over a basin, summarized from an analysis contained in Technical Bulletin 17. Through use of principles set forth in the preceding sections of this chapter, the storage potential of the snowpack is evaluated for each of 5 elevations and is expressed as inches depth of inflow from rain and melt required before runoff appears at the bottom of the snowpack. From assumed rates of rainfall and snowmelt, the time required to condition the snowpack to produce runoff is computed. Table 8-1 lists values of snowpack conditions, liquid-water deficiencies and assumed rates of rainfall and snowmelt, from which components of storage and time delay are computed step by step and combined to show the total effect of the snowpack on runoff. The general equations for computing each value are shown in the table.

8-07.06 For this example, the total storage of water within the snowpack before runoff appeared from the bottom of the snowpack varied from 0.08 inches at 3000 feet msl to 2.86 inches at 6900 feet msl. For an assumed rate of rain plus melt of 0.12 inch per hour, the corresponding total time delay before water was available for runoff ranged from 0.7 hour to 24 hours. For a snowpack about 15 feet deep, after the initial storage requirement of water within the snowpack has been satisfied, additional inflow from rain or melt will pass through the snowpack with a maximum time of transitory storage (t_s) of 4 hours. The average time delay of transitory storage for the basin as a whole would be less, and accordingly, may usually be ignored when considering the magnitude of total basin storage time.

8-07.07 The variation of the storage potential with elevation, for this example, is shown by curve A, figure 2, plate 8-6. Curve B in the same figure illustrates the variation with respect to elevation of the elapsed time from beginning of rain and melt before appearance of flow at the bottom of the snowpack, for the assumed inflow rate of 0.12 inch per hour. Both curves are expressed in terms of unit snowpack water equivalents, from data contained in lines 25 and 26 of table 8-1. These curves apply only to conditions for the illustrative example. Similar curves can be derived for assumed or observed snowpack conditions and for unit rates of inflow. By combining such curves with area-elevation data for a given basin, a general solution for a given snowpack condition can be made.

8-07.08 Total basin storage potential. - In addition to the storage potential of the snowpack itself, storage of water in the soil beneath the snowpack and ponding caused by the presence of a snowpack should be considered in assessing the total basin storage potential. In many cases, the soil has reached its liquid-water-holding capacity in advance of runoff-producing rain or melt, particularly for areas where fall and early-winter rains normally saturate the soil in advance of the major snowpack accumulation. In such cases, the soil remains at or near liquid-water-holding capacity through the winter season. Even for this type of area, however, below normal early-season rainfall may cause a soil moisture deficit which might be carried over through the winter period. The effect of ponding resulting from the presence of a snowpack is important for areas with relatively flat slopes, such as in meadows or plains. Where ponding occurs, the time delay to runoff is increased, and the duration and total amount of infiltration increased as well. The relative magnitude of the effect of ponding is dependent upon the percentage of area on which ponding is likely to occur.

8-08. SUMMARY AND CONCLUSIONS

8-08.01 General. - The effect of varying snowpack conditions on runoff from either rainfall or snowmelt is one of the basic considerations of snow hydrology. Divergent opinions exist as to the storage effect of the snowpack. They range from considering the snowpack to be a vast "sponge" capable of retaining large quantities of liquid water, to the assumption that storage in the snowpack is negligible in any basin study. Actually, there are times when either viewpoint may be correct, and there is no generalization which is universally applicable. The important consideration is that the actual snowpack condition be evaluated in order to properly assess its immediate storage potential. Winter runoff, in particular, is affected by the snowpack condition. In the active spring melt period, on the other hand, within the first week or two of melt, the snowpack becomes fully conditioned to produce runoff so that daily melt or rainfall quantities pass through the pack virtually without loss, except for the minor effect of the nocturnal crust layer. The storage potential within the snowpack, therefore, should be considered in connection with winter or early spring runoff from a rain-on-snow situation. In order to evaluate the storage potential of the snowpack, it is necessary to understand: (1) basic changes in the character of the snowpack through its metamorphism and the processes that cause the changes; (2) heat and water vapor transfer within the snowpack and their relation to meteorologic variables; (3) the cold content of the snowpack; (4) the liquid-water-holding capacity of the snowpack; (5) liquid-water transmission rates; (6) the determination of basinwide snowpack character and evaluation of changes between observations; and (7) methods of analyzing snowpack condition and inflow rates on basin areas for estimating the net effect of the snowpack on runoff.

8-08.02 Snowpack character. - The basin snowpack consists of individual snow crystals, ice planes and communicating air spaces, which may or may not contain liquid water. The volume of air space is at a maximum in a new-fallen snow. This volume decreases very rapidly at first, then gradually reaches a minimum at the end of the season. The rate of settlement and compaction of the pack is primarily a time function of several processes causing changes in form and displacement of crystals within the snowpack. An important function of the air in the snow matrix is the transportation of heat and water vapor from high to low temperature areas. This activity results in rounding off of the snow crystals and growth of some at the expense of others and tends to reduce the depth of penetration of cold into the snowpack. Impermeable ice planes deflect but do not prevent the movement of the air within the snow.

8-08.03 Conditioning of the snowpack. - In general, during the winter period the snowpack loses heat to the atmosphere and gains heat from the ground, resulting in the establishment of a thermal gradient within the pack. (For deep snowpacks, the temperature at the ground surface is usally $0^{o}C$). The snowpack continues to remain cold (sub-freezing) until the melt and rainwater and the diffusion of heat cause the snow temperature to rise to the melting point. Liquid water entering a cold snowpack freezes within the pack, becomes part of it, and increases the temperature within the snowpack. Snow at $0^{o}C$ will impound additional water on crystal surfaces and in air spaces as hygroscopic and capillary water. Such water (held against gravity) also becomes part of the snowpack and is retained until the snow has melted. Pockets or cells of snow which are cold and dry may exist in an otherwise wet snowpack as the result of ice planes which have not yet disintegrated to allow the snow to become fully conditioned. The conditioning of the snowpack from surface water progresses downward in a path of least resistance as one layer after another is completely saturated and yields water to the layer below. Neither the retentivity or permeability of the snow is constant. Therefore, the transmission rates and water storage capacity of the snow vary with the character or the stage of the metamorphism of the individual snow layers. Except for density, no other characteristic of the snowpack element is regularly observed in a project basin. Therefore, density must be used as an index of the general character of the snow as it affects storage.

8-08.04 Evaluation of basin snowpack storage potential. - The storage effect of the snowpack for drainage basins in mountainous regions may be approximated by dividing the basin into homogeneous topographic units or elevation zones. In the lower portions of the basin, the snowpack may be in condition to yield any rain and melt that may enter the pack. At higher levels the snowpack may be at $0^{\circ}C$ but may possess a liquid-water deficit. At still higher levels the snowpack may be cold and dry, a condition for optimum storage of liquid water.

8-08.05 Direct evaluation of storage potential of the snowpack requires observations of depth, density, water equivalent, temperature, moisture content, and character of snowpack for representative areas or zones of elevations within a basin. Such observations are best made by digging snowpits, but cores from snow sample tubes may be utilized. Precise field determinations of liquid-water content of snow are difficult to obtain, but wetness tests may provide qualitative data. Changes in the conditions of snowpack temperature and moisture, subsequent to the time of observation, may be determined on the basis of meteorologic variables.

8-08.06 Evaluation of the snowpack condition for representative zones may be obtained by direct observation or from estimates, and the storage potential and corresponding time delay to runoff may be evaluated on the basis of the example outlined in section 8-07. Combining amounts in each elevation zone, in conjunction with area-elevation relationships, gives the total basin storage effect of the snowpack. The time and frequency with which such evaluations are necessary, varies with the changing conditions of the snowpack. In the active spring melt period, time delay to runoff from storage of liquid water in the snowpack may be ignored.

8-08.07 A snowpack 180 inches deep, with 35 percent density, and an average temperature of $-5^{\circ}C$, would store about 4.0 inches of liquid water before water would become available for runoff. This represents a near-maximum amount of storage for the mountains of western United States. In mountainous areas, slopes are usually sufficient to effect horizontal drainage of the snowpack, and ponding of water within the snowpack is generally minor. In plains areas, however, large amounts of water are frequently ponded within the snowpack as the result of choking of horizontal drainages by snow. The condition of soil moisture is an additional factor affecting runoff and should be evaluated in the total basin storage potential.

8-08.08 A major deficiency in the adequate assessment of storage potential is the lack of observational data on the basic snowpack characteristics involved. In general, snowpack temperatures and liquid-water content are not observed as hydrologic elements in project basins, and until such time as they are, the hydrologist must base his estimates of storage delay on experience and judgment. Additional basic research is also required on metamorphism of the snowpack, as applied to hydrology, in order to understand better the processes affecting the changes of the snowpack condition.

8-09. REFERENCES

1/ BADER, H. and others, Snow and its metamorphism, SIPRE Translation 14, (transl. by J. C. Van Tienhoven from Der Schnee and Seine Metamorphose, Beitrage zur Geologie der Schweiz, Geotechnische Serie, Hydrologie, Lieferung 3, Bern, 1939) Snow, Ice and Perma. Res. Estab., Corps of Engrs., Wilmette, Ill., January 1954.

2/ BESKOW, Gunnar, "Soil freezing and frost heaving with special application to roads and railroads," (Transl. by J. O. Osterberg from Swedish Geol. Soc., Series C, No. 375. 26th year book No. 3) Technological Inst., Northwestern Univ., November 1947.

3/ CARSLAW, H. S. and J. C. Jaeger, "Conduction of heat in solids," University of Tasmania.

4/ DE QUERVAIN, M., "Snow as a crystalline aggregate," SIPRE Translation 21, (transl. by C. M. Gottschalk from "Schnee als kristallines Aggregat," Experientia, Vol. 1, Oct. 1945) Snow, Ice and Perma. Res. Estab., Corps of Engrs., Wilmette, Ill., May 1954.

5/ DIAMOND, Marvin and W. P. Lowry, "Correlation of density of new snow with 700 mb temperature," SIPRE Res. Paper 1, Snow, Ice and Perm. Res. Estab., Corps of Engrs., Wilmette, Ill., August 1953.

6/ GERDEL, R. W., "The transmission of water through snow," Trans. Amer. Geophys. Union, Vol. 35, No. 3, June 1954, pp. 475 -485.

7/ HADLEY, W. A. and Raymond Eisenstadt, "Thermally actuated moisture migration in granular media," Trans. Amer. Geophys. Union, Vol. 36, No. 4, August 1955, pp. 615-623.

8/ HORTON, R. E., "The role of snow, ice and frost in the hydrologic cycle," Proc. Central Snow Conf., Vol. 1, December 1941.

9/ HUGHES, T. P. and Gerald Seligman, "The bearing of snow permeability and retentivity on the density increase of firn and ice band formation in glaciers," Jungfraujoch Research Party, Publication No. 3, Switzerland, 1938.

10/ KONDRAT'EVA, A. S. "Thermal conductivity of the snow cover and physical processes caused by the temperature gradient," SIPRE Translation 22, (transl. by P. P. Kapusta from "Teploprovodnost snegovogo pokrova i fizicheskie protessy, proiskhodiaschie v nem pod vlieniem temperaturnogo gradienta," Akad. Nauk SSSR, 1945), Snow, Ice and Perm. Res. Estab., Corps of Engrs., Wilmette, Ill., March 1954.

11/ NAKAYA, Ukichiro, Snow Crystals, Harvard Univ. Press, 1954.

12/ RIKHTER, G. D., "Snow cover, its formation and properties," SIPRE Translation 6, (transl. by William Mandel from "Snezhnyi pokrov, ego formirovanie i svoistva," Izdatel'stvo Akademiia Nauk SSSR, Moskva, 1945) Snow, Ice and Perm. Res. Estab., Corps of Engrs., Wilmette, Ill., August 1954.

13/ SNOW, ICE AND PERMAFROST RESEARCH ESTABLISHMENT, "Instructions for making and recording snow observations," SIPRE Instruction Manual 1, Wilmette, Ill., May 1954.

14/ WILLIAMS, G. P., "A field determination of free water content in wet snow," National Research Council, Canada, Report No. 69 of the Div. of Building Research, Ottawa, August 1955.

15/ WILSON, W. T., "An outline of the thermodynamics of snowmelt," Trans. Amer. Geophys. Union, Part I, July 1941, pp. 182-195.

TABLE 8-1

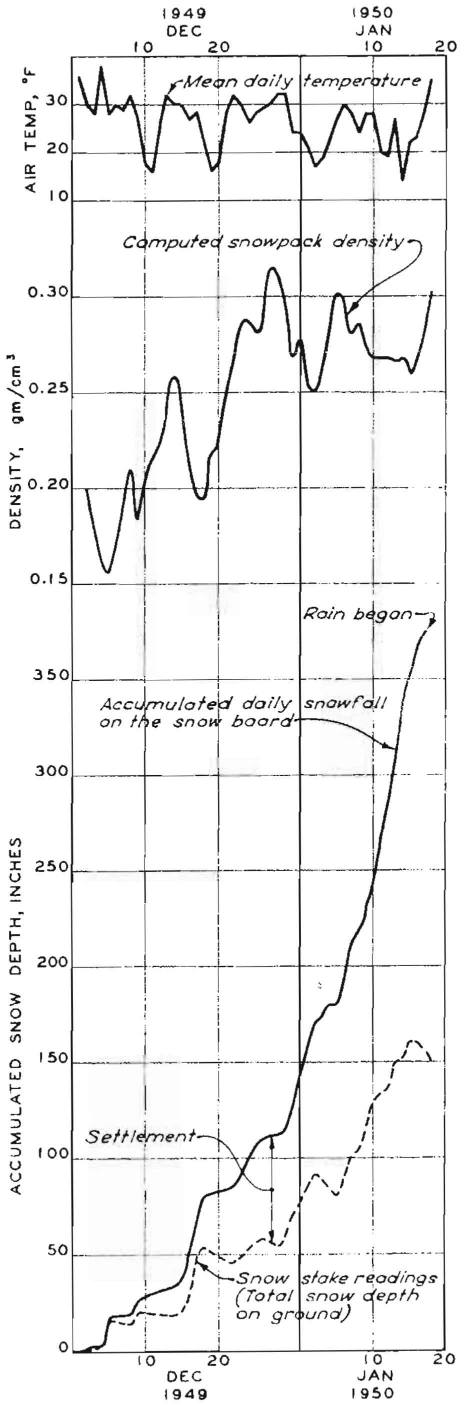
STORAGE POTENTIAL OF THE SNOWPACK 27 JANUARY 1952 1/

American River Basin, California

LINE	ITEM	ELEVATION					UNIT	SYMBOL	EQUIVALENT	REMARKS
		6900	6000	5000	4000	3000				
1	Initial snow depth	172	161	98	58	25	in.	D		Observed
2	Initial snow density	0.28	0.30	0.38	0.36	0.40	g/cc	ρ		Observed
3	Initial snowpack water equivalent	48.5	48.3	37.2	20.9	10.0	in.	W_o	ρD	
4	Initial snow temperature	-3.0	-2.5	0	0	0	°C	T_s		Observed
5	Percentage liquid water holding capacity									
	a. Before drainage	3.0	3.0	3.0	3.0	2.0	pct.	f''_{p_1}		
	b. After drainage	2.0	2.0	2.0	2.0	0	pct.	f''_{p_2}		
6	Initial liquid water content of snow	0	0	0	0.63	0.20	in.	W_f		Assumed
7	Air temperature	5.0	5.0	5.0	6.0	6.0	°C	T_a		Forecast
8	Snowmelt	0.04	0.04	0.04	0.05	0.06	in./hr.	m		Est. from temp. forecast
9	Rainfall	0.08	0.08	0.08	0.07	0.06	in./hr.	i_r		Forecast
10	Inflow	0.12	0.12	0.12	0.12	0.12	in./hr.		$i_r + m$	Line 8 + Line 9
11	Water equivalent of cold content of snow 2/	0.91	0.73	0	0	0	in.	W_c	$-W_oT_s/160$	
12	Time required to raise snow temperature to 0°C	7.6	6.1	0	0	0	hr.	t_c	$W_c/(i_r + m)$	Line 11 / Line 10
13	Water equivalent when snow reaches 0°C	49.4	48.8	37.2	20.9	10.0	in.		$W_o + W_c$	Line 3 + Line 11
14	Liquid water holding capacity	1.48	1.46	1.12	0.63	0.20	in.	S_f	$f_p\,(W_o + W_c)/100$	Line 13 x Line 5a/100
15	Liquid water deficiency	1.48	1.46	1.12	0	0	in.	S'_f	$S_f + W_f$	Line 14 + Line 6
16	Time required to contain liquid water equal to capacity	12.3	12.2	9.3	0	0	hr.	t_f	$S'_f / (i_r + m)$	Line 15 / Line 10
17	Travel rate of water in a primed snow	43	43	40	40	36	in./hr.	v_t		
18	Time of travel through snowpack 3/	4.0	3.7	2.5	1.5	0.7	hr.	t_t	D/v_t	Line 1 / Line 17
19	Transitory storage (water in transit) in snowpack	0.48	0.44	0.30	0.18	0.08	in.	S_t	$t_t\,(i_r + m)$	Line 18 x Line 10
20	Total storage time from beginning of rain to appearance of runoff	23.9	22.6	11.8	1.5	0.7	hr.	t	$t_c + t_f + t_t$	Line 12 + Line 16 + Line 18
21	Total storage in snowpack when runoff appears	2.86	2.71	1.42	0.18	0.08	in.	S	$W_c + S_f + S_t - W_f$	Line 11 + Line 14 + Line 19 - Line 6
	a. From rain	1.91	1.81	0.94	0.11	0.04	in.	P_r	$t\,i_r$	Line 20 x Line 9
	b. From melt	0.95	0.90	0.48	0.07	0.04	in.	M	$t\,m$	Line 20 x Line 8
22	Liquid water present in snow when runoff appears	1.96	1.90	1.42	0.81	0.28	in.		$S_f + S_t$	Line 14 + Line 19
23	Snowpack water equivalent when runoff appears	51.4	51.0	37.7	21.0	10.0	in.	W	$W_o + S$	Line 3 + Line 21
24	Total water deficiency at beginning of rain (includes cold content)	2.39	2.19	1.12	0	0	in.		$W_c + S'_f$	Line 11 + Line 15
25	Storage time	0.493	0.468	0.317	0.072	0.070	hr./in.		t / W_o	Line 20 / Line 3
26	Average water deficiency	0.049	0.045	0.031	0	0	in./in.		W / W_o	Line 24 / Line 3
27	Water released to runoff by the decrease in storage potential	0.49	0.48	0.37	0.21	0	in.		$W_o\,(f''_{p_1} - f''_{p_2})$	(Line 5a - Line 5b) x Line 3

Notes:

1/ See figure 2, plate 8-6 for basinwide variation in the storage potential.
2/ The terms "cold content" and "heat deficiency" are synonymous, 0°C being the reference temperature.
3/ Change in snowpack depth and water equivalent during priming period is ignored in items 18 through 27. This omission is within the accuracy of the approximations.

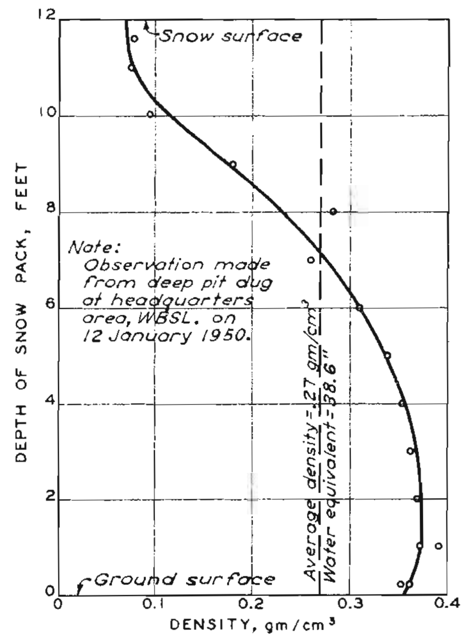


COMPACTION OF WINTER SNOW

FIGURE 1

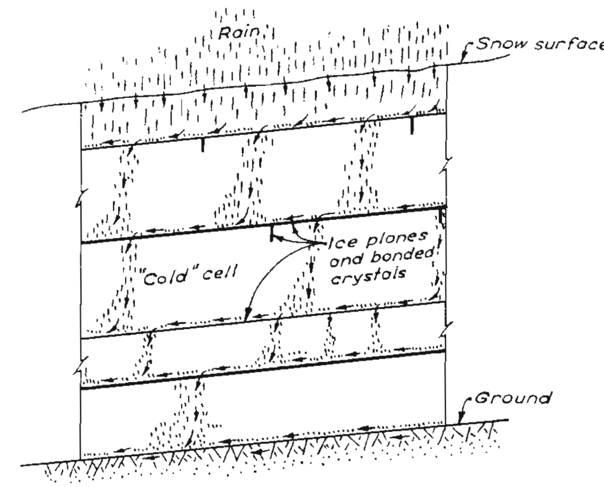
NOTES for FIGURE 1:

1. Observations made at headquarters, Station 1-A, WBSL (El. 4230), during period of snow accumulation, December and January, 1949-50.
2. During period shown, air temperature was continuously below freezing except for very short periods, and precipitation was entirely in the form of snow. Melt of the air-snow interface would be negligible.
3. Effect of ground melt is assumed to be negligible for the period shown.
4. Variation of average density was computed from the summation of daily increments of precipitation catch at Station 1-B, adjusted to the water equivalent of the snow stake computed from the mean density of snow on 12 January.
5. The mean density of new snow for the period 5 January through 18 January, computed from daily snow board measurements of water equivalent and depth, is 0.09 gm/cm³. On the basis of average temperatures during the remainder of the storm periods, the average density of new fallen snow is estimated to be 0.11 gm/cm³ for the entire period.
6. During this storm period, the average precipitation catch at the snow stake was about 115% of the catch at the snow board.



DENSITY VARIATION IN A WINTER SNOW PACK, WBSL 12 JANUARY 1950

FIGURE 2

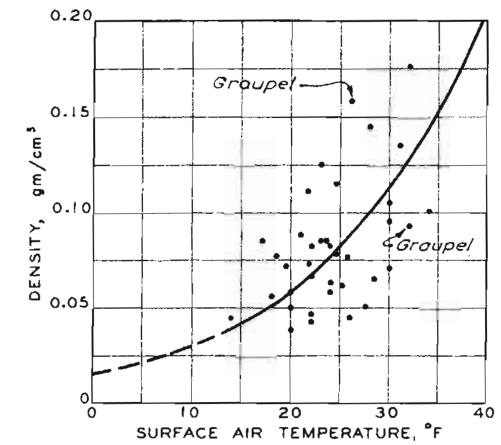


SCHEMATIC DIAGRAM ILLUSTRATING THE FLOW OF WATER WITHIN THE SNOW PACK

FIGURE 3

NOTES:

1. Ice planes and bonded crystals of varying thickness and permeability are usually formed during clear weather between snowstorms as a result of melt water refreezing during the night and also as a result of sublimation process within the snow pack. Crusts may also form from the action of wind, and cooling of surface-snow matrix by evaporation.
2. "Cold" cells of temperature below melting point continue to exist in the snow pack between ice planes and water courses until the ice planes have disintegrated and allowed the percolating water to reach these cells or sufficient heat has penetrated to raise the temperature to 0°C.
3. Capillary water is stored near the ground and above the ice planes and in small voids between snow crystals. In addition water is also held against gravity on snow particles by adhesion or surface tension.



DENSITY OF NEW FALLEN SNOW

FIGURE 4

NOTES:

1. Density measurements were made by SIPRE personnel at CSSL headquarters (El. 6900). Air temperatures were taken at about the time of density measurements at 4 feet above snow surface.
2. Times of accumulation of snow before observation were variable and were always less than 24 hours. The average time was probably in the range between 6 and 12 hours. The results, therefore, cannot be applied directly to the usually observed 24-hour snowfalls.
3. Variability is also introduced into the above relationship as the result of varying rates of snow accumulation which have not been considered.

GENERAL NOTE:

Snow densities in gm/cm³ on this plate are equivalent to commonly used densities in percent, divided by 100.

SNOW INVESTIGATIONS SUMMARY REPORT

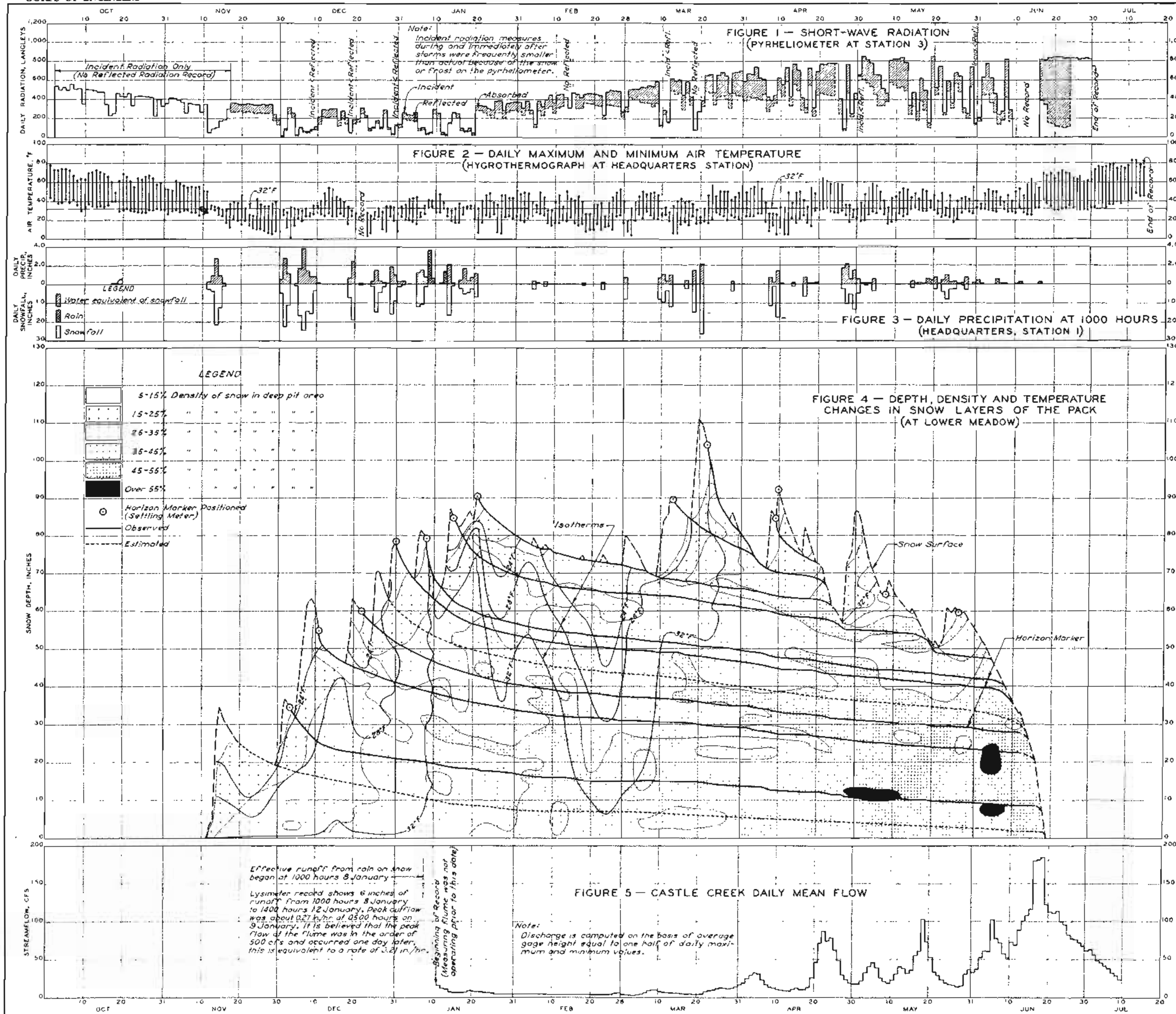
SNOW HYDROLOGY

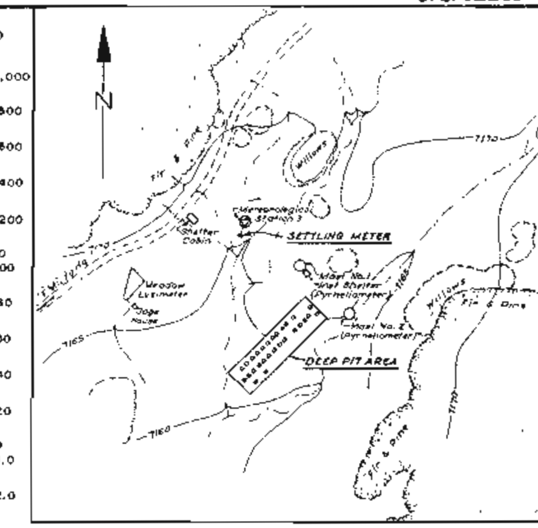
DENSITY AND STRUCTURE OF WINTER SNOWPACK

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U.S. ARMY

TO ACCOMPANY REPORT DATED 30 JUNE 1956

PD-20-25/46





VICINITY MAP — LOWER MEADOW

SCALE IN FEET

Notes:

1. Settling meter readings, showing positions of snow layers in the snow pack continuously through the season were obtained from the Snow, Ice, and Permafrost Research Establishment installation at CSSL Lower Meadow (Station 3) during the 1952-1953 season. The slide wire settling meter is described in SIPRE translation No. 14, "Snow and its Metamorphism" ("Der Schnee und Seine Metamorphose", Beitrage zur Geologie der Schweiz, Geotechnische Serie, Hydrologie, Lieferung 3, Bern (1939)).
2. Temperature and density profiles of the snow pack were obtained from deep snow pits, dug individually at the deep pit area at the Lower Meadow, CSSL, (see Vicinity Map, above), at approximately one week intervals. Observations were made, under supervision of SIPRE, generally between 1000 and 1700 hours.
3. No attempt was made to show the average temperature of the surface snow layer, where daily freezing and thawing is affected by the diurnal change in heat supply at the surface of the snow pack. Note sudden change in thermal character of snow pack during occurance of rain on 8-9 January 1953.
4. Snow densities were determined by weighing horizontal samples taken with standard 500 cc steel cylinders from south wall of snow pits. The average density of the snow pack by this method is one to five percent smaller than the density of a vertical core sampled by a Mount-Rose snow tube, because the horizontal samples from the pack do not contain all ice lenses, while the vertical core in a Mount-Rose tube contains all layers through the entire snow pack.
5. Plots of daily short-wave radiation, maximum and minimum temperatures, precipitation, snowfall, and streamflow in Castle Creek, show the march of hydrometeorologic events during snow accumulation and melt periods.
6. Castle Creek discharge (Figure 5) was computed on the basis of average gage heights equal to one-half of the daily maximum and minimum values.
7. Daily maximum and minimum temperatures and daily precipitation and snowfall were obtained from records for Station 1, CSSL, located approximately 1¼ miles Southwest of the Lower Meadow (Station 3).
8. Deep pit site went bare when basin snow cover was about 50%.
9. See Figure 1, Plate 8-3 for snow classification, temperature, and density profiles.

SNOW INVESTIGATIONS
SUMMARY REPORT

SNOW HYDROLOGY

SNOWPACK CHARACTERISTICS CSSL, 1952 - 53

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U S ARMY

PREPARED: R.S.S. R.M.	SUBMITTED: R.G.S.	TO ACCOMPANY REPORT DATED 30 JUNE 1956
DRAWN: B.V.	APPROVED: D.M.R.	PD-20-25/47

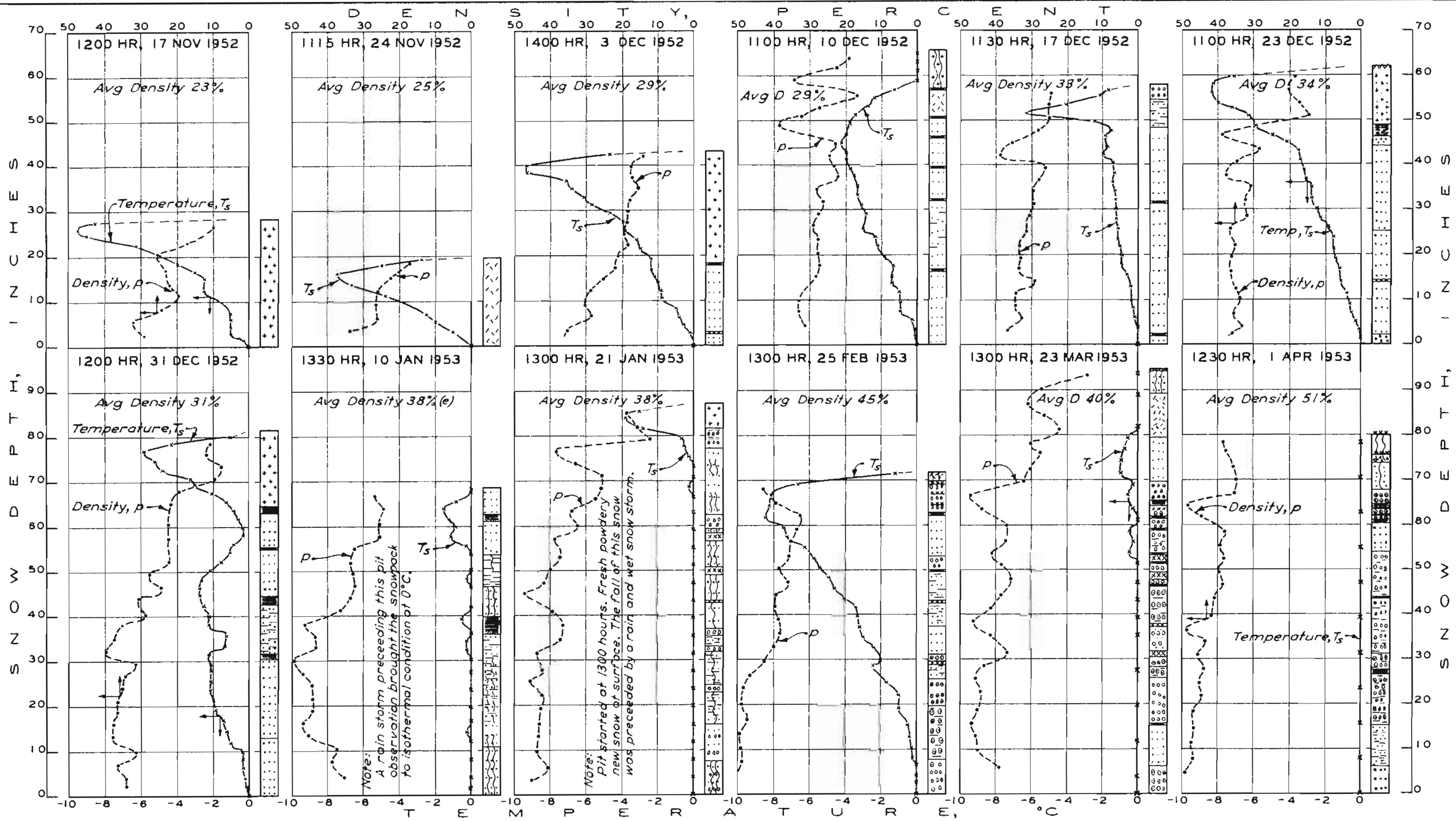


FIGURE 1 — DEEP PIT OBSERVATIONS, CSSL, 1952-53 WATER YEAR

LEGEND

- FEATHERY (DENDRITIC) — "New Snow", "Powder Snow"
- FELTLIKE (INTRANSFORMATION) — "Settling Snow"
- "Old Snow", "Settled Snow" "Firn"
- GRANULAR (ISOMETRIC) — 0 - 1.5 mm Diameter; 1.5 - 3.0 mm Diameter; Over 3.0 mm Diameter; Hoar Frost
- ICE — Compact; Granular
- SURFACE — Wind Erosion; Crust
- WET SNOW — Contains Free Water

NOTES for FIGURE 1:-

1. *Average pack density determined by vertical sample with Mt. Rose sampler.*
2. *Horizontal density samples taken with 500 cc cylindrical sample tubes at positions indicated thus: •. They are taken in homogenous horizons and do not include ice layers.*
3. *Temperature in °C taken with Weston bimatellic dial thermometers at positions indicated thus: ××.*
4. *0°C isothermal condition of the snowpack continued after 1 April except nightly cooling effect of the surface layer by the outgoing longwave radiation.*
5. *Observations made at Lower Meadow, CSSL, under direction of SIPRE.*
6. *Only selected observations are shown to illustrate progress of change, also see Plate 8-1.*

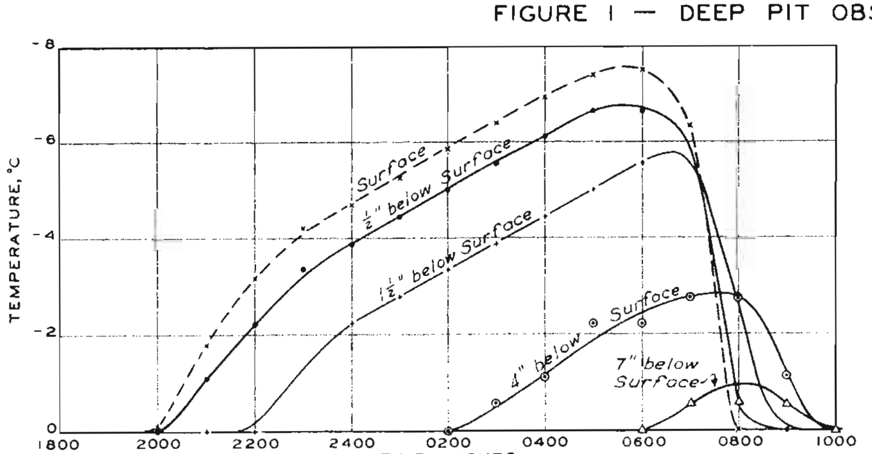


FIGURE 2 — NIGHTTIME COOLING IN SURFACE LAYER OF INITIALLY MOIST SNOW, 12-13 MAY 1954

NOTES for FIGURES 2 and 3:—

1. *Observations made at Lower Meadow, CSSL, in connection with lysimeter studies of snowmelt.*
2. *Air and snow temperatures obtained by use of Thermohms from which continuous records were obtained through the night.*

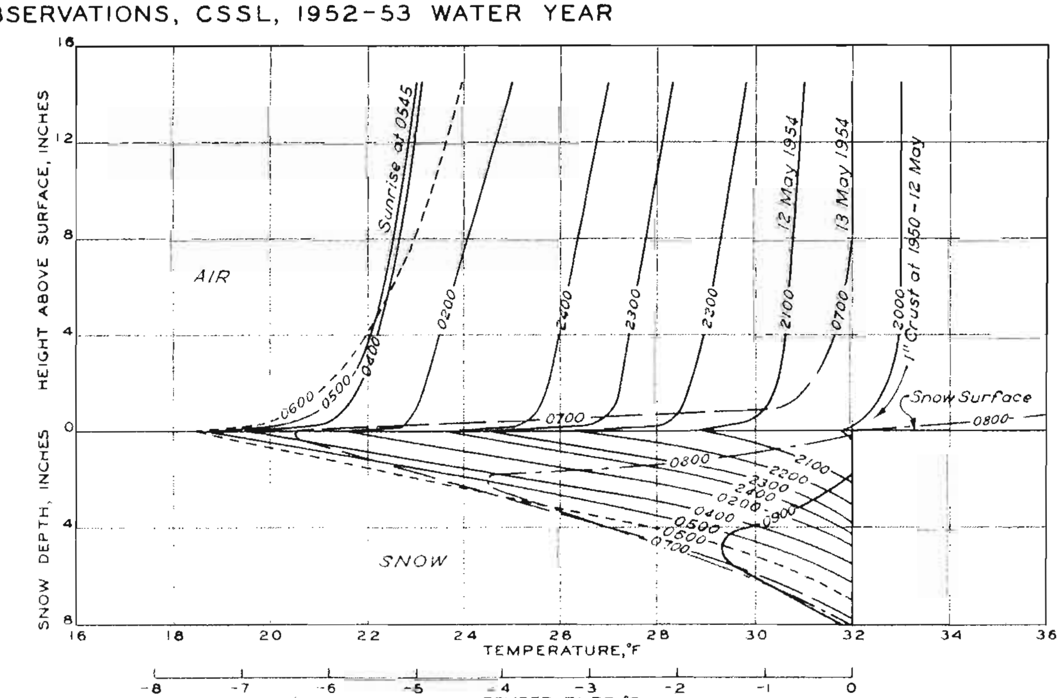


FIGURE 3 — TEMPERATURE PROFILES NEAR SNOW SURFACE DURING NIGHT OF 12-13 MAY 1954

SNOW INVESTIGATIONS SUMMARY REPORT

SNOW HYDROLOGY

SNOWPACK CHARACTERISTICS CSSL

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION CORPS OF ENGINEERS U.S. ARMY

PREPARED DW/P.R.B. SUBMITTED R.H.B. TO ACCOMPANY REPORT DATED 30 JUNE 1955

DRAWN A.K. APPROVED U.S.A.

PD-20-25/48

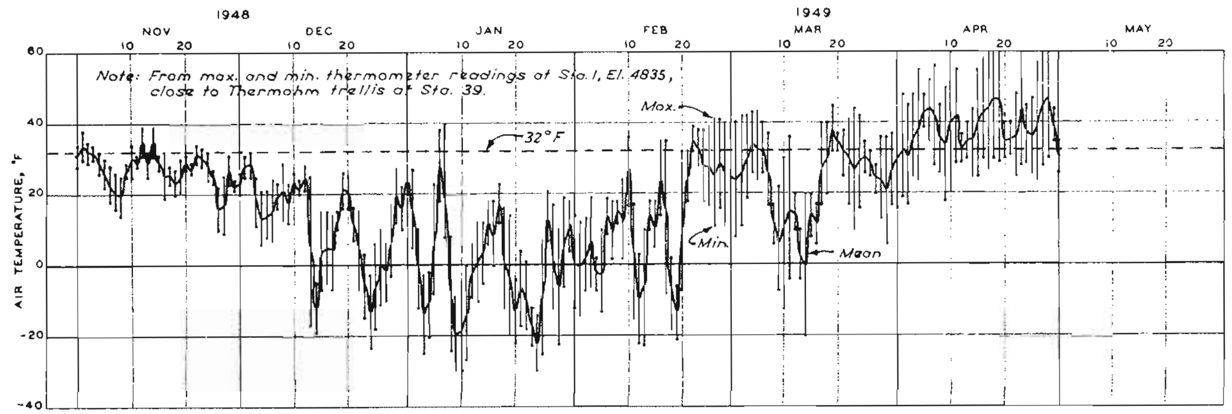


FIG. 1 - AIR TEMPERATURE, UCSL

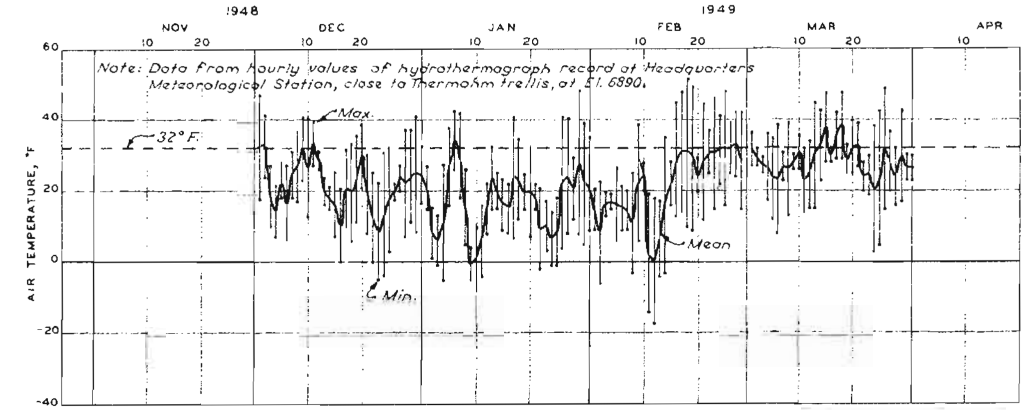


FIG. 3 - AIR TEMPERATURE, CSSL

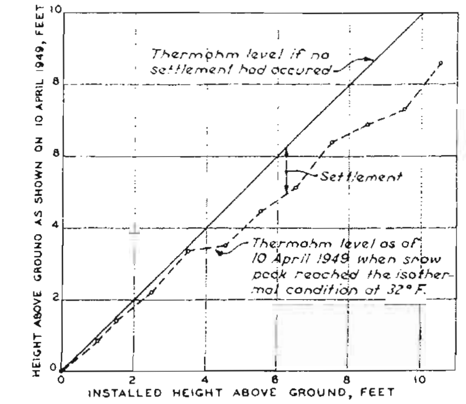


FIG. 5 - POSITION OF THERMOHMS, CSSL

Notes:

1. Temperatures within the snow pack were measured on a trellis by Thermohms suspended at different heights above ground surface and were recorded every fifteen minutes on a recorder chart.
2. Snow depth was noted usually once a day at the trellis. Snow depth at CSSL after 3 March 1949 was estimated from the daily snow stake readings near the trellis.
3. Figure 5 shows the shift in the position of Thermohms at CSSL as of 10 April 1949 when the pack was excavated and the new heights were determined. No such information is available for UCSL.
4. Estimated values are shown by dotted lines.
5. The isotherms shown represent average of daily max. and min. temperatures as obtained at each Thermohm level within the snow pack. The temperature pattern near the surface of the snow is affected by the diurnal pattern of the air temperature above it, and snow surface temperature data are not available to show the temperature variations near the surface. The figures show the seasonal trend and relatively rapid response of snow temperature to that of surface air.

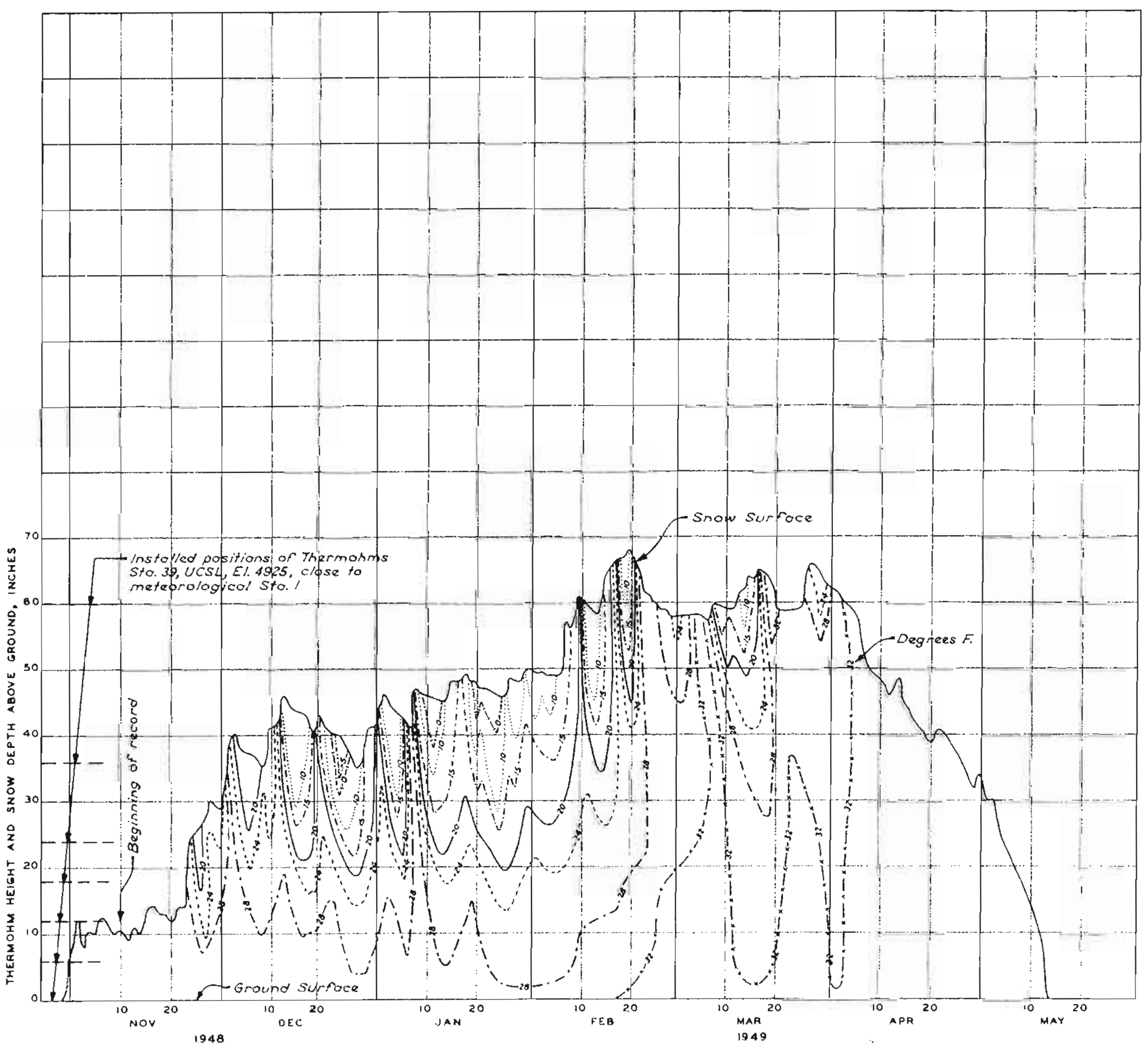


FIG. 2 - ISOTHERMS IN SNOW PACK, UCSL

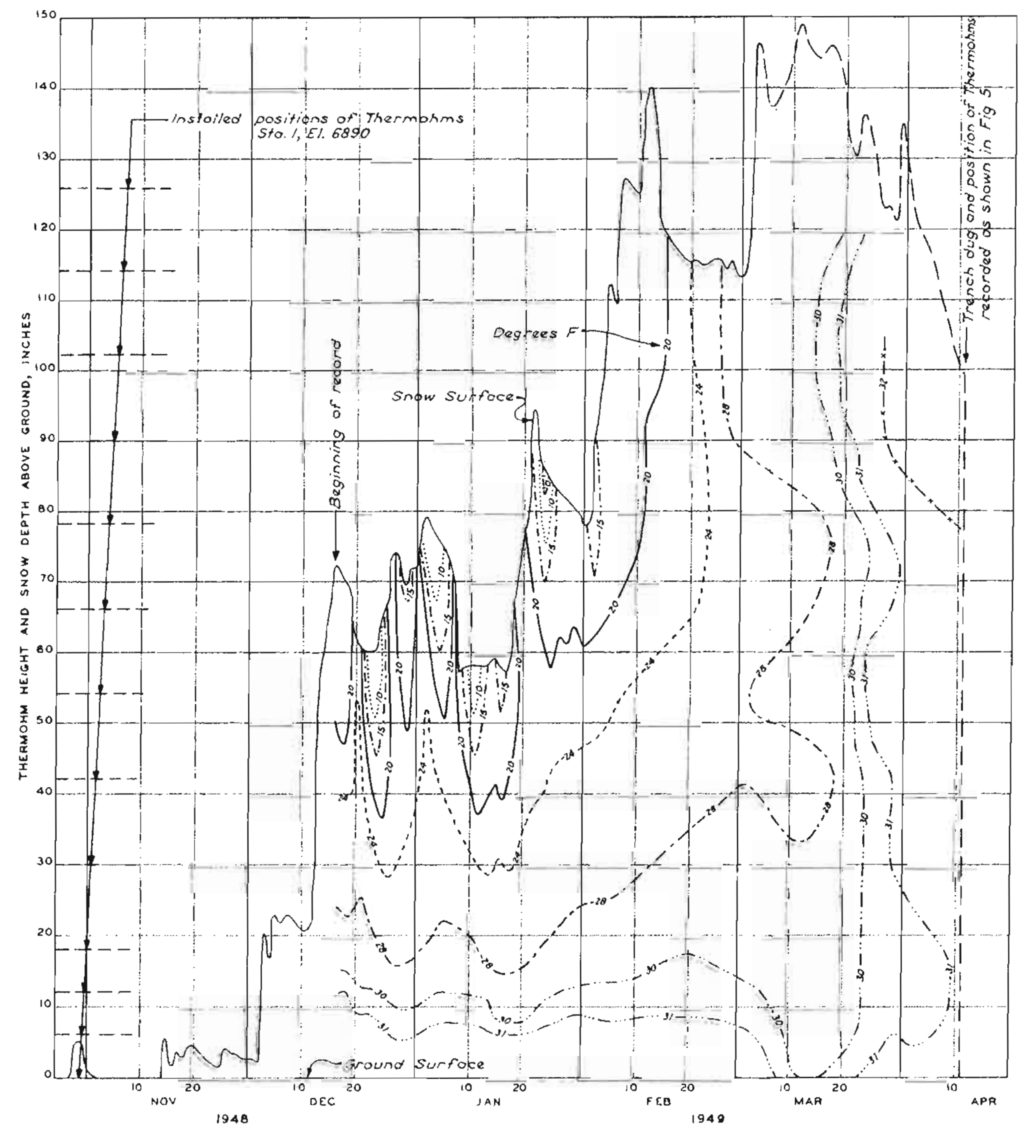


FIG. 4 - ISOTHERMS IN SNOW PACK, CSSL

SNOW INVESTIGATIONS
SUMMARY REPORT

SNOW HYDROLOGY

SNOWPACK TEMPERATURES
UCSL AND CSSL, 1948 - 49

SHEET 1 OF 2

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

PREPARED: PBB-PM SUBMITTED: PBB
DRAWN: WJM APPROVED: DMR

TO ACCOMPANY REPORT
DATED 30 JUNE 1956

PD-20-25/49

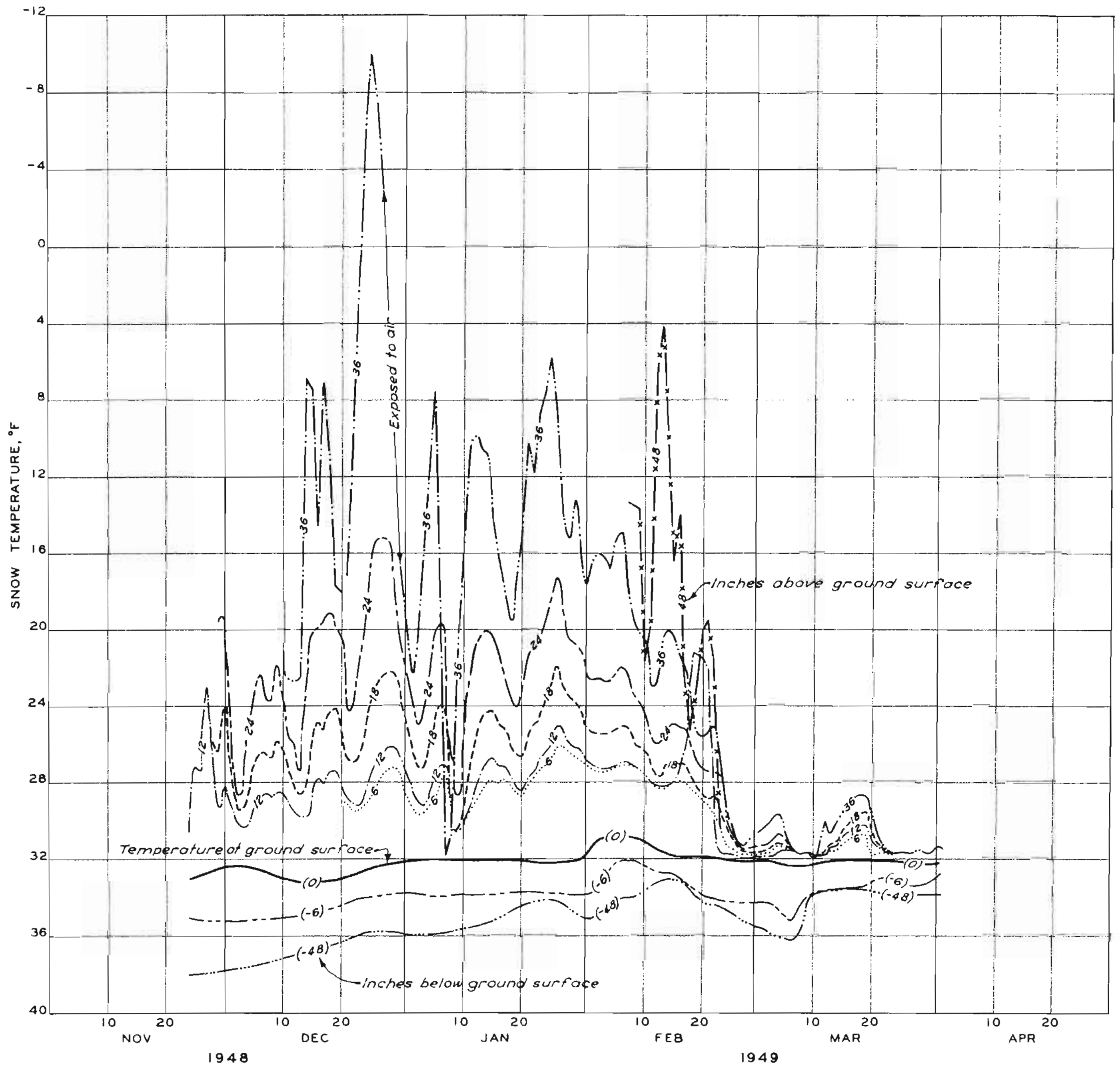


FIGURE 1 — SNOW AND GROUND TEMPERATURES, UCSL

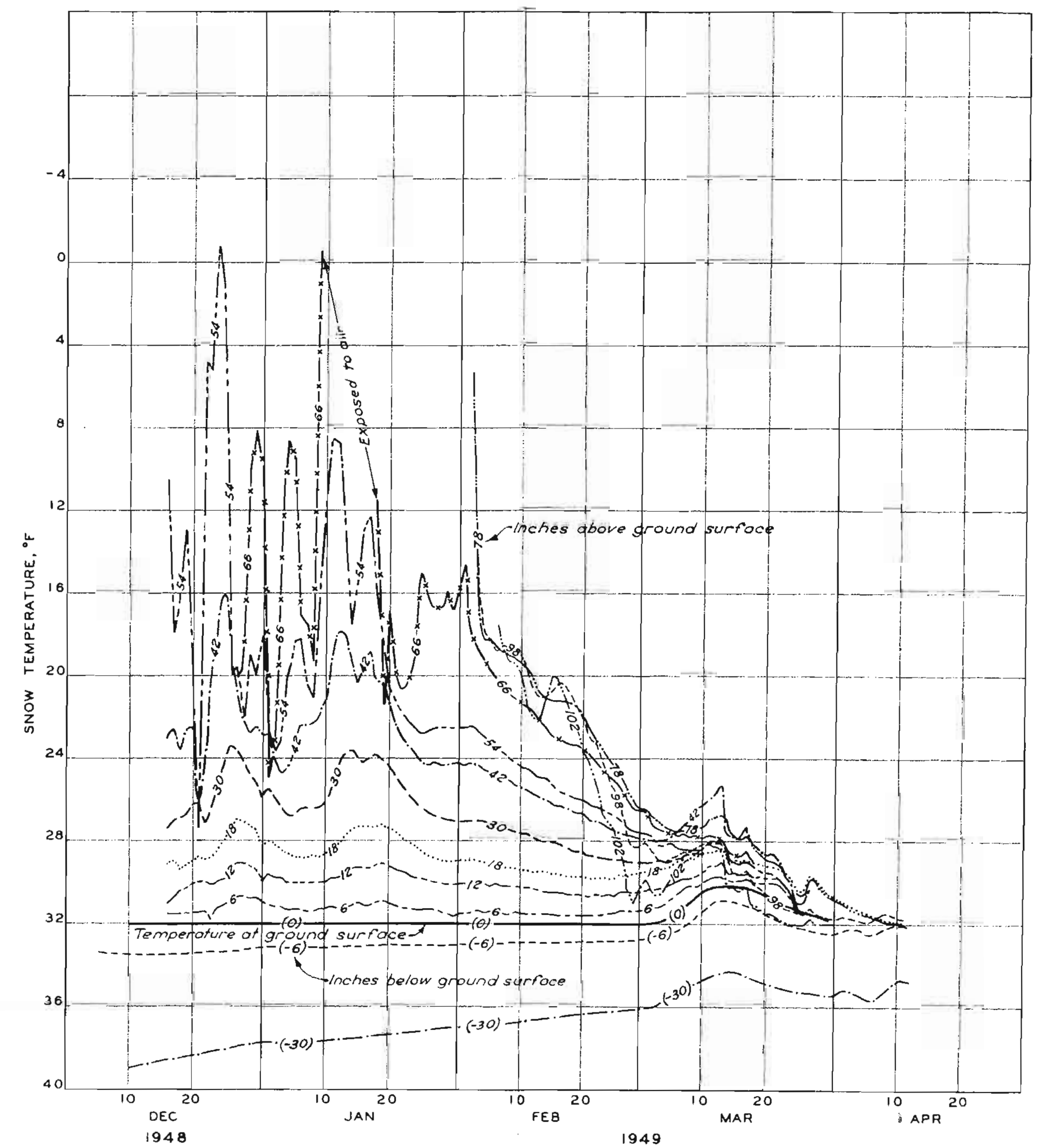


FIGURE 2 — SNOW AND GROUND TEMPERATURES, CSSL

Notes:

1. Data for temperature variation of various levels in snow pack and ground obtained from Thermohm data for Station 39 at UCSL and Station 1 at CSSL.
2. Snow depths and air temperatures for this period as shown on Plate 8-4.
3. All temperatures on these figures are those recorded at 0700 of each day.
4. Surface layers show larger temperature variations with respect to time. The amplitude of these variations diminishes with increasing depth of snow from the surface.
5. When the snow pack reaches the isothermal condition at 32°F, there may still occur diurnal temperature variation in the "crust" layer (averaging about 6" in depth), which freezes by night and thaws during the day.

SNOW INVESTIGATIONS
SUMMARY REPORT
SNOW HYDROLOGY

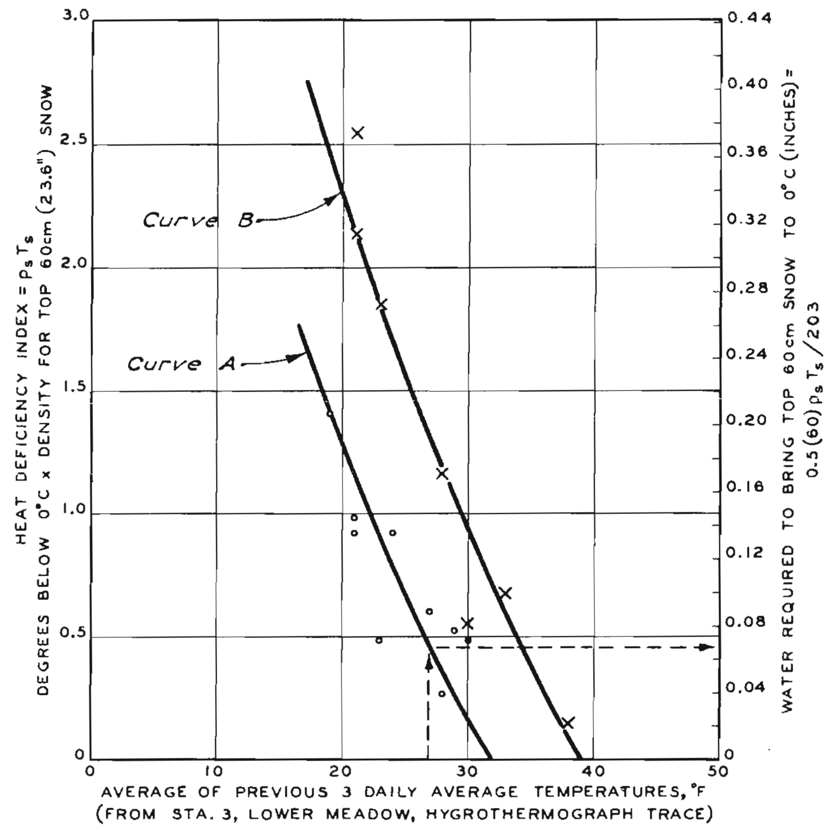
SNOWPACK TEMPERATURES
UCSL AND CSSL, 1948-49

SHEET 2 OF 2

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U.S. ARMY

PREPARED: P.M. SUBMITTED: P.B.B. TO ACCOMPANY REPORT DATED 30 JUNE 1950
DRAWN: B.V. APPROVED: D.M.R.

PD-20-25/50

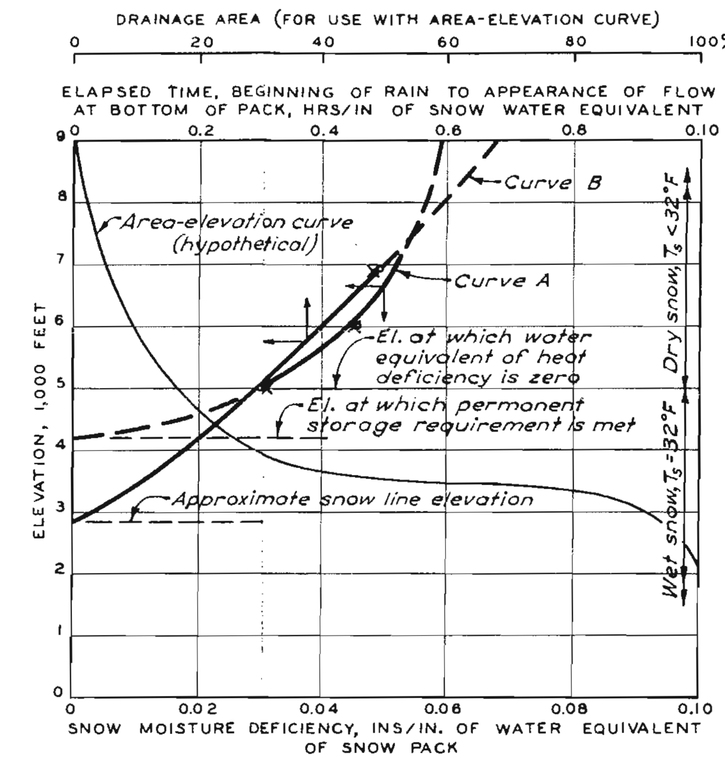


COLD CONTENT AND MOISTURE DEFICIENCY
(TOP 60 CM OF SNOW)

FIGURE 1

Notes:

1. *The curves are based on weekly snow pit observations and temperature data at Lower Meadow for 1952-53 season.*
2. *Curve A - Used when preceding 3 days are generally overcast.*
3. *Curve B - Used when preceding 3 days are generally clear.*
4. *Below 60 cm depth, the snow pack temperature does not change appreciably between observations.*
5. *Example:*
   The average temperature for the 3 days preceding the rainfall is 27°F, and the weather is overcast. From Curve A, water required to raise the temperature of the top 24 inches (60 cm) of snow pack to 32°F is 0.07 inches. The remainder of the pack (60 inches) has a temperature of approximately 29°F and a density of 0.32 and requires 0.22 inches of water (= 60 x 0.32 x 0.5 x 1.7/80). Thus a total of approximately 0.29 inches is required to raise the temperature of the snow pack to 32°F at this site.
6. *Symbols:*
   ρ_s = *snowpack density*, T_s = *snowpack temperature*



ILLUSTRATIVE EXAMPLE OF DETENTION OF RUNOFF BY SNOW PACK OVER BASIN

FIGURE 2

Notes:

1. *Curve A represents evaluation of snow moisture deficiency over basin, as a function of elevation, in terms of unit water equivalent of snow pack.*
2. *Curve B represents time for water to reach bottom of snow pack, including time required to snow moisture deficiency and time for transit of water through pack, on the basis of an assumed rate of inflow at top of pack of 0.12 inches per hour, including rain and snow melt.*
3. *Data for these curves derived from analysis of 27 January 1952 snow surveys along U.S. Highway 40, between Auburn and Soda Springs, California, and apply only to that condition. For other locations and conditions, similar curves can be constructed from observations of snow characteristics which adequately sample basin conditions, particularly with regard to elevation differences.*
4. *Use of a basin area-elevation curve (such as hypothetical one shown) allows direct reading of (1) snow-covered area, (2) area of snowpack primed to yield runoff directly without storage, since temperature and water-holding requirements are satisfied, (3) area of snowpack at 32°F with moisture deficiency, thus requiring storage, and (4) area of snowpack at <32°F, which will store water to satisfy both temperature and water-holding requirements before yielding runoff.*

SNOW INVESTIGATIONS
SUMMARY REPORT

SNOW HYDROLOGY

COLD CONTENT AND MOISTURE DEFICIENCY OF THE SNOWPACK

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U.S. ARMY

PREP.: P.B.B. SUBM.: P.B.B. TO ACCOMPANY REPORT DATED 30 JUNE 1956
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PD-20-25/51

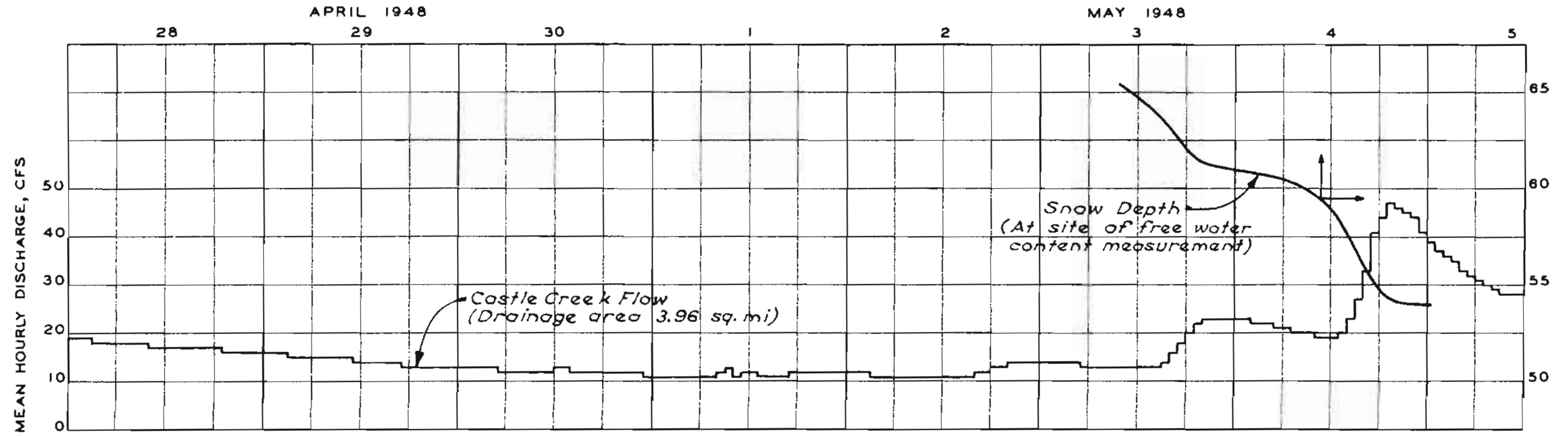


FIGURE 1 - DISCHARGE & SNOW DEPTH

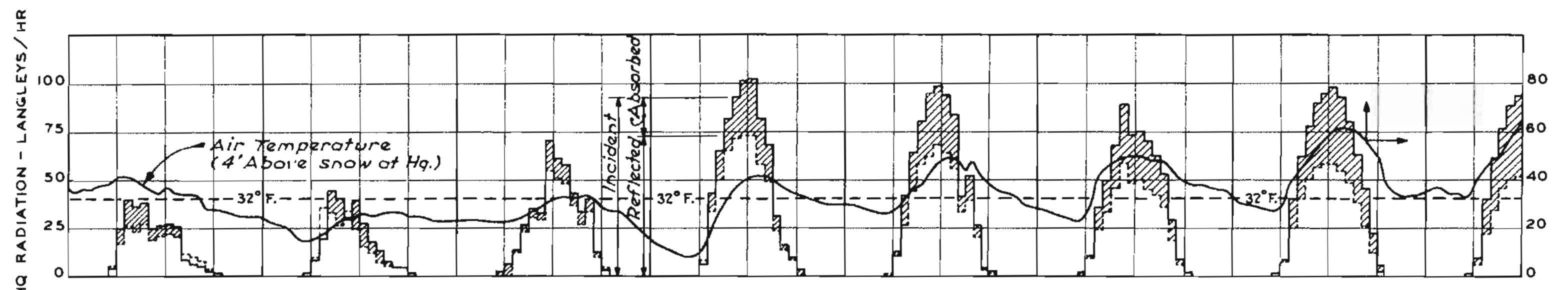


FIGURE 2 - SOLAR RADIATION & TEMPERATURE

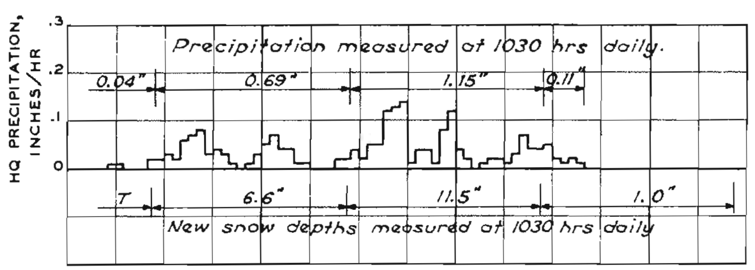


FIGURE 3 - PRECIPITATION

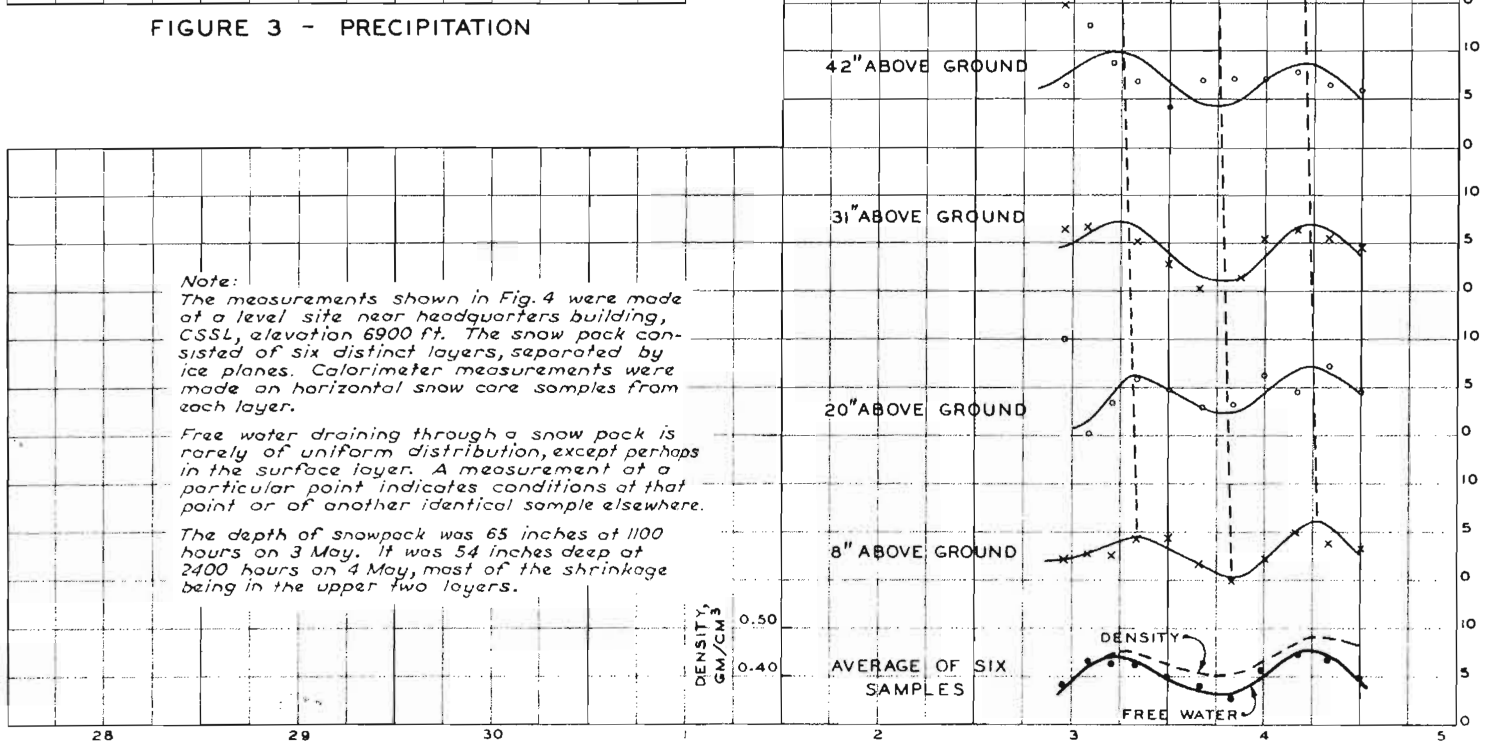


FIGURE 4 - DIURNAL VARIATION IN FREE WATER CONTENT OF SNOWPACK

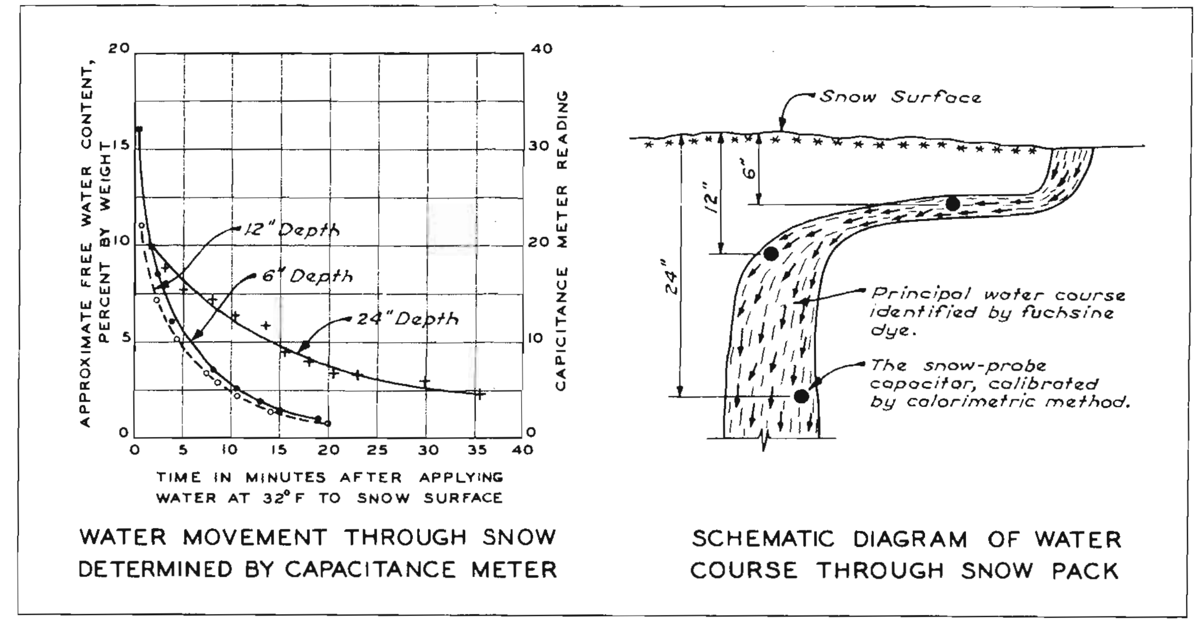
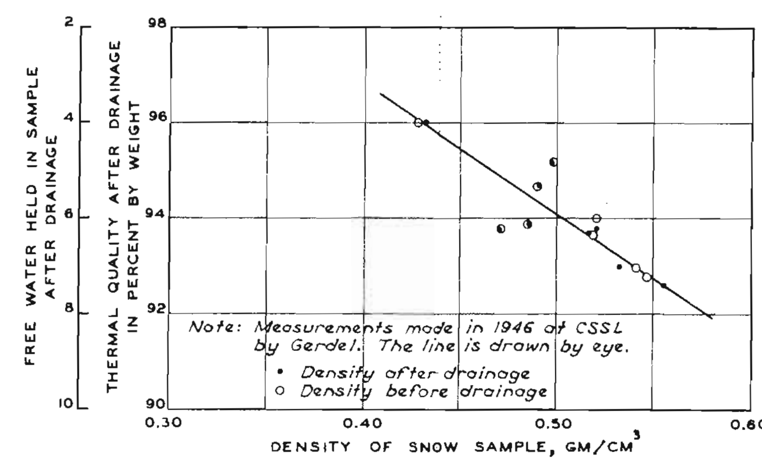


FIGURE 5

Notes:

1. Data for Figure 5 were determined by Dr. R. W. Gerdel from measurements made in May, 1948 at CSSL. See Transactions, AGU, June, 1954 for description of instrument and methods used.
2. Density of snow was 0.46 gm/cm³ before experiment.
3. Temperature of liquid water and snow pack at 32°F before and after experiment.



WATER HOLDING CAPACITY OF "RIPE" SNOW

FIGURE 6

Note:
Snow densities in gm/cm³ on this plate are equivalent to commonly used densities in percent, divided by 100.

SNOW INVESTIGATIONS SUMMARY REPORT

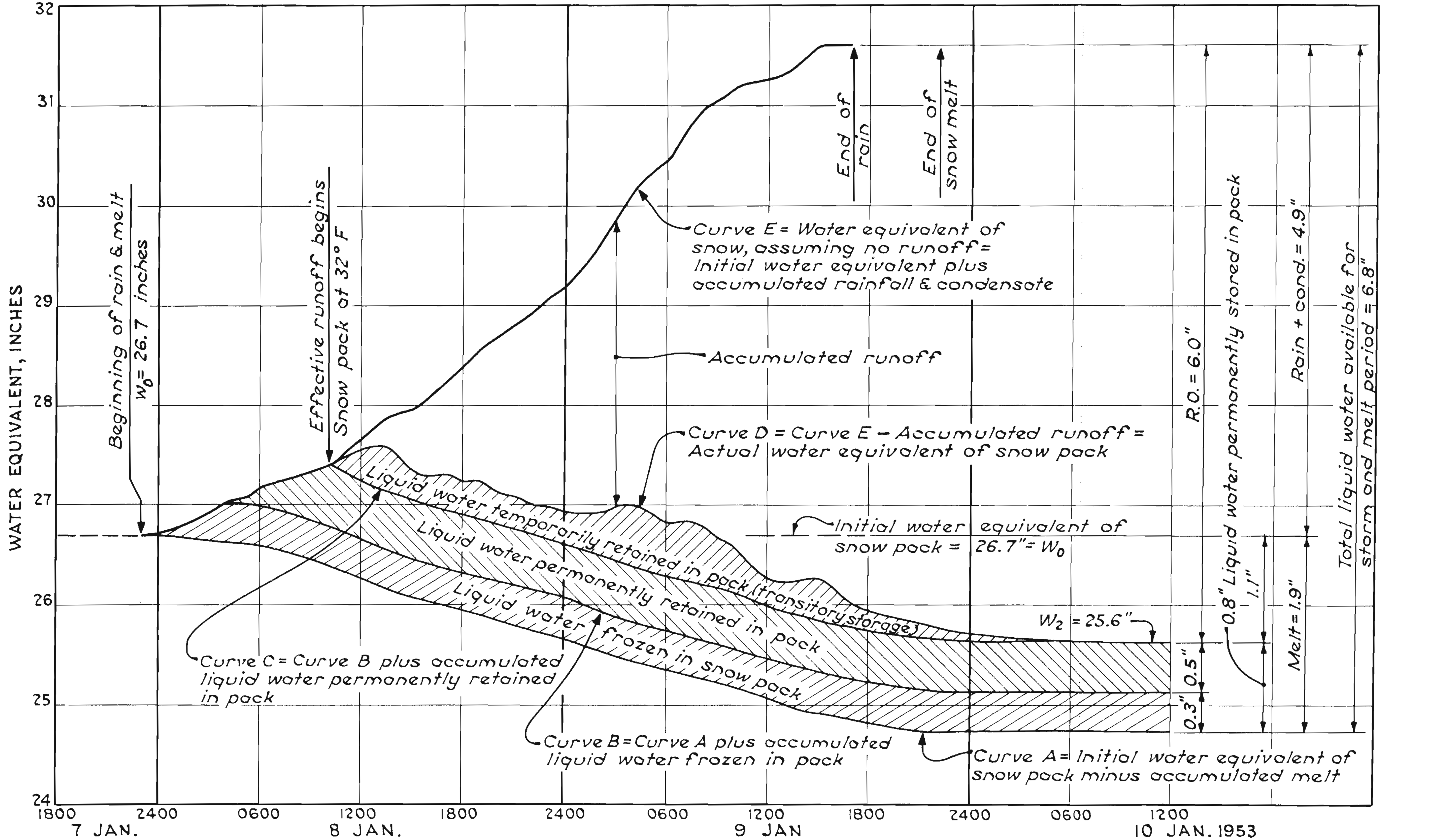
SNOW HYDROLOGY

LIQUID WATER IN SNOW

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

PREPARED: PBB-PM SUBMITTED: PBB
DRAWN: WJM APPROVED: DMR
TO ACCOMPANY REPORT DATED 30 JUNE 1956

PD-20-25/52



Note:
Of the liquid water entering the snow pack, 0.3" is used in raising the temperature of pack to 32° F. and approximately 0.5" is permanently retained in pack. The remainder (6.0") of the inflow appeared as runoff.

SNOW INVESTIGATIONS
SNOW HYDROLOGY
SNOWPACK WATER BALANCE DURING RAIN ON SNOW
CSSL HEADQUARTERS LYSIMETER

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U.S. ARMY

PREP: PBB SUBM: PBB TO ACCOMPANY REPORT DATED 30 JUNE 1956
DRAWN: WJM APPR: DMR PD-20-25/53

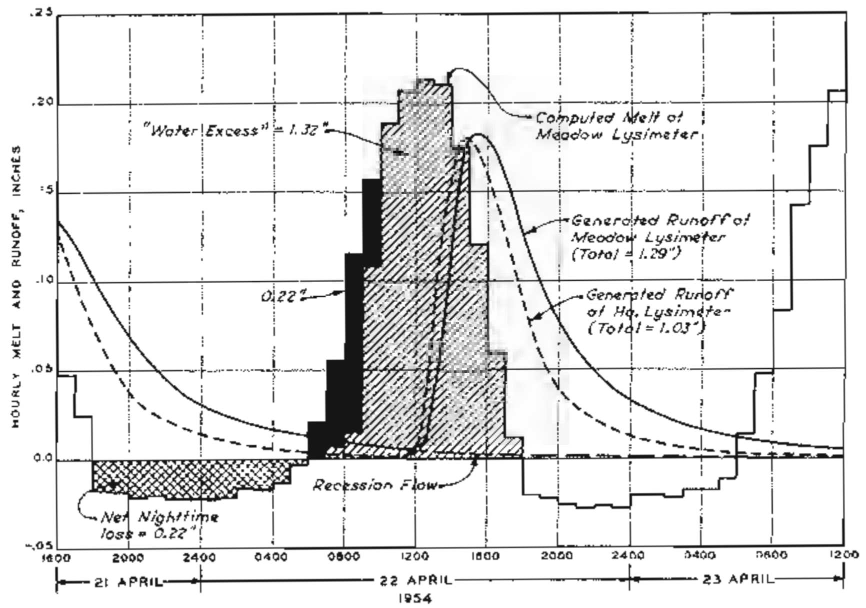


FIGURE 1 - CLOUDLESS DAY

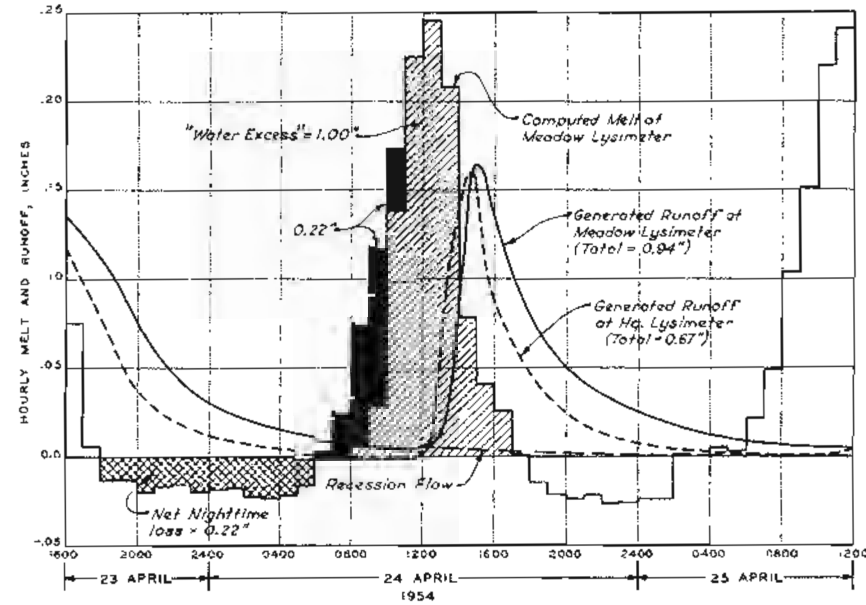


FIGURE 2 - PARTLY CLOUDY DAY

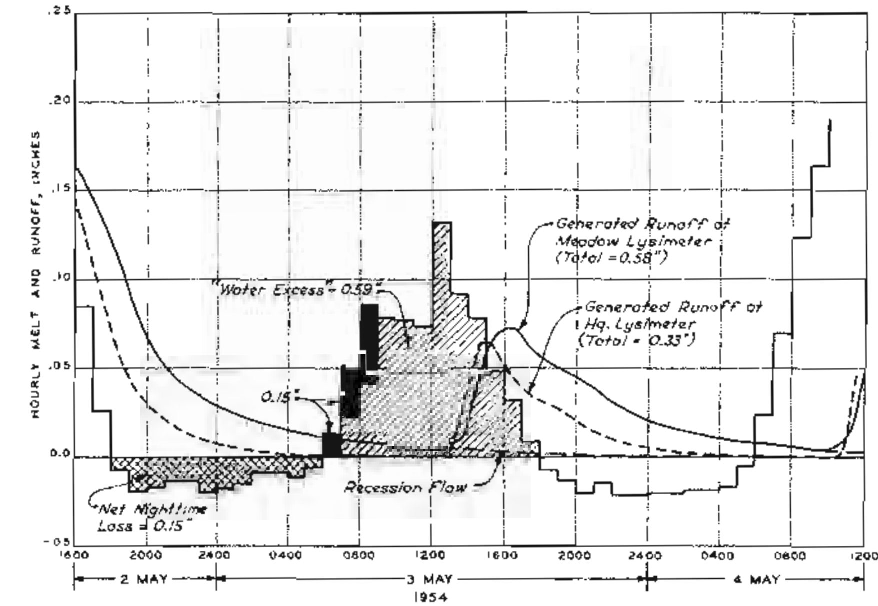


FIGURE 3 - CLOUDY DAY

NOTES for FIGURES 1, 2, & 3

1. Character of snow pack in Meadow lysimeter:

DATE	DEPTH	AVG DENSITY	AVG. TEMP.
22 April	54	0.50	0°C.
24 "	45	0.51	0°C.
3 May	43	0.52	0°C.

2. For the computation of the hourly snow melt by heat balance the 24-hour day was taken from 1900-1800 hours. See Research Note No. 25.

3. The time distribution of the portion of daytime melt going into "permanent storage" or being used up in reducing the cold content of the snow is only an approximation to illustrate the penetration and melting effect of solar radiation below the "cold" surface layer, as well as a likely small amount of surface melt water which may pass through the cold zone without refreezing, before the entire crust reaches the melting point.

4. The following comments are offered to explain the deficiency in runoff of Headquarters snow lysimeter
   a. Snow melt of Headquarters lysimeter is approximately 5 per cent less than that of the Lower Meadow lysimeter.
   b. The snow pack of Headquarters lysimeter was in its natural form on the rock foundation. The continuous ice planes in the snow pack impeded and undoubtedly caused a portion of the melt water to reach outside the area of the lysimeter. In contrast, the Lower Meadow lysimeter snow was separated from the natural snow cover by a half inch slot. Thus all surface melt was led to the impervious floor (18 inches below the bottom of the snow pack) and thence to the gage tank.

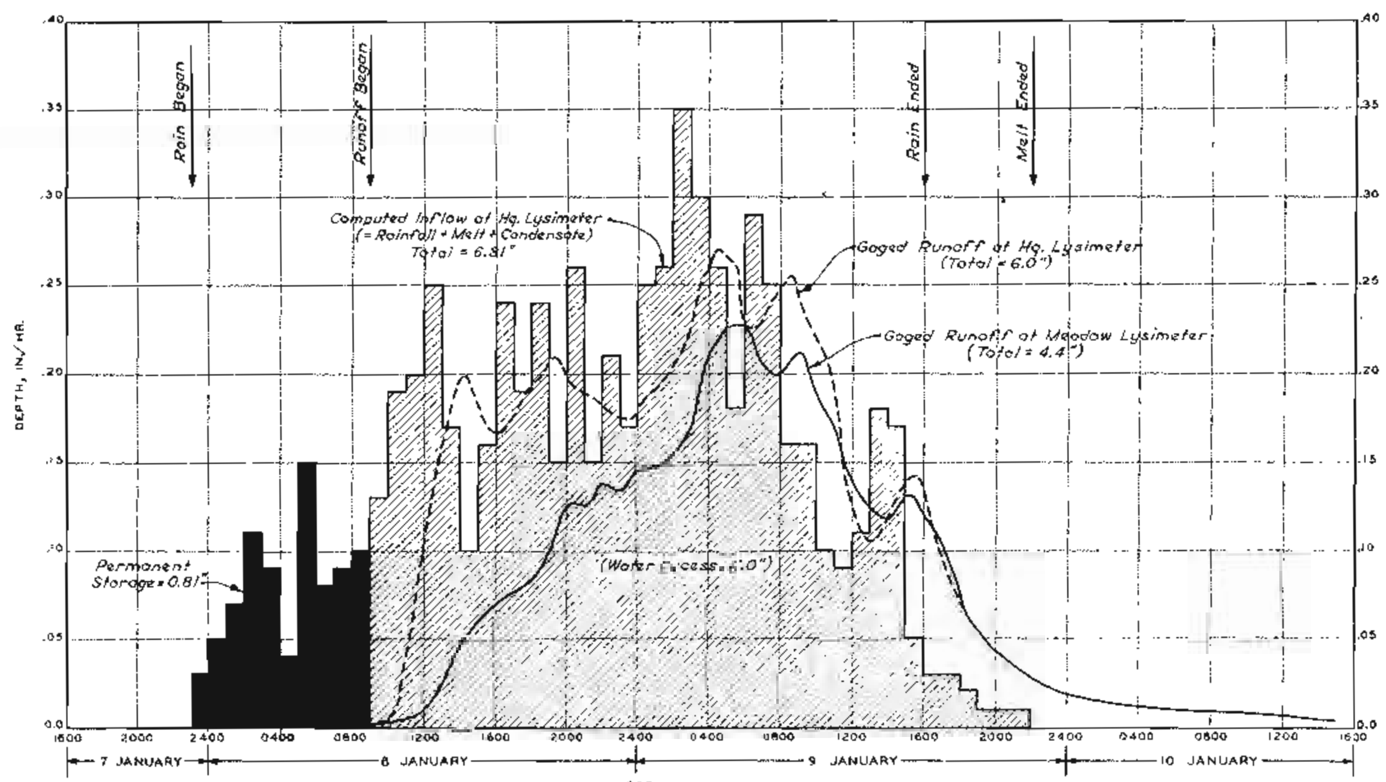


FIGURE 4 - RAIN ON SNOW

NOTES for FIGURE 4

1. Before beginning of rain, depth, average density, and temperature of the snow pack at Headquarters lysimeter were 84", 0.32 and -3°C., respectively.

2. Rainfall at Headquarters Friez gage was 4.8" against 4.3 at Meadow Stevens gage.

3. The deficiency in runoff at the Meadow lysimeter was due to dome-like configuration of the ice planes. Prior to the disintegration of the ice planes, melt and rainwater passing through the pack ran over the ice planes to outside the boundary of the lysimeter. At Headquarters lysimeter, it is believed that the ice planes were not as impervious and contribution from outside the lysimeter area equalled the inflow lost from the lysimeter area.

4. The analysis of this rain on snow event and the reconstitution of the runoff hydrograph at Headquarters lysimeter are in Research Note No. 18, 15 May 1954.

LEGEND

- (cross-hatched) Water equivalent of night-time heat loss from snow
- (solid black) Liquid water from melt or rain which upon refreezing within the "cold" layers of snow pack releases its latent heat of fusion and raises the temperature of snow to the melting point and also provides for liquid water adsorbed on the snow crystals. In spring, it is the melt-water equivalent of heat gain required to replenish night-time heat loss from the surface layer. Such water does not contribute to the "water excess" or runoff until it has melted.
- (hatched) "Water excess" or inflow. On clear days it is that portion of snow melt which passes through the pack and appears as runoff

GENERAL NOTES

1. Drainage areas:
   Meadow lysimeter 600 sq. ft.
   Headquarters lysimeter 1300 sq. ft.
2. Details of construction of lysimeters are shown in Research Notes Nos. 18 and 25.

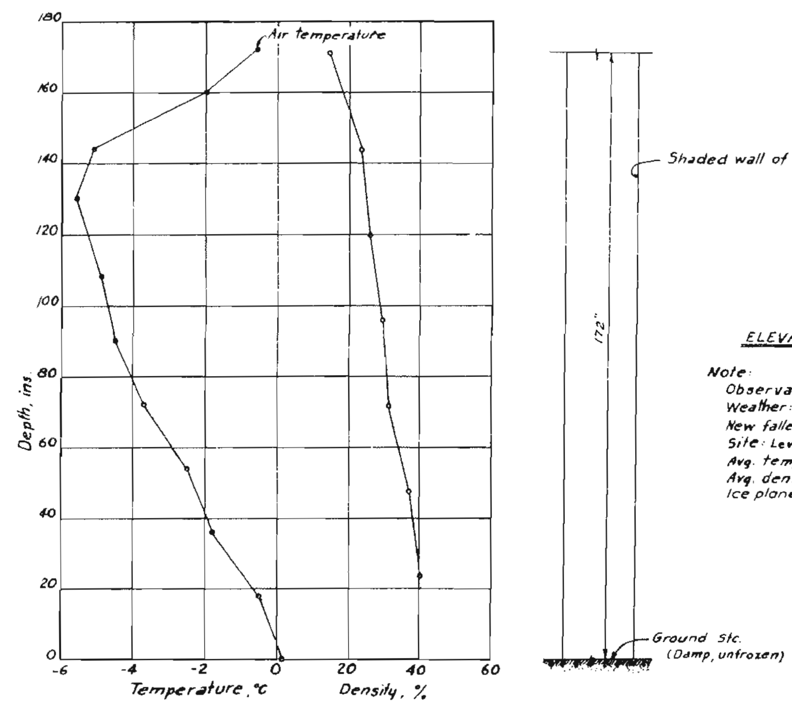
SNOW INVESTIGATIONS
SUMMARY REPORT

SNOW HYDROLOGY

LYSIMETER RUNOFF HYDROGRAPHS

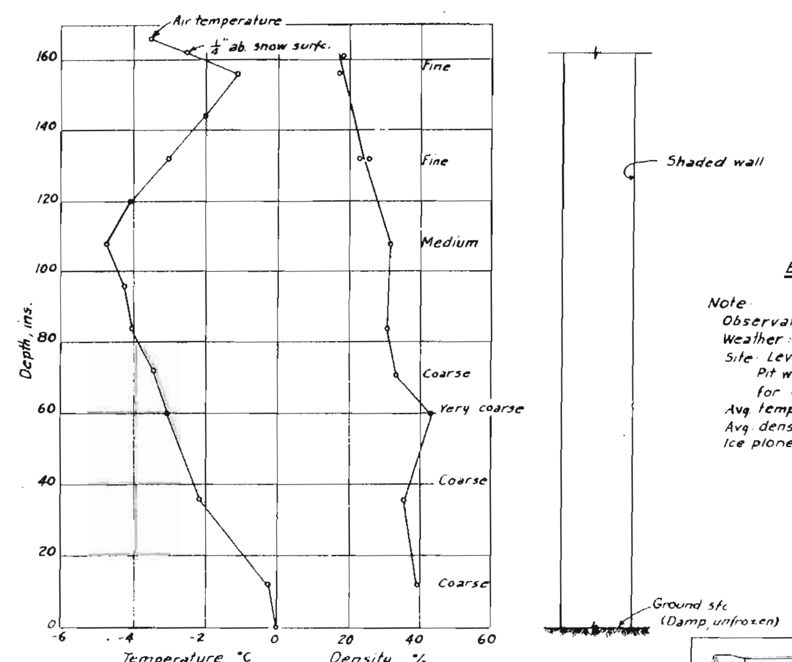
OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

PREPARED PSB	SUBMITTED PSB	TO ACCOMPANY REPORT DATED 30 JUNE 1954
DRAWN MJM	APPROVED SMR	PD-20-25/54



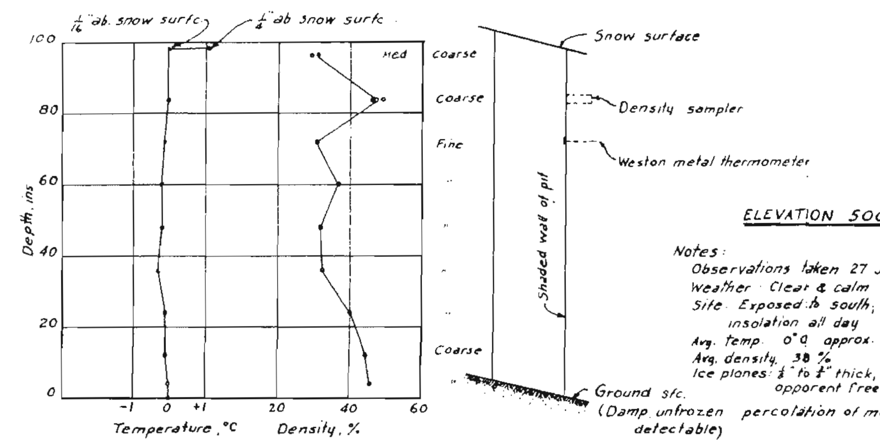
ELEVATION 6900'

Note:
Observations taken 24 Jan. 1952, 1500 to 1600 hrs.
Weather: Continous snow, wind 1-2 mi/hr
New fallen snow density 7.5 to 18.6 % measured
Site: Level, 20' N of HQ lysimeter north wall
Avg. temp. of pack, -3°C
Avg. density 28 %
Ice planes: ⅛" to ¼" thick, not impermeable; no apparent free water



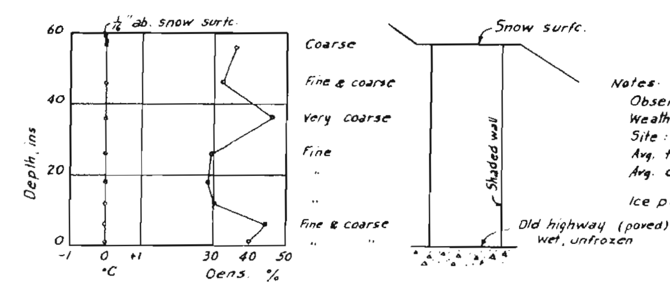
ELEVATION 6000'

Note:
Observations taken 27 Jan. 1952, 1530 to 1700 hrs.
Weather: Clear & calm
Site: Level, exposed to full insolation 0930 to 1600 hrs. Pit was dug on 26 Jan. & new wall exposed for observations on 27 Jan.
Avg. temp. -2.5°C
Avg. density 30 %
Ice planes: ⅛" to ¼" thick, not impermeable; no apparent free water



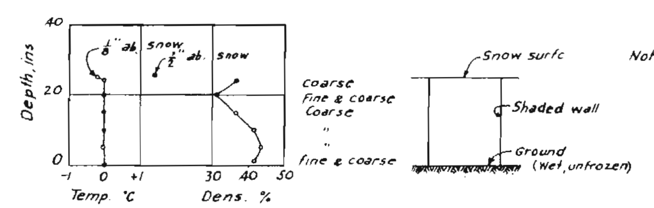
ELEVATION 5000'

Notes:
Observations taken 27 Jan. 1952, 1430 to 1500 hrs
Weather: Clear & calm
Site: Exposed to south; snow sfc. 8% slope, full insolation all day
Avg. temp. 0°C approx.
Avg. density 38 %
Ice planes: ⅛" to ¼" thick, not impermeable; no apparent free water
percolation of melt water was not detectable



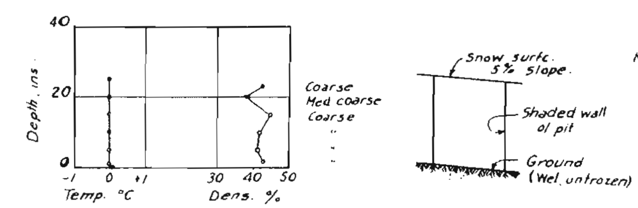
ELEVATION 4000'

Notes:
Observations taken on 27 Jan. 1952, 1150 to 1220 hrs.
Weather: Clear & calm
Site: Level, insolation on snow from 1100 to 1400 hrs.
Avg. temp. 0°C.
Avg. density 36 %. Snow is wet and appears to be at its moisture capacity
Ice planes: ⅛" to ½" thick, disintegrated



ELEVATION 3000'

Notes:
Observations taken on 27 Jan. 1952, 1055 to 1110 hrs
Weather: Clear & calm
Site: Forested area on dirt road, shaded, level
Avg. temp. 0°C
Avg. density 39 %. Snow is wet, and at its moisture capacity.
No visible ice planes.



ELEVATION 3000'

Notes:
Observations taken on 27 Jan. 1952, 1025 to 1040 hrs
Weather: Clear & calm
Site: Open, exposed to full insolation except between 1100 & 1500 hrs
Avg. temp. 0°C
Avg. density 42 %
No ice planes, homogeneous granular snow.
Snow is wet and at its moisture capacity.

Average snow line is approximately at the 3000-foot level.

Classification of the snowpack by visual inspection of the core.

Note: Ice lenses or heavy crusts may not be as numerous or nearly as impervious in forested areas.

Weston bimetallic dial thermometers with stem inserted through slots in the snow tube. The sensitive portion is immersed in snow core but not touching the metal wall of the tube.

Mount Rose snow tube with a core from the snowpack, laid on a shaded snow surface. If a shaded surface is not available, note temperature and the corresponding depth readings as the tube is drawn from the pack and shaded by the observer. Depending on the depth, thermometers may be spaced from 4 to 24 inches apart to obtain the average temperature of the snowpack. Leave tube in snow pack long enough to come to equilibrium with the surrounding snow.

A PROPOSED METHOD OF DETERMINING TEMPERATURE AND STRUCTURAL PROFILE OF THE SNOWPACK WITHOUT DIGGING A PIT.

Note:
Observations were taken along U.S. Highway 40 between Auburn and Soda Springs, California.

SNOW INVESTIGATIONS SUMMARY REPORT

SNOW HYDROLOGY

DENSITY & TEMPERATURE PROFILES FOR EVALUATING SNOWPACK CONDITIONS

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

PREPARED: PAR.-F.M.	SUBMITTED: PAB.	TO ACCOMPANY REPORT DATED 30 JUNE 1956
DRAWN: PAR.-F.M.	APPROVED: D.M.R.	PD-20-25/55

CHAPTER 9 - HYDROGRAPH SYNTHESIS

9-01. INTRODUCTION

9-01.01 General. The foregoing chapters of this report have been concerned with several specialized aspects of snow: the deposition and distribution of the snowpack and methods by which it is measured; the role of snow in the hydrologic cycle; the physical causes and practical indexes of snowmelt; variations in snow cover and methods by which it can be estimated; and the effect of the snowpack on the storage and routing of water. This report is not, however, concerned primarily with the study of snow itself; rather, it is interested in the hydrologic aspects of snow, and the effect snow has on the runoff from basins where snow exists. Consequently, it is only when the separate findings of the previous chapters are considered in relation to their effects upon streamflow that the purpose of this report is realized. There are two aspects to be considered in the problem of determining runoff from snow-covered areas: one is concerned only with the total volume of snowmelt runoff; the other requires that the time distribution of the runoff also be determined. It is the latter aspect that is of concern in this chapter. The problem of forecasting the volume of runoff will be considered separately in chapter 11. In the determination of the time distribution of runoff from snow-covered areas there are, furthermore, two distinct types of hydrograph synthesis involved. One requires that the flow be determined only a few days in advance, current conditions of snow cover and streamflow being known. This type is used in river forecasting and in the operation of reservoirs (see chapter 12). The other requires that the discharge hydrograph for an entire rain-on-snow event or the hydrograph for an entire snowmelt season be determined, with only the initial conditions of streamflow and snow cover being given. This type of hydrograph synthesis is most often used in the development of design floods (see chapter 10). Both types will be dealt with here. Moreover, flood hydrographs resulting both from rain-on-snow events and from snowmelt alone will be considered. These same problems have previously been considered in a report by Snyder 19/ which briefly summarizes much of the work of the Snow Investigations with respect to hydrograph analysis and synthesis. Before examining these specialized aspects of runoff from snow-covered areas, some basic considerations common to all shall first be examined.

9-01.02 Basic considerations. - In any system devised for the synthesis of discharge hydrographs, whether for snow-covered or non-snow-covered areas, there is one paramount consideration: the system must be internally consistent. That is to say, each component of the synthesis must be determined in relation to all the other components. For example, in the determination of losses, the methods by which rainfall and snowmelt were determined play a most important part. An overestimation of either of these must result in a compensating overestimate of losses. While the deliberate use of such compensating errors is not advocated, it must be realized that in a field such as hydrology, where measurements are somewhat inexact to begin with and areal variations in the measured elements make accurate determination impossible, such errors are inherent in the basic data. Recognizing their existence, systems can be worked out to mitigate these effects. Many systems of hydrograph reconstitution currently in use give adequate results even though the magnitudes of some of the factors involved are obviously incorrect. Precipitation amounts may be uncorrected for gage deficiencies and for areal variation, resulting in much too low a total figure. Yet losses, determined as the difference between precipitation and runoff, may here be underestimated, bringing the system into balance. Likewise snowmelt amounts are often underestimated and compensated by an overestimate of the areal extent of the snow cover, the latter being the derived factor which makes agreement between snowmelt and runoff quantities. On the other hand, the fact that such systems can give suitable reconstitutions is not to argue that care should not be taken in determining the variables. The chief danger in any system in which the basic data are not a true representation of the physical facts is that the system may be applied to data outside the range for which it was developed. This extrapolation may produce results which are no longer rational: negative losses or greater than 100 percent snow cover, for example, may be required to bring the system into balance. Thus, while any system of hydrograph synthesis should be consistent within itself, it should at the same time be rational. All variables should be estimated as closely as possible and in a manner such that a balance is possible without undue juggling of the data.

9-01.03 There are two general situations in which snow has an important effect upon streamflow. One is the discharge hydrograph that results from snowmelt alone or from snowmelt abetted by small and scattered amounts of rainfall. The snowmelt may extend over a period of several months, as it does in the mountainous drainage basins of the western United States, melting at moderate rates with only part of the total drainage basin contributing at any one time, or it may last for only a few days, as is generally the case in the Great Plains area, and be characterized by more intense rates and basinwide melt. The other general situation where snow plays an important role is where an intense rain storm falls on a snow-covered area. Here the rain may be abetted by melting snow, thereby increasing its effect, or, on the other hand, it may be partially stored or detained by the snow cover, mitigating its effect. Both situations will be considered in this chapter.

9-02. GENERAL APPROACH

9-02.01 Elevation effects. - The synthesis of hydrographs which result from snowmelt or from rain-on-snow differ in several respects from those resulting from rain alone. For one thing, a drainage basin cannot be considered simply as a homogeneous unit; the areal extent of the snow cover is involved. In snowmelt floods this limits the contributing area; in rain-on-snow floods, loss rates may differ markedly between the snow-free and snow-covered areas. Because the snowpack exhibits its principal variation with elevation (see chapter 3), it becomes necessary to include, in some manner or another, elevation effects in any system of hydrograph synthesis for basins in which snow is a factor and which have a sufficient range in elevation to warrant it. This consideration of elevation is also pertinent in the determination of form and intensity of precipitation--a problem not limited to snow-covered areas but frequently involved in rain floods in general. As was shown in chapter 3, the intensity of precipitation generally increases with increasing elevation, and it is more frequently in the form of snow at higher elevations that it is at lower elevations. Then too, snowmelt rates tend to decrease with increasing elevation as is subsequently discussed. In view of all this, the importance of elevation effects in the synthesis of runoff hydrographs may readily be seen.

9-02.02 There are two general approaches to the problem of computing the runoff from snow-covered areas. They differ in the manner in which elevation effects are handled. The first method to be considered consists simply of dividing the drainage basin into bands of equal elevation and computing the snowmelt, rainfall, and losses for each such elevation band separately; the net runoff from all bands is then combined to arrive at the net basin runoff. Such a method was used in the synthesis of discharge hydrographs for the Columbia River basin above McNary Dam. 8/ Sufficient bands should be selected so as to smooth any incremental changes, as no finer sub-division is made than the elevation band as a whole. The entire band is considered to be snow-covered or not, the entire band to be effectively melting or not, etc. Bands may be determined either on the basis of equal increments of elevation or on the basis of equal areas. The former method is advantageous from the standpoint of melt and form-of-precipitation computations, since the temperature decrease with elevation may then be assumed to be uniform between bands. For this reason it is the more commonly used method; however, the use of equal areas is superior in almost all other respects. With the exception of the highest and lowest bands, however, both methods are generally similar due to the usual approximate linear relationship between area and elevation for intermediate elevations.

9-02.03 The other general approach to determining runoff from snow-covered areas is to treat the basin as a unit, making corrections for the non-snow-covered area and other non-contributing areas (e.g., areas of no snowmelt and areas having precipitation in the form of snow during rain-on-snow storms). In this approach the assumption is usually made that the basin snow cover is depleted in a regular manner with elevation: the lowest portions of the basin are the first to become bare and the snow-cover depletion with time progresses regularly upwards. In the situation where low-elevation melt is taking place but no melt occurs at the higher elevations within the basin, even though they are snow-covered, only the area in the band between the "snowline" (average elevation of the lower limit of the snow-covered area) and the "melt line" (average elevation of the upper limit of snowmelt) contributes to to snowmelt runoff. In the case of rain-on-snow, the contributing area for rainfall runoff has an upper limit at the elevation where the rain becomes snow. There may be three distinct elevation bands marked by different situations in rain-on-snow events: (1) an upper band where snow falls on a snowpack; (2) an intermediate band where rain falls on the snowpack; and (3) a lower band where rain falls on bare ground. The deviations of these bands may change with time making for a complex situation. This situation will be considered further in section 9-03.

9-02.04 Areal effects. - In the foregoing the basin has been considered as a unit in an areal sense. Even when broken up into elevation bands, portions of the basin having similar elevations were considered to have the same snowmelt, precipitation, snow cover, and losses. This is not always a good approximation, especially in the case of large basins covering a range of climatic factors. Temperatures may be considerably warmer, the snowline higher and the precipitation less for a given elevation in one part of such a large basin than in another. In such basins it sometimes becomes necessary to consider separately the several sub-areas having different characteristics. Separate hydrograph reconstitutions may be made for each sub-area and these combined by proper streamflow routing techniques. Such was the approach used in the design flood determinations for the Gila River basin above Painted Rock Damsite in Arizona, an area of some 50,000 square miles. 3/ On the other hand, it is sometimes possible to treat such large basins as a unit, as was done for the entire Columbia River drainage above McNary Dam. 8/ By using average values of the several variables which include an areal sampling it is possible to arrive at mean runoff values which are adequate. This is possible since the large basin itself effectively averages out its extremes.

9-02.05 Melt period. - Another important general consideration in the reconstitution of streamflow hydrographs for snow-covered areas is the melt period, and hence the routing interval, selected. Since snowmelt is diurnal in character, daily melt amounts are customarily computed for all but the smallest, flashiest basins. However, as has been domonstrated in chapter 5, the daily snowmelt quantity is usually generated in something less than one-half day, the remainder of the day being a period of heat loss. In large basins where the runoff is relatively sluggish and no regular diurnal pattern is discernible in the discharge hydrograph, this fact is of small consequence. Daily melts may be routed using a daily time interval to get the resulting discharge hydrograph. For smaller, hydrologically-faster basins having a regular diurnal snowmelt runoff pattern, such an approach is not always adequate. Yet even here if it is not desired to reconstitute the diurnal fluctuation but only to reconstitute mean daily flows, a daily routing interval is adequate. On the other hand, where it is desired to reconstitute the diurnal rise and fall of the stream, routing intervals of twelve or eight hours may be used. All the melt is attributed to one of the periods, the other(s) contributing nothing to runoff except during periods of rainfall. In the case of rainfall, the shorter routing intervals (less than one day) are generally better suited to the reconstitution of hydrographs because of the possible extreme variation in precipitation rates with time. Such data are not always available, however. For large basins daily rainfall amounts are usually adequate.

9-03. RAIN-ON-SNOW FLOOD HYDROGRAPHS

9-03.01 General. - The methods used in the synthesis of flood hydrographs which are predominantly the result of rain-on-snow are similar to those which are used for rain floods in general. In consequence of the high rates of rainfall encountered in these situations, only direct runoff is explicitly considered; base flow rates are usually estimates since they contribute relatively little to the flow at or near the time of peak flow. Also, the relatively high rate of rainfall makes it the primary variable in the analysis; the added increment of snowmelt is considered more as an additive factor than as a primary variable. In rain-on-snow situations, the effects of the snowpack are twofold: (1) to add an increment of melt water to the rainfall and (2) to store and detain, in varying degrees, the melt and rain water generated. It is the latter effect that makes the reconstitution of rain-on-snow floods most complex. Rain falling on a snowpack may be stored by the pack or pass through without depletion, depending upon the condition of the pack. A considerable quantity of rain water may be stored by a dry, sub-freezing, snowpack. Moreover, a deep snowpack that has previously experienced little or no melt or rainfall of consequence may add an additional increment of storage by virtue of its delaying effect upon runoff. Impenetrable ice planes within the snowpack may give a large horizontal component to the flow of water through the pack itself (along the ice planes seeking a pervious area). Then too, water may be perched above such impenetrable layers. In addition, such a snowpack effectively chokes or dams the natural surface drainage channels to high rates of flow. Thus a sudden occurrence of heavy rainfall on such a pack is quite effectively retarded. Delays as long as two days have been noted in situations where rain fell on such a snowpack. (This storage action of the snowpack is discussed in detail in chapter 8.) On the other hand, the snowpack may be quite pervious to the high rates of rainfall. Prior rains or melt may have caused it to become isothermal at 32°F, may have satisfied its liquid-water-holding capacity, may have established percolation paths through the pack, and may have melted and scoured out adequate surface drainage channels beneath the snowpack. If such be the case, not only is there practically no delay in runoff as a result of the snowpack, but the snowpack may abet runoff by adding an increment of melt water. Even more important, its melt may have maintained the soil moisture and depression storage so that even those usual losses are reduced. Between the two extreme conditions cited above range an indefinite number of intermediate conditions. From the foregoing discussion, one important fact stands out: the condition of the snowpack has a dominant effect upon the initial basin discharge of a rain-on-snow event. Because of this, rain-on-snow floods are difficult to synthesize; some knowledge of the initial condition of the snowpack is mandatory.

9-03.02 Because of the general similarity of methods employed in the synthesis of rain-on-snow floods and rain floods in general, no actual synthesis of a rain-on-snow event will be made. Examples of some outstanding occurrences will be given, followed by a discussion of the principal factors involved in such events. Some problems pertinent to rain floods in general will be considered as well as those peculiar to rain-on-snow flood events.

9-03.03 Examples. - Two outstanding examples of rain-on-snow floods occurred during the period of operation of CSSL, affording an excellent opportunity for study. One, actually a series of separate flood events, occurred in November-December 1950. Intense rains falling on a relatively shallow snowpack were abetted by melting snow. A record peak discharge estimated at 1200 cfs was observed from this small, 4-square mile basin. Except for two of the flood waves (in the series of five) which were preceded by new-fallen snow, little of the rain or melt water was stored by the snowpack; its action was rather to add to runoff by melt and by decreasing basin losses. Data on this flood series may be found in the Hydrometeorological Log for the 1950-1951 water year at CSSL. Analyses of the event at CSSL were made in Technical Bulletin 14 and in Snow Investigations Miscellaneous Report 3. A reconstitution of this same flood event in the American River basin, California (adjacent to the CSSL area), is contained in the previously cited report by Snyder 19/. The other outstanding rain-on-snow flood at CSSL occurred in January 1953; it is described in Research Note 18. Here a considerable amount of rain fell on a moderately deep and cold snowpack. About 12 percent of the water supply was lost to runoff as a result of the snowpack and the runoff was delayed due to the damming and channel-choking action of the snowpack. Unfortunately, the flume used to gage runoff from the CSSL basin area was also choked by snow and no discharge record was available for the basin. A detailed analysis of the runoff from the headquarters lysimeter was made (see Research Note 18) to investigate the storing and delaying action of the snowpack. The results of this analysis are given in chapter 8.

9-03.04 WBSL was established primarily for the express purpose of obtaining data on rain-on-snow events; however, many of the storms for that laboratory involved both (1) rain and snow falling simultaneously at different elevations, and (2) partial basin snow covers, adding sufficient complications to make most cases not readily amenable to analysis. During February of 1951, however, a sequence of storms occurred in which the precipitation was almost entirely in the form of rain; moreover, the basin was virtually 100 percent snow covered. An analysis was made of this sequence in Research Note 24; the results of this analysis are presented in plates 9-1 and 9-2. No actual reconstitution of the discharge hydrograph was made; rather a recession analysis was made of the hydrograph, separating periods of significant change and determining the water generated during the periods. Ten periods were thus defined (see Fig. 1, plate 9-1). Estimates of the water generated during each period were made by the techniques subsequently discussed in this section, and these amounts were compared with those from the recession analyses (see fig. 3, plate 9-2).

9-03.05 Rainfall. - It is not the purpose of this paragraph to describe methods by which basinwide rainfall amounts may be computed. That has been done elsewhere in this report (see chapter 4). The principal concern here is in the manner in which rainfall amounts, once computed may be fitted into a comprehensive scheme for determining the resulting runoff. A few remarks regarding the determination of basin rainfall are fitting, however. It is often the case that inadequate consideration is given to the form of the precipitation; in many storms where rain is falling at the observation sites, the precipitation at the higher elevations is in the form of snow. This fact is often ignored and all precipitation is considered to be of the same form observed at the precipitation stations. A study has been made relating the form of precipitation to surface air temperature (see 3-02.03), and the data are presented diagrammatically in figure 1 of plate 3-1. It will be noted that with a surface air temperature of 34°F, the frequency of occurrence of snow is greater than is the frequency of rain; with an air temperature of 35°F the frequency of rain is greater than that of snow. Using this dividing line between rain and snow and a standard lapse rate of 3°F per 1000 feet (or, better, the pseudo-adiabatic lapse rate, since precipitation is occurring), it is possible to estimate surface air temperatures and hence form of precipitation at different elevations within the basin from the temperature stations usually found at the lower elevations. If the basin is sub-divided into elevation bands, it is a simple matter to estimate the temperature at the mean elevation of each band and hence the form of precipitation for the band. If the basin is being treated as a unit, corrections must be made for the portion of the precipitation that occurs in the form of snow.

9-03.06 A correction for the variation of precipitation with elevation is a refinement that is seldom made in studies of rainfall runoff for snow-covered areas or otherwise. This is mainly because its inclusion in any scheme of hydrograph synthesis complicates it greatly while making little difference in the results. Usually basin rainfall may be determined by using a fixed ratio between it and the rainfall at some index station(s). However, when the precipitation changes from one form to another over large portions of the drainage basin and deficiencies in loss rates between snow-covered and snow-free ground also are important in the synthesis, (also an elevation function since snow cover varies with elevation), it may be that some consideration should be given to the variation of precipitation with elevation. The only practical method of accomplishing this is to use the elevation-band method and to relate normal precipitation in each band to the normal basin precipitation or directly to the normal for the precipitation index stations, thus determining a factor which then may be used to determine storm precipitation amounts in each band.

9-03.07 Snowmelt. - The computation of snowmelt during periods of significant rainfall is a problem quite different from the computation of melt during non-rain periods. Because of the generally overcast conditions, solar radiation has but a minor role in the melt scheme; longwave radiation losses are small, and at times there is even a net heat gain from this source. Because of the turbulent conditions which usually accompany rainstorms, convection and condensation melts are relatively large. In addition, fairly high vapor pressures result from the high relative humidities encountered in this situation, tending further to increase condensation melt. This problem was considered at some length in chapter 6 and an equation was developed for the computation of snowmelt during periods of rainfall. It is repeated here.

$$M = (T_a - 32)(0.029 + 0.0084kv + 0.007P_r) + 0.09 \qquad (9\text{-}1)$$

where M is the total daily snowmelt in inches for open or partly-forested basins, T_a is the mean daily air temperature at the 10-foot level in degrees F, v is the mean daily wind speed at the 50-foot level in mph, P_r is the total daily rainfall in inches, and k is the basin constant expressing its exposure to wind (see par 6-04.13). In application, this equation may be further simplified by several assumptions. For one, the variation of melt with wind speed may be ignored by considering the wind to be constant. This assumption is especially suited to areas of heavy forest cover where the 50-foot level wind is relatively light and constant. Assuming, for example, as in chapter 6, a mean wind speed of about 5 mph, the foregoing equation becomes:

$$M = (T_a - 32)(0.074 + 0.007\, P_r) + 0.05 \qquad (9\text{-}2)$$

the convection-condensation term being combined with the term representing long-wave radiation.

9-03.08 Snowmelt, of course, occurs only over the snow-covered portions of the drainage basin. Moreover, it is possible that no melt may be occurring at the higher elevations within the basin at the same time the pack is melting at lower elevations. The resulting intermediate contributing area varies in time with both snow cover and temperature. As with rainfall, melt may be computed by elevation bands or by considering the basin as a unit and making corrections for non-contributing areas. A simple assumption which may be used in the synthesis of rain-on-snow events is to consider the melt as being uniform over the entire snow-covered area on which rain is falling (area having temperatures in excess of 34°F) and to assume no melt in areas over which snow is falling. Further refinement than this is seldom warranted in rain-on-snow situations since snowmelt is usually a relatively small contribution to the total storm runoff, particularly for design floods.

9-03.09 Losses. - Losses in rain-on-snow situations consist not only of water permanently lost to runoff by evapotranspiration, deep percolation, etc., but also of water that is detained both by the snowpack and by the basin itself. This is the more usual connotation of the term losses as used in rain storms in general, where the term losses implies water lost to direct runoff. It is distinct from the use of the term in the synthesis or reconstitution of hydrographs from snowmelt alone, where only water that is permanently lost to runoff is included in the term (as will be discussed subsequently). Since in the synthesis of rain-on-snow flood hydrographs, as with all rain flood hydrographs, the high rates of water generated result in high peak flows from direct runoff, the water discharged more slowly over a longer period by base flow is of little concern in the hydrograph synthesis, even though a considerable volume of water may be included in the base flow. Hence the water which goes to make up the base flow is considered to be a part of the losses.

9-03.10 Of primary importance in rain-on-snow events are the initial losses; that is, non-recurrent losses that need be satisfied only once at the start of a given rain-on-snow event. These losses may be further categorized as those which result from the basin itself. The former includes water which may be frozen within a sub-freezing snowpack, water which may be permanently held by capillarity and absorption within the pack, and water perched above impermeable ice planes or otherwise dammed by the snowpack. As was previously mentioned, it is the extreme variability of this loss that makes the reconstitution of rain-on-snow flood hydrographs most difficult. Such a loss may be non-existent in the case of saturated snowpack, isothermal at 32°F, or may amount to several inches of water in the case of a deep and cold snowpack. Moreover, this snowpack loss may be compounded by basin losses. Once snowpack losses are satisfied, in the case of a sub-freezing snowpack, it still may be necessary to supply water to soil moisture and depression storage before appreciable direct runoff occurs. Beneath a melting snowpack, on the other hand, soil moisture and despression storage losses are generally satisfied, so that rain falling on such a pack not only suffers no losses within the snowpack but losses that usually exist even in snow-free areas are non-existent. It is because of this possible extreme variation in initial losses that a knowledge of pre-existing conditions of the snowpack and the basin is required in the synthesis or reconstitution of discharge hydrographs resulting from rain-on-snow. Methods of determining losses within the snowpack were discussed in chapter 8; other initial basin losses are considered in chapter 4. (Reference is made to Research Notes 4, 18, and 24 and to Technical Bulletin 17 for further and more detailed discussion of the storage effect of the snow cover.)

9-03.11 After the initial basin losses are satisfied, still other losses to direct runoff continue; however, these losses are relatively constant in time and easy to evaluate. Evapotranspiration removes some water from the soil, thus allowing further soil moisture loss. Some water percolates downward through the soil to the ground-water level to be discharged very slowly, for the most part subsequent to the period of interest. Some water merely takes devious routes through the ground (inter-flow) not reaching the stream channel until after the period of interest. Yet all these are considered losses in the sense in which the term is applied to rain-flood hydrograph reconstitutions. It is to be pointed out that there is a difference between the losses previously enumerated and the ultimate disposition of the water involved. Water initially lost in the snowpack is released with the melting of the snow; water initially lost to depression storage may subsequently add to the soil moisture and to the ground-water storage, other water taking its place in filling the depressions; water held in soil moisture is, in turn, evaporated and transpired, mostly after the cessation of the storm event; and water which percolates to the ground-water table is discharged slowly over a long period. Methods of determining basin losses are many and varied and shall not be gone into here. Reference is made to an article by Snyder 17/ and to standard hydrology text books (e.g., Applied Hydrology 15/ and Hydrology Handbook 1/) for a discussion of those losses and methods by which they are included in the over-all routing procedure. When losses are deducted from the total water generated (rainfall plus snowmelt) the residual is termed water excess. Methods by which this water excess is distributed to determine the resulting discharge hydrograph will now be considered.

9-03.12 Time distribution of runoff. - The incremental quantities of water excess (water generated minus losses), determined as explained in the preceding paragraphs of this section, are translated into the resultant discharge hydrograph by any of the standard methods used in conjunction with rain floods in general. Unit hydrographs (or distribution graphs) and methods of storage routing are used. It is to be emphasized that here, as with all rain floods, the water excess routed consists only of that water that reaches the stream channels by the more direct routes and results in the high peak flow characteristic of rain floods. Water travelling by indirect routes is included in a base flow curve which is generally estimated. This approach is familiar to hydrologists and will not be considered further here. Reference is made to the manual, "Flood Control" 12/ and to chapter 5 (Flood-hydrograph analyses and computations 4/) and chapter 8 (Routing of floods through river channels 5/) of the Engineering Manual for Civil Works Construction for descriptions of standard methods of flood-hydrograph determination.

9-03.13 A few general remarks regarding the routing interval to be used in determining rain-on-snow flood hydrographs follow. Since it is the rainfall that is of primary concern here, and not the snowmelt, the availability of rainfall data should determine the time periods selected. Thus if 6-hourly rainfall data are available from Weather Bureau stations, the routing interval should be made to correspond to these times of measurement. If only daily data are available from cooperative observers and these readings are made during the morning or evening, as is customary, then the routing period must agree with these measurements. Only where hourly precipitation data are available may the routing interval be selected as desired. Here it is usually advantageous to subdivide the day into a fixed number of even periods beginning at midnight (e.g., 00-08, 08-16, 16-24 hours). The length of the periods selected should be short enough to adequately define the rapid rise and fall of the hydrograph yet should not be so short as to require unnecessarily detailed computations. Computations of snowmelt are made to agree with the periods selected for the rainfall determinations.

9-03.14 A good example of the synthesis of a rain-on-snow event is contained in Design Memorandum No. 2 for Cougar Dam and Reservoir in Oregon 9/. The syntheses of standard project and maximum probable floods given therein make use of elevation bands and maintain inventories of snowpack water equivalent in each band to determine the areal extent of the snow cover. Snowmelt computations are based upon the equation given in this section. Plate 10-1 gives the results of the Standard Project Flood derivation; it is discussed further in chapter 10.

9-04. SPRING SNOWMELT FLOOD HYDROGRAPHS

9-04.01 General. - So far this chapter has been concerned with the synthesis and reconstitution of floods that result from rain falling on snow. As was previously pointed out, these flood events are for the most part like any other rain floods with only the added effects of the storage and the melt of the snowpack. In this section the synthesis of springtime snowmelt floods is to be considered. The methods used in the synthesis of these floods are markedly different from those used in rain-on-snow floods. Special techniques quite different from those used in rain or rain-on-snow flood synthesis are required in their synthesis. The Boise River basin above Twin Springs, Idaho will be used as an example to illustrate the points raised herein. This basin was selected as being typical of areas in which springtime snowmelt is of major importance. In chapter 6 snowmelt indexes were determined for this basin, one of which will be used to compute snowmelt for the sample presented here. In chapter 7 basin snow-cover relationships for the Boise River basin were determined which will also be used here. Reference is made to two other studies of snowmelt runoff for this same general area, by Summersett 20/ and Zimmerman 21/ and to the Definite Project Report on Lucky Peak Dam, on the Boise River, Idaho, 11/ which are of interest to what follows. The 1955 melt season as a whole will be reconstituted on the basis of mean daily flows; also forecasts of mean daily snowmelt runoff, one, two, and three days in advance will be made throughout the season. In addition, reconstitutions will be made showing the diurnal variation of the snowmelt runoff for a portion of the season.

9-04.02 In the mountainous areas of the western United States, the annual spring snowmelt flood results more from the sustained melting of deep snow packs over a long period of time than it does from high rates of generation of water. Continued melting over a period of a month or two results in a piling up of runoff as a result of slow recession until relatively high flood flows result even from the moderate

rate at which snow melts. For example, during the 1952 spring snowmelt season at CSSL, a peak discharge of 306 cfs was attained, which amounts to a flow of 77 cfs/sq.mi., for this small, 4-square-mile basin. Reference is made to Technical Bulletins 4, 8, 9, and 10 for some early studies of the synthesis of mountain snowmelt hydrographs at CSSL. In most areas rain usually adds an additional increment to the basin discharge; however, it is of secondary importance to the snowmelt in these situations. While what follows in this section is concerned mainly with the spring snowmelt floods which result from the deep snow packs of the mountainous areas of the western United States, it is also generally applicable to other areas and other flood situations which are primarily the result of snowmelt, providing allowance is made for characteristics peculiar to the area. The northern Great Plains area, for example, often experiences large floods early in the spring from the comparatively rapid and simultaneous melting of the extensive but relatively shallow snow cover of that area. Elevation effects are nonexistent here, as are forest cover effects. Runoff from the area is also frequently affected by frozen soil. These characteristics of snowmelt in the plains area will be considered later in this section.

9-04.03 Three approaches to the problem of synthesizing springtime snowmelt hydrographs from deep mountain snowpacks will be considered: (1) a rational approach wherein snowmelt, rainfall, snow cover, losses, etc., are evaluated separately and combined to arrive at the resulting discharge hydrograph; (2) the elevation-band method where separate computations of all elements are repeated for each band; and (3) a one-step method whereby snowmelt runoff is computed directly from a single diagram relating a temperature index and areal snow cover to snowmelt runoff. Other approaches using different combinations are of course also possible. Before considering these approaches, however, a discussion of some of the components involved is in order.

9-04.04 Snowmelt. - Basic to the synthesis of snowmelt hydrographs is the computation of snowmelt itself. Snowmelt may be computed by any of the methods previously discussed in chapters 5 and 6. For design floods the thermal-budget approach may best be used, while for operational forecasting some index method is probably better suited to the purpose. In chapter 6 analyses were made of snowmelt in the Boise River basin and thermal-budget indexes were developed relating snowmelt runoff to the parameters of air temperature, vapor pressure and radiation. One of these relationships is used in a reconstitution to be made later in this chapter. It is repeated below:

$$M = 0.0267T_{max} + 0.00227G - 2.00 \qquad (9\text{-}3)$$

where M is the snowmelt runoff for Boise River above Twin Springs, Idaho, in in./day, T_{max} is the daily maximum temperature at Idaho City, and G is the net radiation absorbed by the snowpack (including both absorbed shortwave radiation and net longwave radiation loss). Temperature indexes are the most widely used method of computing snowmelt and snowmelt runoff. They are used in the elevation-band and the one-step methods described herein. A few remarks concerning them follow.

9-04.05 In the computation of basinwide snowmelt or snowmelt runoff by temperature indexes, the temperature is usually adjusted to the mean elevation of the contributing area (see chapter 6). This presupposes a direct variation in melt rate with temperature and hence with elevation within a given basin. Now, from the thermal budget considerations of chapter 5, it would seem this implied decrease in melt rate with elevation is not strictly valid. While it is true air temperature and vapor pressure tend to decrease with increasing elevation, thus tending to reduce convection-condensation melt with increasing elevation, these melt components are but a part of the total melt. Solar radiation, the single most important source of heat in melting snow, tends to increase with increasing elevation due to the lesser scattering and absorption by the air at higher elevations. Yet a comparison of snowmelt rates at different elevations over a period of time indicates generally less melt at the higher elevations, especially early in the melt season. A partial explanation of these apparently contradictory statements may be had when the variation of albedo with elevation is taken into account. As a result of the greater frequency of new snowfalls at higher elevations of a basin, there is an increase in the mean albedo with elevation; it is primarily in consequence of this higher albedo at high elevations that the melt is reduced rather than as a direct result of the decreased air temperature. (Over extreme ranges, air temperature itself certainly also has an effect. Very little melt occurs with marked sub-freezing temperatures.) Thus for periods having no new snowfall and late enough in the melt season to assure a ripe snowpack of uniform albedo throughout a drainage basin, melt is largely independent of elevation. Practically, however, for the melt season as a whole, the delay in ripening, the higher albedo of the higher-level snow, and the decrease in air temperature and vapor pressure with elevation all result in decreasing melt rates with increasing elevation. This is allowed for in temperature-index melt computations by the lapse rate correction which thus represents average climatic characteristics more than an actual physical relationship between air temperature and snowmelt.

9-04.06 In the computation of snowmelt and snowmelt runoff for use in hydrograph synthesis and in short-term forecasts of runoff, a high degree of accuracy is not ordinarily essential. Random errors which result from the imperfect relationship between various indexes and snowmelt are of small consequence. Since, in hydrograph synthesis, a given day's observed runoff is the integrated result of several days' snowmelt, these random errors tend to compensate one another. It is important, however, that no consistent errors or errors of bias occur in the computation of melt amounts. Melt amounts consistently too high or consistently too low cannot be self compensating. While adjustments in loss rates or areal snow-cover amounts can correct for incorrect melt rates (as was previously discussed) such a procedure is not recommended. It is better that each factor in a snowmelt snythesis be a rational approximation of the actual event as it occurs in nature. Still more important, the relationship between snowmelt and the snowmelt index used should be consistent for all ranges of melt. An index that gives, say, too high estimates of melt at low rates of melt and too low estimates at high rates of melt is not satisfactory. Assuming no such errors of bias to exist, random errors do not seriously affect either seasonal hydrograph synthesis or short-term forecasts.

9-04.07 Units. - Snowmelt may be expressed in units of acre-feet, day-second-feet, inches or other measure of volume. If expressed in inches, it may be given in terms of either inches over the basin or inches over the snow-covered area. Because of the variation in snow-covered area, melt rates are usually given in terms of the latter; however, the computation of the discharge hydrograph requires that inches of melt be given for the basin as a whole. The use of inches over the basin is convenient in that it makes it possible to readily combine snowmelt and rainfall amounts (which are ordinarily given in those units) and to deduct losses. In addition, this unit is easy to visualize and to compare even for basins of markedly different size. For these reasons, it is used here and elsewhere in this report. The use of day-second-feet, while convenient in routing the generated runoff, is not amenable to the several other steps in the reconstitution.

9-04.08 Snow-cover depletion. - The actual areal extent of the snow cover is often omitted in calculations of snowmelt runoff, the effects of varying cover being integrated with other factors in establishing the relationship between melt rates and snowmelt runoff. In other cases it is included as a derived factor relating point melts to basinwide snowmelt or snowmelt runoff. Such practices are undesirable; in keeping with the objectives of this report--that only indexes which logically explain the physical phenomena be used--the effect of each important factor should be evaluated rationally wherever possible. In chapter 7, methods whereby the areal extent of the snow cover could be estimated were given. In addition, the depletion of the snow cover was related to other variables so that the areal cover could be estimated throughout the season. Two distinct problems exist in the determination of areal snow cover as a result of the different requirements of seasonal hydrograph reconstitutions and short-term forecasts of runoff. Seasonal reconstitutions require that, starting with a known initial snow cover (and depth-elevation characteristics), the increase and decrease in extent of cover for the remainder of the season be obtainable from regularly available hydrologic data. Short-term forecasts generally require only few determinations of areal snow cover from observed data and require little, if any, interpolation of snow-cover data. When the elevation-band approach to hydrograph synthesis is used, it is possible to maintain an inventory of the mean snowpack water equivalent in each band, thereby determining when the band becomes bare of snow. All new accretions of snow are added as well as melt amounts subtracted. This approach is especially amenable to design flood computations; it will be discussed further in a subsequent paragraph.

9-04.09 Rainfall. - Since there are few areas that do not have some rain during the spring snowmelt season, the inclusion of rainfall in spring snowmelt flood hydrograph synthesis is usually necessary. The special problems encountered in the determination of basinwide rainfall amounts in mountainous areas have already been discussed previously (see paragraph 9-03.05) and little need be added here. Differences in the form of the precipitation (rain or snow) must be considered here also. In addition, rainfall on bare ground and rainfall on the snowpack must be treated separately. In the case of springtime snowmelt, it is likely that a considerable portion of the basin (at the lower elevations) will be snow free, the portion increasing as the melt season progresses. In general, only the zone having rain-on-snow contributes appreciably to runoff, with both snowmelt and rainfall runoff occurring in this zone. Assuming, as is explained in the following paragraph, that all rain that falls on bare ground is lost, there is no contribution to runoff from the lowest zone. Moreover, since little or no melt occurs at elevations where snow is falling, the contribution from the top zone is practically nil.

9-04.10 Losses. - The term losses, as used here, means permanent losses, that is, water which will never show up at the gaging station. This differs from the use of the term with respect to rainfall hydrograph studies where losses are usually considered to consist of all water that does not show up directly as runoff. The reason for this difference is that in the analysis of rainfall flood hydrographs much of the runoff is found to be direct runoff. This runoff, being quickly discharged from the drainage basin, produces the sudden rise and sharp peak characteristic of most rain floods. While a considerable portion of the total rainfall runoff is also discharged by more indirect routes, this flow is relatively sluggish and is hence distributed over a long time interval. As a result, these rates of flow are very low compared to those of the direct runoff and, consequently, the "base flow" is merely estimated in rain-flood hydrograph analysis. Since at the time of peak flow, the base flow component is usually less than 10 percent of the total discharge, it is evident that even large errors in its estimation have little effect on the final result. No strict accounting is kept of the rainfall that goes into losses and the water that is subsequently discharged as base flow. In the analysis of snowmelt flood hydrographs, on the other hand, such an approach is not applicable. As a result of the comparatively low rates of snowmelt(as compared to rainfall), almost all of the snowmelt would go into losses should such a rain-flood hydrograph procedure be used. Little or no direct runoff would remain to produce a hydrograph peak; there would be only a base flow curve to be estimated. Obviously, snowmelt hydrograph syntheses require a different approach from those used for rain floods.

9-04.11 Since sub-surface flow is more important in snowmelt flood hydrographs than in rain-flood hydrographs, all water not permanently lost to runoff must be routed. Permanent losses consist basically of evapotranspiration and deep percolation. In addition, however, some melt water is lost to soil moisture recharge and depression storage; it is eventually disposed of by evapotranspiration and deep percolation occurring subsequent to the melt season; hence it too is a permanent loss. The losses due to soil-moisture recharge and depression storage are not ordinarily recurring losses but constitute an initial loss at the beginning of the melt season only. After the initial loss is satisfied, only losses due to evapotranspiration and deep percolation remain. For this reason snowmelt losses are relatively simple to estimate when compared to losses from rain storms. Once the initial loss is satisfied at the start of the snowmelt season--soil moisture recharge and depressions filled--an equilibrium is reached between water available and losses. This condition prevails throughout the melt season. Since evapotranspiration is somewhat proportional to heat supply and deep percolation to available water (see chapter 4), losses may conveniently be included in the snowmelt index. When an index of snowmelt runoff is computed, it is assumed that losses are a fixed percentage of the heat supply and hence of the snowmelt. One fault in assuming a fixed percentage loss for snowmelt floods becomes apparent from a consideration of extreme melt rates. At extremely low melt rates all melt may be lost; on the other hand, it is unreasonable to suppose losses to keep increasing directly with melt rates up to their highest possible values. An alternative method (to a fixed percentage loss) has been to use a constant loss rate for all ranges of melt. Here some critical snowmelt rate is required before any snowmelt runoff is realized and the greater the melt rate the higher is the percentage runoff. Still more logical than either of the foregoing is use of a curve combining the best features of the two: A fixed loss rate that must be exceeded by supply before any direct runoff results, with losses then increasing with increasing water generated until maximum loss rate is attained, beyond which losses are constant regardless of increasing water generated.

9-04.12 Generated runoff. - Many different approaches have been used by hydrologists concerned with snowmelt runoff for determining the combined effect of snowmelt, snow cover, rainfall, losses etc., on runoff. As previously mentioned, these may be grouped into three general methods for determination of generated runoff: (1) Method A - a rational method wherein the basin is considered as a unit and adjustments are made for variations in melt rates, snow cover, form of precipitation, and losses; (2) Method B - the elevation-band method where the basin is subdivided into elevation bands, each band being considered to have a uniform melt rate, be entirely snow covered or bare, have the same form and intensity of precipitation throughout, and have uniform losses; and (3) Method C - a one-step method whereby generated runoff is determined directly from temperature index and snow-cover data. Each of these methods have something to recommend it; they will be discussed in detail in what follows. To illustrate the methods, the drainage basin of the Boise River basin above Twin Springs, Idaho is used as an example. Particulars regarding each of these methods follow.

9-04.13 Method A. - In this method the basin is treated as a unit; melt is computed in inches over the snow-covered area, to which is added the mean basin precipitation in inches. The sum is then multiplied by the contributing area (the percentage of snow-covered area below the freezing level) to arrive at the total water generated in inches over the basin. Losses are then deducted to arrive at the water excess. The device of combining snowmelt and rainfall before multiplying by the percentage contributing area presupposes that all rain falling on bare ground is lost and that the dividing line between rain and snow is at the freezing level. Both of these are approximations that are sufficiently accurate for most areas of spring snowmelt providing no considerable amount of rain is involved. While this method is essentially simple, there are a few complicating factors which should be pointed out. The temperature index is ordinarily corrected to the median elevation of the contributing area. This may be done handily with the aid of figure 3 of plate 9-3 where standard lapse rate curves of 3°F per thousand feet are superimposed on an area-elevation curve of the Boise River basin (see figure 1, plate 9-3). From the temperature index, it is possible to determine the elevation of the freezing level. By assuming all of the snow-covered area to be above all of the snow-free area, it is possible to determine the mean lower limit of the snow cover, or the snowline from the basin snow-cover data. By bisecting the contributing area (area between freezing level and snowline), the median elevation of the contributing area is found. An example, using this method to reconstitute the spring 1955 snowmelt hydrograph for the Boise River above Twin Springs, Idaho, is given later in the chapter.

9-04.14 Method B. - In this method the basin is subdivided into equal-elevation bands and snowmelt, rainfall, and losses are computed separately for each band. An example of such a subdivision is given in figure 2 of plate 9-3. Melt and rain are considered to be uniform throughout the band and the entire band is considered to be either snow covered or bare. For simplicity, temperature indexes are generally used to compute snowmelt, although other indexes could be used. Two variations of this method have been used. In one, rain and melt amounts are computed in terms of inches over the elevation band, the melts then being combined as a weighted average for the basin as a whole, the weighting being in accord with the proportionate share of the basin being contained within each band. In this method, a separate inventory may be kept of the snowpack in each band. In the other variation, melt is computed from curves which automatically give melt in terms of inches over the basin as a whole. Whether or not the band is snow covered must be determined, in this case, from a separate snow-cover index. Such curves expressing snowmelt in the various bands in terms of inches over the basin are shown in figure 4 of plate 9-3 for the Boise River basin above Twin Springs, Idaho. These curves reflect both the area included in the particular band and the elevation of the band with respect to the temperature index station. They are based on an assumed melt rate of 0.1 inch per degree-day and must be multiplied by an appropriate conversion factor. This variation of the basic method was originally used by the Office, Chief of Engineers, in computing the spillway design flood for McNary Dam on the Columbia River. 8/ The other variation whereby the melt in each elevation band is weighted and averaged and an inventory of snowpack water equivalent in each band is maintained has been used in the synthesis of the maximum probable flood for the Painted Rock Reservoir on the Gila River, Arizona. 3/

9-04.15 Method C. - This method makes use of a diagram which gives snowmelt as a function of temperature index and mean elevation of the snow line. Such a diagram for the Boise River basin above Twin Springs, Idaho is included as figure 5 of plate 9-3. (A similar diagram for computing monthly melt on the North Santiam River in Oregon is shown in figure 3, plate 11-4.) The diagram takes into account the contributing area, assuming no melt from areas having temperatures below freezing and from snow-free areas. The method may be used to compute either snowmelt or snowmelt runoff. In the appendix to the Definite Project Report on McNary Dam, it is used in the former sense, while Linsley, in a discussion of the method, 14/ makes use of it to compute snowmelt runoff. Moreover, since Linsley relates it directly to observed (rather than generated) runoff, he found it necessary to include an auxiliary diagram to explain the apparent increase in melt rate through the melt season. In the example given in figure 5, a melt rate of 0.1 inch per degree-day was adopted as a convenience. Hence the results from this diagram must be multiplied by an appropriate factor. As a rule, this method is more amenable to day-to-day forecasting than it is to computation of design floods although it may be used for either.

9-04.16 Time distribution of runoff. - In the preceding paragraph, methods were presented whereby the water generated within a drainage basin during a given time interval could be computed. In order to determine the resultant discharge hydrograph, some method of distributing these amounts of water with time must be employed. For snowmelt runoff hydrographs, the techniques employed in making time distributions of runoff are quite different from those used for rain-flood hydrographs, although the general principles are the same. Those techniques are described in what follows, along with a general discussion of the effect of the snowpack on the time distribution of runoff.

9-04.17 The delay in runoff caused by the snowpack in areas of snow cover has been discussed in some detail in chapter 8. Methods whereby snowmelt or rainwater could be routed from the snow surface to the snow-ground interface were presented. It was shown that most of the rain falling on a subfreezing pack of low density would be absorbed within the pack and the time for the remainder to pass through the pack could be large. This same effect was also shown to apply to early season snowmelt, occurring before the pack was thoroughly ripened and drainage channels had been established. Once the initial losses were satisfied--the pack isothermal at 32°F and liquid-water-holding requirements met--the time delay to runoff caused by the snowpack became relatively small and constant. The time required for melt (or rain) water to percolate through the snowpack varies, of course, with the depth of the pack. Thus even thoroughly ripe packs show some variation with time in the time delay to runoff. This effect, which reduces the time delay as the melt season progresses, tends to be offset by the fact that, as the melt season progresses, the remaining snow cover becomes more and more remote from the basin outlet. Only in very small basins (the size of the laboratory areas) does the time delay becomes shorter as the melt season progresses. In larger basins no such change is discernible. In very large basins, it would seem the reverse should be true: the time delay should increase as the melt season progresses and the remaining snow cover becomes more and more remote from the basin outlet. It has been found, however, that in large basins where the travel time is large compared to time required for melt water to percolate through the snowpack, there is no need to make allowances for variations in basin storage once the spring melt season is actively progressing.

9-04.18 Two methods have been employed in reconstituting discharge hydrographs from given amounts of water excess for snow-covered areas. One employs the method of storage routing; the other makes use of the unit-hydrograph approach. Both methods give satisfactory reconstitutions; however, the latter is generally preferred because of its greater simplicity of computation. (Examples of reconstitutions made by both methods are given in the paragraphs which follow.) For large basins where no regular diurnal hydrograph rise as a result of snowmelt is discernible, or even for smaller basins where only mean daily flows are to be determined, the time distribution of melt is relatively simple and either of the foregoing methods may be employed. If, on the other hand, it is desired that the diurnal rise as a result of snowmelt also be shown, the unit-hydrograph method is definitely superior to that of storage routing. Considering first the situation where only mean daily flows are desired in the reconstitution, standard storage routing and unit hydrograph techniques may be used with a few modifications.

9-04.19 The time distribution of snowmelt runoff by the storage routing method requires that the total runoff be separated into two components: surface runoff and ground-water discharge. (Interflow is included in these two to varying degrees depending upon the method of separation used.) Each component is then routed separately and the two added to determine the resultant streamflow. The surface component has a relatively short storage time while the time of storage for the subsurface component is very long for most basins. If the unit-hydrograph approach is used, on the other hand, it is not necessary to separate the runoff into two components. The unit hydrograph used has an exceptionally long recession limb which effectually represents the slow subsurface and ground-water flows. The special techniques necessary in the derivation and application of such snowmelt unit hydrographs are discussed in section 9-05. Also contained in this section is a further discussion of the storage routing techniques in the synthesis of snowmelt hydrographs.

9-04.20 This section has so far been concerned exclusively with snowmelt and snowmelt runoff in mountainous areas. Because of the effect of elevation on both snowmelt and snow cover, only rarely do mountainous drainage basins contribute melt from all levels simultaneously. A more limited contributing area is usually found which decreases in areal extent and retreats to higher elevations as the melt season progresses. Moreover, these mountainous areas are generally forested to varying degrees, a fact that is of importance in the computation of snowmelt (see chapter 6). While the mountainous areas of the western United States were the particular concern of what has gone before, the relationships presented are also generally applicable to any mountainous area where snow accumulates in deep and lasting snowpacks. Such is not the case with the northern Great Plains area of the United States.

9-04.21 Snowmelt in the Great Plains. - While snowmelt floods are of great importance in this area, they present a problem distinct in many respects from what has so far been considered. The snow cover is relatively shallow; snowmelt floods last but a few days. Due to the flat terrain and lack of forest cover, melting occurs practically simultaneously over entire drainage basins and melt rates are higher than those ordinarily found in mountainous areas. Both these factors tend to produce high, sharp flood waves of the type characteristic of rain floods. In addition the ground beneath the snowpack is often frozen, which encourages surface runoff. The flat terrain and shallow stream gradients allow melt water to be dammed up for some time before being released with a rush.

9-04.22 An extreme example of such a flood was the spring 1950 flood on the Cannonball River near New Leipzig, North Dakota which is illustrated in figure 3 of plate 9-6. The general situation which produces these floods is described in the following excerpt from an unpublished report by K. A. Johnson (see Appendix II, No. 44):

"The plains area of the Missouri River basin may be described as predominantly rolling country with relatively few trees. The temperature range is large with hot dry summers and cold winters. The snow accumulation season generally begins in December and extends through March. Although some melting generally occurs as a result of short warming periods during the winter months, the major portion of the snowmelt occurs in a period of 10 days or less during late March or early April. Flooding as a result of snowmelt does not occur every year, but it is the most prevalent type of flood for the western tributaries of the Missouri River down to the Nebraska-South Dakota State line.

"When a flood potential does exist as the result of snow accumulation the following conditions are frequently found at the beginning of the melt period: (1) moderate to severe drifting, (2) frozen ground with top few inches likely to be quite moist, (3) layer of solid ice and/or granular ice crystals next to ground. Items (1) and (2) are almost always present but the existence of item (3) is likely to be quite spotty even though fairly widespread. As the temperature rises and melt begins, the snowpack increases in density and bare ground becomes evident in numerous cases. A layer of slush is likely to be found at the bottom of the remaining snow. Appreciable runoff into the streams is not likely to occur before these conditions are noted. Then, as warm temperatures continue, the ground becomes progressively free of snow cover, and runoff into the streams increases rapidly, continuing until the area is free of snow, with the exception of that part which is covered by deep drifts or protected from the melting factors."

Meteorological conditions antecedent to and during this flood are given in figure 1 of plate 9-6. Concerning this flood occurrence is the following quotation from an unpublished report by C. W. Timberman (see Appendix II, No. 38):

"Description of Basin. The Cannonball River basin is located in southwestern North Dakota and drains an area of approximately 4300 square miles. The river flows in an easterly direction to enter the Missouri River about 45 miles downstream from Bismarck, North Dakota. The D. A. of the sub-basin above New Leipzig is approximately 1180 square miles and is of predominantly rolling topography. Elevations range from about 2250 feet at the New Leipzig gage to over 3000 feet along the rim of the basin with scattered buttes extending up to 3300 feet. Characteristic low bluffs occur along the stream channels. Very few trees grow in the area even along the main stream channels.

"Climate. The climate is sub-humid. Average annual precipitation is slightly over 15 inches, of which about 11 inches occur from April to September, inclusive. General rains with average depths of over two inches are rare. Amounts vary greatly from the normal. Snow accumulation during the winter is usually moderate. The mean annual temperature at Mott, North Dakota, is 41.7°F; normal mean for January is 10.2°F. Extreme temperatures vary between 114 and minus 47°F.

"Floods. Spring floods, caused by the melting of the winter snow accumulation, are the most frequent type and occasionally cause extensive damage. Summer rains causing extensive floods are rare. Average discharge over a nine-year period of record is 107 cfs at New Leipzig. The 1950 flood, the reconstitution of which is the object of this report, is by far the largest of record throughout the basin except possibly in some of the small headwater areas above the heavy 1950 snowfall area. The preceding winter was one of the coldest of record while precipitation was considerably above normal. No appreciable runoff occurred from first week in December through the end of March because of the low temperatures. The melt that did occur during the winter is believed to have infiltrated into and saturated the upper layer of ground which subsequently was frozen into an impervious layer. From 23 to 27 March a severe blizzard occurred, with precipitation amounts ranging to over 3 inches of water content. From 7 to 10 April approximately one additional inch (water content) fell. Warm temperatures during the first week of April over the lower portion of the basin resulted in approximately 100,000 acre feet of runoff passing the gaging station at Breien; however, no appreciable flow was evidenced at the New Leipzig or the Pretty Rock gaging stations, above which lay the heaviest accumulation of snow. At the end of the second week in April much warmer temperatures occurred (with maximums approaching 70°F on 17 April) causing rapid melting of the entire snow cover, except the deeper drifts."

9-04.23 Another basin in this area that has been extensively investigated is that of Spring Creek near Zap, North Dakota. It also experienced a severe flood in 1950 and again in 1952. These flood events, along with the pertinent hydrometeorological data, are illustrated in figures 1, 2, and 4 of plate 9-6 and discussed in the following excerpt from an unpublished report by C. A. Burgtorf (see Appendix II, No. 42):

"Description - Spring Creek lies in west central North Dokata and is a tributary of Knife River which it enters about 40 miles above its mouth. Zap, North Dakota lies on Spring Creek about 11 miles upstream from its confluence with Knife River. The drainage area above Zap is 545 square miles. The topography is predominantly rolling with very few trees and elevations ranging from about 1750 to 3000 feet.

"Climate - The climate is relatively dry with a large temperature range. The average annual precipitation is about 16 inches of which about 60 percent falls during the summer months. Snow usually accumulates through the winter months and melts off in March or early April. The major portion of the melt usually occurs during a period of less than ten days and often results in some flooding. Rainfall floods in the area are rare.

"1950 and 1952 Floods - The two largest floods of record for Spring Creek at Zap occurred in 1950 and 1952. From snow surveys it was estimated that the water content of the 1950 snow cover at the beginning of the melt period was 2.5 inches and for 1952 was 3.4 inches. Discharge records show that the runoff from the principal melt period was 1.05 inches in 1950 and 2.08 inches in 1952. The crest discharge in 1950 was, 4580 cfs and in 1952 was 6130 cfs...."

9-04.24 The reconstitution of such plains area floods is a difficult procedure. For one thing, the lack of forest cover makes temperature and temperature-vapor pressure indexes of snowmelt generally unsatisfactory. As a result of rapid changes in albedo of the snowpack during such a flood event and the difficulty inherent in the estimating of albedo, the inclusion of radiation in a thermal-budget index is of questionable value. The initial condition of the snowpack and the underlying soil have most important, yet difficult to assess, effect upon the resulting runoff. Akin to rain-on-snow events, the runoff is largely dependent upon snow and ground conditions, and because of the rapid rise in both these types of floods, this initial condition must be known. Unlike spring snowmelt floods from the deep mountain packs where the pack is ripened and initial losses satisfied long before peak flows are reached, the rapid rise immediately follows satisfaction of these losses. Rapid variation in the areal extent of the shallow snow cover serves to further confound the problem. The problem will not be dwelt on further here; no ready solution is known. Reference is made to the Definite Project Reports for Garrison, and Oahe and Fort Randall Reservoirs for attempts at practical solution. 6/ 7/

9-05. TIME DISTRIBUTION OF RUNOFF

9-05.01 General. - The time distribution of runoff from rain-on-snow events is customarily made using the methods applicable to rain floods in general: the unit-hydrograph method is commonly employed to distribute the water excess, the base flow being estimated. Because these methods are common to hydrology in general, they will not be examined here. Of concern to this report are methods whereby the runoff from melting snow may be distributed into a discharge hydrograph. This problem is peculiar to the field of snow hydrology and little is said concerning it in the general literature of hydrology. For this reason, it shall be treated here in some detail.

9-05.02 One thing that makes the time distribution of snowmelt water different from that of rain floods is the difference in the rates of generation. In rain-flood control problems, since only relatively high rates of runoff are usually considered, it has been customary to deal only with the direct runoff in the unit hydrograph application, that is, the water that reaches the stream channels by the most direct routes. The other runoff--that which is more delayed in reaching the stream channels--is considered only as base flow, its time distribution being merely estimated. While a considerable volume of water may eventually be discharged as base flow (even exceeding the volume of the direct runoff), because of the relatively sluggish flow, compared to the direct runoff, it adds relatively little to the rate of flow at the time of peak flow. Thus even a large error in estimating the base flow component has relatively little effect upon the peak discharge. The total water generated, in the case of rain floods, is separated into water excess (direct runoff) and "losses" usually by means of an infiltration curve which gives infiltration rates corresponding to different accumulations of water generated. "Losses," so defined, thus include much of the water ultimately discharged by the base flow curve. Usually no strict accounting is kept between the "losses" and the water discharged by the base flow curve. The unit hydrograph approach may be used for snowmelt runoff hydrographs by routing separately the subsurface and surface runoff components. Also, storage routing techniques are applicable to the time distribution of runoff from snowmelt. Both of these approaches will be considered.

9-05.03 Storage routing. - Storage routing, while most commonly associated with the routing of flood waves through river reaches or through reservoirs, may also be applied to the determination of discharge hydrographs from drainage basins. The rainfall and snowmelt water generated within the basin is the inflow to be routed; the hydrologic characteristics of the basin are reflected in the time of storage used in the routing procedure. The following paragraphs will describe briefly the methods employed. No detailed discussion of storage routing in general will be given here, it being assumed the reader is familiar with the basic principles.

9-05.04 In the solution of the general storage equation, two assumptions are commonly made: (1) that the storage is directly proportional to the outflow, $S = t_s(O)$; and (2) that the storage is a function of both inflow and outflow, $S = K\ (xI + (1-x)\ O\ )$. The former is termed reservoir-type storage, the assumption being that outflow and storage vary together. The latter assumption is the basis for the Muskingum system of flood routing; here it is assumed that the storage varies directly with a weighted inflow and outflow. (As x approaches zero, the conditions of reservoir-type storage are approached.) Simple reservoir-type storage is limited in application, since no allowance can be made for time of travel through a river reach or in a drainage basin. Inherent in the method is the fact that the peak of the outflow discharge hydrograph always occurs at the time at which the outflow graph crosses the recession limb of the inflow hydrograph. This follows from the fact that subsequent to this time outflow exceeds inflow and hence the amount of storage (and hence the outflow also) must decrease. This restriction does not hold for the Muskingum method of routing since here storage is also, partially, a function of inflow. This makes this method somewhat more flexible, allowing for a possible travel time for the flood wave.

9-05.05 The storage routing approach has been used by the U. S. Weather Bureau River Forecast Center at Portland, Oregon for reconstitutions of spring snowmelt floods. Flood hydrographs for the Payette River basin, Idaho, have been determined by this method as reported on by Zimmerman. 21/ Briefly the method consists of separating the total water excess into two components: surface runoff and ground water. Each is then routed separately using different times of storage. These times of storage are empirically determined to give the best fit to historical data. The time of storage for the ground-water component is, of course, quite long compared to that for surface runoff. To facilitate storage routing in general, the U. S. Weather Bureau has developed an electronic routing analogue which solves the Muskingum routing equation. 13/ It is especially useful in reconstituting an entire season's snowmelt flood hydrographs. Basically the method is simple. Practically, the division of the runoff into the two components, the determination of the best storage times, and the relative weighting of inflow and outflow in determining the storage time for the two components, are more difficult.

9-05.06 Another method of storage routing, based on the reservoir-type storage equation, may also be used to determine outflow-hydrographs from drainage basins. This method consists of multiple-stage reservoir-type storage routing. That is, the inflow hydrograph is successively routed through two or more stages of reservoir-type storage. An example of this type of routing is given in figure 4, plate 9-7, which shows a rectangular inflow hydrograph routed through one, two, and three 6-hour stages of reservoir-type storage. Also shown is the same inflow hydrograph routed through one 18-hour stage of reservoir-type storage. As may be seen, a unit-hydrograph-shaped outflow wave results from the square inflow wave. By this method any desired time of travel delay may be had by the proper selection of number of stages and times of storage for each stage. Moreover, by varying the times of storage between stages, an extremely flexible system of time distribution of runoff is obtained, capable of reconstituting discharge hydrographs from basins of widely differing hydrologic characteristics. The disadvantage of such a procedure is that the computations involved are laborious. Even one stage of reservoir-type storage requires a considerable amount of computation; multiple stages make the computations even more formidable. Recently, however, an electronic analog has been developed which solves this type of storage routing (see Tech. Bull. 18). The analog permits the use of as many stages as desired with different times of storage in each stage if desired. In addition, this device makes it possible to vary the time of storage during the routing operation, a feature that further increases its flexibility. Multiple-stage routing is, of course, also possible for the Muskingum type of storage routing, and an electronic analog to solve this type of routing appears feasible. This type of routing is discussed by Clark 2/ who also points out the expedient of translating the inflow hydrograph in time and then routing it through a single stage of reservoir-type storage in order to simplify the computations involved. The use of multiple-stage routing in conjunction with the Muskingum method seems hardly warranted in view of the adequacy of the alternative approach of multiple stage storage-type routing.

9-05.07 Unit hydrographs. - The unit-hydrograph approach to the time distribution of runoff has been the method most extensively used in connection with runoff from snowmelt. However, as was previously pointed out, unit hydrographs developed for snowmelt are not applicable to rainfall and vice versa. Yet the difference between a rainfall unit hydrograph and one for snowmelt for a given area is one of degree only. Base flow curves can be eliminated for snowmelt runoff as for rainfall runoff; however, in the former case, the rate used to separate the two components of flow must be considerably less than in the case of rainfall. No absolute division exists between water excess and losses. As a greater percentage of the total water available is considered to be water excess, the recession limb on the unit hydrograph increases in length and rate of flow and the base flow curve decreases its contribution to runoff. On the other hand, as a greater quantity of water is considered to be losses and less as water excess in the infiltration curve separation, the unit-hydrograph tail shortens, and the unit hydrograph approaches that used for rain-flood synthesis. At the same time the base-flow curve increases in magnitude, becoming progressively more difficult to estimate. It is because of this that the method given herein was devised: all water not permanently lost to runoff is distributed by means of the snowmelt unit hydrograph. This assumes a fixed and constant quantitative relation between direct and base flow, which also has certain weaknesses. No base-flow curve need be estimated; only the recession from the existing streamflow at the start of the reconstitution is added to the computed hydrograph. The exceptionally long recession limb of the resulting unit hydrograph poses a special problem in its application. This and other problems involved in the derivation and application of unit hydrographs will now be considered.

9-05.08 Rainfall unit hydrographs may best be derived from a brief, isolated period of intense rainfall, falling at a uniform rate on a surface which has previously has its initial loss capacity satisfied so that the water excess rate is high and uniform. For snowmelt runoff, where the rate of water excess is low and is more or less continuous, the use of the S-hydrograph approach to the derivation of unit hydrographs for snowmelt runoff is appropriate.

9-05.09 The S-hydrograph approach to the derivation of unit hydrographs 16/ is useful in many respects: (1) it provides a method of adjusting the derived unit hydrograph for non-uniform rates of generation of water during the period used in computing the unit hydrograph; (2) it provides a means of adjusting the observed period to the period desired for the derived unit hydrograph; (3) it provides a convenient method of adjusting the area under the unit hydrograph to unit volume; (4) it allows several unit hydrographs to be averaged in order to arrive at a mean unit hydrograph; and (5) it provides a method of seperating a given unit hydrograph into two unit hydrographs of unequal periods of generation. It is this last aspect that makes the S-hydrograph approach especially valuable to snowmelt runoff. The aspects enumerated above are considered in detail in Technical Bulletin 14; they shall be discussed only briefly here.

9-05.10 Figure 1 of plate 9-7 shows an example of the use of an S-hydrograph in adjusting the unit hydrograph for non-uniform rates of generation of water excess. Suppose the hydrograph given by the solid curve results from the 6-hour period of non-uniform water excess illustrated. If this pattern of water excess is repeated every six hours, the resulting hydrograph is likewise repeated, the total flow being given by the wavy solid curve. This curve was determined by repeating the observed hydrograph every six hours and summing its ordinates. A smooth curve drawn through the wavy one represents an S-hydrograph which results from a constant rate of water excess equivalent to the mean rate during the 6-hour period (0.2 in./hr.). Expressing the ordinates of this curve in percent of its equilibrum rate, a percentage S-hydrograph results which is independent of the rate of water excess and reflects only the basin characteristics. From the percentage S-hydrograph, unit hydrographs of any period may be derived as illustrated in figure 2 of plate 9-7. The equilibrim rate corresponding to the rate of water excess for the period selected is determined, and the ordinates of the S-hydrograph expressed relative to this rate. Differences in ordinates, (e.g., bc, b'c', b"c") separated by the desired unit-hydrograph period (ab, a'b, a"b") define the unit hydrograph.

9-05.11 In the application of the unit hydrographs of snow-melt runoff, the selection of the best unit-hydrograph period becomes a problem. The early portion of the unit hydrograph with its comparatively rapid changes requires a relatively short period in order that it be adequately defined, while with the long recession limb this would result in unnecessarily detailed computations. Here a longer period is better suited to the job. It is possible, for a single S-hydrograph, to define two unit hydrographs of unequal periods, a shorter period being used for the earlier, more rapidly changing portion and a longer period for the long, more slowly changing recession limb. This may be seen by reference to figure 3 of plate 9-7. Supposing it is desired to separate the S-hydrograph CDE at time D. If a horizontal line AB is drawn to intersect the curve CDE at point D, two S-hydrographs are thus defined: CDB and ADE. By taking incremental differences on these two S-hydrographs, it is possible to determine unit hydrographs of any desired period. For example, supposing a 3-hour unit hydrograph is desired for the portion to time D and a 6-hour unit hydrograph thereafter. Three-hour incremental differences are thus determined for the first portion and 6-hour incremental differences for the latter, each being multiplied by the proper conversion factor. Such unit hydrographs are illustrated in figure 3 of plate 9-7 along with the S-hydrographs from which they were derived. This example, however, is not representative of the actual situation, where a greater contrast in time periods would exist as well as a longer recession limb on the S-hydrograph. The example was chosen for clarity in illustrating the method of diversion rather than its representativeness of an actual situation. It is to be pointed out that the same water generated is distributed by both these unit hydrographs. No separation into two components, as was the case in the storage-routing approach, is necessary. In combining the two flow components resulting from the time distributions by the two unit hydrographs, flows corresponding to the finer time increment are determined from the coarser by interpolation.

9-05.12 By rainfall unit-hydrograph standards, the two unit hydrographs above are both predominantly for base flow. If it is also desired to distribute rainfall-type water excess, still another surface runoff unit hydrograph is required; however, the detailed base flow analysis just discussed is hardly warranted in this case. The method of rainfall unit hydrographs (with estimated base flow) is generally used. Finally, it is pointed out that the separation of water generated into two components is quite arbitrary; the important thing is that the unit hydrograph used to distribute the water excess and the method used to estimate the base flow be consistent with the method of separation--a precaution that was previously discussed in connection with the synthesis of discharge hydrographs in general.

9-06. BOISE RIVER HYDROGRAPH RECONSTITUTIONS

9-06.01 General. - To illustrate some of the methods of snowmelt hydrograph synthesis previously discussed, the example of the 1955 spring snowmelt hydrograph rise on the Boise River near Twin Springs, Idaho is used. Melts for the period were computed by the thermal-budget index discussed in paragraph 9-04.04. They are shown diagramatically in figure 1 of plate 9-4 which also shows the pertinent meteorological data. One-, two-, and three-day forecasts of mean daily flow were made from these melt data (and gaged rainfall amounts), as well as reconstitutions of virtually the entire melt season. The reconstitutions and forecasts are discussed in the paragraphs which follow.

9-06.02 Seasonal reconstitutions. - In figure 2 of plate 9-5 the actual discharge hydrograph for the 1955 spring snowmelt season on the Boise River above Twin Springs, Idaho is given in terms of mean daily flows. Also shown on the same figure are reconstitutions made using the unit hydrograph of figure 3, plate 9-4 and by storage routing. The storage-routing reconstitution was accomplished by separating the total water excess into two components: 70 percent to direct runoff and 30 percent to ground-water discharge. Both components were routed using multiple-stage reservoir-type storage routing, the direct runoff being routed by two 36-hour stages and the ground-water discharge by three 10-day stages. Prior to the beginning of the reconstitutions on 1 May, several weeks of melt had occurred. Thus the pack was thoroughly ripe so that a constant melt factor could be used, and also all initial losses, both in the snowpack and in the basin, were satisfied. Recession flows, from the flow at the start of the reconstitution, were added to the flows determined by unit-hydrograph and storage-routing methods. (The recession flow curve for the Boise River basin is shown in figure 2, plate 9-4.) In making the foregoing reconstitutions, it is assumed that the actual temperature sequence, basin snow cover, and precipitation are known throughout the season. The agreement between the actual discharge hydrograph and the reconstitutions is actually a test of how well the discharge hydrograph can be reconstituted from known hydrometeorological conditions. This is what is pertinent in the synthesis of design floods. It is not a test of ability in forecasting meteorological conditions.

9-06.03 In figure 4 of plate 9-5, a portion of the 1955 spring snowmelt hydrograph for the Boise River basin is snown in expanded time scale. Hourly flows are plotted and the diurnal variation in flow is apparent. This diurnal variation in flow was reconstituted using the 8-hour unit hydrograph of figure 3, plate 9-4. All snowmelt was assumed to have occurred during one 8-hour interval (1000 - 1800 hours), the other two such intervals having no water generated except in the instances where rain occurred during these times. While such a reconstitution is seldom required, except possibly for very small drainage basins, it is included here to demonstrate that such reconstitutions are possible.

9-06.04 Short-term forecasts. - The reconstitution of an entire melt season, or portion thereof, as in the preceding paragraphs, is usually made only in connection with design-flood determinations. More usual are short-term forecasts of runoff made for the operation of reservoirs and other purposes. Such forecasts are generally more accurate than are the predicted flows for the same days taken from a reconstitution of an entire season's runoff hydrograph. The chief reason for the better accuracy of short-term forecasts is that when short-term forecasts are made, the current streamflow is known and only the increment of flow above the recession from the known flow need be estimated. When an entire snowmelt season's hydrograph is reconstituted, all water generated from the beginning of the season through the day in question has an effect upon the discharge for a given day. Presented in figure 1 of plate 9-5 are the results of one-, two-, and three-day forecasts of runoff for the 1955 spring snowmelt season on the Boise River. These forecasts were made using the same basic data, and a distribution graph derived from the same S-hydrograph was used to make the seasonal reconstitutions. Actual values of temperature, vapor pressure, precipitation and snow cover were used in the examples given; however, these data would not be available for actual runoff forecasts, and forecast values would have to be used, introducing another possible source of error. Thus figures 1a, 1b, and 1c represent what could be done in the way of one- to three-day forecasts, providing the forecasts of the meteorological conditions were correct. In making such short-term forecasts of runoff, the long recession flow of the unit hydrograph need not be run out; only as many days flow need be used as the length of the forecast. Thus there is a considerable saving in computation over that required to reconstitute an entire season.

9-07. SUMMARY

9-07.01 Two kinds of syntheses of runoff hydrographs are encountered in snow hydrology: (1) short-term forecasts and (2) the synthesis of an entire melt season or rain-on-snow event. The former is ordinarily used in the operation of reservoirs and in the making of streamflow forecasts, while the latter is ordinarily involved in the determination of design floods. With respect to the first kind of hydrograph synthesis, current conditions of streamflow, snow cover, etc., are known; only the increment of flow above the recession from the current flow need be estimated, and that only a few days in advance. Forecast values of meteorological parameters are required if the forecast period exceeds the lag time for the basin. With respect to the second kind of hydrograph synthesis, an entire flood hydrograph must be determined with only the initial conditions of streamflow and snow cover known. The actual values of the meteorological parameters necessary to the computation are known in the case of the reconstitution of a historical flood, and the assumed parameters in the case of a design flood synthesis.

9-07.02 Elevation has an important effect upon both snowmelt and precipitation excesses. Snowmelt rates vary inversely with elevation as a result of the general decrease in net heat supply with increasing elevation. The form of the precipitation (rain or snow) is a function of air temperature and hence also of elevation, while the total quantity of precipitation also increases with elevation, due to the orographic effect. Snow cover ordinarily exhibits a marked increase with elevation, in consequence of the precipitation increase with elevation, the greater likelihood of it occurring in the form of snow with increasing elevation, and the greater melt rates at lower elevations. Consequently, basinwide snowmelt must first increase with increasing elevation as the snowline is approached and the areal extent of the cover increases, and then must decrease with the decreasing melt rates at the highest elevations over the virtually 100 percent snow-covered areas. Moreover, snow cover has an important effect upon the runoff which results from rainfall. Generally speaking, very little runoff results from the usually light to moderate rains which fall on the snow-free portions of the basin, during the spring snowmelt season, while the percentage runoff is quite high for the snow-covered portions. During the winter season, however, this effect may be reversed, with the more intense winter rains producing considerable runoff from the snow-free areas, at the same time being stored to a greater degree over the snow-covered areas. In consequence of all these things, it becomes apparent that elevation must be considered in any general scheme of hydrograph synthesis for snow-covered areas.

9-07.03 There are two general methods by which elevation effects may be incorporated in a scheme of hydrograph synthesis. One is simply to divide the drainage basin into elevation bands and compute the water excess for each band separately--snow cover, precipitation, snowment and losses to be uniform over each band. The other method is to treat the basin as a unit, making corrections for variations in the form of precipitation, snow-covered area, melt rates, etc., with elevation.

9-07.04 The basic components of any method of hydrograph synthesis are: (1) snowmelt, (2) rainfall, (3) losses, and (4) time distribution of runoff. Pertinent comments on each of these follow:

(1) Snowmelt. - In general, the thermal-budget method of snowmelt computation is more amenable to design floods and the index method to the forecasting of streamflow. Because of the different meteorological conditions, different methods are necessary for the computation of melt during spring snowmelt periods and winter rain-on-snow events. (See sections 6-04 and 6-07.)

(2) Rainfall. - In determining basin rainfall from precipitation gage data, corrections must be made for gage deficiencies and form of precipitation. In separating out snowfall amounts, it should be remembered that gage deficiencies are commonly large in areas of snowfall.

(3) Losses. - The concept of losses is different for synthesizing rain-on-snow hydrographs and predominantly snowmelt hydrographs. Rain-on-snow synthesis uses the conventional rainfall-loss concept, wherein all water is considered "lost" (to direct runoff) which is delayed in reaching the gaging station through varying degrees of subsurface flow, so that it contributes little to the hydrograph peak. Snowmelt snythesis considers only that water to be a "loss" which is permanently stored in the snowpack (as free or refrozen water) or is permanently lost to runoff (by evapotranspiration and deep percolation).

(4) Time distribution. - Either unit hydrographs (including distribution graphs) or storage routing methods may be used in the time distribution of runoff from snow-covered areas. Conventional rainfall-type unit hydrographs may be used for rain-on-snow events; special "long-tailed" unit graphs are used to distribute spring snowmelt excess. Storage routing is most amenable to spring snowmelt hydrograph synthesis where the total water excess is divided into two (or more) components and routed separately using relatively short times of storage for the more direct component and relatively long times of storage for that water which is more delayed in reaching the basin outlet.

9-08. REFERENCES

1/ ASCE, Hydrology Handbook, Manuals of Engineering Practice, No. 28, New York, 1949.

2/ CLARK, C.O., "Storage and the unit hydrograph," Trans. ASCE, Vol. 110, pp 1419-1446, 1945.

3/ CORPS OF ENGINEERS, Los Angeles District, "Hydrology for Painted Rock Reservoir, Gila River, Arizona," Design Memorandum No. 1, 1 August 1954.

4/ CORPS OF ENGINEERS, Office, Chief of Engineers, "Flood hydrograph analyses and computations," Part CXIV, Chap. 5, Engineering Manual, Civil Works Construction, March 1948.

5/ CORPS OF ENGINEERS, Office, Chief of Engineers, "Routing of floods through river channels," Part CXIV, Chapter 8, Engineering Manual, Civil Works Construction, September 1953.

6/ CORPS OF ENGINEERS, Omaha District, "Garrison Reservoir, North Dakota, Definite Project Report--Appendix I-A; Hydrology and general hydraulics, Spillway Design Flood," January 1946.

7/ CORPS OF ENGINEERS, Omaha District, "Oahe Reservoir-Fort Randall Reservoir, North Dakota, Definite Project Report--Appendix I; Hydrology and general hydraulics, Supplement--Spillway Design Flood," July 1947.

8/ CORPS OF ENGINEERS, Portland District, "Definite project report on McNary Dam, Columbia River, Oregon-Washington," Basis of Design, Vol. I (Appendix A-2--Hydrology), 6 May 1946.

9/ CORPS OF ENGINEERS, Portland District, "Hydrology and meteorology, Cougar Dam and Reservoir, South Fork McKenzie River, Oregon," Design Memorandum No. 2, 15 December 1955.

10/ CORPS OF ENGINEERS, Seattle District, "Derivation of spillway design flood inflow and Appendix A, Libby Project, Kootenai River, Montana," Design Memo. No. 2, 29 July 1952.

11/ CORPS OF ENGINEERS, Walla Walla District, "Definite Project Report on Lucky Peak Dam, Boise River, Idaho, " Appendix A--Hydrology, 3 October 1949.

12/ THE ENGINEER SCHOOL, Fort Belvoir, Va., "Flood Control," Engineering Construction, 1946.

13/ KOHLER, M.A., "Application of electronic streamflow routing analogue," U.S. Weather Bureau, Washington D.C., 25 October 1950, (mimeo. report).

14/ LINSLEY, R.K., "A simple procedure for the day-to-day forecas of runoff from snowmelt," Trans. Amer. Geophys. Union, 24, Pt. III, pp. 62-67, 1943.

15/ LINSLEY, R.K., Jr., M.A. Kohler and J.L.H. Paulhus, Applied Hydrology, McGraw-Hill Book Co., Inc., New York, 1949.

16/ MORGAN, R. and D.W. Hulinghorst, "Unit hydrographs for gaged . ungaged water sheds," U.S. Engineer Office, Binghamton New York, July 1939, (Rev. 1943).

17/ SNYDER, F. F., "A conception of runoff phenomena," Trans. Ame Geophys. Union, Pt. IV, pp. 725-738, 1939.

18/ SNYDER, F.F., "Discussion of storage and the unit hydrograph C.O. Clark," Trans, ASCE, Vol. 110, pp. 1465-1471, 194

19/ SNYDER, F.F., "Large floods from melting snow and rain," Symp on Hydrology of Floods, Ninth General Assembly of The International Union of Geodesy and Geophysics, Brussel 1951.

20/ SUMMERSETT, John, "Reproduction of snow melt floods in the Bo River," Proc. West. Snow Conf., Boise, Idaho, April 19

21/ ZIMMERMAN, A.L., "Reconstruction of the snow-melt hydrograph Payette River Basin," Proc. West. Snow Conf., Portland Oregon, April 1955.

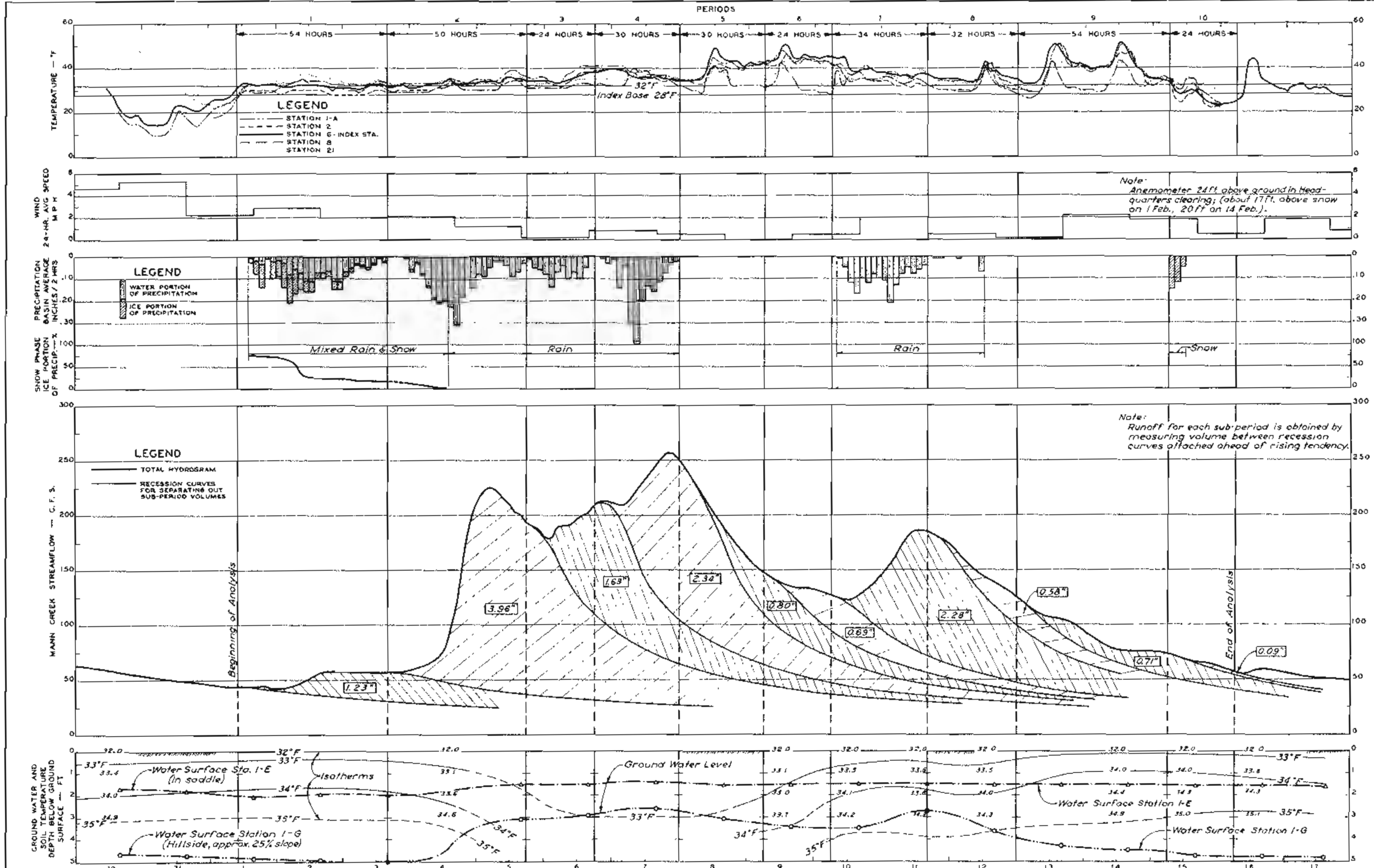


FIGURE 1 — HYDROLOGICAL AND METEOROLOGICAL LOG — JAN.-FEB. 1951

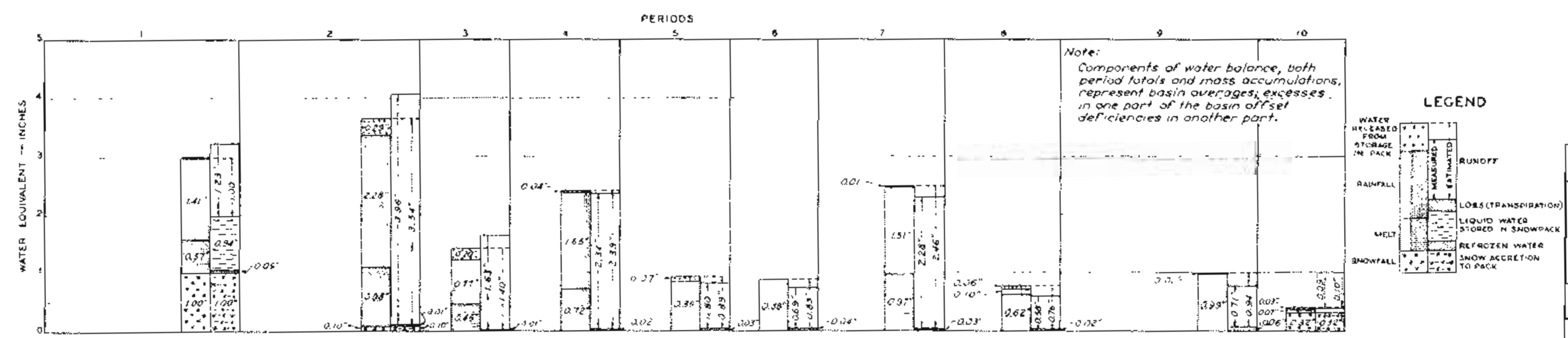
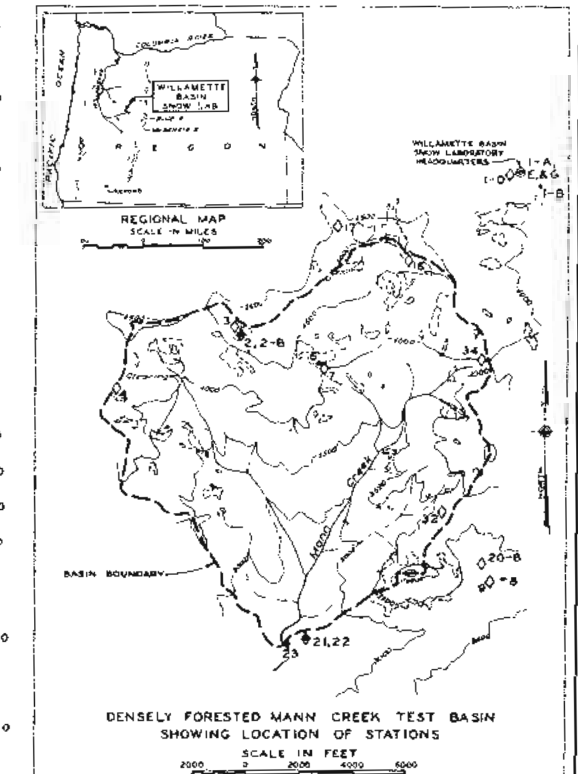


FIGURE 2 — COMPONENTS OF WATER BALANCE FOR EACH PERIOD



OBSERVATIONS		
DATA ELEMENTS	STATIONS	SYMBOL
PRECIPITATION (RECORDERS)	1-B, 2, 6*, 8, 21	•
AIR TEMPERATURE AND RELATIVE HUMIDITY	1-A, 2, 6*, 8, 21	⊸
SOIL AND SNOW TEMPERATURE, WIND MOVEMENT, AND TWO GROUND WATER WELLS	HEADQUARTERS (1-A, E, & G)	◎
STREAM FLOW	23	▲
SNOW DEPTH AND WATER EQUIVALENT	1-D, 2-B, 3, 4, 7, 9, 16, 17, 20-B, 22, 32, 34	◇
*STATION 6 IS THE TEMPERATURE INDEX STATION AS IT IS IN RESEARCH NOTE NUMBER 19.		

FIGURE 3

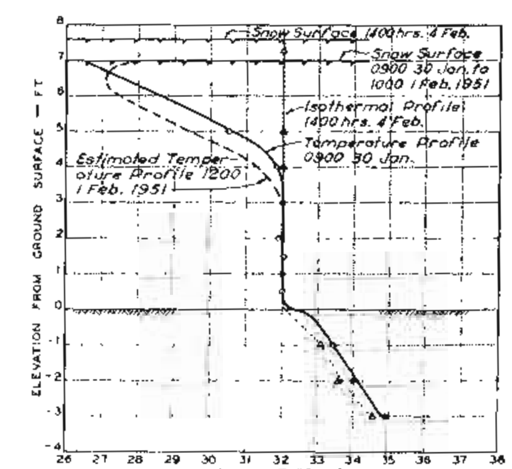


FIGURE 4 — GROUND AND SNOW-PACK TEMPERATURE PROFILES — HEADQUARTERS

SNOW INVESTIGATIONS
SUMMARY REPORT

SNOW HYDROLOGY

RAIN ON SNOW ANALYSIS

FEBRUARY 1951 WILLAMETTE BASIN SNOW LABORATORY
MANN CREEK DRAINAGE AREA 5.2 SQUARE MILES

SHEET 1 OF 2

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

TO ACCOMPANY REPORT DATED 30 JUNE 1956

PD-20-25/56

PLATE 9-1

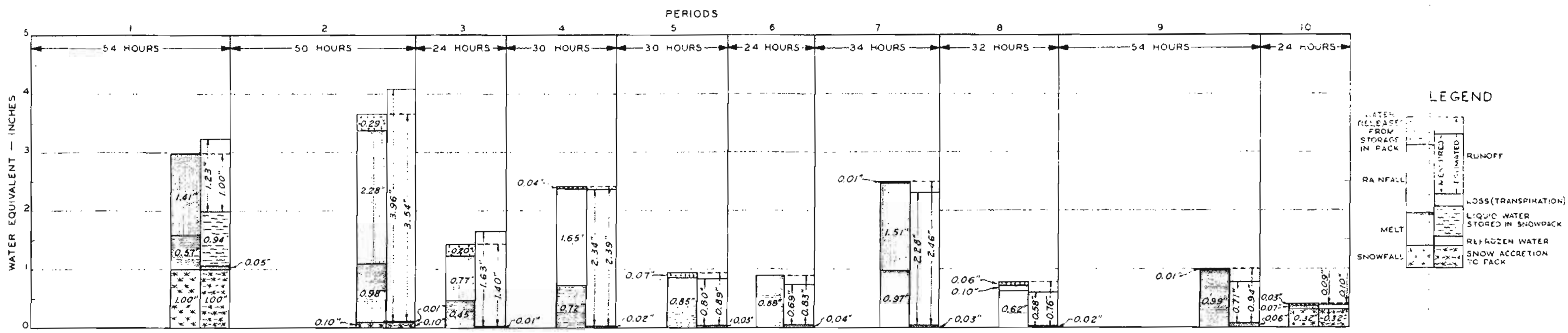


FIGURE 1 — COMPONENTS OF WATER BALANCE FOR EACH PERIOD①

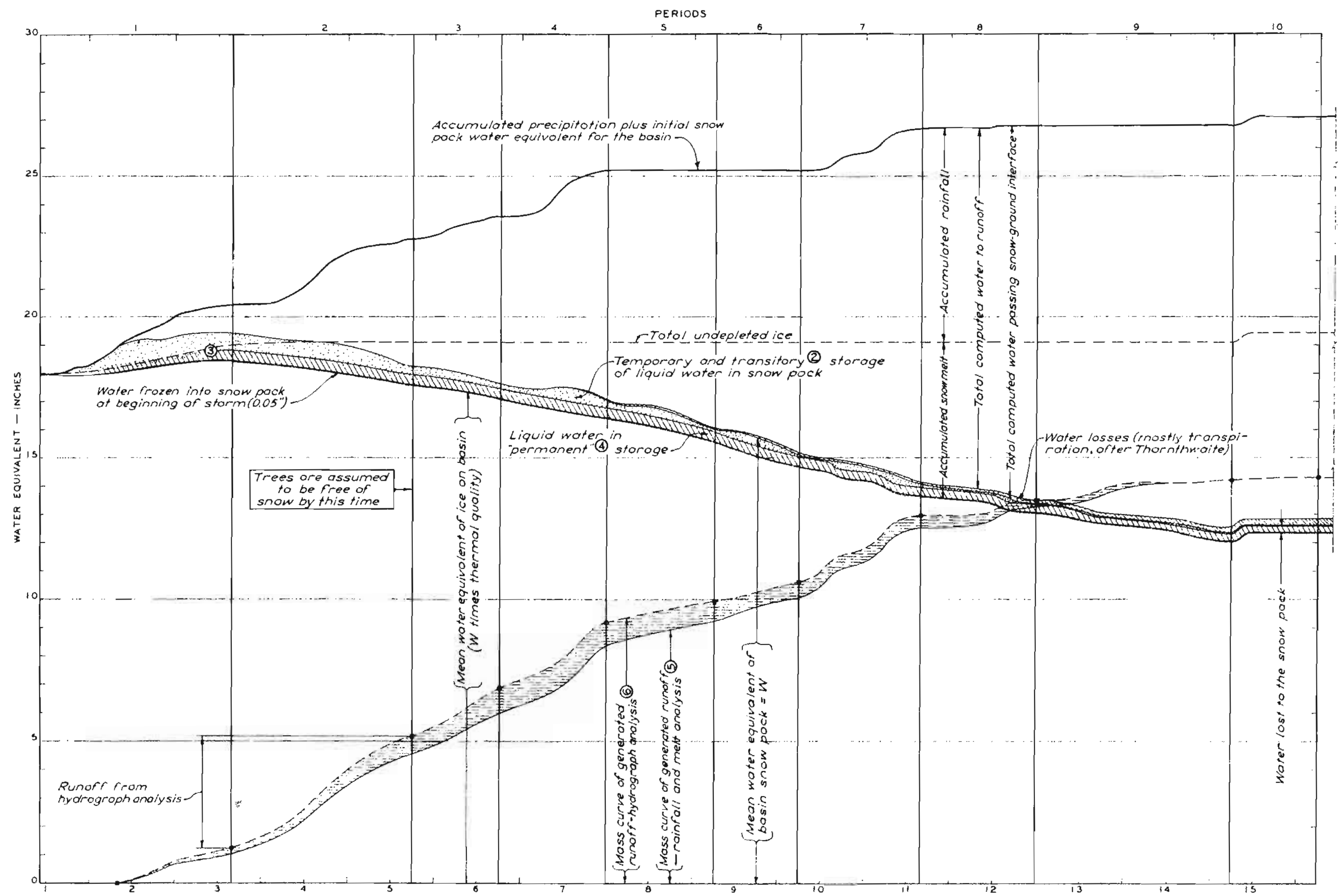


FIGURE 2 — ACCUMULATION AND DEPLETION OF HYDROLOGIC ELEMENTS (MASS CURVES)①

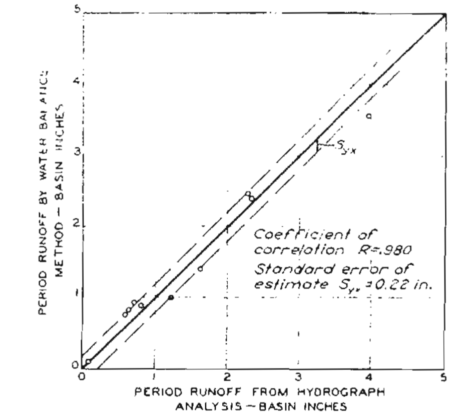


FIGURE 3 — RUNOFF COMPUTED BY WATER BALANCE METHOD COMPARED WITH THAT DERIVED FROM HYDROGRAPH ANALYSIS

Notes:

① Components of water balance, both period totals and mass accumulations, represent basin averages; excesses in one part of the basin offset deficiencies in another part.

② Transitory storage is computed by regarding the snow cover as a channel and routing the rain plus melt through it to the ground surface by the Muskingum method with $x=0.3$ and $t=2.5$ hours, while the water equivalent of the snow cover depletes from 19 to 15 inches; and with $x=0.3$ and $t=1.5$ hours as the snow water equivalent decreases from 15 to 12 inches.

③ Distinction between temporary and permanent storage during first sub-period is not significant and therefore not shown.

④ Basin snow pack is assumed to have the potential of holding 2% by weight as liquid water adsorbed to the snow crystals. This water that resists gravitational drainage is designated as "permanent" storage, although it becomes available for runoff when the ice matrix to which it is adsorbed is melted.

⑤ Curve ⑤ was drawn from the 2-hourly values of generated runoff at snow-ground interface, computed from precipitation, melt, and storage in the snow pack. Sta. 6 temperatures were used for melt.

⑥ Curve ⑥ was drawn to plotted points (o's) computed from hydrograph analysis, but between points was patterned in accordance with curve ⑤.

SNOW INVESTIGATIONS
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FEBRUARY 1951 WILLAMETTE BASIN SNOW LABORATORY
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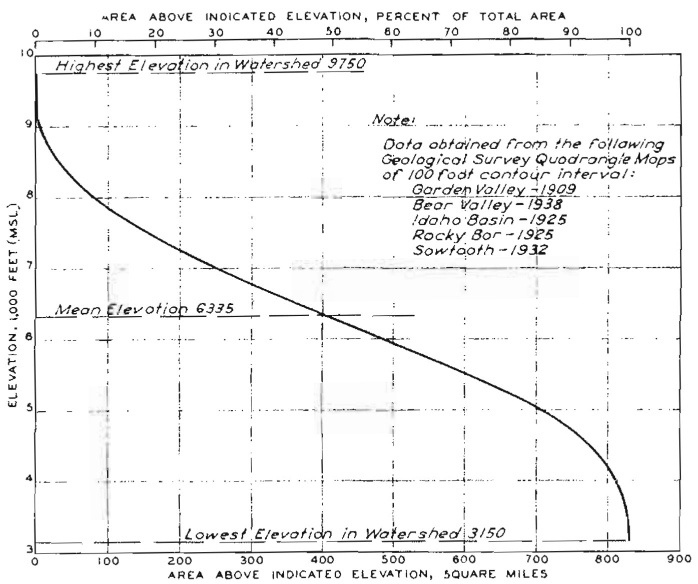
SHEET 2 OF 2

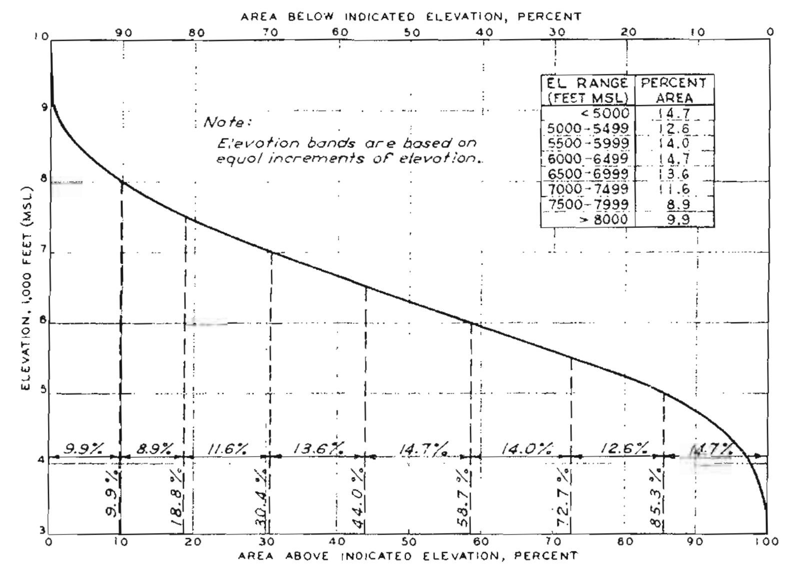
OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

PREPARED: C.E.J.	SUBMITTED: F.M.B.	TO ACCOMPANY REPORT DATED 30 JUNE 1956
DRAWN: B.V.	APPROVED: J.M.R.	PD-20-25/57



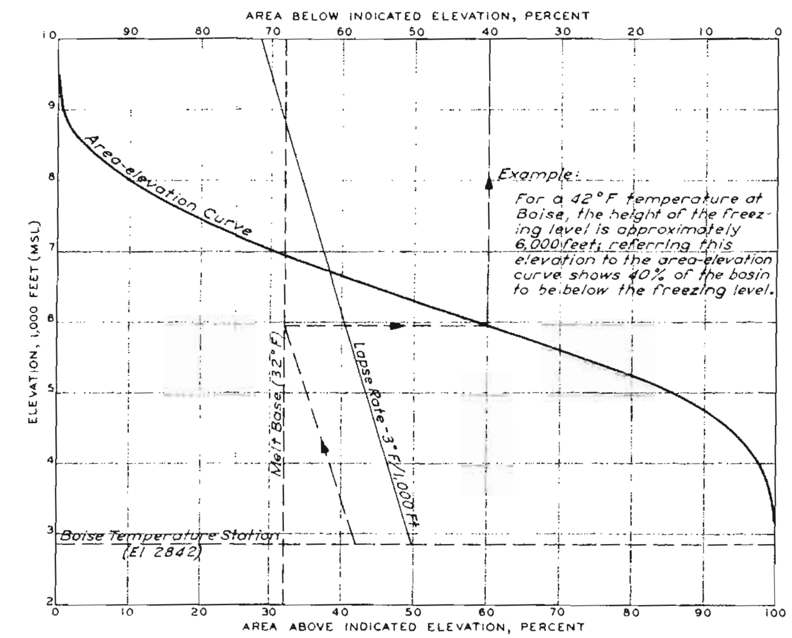
AREA-ELEVATION CURVE

FIGURE 1



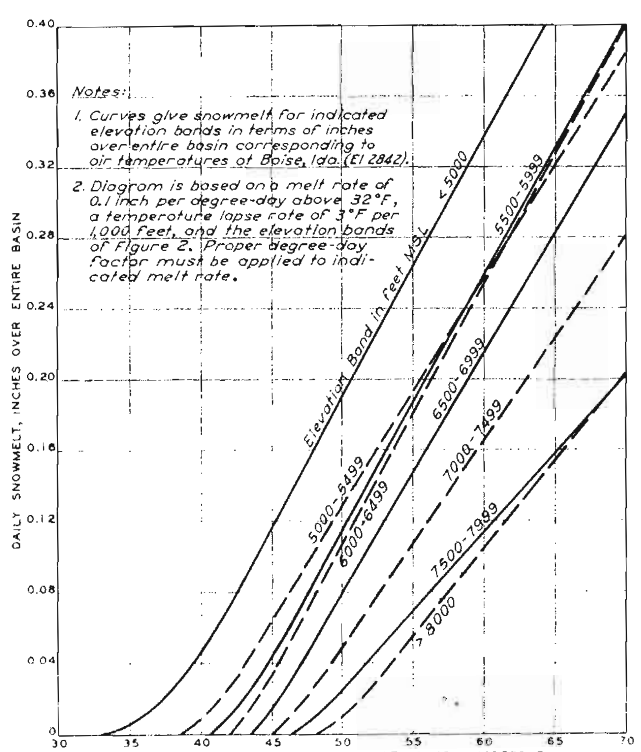
BASIN SUB-DIVISION INTO ELEVATION BANDS

FIGURE 2



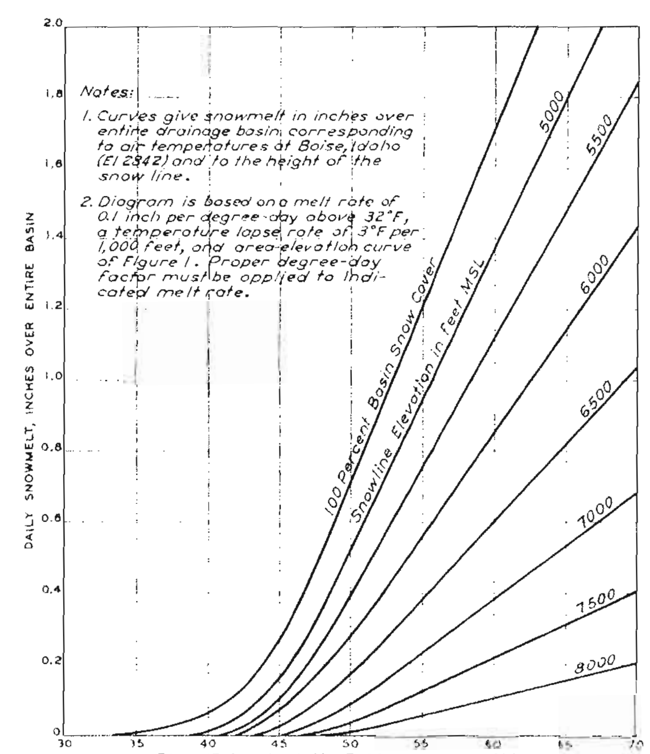
CONTRIBUTING AREA DIAGRAM

FIGURE 3



TEMPERATURE-SNOWMELT CURVES
FOR INDIVIDUAL ELEVATION BANDS

FIGURE 4



TEMPERATURE-SNOWMELT CURVES
FOR BASINWIDE SNOWMELT

FIGURE 5

SNOW INVESTIGATIONS
SUMMARY REPORT

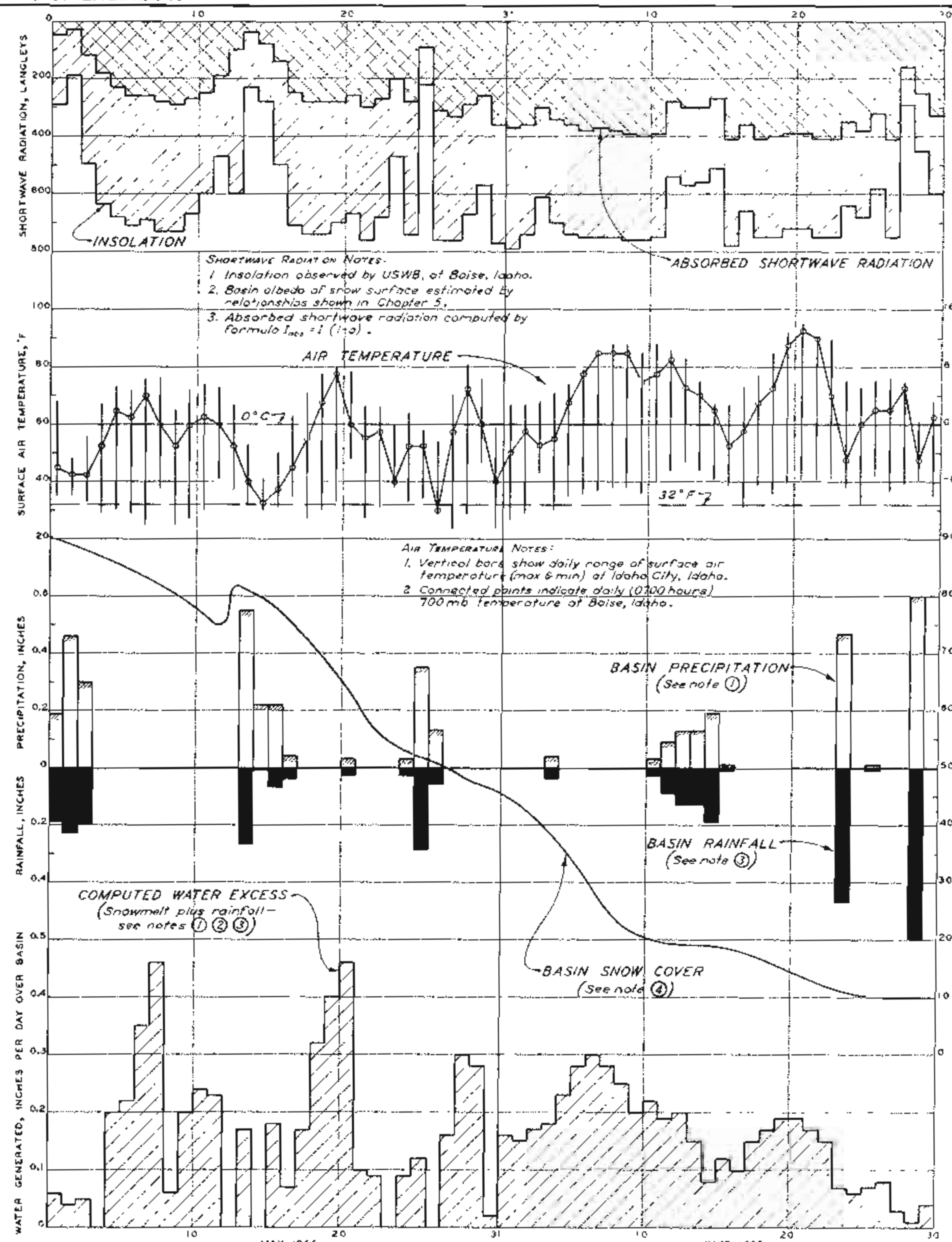
SNOW HYDROLOGY

TEMPERATURE-INDEX COMPUTATION OF SNOWMELT

BOISE RIVER BASIN ABOVE TWIN SPRINGS, IDAHO
DRAINAGE AREA 830 SQUARE MILES

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

PD-20-25/58



HYDROMETEOROLOGICAL DATA AND COMPUTATION OF WATER GENERATED

FIGURE 1

NOTES for FIGURE 1.—

① Basin precipitation computed as follows precipitation index equals sum of Arrowrock Dam, Idaho City, Obsidian (4 NNE), and two times Atlanta (1E) precipitations. Normal annual precipitation (NAP) for basin equals 33.3 inches, while precipitation index for station NAP equals 113.0 inches hence,

$$\text{Basin Precip} = \frac{13.0}{113.0}\ \text{Precip index}$$

$$= 0.30 \times \text{Precip index}$$

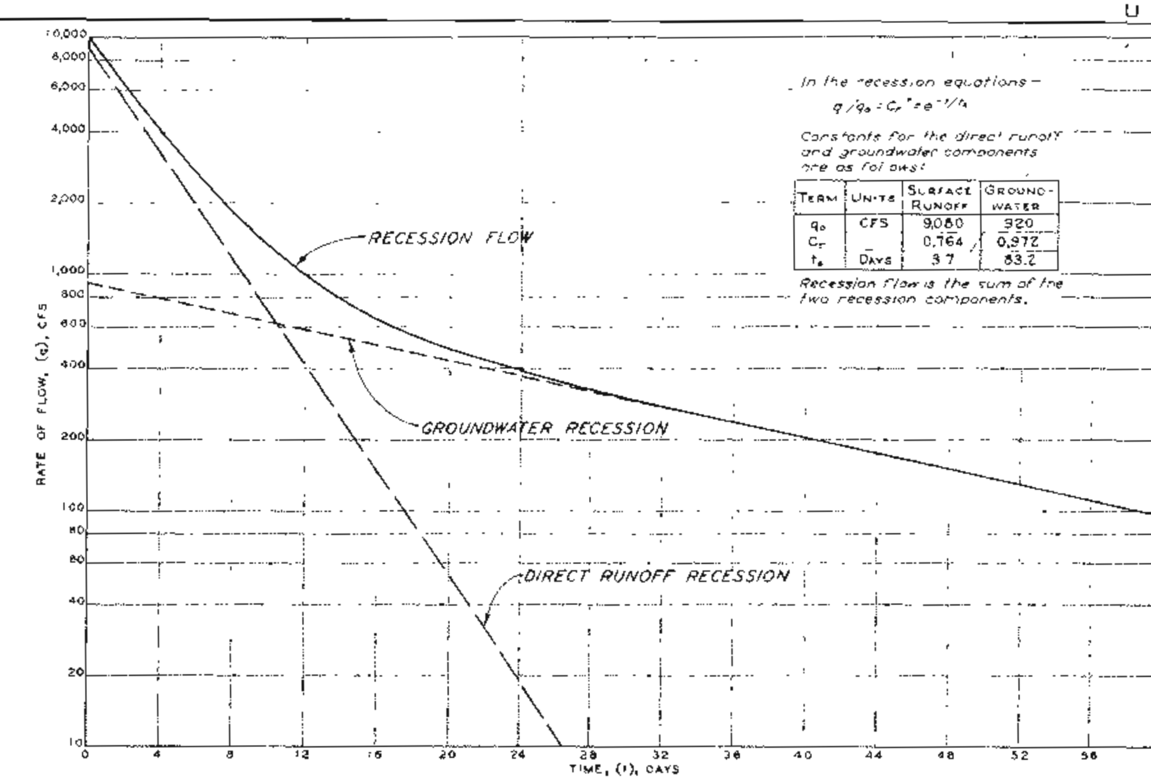
② Snowmelt excess computed from equation,

$$M = 0.0267\ T_{max} + 0.00227G - 2.00$$

where $\underline{M}$ is the snowmelt in inches over the basin, $\underline{T_{max}}$ is the daily maximum temperature of Idaho City, in degrees F, and $\underline{G}$ is a radiation parameter in langleys (equals absorbed shortwave radiation plus estimated longwave loss—longwave loss estimated from Boise 700 mb temperature and Idaho City minimum temperature using diagram of (Figure 2, Plate 6-9).

③ Rainfall computed from basin precipitation assuming lapse rate of 3 degrees F per 1,000 feet and using Idaho City mean daily temperature. (Precipitation in form of snow at temperatures less than 35 degrees F). Rainfall excess assumes all rain falling on bare ground is lost while 100 percent runoff results from snow-covered area.

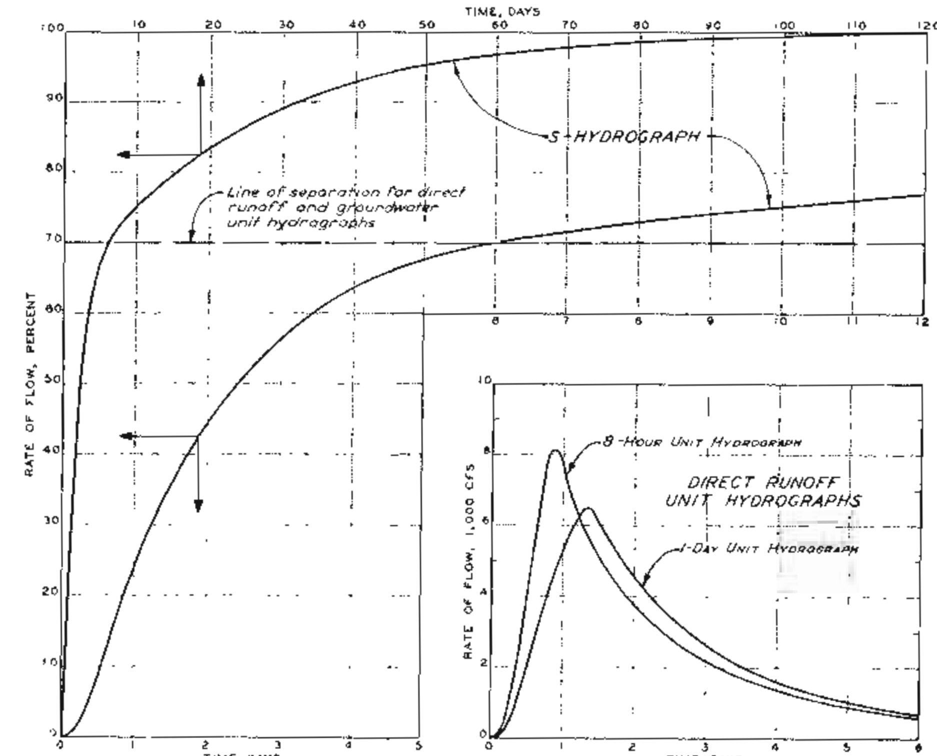
④ Basin snow cover obtained from observations by aerial reconnaissance, Walla Walla District Office. See Plates 7-5 and 7-6.



TERM	UNITS	SURFACE RUNOFF	GROUND-WATER
q_0	CFS	9,080	320
C_r	—	0.764	0.972
t_k	DAYS	3.7	83.2

RECESSION CURVES

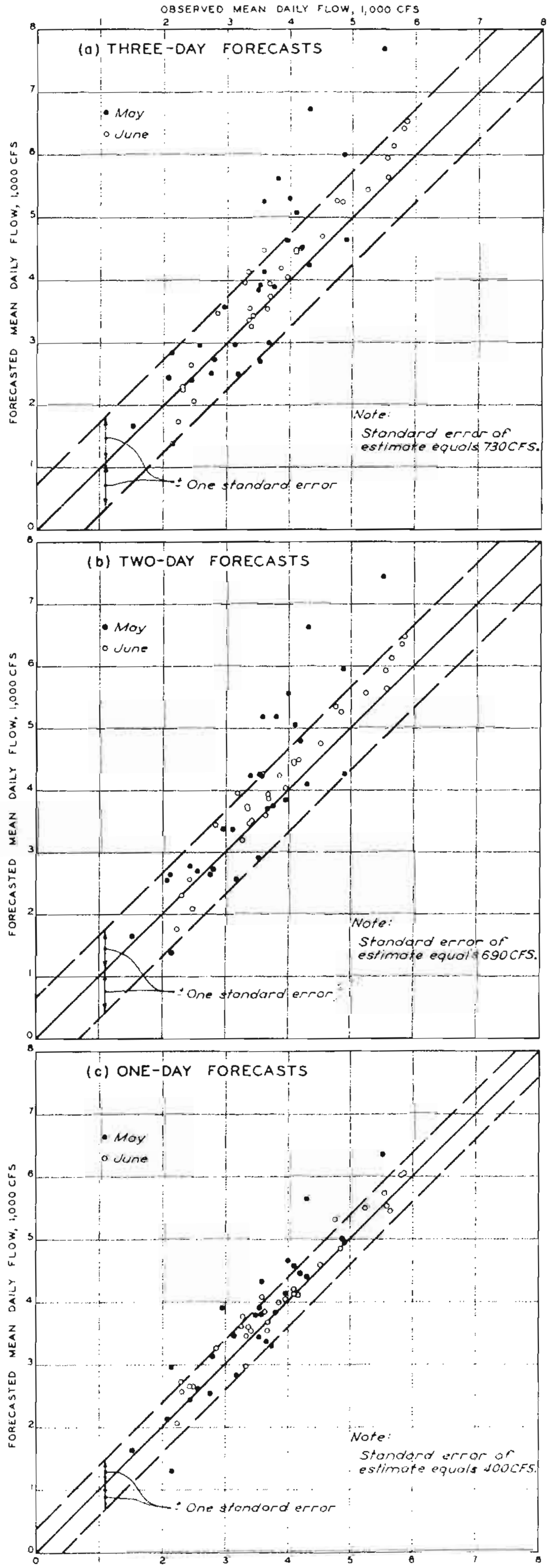
FIGURE 2



S-HYDROGRAPH AND UNIT HYDROGRAPHS

FIGURE 3

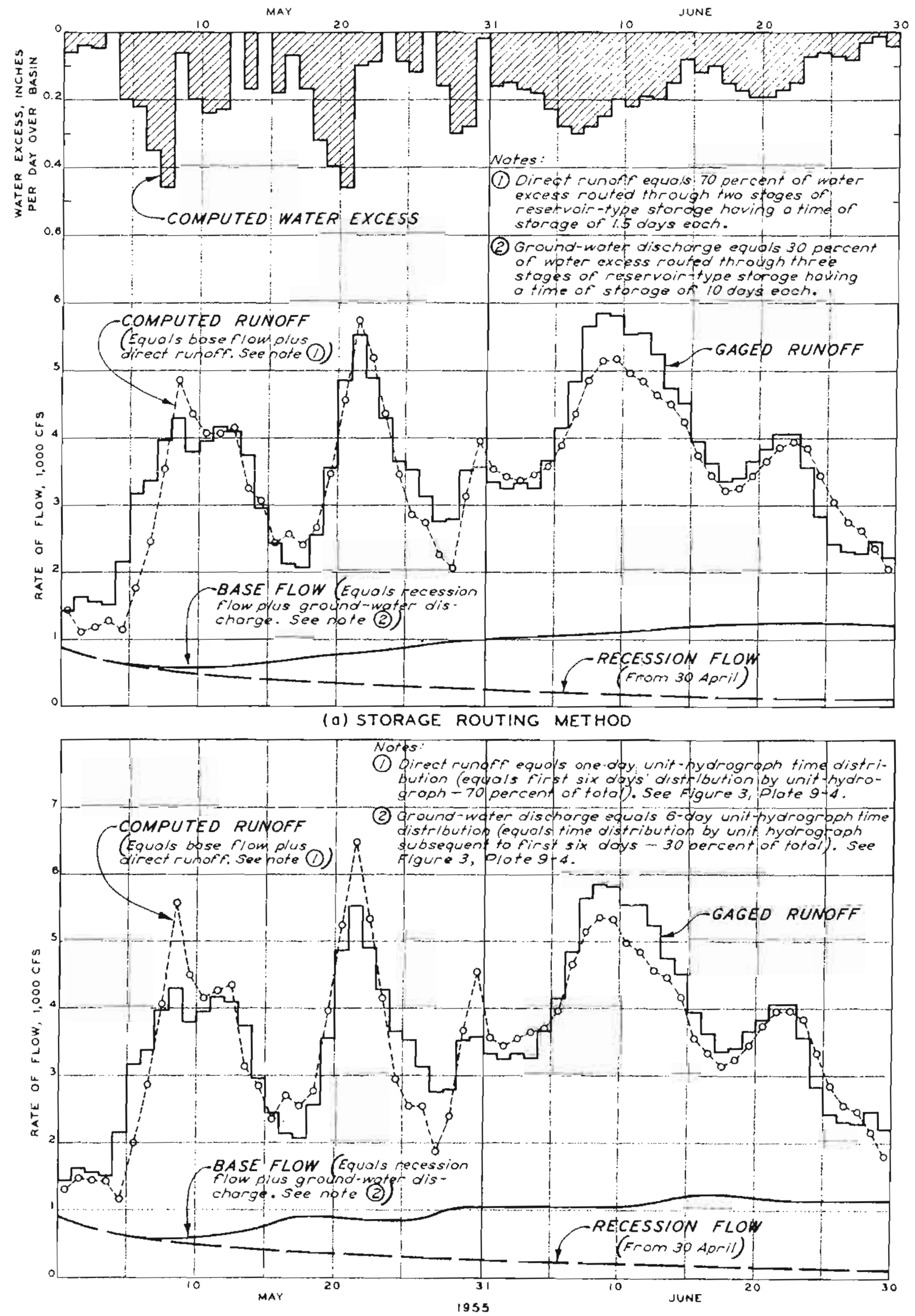
SNOW INVESTIGATIONS SUMMARY REPORT		
SNOW HYDROLOGY FLOW FORECASTS AND RECONSTITUTION BOISE RIVER NEAR TWIN SPRINGS, IDAHO 1955 SPRING SNOWMELT SEASON DRAINAGE AREA 830 SQUARE MILES SHEET 1 OF 2		
OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION CORPS OF ENGINEERS U.S. ARMY		
PREPARED C.E.H. DRAWN LV	SUBMITTED C.E.H. APPROVED D.M.R.	TO ACCOMPANY REPORT DATED 30 JUNE 1956 PD-20-25/59



MEAN DAILY FLOW FORECASTS

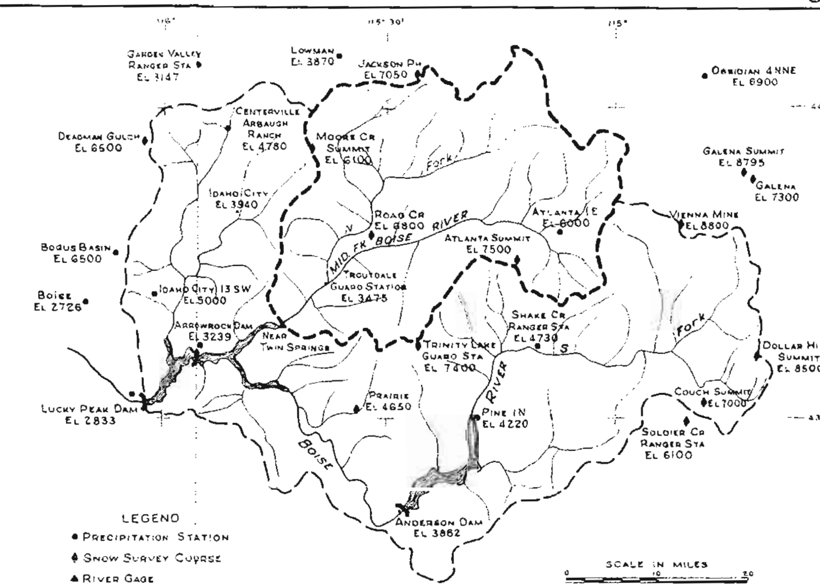
FIGURE 1

NOTE for FIGURE 1: One-, two-, and three-day forecasts computed from values of snowmelt and rainfall described on Figure 1, Plate 9-4, utilizing distribution graph derived from S-hydrograph shown on Figure 3, Plate 9-4, and total recession flow from date of forecast based on recession curve shown on Figure 2, Plate 9-4. Actual observed melt parameters and rainfall were used in melt equation.



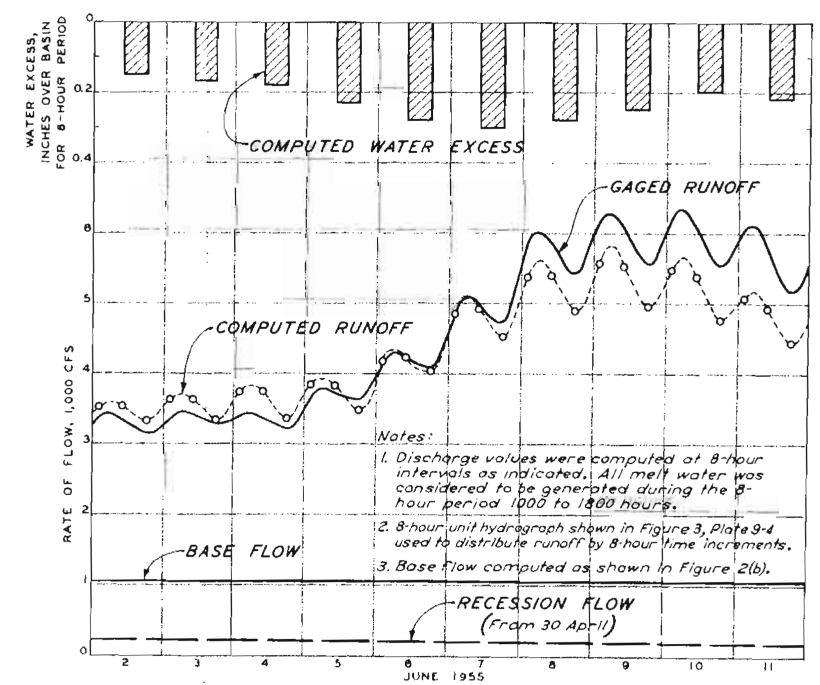
RECONSTITUTION OF SEASONAL HYDROGRAPH OF MEAN DAILY FLOWS

FIGURE 2



HYDROMETEOROLOGICAL STATIONS

FIGURE 3



RECONSTITUTION OF HYDROGRAPH SHOWING DIURNAL FLOW VARIATION

FIGURE 4

SNOW INVESTIGATIONS
SUMMARY REPORT
SNOW HYDROLOGY

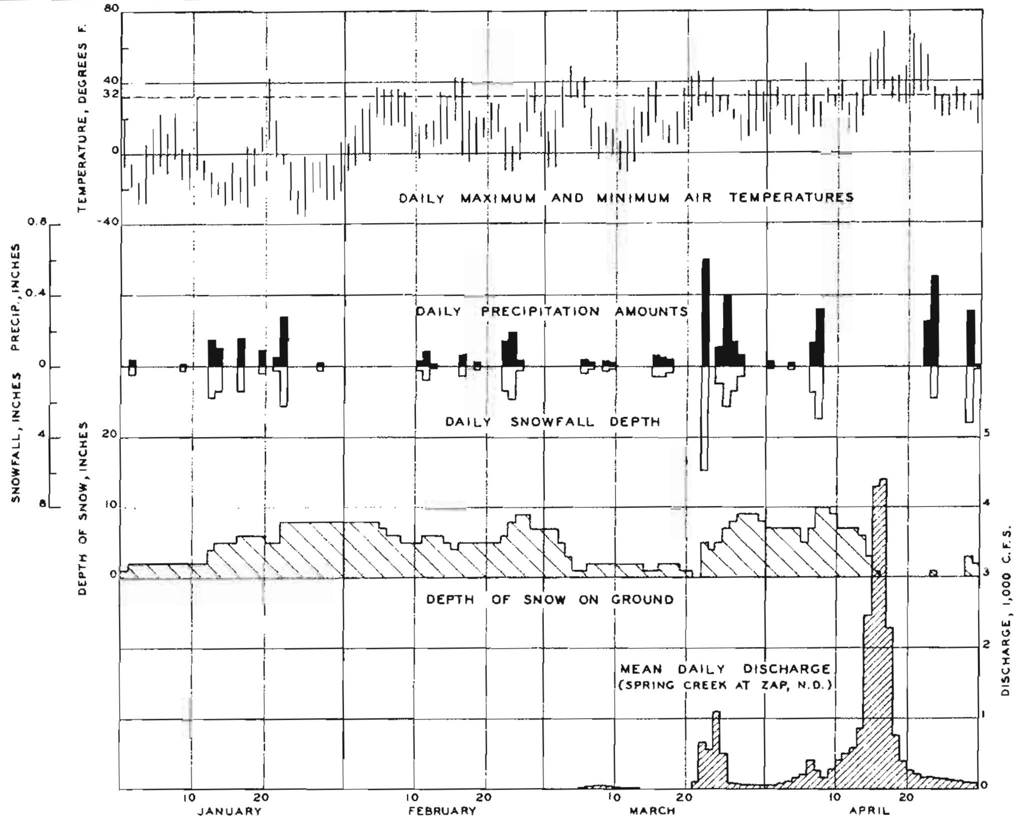
FLOW FORECASTS AND RECONSTITUTION

BOISE RIVER NEAR TWIN SPRINGS, IDAHO
1955 SPRING SNOWMELT SEASON
DRAINAGE AREA 830 SQUARE MILES

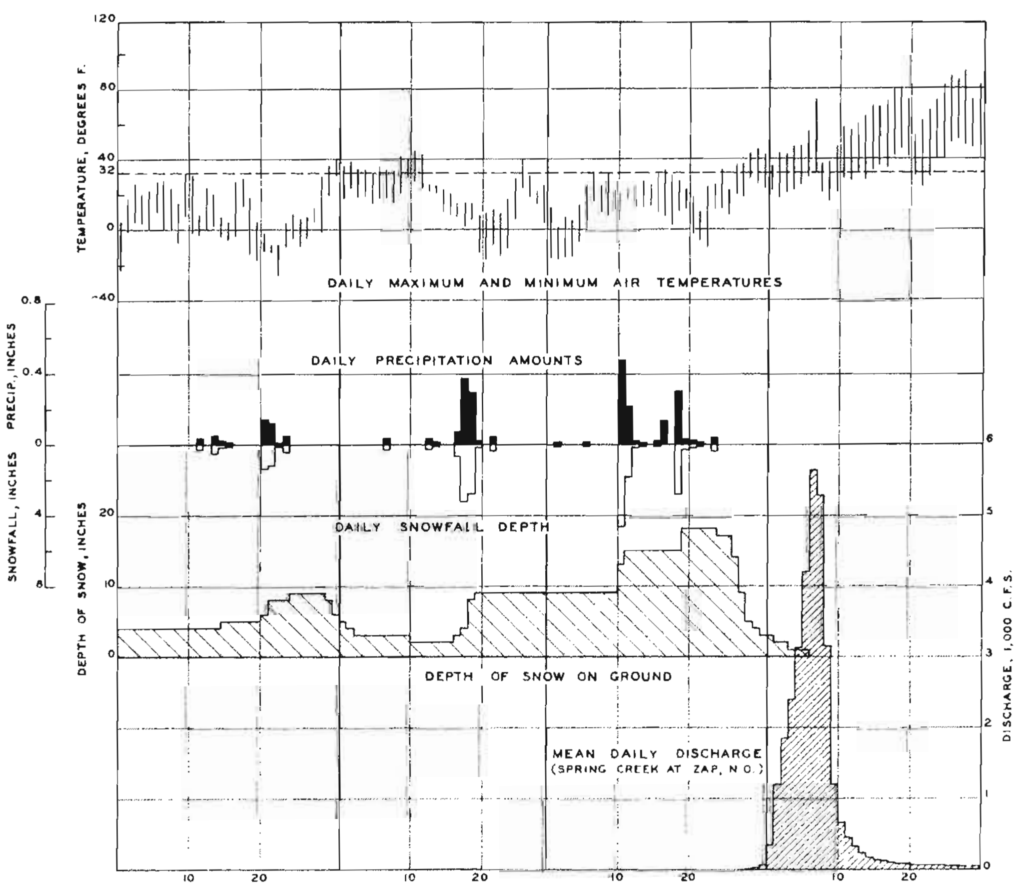
SHEET 2 OF 2

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U.S. ARMY

PREPARED: C.E.K.	SUBMITTED: C.C.H.	TO ACCOMPANY REPORT DATED 30 JUNE 1956
DRAWN: M.	APPROVED: D.M.R.	PD-20-25/60



1950 DATA

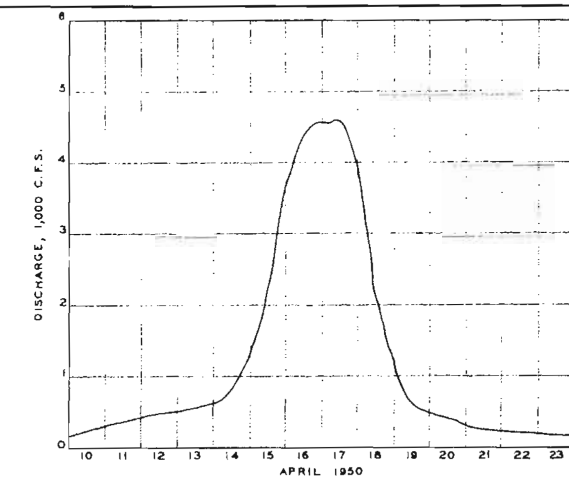


1952 DATA

HYDROMETEOROLOGICAL DATA

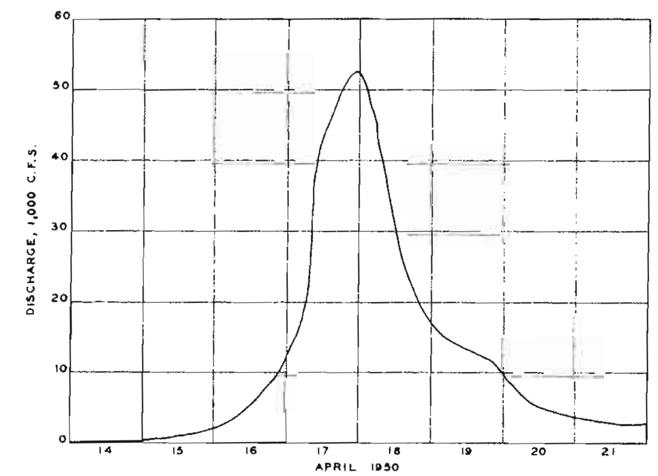
FIGURE 1

NOTE
METEOROLOGICAL DATA IN FIGURES ARE FOR DICKINSON CAA AIRPORT STATION (EL 2587)



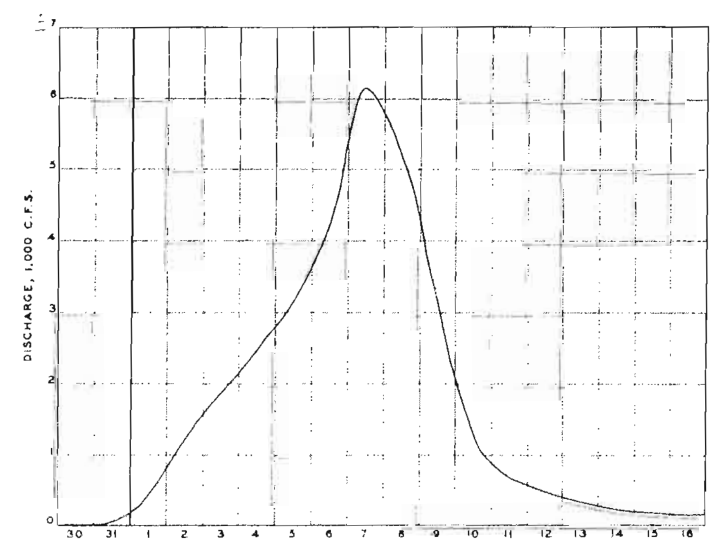
SPRING 1950 FLOOD HYDROGRAPH
FOR SPRING CREEK AT ZAP, N.D.
DRAINAGE AREA, 545 SQ. MI.

FIGURE 2



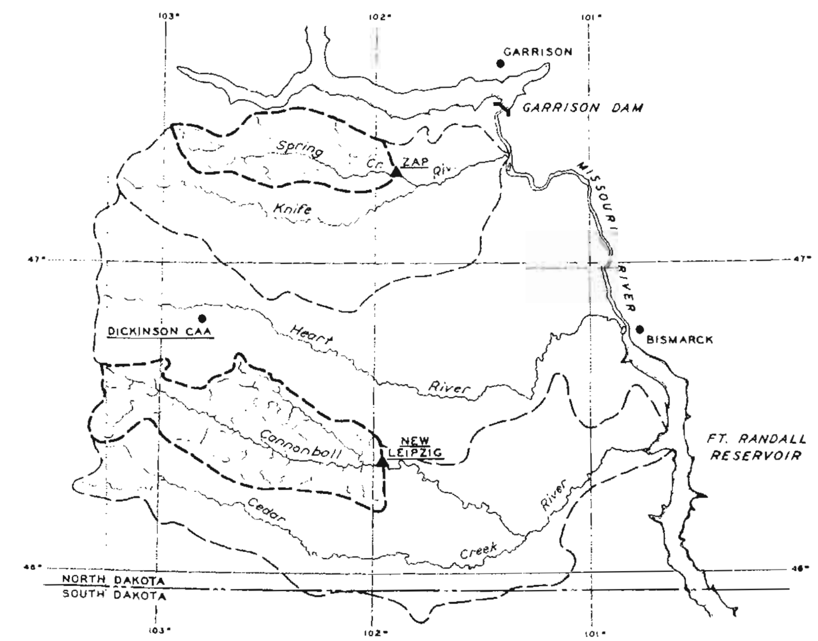
SPRING 1950 FLOOD HYDROGRAPH
FOR CANNONBALL RIVER NEAR NEW LEIPZIG, N.D.
DRAINAGE AREA, 1,160 SQ. MI., APPROXIMATELY

FIGURE 3



SPRING 1952 FLOOD HYDROGRAPH
FOR SPRING CREEK AT ZAP, N.D.
DRAINAGE AREA, 545 SQ. MI.

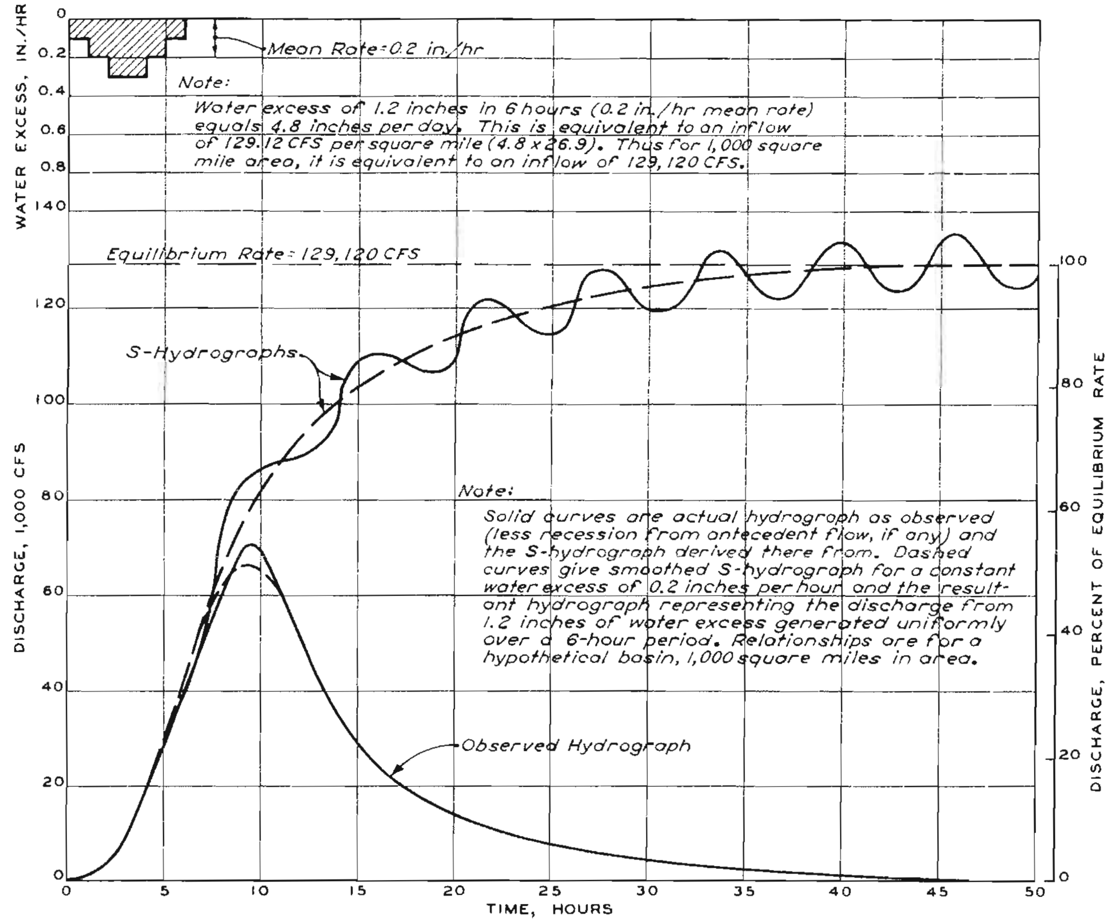
FIGURE 4



LOCATION MAP

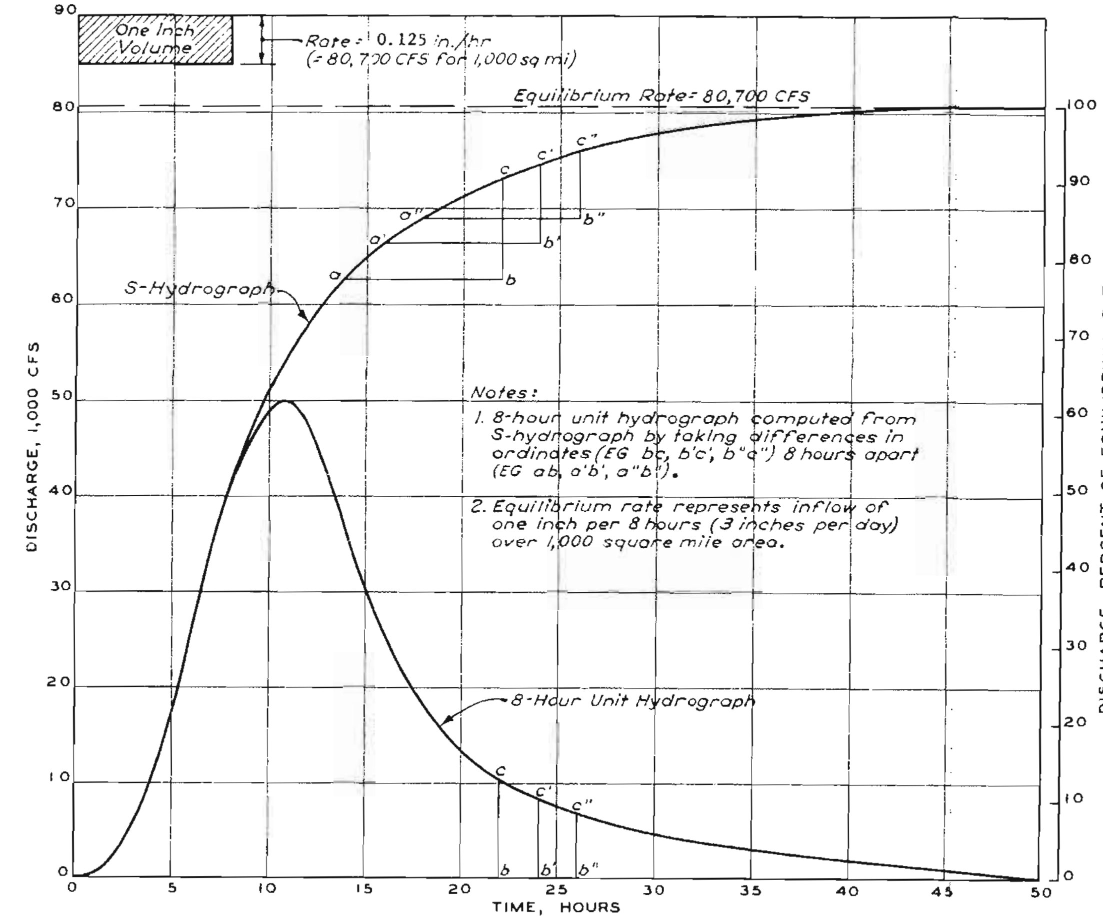
SCALE IN MILES

SNOW INVESTIGATIONS SUMMARY REPORT	
SNOW HYDROLOGY	
SNOWMELT FLOODS IN GREAT PLAINS AREA	
OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION CORPS OF ENGINEERS U. S. ARMY	
PREPARED C.E.H. / DRAWN H.E.B. / SUBMITTED C.E.H. / APPROVED D.M.R.	TO ACCOMPANY REPORT DATED 1 JUNE 1956
	PD-20-25/61



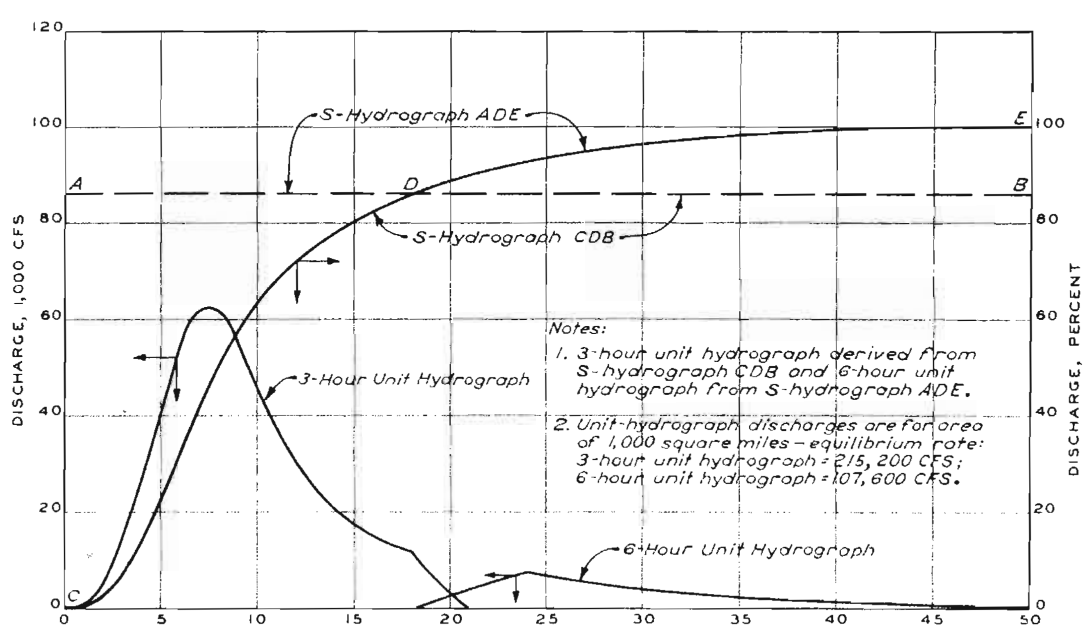
S-HYDROGRAPH — UNIT-HYDROGRAPH RELATIONSHIPS

FIGURE 1



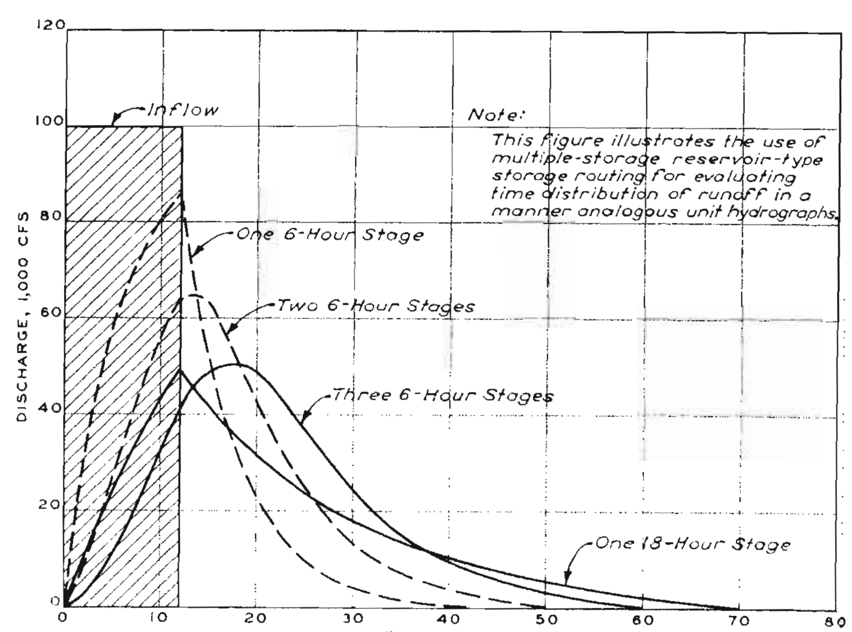
DERIVATION OF UNIT HYDROGRAPH FROM S-HYDROGRAPH

FIGURE 2



DERIVATION OF UNIT HYDROGRAPHS HAVING DIFFERENT PERIODS FROM A DIVIDED S-HYDROGRAPH

FIGURE 3



EXAMPLE OF MULTIPLE-STAGE RESERVOIR-TYPE STORAGE ROUTING

FIGURE 4

SNOW INVESTIGATIONS SUMMARY REPORT

SNOW HYDROLOGY

ILLUSTRATIVE DIAGRAMS OF TIME DISTRIBUTION OF RUNOFF

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION CORPS OF ENGINEERS U.S. ARMY

PREPARED: G.E.M.	SUBMITTED: G.E.M.	TO ACCOMPANY REPORT DATED 30 JUNE 1956
DRAWN: R.V.	APPROVED: D.M.R.	PD-20-25/62

CHAPTER 10 - DESIGN FLOOD DETERMINATION

10-01. INTRODUCTION

10-01.01 General. - No general all-inclusive rules of universal applicability can be given for use in hydrologic design. Every basin, every stream, is an individual and separate problem unique in its flood-producing characteristics. Each requires careful study to establish hydrometeorological relationships by which estimates of probable optimum conditions can be translated into the rates of streamflow (or volume of runoff) for the several different design requirements. Optimum conditions of weather, ground, and snowpack must be considered in combination to arrive at estimates of the basic flood magnitudes which form the basis of design of projects. Observed floods usually reflect compensating variations in the several factors affecting flood runoff, so that the runoff rates and volumes are far below those that would result from more critical combinations of the factors. Statistical studies provide a means of estimating the magnitude of flood potential and average flood frequencies for streams having relatively long periods of record, particularly where records of flow for many streams in a region of reasonably comparable hydrologic and meteorologic influences can be analyzed. However, because of the number and range of variation in independent variables involved in floods, and the wide range between flood magnitudes that would result from optimum combinations of critical flood-producing factors as compared with combinations generally observed, statistical analyses of actual stream flow records seldom, if ever, provide a reliable indication of extraordinary flood potentialities of a specific drainage basin.

10-01.02 Basic flood estimates. - In Corps of Engineers practice, there are two classes of floods for which hydrographs are usually synthesized: (1) maximum probable flood, which is used primarily for the design of spillways and appurtenant structures for virtually complete security of major projects against structural failure, and is defined as the flood discharge that may be expected from the most severe combination of critical meteorologic and hydrologic conditions that are reasonably possible in the region; and (2) standard project flood, which represents a "standard" against which the degree of protection finally selected for a project may be judged and which thus will serve as a basis for comparison with protection provided at similar projects in other localities. The standard project flood is defined as the flood discharge that may be expected from the most severe combination of meteorologic and hydrologic conditions that are considered reasonably characteristic of the geographical region involved, excluding extremely rare combinations. The standard project flood is based on less severe conditions than the maximum probable flood; in practice it has been found to be generally equal to 40 to 60 percent of the maximum probable flood for the same basins.

10-01.03 Design Flood. - The term design flood has been applied to the flood volume or peak discharge finally adopted for which full protection is being provided in a particular project or section thereof. It may be either greater or less than the basic flood estimate, depending to an important extent upon flood characteristics, frequencies, and potentialities, and upon economic factors and other practical considerations. The preceding definitions of basic and design floods have been summarized from Civil Engineer Bulletin No. 52-8. 2/ More complete listings and definitions of design criteria for these floods may be found therein.

10-01.04 The rational procedure. - The principal factors to be considered in determining the magnitude of design floods are discussed in chapter 5, "Flood-hydrograph analyses and computations," of Part CXIV of the Engineering Manual for Civil Works Construction. 3/ The rational procedure involves consideration of the optimum meteorologic and hydrologic conditions which are likely to occur simultaneously to produce maximum runoff. Rational determinations of design floods involving snow require knowledge and use of the combined effect of snow accumulation, snowmelt, and effect of the snowpack on runoff, as described in the preceding chapters. In general, the limited period of record of snow accumulation data precludes their use for application to design floods. The estimate of snow accumulation for a given design condition may, however, be based on a function of normal annual precipitation for cases where winter precipitation is nearly all in the form of snow. Extrapolation of precipitation to design condition amounts is possible because of the many years of record of storm experience which are usually available. Snowmelt is determinable from thermal-budget indexes appropriate to the type of area and the specific flood condition. The melt coefficients may be derived from historical data or from generalized melt equations as presented in chapter 6. Since snowmelt is a direct function of thermal energy input to a basin, there are definite upper limits to the amount of heat exchange that may be experienced by radiative processes or by advection of heat by airmasses. The effect of the snowpack on runoff varies through a wide range of conditions; the pack may be initially "primed" or "ripe" (conditioned to produce runoff); or it may be initially "cold" and dry. In addition to the effects of snow in the development of rationally derived design floods, other hydrologic factors must be evaluated, including rainfall, soil moisture recharge, ground water condition, and evapotranspiration loss. Also, the natural storage time of the basin as expressed by such standard routing techniques as unit graphs or storage-routing procedures must be determined. Having determined the basin hydro-meteorological characteristics, it is then possible to maximize each variable on the basis of optimum runoff-producing conditions, combined with the optimum meteorological sequence, for the specified condition of design.

10-01.05 Simplified design-flood estimates. - In some cases preliminary estimates of design floods or flood estimates may be required for minor engineering works which do not warrant a complete flood analysis by the rational procedure outlined in the preceding paragraph. For such determinations, the judgment and experience of the hydrologist is relied upon; short cuts and subjective analysis of the factors affecting runoff are used. Previously derived floods for areas of similar hydrologic character may be used as guides, as for example charts 1 and 2 contained in Appendix "M", Columbia River and Tributaries, 6/ which show curves for estimating spillway-design-flood peak discharges resulting from snowmelt in the Columbia River basin on the basis of drainage area and normal annual precipitation. Many factors affecting runoff are not directly evaluated by these curves. For this reason, it is especially important that the hydrologist have a full understanding of the basin to which they are applied, in order to account for differences in conditions from those for which the curves were derived. Estimates so derived should be considered to be preliminary and/or approximate, subject to revision when and if more complete analysis is warranted. The use of historical streamflow data alone should not be considered as a basis of derivation for design-flood determinations (e.g., the arbitrary use of a multiplication factor applied to the maximum flood of record). For the short period of streamflow records normally available, there is little likelihood that a constant relationship between the maximum flood of record and a specific design condition exists.

10-01.06 Design floods involving snow. - There are two general types of design floods involving snow: (1) winter rain-on-snow floods, which are of relatively short duration and for which snowmelt usually constitutes the minor contribution to runoff; and (2) spring snowmelt floods, which are the result of the melting of the accumulated snowpack, are usually several months in duration, and for which rainfall is usually of lesser consequence. For both types of floods, the snowpack accumulation, snowpack condition, and snowmelt rates must be evaluated, plus all other factors affecting runoff.

10-01.07 Factors in design flood derivation. - In the rational derivation of design floods involving snow, certain general procedures should be outlined and certain basic factors evaluated before detailed studies are begun. They are summarized as follows:

I. Review of general hydrologic features

A. Location of drainage area with respect to major topographic features, airmass types, and general airmass circulation during storms and periods of melt.

B. Physical characteristics of the watershed.

1. Drainage area.
2. Area-elevation relationship.
3. Normal annual basin precipitation.
4. Normal annual runoff.
5. Normal annual loss.
6. Normal snowpack accumulation and seasonal distribution.
7. Soil conditions and seasonal change in soil moisture.
8. Ground water geology and ground water storage.
9. Vegetative cover.
10. Artificial regulation of streamflow.
11. Streamflow characteristics from analysis of past record.
12. Natural basin storage time, with or without snow cover, expressed by unit hydrograph or storage-routing constants.

II. Evaluation of specific conditions pertinent to winter rain-on-snow design floods, according to established design criteria.

A. Initial snowpack characteristics.

1. Snow-covered area.
2. Snowpack depth and water equivalent and distribution with respect to elevation (slope of snow wedge).
3. Snowpack condition with respect to temperature, free water, and density-elevation variation.

B. Determination of sequence of meteorological factors affecting melt.

C. Selection of snowmelt rates (snowmelt indexes or generalized snowmelt equations appropriate to area and rain-on-snow conditions).

D. Determination of rainfall.

1. Time distribution.
2. Total amount.

E. Determination of loss and runoff conditions.

F. Synthesis of all factors affecting runoff into a design flood hydrograph.

III. <u>Evaluation of specific conditions pertinent to spring snowmelt design floods, according to established design criteria</u>.

A. Initial snowpack characteristics.

1. Snow-covered area.
2. Snowpack water equivalent and distribution with respect to elevation (slope of snow wedge).
3. Albedo of snow surface (for areas with significant open areas).

B. Determination of critical sequence of meteorological factors affecting melt.

C. Selection of snowmelt rates utilizing thermal-budget indexes or generalized snowmelt equations appropriate to area.

D. Evaluation of effects of rainfall at time of maximum snowmelt flood, considering changes in snowmelt conditions during rain.

E. Determination of loss and runoff conditions.

F. Synthesis of all factors affecting runoff into a design flood hydrograph.

10-02. OPTIMUM CONDITIONS FOR DESIGN FLOODS

10-02.01 General. - In the derivation of design floods it is necessary to consider the optimum runoff-producing conditions, with regard to: (1) the snowpack; (2) the meteorological sequence affecting melt and rainfall; (3) the effect of losses to soil moisture and evapotranspiration; (4) changes in ground-water storage; and (5) time delay to runoff. The following paragraphs describe the derivation of optimum runoff conditions for maximum probable and standard project floods, in connection with both winter- and spring-type floods.

10-02.02 Optimum snowpack conditions. - The three basic considerations of optimum snowpack condition are (1) water equivalent and its distribution, (2) areal cover, and (3) structural character. For spring snowmelt design-flood hydrographs, the structural character of the snowpack is unimportant (it is assumed the snowpack is isothermal at 32°F and saturated with free water). Generally, only the total water equivalent of the snowpack and its distribution with elevation and area must be considered; for basins with significant open areas, snow-surface albedo must also be evaluated. For winter rain-on-snow floods, however, the stage of metamorphism of the snowpack must be taken into account as set forth in chapter 8. For winter floods, the total water equivalent of the snowpack may not be critical. The principal consideration for winter rain-on-snow floods is that possible storage of liquid water in the snowpack must be satisfied before runoff occurs.

10-02.03 For spring snowmelt design floods, the maximum possible snowpack water equivalent is generally based upon detailed studies of the potential total winter-season precipitation, with assumed percentages of total winter precipitation falling in the form of snow. The studies may relate maximum winter-season precipitation to size of drainage area and normal annual precipitation, as was done for the Columbia River basin by the Hydrometeorological Section of the U. S. Weather Bureau. 7/ From such studies the maximum winter snowfall for specific basins may be derived. The increase of the snowpack with elevation is determined on the basis of normal increase of precipitation with elevation. The snow wedge so derived represents the maximum possible flood-producing snowpack. For standard project flood conditions, the snowpack water equivalent determination is based on less severe conditions than that for the maximum possible and conforms to the maximum which is reasonably characteristic of the region involved.

10-02.04 The initial snowpack condition for winter rain-on-snow floods is important both from the consideration of snowmelt and for storage and delay of liquid water in the snowpack. For maximum probable rain-on-snow flood conditions, in some cases it may be assumed that sufficient water equivalent exists to provide snowmelt continuously through the storm period throughout the entire range of elevation. In other cases, a derived maximum snow wedge is required. Also for the maximum probable flood, it may be assumed in most cases that the preceding melt or rainfall has provided drainage channels through the snowpack and has conditioned it to produce runoff without significant delay, so that water excesses from rain and snowmelt during the storm period are immediately available for runoff. In unusual circumstances, however, especially where a significant portion of the basin is at high elevations, it may be necessary to ascribe snowpack storage and delay to a portion of the water excess (see discussion of liquid-water-holding capacities of the snowpack in section 8-05.). Evaluation of this snowpack condition may be established on the basis of meteorological events preceding the design storm. For standard project flood determinations, the storage and delay of liquid water in the snowpack should be evaluated for all ranges in elevation, based on preceding meteorological events. The maximum-runoff condition in this case is one where there is (1) sufficient snow on the basin initially to provide melt contribution to runoff over the entire area for the storm period, and yet (2) a minimum depth of snow, especially at high elevations, to provide the least possible storage required in conditioning the snowpack to produce runoff. Thus the flattest possible snow wedge having sufficient snow to just equal the total melt at the lowest elevation in the basin, is the optimum condition for winter rain-on-snow floods.

10-02.05 Optimum meteorological conditions. - Meteorological conditions during both the pre-flood and flood periods affect design floods involving snow. Pre-flood conditions determine the snowpack soil-moisture and ground water conditions, as well as the recession flow. Rates of snowmelt and rates of rainfall during the flood period are governed by meteorological conditions. For spring snowmelt floods, the optimum condition is that in which winter snowpack accumulation occurs with no significant melt, followed by a cold spring with minimum snowmelt and continued increase in the snowpack, and finally by a sudden change to a sustained high heat input to the basin at a time when the seasonal energy input may be near maximum. Rainfall occurring near the snowmelt peak may be superimposed upon the critical snowmelt sequence to augment the maximum probable flood peak discharge. For standard project flood conditions, a similar but less severe sequence of snowmelt conditions may be assumed, which would be reasonably characteristic of the maximum for the region involved.

10-02.06 During winter rain-on-snow design floods, the optimum meteorological sequence for the maximum probable flood requires sufficient water equivalent accumulation in the pre-flood period to provide active snowmelt for the entire flood period accompanied by heat supply and rainfall sufficient to condition the pack for runoff prior to the occurrence of the design storm. During the design storm period, maximum possible snowmelt rates commensurate with the meteorological conditions accompanying the rainfall are assumed. For standard project floods, the pre-flood meteorological sequence must be carefully analyzed, to determine the initial snowpack condition. Air temperatures may be such that part of the precipitation falling during this period will be in the form of snow in the higher elevations, and part in the form of rain in the lower areas. Separation of these effects must be made in order to arrive at a reasonable snowpack condition for the basin as a whole. During the period of the design storm, snowmelt rates are assumed which are reasonably near maximum for the region considering the meteorological conditions accompanying the rainfall.

10-02.07 Meteorological factors which are pertinent to the computation of snowmelt for design floods are subdivided as follows:

Type of area	Spring snowmelt design flood	Winter rain-on-snow design flood
Open	Incident radiation Air temperature Dewpoint temperature Wind speed Cloud cover	Air temperature* Wind speed Rainfall
Partly forested	Incident radiation Air temperature Dewpoint temperature Wind speed	Air temperature* Wind speed Rainfall
Heavily forested	Air temperature Dewpoint temperature	Air temperature* Rainfall

* Air temperature function accounts for condensation melt under a saturated air condition.

The meteorological factors shown in the above tabulation appropriate to the type of area and design flood should be considered in setting up optimum meteorological conditions for determining snowmelt for design flood synthesis.

10-02.08 <u>Optimum ground conditions</u>. - Evaluation of loss through the processes of soil-moisture and ground-water recharge must be made for design-flood determinations. For spring snowmelt floods, the soil-moisture deficit from the preceding summer season must be assumed at the beginning of the accumulation of winter precipitation. Usually this amount is assumed to be equal to the difference between the wilting point and field moisture capacity for the average basin soil mantle. Part or all of this deficit will be satisfied by fall rains and minor melting of the snowpack during the winter. For winter rain-on-snow floods, soil-moisture deficits are usually assumed to be satisfied by snowmelt or rainfall prior to the occurrence of the design storm. In the case of standard project floods, a less severe runoff assumption as to loss by soil-moisture requirements is made, depending upon conditions which may reasonably prevail over the basin area. For cases where shallow snow depths and low temperatures prevail prior to the design storm, it is possible to have solidly frozen ground which would prevent any loss of water to the soil and also provide less delay to water in transit than occurs with unfrozen ground. With a deep snowpack, however, there is generally sufficient flow of ground heat to keep the soil unfrozen, regardless of the air temperature above the snow.

10-02.09 Ground-water recharge may be accounted for by the separation of flows through streamflow recession analysis as explained in chapter 4. Transitory storage in the soil and ground results in time delay to runoff, which may be accounted for by unit graph or storage routing techniques, as explained in chapter 9. For design flood computations, minimum time delay to runoff commensurate with the design criteria and basin characteristics is assumed, thereby maximizing peak flow conditions.

10-02.10 Evapotranspiration and interception loss. - Spring snowmelt design floods must account for loss of water by evapotranspiration to the atmosphere. During the snow accumulation season, there is a small loss by evaporation from the snow surface and transpiration from the forest. Under assumptions of maximum snow accumulation, however, air temperatures would be low, and these amounts would be negligibly small. Loss by interception can be estimated as a constant percentage of the precipitation. During the snowmelt season, the energy consumed in the evapotranspiration process is directly proportional to the energy used in melting the snowpack; therefore, the loss by evapotranspiration can be considered as a fixed percentage of the snowmelt for the snow-covered portions of the basin. For winter rain-on-snow floods, evapotranspiration loss is negligible during the storm period.

10-03. COMPUTATION OF SNOWMELT FOR DESIGN FLOODS

10-03.01 General. - Synthesis of design floods requires (1) the determination of the optimum meteorological flood-producing sequence, and (2) the use of snowmelt equations to compute the snowmelt runoff (as outlined in chaps. 5 and 6). The meteorological factors pertinent to such design-flood snowmelt computations differ according to varying forest cover, and are listed in paragraph 10-02.07. The necessary snowmelt equations may be derived from a rational analysis of historical records of the particular basin under consideration by use of the thermal-budget index technique (as explained in chap. 6) and tested by reconstitution of historical flood hydrographs (as shown in chap. 9). For cases where it is impossible or impractical to derive particular basin melt coefficients, the generalized snowmelt equations listed in section 6-07 may be applied. As indicated in section 9-02, there are two general procedures for computing runoff from snow-covered areas, depending upon the manner in which elevation effects are handled. The basin may be either (1) subdivided into elevation bands or (2) treated as a whole, making corrections for non-snow-covered areas and other non-contributing areas. For the computation of snowmelt, the first method requires the application of appropriate generalized snowmelt equations, while for the second, either generalized snowmelt equations or particular basin melt coefficients derived from historical record may be used.

10-03.02 Snowmelt during winter rain-on-snow design floods. - Having adopted the optimum weather and basin conditions for design and the method of subdividing the watershed, the snowmelt portion of winter rain-on-snow design floods may be determined from the following general equations described previously in chapter 6:

Open or partly forested area:

$$M = (0.029 + 0.0084kv + 0.007P_r)\ (T_a - 32) + 0.09 \qquad (10\text{-}1)$$

Heavily forested area:

$$M = (0.074 + 0.007P_r)\ (T_a - 32) + 0.05 \qquad (10\text{-}2)$$

where M is the total daily snowmelt in inches per day, T_a is the temperature of air (assumed to be saturated) at the 10-foot level in °F, P_r is the daily rainfall in inches, v is the wind speed at the 50-foot level in miles per hour, k is the basin convection-condensation melt factor expressing the relative exposure of the area to wind and is affected principally by forest cover. The value of k is 1.0 for plains areas with no forest cover. It may be slightly greater than 1.0 for exposed ridges and mountain passes, and for heavily forested areas it approaches a minimum value of about 0.2. The 50-foot level wind value for forested areas is assumed as the average wind in an open area resulting from the general airmass circulation prevailing at the time. The constants 0.09 and 0.05 represent average maximum daily melt under rain-on-snow conditions, which would result from absorbed shortwave radiation and ground heat. For heavily forested areas such as WBSL, it has been shown that wind is damped out to a great extent and that heat transfer by convection and condensation may be expressed by an average constant wind, so that wind variation need not be considered. The melt equation for rain-on-snow conditions in heavily forested areas involves only air temperature and rainfall intensity. The above equations are for saturated air conditions, and assume linear variation between dewpoint temperature and saturation vapor pressure.

10-03.03 Design-flood snowmelt during rain-free periods. - Computation of snowmelt for design floods during rain-free periods, which is generally required for spring snowmelt-type floods, is somewhat more complex than that for rain-on-snow type floods. Because of the variation in dewpoint, radiation exchange, and cloud cover, clear-weather melt cannot always be expressed by the simple temperature functions used during rain periods, especially for open or partly forested areas. Reference is again made to chapter 6, for a discussion of the generalized snowmelt equations applicable to clear-weather (rain-free) melt periods, and the equations are repeated below for use in design-flood derivation.

Heavily forested area:

$$M = 0.074\ (0.53T'_a + 0.47T'_d) \tag{10-3}$$

Forested area:

$$M = k(0.0084v)\ (0.22T'_a + 0.78T'_d) + 0.029T'_a \tag{10-4}$$

Partly forested area:

$$\begin{aligned} M &= k'(1 - F)(0.0040\ I_i)\ (1 - a) \\ &+ k(0.0084v)(0.22T'_a + 0.78T'_d) + F(0.029T'_a) \end{aligned} \tag{10-5}$$

Open area:

$$\begin{aligned} M &= k'(0.00508\ I_i)\ (1 - a) + (1-N)(0.0212T'_a - 0.84) \\ &+ N(0.029T'_c) + k(0.0084v)\ (0.22T'_a + 0.78T'_d) \end{aligned} \tag{10-6}$$

where:

M is the snowmelt rate in inches per day.

T'_a is the difference between the air temperature measured at 10 feet and the snow surface temperature, in ^{o}F.

T'_d is the difference between the dewpoint temperature measured at 10 feet and the snow surface temperature, in ^{o}F.

v is the wind speed at 50 feet above the snow, in miles per hour.

I_i is the observed or estimated insolation (solar radiation on horizontal surface) in langleys. (See plates 5-1 and 6-1)

a is the observed or estimated average snow surface albedo. (See figures 3-4, plate 5-2 for estimating albedo of the snow.)

k' is the basin shortwave radiation melt factor. It depends upon the average exposure of the open areas to shortwave radiation in comparison with an unshielded horizontal surface. (See figure 6, plate 5-1, for seasonal variation of k' for North and South 25° slopes).

F is an estimated average basin forest canopy cover, effective in shading the area from solar radiation, expressed as a decimal fraction.

T'_c is the difference between the cloud base temperature and snow surface temperature, in $^{\circ}F$. It is estimated from upper air temperatures or by lapse rates from surface station, preferably on a snow-free site.

N is the estimated cloud cover, expressed as a decimal fraction.

k is the basin convection-condensation melt factor, as defined in paragraph 10-03.02. It depends on the relative exposure of the area to wind.

The melt coefficients given in the above equations express melt rates in inches per day. For those equations where wind is included in the convection-condensation term, it may be necessary to subdivide the day into smaller time increments, especially if there is marked variation in both wind and temperature or dewpoint. The coefficients also express melt for ripe snowpacks (isothermal at $0^{\circ}C$ and with 3 percent initial free water content — see chap. 8). Except for loss by transpiration from forested areas, the melt determined by the above equations represents the actual melt of the snowpack averaged over a basin area (or zone), expressed as ablation of the snowpack in inches of water equivalent. The equations are based on linear approximations between saturation air-vapor pressure and dewpoint, and between longwave radiation and the temperature of the radiating surface for the ranges ordinarily experienced (see chap. 6). Substitution of values for design conditions is made in accordance with the optimum meteorological sequence for each of the meteorological factors, either on the basis of the average for the whole snow-covered area of the basin, or of varying values for increments of elevation. For cases where elevation zones are evaluated separately, it is necessary to describe the meteorological sequence and melt factors characteristic of each zone. This requires lapsing air temperature, dewpoint, and wind to the specified elevation level. An additional consideration, when applying any design-flood snowmelt equations to forested or partly forested areas, should be given to the possibility of change in forest condition by subsequent timber removal and consequent change in the basin convection-condensation melt factor, k.

10-03.04 Basin clear-weather snowmelt coefficients. - For those basins with adequate hydrometeorological records for synthesizing historical streamflow data, basin melt coefficients using appropriate thermal budget indexes may be derived as outlined in chapter 6. It is necessary, of course, to treat the basin or component sub-basins as a whole rather than a series of elevation bands. The derived basin snowmelt coefficients integrate the basin characteristics with regard to factors affecting snowmelt, and relate the snowmelt to a fixed condition of observation. It is then necessary to relate the meteorological factors to the conditions of measurement for which the coefficients have been derived.

10-03.05 Elevation variation of snowmelt. - The use of elevation zones for snowmelt computations leads to consideration of the variation of snowmelt with elevation. It is a generally held opinion that snowmelt decreases with elevation because of the normal decrease of temperature with height. It has been shown for WBSL that, during active melt periods, the decrease of snowmelt with elevation is very slight, considering average basin characteristics in mountainous regions. Although the average snow surface albedo tends to increase with height, there is normally less dense forest cover at higher elevations, so that there is likelihood of greater energy input to the snowpack directly by solar radiation. Wind speeds, also, are generally greater at high elevation areas. These factors tend to balance the normal air temperature decrease with elevation, as it affects snowmelt. It is emphasized that this situation prevails only during clear weather periods in the active melt season; limited studies of water equivalent ablation under these conditions tend to verify nearly uniform melt rates with respect to elevation. In the derivation of design floods, the separation of the basin into elevation zones is important from the standpoint of defining the snow wedge and subsequent depletion of the snow cover. If a simple temperature index is used to evaluate melt for spring snowmelt design floods, an increase in the melt factor with elevation, which would partially compensate for the normal decrease in temperature with elevation, is appropriate.

10-04. DESIGN FLOOD HYDROGRAPH SYNTHESIS

10-04.01 The derivation of design flood hydrographs requires combining the effects of snowmelt, rainfall, losses by evapotranspiration and soil-moisture recharge, and total time delay to runoff by storage in the snowpack, ground, and channel. All must be evaluated on a time-rate basis over the effective runoff-producing areas. The methods of hydrograph synthesis presented in chapter 9 apply directly to design flood analysis, and accordingly the information presented there will not be repeated. Wherever possible, the method of hydrograph synthesis should be checked against historical data by the reconstitution of major floods of record.

10-04.02 The extension of the hydrologic variables to design-flood conditions can be accomplished as set forth in section 10-02. Having arrived at the optimum meteorological sequence, rational snowmelt rates may be determined (section 10-03.) and water excesses from rain and snowmelt may be routed through the optimum basin storage condition consistent with the design condition to produce the maximum peak discharge. The principles outlined above apply to both winter and spring floods. The storage effect of the snowpack must be taken into account for winter floods. For spring floods, it is usually assumed that the snowpack is primed prior to the flood event.

10-05. EXAMPLES OF DESIGN FLOODS INVOLVING SNOWMELT

10-05.01 General. - Under Project CW-171, the Snow Investigations Unit has assisted participating district offices in the derivation of a number of the design floods involving snowmelt for a number of reservoir projects. The following paragraphs contain brief descriptions of some of the design floods so derived by district offices. Also shown are excerpts from the plates prepared for illustrating the procedures.

10-05.02 Painted Rock maximum probable flood. - The maximum probable flood for the design of the spillway at Painted Rock Reservoir was derived by hydrologists in the Los Angeles district office. The details of design are reported in Design Memorandum No. 1 for the project. 1/ This flood is an example of a winter rain-on-snow type, in which the major contribution to runoff is from rainfall, but snowmelt significantly augments the runoff volume as well as peak discharge. The project is located on Gila River, near Gila Bend, Arizona. The drainage area of

50,800 square miles was divided into 12 sub-areas, each of which were further divided into 8 elevation zones. The initial snowpack condition was determined from analysis of climatological records involving snow depths, to which were applied an assumed density consistent with the time of year. A snow wedge, based on an enveloping line of water equivalent vs. elevation, was determined for each sub-basin. Snowmelt was computed for each sub-area and elevation zone by six-hour increments, utilizing the methods outlined in section 10-03. The values of snowmelt were added to the six-hour rainfall increments for the maximum probable storm. Losses to direct runoff were computed on the basis of assumed infiltration rates by zones and sub-areas, and water excesses contributing to direct runoff were routed by synthetic unit hydrographs. The hydrograph for each watershed was in turn routed through upstream channel and reservoir storage to the Painted Rock reservoir site, and a composite design flood hydrograph was derived for the project. The snowline was initially at 3000 feet and receded to 5000 feet by the end of the storm period.

10-05.03 Cougar standard project flood. - The standard project flood for Cougar Dam site on the South Fork, McKenzie River, Oregon, is an example of a winter rain-on-snow standard project flood and was derived in the Portland District office, as reported in Design Memorandum No. 2, Cougar Dam and Reservoir. 4/ Plate 10-1, which is extracted without change from the design memorandum, illustrates the pertinent information used in the derivation of the standard project flood. The 210-square-mile drainage area was divided into 5 elevation bands which varied from 4 to 35 percent of the basin area. Figure 1 illustrates the components of the hydrologic balance for the standard project storm. Values of rainfall, snowmelt, water stored in the snowpack, surface losses, and water excesses are given, together with the assumed temperature distribution. Figure 2 shows the snow-wedge condition before and after the design storm. The initial snow wedge was derived from analysis of water-equivalent data for snow courses in the surrounding regions. Figure 3 is the standard project flood series, showing the inflow and outflow hydrographs derived from the assumed pre-flood storm and the standard project storm. Figures 4, 5, and 6 are depth-duration curves, a six-hour unit hydrograph, and loss curves, respectively. In the derivation of this flood, snowmelt was computed by zones, using a melt rate of 0.08 inch per degree-day above 32°F applied to appropriate air temperatures for each zone. The melt rate conforms to that previously described for the condition of rain-on-snow in heavily forested areas. Melt from rain itself was added separately. Storage in liquid water in the snowpack during the pre-flood storm was computed in accordance with the liquid-water-holding capacities of the snowpack presented in chapter 8. Reference is made to the previously referenced design memorandum for a more complete description of the standard-project-flood analysis for this site.

10-05.04 Libby spillway design flood. - The derivation of a maximum probable flood for the design of the spillway for Libby project was completed by the Seattle District office and reported on in the Design Memo No. 2 for that project. 5/ This flood is the spring snowmelt type, augmented by rainfall assumed to occur near the crest of the flood. Evaluation of the basin runoff characteristics and empirical snowmelt rates was first accomplished by reconstitution of flood season hydrographs for five years of historical record. The procedures were then applied to the optimum flood-producing sequence as determined from a study of the maximum flood producing meteorological conditions in the Columbia River Basin, by the Hydrometeorological Section of the U. S. Weather Bureau. 7/ The Kootenai River, upon which the project is located, drains 10,240 square miles at the gaging station at Libby, Montana. In the derivation of the maximum probable flood, the basin was treated as a whole, rather than subdividing the area into zones of elevation or homogeneous units. Corrections for snow-covered area were made progressively through the melt season. Snowmelt rates were computed using degree-day indexes, the degree-day factors being varied according to season. Runoff excesses were routed to the project site by a single unit hydrograph. As an independent check upon results, snowmelt by the thermal-budget method was computed, and a separately derived inflow hydrograph was obtained. Plate 10-2 shows the spillway design flood inflows computed by each method, as well as pertinent data used in the flood derivation.

10-06. SUMMARY

10-06.01 The technique of determining either a maximum probable or a standard project flood is essentially the same for both snow-free and snow-covered areas. The existence of snow merely introduces additional complicating factors. Two principal types of floods occur: (1) winter floods resulting from rain-on-snow events where the air temperature is relatively low and the snowmelt contribution to flood is relatively small; and (2) spring floods from melting of the accumulated winter snowpack. Rain falling on snow at a time when the streams and melt rates are high may also contribute to a spring snowmelt flood. Winter floods are generally of short duration and exhibit a rapid rise and fall in the runoff hydrograph, because of the relatively intense rates of rainfall compared to those of snowmelt. In contrast, spring snowmelt floods are of long duration and the runoff hydrograph is generally flat-crested.

10-06.02 Hydrologic design requirements for reservoir projects include the control of a selected design flood and the ability to pass safely the maximum probable flood inflow. A design flood of maximum volume does not necessarily produce a maximum peak discharge. Both volume and peak discharge are evaluated from a certain optimum combination of weather, snow and soil moisture conditions. In evaluating the factors for design conditions, the selected values must be compatible with the other factors affecting runoff or peak discharge.

10-06.03 Assurance of economical and safe design can be best obtained through use of a rational approach to the problem, based on known physical laws concerning the processes affecting streamflow and runoff, and extension of those relationships to given conditions of design. The use of simplified or short-cut methods is warranted only for preliminary use or for projects whose safety and economic justification do not require detailed flood analyses. For such cases, the judgement of the engineer responsible for selection of design floods is relied upon to evaluate the flood potential properly. His background and experience in applied hydrology should include such a knowledge of hydrograph analysis and synthesis as is indicated in this chapter.

10-07. REFERENCES

1/ CORPS OF ENGINEERS, Los Angeles District, "Hydrology for Painted Rock Reservoir, Gila River, Arizona," Design Memo. No. 1, 1 August 1954.

2/ CORPS OF ENGINEERS, Office of the Chief of Engineers, "Standard project flood determinations," Civil Engineer Bulletin No. 52-8, Washington, D. C., 26 March 1952.

3/ CORPS OF ENGINEERS, Office of the Chief of Engineers, "Flood-hydrograph analyses and computations," Part CXIV, Chap. 5, Engineering Manual, Civil Works Construction

4/ CORPS OF ENGINEERS, Portland District, "Hydrology and meteorology, Cougar Dam and Reservoir, South Fork McKenzie River, Oregon," Design Memo. No. 2, 15 December 1955.

5/ CORPS OF ENGINEERS, Seattle District, "Derivation of spillway design flood inflow, and Appendix A, Libby Project, Kootenai River, Montana," Design Memo. No. 2, 29 July 1952.

6/ CORPS OF ENGINEERS, U. S. Army "Columbia River and tributaries, northwestern United States," House Doc. 531, 81st Cong., 2nd sess., (8 vols.), Government Printing Office, Washington, D. C., 1952.

7/ U. S. WEATHER BUREAU, Hydrometeorological Section, "Tentative estimate of maximum-possible flood-producing meteorological conditions in the Columbia River basin," 25 January 1945.

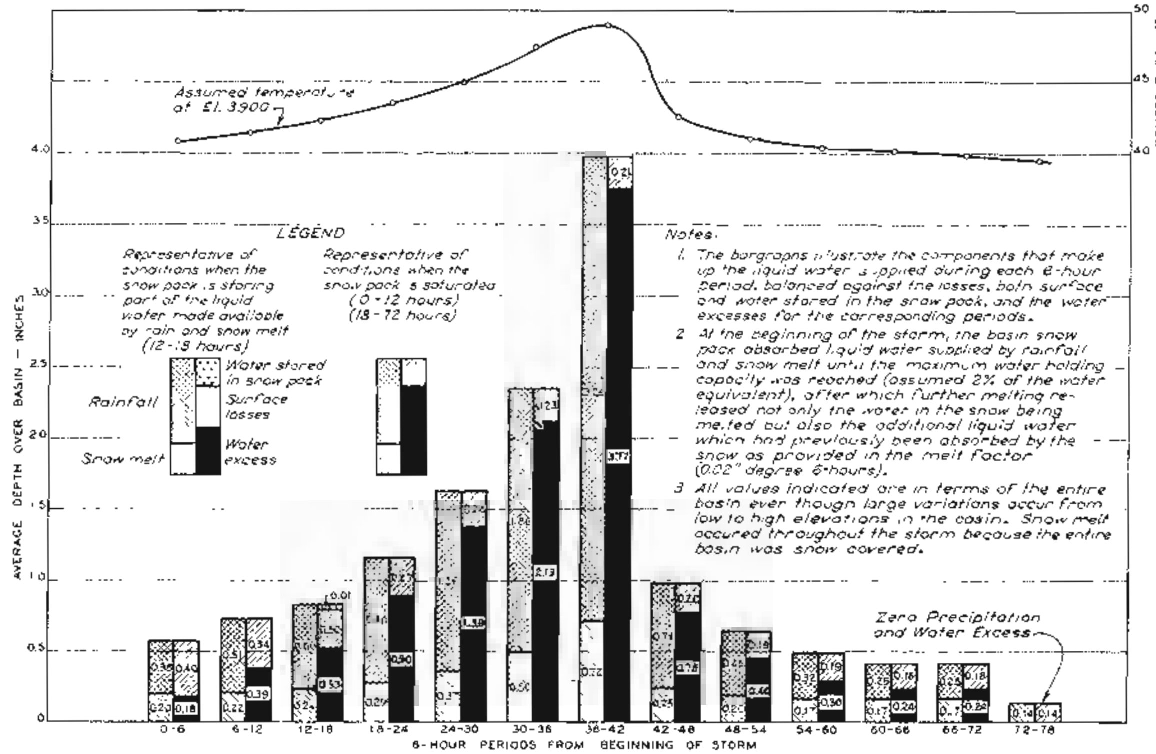


FIGURE 1

STANDARD PROJECT STORM HYDROLOGIC BALANCE

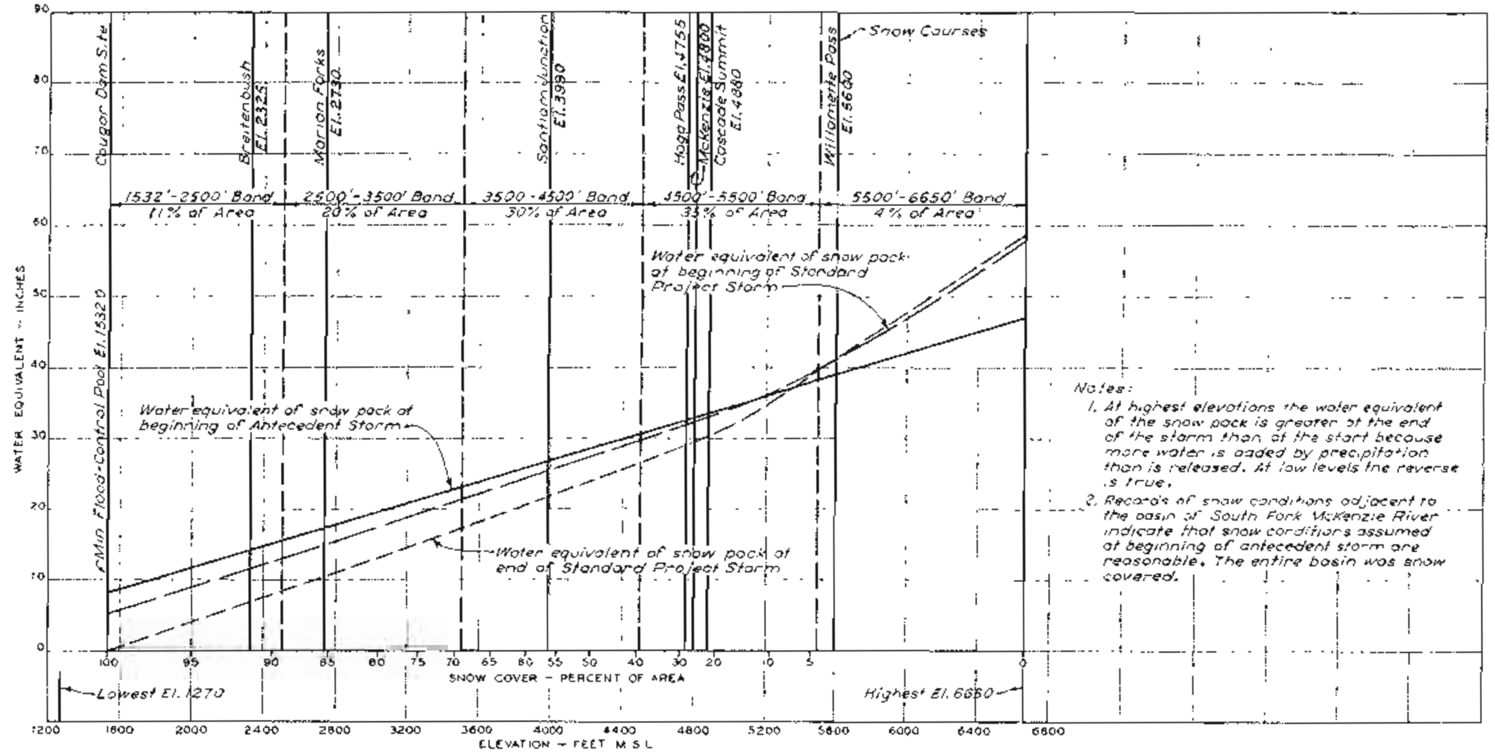


FIGURE 2

SNOW COVER DISTRIBUTION

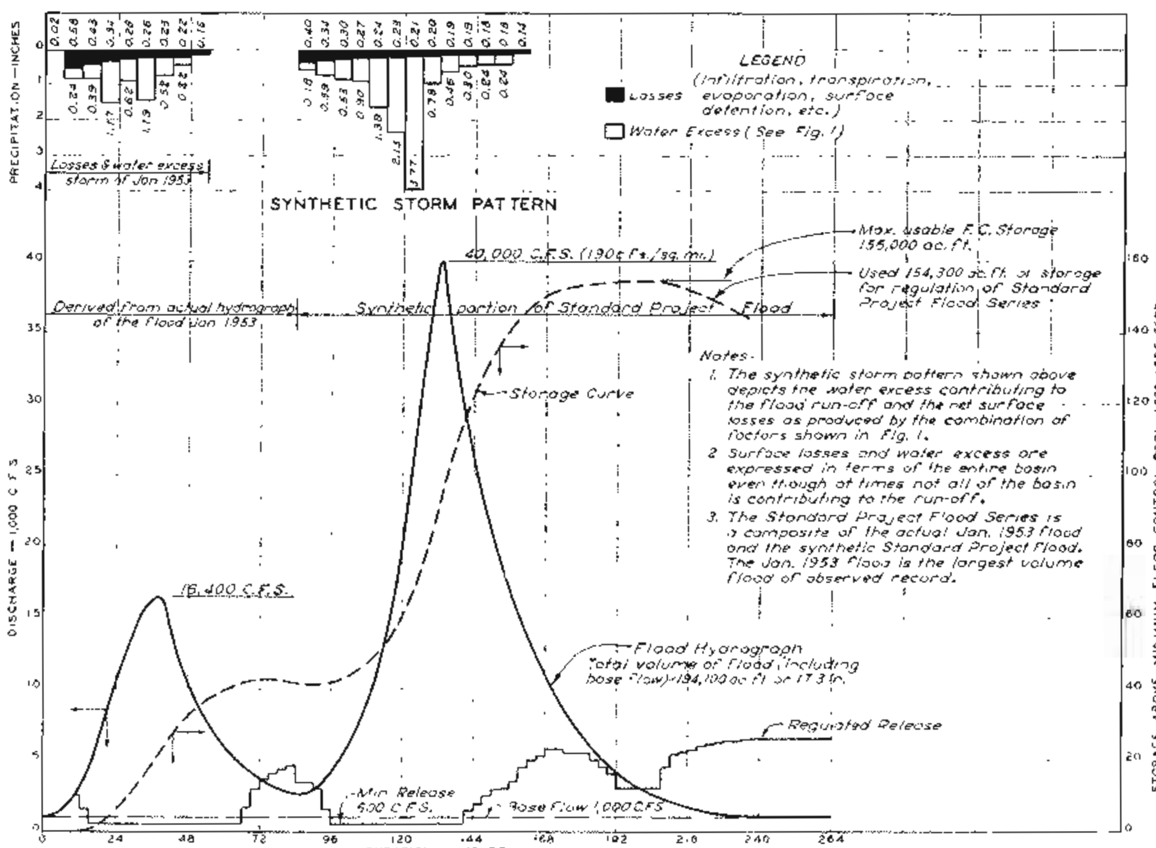


FIGURE 3

STANDARD PROJECT FLOOD SERIES

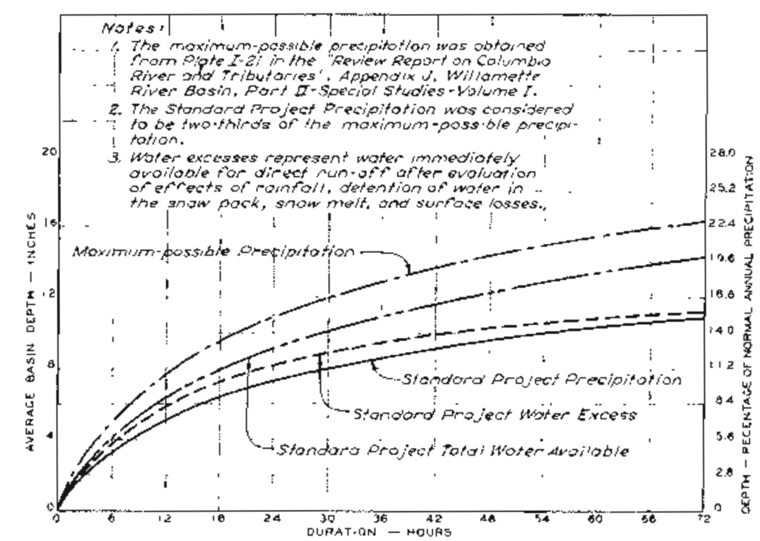
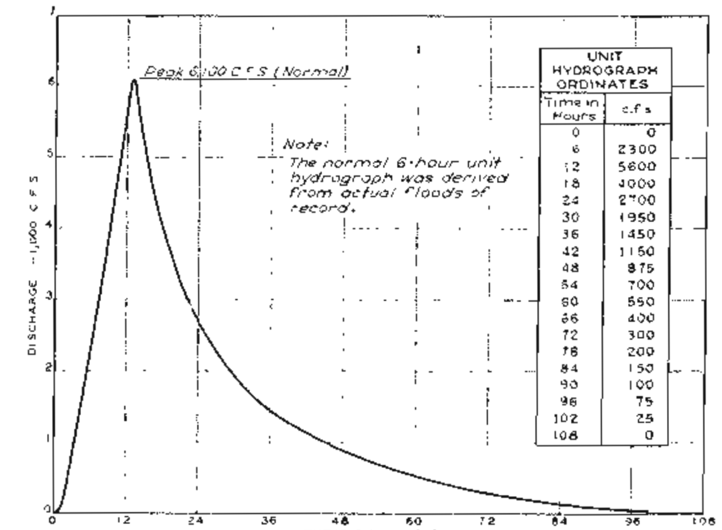


FIGURE 4

DEPTH-DURATION CURVES



UNIT HYDROGRAPH ORDINATES

Time in Hours	c.f.s.
0	0
6	2300
12	5600
18	4000
24	2700
30	1950
36	1450
42	1150
48	875
54	700
60	550
66	400
72	300
78	200
84	150
90	100
96	75
102	25
108	0

FIGURE 5

6-HOUR UNIT HYDROGRAPH

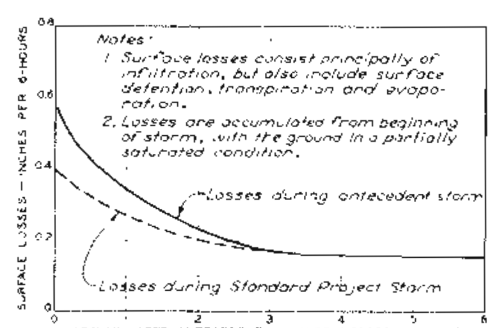


FIGURE 6

SURFACE LOSSES

WILLAMETTE RIVER BASIN, OREGON
SOUTH FORK McKENZIE RIVER
COUGAR DAM

STANDARD PROJECT FLOOD

SCALES AS SHOWN
PORTLAND DISTRICT, CORPS OF ENGINEERS NOV. 15, 1955

CU-20-2/12

PLATE 10-1

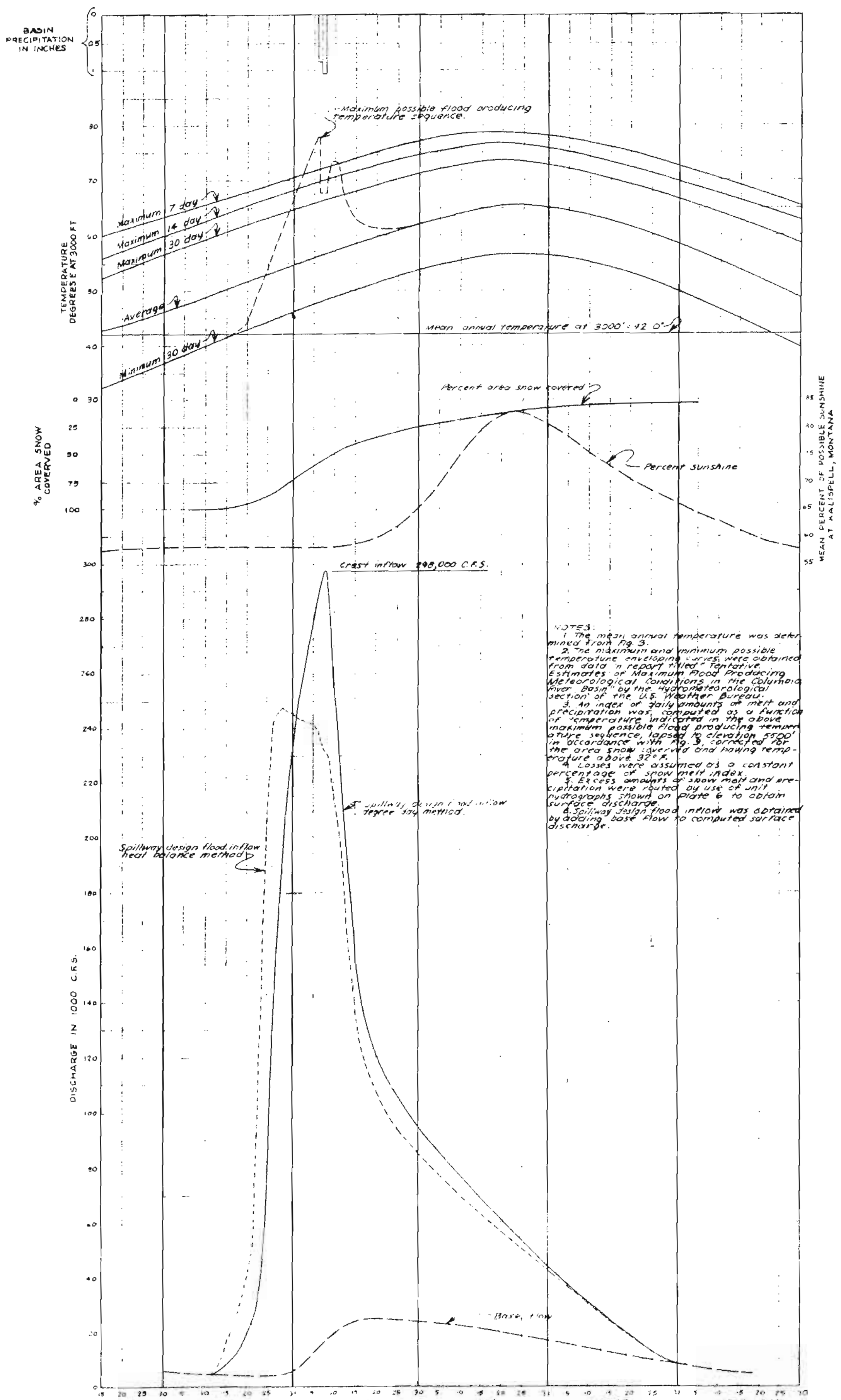


FIG. 1 SPILLWAY DESIGN FLOOD INFLOW TO LIBBY RESERVOIR

NOTES:
1. The mean annual temperature was determined from Fig. 3.
2. The maximum and minimum possible temperature enveloping curves were obtained from data in report titled "Tentative Estimates of Maximum Flood Producing Meteorological Conditions in the Columbia River Basin" by the Hydrometeorological section of the U.S. Weather Bureau.
3. An index of daily amounts of melt and precipitation was computed as a function of temperature indicated in the above maximum possible flood producing temperature sequence, lapsed to elevation 5500' in accordance with Fig. 3, corrected for the area snow covered and having temperature above 32° F.
4. Losses were assumed as a constant percentage of snow melt index.
5. Excess amounts of snow melt and precipitation were routed by use of unit hydrographs shown on Plate 6 to obtain surface discharge.
6. Spillway design flood inflow was obtained by adding base flow to computed surface discharge.

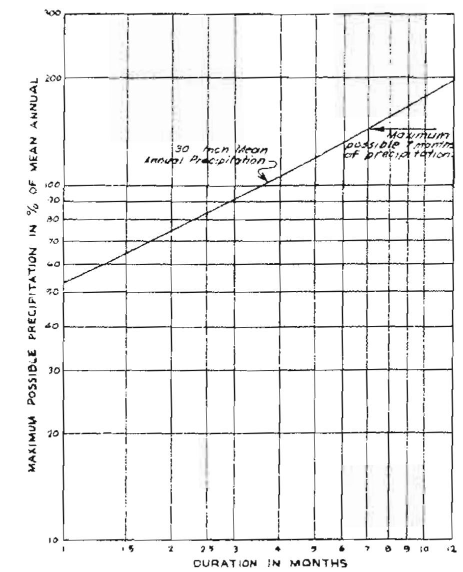


FIG. 2 PRECIPITATION DURATION CURVE

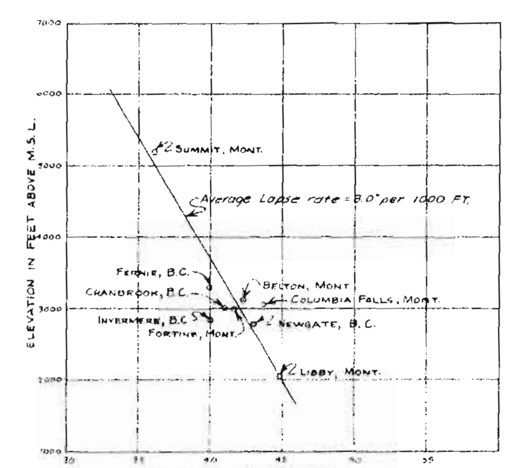


FIG. 3 TEMPERATURE-ELEVATION CURVE

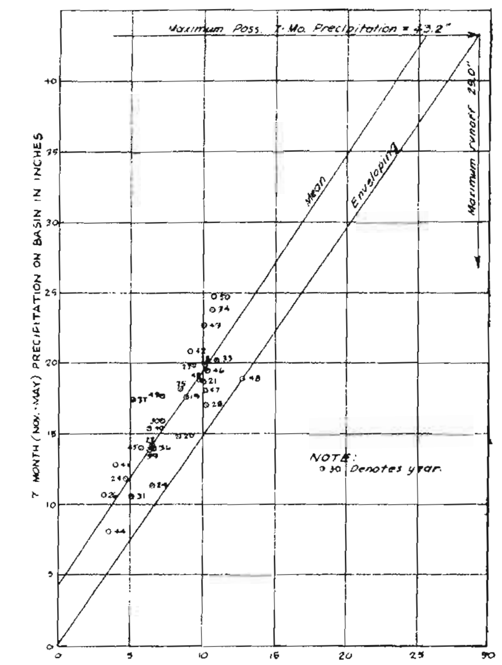


FIG. 4 PRECIPITATION RUN-OFF CURVE

KOOTENAI RIVER, MONTANA
LIBBY PROJECT

SPILLWAY DESIGN FLOOD INFLOW

In 1 Sheet Sheet No. 1 Scale: As Shown
Seattle District, Seattle, Washington 29 July 1952
Prepared: Submitted:

Recommended: Approved:

Chief, Planning and Reports Branch Chief, Engineering Division
Drawn by: M.E.T.
Traced by: W.E.L.
Checked by: M.E.T.
Transmitted with report dated 29 July 1952
File No. E-53-3-8

CHAPTER 11 - SEASONAL RUNOFF FORECASTING

11-01. INTRODUCTION

11-01.01 General. - The continued expansion in the use of water resources has emphasized the need for reliable seasonal runoff forecasts. These forecasts are useful for operational planning both in areas in which the streamflow is regulated by reservoirs and in areas where no such controls exist. In the case of uncontrolled flows, an advance knowledge of the anticipated volume of runoff is most useful in making advance plans for irrigation diversions, power generation, flood protection, etc. Where storage reservoirs exist, optimum use of the storage space for such conservation uses as power production, irrigation, navigation, industrial and domestic needs, preservation of fish and wildlife, pollution abatement, and recreation requires an advance knowledge of runoff volume. The flood-control operation of reservoirs is also benefited by volume-of-runoff forecasts. The most important need for such forecasts is to be found, however, in the operation of multiple-purpose reservoirs, where the contradictory requirements of flood control and conservation require accurate forecasts of seasonal runoff volume. As demands for additional water and additional flood protection continue to increase, the need for greater efficiency in the control of water, and thus for reliable seasonal runoff forecasts, becomes increasingly important.

11-01.02 Limitations. - The accuracy of seasonal runoff forecasts from snowmelt basins is limited by a number of factors. As was emphasized in chapter 3, the problem of evaluating the actual amounts of precipitation and basin snowpack water equivalent is a complex one, particularly for areas where hydrologic stations are sparse. As pointed out in chapter 4, the problem of evaluating other factors such as soil moisture, evapotranspiration loss, and ground water supply is also a difficult one. Even with a knowledge of conditions throughout the forecast period, forecasts are subject to error resulting from improper evaluation of the important factors affecting runoff.

11-01.03 Also contributing toward inaccuracy of seasonal runoff forecasts are those hydrologic events which occur after the initial date of forecast, the most important of these being spring precipitation. Their importance is largely due to the fact that they cannot be forecast accurately by presently available techniques. Seasonal runoff forecasting is particularly difficult in areas where significant proportion of the runoff results from widely varying amounts of spring precipitation from year to year. Factors such as loss by evapotranspiration and variations in soil moisture retention are generally of lesser importance. These factors may increase the error in runoff forecast significantly if their effects are additive to errors resulting from improper evaluation of other factors. On the other hand their effects may cause an apparent increase in accuracy of a given forecast by compensating for errors in other factors. Such random improvement cannot be depended upon to produce forecasts of equal reliability in the future.

11-01.04 Unfortunately, on many project basins, runoff resulting from conditions occurring after the date of forecast is of such magnitude and variation that forecasts of an acceptable degree of accuracy are not presently possible. At some future date the accuracy of seasonal runoff forecasts may be improved by use of long-range weather forecasts. Meanwhile, special consideration must be given to effects of conditions occurring subsequent to the date of forecast. It is often necessary to revise the forecast in keeping with conditions which occur after the initial forecast is made. Such a situation emphasizes the need for developing forecast procedures that permit easy and logical revision of a given forecast where necessitated by the occurrence of unusual weather conditions.

11-01.05 Feasibility. - Despite the difficulties encountered in forecasting seasonal runoff, forecasts of acceptable accuracy can be made if due consideration is given to all important factors affecting runoff. On most snowmelt basins, a large proportion of the spring snowmelt runoff can be evaluated at the time the forecast is made. This is particularly true on basins where the snowpack water equivalent on the date of forecast represents a large percentage of the seasonal runoff. Accurate evaluation of factors existent on the date of the forecast restricts errors in the forecast to those caused by occurrence of subsequent unusual weather conditions. On basins where conditions subsequent to the forecast date account for only a small proportion of the runoff and do not vary greatly from year to year, errors may be quite small. Under such circumstances seasonal runoff forecasting may be accomplished with reasonable assurance that the deviation of the forecasted amount from the true amount will be confined within certain prescribed limits. Such prescribed limits, of course, vary in accordance with the use for which the forecast is intended; a forecast acceptable for one purpose may be entirely inadequate for other purposes.

11-01.06 <u>Factors affecting runoff</u>. - Factors affecting runoff may be logically classified under the two categories, supply and loss. Supply for a given season is comprised largely of precipitation, minor sources being condensation and carry-over of water from preceding seasons in various forms such as ground water, channel and lake storage, and snow. A possible additional source of supply is underground flow from adjacent basins. Soil moisture is not considered a source of supply because it is not available for runoff.

11-01.07 Loss occurs in a number of ways, the importance of each varying in accordance with meteorological and basin characteristics. Generally, the greatest proportion of loss on a basin results from evapotranspiration, comprised of evaporation from the ground and snow surfaces and transpiration from leaves of vegetation. Considerable loss also occurs through evaporation of intercepted snow and liquid water from the external surfaces of vegetation. Such losses occur before the precipitation reaches the ground and is generally called interception loss, since its occurrence is dependent upon interception of precipitation by vegetation. The remaining sources of loss on a basin during a given season are deep percolation, retention as soil moisture, and carryover of moisture into the next season in various forms, such as ground water, channel and lake storage, and snow. Loss by deep percolation is difficult to evaluate. It is generally assumed that such loss is either negligible, a constant amount, or a fixed percentage of the total loss, and its effect upon runoff is integrated into one or more of the other factors affecting runoff.

11-01.08 Factors affecting the quantity of precipitation were discussed in chapter 3, and, since the relationship of quantity of precipitation to runoff is clearly defined, further discussion is deemed unnecessary. Supply resulting from condensation is dependent upon the vapor pressure gradient between the surface and the air; the gradient, in turn, is dependent upon the vapor pressure of the air and of the snow surface. Although the addition of condensation to the quantity of runoff is negligible, the resultant heat of condensation has a significant effect upon rate of snowmelt and, consequently, upon the distribution of runoff. Supply through underground flow from adjoining basins cannot be precisely determined, but qualitative evaluation can be made from detailed geologic and hydrologic investigations. Factors affecting carryover from preceding seasons are those which determine the supply and loss during the antecedent seasons. Direct evaluation of ground water is impractical because of general unrepresentativeness of ground water well observations and great variability of ground water conditions over basin areas. Indirect evaluation based on recession analysis is generally satisfactory as a method for estimating changes in ground water storage. Carryover between years in the form of snow, channel, ground or lake storage may be computed where necessary. Factors affecting the amount of runoff as a result of variation in loss are interrelated with factors associated with supply of moisture. Since evaporation and transpiration losses vary largely in accordance with temperature, the latter is considered an important factor affecting runoff. A related factor affecting loss by evapotranspiration is supply of water during periods when evapotranspiration is occurring.

11-01.09 High rates of rainfall are conducive to high runoff per unit volume of precipitation. On the other hand, precipitation is less effective in producing runoff if it occurs in light storms, particularly if it is associated with high temperatures during or between storms. Precipitation falling in the form of rain on bare ground is subject to greater loss than that falling on snow. For light rainfall intensities, precipitation falling on bare ground during the spring melt season may be considered to be lost to runoff, while precipitation falling on snow-covered areas may be considered to be 100 percent effective in producing runoff. Thus the areal extent of snow cover during periods of spring precipitation and during periods of high evaporation rates has an effect upon seasonal runoff.

11-01.10 The soil moisture content is affected by the climatic regime of a given area. Where autumn or winter rains are sufficient to provide full field capacity of the soil, the year-to-year variation in soil moisture at the beginning of the spring snowmelt runoff season is negligible. However, lesser amounts of precipitation will result in corresponding deficits in soil moisture, up to full field capacity of the soil. Such deficits must be made up by melt or rainfall contribution during the melt season, resulting in a corresponding loss to runoff.

11-01.11 Soil moisture deficits may be accounted for in a number of ways, as will be explained later in connection with runoff indexes. Consideration should be given to the probable condition of soil moisture at the time of the forecast. After the melt season is well underway and the soil has attained field moisture capacity throughout the entire range of elevation within the basin, no further consideration need be given to losses due to soil moisture deficiency. Once the soil reaches field moisture capacity as the result of fall or winter rains, soil beneath the snowpack will remain saturated throughout the period of snow cover, because any loss by transpiration will be supplied by melt. When soil moisture deficits exist, however, they vary widely with elevation within a basin.

11-01.12 Methods. - Methods of forecasting seasonal runoff may be broadly classified as two main types, water-balance and index. A third type consists of a combination of the two main types. The index method assumes a fixed relationship between volume of runoff and causative indexes representing factors. In the index method, no implication is made that the factors are quantitatively evaluated. The water-balance method, on the other hand, implies that each factor is quantitatively measured, the algebraic sum of all the factors being equal to the runoff. In water-balance procedures the factors determining runoff are referred to as components since they are actually the component parts of runoff.

11-01.13 Regardless of the method, the factors used should be selected on the basis of the hydrologic balance of the area involved. The water balances shown in chapter 4 for each snow laboratory provide a guide for selecting pertinent factors. The forecasting procedure should utilize all important variables affecting runoff. The effects of the variables before and after the date of forecast should be evaluated separately, to insure proper weighting of each variable and provide a means of revising forecasts to suit conditions subsequent to the date of forecast. Direct correlations of early-season precipitation or snow accumulation with total seasonal runoff should be avoided, because of the likelihood of unrepresentative weightings of the variables caused by unaccounted random variance in late-season precipitation and losses.

11-02. INDEX PROCEDURES FOR FORECASTING SEASONAL RUNOFF

11-02.01 General. - Procedures for forecasting runoff by the index method basically involve correlations of historical records of runoff with indexes of important determinants of runoff for the area. Forecasting procedures may be based on either mathematical or graphical correlations, or a combination of the two. The regression functions so derived effectively weight the variables corresponding to their effect on runoff. An adequate period of record is essential to proper evaluations of runoff coefficients, and in general, the greater the number of variables involved, the longer is the period of record required. The use of statistical procedures for deriving mathematical relationships has been widely used by hydrologists for obtaining the best fit of historical data. Statistical procedures provide standard methods for evaluating effectiveness of runoff parameters and for comparing relative reliability of forecasting methods. The use of statistical methods, however, should not be attempted on a casual basis without full knowledge of the hydrologic factors involved, the statistical techniques, and the limitations of the methods.

11-02.02 The simplest mathematical procedure for estimating seasonal runoff is by means of a simple linear regression,

$$Y = a + bX,$$

where $\underline{Y}$ is the dependent variable, runoff; $\underline{X}$ is an index representing the principle determinant of runoff, and $\underline{a}$ and $\underline{b}$ are the derived constants. A comparable graphic procedure consists of simply plotting the values of each of the variables on rectangular graph paper and drawing a linear regression line of best fit by eye through the plotted points. However, numerous factors usually affect the volume of runoff, necessitating the introduction of additional parameters into the forecast procedure. Thus, mathematical relationships may involve equations varying from simple two-variable regressions to multi-variable linear and curvilinear regressions. In some instances variables used in the basic forecast equation are derived by correlating component parts of a given independent variable with the dependent variable. As an example, monthly precipitation values are often correlated with runoff in a multiple regression equation to aid in determining the weight to be assigned to monthly values to obtain the best possible precipitation index for inclusion in the basic forecast equation. Likewise graphic procedures may vary from simple single straightline relationships to complex coaxial graphs involving numerous variables, with relationships of variables represented by curves of varied shapes.

11-02.03 Indexes used in runoff forecasting. - A given factor can often be represented by more than one index. The supply of water stored in a snowpack on a given date, for example, may be represented by an index of either precipitation or snowpack water equivalent.

11-02.04 Water supply index. - Since the supply of water in a snowpack is the most important factor affecting runoff in areas of snow accumulation, the selection of an index to represent this factor has been given much attention. The relative reliability of precipitation and snow course measurements has been discussed in chapter 3, where it was shown that both types of measurements include errors due to the method of sampling as well as errors due to non-representativeness of point measurements. It was also pointed out (in chap. 3) that measurements of precipitation are generally more suited to early-season forecasts, while snow accumulation measurements are generally more reliable for late-season forecasts. In either case, the availability of historical record is a vital factor in the selection of indexes of water supply. If the period of record of both types of data are approximately equal, each should be tested to determine which yields the best results for historical data. In addition to the foregoing indexes, measurements of quantities such as low-elevation winter runoff, atmospheric moisture inflow, or heat supply-runoff relationships, may be used.

11-02.05 Hydrologic network. - When considering the development or expansion of a network of hydrologic stations in snowmelt basins, the choice of whether to establish precipitation stations or snow survey courses depends on varying needs. For proper evaluation of rainfall effects, especially during the fall or spring, precipitation stations are necessary. They are more economical from the standpoint of cost and time where the services of an observer are available. On the other hand, if taking the observations necessitates field trips, the economy is no greater than that of making snow surveys, and involves, moreover, the particular data complications at unattended sites arising from the variability in gage catch due to the deficiencies discussed in chapter 3. Snow surveys, while subject to disadvantages of their own, have the advantage of providing a direct estimate of actual snowpack conditions at a given date, and consequently they provide a measure of the residual water supply which remains in storage in the snowpack. Precipitation gages which are attended daily can provide data for evaluation of incremental changes in moisture supply from the time of a comprehensive snow survey of a basin, by which short-term changes in forecasts may be made. Whichever source of water-supply data is used (precipitation gages or snow courses), emphasis should be placed on proper site selection, in accordance with the requirements outlined in chapter 3, in order to provide reliable and representative basic data. Since the adequacy of a newly-established network cannot be fully appraised until a number of years have elapsed, it may be desirable to establish both precipitation-gage and snow-course networks, and maintain both until it becomes conclusive that one or the other provides the data most suitable for an index. The final appraisal of the indexes is largely determined by the results obtained from their use in the development and testing of forecast procedures.

11-02.06 Precipitation index. - A simple type of precipitation index is the average of measurements at a number of stations considered representative of the basin. Such an index assumes that the data at all stations have equal weight in determining the volume of runoff. Often, however, the distribution of stations is such that weighting of station values is necessary to obtain proper results. Furthermore, the method of computing the index is dependent upon whether or not loss is treated in a separate index; if loss is not treated separately, its effect upon runoff is generally included in the precipitation index.

11-02.07 A method developed by Kohler and Linsley 6/ and used by the U. S. Weather Bureau consists of performing a multiple correlation, using annual basin runoff as the dependent variable and annual precipitation at representative stations as independent variables. Resultant regression coefficients are used as guides in establishing station weights. Stations having high negative values are considered unrepresentative and are excluded. A weighted basin value for each month is determined by summing the products of station weight and respective monthly precipitations. This weighted basin value is often referred to as effective precipitation; however, it is not a quantitative evaluation but an index of the precipitation effective in producing runoff. Since precipitation occurring in various months is not equally effective in producing runoff, further weighting is necessary if greater refinement is desired in establishing the effective precipitation index. This phase of weighting consists of performing a multiple correlation, with annual runoff as the dependent variable and the preliminary effective precipitation index by months as the independent variables. Resultant coefficients are used as guides in establishing weights to be assigned to the various months, the procedure being similar to that used for determining station weights. Final adjustment of the values is facilitated by plotting the monthly values as a function of time and drawing a curve through the plotted points.

11-02.08 Another method of weighting precipitation stations consists of qualitatively assigning weights, using graphs of station precipitation versus runoff as a guide. Preparation of the graphs consists simply of plotting water-year precipitation at individual stations versus water-year runoff and drawing a curve of best fit through the plotted points. The process is performed for each station considered representative of the basin. Deviations of the plotted points from the curves are indicative of the correlation of precipitation values with runoff. Weighting of stations consists of assigning high weights to stations showing the best correlation with runoff, and assigning progressively lower weights to stations for which the deviations of the points from the curve are progressively greater. No set rule can be established regarding the magnitude of the weights, but it is customary to make the sum of the station weights equal to unity. Normally the highest weighting factors are not greater than 3 times the smallest factors, though in some cases, the Weather Bureau has assigned weights as high as 5 times the smallest weight for stations on a given basin. Further refinement can be made by assigning weights to monthly values. These weights are based largely on loss rates; highest weights are generally assigned to coldest months when losses are at a minimum, with decreasing weights being assigned to months with increasing mean temperatures. Months are omitted in which the runoff resulting from precipitation is insignificant. As in the case of station weights, it is customary to make the sum of the weights equal to unity.

11-02.09 Basin precipitation amounts determined by the method described in paragraph 3-06.03 may be used as an index. In this connection, the index is one of basin precipitation as distinguished from effective basin precipitation. Accordingly, such an index should be used only if the loss factor is treated separately.

11-02.10 Snowpack water equivalent index. - A snowpack water-equivalent index of water supply can be derived in a number of ways. In general, methods for determining the effective precipitation index are applicable to the water-equivalent index; however, monthly weightings are not necessary. Since the basin snowpack water equivalent is a measure of the snow accumulation on a given date rather than of the amount occurring during a given period of time, the time of occurrence is unimportant. Because of the limited period of record of snow-course data on most basins, statistical procedures have not been generally used in weighting snow courses. The most commonly used index for expressing water supply in the snowpack is the average of water-equivalent measurements at a number of courses representative of the basin. The method is highly favored because of its simplicity. However, on many project basins snow courses are not representatively distributed, particularly with regard to elevation. Because of the pronounced effect of elevation upon depth of snow it is often advantageous to segregate snow courses by elevation zones, weighting each group in accordance with the percentage of basin area represented by each zone.

11-02.11 The snow chart described in paragraph 3-08.04 is a useful tool for computing indexes of snowpack water equivalent. Other means of weighting snow course measurements to obtain a basin index include assigning of weights in accordance with the area represented by each snow course or assigning weights in accordance with the hydrologist's subjective estimate of the representativeness of each snow course with respect to the basin snowpack water equivalent. Estimation of weights to be assigned to snow courses may be facilitated by plotting runoff versus snowpack water equivalent at individual snow courses, and comparing the degree of scatter of plotted points for each snow course.

11-02.12 Indirect indexes of water supply. - Other indexes of water supply exist, which represent less directly than precipitation or snowpack water equivalent the amount of stored water on a given area. Among these are (1) area of snow cover, (2) accumulated heat supply and runoff relationships, and (3) low-elevation winter streamflow. The area covered by snow can be determined in several ways, as described in chapter 7. It was pointed out that the usefulness of this index lies in evaluation of late-season residual runoff, when basin snow cover is less than, say, 50 percent of the initial snow-covered area. The error of the forecast represents a correspondingly smaller percentage of the total runoff than that of forecasts made earlier in the season (e.g., April first forecast). Photographic indexes of snow cover for forecasting runoff volumes have also been developed.8/ However, early season forecasts of runoff based solely on observations of snow-covered areas are usually unreliable because of the varying slope of the snow wedge from year to year. Relationships between heat supply and runoff have been tested for various basins, the form of the relationship usually being expressed in terms of an accumulated temperature melt index and accumulated runoff. Such relationships are based primarily on the relation between water supply and area of snow cover, as indicated by the runoff produced for a given condition of seasonal heat supply. Such relationships also integrate a variety of other effects of water supply, runoff, and loss. Koelzer 5/ devised such a procedure for the Seminoe River, Wyoming, and a somewhat similar procedure was developed for the Columbia River near The Dalles, Oregon under project CW 171. The usefulness of the method is that it provides an independent check upon forecasts made by other methods. Also, it evaluates runoff potential through the melt period. Its application, however, is limited to periods after the melt season is underway. The use of low-elevation winter runoff as an index of snowpack water equivalent is confined to situations where the area on which the runoff index is measured is in the path of the airflow carrying the moisture to the high-elevation areas where the snowpack forms. This method has been applied to the Columbia River near The Dalles, Oregon, as is reported on in Research Note 23. It is discussed in paragraph 11-03.09.

11-02.13 Soil moisture indexes. - Soil moisture can be represented by a variety of indexes. Correlations between precipitation indexes and runoff implicitly evaluate soil moisture conditions, since a relatively constant soil moisture deficit from the previous summer period must be satisfied before significant runoff occurs. Procedures involving water equivalent of the snowpack, on the other hand, must consider possible variations in soil moisture deficits. One of the most commonly used indexes of soil moisture is fall precipitation. An index of soil moisture deficit based exclusively on fall precipitation, however, is not entirely realistic, because of the variation in form of precipitation that may occur in the fall, and the possibility that winter rains or snowmelt may penetrate through the snowpack. Varying amounts of snowpack melt from ground heat may also affect the condition of soil moisture. Also, the effect of elevation variation of soil moisture should be taken into account, since an index at one elevation level may not be representative of other levels. Another commonly used index of soil moisture is winter runoff, since greater winter flows are generally associated with higher soil moisture content. However, this is more directly an index of ground water.. Actual measurements of moisture content of soil samples may be used as indexes of soil moisture. They are obtained by direct measurement of the moisture in soil samples by laboratory techniques, or by electrical resistance methods, using either Bouyoucos or Colman blocks. At present, electrical resistance methods are unreliable because of the difficulties in calibration (see chapter 4).

11-02.14 Ground water indexes. - Various indexes may be employed to represent the amount of ground-water storage on a given date. A commonly used index is volume of runoff occurring during a given period, higher runoff volumes generally being associated with higher ground water storage. Properly located wells provide data for a useful index of ground-water storage. Another highly useful index of ground water is base flow; however, inability to separate base flow from total flow sometimes imposes a limitation on the use of the base flow index.

11-02.15 Evapotranspiration indexes. - A separate index of evapotranspiration loss is seldom used in index forecasting procedures. Because the meteorological factors affecting evapotranspiration are generally the same as those causing snowmelt, the loss tends to be a direct function of melt for the snow-covered portions of the basin. Light spring precipitation falling on bare areas may usually be considered to be lost. Therefore, evaluation of evapotranspiration in a procedure involving primarily the water equivalent of the snowpack is not warranted. Methods based on an index of total precipitation throughout the period of snow accumulation and melt could logically include an index of evapotranspiration to account for variation in water loss during the fall and winter season. A temperature index function, based on mean monthly air temperature at a station representative of the basin area, could most easily serve this purpose.

11-02.16 Statistical methods. - Statistical techniques may be used to determine the effect of each index upon runoff and its relative importance in explaining the variance of runoff. Various indexes for a particular variable may be tried independently, to determine from historical record the ones which provide the best correlation. It is emphasized that the use of either graphical or mathematical methods of statistical analysis, whether they be used for simple linear two-variable correlations or complex multivariable relationships, should be considered simply as a tool to aid the hydrologist in evaluating indexes. The selection of variables used in the statistical analyses should be based on sound and thorough reasoning with regard to the conditions affecting runoff on the particular basin involved. Statistical methods may easily lead to a false sense of knowledge if results are used blindly without regard to hydrologic significance. This is particularly true in the case of procedures for forecasting seasonal runoff volumes, where historical data usually limit the number of observations in the sample to less than 20. Little confidence can be placed in a statistically derived forecasting procedure if the cause and effect relationships are either unknown or poorly understood.

11-02.17 Graphical methods. - Details pertinent to development of graphic correlations are given in various standard texts on hydrology and statistical methods; only a brief discussion of the principal methods of graphic analysis is presented here for the purpose of general appraisal of the method. One of the simplest methods of determining graphically the effects of a number of factors upon a given dependent variable is the method of deviations described by Ezekiel.4/ The first step consists of plotting scatter diagrams relating each independent variable to each of the remaining ones, and eliminating one of any pair of variables that show a high degree of correlation; such a correlation indicates that the variables are so closely related that their effects upon the dependent variable are inseparable. Of the remaining independent variables, the one considered most important (labeled X), is plotted against the dependent variable (Y), and a line of best fit is drawn through the plotted points. The deviations, Y-Y' (where Y' is the ordinate of the line of best fit corresponding to a given value of X.) of each point are then plotted against the next most important independent variable (X_2), and a curve of best fit is drawn through the plotted points. Deviations of each point from this curve are then plotted against a third variable (X_3). The process is repeated for each factor considered to have an effect upon the dependent variable. The completed curves are first-approximations, subject to revision inasmuch as each curve is drawn without consideration of the factors treated in subsequent curves. Having completed the first-approximation curves, the deviations in the last curve drawn are plotted as deviations from the initial first-approximation curve, $Y = f'(X_1)$. A revised curve, $Y = f''(X_1)$, is drawn through the plotted points. The deviations from the new curve are then plotted as deviations from the first-approximation to $Y = f'(X_2)$ and a revised curve is drawn through the plotted points. The process is repeated for each of the first approximation curves. Third-approximation curves are generally unnecessary, but if they are considered desirable, they may be made by the procedure used for the second-approximation curves. Although this method is relatively simple, its usefulness is limited by its lack of consideration of joint relationships between variables. As an example, the method implies that runoff resulting from a given amount of spring precipitation would be the same regardless of the extent of basin snow cover. Water-balance computations, as well as actual observations, indicate that such an implication is erroneous.

11-02.18 Another graphic method of determining the effect of variables upon runoff is the coaxial method (described in Applied Hydrology 7/). While more complex than the method of deviations, it is better adapted to the representation of joint functions. In one of the common variations of the method, the first step consists of plotting runoff, Y, along the ordinate versus the most important independent variable, X, along the abscissa in the first of four quadrants on a graph. The indexes representing a second important variable, X_2, are shown at each plotted point and a family of curves representing the index values is drawn. Runoffs determined from the curves in the first quadrant are then plotted on the ordinate of the second quadrant versus the observed runoff along the abscissa. Each of the plotted points is labeled with an index representing a third independent variable, X_3, and a family of curves is constructed to fit the plotted points. Similarly, additional variables are introduced in the third and fourth quadrants. Another graph of four additional quadrants may be utilized if necessary to consider all the important variables. As in the method of deviations, the first-approximation curves are subject to revision. Deviations of observed runoff values from the curves in the final quadrant are plotted against the first independent variable, X_1, and a curve of best fit is drawn through the plotted points. Deviations of this curve from the zero axis at given values of the variable X_1 denote the change to be made in the curves of the first quadrant at corresponding values of X_1. Following revision of the curves in the first quadrant, the curves in all successive quadrants must be revised before proceeding with refinements in the second quadrant. The revised deviations in the final quadrant are plotted against the second independent variable, X_2, and a line of best fit is drawn through the plotted points. Deviations of this line from the zero axis are used for adjusting the curves in the second quadrant, using the procedure described for the first quadrant. The process is repeated until all variables have been considered. Although generally unnecessary, a third approximation may be made, using the procedure described for the second approximation.

11-02.19 Numerical statistical methods. - It is not within the scope of this report to present a discussion of statistical techniques. Reference is made to standard textbooks or references on statistical analysis for detailed presentations of the methods commonly used. (e.g., Ezekiel 4/, Snedecor 9/, Brooks and Carruthers 3/, Arkin and Colton 1/, and Wilm 11/) Full understanding of the capabilities and limitations of least squares techniques, familiarity with the statistical nomenclature and significance of the concepts involved in statistical analysis, are requirements for intelligent application of statistical methods to forecasting procedures. In establishing a forecasting procedure for a given area, indexes of all variables known to have a significant effect on runoff during the forecast period should be incorporated in the multiple correlations, in order to determine the reliability of the method as a whole. The least significant variables may then be dropped, depending upon requirements, and incremental effects of variables may be determined. Data for regression analysis may be transformed logarithmically or exponentially to provide curvilinear rather than linear relationships. Such transformation is not recommended, however, unless curvature is known to exist from physical considerations of the variables involved. Computations performed in connection with multiple regression analysis involving extensive hydrologic data are laborious and time consuming. With the advent of high-speed electronic computing machines, however, the time and labor involved in performing the computations may be reduced to a small fraction of that required using desk calculators. Special programs for electronic computers are available which may be used to perform automatically all computations involved in the solution of the normal regression equations.

11-03. EXAMPLES OF INDEX METHODS

11-03.01 General. - Index methods for forecasting seasonal snowmelt runoff have been developed for a wide variety of conditions. Many of the procedures have been reported on in various technical journals dealing with hydrologic problems, while others, although in operational use, have not been generally disseminated. A complete review of all such forecasting procedures is not practical here. Reference is made to examples of graphical correlations for forecasting seasonal runoff for California drainages, as described by Strauss 10/. A brief discussion of a report on procedures for forecasting seasonal runoff for Columbia River near

The Dalles, Oregon, issued by the Water Management Subcommittee, Columbia Basin Inter-Agency Committee,13/is presented to illustrate some of the basic techniques involved. A total of four procedures prepared by various federal agencies were reviewed in connection with this report.

11-03.02 The Columbia River basin (D.A. = 237,000 sq. mi.) is characterized by wide variations in both meteorological and topographical features. A large proportion of the winter precipitation is in the form of snow, with the maximum accumulation of snow occuring on about April 1st of each year. A high proportion of the runoff occurs during the late spring and early summer months, largely as a result of snowmelt. Basic hydrologic data used in the development of forecast procedures are comprised of runoff, precipitation, and snowpack water equivalent data. Adequate precipitation records are available as far back as 1927; in addition there are a number of stations in the Columbia Basin whose records extend back before the turn of the century. A complete record of discharge, as gaged near The Dalles, Oregon, is available from 1879 to date. Adequate records of snowpack water equivalent are generally confined to years subsequent to 1938.

11-03.03 U. S. Weather Bureau procedure. - A procedure which uses a precipitation index as the principal parameter has been developed by the U. S. Weather Bureau for forecasting seasonal runoff on the Columbia River near The Dalles, Oregon. The procedure consists essentially of forecasting the runoff on each of 22 sub-basins and equating the runoff from the sub-basins to runoff at successive downstream points. Procedures for sub-basins consist generally of establishing a relationship between water-year runoff and precipitation for the period September through June. The precipitation period is longer for some sub-basins, the objective being to include all months having significant amounts of precipitation for any area. A total of 78 precipitation stations are used for the basin as a whole, the number per square mile for each sub-basin varying widely as a result of over-all variation in density of stations having adequate records. Precipitation values are weighted with regard to both station and month, multiple correlations of runoff and precipitation being used as guides in assigning the weights for various stations and months. Weighting of precipitation by months serves as an indirect means of accounting for losses, less weight being assigned to months in which greater losses are normally incurred.

11-03.04 In most of the sub-basin forecasts, effects of conditions occuring in previous years are accounted for either by a carry-over factor incorporated in the precipitation index or by a carry-over adjustment to the forecasted runoff, the latter generally being used when consideration is given to the conditions occuring in several antecedent years. A carryover factor in the precipitation index is normally used when carryover effects are considered for only the preceding year. The value of the factor is determined by multiple correlation and usually varies from one-tenth to two-tenths of the previous year's partial precipitation index. that is, the index exclusive of carryover effects. Although the runoff for the full water year is used in the statistical correlations, forecasts of seasonal runoff can be made by simply deducting the flow prior to the date of forecast from the amount forecast for the water year as a whole. Observed precipitation values are used for months prior to the date of forecast; assumed, forecast, or normal values are used for subsequent months.

11-03.05 An outstanding characteristic of the Weather Bureau procedure is the extensive use of statistical analyses. In conjunction with statistical derivations, it has been noted that in some instances stations located outside of a given sub-basin are used in preference to a station located within the sub-basin, the latter, however, being used for another sub-basin. Although forecast results were improved by use of the carry-over adjustment, it is believed that an adjustment based on the flow at the end of the preceding water-year would yield results comparable to those obtained by the laborious statistical procedure used by the Weather Bureau.

11-03.06 Corps of Engineers (Portland District) procedure. - An example of an index method using snowpack water equivalent as the independent variable is that derived by the Corps of Engineers, Portland District, for forecasting seasonal volume of runoff on the Columbia River near The Dalles, Oregon. As in the Weather Bureau method, forecasts were prepared for sub-basins. Because of the relatively large size of the Columbia River basin, some difficulty is experienced in selecting stations that properly represent the basin as a whole. The general form of the forecast equation used is

$$y = a\ (x_1 + x_2 \ldots\ldots x_9) + b$$

where the $\underline{x}$ values are forecasts for the sub-basins. The equations for the sub-basins are of the form

$$y = a\ x + b$$

where $\underline{x}$ is now an index representing the April 1st snowpack water equivalent. All relationships for the sub-basin forecasts are derived by graphical correlations. The period 1938 through 1953 was used for verification of the Columbia River forecast, this being the longest period for which adequate water equivalent data were available. Although the procedure was primarily developed for preparation of a forecast on April 1st, earlier forecasts can be made by extrapolating existing conditions to April 1st. The effects of spring precipitation or other factors were not directly included as parameters in the forecast procedure. Accordingly, forecasts made after April 1st do not directly take into account the effects of abnormal spring precipitation. Since the derivation of the procedure does not differentiate between effects of spring precipitation and effects of other factors, only subjective adjustments for abnormal spring precipitation can be made. The outstanding feature of the forecast method is its simplicity. Results could probably be improved by inclusion of other parameters which would evaluate spring precipitation and soil moisture deficits. However, such refinements would detract from its simplicity. The graphical derivation of the relationship between water equivalent and runoff permits subjective visual evaluation of the data, by which allowances may be made for unrepresentative conditions of precipitation or known deficiencies in the data.

11-03.07 Soil Conservation Service procedure. - A method utilizing both snowpack water equivalent and precipitation indexes has been developed by the Soil Conservation Service for forecasting seasonal runoff on the Columbia River near The Dalles, Oregon. Indexes used in this method are measures of the amount of water in storage in the snowpack on the date of forecast, usually April 1st, and the amount of water stored in the soil as the result of autumn precipitation. Basically, the forecast procedure consists of correlating April-through-June runoff with these indexes of water supply. Selection of the April-through-June runoff period was made with the objective of correlating volume of runoff with peak flow (see chapter 12). The forecast equation, developed from data for the period 1937 through 1950, is of the general form

$$Y = aX_1 + bX_2 + c$$

where X_1 is the snowpack water equivalent index and X_2 is the autumn precipitation index. For the May 1st forecast the equation is expanded to include an April precipitation index, X_3. A similar equation for forecasts issued on May 15th, uses an April 1st-to-May 15 precipitation index instead of the April index. The Y value in all cases is the April-through-June runoff. Snowpack water equivalent and spring precipitation indexes are determined for each of 8 sub-basins and then weighted in accordance with the average runoff contribution of each sub-basin to obtain the index for the Columbia River basin. Spring precipitation indexes are based on departures from normal published in USWB Climatological

Bulletins, the index being the average of the departures at stations representative of the sub-basin. It is noted that the effect of spring precipitation upon the seasonal runoff was not considered when correlating runoff with water supply in the derivation of the equation for the April 1st forecast. On the other hand, spring precipitation was considered important enough to warrant its inclusion as a variable in deriving the equations for the May 1st and May 15th forecasts. The omission of the spring precipitation parameter in the development of the equation may have a significant effect upon the coefficients of the $\underline{X_1}$ and $\underline{X_2}$ terms, thus significantly affecting the runoff values computed by the equation. The usefulness of a runoff forecast for the period April through June for the Columbia River near The Dalles, Oregon, is limited because of the variability of distribution of runoff in individual years. The average April-June runoff is 61 percent of the April-September runoff, but values for individual years range from 47 to 70 percent, depending upon the meteorologic sequences during the melt season. Since the sequence cannot be forecast on a long range basis, an additional variable which cannot be evaluated is introduced when forecasting for the April-through-June period.

11-03.08 Soil Conservation Service-Geological Survey procedure. - A method developed jointly by the Soil Conservation Service and Geological Survey incorporates the use of base flow as an index of the soil-moisture content. The method is similar to that described in the previous paragraph, the principal difference being the use of base flow instead of autumn precipitation for the soil-moisture index. Base flow is generally considered to be a good index of soil-moisture content because it integrates conditions over the entire basin. A disadvantage of using base flow is that it cannot always be accurately determined, particularly when it is necessary to separate base flow from that resulting from recent rain and/or snowmelt. Regardless of whether autumn precipitation or November 1st base flow is used as an index of soil moisture, it is assumed in these methods that no significant change in soil moisture occurs during the period from November 1st to the date of forecast. It should be recognized that a soil-moisture index which accounts for varying soil moisture deficits as of 1 November of each year does not necessarily represent the deficit which would occur on April 1st, the effective date of the forecast for which snow survey data are generally available. Also, there is some ambiguity as to whether a base-flow index is representing soil moisture or ground water deficits, or a combination of the two.

11-03.09 Coastal winter-flow index method. - An index method based primarily upon the relationship between winter runoff of low-elevation drainages in western Washington and Oregon, and the spring snowmelt runoff of the Columbia River was reported in Research Note 23. Indexes of winter temperature and spring precipitation are included in the forecast procedure as secondary parameters. The use of low-elevation winter flow as an index is confined to regions where the low-elevation and high-elevation areas have a common source of moisture. Such a situation exists in the region comprised of the Columbia River basin and western Washington and Oregon, the entire region being well centered in the belt of prevailing westerlies. Moisture is carried in a generally eastward direction from the Pacific Ocean, the amount being largely dependent upon the rate of the flow and precipitable water content of the air. The amount of moisture deposited over the region is a function of the moisture supply in the atmosphere and is reflected by both winter streamflow at low elevations and accumulation of snow at high elevations. If it is assumed that a given supply of moisture results in a fixed winter precipitation pattern over the entire region, precipitation at lower coastal mountains may be correlated with that at higher levels in inland mountain ranges. However, winter streamflow and accumulation of snow, as well as transpiration losses vary with temperature, necessitating the introduction of a temperature parameter. Likewise, amounts of seasonal runoff associated with given amounts of snow accumulation, vary with amounts of spring precipitation occurring over the high-elevation area, necessitating the introduction of a spring precipitation parameter. Indexes used in the forecast procedure were averages of observations for several representative stations. The relationship of the parameters to the runoff of the Columbia River was determined graphically, using a coaxial method similar to the one described in paragraph 11-02.18. A comparison of the reliability of various index procedures developed for forecasting seasonal runoff for the Columbia River shows that the low-elevation winter-flow index method is as accurate with regard to historical data as those which use precipitation and snow-course data for the principal index.

11-03.10 Plate 11-1 is a map of the Columbia River basin, and shows the location of the index streams and the spring precipitation station used in the winter-flow index method. Plate 11-2 shows the forecasting diagrams and scatter diagrams illustrating the relative reliability of forecasts made as of 1 March and continuing through 1 July. The procedure was developed by utilizing all known hydrologic data for the water year, as of 1 July. Forecasts made for earlier dates were derived by assuming average conditions of precipitation for the period subsequent to the date of forecast.

11-04. EXAMPLES OF WATER-BALANCE METHODS

11-04.01 General. - Procedures for forecasting runoff by the water-balance method consist of evaluating each of the water-balance components and summing them algebraically to determine runoff. In using historical records to develop the procedure, hydrologic events occurring both previous and subsequent to the date of forecast are evaluated. Application of the water-balance method, as well as any other method of seasonal runoff forecasting, necessitates the use of normal, forecast, or assumed values for events which occur after the date of forecast. The distinguishing feature of the water-balance procedure is that the effect of each factor upon runoff is in accordance with its actual value. It will be remembered that index procedures involve use of coefficients by which the index is multiplied to obtain the effect of a given factor upon runoff, the coefficients being evaluated in accordance with the integrated effect of all factors collectively.

11-04.02 Basically, all water-balance procedures for forecasting runoff are similar, differences being largely confined to the number of components considered and the method of their evaluation. The simplest type of water-balance procedure is one in which only the principal component is evaluated separately, the remaining components being evaluated collectively. In more complex procedures, more than one component is evaluated, and collective evaluations are confined to minor components only. A highly developed procedure is one in which all the significant components are evaluated separately by the best available means. An example of such a procedure is the development of the water balance for each of the snow laboratory areas, as described in chapter 4.

11-04.03 Example of simple water-balance procedure. - A simple water-balance procedure is that developed by Bean and Thomas 2/ primarily for forecasting minimum volume of runoff on the Androscoggin River basin in Maine (D.A. = 3430 sq. mi.). A computed volume of snowpack water equivalent was used as the primary determinant of volume of seasonal runoff. A relatively high density network of snow courses (approx. one per 50 sq. mi.) located through a wide range of elevation was used in computing basin snowpack water equivalent. The basin area was divided into elevation zones bounded by 500-ft. contours and mean water equivalent depths within each zone were determined. For high elevations where snow course data were lacking, values were extrapolated. Total basin values were obtained by summing the products of water-equivalent depth and the area of each zone. Losses were estimated to be 25 percent of the total amount of water contained in the snowpack. Thus, 75 percent of the snowpack water equivalent was considered to be a firm source of water supply. Although precipitation that occurs subsequent to the date of forecast cannot be accurately forecast, the additional runoff from this source can be estimated on the basis of past records. The feature of this method is that snowpack water equivalent, the component of prime importance, is computed with a relatively high degree of accuracy, whereas those of lesser importance are estimated.

11-04.04 Combination water balance-index procedures. - In cases where the use of index procedures is limited by a combination of short record and many variables, the number of variables may be reduced by introduction of water-balance evaluations. A typical example of such a procedure is that developed by the Walla Walla District, Corps of Engineers in 1953 for Boise River at Lucky Peak Dam (D.A. = 2650 sq. mi.).12/ A preliminary study indicated that the important variables to be considered were winter precipitation, April 1st snowpack water equivalent, and spring precipitation. Water stored in the snowpack was evaluated in accordance with methods described in chapter 3, using a snow chart constructed for the basin. Because of the limited number of years with adequate snow course records, it was believed that results could be improved by reducing the number of independent variables in the statistical correlation to two. Variables selected for inclusion in the regression equation were winter precipitation and April 1st snowpack water equivalent. The contribution of spring precipitation to runoff was found to be dependent to a great extent upon percent of area covered by snow. It was noted that precipitation falling on bare ground during the spring months did not produce significant rises in streamflow; it was therefore assumed that this precipitation was lost by evapotranspiration. Precipitation falling on snow was considered to be fully effective in producing runoff; that is, losses normally incurred by the snowpack are not increased as a result of precipitation falling on the snow. The contribution of effective spring precipitation to runoff was, therefore, considered to be the amount falling on the snow field. The dependent variable used in the correlation was observed generated runoff minus runoff from effective spring precipitation.

11-04.05 A feature of this method is that the independent variables (April 1st water equivalent and winter precipitation) in the regression equation are indexes whose values are known on the date of the forecast. Thus, revisions necessitated by occurrences of unexpected conditions during the forecast period may be made as conditions warrant. The amount of runoff expected from spring precipitation is computed separately, based upon occurrence of normal, assumed, or forecast spring precipitation and temperature. Separate computation of runoff resulting from factors effective during the forecast period permits easy revision of the runoff forecast where necessitated by the occurrence of unexpected conditions. Furthermore, the weighting of the prime variables in the regression equation is not affected by occurrences of unusual spring precipitation.

11-04.06 Forecasts for partial season. - Forecast procedures discussed thus far are for seasons ending after the snowpack water equivalent remaining on the ground is negligible. However, it is sometimes necessary to have a runoff forecast for a period ending prior to the end of the snowmelt season. Such a forecast, of course, necessitates determination of the runoff resulting from snowmelt during the forecast period. It is apparent that the accuracy of runoff forecasts for periods ending before all snow is depleted is largely dependent upon ability to forecast weather conditions subsequent to the date of forecast. With presently available means of forecasting weather, forecasts for periods of more than a few days are not sufficiently reliable to warrant their general use for forecasting seasonal runoff. Runoff resulting from conditions occurring after the date of forecast is best determined on the basis of normal or assumed weather conditions.

11-04.07 Because of limited accuracy of forecasts of weather for extended periods, direct computation of resultant runoff for periods ending before all snow is depleted is not justified. Equally good results can be obtained by preparing the forecast for the full melt season and subtracting the flow expected to occur after the termination of the period for which the forecast is desired. Such subsequent flow may be determined on the basis of past records. It is generally expressed in terms of percentage of total seasonal flow remaining after a given date. Obviously, such percentages will vary in accordance with conditions occurring during the melt season, and selection of the percentage used in the forecast is usually the normal percentage.

11-04.08 Application of water balance method to Detroit Project basin. - The most refined water-balance procedure for forecasting runoff is that in which each component is evaluated by the best available means. The water-balance derivations for the laboratory areas, described in chapter 4, are illustrative of such refined methods. However, instrumentation and observational facilities on the laboratory areas are far better than those on the average project basin. The water-balance procedure for forecasting seasonal runoff on the North Santiam River above Detroit Reservoir, Oregon, reported in Research Note 22, is considered representative of a method adaptable to an average project basin. Although the method is basically the same as that used on the small laboratory areas, deviations from these procedures are significant enough to warrant some explanation. For example, it will be noted that in the development of the procedure for Detroit Reservoir no mention is made of losses by interception. It will also be noted that no direct calculations of gage-catch deficiency due to wind were made in computing the basin precipitation. Omission of these items from the water-balance computations is not to be interpreted as failure to recognize their importance; their effects were considered in the computation of the net precipitation occurring over the basin. Wind records applicable to the precipitation gages were lacking, necessitating computation of net basin precipitation by indirect means. Net precipitation on the laboratory areas was obtained by subtracting interception loss from total basin precipitation, the latter having been computed by the isopercentual method, utilizing station values adjusted for gage-catch deficiency due to wind effect. For the Detroit Reservoir area, net precipitation for each water year was obtained by summing the generated runoff and evapotranspiration loss. Month-to-month variation in gage catch was accounted for by varying the ratio of basin to station precipitation, the ratios being derived from water-balance studies. Since no differentiation was made between total and net precipitation, the sum of runoff and evapotranspiration loss was designated simply as basin precipitation, a term comparable to net precipitation as used in the laboratory studies. Likewise, since interception loss was not computed as a separate component, the term <u>loss</u> refers to that resulting from evapotranspiration and change in soil moisture; that is, it does not include interception loss, as in the laboratory studies.

11-04.09 <u>Description of area</u>. - The North Santiam River basin above Detroit Reservoir (D.A. = 438 sq. mi.), is located on the west slope of the Cascade Mountains about 60 miles southeast of Portland, Oregon. Elevations range from 1200 feet at the damsite to 10,495 feet at the top of Mount Jefferson, the mean basin elevation being 3718 feet. A location map and area-elevation curve for the basin is shown on plate 11-3. A large percentage of the area is comprised of valleys and ridges with steep slopes, and a heavy stand of coniferous timber covers most of the area. In general, the area is underlain with rock of basalt formation which outcrops on many of the steep slopes, particularly at higher elevations. Soil cover is relatively thin, but there is considerable duff and litter under the heavy forest canopy.

11-04.10 Because of its location on the windward slope of the Cascade Range, the climate of the area is dominated by maritime influences during the entire year, except during short periods of continental airmass control. The climate is characterized by wet, moderately cold winters and dry, warm summers. Snow accumulates to great depths at higher levels during the winter months, temperatures being near freezing at these levels during most of the winter. Normal annual precipitation over the basin is estimated at 82 inches and ranges from less than 70 inches near Detroit to over 100 inches near Mount Jefferson. Records at Detroit indicate that about 60 percent of the annual precipitation occurs during the November-through-February period, largely in conjunction with the widespread storm activity. Precipitation during the June-through-September period comprises only about 10 percent of the annual amount, much of it occurring in convective-type storms. The percentage of precipitation occurring as snow is small at Detroit, but increases with elevation to approximately 75 percent at the 7000-foot level. The accumulation of snow over the basin as a whole generally increases from the beginning of the water year until April. At low levels, periods of depletion as well as accumulation occur throughout the snowfall season. Reference is made to the water balance for WBSL as presented in chapter 4, for a hydrologic summary of an area similar in character to that of the North Santiam River basin above Detroit Dam.

11-04.11 Hydrologic data available. - Precipitation, snowfall, streamflow, air-temperature, and snowpack water-equivalent data are available for varying periods. The streamflow record for North Santiam River above Mayflower Creek is directly applicable to the area above Detroit Reservoir, the drainage areas being nearly identical. The only adequate temperature record available is that at Detroit, necessitating use of lapse rates to obtain estimated temperatures at higher levels. Precipitation data are available for five stations of which one, Detroit, has a virtually continuous record since 1909. The remaining four stations, Santiam Pass, Santiam Junction, Marion Forks, and Breitenbush have short records with significant periods of missing data. Water equivalent is measured at four snow courses having records since 1941. Depth of snow on the ground is measured at Detroit, and supplementary snow surveys have been obtained since 1950 at two low-level stations, Detroit and Whitewater Bridge. Because of regulation of streamflow during the construction phase of Detroit Dam, the streamflow record subsequent to 1951 is not considered usable for study, thus limiting the hydrologic study to prior years. Since adequate snow-course data are not available for years prior to 1941, the period of record suitable for study is confined to the water years 1940-41 through 1950-51. Locations of hydrologic stations are shown on plate 11-3.

11-04.12 Analysis for forecast period ending August 31. - This phase of the analysis is applied to forecast periods ending on

August 31 at which time the snowpack remaining on the basin is negligible. The forecast procedure was developed for three periods: February through August, March through August, and April through August. The basic equation used for all periods is as follows:

$$Q_{gen} = P + (W_1 - W_2) - L \qquad (11\text{-}1)$$

in which $\underline{Q_{gen}}$ is generated runoff, $\underline{P}$ is precipitation, $\underline{W_1}$ and $\underline{W_2}$ are the initial and final snowpack water equivalents respectively, and $\underline{L}$ is loss. The final snowpack water equivalent, $\underline{W_2}$, is equal to zero in this case. Methods of evaluation of the terms of the equation are discussed in subsequent paragraphs.

11-04.13 The basin snowpack water equivalent was computed by use of a snow chart. Figure 1, plate 11-4 shows the snow chart and a sample determination of the snowpack water equivalent on February 1, 1954. Using the three key stations, Santiam Junction, Marion Forks, and Hogg Pass, a line was drawn representing the unadjusted mean depth of water equivalent over the basin. The line is drawn through points A and B, representing the mean depths and elevations of Marion Forks and Santiam Junction, and Santiam Junction and Hogg Pass, respectively. The unadjusted basin water equivalent is obtained by summing the zonal depths and dividing by 10. The values are shown in the tabulation accompanying the figure.

11-04.14 The factor by which the unadjusted value is multiplied to obtain the actual basin water equivalent, is derived from computed 11-year averages of precipitation, loss and runoff for the period September through December. These data are shown in the following tabulation:

Precipitation (P)	37.8 inches
Loss (L)	7.1 "
Generated Runoff (Q_{gen})	23.6 "

Substituting these values in equation 11-1, and considering the September 1 water equivalent (W_1) to be zero, the January 1 water equivalent (W_2) is calculated to be 7.1 inches. For the corresponding 11-year period, the average unadjusted water equivalent on January 1, obtained from the snow charts, is 9.4 inches. The adjustment factor is therefore 0.75 (7.1 divided by 9.4). Accordingly, the basin snowpack water equivalents indicated by the charts must be multiplied by 0.75 to obtain the actual basin values. Computed basin water equivalents for February 1, March 1, and April 1 of the 1941-51 period are shown in table 11-1.

11-04.15 Generated runoff, Q_{gen}, for each of the three forecast periods is computed by the method previously discussed. Observed runoff is converted to generated runoff by subtracting the recession volume of the initial flow and adding the recession volume of the terminal flow to the observed runoff during the period. Recession volumes are obtained from the scale on the right-hand side of figure 3, plate 11-5. Calculated generated flows are shown in table 11-1.

11-04.16 Losses, L, were computed by Thornthwaite's method for each of the years of the 11-year study period. Temperatures used in the computations are temperatures at the mean elevation of the basin and were obtained by applying estimates of lapse rate to the temperature at Detroit. As previously defined, loss is that portion of water supply which is lost to runoff and includes water retained in soil as well as that lost by evapotranspiration. Distribution of losses, averaged over the 11-year period, is shown in the following tabulation:

Period	Evapotranspiration (inches)	Retention in soil (inches)	Loss to runoff (inches)
September through December	3.3	3.8	7.1
January	0	0	0
February	0	0	0
March	0.3	0	0.3
April	1.3	0	1.3
May	2.5	-0.2	2.3
June	3.1	-1.0	2.1
July	3.3	-2.0	1.3
August	1.7	-0.6	1.1
Annual Total	15.5	0	15.5

Losses for the February-August, March-August and April-August periods for each year are shown in table 11-1 and in the bar diagrams in figure 4, plate 11-6.

11-04.17 Evapotranspiration losses during the period February through June are largely a function of monthly heat index, since there is sufficient water available to meet the potential demand. An estimate of losses for forecasting purposes can be obtained by using the relationship of previously computed evapotranspiration losses and corresponding monthly heat indexes at a representative station. Plotted points in figure 2, plate 11-5 show the relationship of heat index and evapotranspiration loss during March, April, May, and June of the 11-year study period, and the curves show the most probable amount for given heat indexes for each of the months. In the preparation of forecasts, the heat index, expressed as degree-days,* is based on occurrence of either forecasted, assumed, or normal temperatures during the forecast period. Table 11-2 shows the most probable heat index at Detroit for each of the ranges of temperature used in the U. S. Weather Bureau's Average Monthly Weather Résumé and Outlook. The Detroit temperature for each forecast range was determined by plotting long-term records of Portland temperatures versus Detroit temperatures and establishing the ranges for Detroit in accordance with those established for Portland by the U. S. Weather Bureau.

11-04.18 The average annual precipitation for the selected 11-year record was determined by adding the annual computed loss, 15.5 inches, to the annual runoff, 67.7 inches, to obtain the average annual basin precipitation of 83.2 inches. In computing monthly values of the water balance, it was found that the basin precipitation for January, February, and March, as determined from the single station at Detroit, weighted in accordance with the basin normal annual precipitation, produced more reliable results than those obtained using several stations. The recording gages at Marion Forks, Santiam Pass, and Santiam Junction all had significant periods of missing data during the winter months. However, during the period of April through August, the records of all stations appear reliable and were, therefore, used in computing basin precipitation for this period.

11-04.19 Bar diagrams illustrating the water balance for the various forecast periods of each of the years, 1941 through 1951, are shown in figure 4, plate 11-6. The first bar in each group represents the snowpack water equivalent in inches over the basin at the beginning of the forecast season. Total precipitation occurring during the forecast season is represented by the second bar. Total length of the third bar represents the sum of the water equivalent and precipitation; the hatched portion represents loss, and the unhatched portion shows the amount available for runoff. Actual generated runoff is depicted by the fourth bar. Figure 2, plate 11-6, shows graphically the correlation between computed and actual generated runoff values.

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* Daily maximum temperatures above 32°F.

11-04.20 As mentioned previously, the water balance method of forecasting runoff permits use of the U. S. Weather Bureau's Average Monthly Weather Résumé and Outlook. The expected precipitation given in the Outlook may be used for the first 30 days of a forecast period, assumed or normal values being used for the remainder of the period. Table 11-3 shows the most probable basin precipitation for each of the ranges used in the Outlook for each of the months of February through June. The Detroit precipitation for each forecast range for each month was determined by plotting long-term records of Portland precipitation versus Detroit precipitation, and establishing the ranges for Detroit in accordance with those established for Portland by the U. S. Weather Bureau. Most probable basin amounts for given amounts at Detroit for the months of April through August were derived from records of all stations in the basin, adjustments for gage catch being made in accordance with meteorological characteristics of each month. Adjustments are such that the sum of the precipitation amounts for these months is in agreement with the April-through-August total computed by the water balance equation. Because of the relative insignificance of the July and August precipitation, no effort is made to classify the amounts by ranges. Instead, the average of 1.9 inches of basin precipitation for the two months is used for the expected amount.

11-04.21 Charts depicting the water balance for the entire year based on averages for the 11-year study period are shown in figures 1 and 3, plate 11-6. The graph at the bottom of figure 1 shows the change in the amount of water in ground storage; positive values indicate increases and negative values indicate decrease in ground and channel storage. Cumulative totals of the water-balance components for the year beginning on September 1 are shown in figure 3.

11-04.22 Analysis for forecasting by months. - As previously stated, the water-balance method of forecasting is developed with the objective of forecasting runoff for periods terminating before the end of the melt season as well as for periods ending at the completion of snow melt. For forecast periods ending prior to the end of the melt season, specifically with the months of January, February, and March, the water balance equation used is

$$Q_{gen} = P_r + M - L \qquad (11\text{-}2)$$

where Q_{gen} is generated runoff, P_r is basin rainfall, M is basin melt, and L is basin loss, all in inches. The basin melt is obtained by applying the monthly degree-days at Detroit and percentage of basin covered by snow to the charts shown in figures 3 and 4, plate 11-4. The percentage of snow cover is determined by a correlation of snow-course data and aerial reconnaissance data. Melts based on an arbitrary 0.01 inches per degree-day melt rate are obtained from the melt charts, using the degree-days* and percentage of cover for the given month. Corrected melt is obtained by multiplying the value from the chart by factors derived from runoff and degree-day relationships during rain-free periods in the basin. Approximate factors are as follows:

Month	Correction factor
January	1.0
February	1.4
March	1.8

11-04.23 To compute the probable percentage of the precipitation that will occur as rain during a given forecast period, the amount occurring as rain was determined for each of the months, January through March, of the 11-year study period. Snowfall, obtained by adding algebraically the melt and the change in snowpack water equivalent, was subtracted from basin precipitation to obtain rainfall, which, in turn, was expressed in terms of percent of basin precipitation. The percentages are plotted as a function of the number of degree-days* at Detroit as shown in figure 1, plate 11-5. The most probable percentages for use in forecasting are indicated by the curves drawn on the chart. Computations of the water balance for each of the months are shown in table 11-4.

11-04.24 Preparing the forecast. - Having used historical data to establish criteria for evaluating the components of the water balance, the criteria may be applied to forecasting runoff for a given period. Forecasts for the Detroit project are confined to seasons ending on August 31, at which time the snowpack is negligible. Steps in the preparation of the forecast are as follows: (1) evaluate snowpack water equivalent on the initial day of the forecast period; (2) determine precipitation expected during the forecast period; (3) determine loss expected during the forecast period; (4) take algebraic sum of (1), (2), and (3) above; (5) add antecedent recession volume and subtract estimated

* Daily maximum temperatures above 32°F.

terminal-recession volume to obtain runoff for forecast period. Recession volumes for given flows are shown in figure 3, plate 11-5.

11-04.25 Conclusion. - The foregoing computations of the components of the water balance for the North Santiam River above Detroit illustrate the adaptability of the water-balance principle to the development of forecast procedures. Since runoff forecasts made by any method are subject to inevitable errors arising from the inability to foresee unusual hydrometeorological events that will occur during the forecast period, any forecast method should be flexible enough to permit easy revision of the forecast to account for such unusual events as they occur. The water-balance method, by virtue of its inherent adaptability to revision, meets this important requirement.

11-05. SUMMARY

11-05.01 Index procedures rely upon the variance of the independent variables to establish their relationship with the dependent variable. The magnitude of the derived coefficients is a function of the units of measurement as well as the conditions of measurement at the point of observation in relation to basin averages. Therefore, the coefficients do not necessarily have any physical significance in the relationship. In addition, the coefficients which provide the best solution for the years of record used in developing the equation are not necessarily the best for application to other years. This results from improper weightings of the variables in arriving at a best fit of the historical data. Indexes should be selected on the basis of representing known physical processes. Since the coefficients have no physical significance, there is little possibility to check them rationally, except in extreme cases. The use of index relationships is valuable, however, in establishing weightings of variables known to represent physical processes, but it should be recognized that such weightings may vary with different periods of record used in their derivation. The weightings of the variables should be based on complete indexes of water-balance components for the entire water year. Forecasts should use these weightings both for conditions known at the time of the forecast, and for normal or assumed conditions subsequent to the forecast date. The principal limitation of index procedures results from inadequate lengths of record of basic data for statistical analysis. Although it is desirable to include indexes of all important variables affecting runoff, the number of variables that can be used with confidence is limited by the length of the historical record. By contrast, evaluation of the components in the water-balance method is not dependent upon length of record. Although historical data are used in the development of the method, the forecast of runoff is based on an appraisal of each component for the current year rather than upon the effect produced by a given set of conditions in past years.

11-05.02 It has been pointed out that the volume of seasonal runoff is dependent not only upon the magnitude of individual components, but also upon the interrelationship of these components. For example, losses from spring and summer precipitation are a function not only of total moisture supply, but are also dependent upon the areal extent of the snow cover during the spring and summer and, hence, indirectly upon the maximum snowpack accumulation. In water-balance computations, such interrelationships where they are important enough to warrant consideration, are taken into account in a rational way in the computation of the individual components. Similarly, in the index approach to seasonal runoff forecasting, the individual indexes which determine runoff should each be a rational expression of the particular parameter, including any interrelationships that exist. Neither the water-balance method nor the index method of weighting the several components will, in itself, evaluate such interrelationships.

11-05.03 The sparsity of data on many project basins imposes limitations upon the accuracy with which water-balance computations can be made. Although the true values of the components may never be exactly known, satisfactory results are usually obtainable by use of computed values. Errors in the computation of the components of the water balance are known to exist when the values fail to show a balance in the application of the water balance equation to past data. Although it is recognized that the existence of a balance does not necessarily indicate correct evaluation of each component, it is highly probable that the component values are reasonably accurate if they consistently provide a balance under varying conditions. Failure of the components to balance indicates that further refinement is necessary.

11-05.04 The reliability of both water-balance and index methods is largely dependent upon the hydrologic data available for development and application of the methods. No definite rules can be made regarding the reliability of each of the methods; final appraisal of the methods is made largely on the basis of results obtainable by each. It is probable that better results would be obtained by the index method on project basins having records of 25 or more years duration, particularly if the areal coverage instrumentation is not good. On the other hand, records of less than 10-years duration are generally inadequate for development of forecast procedures by index methods.

11-05.05 The usual criterion of accuracy for forecasting procedures is the relative degree of correlation obtained by each procedure on the basis of historical record. Although it is desirable to obtain a high degree of correlation with historical data for a derived relationship, that should not be the only basis of judgement. Of even greater importance is the rational selection of variables affecting runoff. Unless it can be shown that all of the variables which significantly affect runoff are accounted for in the forecast equation, and that the effect of each variable is in the correct order of magnitude from the standpoint of known physical relationships, little reliance can be placed on the statistically derived relationship regardless of the degree of correlation. A line of best fit for a relatively few years of historical data for a relationship derived from incomplete indexes of the water-balance components will sometimes show a higher degree of correlation for an early-season forecast than a procedure derived from complete water-balance indexes and applied to the early date of forecast. The greater accuracy of the former is meaningless and reflects only the forcing of the regression to obtain the best fit of data which do not adequately represent the entire runoff process.

11-05.06 Because of the wide variation in problems associated with seasonal runoff forecasting, definite recommendations regarding choice of forecast methods to be used cannot be made. The adoption of certain methods may be immediately ruled out by lack of adequate data. In some instances the data may be inadequate for development of acceptable forecast procedures regardless of the method employed, necessitating development or expansion of a hydrologic network to provide the required data.

11-05.07 Although forecast procedures of limited refinement may be adequate for given projects, consideration should be given to possible future development of water uses. Since length of hydrologic records is an important factor in the development of forecast procedures, future needs should be anticipated far enough in advance to permit establishment of a hydrologic network for providing an adequate record of hydrologic data. The requirements for the hydrologic network should be considered in the light of the hydrologic character of the area involved and anticipated requirements for forecasts. Site selection for obtaining point observations of the principal elements should be made on the basis of obtaining representative samples for the area involved, as set forth in chapters 3 and 4. A final incentive for improving forecast techniques is the knowledge that better seasonal runoff forecasts make possible better utilization of water supply, thus contributing toward development of additional uses of water resources.

11-06. REFERENCES

1/ ARKIN, Herbert and Raymond R. Colton, An Outline of Statistical Methods, (College Outline Series), Barnes & Noble, Inc., New York.

2/ BEAN, Paul L. and Philip W. Thomas, "A quantitative forecast-system for power- and flood-warning in the Androscoggin River basin, Maine," Trans. Amer. Geophys. Union, Pt. III, September 1940, pp. 835-846.

3/ BROOKS, C.E.P. and N. Carruthers, Handbook of Statistical Methods in Meteorology, (M.O. 538, Air Ministry, Meteorological Office), H. M. Stationery Office, London, 1953.

4/ EZEKIEL, Mordecai, Methods of Correlation Analysis, John Wiley & Sons, Inc., New York, 1941.

5/ KOELZER, Victor A., "Cumulative snowmelt runoff distribution graphs and their use in runoff forecasting," Proc. Western Snow Conference, April 1951, pp. 30-39.

6/ KOHLER, M. A. and R. K. Linsley, Jr., "Recent developments in water supply forecasting from precipitation," Trans. Amer. Geophys. Union, Vol. 30, No. 3, June 1949, pp. 427-436.

7/ LINSLEY, Ray K., Max A. Kohler, and Joseph L. H. Paulhus, Applied Hydrology, McGraw-Hill Book Company, Inc., New York, 1949.

8/ POTTS, H. L., "A photographic snow-survey method of forecasting runoff," Trans. Amer. Geophys. Union, Vol. 25, September 1944, pp. 149-153.

9/ SNEDECOR, George W., Statistical Methods, (4th Ed.), The Iowa State College Press, Ames, Iowa, 1946.

10/ STRAUSS, Fred A., "A revision of forecasting methods as practiced by the California Cooperative Snow Surveys," Proc. Western Snow Conference, April 1950, pp.49-59

11/ WILM, H. G., "Statistical control in hydrologic forecasting," Pacific Northwest Forest and Range Experiment Station, Research Note No. 61, January 1950.

12/ WATER MANAGEMENT SUBCOMMITTEE, CBIAC, "Review of procedures for forecasting seasonal runoff of Boise River above Diversion Dam, Idaho," January 1954.

13/ WATER MANAGEMENT SUBCOMMITTEE, CBIAC, "Review of procedures for forecasting seasonal runoff of Columbia River near The Dalles, Oregon," August 1954.

TABLE 11-1

WATER BALANCE BY FORECAST SEASON
North Santiam River above Detroit Dam

ITEM	1941	1942	1943	1944	1945	1946	1947	1948	1949	1950	1951	TOTAL	MEAN
						APRIL Thru AUGUST							
1. Precipitation 1/	17.42	13.22	16.10	10.00	14.27	8.60	14.63	16.18	8.78	11.05	5.98	136.23	12.38
2. April 1st W 2/	4.05	12.08	24.82	7.72	11.48	27.00	11.85	19.13	33.49	28.35	24.08	204.05	18.55
3. Supply (1)+(2)	21.47	25.30	40.92	17.72	25.75	35.60	26.48	35.31	42.27	39.40	30.06	340.28	30.93
4. Loss 3/	11.09	8.31	9.08	6.02	5.13	8.42	11.96	9.25	6.45	8.36	5.10	89.17	8.11
5. Computed R.O. (3)-(4)	10.38	16.99	31.84	11.70	20.62	27.18	14.52	26.06	35.82	31.04	24.96	251.11	22.82
6. Generated R.O. 4/	10.43	14.41	29.65	12.74	21.90	24.43	16.11	29.37	35.20	35.44	21.33	250.01	22.73
7. Percent Deviation	-0.5	17.9	7.4	-8.2	-1.3	11.2	-9.9	-11.3	1.8	-12.4	17.0	0.4	0.4
						MARCH Thru AUGUST							
1. Precipitation	20.25	17.89	27.04	15.52	27.19	18.44	23.40	24.12	14.91	25.77	15.79	230.32	20.94
2. March 1st W	6.30	11.64	24.90	7.20	5.89	23.81	11.18	15.19	34.69	26.74	18.45	185.99	16.91
3. Supply	26.55	29.53	51.94	22.72	33.08	42.25	34.58	39.31	49.60	52.51	34.24	416.31	37.85
4. Loss	12.39	8.91	9.08	6.32	5.33	8.52	12.76	9.25	6.65	8.36	5.10	92.67	8.43
5. Computed R.O.	14.16	20.62	42.86	16.40	27.75	33.73	21.82	30.06	42.95	44.15	29.14	323.64	29.42
6. Generated R.O.	13.28	19.26	38.31	17.75	27.40	31.19	23.36	34.90	43.19	46.97	27.02	322.63	29.33
7. Percent Deviation	6.6	7.1	11.9	-7.7	1.3	8.1	-6.6	-13.9	-0.6	-6.0	7.8	0.3	0.3
						FEBRUARY Thru AUGUST							
1. Precipitation	23.89	26.58	36.47	23.28	42.83	28.70	27.90	39.34	38.32	38.94	28.58	354.83	32.26
2. February 1st W	5.85	6.75	24.68	4.05	3.19	18.68	12.04	7.91	21.45	23.92	17.25	145.77	13.25
3. Supply	29.74	33.33	61.15	27.33	46.02	47.38	39.94	47.25	59.77	62.86	45.83	500.60	45.51
4. Loss	12.39	8.91	9.08	6.32	5.33	8.52	12.76	9.25	6.65	8.36	5.10	92.67	8.43
5. Computed R.O.	17.35	24.42	52.07	21.01	40.69	38.86	27.18	38.00	53.12	54.50	40.73	407.93	37.08
6. Generated R.O.	16.23	24.68	48.61	21.73	37.96	36.89	31.02	43.77	52.50	56.48	37.08	406.95	37.00
7. Percent Deviation	6.9	-1.1	7.1	-3.3	7.2	5.3	-12.4	-13.2	1.2	-3.5	9.8	0.2	0.2

1/- Precipitation values are inches over basin.
2/- W is snowpack water equivalent expressed in inches over basin.
3/- Loss is computed by Thornthwaite's method, and is expressed in inches over basin.
4/- Generated runoff is actual runoff minus initial recession-volume plus terminal recession-volume.

TABLE 11-2

MONTHLY TEMPERATURE AND HEAT INDEX RANGES

DETROIT, OREGON

Forecast 1/	Range		Most Probable	
	Temperature (°F)	Heat Index (Degree-Days) 2/	Temperature (°F)	Heat Index (Degree-Days)
		February		
Much above	41.4 or more	558 or more	42.6	650
Above	39.1 - 41.3	472 - 557	40.2	500
Normal	37.2 - 39.0	382 - 471	37.6	400
Below	37.1 - 33.2	381 - 242	35.0	300
Much below	33.1 or less	241 or less	32.0	200
		March		
Much above	45.6 or more	801 or more	47.0	850
Above	42.1 - 45.5	701 - 800	43.8	750
Normal	39.6 - 42.0	550 - 700	41.3	600
Below	39.5 - 38.0	549 - 400	38.8	500
Much below	37.9 or less	399 or less	37.2	350
		April		
Much above	50.6 or more	1001 or more	51.2	1050
Above	48.3 - 50.5	901 - 1000	49.4	950
Normal	46.2 - 48.2	800 - 900	47.3	850
Below	46.1 - 44.0	799 - 700	45.2	750
Much below	43.9 or less	699 or less	43.6	650
		May		
Much above	57.7 or more	1301 or more	58.0	1350
Above	54.5 - 57.6	1201 - 1300	56.1	1250
Normal	52.5 - 54.4	1100 - 1200	53.8	1150
Below	52.4 - 51.0	1099 - 1000	51.7	1050
Much below	50.9 or less	999 or less	50.4	900
		June		
Much above	61.1 or more	1401 or more	61.6	1450
Above	59.1 - 61.0	1301 - 1400	60.0	1350
Normal	57.9 - 59.0	1200 - 1300	58.6	1250
Below	57.8 - 56.6	1199 - 1100	57.2	1150
Much below	56.5 or less	1099 or less	56.0	1050

1/ Forecast designations are those used in the U. S. Weather Bureau Average Monthly Weather Résumé and Outlook, and corresponding temperatures are mean monthly values.

2/ Heat indexes are based on computations for each of the months of the years 1941 through 1951 and are monthly totals of maximum temperatures above base 32°F.

TABLE 11-3

MONTHLY PRECIPITATION RANGES

NORTH SANTIAM RIVER ABOVE DETROIT DAM

Forecast 1/	Range		Most Probable	
	Detroit (inches)	Basin 2/ (inches)	Detroit (inches)	Basin (inches)
		February		
Heavy	11.2 or more	13.4 or more	13.7	16.4
Moderate	11.1 - 6.1	13.3 - 7.3	8.7	10.4
Light	6.0 or less	7.2 or less	3.8	4.5
		March		
Heavy	8.9 or more	9.9 or more	12.8	14.3
Moderate	8.8 - 4.8	9.8 - 5.3	7.4	8.3
Light	4.7 or less	5.2 or less	3.8	4.2
		April		
Heavy	7.0 or more	6.3 or more	8.5	7.6
Moderate	6.9 - 4.3	6.2 - 3.8	5.0	4.5
Light	4.2 or less	3.7 or less	2.6	2.3
		May		
Heavy	5.1 or more	4.5 or more	6.0	5.3
Moderate	5.0 - 3.6	4.4 - 3.2	3.9	3.4
Light	3.5 or less	3.1 or less	1.7	1.5
		June		
Heavy	3.2 or more	2.8 or more	4.3	3.8
Moderate	3.1 - 1.6	2.7 - 1.4	2.4	2.1
Light	1.5 or less	1.3 or less	0.9	0.8

1/ Forecast designations are those used in the U. S. Weather Bureau Average Monthly Weather Resume and Outlook.

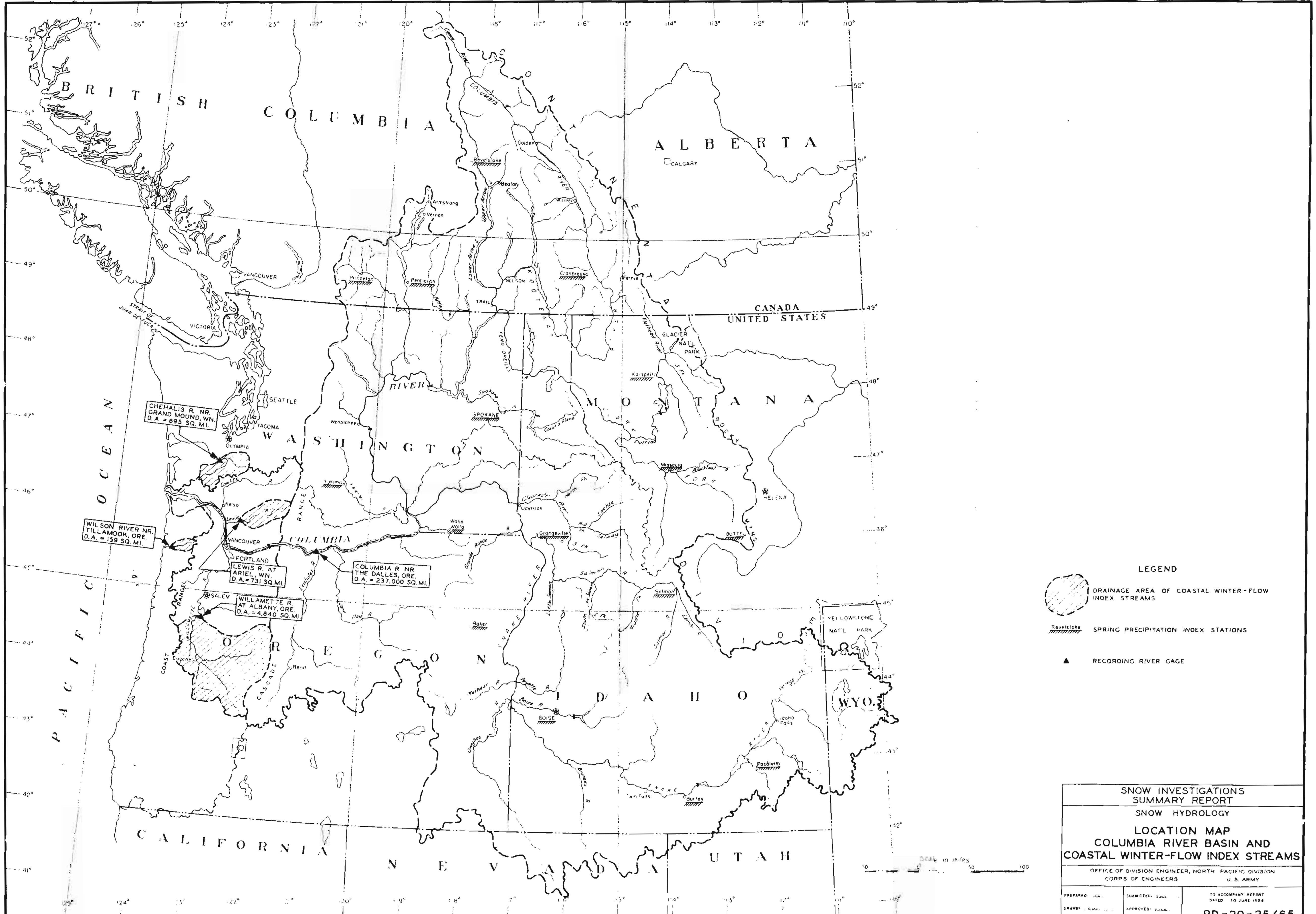
2/ Basin values are determined by water-balance studies for the period 1941 through 1951.

TABLE 11-4

WATER BALANCE BY MONTHS
North Santiam River above Detroit Dam

ITEM 1/	1941	1942	1943	1944	1945	1946	1947	1948	1949	1950	1951	TOTAL	MEAN
						JANUARY							
(1) Degree Days 2/	420.00	353.00	239.00	303.00	385.00	266.00	183.00	368.00	154.00	78.00	218.00		
(2) Percent Snow Cover	54.00	73.00	87.00	64.00	58.00	90.00	60.00	65.00	92.00	98.00	82.00		
(3) Melt	1.07	1.23	0.83	0.80	1.05	1.02	0.25	1.20	0.45	0.15	0.68	8.73	0.79
(4) Change in Water-Equivalent	3.41	3.77	10.46	1.84	0.90	7.24	6.82	3.71	3.90	13.88	11.81	67.74	6.16
(5) Snowfall (3)+(4)	4.48	5.00	11.29	2.64	1.95	8.26	7.07	4.91	4.35	14.03	12.49	76.47	6.95
(6) Basin Precipitation	10.06	8.02	15.64	7.08	10.31	15.93	13.13	15.60	3.74	25.48	21.73	146.72	13.34
(7) Snowfall	4.48	5.00	11.29	2.64	1.95	8.26	7.07	4.91	4.35	14.03	12.49	76.47	6.95
(8) Rainfall (6)-(7)	5.58	3.02	4.35	4.44	8.36	7.67	6.06	10.69	-0.61	11.45	9.24	70.25	6.57
(9) Supply (3)+(8)	6.65	4.25	5.18	5.24	9.41	8.69	6.31	11.89	-0.16	11.60	9.92	78.98	7.18
(10) Loss 3/	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
(11) Computed Runoff (9)-(10)	6.65	4.25	5.18	5.24	9.41	8.69	6.31	11.89	-0.16	11.60	9.92	78.98	7.18
(12) Generated Runoff 4/	6.57	5.24	7.08	4.38	8.39	9.85	7.12	11.64	1.50	6.80	9.91	78.48	7.13
(13) Percent Rainfall	55.00	38.00	28.00	63.00	81.00	48.00	46.00	69.00	-	45.00	43.00		
						FEBRUARY							
(1) Degree Days	536.00	423.00	538.00	403.00	371.00	309.00	514.00	306.00	278.00	281.00	369.00		
(2) Percent Snow Cover	49.00	90.00	88.00	54.00	66.00	87.00	62.00	75.00	93.00	100.00	87.00		
(3) Melt	2.24	2.52	3.50	1.09	1.26	1.47	1.99	1.15	1.40	1.75	1.89	20.26	1.84
(4) Change in Water-Equivalent	0.45	4.88	0.23	3.15	2.70	5.14	-0.86	7.28	13.24	2.82	1.20	40.23	3.66
(5) Snowfall (3)+(4)	2.69	7.40	3.73	4.24	3.96	6.61	1.13	8.43	14.64	4.57	3.09	60.49	5.50
(6) Basin Precipitation	3.64	8.69	9.43	7.76	15.64	10.26	4.50	15.22	23.41	13.17	12.79	124.51	11.32
(7) Snowfall	2.69	7.40	3.73	4.24	3.96	6.61	1.13	8.43	14.64	4.57	3.09	60.49	5.50
(8) Rainfall (6)-(7)	0.95	1.29	5.70	3.52	11.68	3.65	3.37	6.79	8.77	8.60	9.70	64.02	5.82
(9) Supply (3)+(8)	3.19	3.81	9.20	4.61	12.94	5.12	5.36	7.94	10.17	10.35	11.59	84.28	7.66
(10) Loss	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
(11) Computed Runoff (9 -(10)	3.19	3.81	9.20	4.61	12.94	5.12	5.36	7.94	10.17	10.35	11.59	84.28	7.66
(12) Generated Runoff	2.95	5.42	10.30	3.98	10.56	5.70	7.66	8.87	9.31	9.51	10.06	84.32	7.67
(13) Percent Rainfall	26.00	15.00	60.00	45.00	75.00	36.00	75.00	45.00	37.00	65.00	76.00		
						MARCH							
(1) Degree Days	966.00	707.00	578.00	592.00	465.00	510.00	767.00	499.00	545.00	408.00	422.00		
(2) Percent Snow Cover	27.00	72.00	88.00	57.00	68.00	91.00	57.00	82.00	88.00	100.00	82.00		
(3) Melt	2.80	5.56	5.16	2.90	2.16	4.20	4.70	3.42	4.70	3.14	2.40	41.14	3.74
(4) Change in Water-Equivalent	-2.25	0.45	-0.08	0.52	5.59	3.19	0.68	3.94	-1.20	1.60	5.55	17.99	1.64
(5) Snowfall (3)+(4)	0.55	6.01	5.08	3.42	7.75	7.39	5.38	7.36	3.50	4.74	7.95	59.13	5.38
(6) Basin Precipitation	2.83	4.67	10.94	5.52	12.92	9.84	8.77	7.94	6.13	14.72	9.81	94.09	8.56
(7) Snowfall	0.55	6.01	5.08	3.42	7.75	7.39	5.38	7.36	3.50	4.74	7.95	59.13	5.38
(8) Rainfall (6)-(7)	2.28	-1.34	5.86	2.10	5.17	2.45	3.39	0.58	2.63	9.98	1.86	34.96	3.18
(9) Supply (3)+(8)	5.08	4.22	11.02	5.00	7.33	6.65	8.09	4.00	7.33	13.12	4.26	76.10	6.92
(10) Loss	1.30	0.60	0.00	0.30	0.20	0.10	0.80	0.00	0.20	0.00	0.00	3.50	0.32
(11) Computed Runoff (9)-(10)	3.78	3.62	11.02	4.70	7.13	6.55	7.29	4.00	7.13	13.12	4.26	72.60	6.60
(12) Generated Runoff	2.85	4.85	8.66	5.01	6.50	6.76	7.25	5.53	7.99	11.53	5.69	72.62	6.60
(13) Percent Rainfall	81.00	-	54.00	38.00	40.00	25.00	39.00	7.00	43.00	68.00	19.00		

1/- Items (3) through (12) are expressed in inches over basin.
2/- Degree days are monthly totals of degrees of maximum temperatures above base 32° F.
3/- Loss is computed by Thornthwaite's method.
4/- Generated runoff is actual runoff minus initial recession volume plus terminal recession volume.



SNOW INVESTIGATIONS SUMMARY REPORT

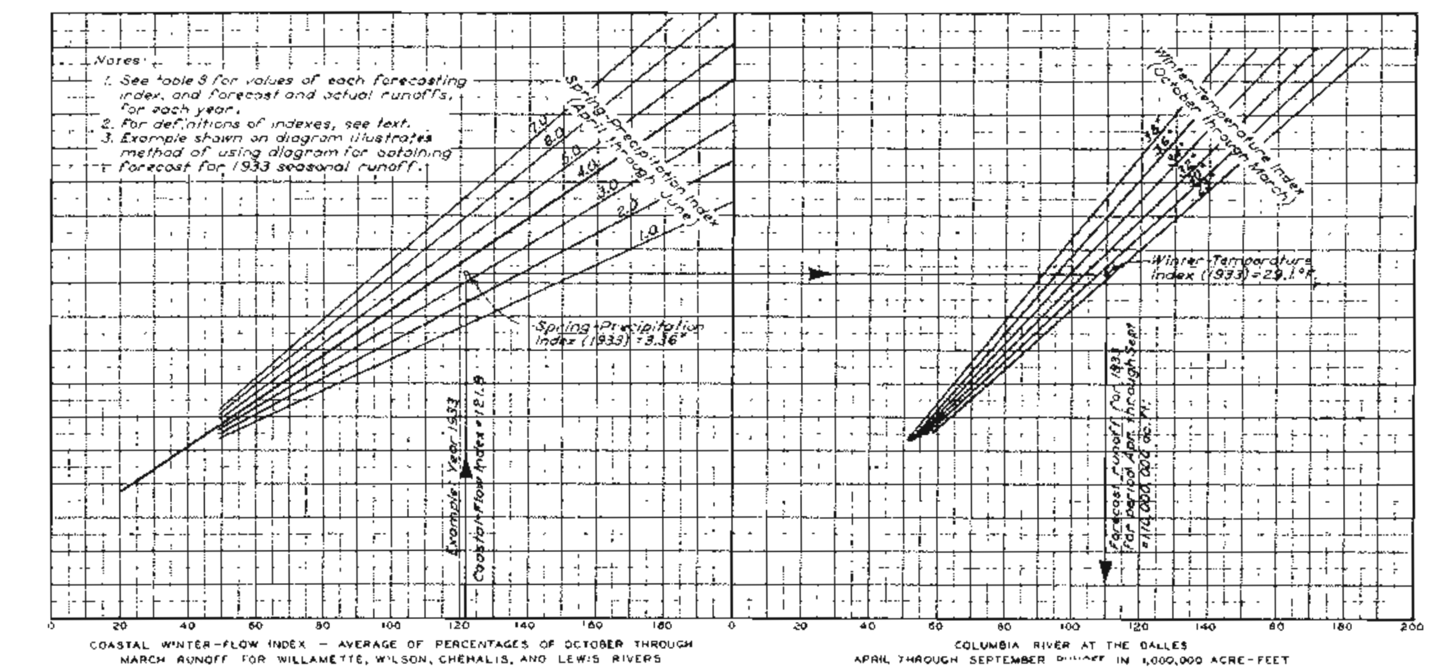
SNOW HYDROLOGY

LOCATION MAP
COLUMBIA RIVER BASIN AND
COASTAL WINTER-FLOW INDEX STREAMS

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

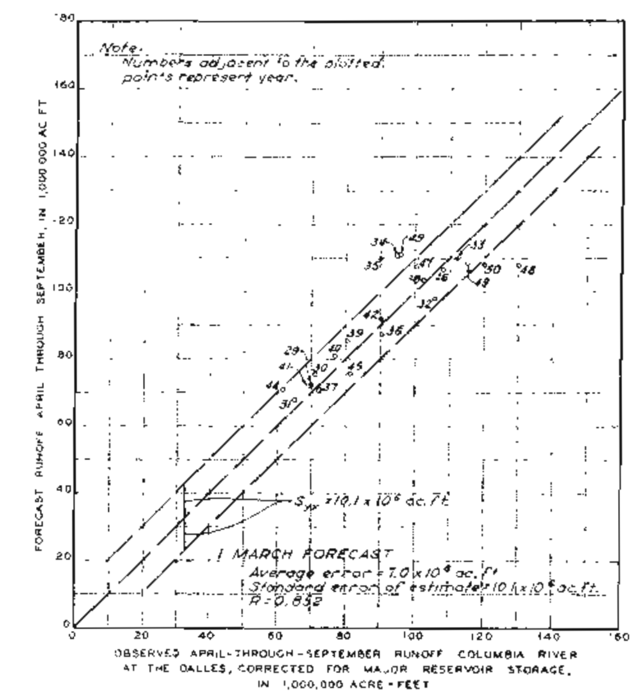
TO ACCOMPANY REPORT DATED 30 JUNE 1956

PD-20-25/65



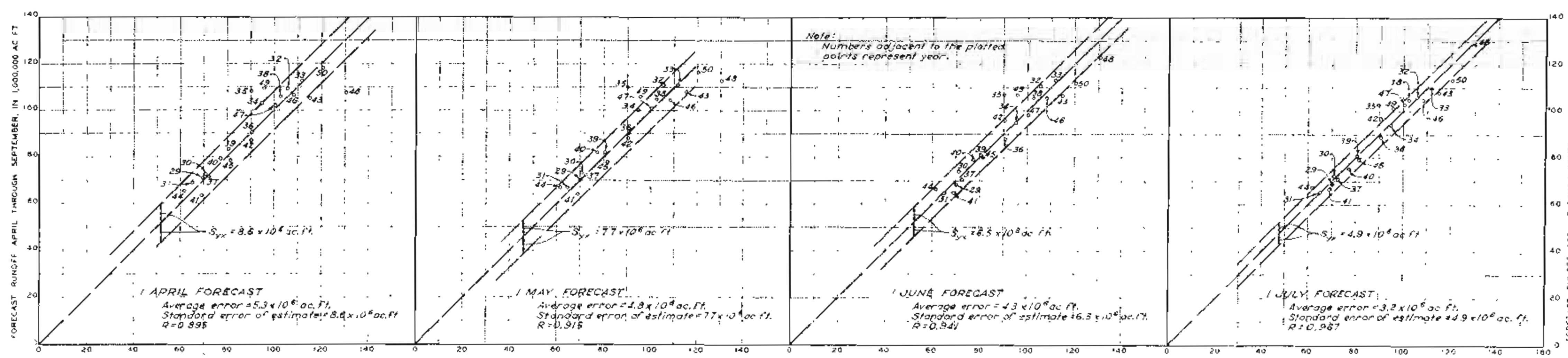
COAXIAL DIAGRAM FOR FORECASTING COLUMBIA RIVER APRIL-THROUGH-SEPTEMBER RUNOFF AT THE DALLES

FIGURE 1



FORECAST VS ACTUAL RUNOFF
(1 MARCH FORECAST)

FIGURE 2



FORECAST VS ACTUAL RUNOFF
1 APRIL, 1 MAY, 1 JUNE, AND 1 JULY FORECASTS

FIGURE 3

SNOW INVESTIGATIONS
SUMMARY REPORT

SNOW HYDROLOGY

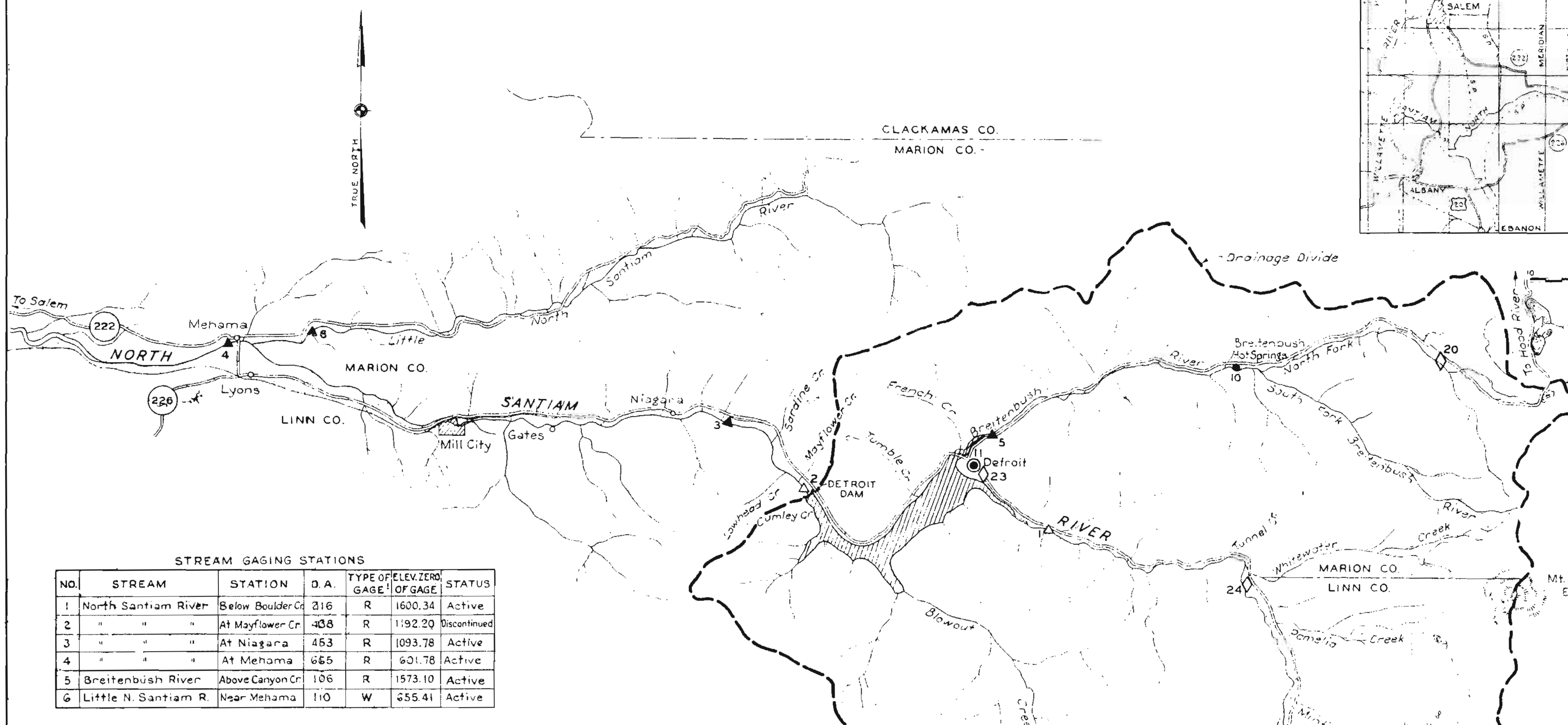
FORECASTING DIAGRAMS
COASTAL WINTER-FLOW INDEX METHOD

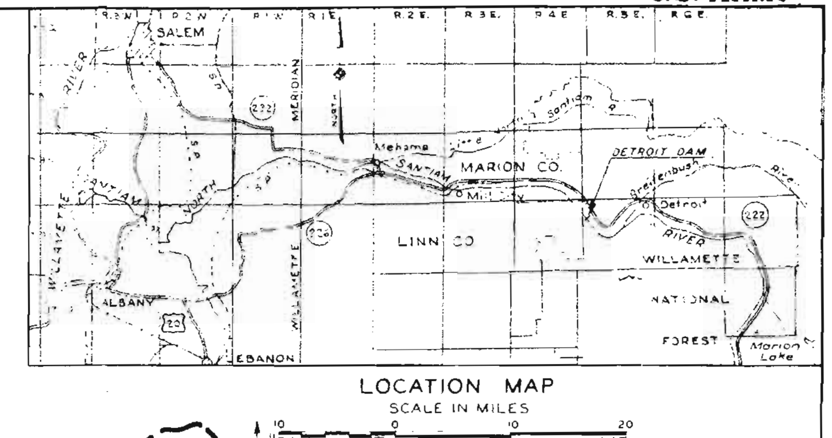
COLUMBIA RIVER NEAR THE DALLES

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U.S. ARMY

PD-20-25/66

PLATE 11-2





LOCATION MAP
SCALE IN MILES

LEGEND

- ▲ Active Stream Gaging Station
- △ Discontinued Stream Gaging Station
- ● Recording Precipitation Gage
- ○ Non-recording Precipitation Gage
- ◇ Snow Survey Courses
- Detroit Reservoir

Notes:
1. Types of Gages
   Stream
   R - Recorder
   W - Wire weight
   Precipitation
   N - Non-recorder
   R - Recorder
2. Hogg Pass not shown on map; located about 8 miles south of Three Fingered Jack.

STREAM GAGING STATIONS

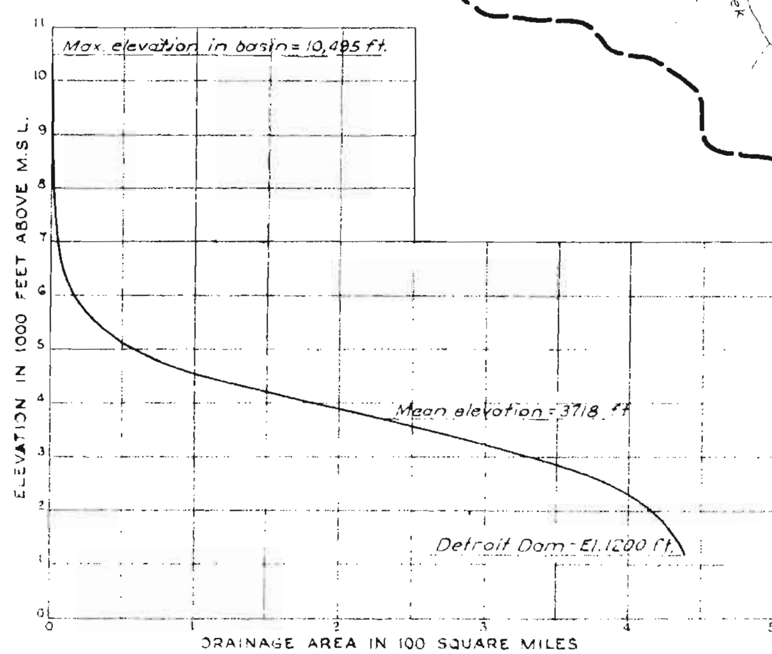
NO.	STREAM	STATION	D.A.	TYPE OF GAGE[1]	ELEV. ZERO OF GAGE	STATUS
1	North Santiam River	Below Boulder Cr.	216	R	1600.34	Active
2	" " "	At Mayflower Cr.	438	R	1192.20	Discontinued
3	" " "	At Niagara	453	R	1093.78	Active
4	" " "	At Mehama	655	R	601.78	Active
5	Breitenbush River	Above Canyon Cr.	106	R	1573.10	Active
6	Little N. Santiam R.	Near Mehama	110	W	655.41	Active

PRECIPITATION STATIONS

MAP NUMBER	STATION	ELEVATION	TYPE OF GAGE
10	Breitenbush	2225	R
11	Detroit	1452	N & R
12	Marion Forks	2445	R
13	Santiam Junction	3750	R
14	Santiam Pass	4800	R

SNOW SURVEY COURSES

MAP NUMBER	NAME	ELEVATION
20	Breitenbush	2325
2	Hogg Pass	4755
21	Marion Forks	2730
22	Santiam Junction	3990
23	Detroit	1500
24	Whitewater Bridge	2175



AREA ELEVATION CURVE — DETROIT DAM
DRAINAGE AREA 438 SQUARE MILES

SNOW INVESTIGATIONS SUMMARY REPORT
SNOW HYDROLOGY

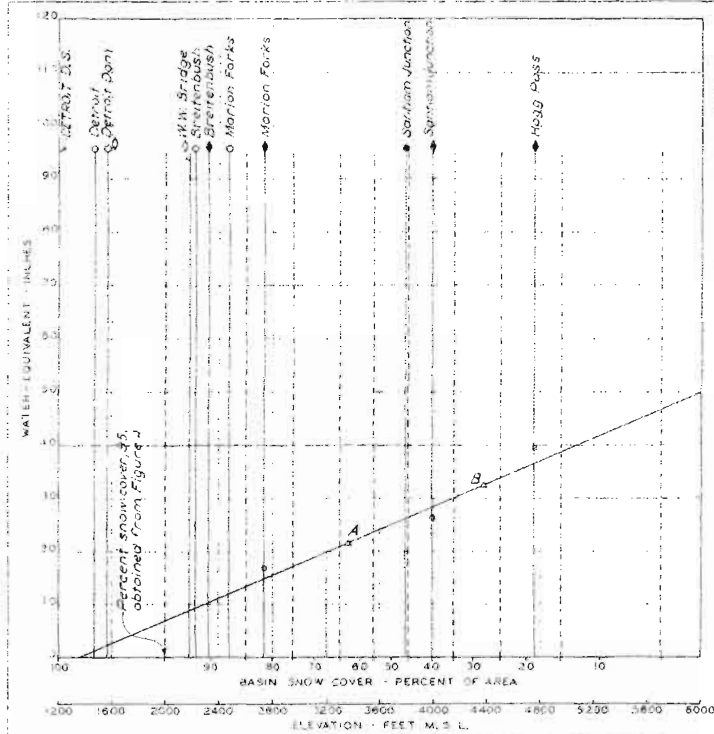
BASIN MAP

NORTH SANTIAM RIVER ABOVE DETROIT DAM
DRAINAGE AREA 438 SQ. MI.

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

TO ACCOMPANY REPORT DATED 30 JUNE 1952

PD-20-25/67



LEGEND

- o Precipitation, manual
- ● Precipitation, recorder
- ◊ Snow Course, supplementary
- ♦ Snow Course, regular

SNOW WATER-EQUIVALENT

North Fork, Santiam River Above Detroit

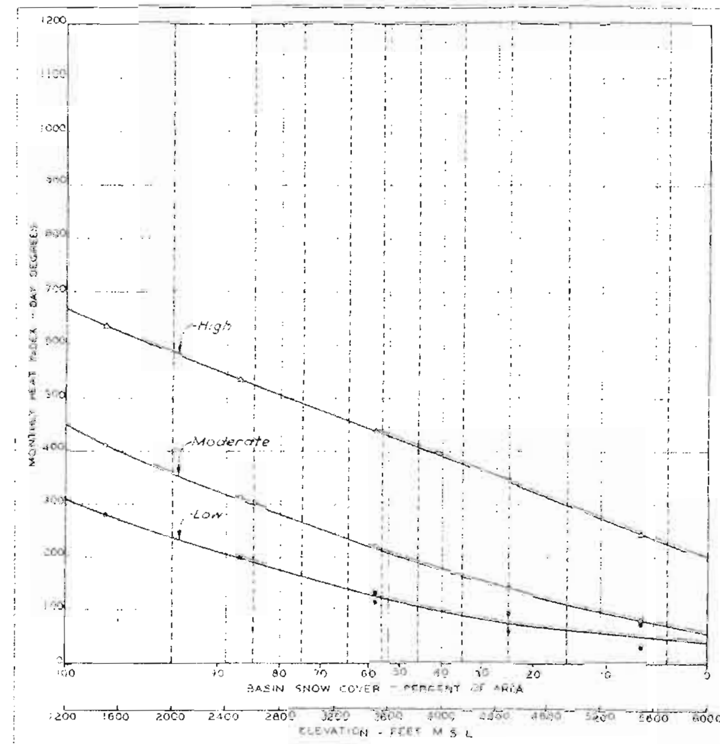
ZONE	INCHES
90 - 100	6.8
80 - 90	13.2
70 - 80	17.0
60 - 70	20.8
50 - 60	23.5
40 - 50	26.0
30 - 40	30.0
20 - 30	33.7
10 - 20	38.6
0 - 10	46.7
Wtd. Total	25.6
% Cover	95
Date	1 Feb. 1954

Corrected water-equivalent = 25.6 (0.75) = 19.2"

Notes:

1. Plotted points show depth of snow at named snow courses at given elevations.
2. Points A and B show mean snow water-equivalent depth and mean elevation at Marion Forks and Santiam Junction and Santiam Junction and Hogg Pass, respectively.
3. Line through A and B represents unadjusted depth of snow water-equivalent at all elevations.
4. Dashed lines indicate midpoint of each zone of 10 percent of basin area.
5. Intersections of dashed line and line through A and B show unadjusted mean depth of water-equivalent for each zone.
6. Summation of water-equivalent for each zone divided by 10 gives unadjusted basin snow water-equivalent.
7. Corrected basin snow water-equivalent is obtained by multiplying unadjusted basin snow water-equivalent by 0.75, a factor derived from water-balance studies.

FIGURE 1 - SNOW WATER-EQUIVALENT



Notes:

1. Chart is sample determination of day degrees over basin for high, moderate, and low temperatures during February.
2. Plotted points represent monthly total of day degrees at each elevation for given number of day degrees at Detroit, based on lapse rate determined from Medford radiosonde data, 3.5°F/1000 feet.
3. Dashed lines represent mean of each 10 percent zone of basin.
4. Intersections of dashed lines with curves represent day degrees for corresponding zones.
5. Melt in a given zone may be obtained by multiplying day degrees by rate of melt in inches per day degree.

FIGURE 2 - FEBRUARY HEAT SUPPLY

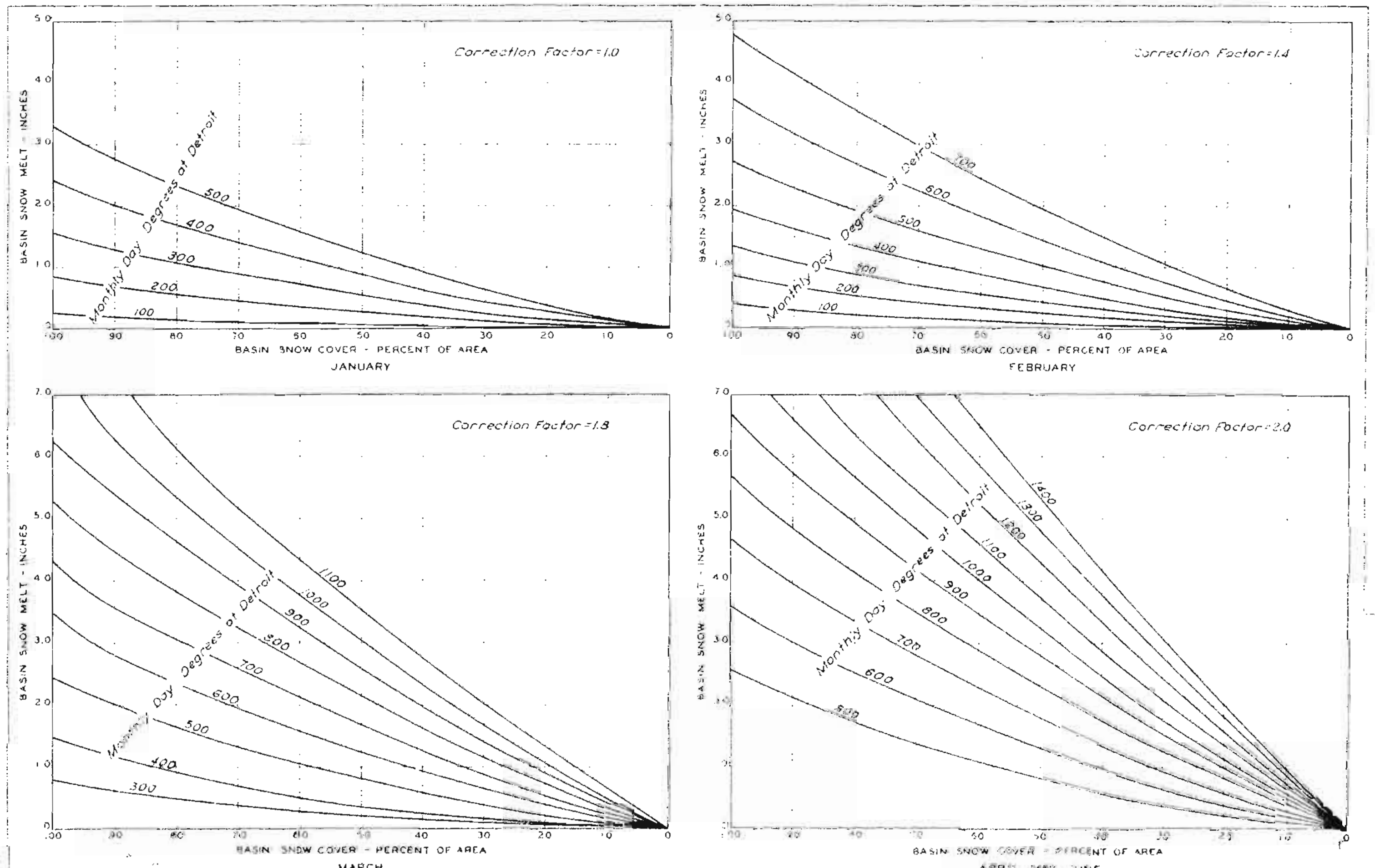
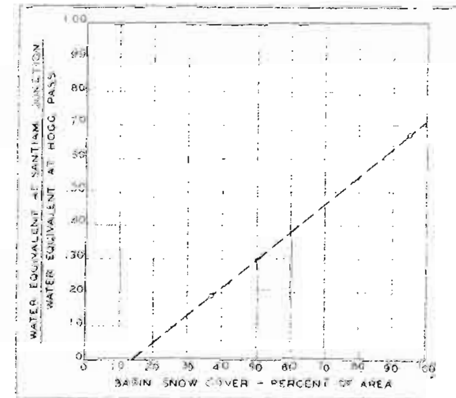


FIGURE 3 - BASIN SNOW MELT



Notes:

1. Plotted points show percent cover determined by aerial reconnaissance and ratio of water-equivalents of two key snow courses, Santiam Junction and Hogg Pass on same date.
2. Line connecting plotted points gives estimated percent snow cover for given ratio of depths of water-equivalent at the key snow courses.
3. Chart values not considered conclusive because of insufficient data and are shown principally for illustrating method of estimating percent snow cover.

FIGURE 4 - BASIN SNOW COVER

Notes:

1. Curves represent monthly totals of day degrees at Detroit.
2. Ordinate at intersection of given percent snow cover and line representing given day degrees at Detroit gives melt in inches over basin, assuming 0.01 inch per day degree melt rate.
3. Corrected basin melt is obtained by multiplying melt from chart by factor based on melt studies in basin and shown on each chart.

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SNOWPACK WATER EQUIVALENT AND MELT

NORTH SANTIAM RIVER ABOVE DETROIT DAM
DRAINAGE AREA 438 SQ MI

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

PD-20-25/68

PLATE 11-4

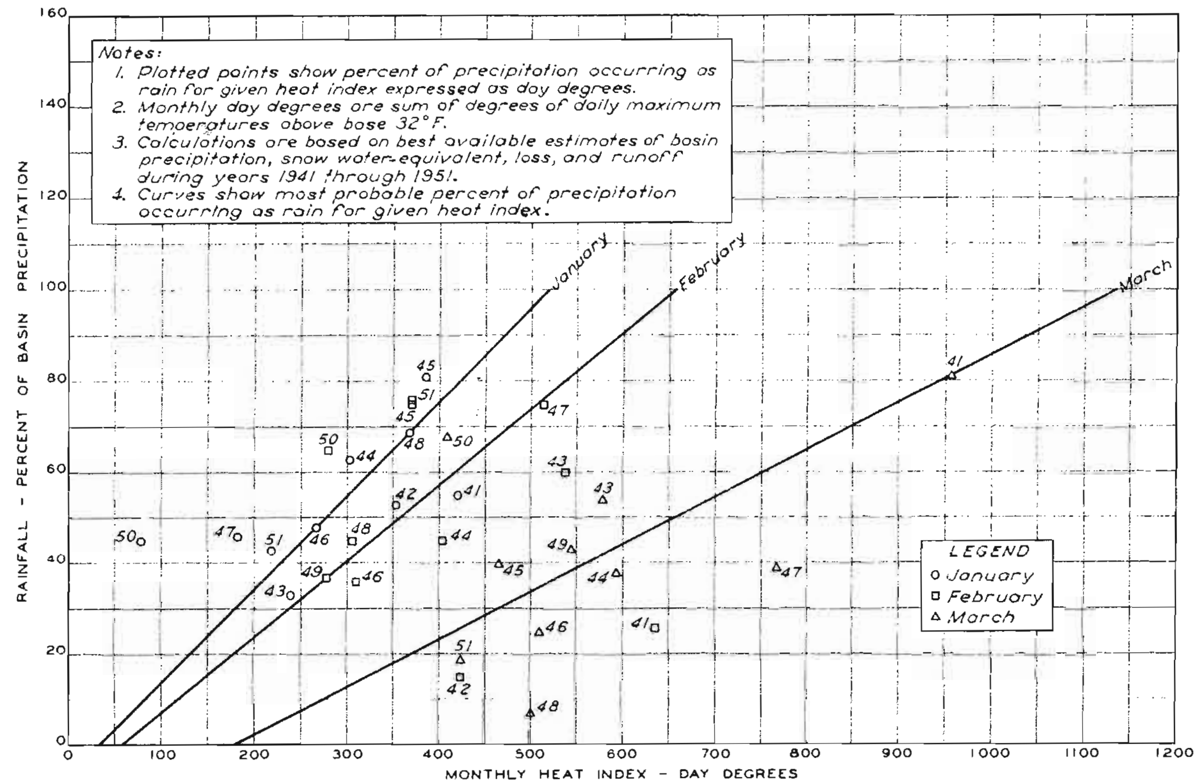


FIGURE 1 - PERCENT OF PRECIPITATION OCCURRING AS RAIN

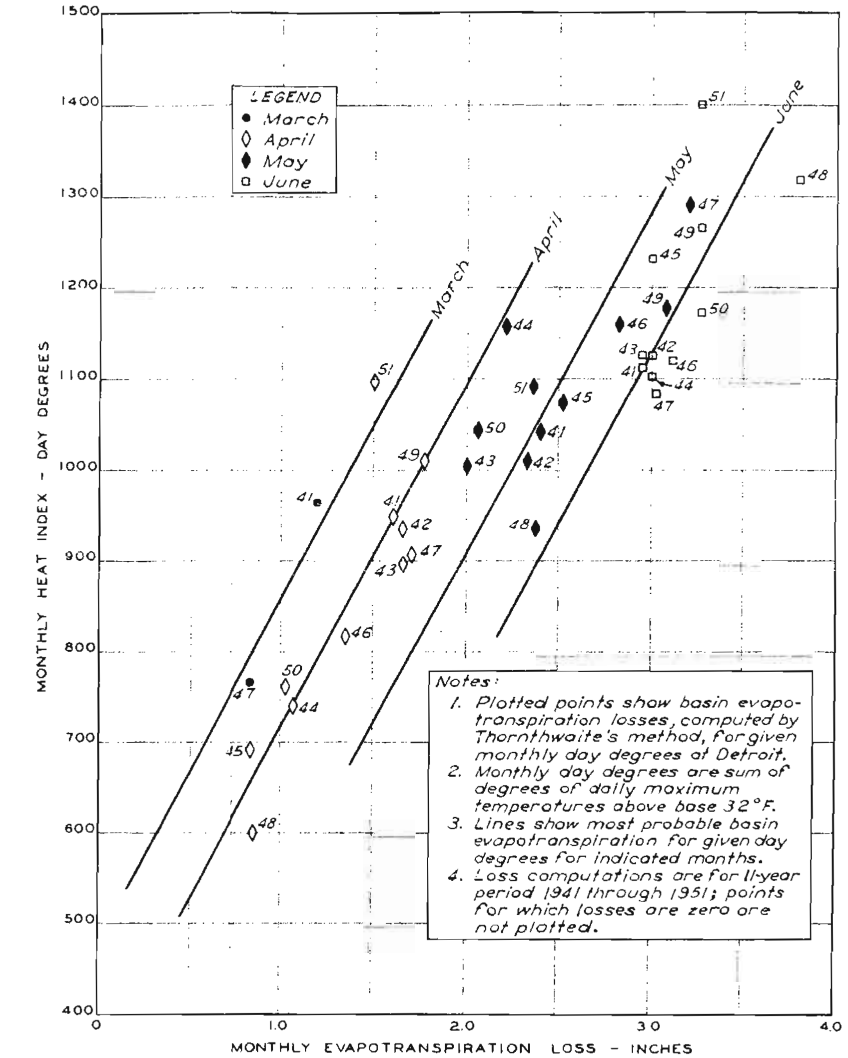
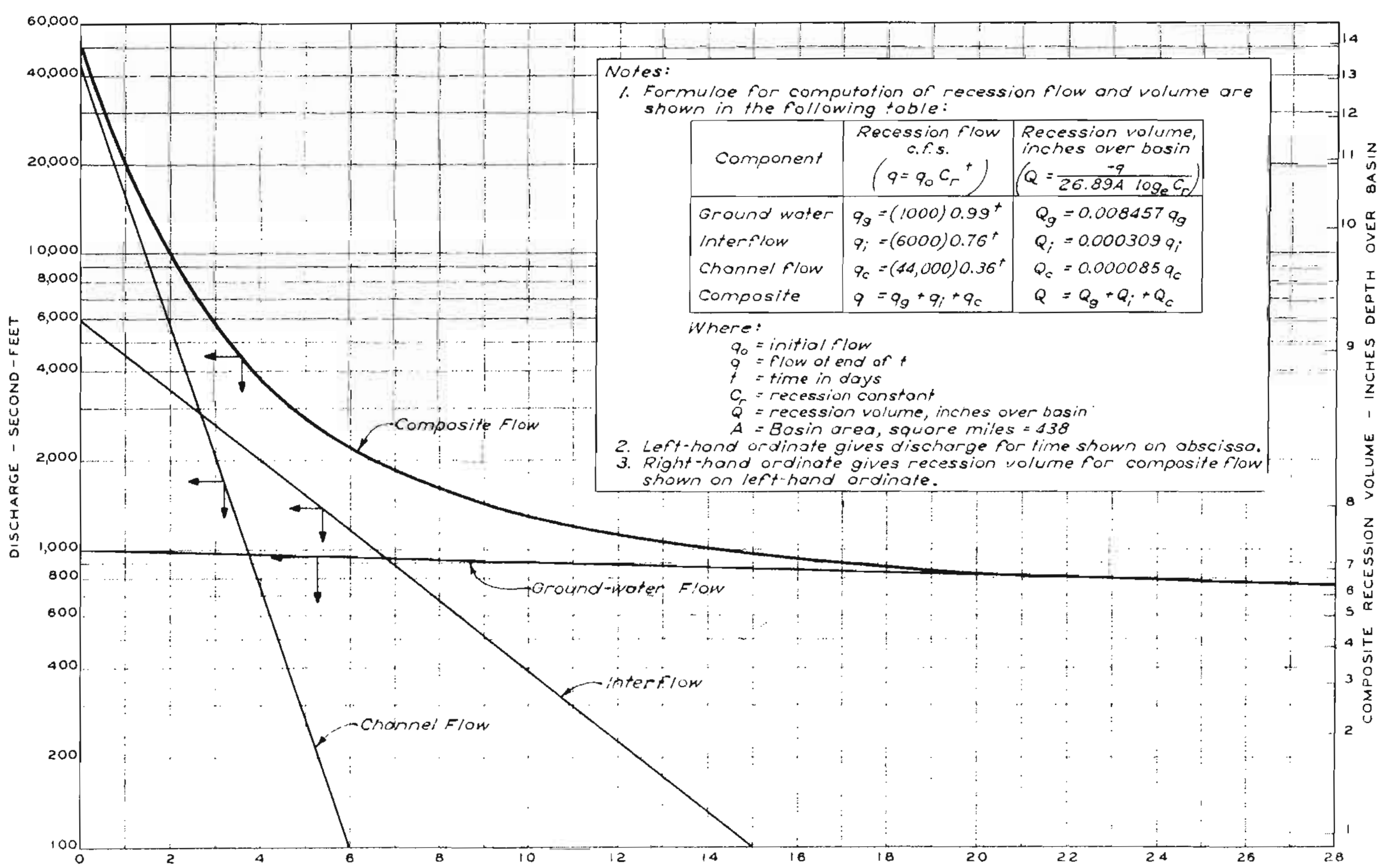


FIGURE 2 - LOSS BY EVAPOTRANSPIRATION



Notes:
1. Formulae for computation of recession flow and volume are shown in the following table:

Component	Recession flow c.f.s. ($q = q_o C_r^{\,t}$)	Recession volume, inches over basin ($Q = \frac{-q}{26.89A \log_e C_r}$)
Ground water	$q_g = (1000)\,0.99^t$	$Q_g = 0.008457\,q_g$
Interflow	$q_i = (6000)\,0.76^t$	$Q_i = 0.000309\,q_i$
Channel flow	$q_c = (44{,}000)\,0.36^t$	$Q_c = 0.000085\,q_c$
Composite	$q = q_g + q_i + q_c$	$Q = Q_g + Q_i + Q_c$

Where:
q_o = initial flow
q = flow at end of t
t = time in days
C_r = recession constant
Q = recession volume, inches over basin
A = Basin area, square miles = 438

2. Left-hand ordinate gives discharge for time shown on abscissa.
3. Right-hand ordinate gives recession volume for composite flow shown on left-hand ordinate.

FIGURE 3 - STREAMFLOW RECESSION CHARACTERISTICS

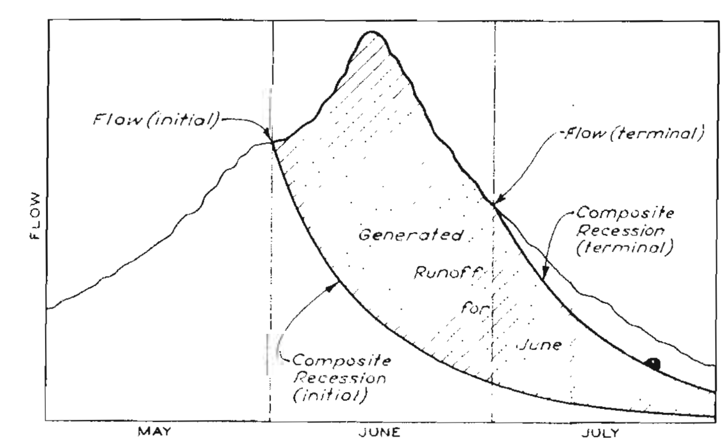


FIGURE 4 - DIAGRAM ILLUSTRATING GENERATED RUNOFF

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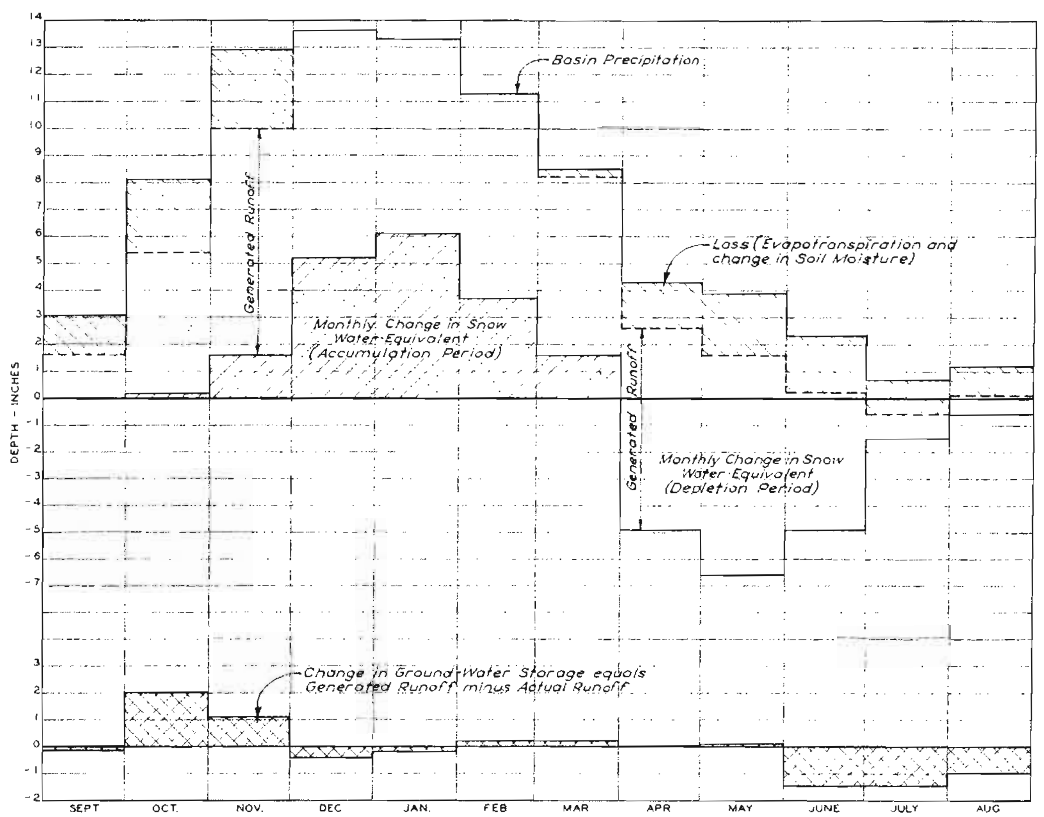


FIGURE 1 - MONTHLY COMPONENTS OF WATER BALANCE

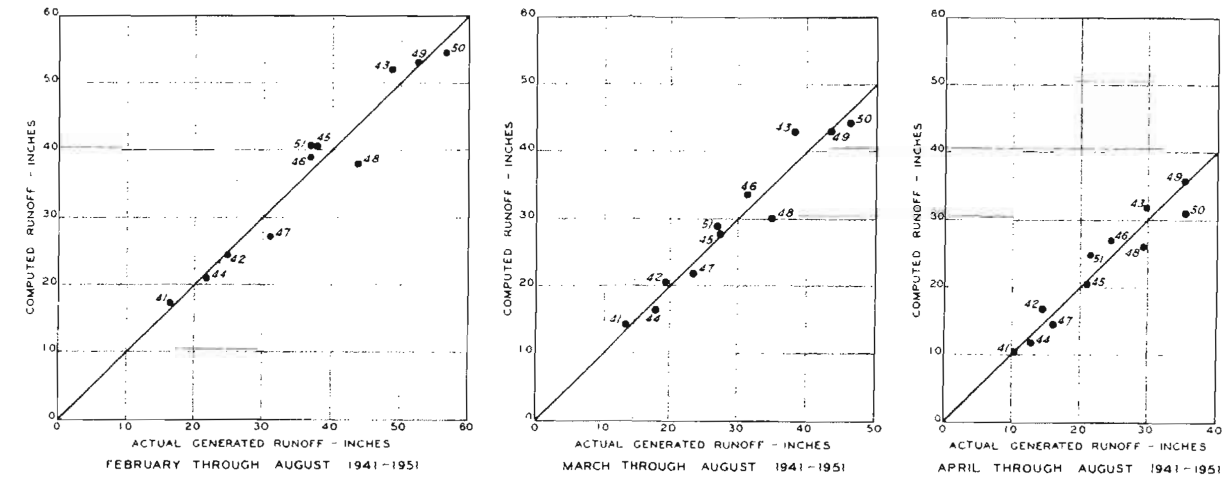
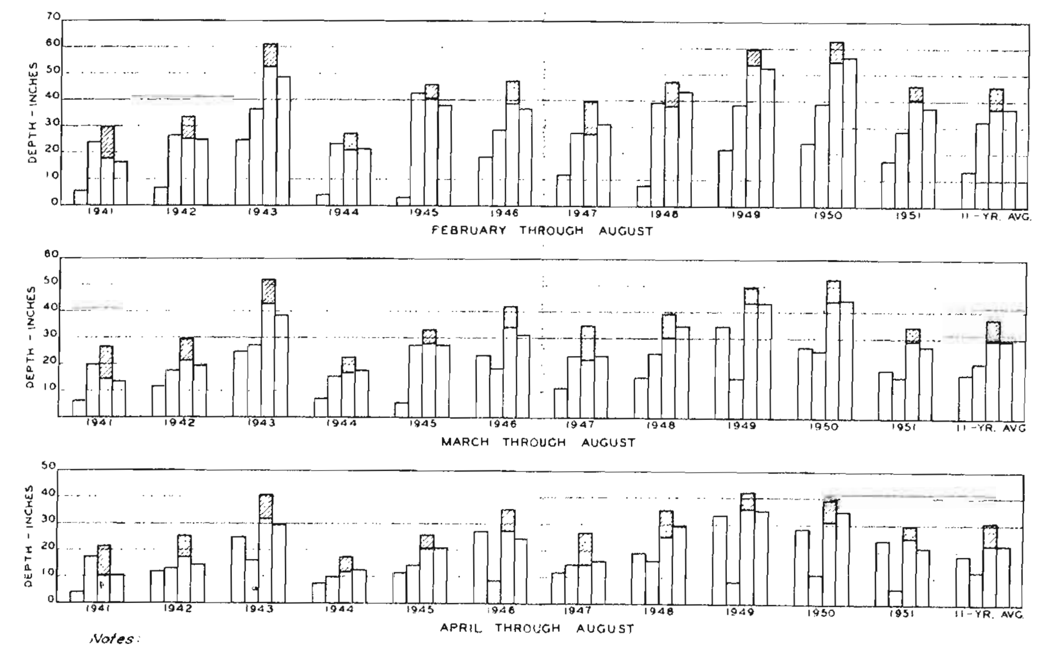


FIGURE 2 - COMPUTED VERSUS ACTUAL GENERATED RUNOFF



Notes:

1. *Bar diagrams show water balance for indicated seasons and years.*
2. *Bar at left shows initial depth of basin snow water-equivalent. Second bar shows basin precipitation. Total length of third bar gives sum of precipitation and water-equivalent; hatched portion represents amount deductible for loss, remainder representing expected runoff. Fourth bar shows actual generated runoff.*

FIGURE 4 - WATER BALANCE FOR FORECAST PERIODS

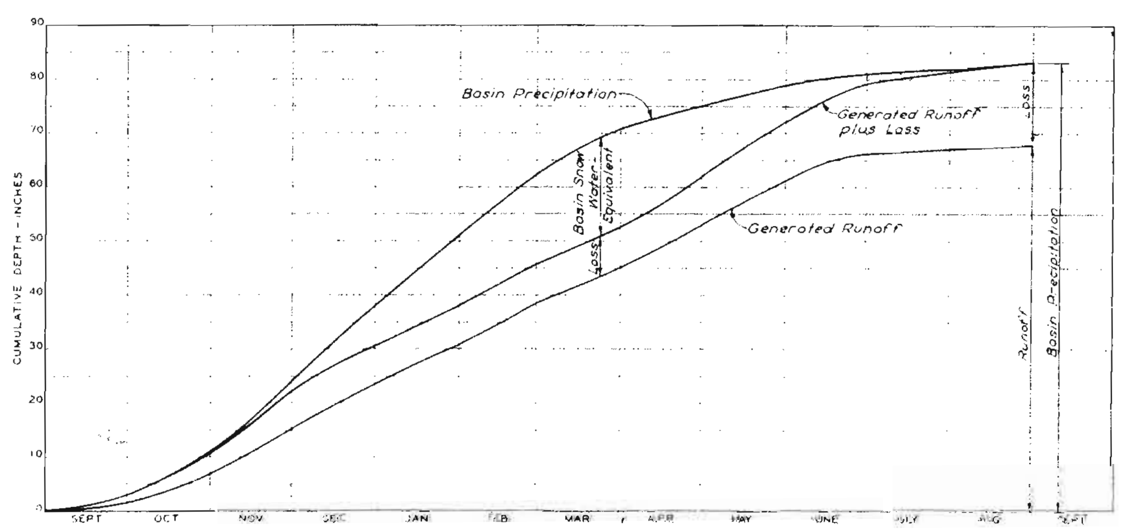


FIGURE 3 - CUMULATIVE COMPONENTS OF WATER BALANCE

Note:

Figures 1 and 3 represent monthly components of water balance for North Santiam River above Detroit Dam, based on mean values for period 1941 through 1951.

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CHAPTER 12 - RESERVOIR REGULATION

12-01. INTRODUCTION

12-01.01 General. - In the operation of reservoirs and other engineering works for the regulation of streamflow, two general types of problems are involved. One is concerned with short-term forecasts of inflow for the day-to-day planning of reservoir releases for power generation, flood control, etc. The other problem is concerned with seasonal runoff volume and is encountered in the determination of seasonal regulation schedules and flood control storage allocation and reservation diagrams. It is in the solution of this latter problem that the application of snow hydrology is especially useful and permits flexible and efficient use of multiple-purpose storage, by making flood control and conservation use of storage compatible on a seasonal basis. Ordinarily, in the case of rainfall, it is possible to estimate runoff volumes, with any degree of certainty, only a few days in advance of their occurrence. This follows from the fact that rainfall volumes can be accurately estimated only after the rainfall has actually fallen and been gaged. Then, only the natural lag time of the drainage basin remains before the resulting runoff is realized. This period may be extended somewhat by the use of 24- and 48-hour quantitative precipitation forecasts; however, the accuracy of such forecasts does not warrant their use without qualification. On the other hand, in the mountainous areas of the western United States (and elsewhere for areas having similar climatological conditions), it is possible to estimate accurately the volume of snowmelt runoff months in advance of its actual occurrence. Since the snowpack is, for the most part, deposited well in advance of its subsequent ablation by melting, it is, in effect, an immense natural reservoir. Its water content can be gaged quite accurately (either directly or indirectly) by any of the several methods outlined in the previous chapter. In this chapter, the manner in which runoff volume forecasts are utilized in the operation of reservoirs will be presented. Also, methods used in the day-to-day operation of reservoirs, based on short-term forecasts (see chap. 9), will be considered briefly.

12-01.02 Multiple-purpose reservoirs. - The climatic regime of the western mountain areas of the United States is such that the same reservoir storage space can be used for the usually incompatible requirements of flood control and conservation. The varied condition of rainfall and snowfall in this region are shown in plate 3-1, which gives the relationship between form of precipitation and elevation. Drainage basins whose runoff-producing areas are predominantly above 5000 to 6000 feet in elevation receive precipitation almost entirely in the form of snow. In these areas, winter rain runoff is usually negligible, with most of the annual runoff volume occurring during the spring and early summer as a result of melting of the accumulated snow. For basins lying within low and intermediate elevation ranges (below 5,000 feet), precipitation falls predominantly in the form of rain, and winter runoff from rainfall constitutes the primary source of streamflow. In the higher portions of these basins, however, a portion of the precipitation is accumulated in the form of snow, so that there is an appreciable contribution to runoff resulting from snowmelt during the spring. For both cases, reservoir storage regulation schedules may take advantage of the known storage of water in the snowpack for beneficial use on a seasonal basis. Reservoirs used in this manner are thus multiple-purpose in a true sense, unlike reservoirs where different portions of the total storage space are allocated for power generation, flood control, irrigation, etc., on a fixed and inflexible basis.

12-01.03 For reservoirs on streams whose drainage areas are low to intermediate in elevation (as in the case of tributaries along the coastal regions in western United States), the marked seasonal variation in precipitation allows the winter-rain-flood season to be rather definitely defined; generally speaking, by the time the spring snowmelt season begins, the threat of rain floods has passed for the year. The same reservoir space that was evacuated for control of winter rain floods may be filled from the volume of spring snowmelt, augmented by occasional runoff from spring rainfall, and thereby result in a full reservoir with non-damaging streamflow releases in downstream channels. The stored water may then be released to augment streamflow in the dry summer months and for power production during the fall in anticipation of the ensuing winter flood season. The spring filling of these reservoirs may be accomplished in accordance with a fixed seasonal regulation schedule as shown in plate 12-1, which was extracted from the reservoir regulation manual for Detroit and Big Cliff Reservoirs. 2/ Optimum use of the available storage for conservation as well as flood-control storage, however, requires that the possible variation in volume of snowmelt runoff also be considered in the filling schedule.

12-01.04 For reservoirs controlling flows from relatively high elevation areas, drawdown of the reservoir level is accomplished in accordance with the requirements for use of the stored water, either in the summer or through the fall and winter seasons. In the winter (usually beginning on the first of January), schedules may be prepared for providing flood control storage space on the basis of conditions known at that time, and revisions in the schedule may be made as the runoff potential develops through the winter and early spring. A storage allocation diagram giving the flood control storage required for different seasonal runoff volumes is shown as figure 1, plate 12-2. The use of such a diagram with its seasonal runoff parameter results, at all times, in the maximum possible flood control storage reservation compatible with the filling of the reservoir. Details of the construction of this diagram are given later in this chapter.

12-01.05 Peak flow forecasts. - Reservoir regulation for flood control requires predictions of peak flow as well as of volume. In the case of spring snowmelt floods, the peak rate of flow is, to a great extent, dependent upon variations in the rate of melt and hence upon the melt-producing meteorological conditions. Nevertheless, there exists a certain correlation between seasonal volume of runoff from snowmelt and the peak rate of flow. This is illustrated in figure 2, plate 12-2, which shows the relationship between peak flow and seasonal volume of runoff for the Columbia River near The Dalles, Oregon. The use of such a relationship in estimating peak flows requires, of course, a method of estimating volume of runoff.

12-01.06 Incidental relationships. - Figure 3 of plate 12-2 presents some frequency distributions of seasonal runoff volume, peak discharge, and date of peak discharge for the Columbia River near The Dalles, Oregon. These data are of incidental value in reservoir regulation. In figure 4 of plate 12-2, a flood control storage reservation curve for the Columbia River near The Dalles is shown which gives the amounts of storage required to control to specified discharges the various seasonal runoff volumes, or, conversely, the controlled discharges that would result from various seasonal runoff volumes and available amounts of storage.

12-02. DAY-TO-DAY REGULATION

12-02.01 The day-to-day regulation of reservoirs in accordance with short-term forecasts of reservoir inflow is, for the most part, connected with the regulation of flood flows and the generation of power. The regulation of reservoirs for such other conservation uses as irrigation, navigation, recreation, pollution abatement, and domestic water supplies, is usually planned on a longer-term or seasonal basis, and changes in outflows are required infrequently. While seasonal operation schedules are used for the long-term planning of power releases and flood control reservations, as was previously mentioned, the fact that the rates of reservoir inflow and regulated outflows cannot be foretold much in advance necessitates that the operation also be based on short-term forecasts of inflow. Short-term forecasts of reservoir inflow from either rain-on-snow events or from snowmelt alone can be made as described in chapter 9. For the generation of power, such inflow forecasts, combined with the power requirements for the project in conjunction with the system as a whole, determines the schedule of releases. For flood control operation, such other factors as inflow from uncontrolled downstream areas and available storage capacity in the reservoir also influence the releases. Because of the complex relationships involved, flood control regulation schedules are drawn up on a basis of historical data to best accomplish the desired flood control regulation.

12-03. SEASONAL REGULATION

12-03.01 Storage allocation for flood control. - In the multiple use of reservoir space for the contradictory requirements of flood control and conservation of spring snowmelt floods, storage allocation diagrams are customarily derived from historical data, as previously mentioned. Such a diagram for Hungry Horse reservoir on the South Fork of the Flathead River, Montana, 1/ is given as figure 1 of plate 12-2. Such diagrams are determined by computing the storage required, both before and during the melt season, to control to a given outflow, the maximum and other critical historical flood events. Parameters of the remaining runoff from any given date to the end of the snowmelt runoff season (usually 30 September) are drawn to envelop these historical flood data. It is customary to limit the slope of these parameter lines to the maximum permissible rate of drawdown of the reservoir (maximum permissible discharge) as governed either by outlet capacity or downstream channel capacities. Thus, in the diagram of figure 1, plate 12-2, the slope of the pre-melt-season drawdown curves is equivalent to 20,000 cfs (approximately 1.2 million acre feet per month) which is the approximate maximum outlet capacity of the reservoir (outlet valves plus allowable flow through power turbines). The enveloping curves during the flood season proper (1 May to 30 June in fig. 1, pl. 12-2) also indicate an increasing storage requirement with time for a given parameter value. This is in consequence of the increase in the potential flood flows from the same volume of runoff, that occurs as the melt season progresses. The Hungry Horse flood-storage allocation diagram is designed to provide flood control for the lower Columbia River and for the reach of the Flathead River immediately downstream from the dam and above Flathead Lake in Montana. It is based on the criteria of (1) restricting the reservoir releases to 3,000 cfs during the period beginning five days before the natural flow of the Columbia River at The Dalles, Oregon reaches 500,000 cfs and ending five days before it again decreases to 500,000 cfs (five days being the time of travel between Hungry Horse dam and The Dalles), (2) restricting the releases to control the Flathead River, as gaged at Columbia Falls, Montana, to certain non-damaging flows, the permissible flows depending partially upon the backwater effect in the river resulting from varying lake stages, and (3) maintaining a minimum release of 500 cfs at all times.

12-03.02 Safety factors. - Factors of safety, beyond what is actually required to envelop the plotted historical flood data, may be included in storage allocation diagrams. Thus, in figure 1 of plate 12-2, a factor of safety of 200,000 acre-feet was incorporated in the parameters prior to 1 May, decreasing, from that date, at a uniform rate such that it equals zero on 30 June. This factor of safety allows for errors in the forecast volume of runoff, thereby assuring adequate flood-control reservation. An additional factor of safety was incorporated in the Hungry Horse flood-storage allocation diagram for those parameters outside the range of the historical data. An analysis of the parameters of 2.0 million acre-feet and less, which are based on historical data, indicated an increase of 0.83 acre-foot in flood-control allocation for each acre-foot increase in volume of runoff. For the parameters in excess of 2.0 million acre-feet, no historical data were available; consequently, it was considered prudent to increase the incremental changes in the flood control allocation for these large floods to an amount equal to the increase in the volume of runoff. This change is apparent in the change in spacing of the parameter lines of the figure.

12-03.03 In the foregoing example, the factor-of-safety allowances were made to assure adequate flood-control allocations, at the expense of conservation storage, for situations more critical (from the flood control viewpoint) than those given by historical data or to allow for possible errors in the volume forecasts. Consequently, there is this added risk of not filling the reservoir, especially where errors in volume forecasts result in over-estimates of runoff volume. It is to be pointed out that factors of safety may also be provided from the viewpoint of conservation of water. There is also included in the storage allocation diagrams derived for Hungry Horse project, a factor of safety for refilling the reservoir at the expense of some flood control storage. By establishing a minimum release at the project of 3000 cfs for downstream flood control as measured at The Dalles, a flexibility of regulation is established. If late season forecasts indicate that original volume inflow forecasts were too high, release from the reservoir may be reduced to the minimum discharge of 500 cfs, and thereby refill storage at a faster than normal rate so as to assure the refilling of the reservoir by the season's end. A study of the Hungry Horse flood control storage allocation diagram 3/ indicates the factors of safety incorporated therein do not seriously affect the refilling of the reservoir even when possible errors in the forecast runoff volumes are considered. Moreover, forecasts which are some 200,000 acre-feet too low (approximately average error of Hungry Horse inflow forecasts) 4/, do not seriously affect the flood control operation of the reservoir. Concerning the testing of the flood control storage allocation diagram for Hungry Horse reservoir, the following excerpt from the previously cited study 3/ is quoted:

"The summary indicates that, with completely accurate forecasts, the reservoir would have refilled in every year of the 31 years studied. In 1931 and in 1942, both of which were very dry years, the reservoir would have refilled prior to the date of the last significant peak at The Dalles. The time required for the effect of spills at Hungry Horse to reach The Dalles is such that the latest significant peak at The Dalles would have been reduced by storage in Hungry Horse Reservoir in both years. If forecasts 200,000 acre-feet too low had been used, the reservoir would have refilled in every year of the 31 years, but would have refilled prior to the date of the latest significant peak in 10 of the 31 years. Of these ten years, only 1911, 1936, and 1948 were years in which the natural peak flow at The Dalles exceeded 500,000 cfs, and in each of these three years the time required for spilled flows at Hungry Horse to reach The Dalles would have been such that the latest significant peak at The Dalles would have been reduced by storage in Hungry Horse Reservoir. If forecasts 200,000 acre-feet too high had been used, the reservoir would have failed to refill in only four years of the 31 years studied and would not have refilled prior to the date of the latest significant peak at The Dalles in any year. The four years in which the reservoir would have failed to refill were 1931, 1937, 1941, and 1944, all of which were dry years, but the greatest deficiency would have been only 32,000 acre-feet in 1931 which is only slightly more than one percent of the live storage capacity of the reservoir. Therefore, such failure to refill under these assumed conditions has little significance."

12-03.04 Volume forecasts. - Forecasts of seasonal volume of runoff are, of course, necessary in the application of flood-control storage allocation diagrams (in the place of the observed historical values which were used in the derivation of the diagrams). Errors inherent in these forecasts may possibly result in the undesirable operation of a reservoir, as was discussed in the previous paragraph. Methods by which seasonal volume forecasts can be made were discussed in the preceding chapter. For situations where a definite method of seasonal-runoff forecasting is used in conjunction with the storage-allocation diagram in the operation of a reservoir, it is possible to assess, rather definitely, the effect of errors in the forecasting method upon the operation of the reservoir. The effect of errors in volume forecasts is also pertinent to the discussion in the section which follows, where volume forecasts are used to estimate peak flows.

12-04. PEAK-TO-VOLUME RELATIONSHIP

12-04.01 General. - As previously mentioned, there exists a general relationship between the peak snowmelt discharge and the seasonal snowmelt runoff volume for most basins which have appreciable winter snowpack accumulations. Since the volume of runoff from spring snowmelt can be estimated quite accurately some months in advance, it is likewise possible to make forecasts of peak flows resulting from springtime snowmelt well in advance of their actual occurrence. Intelligent application of long-range forecasts of unregulated peak discharges resulting from snowmelt requires full understanding of (1) the significance of peak-to-volume ratios, (2) the best method of applying them to specific cases, and (3) the probable accuracy of the estimates. Closely allied to the peak-to-volume determination is that of evaluating flood-control storage reservation requirements. Examination of the peak-to-volume relationship in this section is accompanied by an illustration of the relationship for the Columbia River near The Dalles, Oregon, one of the major snowmelt runoff rivers in the United States. Reference is made to the report, "Relationship between peak discharge and volume runoff of the Columbia River near The Dalles, Oregon" by the Water Management Subcommittee of the Columbia Basin Inter-Agency Committee (CBIAC), 6/ for a more complete discussion of peak-to-volume relationship for Columbia River near The Dalles.

12-04.02 Peak-to-volume diagram. - Figure 2 of plate 12-2 gives the basic relationship between peak flows and volume of snowmelt runoff for the Columbia River near The Dalles, Oregon. The peak flows given there are mean daily values and include adjustments for relatively minor flood control regulation by Grand Coulee and Hungry Horse dams during recent years. The seasonal runoff volumes used in the relationship were for the period April through September, and adjustments for storage in six major reservoirs were made. 5/ The entire 77 years of available record of flows for the Columbia River at The Dalles (1879 through 1955) were used in the determination of the relationship of figure 2. The regression line fitted to these data is as follows:

$$Y = 6.77X - 118 \qquad (12\text{-}1)$$

where Y is the peak daily flow in thousand cfs and X is the April through September runoff in million acre-feet. The standard error of estimate of the relationship amounts to 76.2 thousand cfs in contrast to the standard deviation of 172.6 thousand cfs for the peak flows. The resulting correlation coefficient is 0.90.

12-04.03 Time changes in relationship. - In recent years there has been a tendency for higher peak flows to be associated with a given volume of runoff for the Columbia River near The Dalles, Oregon. A study of the peak-to-volume relationship, analyzing the periods from 1879 through 1916 and from 1917 through 1955 separately, resulted in the following regression equations:

PERIOD	EQUATION	
1879 - 1955	$Y = 6.77X - 118$	(12-1)
1879 - 1916	$Y = 7.10X - 179$	(12-2)
1917 - 1955	$Y = 7.63X - 177$	(12-3)

where Y is the peak discharge in thousand cfs and X is the April-September volume of runoff in million acre-feet. Equation 12-1 is also repeated in the above tabulation for comparative purposes. Although this change in the relationship with time could be attributed to man-made changes in the basin, careful consideration of the nature and order-of-magnitude of such changes shows that such is not likely. The change in the relationship appears to be associated with the natural changes in climate that occurred within the period and therefore is characteristic of large-scale climatic variations.

12-04.04 Errors of estimate for prediction of peak discharge. - It is possible to combine the effect of errors in forecasts of runoff volume and of the historical peak-to-volume ratio by statistical relationships (see Wilm 7/ for a discussion of the statistical derivation of such relationships), whereby comparisons of reliability of estimates of peak discharge through use of differing periods of runoff may be determined. Tests of the relative accuracy of peak discharge forecasts, using total and residual volume forecasts, were made by the Technical Staff of the Water Management Subcommittee using data for the Columbia River near The Dalles. 6/ The results of these tests are tabulated below:

STANDARD ERROR OF ESTIMATE OF AN INDIVIDUAL MEAN PREDICTION
OF PEAK DISCHARGE
(In thousands of second feet)

For peak-volume relationships based on total and on residual runoff

Forecast Date	Period for which runoff volume forecast is made					
	April through June		April through July		April through Sept.	
	From Total Volume Forecast	From Residual Volume Forecast	From Total Volume Forecast	From Residual Volume Forecast	From Total Volume Forecast	From Residual Volume Forecast
1 April	106.3	106.3	101.8	101.8	102.8	102.8
1 May	102.8	104.1	90.6	86.6	95.4	94.4
16 May	98.9	107.0	86.5	83.6	85.7	83.3

In general it may be stated from this study that forecasts of peak discharge for Columbia River near The Dalles based on April through July runoff are most reliable, but that little difference exists in using the April through September period. The April through June period gives consistently poorer results. It is also seen that for both 1 May and 16 May forecast dates, there is a slight improvement by using residual rather than total volume forecasts; however, the differences are generally small and of little significance.

12-05. FLOOD CONTROL STORAGE RESERVATION

12-05.01 From what has been stated, it is apparent that a relationship exists between the seasonal runoff volume and the amount of storage which would be required to control the peak discharge near The Dalles, Oregon to some given regulated outflow. Diagrams giving this relationship may be determined from an analysis of historical data wherein the volume of runoff in excess of the desired regulated discharge rate is plotted as a function of the seasonal runoff volume and lines drawn to envelop these data. Several parameters of regulated outflow may thus be determined to give the flood control storage reservation associated with various regulated discharges. Such a diagram for the Columbia River near The Dalles, Oregon is included as figure 4 of plate 12-2. Also shown on this diagram is a line representing the volume of the record 1894 spring snowmelt flood.

12-05.02 The entire 77-year record for the Columbia River near The Dalles was used in the derivation of the diagram; however, only a few of the years were critical in determining the parameters. For example, those years whose peak discharges were less than the parameters obviously could not enter in their determination. The parameters were not drawn to envelop the 1948 flood data. With the exception of this year, all other pertinent data, including the 1894 flood, gave a consistent relationship which defined the parameters quite well. The data for 1948 were, however, so far out of line that to envelop them would result in grossly inefficient use of flood control storage space in all other years. It is necessary, therefore, that in utilizing this set of curves, provision must be made for the occurrence of exceptionally high peak-to-volume ratios, such as occurred in 1948. With the repetition of such an occurrence, it would be necessary to adjust upward the regulated discharge in the lower Columbia River during the progress of the flood. It is pointed out that the curves shown in figure 4 of plate 12-2 are provisional in nature and are presented as a guide for over-all flood control regulation of the Columbia River. The diagram assumes flood control storage which is 100 percent effective in controlling discharges in the lower Columbia River. Much of the present and planned flood control storage in the basin is so located that its effectiveness is considerably less than 100 percent, and appropriate factors must be applied to determine the amount at each project which is effective for downstream flood control.

12-06. SUMMARY

12-06.01 One of the most useful applications of snow hydrology is to be found in the reservoir regulation of snowmelt runoff. For areas where much of the winter precipitation is stored in deep snowpacks, there is an interval of several months between the time the precipitation falls and the time it melts and contributes to runoff. Since this portion of the total runoff can be gaged well in advance of its realization as streamflow, allowances can be made in the operation of reservoirs in anticipation of this runoff volume. For flood control operation, the reservoir can be drawn down in advance to allow for the estimated volume of inflow. At the same time, this forecast of future inflow volume assures that the reservoir storage space evacuated for flood control can be refilled for conservation uses from the spring snowmelt flood. Reservoirs operated in such a manner are multiple-purpose reservoirs in the true sense of the term.

12-06.02 For basins having deep winter snowpack accumulations, there exists a relationship between the peak discharge and the spring snowmelt flood volume. This peak-to-volume relationship is useful in advance flood-control planning. Like the volume forecast, estimates of peak flow can be made many months in advance of their realization.

12-06.03 Diagrams which serve as guides in the operation of reservoirs are prepared from historical streamflow data. Examples of such diagrams are: (1) seasonal regulation schedules, (2) flood-control storage allocation diagrams, and (3) flood-control storage reservation diagrams. The first of these is, basically, a curve showing the maximum allowable reservoir content as a function of the time of year (see plate 12-1). During the winter rain flood season, the reservoir is held in an evacuated condition, insofar as is possible, to provide storage space for the control of rain-on-snow floods. It is filled during the spring, as the danger of rain floods diminishes, by utilizing snowmelt runoff augmented by spring rains, thereby conserving water for use during the summer and fall months. It is drawn down in the fall to again provide flood-control storage space. Filling and drawdown rates are in accordance with channel capacities and available water. The second diagram, which makes use of forecasts of spring snowmelt runoff volume, indicates, as a function of time of year, the reservoir storage space that must be allocated to flood control for different parameters of seasonal runoff volume (fig. 1, plate 12-2). Rate of drawdown is controlled by existing downstream channel and outlet capacities. The required storage allocations are also governed by given permissible releases during flood-control operation. The third diagram, unlike the first two, does not include the time of year as a factor. It shows the amount of storage, as a function of flood volume, required to control snowmelt floods to various parameters of regulated outflow (see fig. 4, plate 12-2). Nothing is said of when or where the storage reservation must be available. With an existing flood control reservation, the diagram gives the regulated outflow which may be attained for various floods or the flood volume that can be controlled to a given outflow.

12-07. REFERENCES

1/ BUREAU OF RECLAMATION - CORPS OF ENGINEERS, "Reservoir regulation manual, Hungry Horse Dam, South Fork of Flathead River, Montana (U.S.B.R. Project)" Corps of Engineers, U. S. Army, Seattle District, December 1952.

2/ CORPS OF ENGINEERS, Portland District, "Reservoir regulation manual for Detroit and Big Cliff reservoirs, North Santiam River," September 1953.

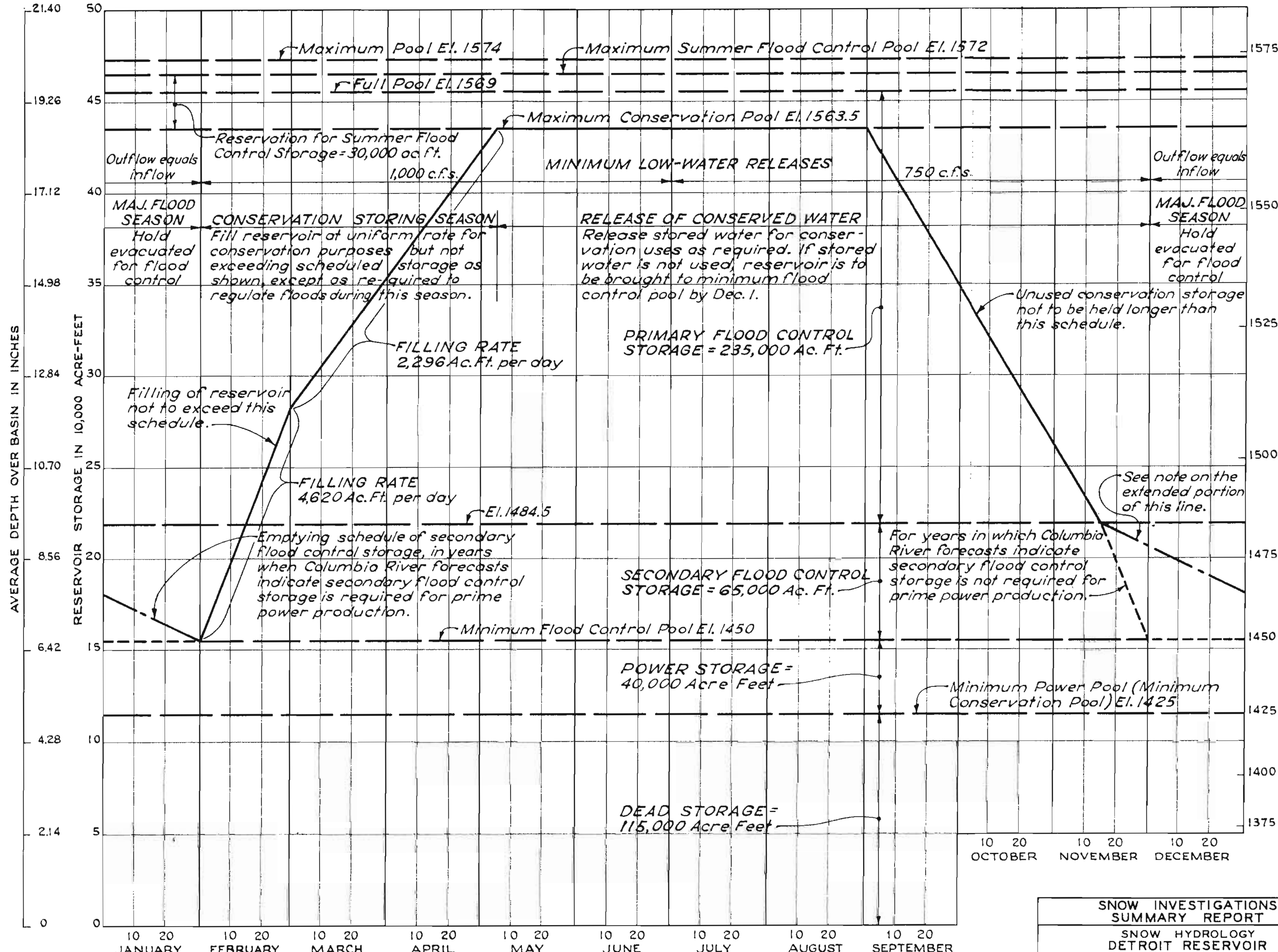
3/ DONLEY, David E., "Operation of Hungry Horse Reservoir for flood control," Bureau of Reclamation, Hydrologic Studies Office, Boise, Idaho (Dittoed report), 29 December 1951.

4/ WATER MANAGEMENT SUBCOMMITTEE, CBIAC, "Review of procedures for forecasting inflow to Hungry Horse Reservoir, Montana," (Mimeo. report). June 1953.

5/ WATER MANAGEMENT SUBCOMMITTEE, CBIAC, "Recommended reservoir storage adjustments to seasonal runoff volume forecasts in the Columbia River basin," (Mimeo. report). February 1954.

6/ WATER MANAGEMENT SUBCOMMITTEE, CBIAC, "Relationship between peak discharge and volume runoff of the Columbia River near The Dalles, Oregon," (Mimeo. report). June 1955.

7/ WILM, H. G., "Statistical control in hydrologic forecasting," Research Note No. 61, Pacific Northwest Forest and Range Experiment Station, Forest Service, Portland, Oregon, January 1950.



This diagram taken from Plate 20, Reservoir Regulation Manual for Detroit and Big Cliff Reservoirs, North Santiam River, prepared by Portland District, Corps of Engineers, 1 Sep 1953.

SNOW INVESTIGATIONS
SUMMARY REPORT

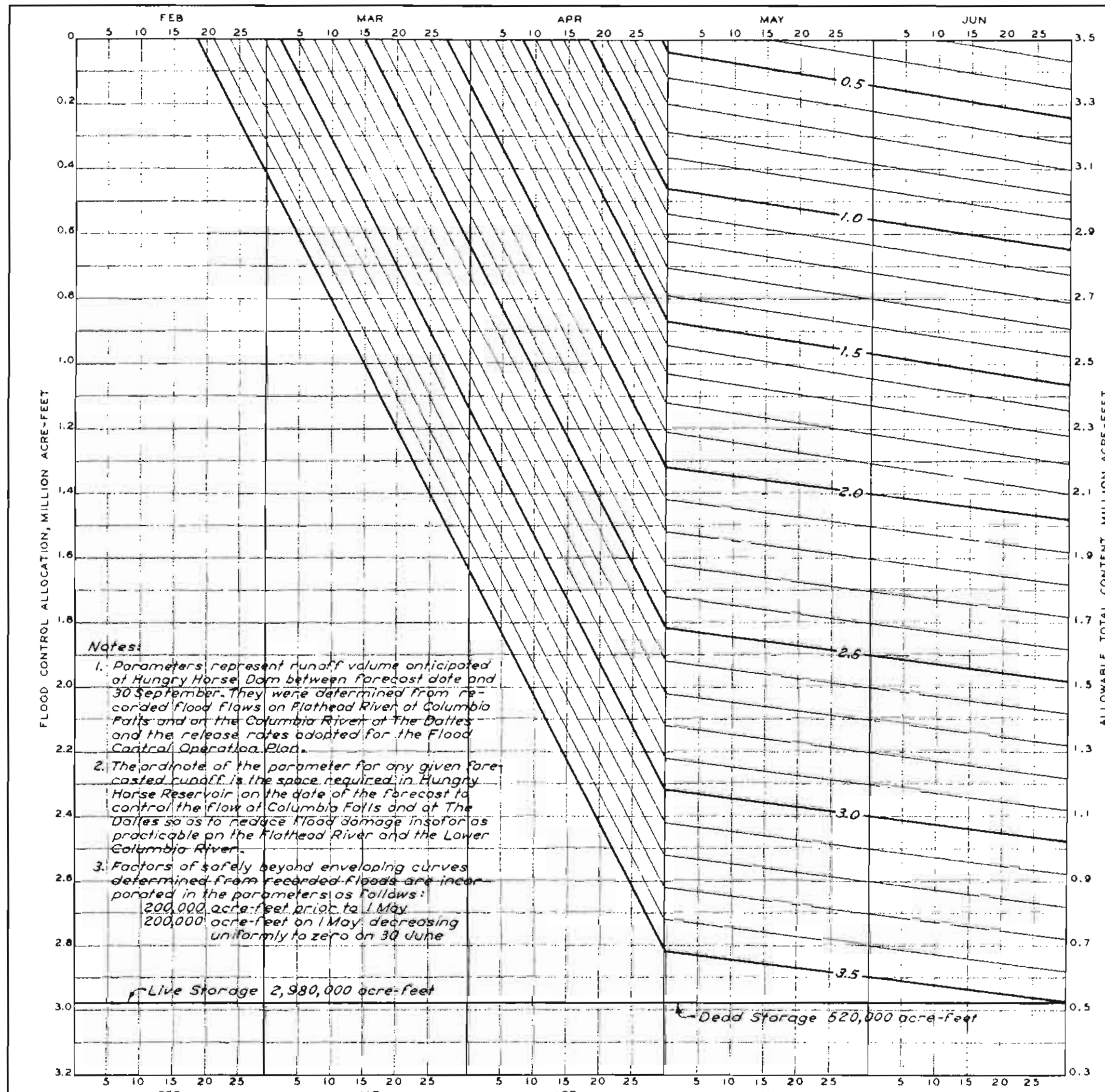
SNOW HYDROLOGY
DETROIT RESERVOIR
NORTH SANTIAM RIVER, OREGON

SEASONAL REGULATION SCHEDULE

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U.S. ARMY

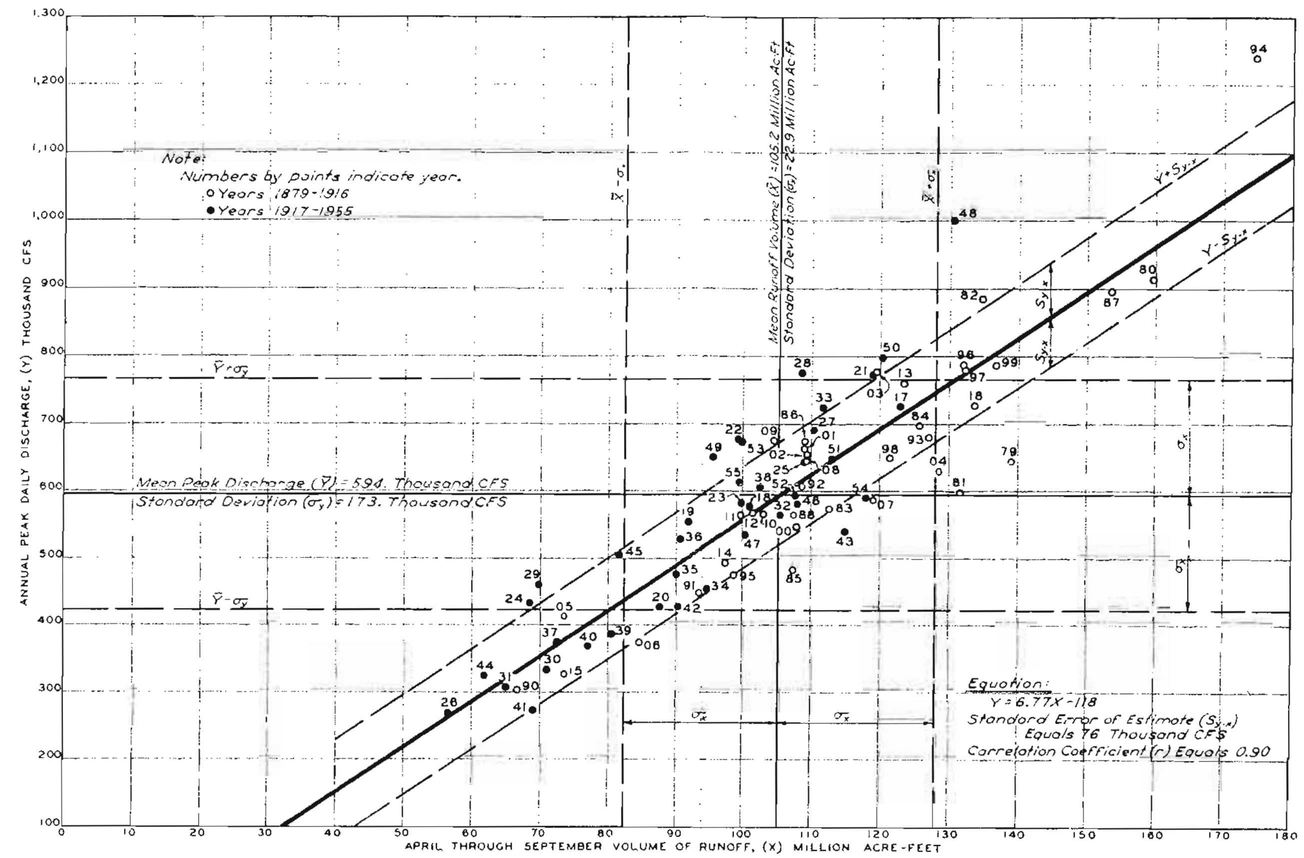
PREP: WAM/SSS SUBM: C.E.M. TO ACCOMPANY REPORT DATED 30 JUNE 1956
DRAWN: L.S.M. APPR: D.M.S.

PD-20-25/71



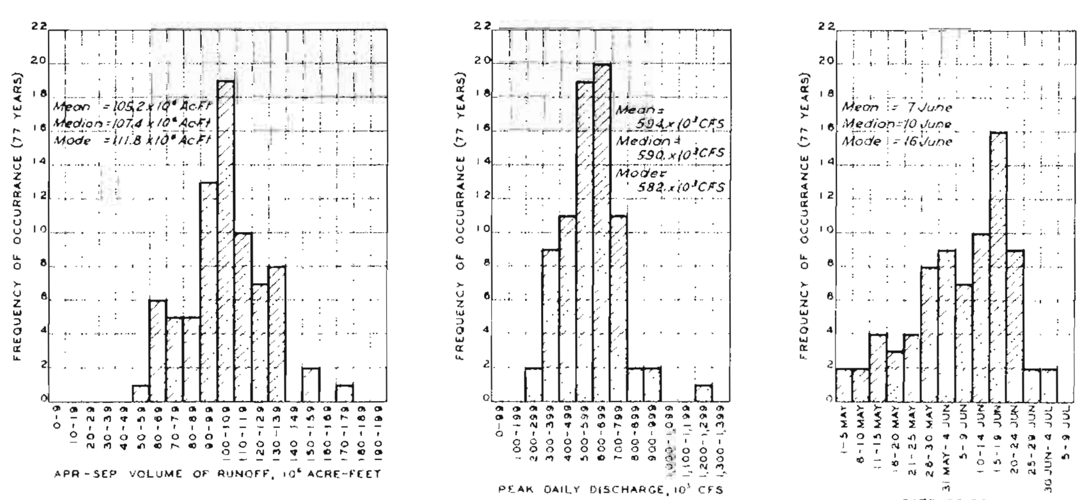
FLOOD CONTROL STORAGE ALLOCATION, HUNGRY HORSE RESERVOIR, MONTANA

FIGURE 1



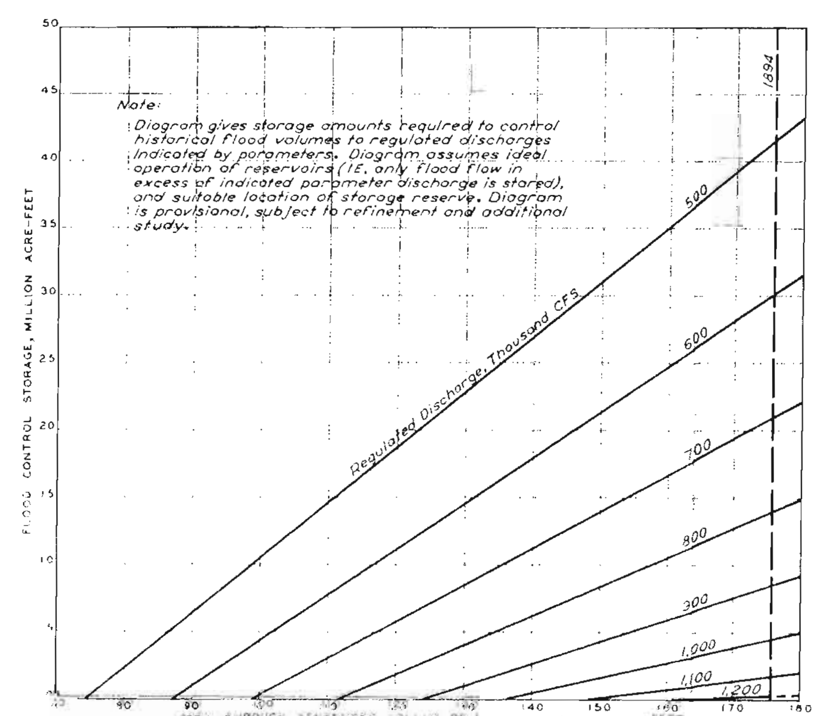
RELATIONSHIP BETWEEN PEAK DISCHARGE AND VOLUME OF RUNOFF FOR THE COLUMBIA RIVER NEAR THE DALLES, OREGON

FIGURE 2



FREQUENCY DISTRIBUTIONS FOR THE COLUMBIA RIVER NR THE DALLES, OREGON

FIGURE 3



FLOOD CONTROL STORAGE RESERVATION DIAGRAM, COLUMBIA RIVER NEAR THE DALLES, OREGON

FIGURE 4

Figure 1 was prepared by Hydrologic Studies Office, Bureau of Reclamation, Boise, Idaho. It is taken from the Reservoir Regulation Manual for Hungry Horse Dam, issued by Seattle District, Corps of Engineers, dated December 1952.

SNOW INVESTIGATIONS
SUMMARY REPORT

SNOW HYDROLOGY

RESERVOIR REGULATION DIAGRAMS

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U. S. ARMY

TO ACCOMPANY REPORT DATED 30 JUNE 1956

PD-20-25/72

PLATE 12-2

APPENDIX I

PUBLICATIONS OF THE COOPERATIVE SNOW INVESTIGATIONS

TECHNICAL REPORTS
Cooperative Snow Investigations

1. Instructions for the determination of snow quality. December 1944.

2. Bibliography of snow and ice (preliminary). June 1945.

3. Heat transmission constants of snow. (Draft) 9 December 1946.

4A. Terrain characteristics, Central Sierra Snow Laboratory Basin. June 1951.

5. Hydrometeorological Log of the Central Sierra Snow Laboratory 1945-1946. September 1947.

6-1. Classified outline of analytical program, Processing and Analysis Unit. 10 March 1947.

6-2. Progress Report of the Processing and Analysis Unit to 31 March 1948. August 1948.

6-3. Progress Report of the Processing and Analysis Unit to 31 March 1949. June 1949.

6-4. Progress Report, 1945-1950. March 1950 (Revised as of 1 June 1950).

6-5. Annual Progress Report, 1950-1951. November 1951.

6-6. Annual Progress Report, 1951-1952. July 1952.

7. Hydrometeorological Log of the Upper Columbia Snow Laboratory 1945-1946. July 1948.

8-1. Penetration of Solar Radiation into the Snowpack. March 1948.

13. Annotated brief bibliography of snow hydrology. 20 January 1950.

15-1. The storage and transmission of liquid water in the snowpack as indicated by dyes. March 1948.

16-2. Empirical methods of estimating snow melt runoff from temperature. March 1948.

17. Hydrometeorological Log of the Upper Columbia Snow Laboratory 1946-1947. May 1952

TECHNICAL REPORTS - Continued

18. Hydrometeorological Log of the Central Sierra Snow Laboratory 1946-1947. May 1952.

20-1. Hydrometeorological Log of the Upper Columbia Snow Laboratory 1947-1948. August 1949.

21. Hydrometeorological Log of the Willamette Basin Snow Laboratory 1947-1948, 1948-1949. August 1951.

22. Hydrometeorological Log of the Central Sierra Snow Laboratory 1947-1948. February 1952.

23. Hydrometeorological Log of the Central Sierra Snow Laboratory 1948-1949. November 1951.

24. Hydrometeorological Log of the Upper Columbia Snow Laboratory 1948-1949. January 1952.

25. Hydrometeorological Log of the Central Sierra Snow Laboratory 1949-1950. April 1952.

26. Hydrometeorological Log of the Upper Columbia Snow Laboratory 1949-1950. March 1952.

27. Hydrometeorological Log of the Willamette Basin Snow Laboratory 1949-1951. November 1952.

28. Hydrometeorological Log of the Central Sierra Snow Laboratory 1950-1951. August 1952.

29. Hydrometeorological Log of the Upper Columbia Snow Laboratory 1950-1951. June 1952.

30. Hydrometeorological Log of the Central Sierra Snow Laboratory 1951-52. April 1953.

APPENDIX I - Continued

RESEARCH NOTES

Corps of Engineers Analytical Unit, Cooperative Snow Investigations; and Corps of Engineers, Snow Investigations

1. MILLER, D. H., Albedo of the snow surface as related to weathering factors and stage of the season, December, 1950.
2. MIXSELL, J. W. and others, Influence of terrain characteristics on snowpack water equivalent, February 1951.
3. MONDRILLO, G., Estimating insolation from atmospheric conditions, March 1951.
4. HIMMEL, J. M., Lysimeter studies of rain-on-snow phenomena, June 1951.
5. BRECHEEN, K. G., Transmission of shortwave radiation through forest canopy, October 1951.
6. BERGER, P., Trial estimates of net longwave radiation from snowpacks, February 1952.
7. BERGER, P., Estimation of net longwave radiation from snow, October 1952.
8. McCLAIN, M. H. (tr.), Evaporation from the snowpack, by M. de Quervain, October 1952.
9. MILLER, D. H., Some forest influences on thermal balance over the snowpack, (reprint), December 1952.
10. BOTTORF, W. L. D. and C. E. Hildebrand, An empirical method of forecasting critical snowmelt inflows to Pine Flat Reservoir, December 1952.
11. ARNOLD, B. and P. Boyer, Heat exchange and melt of late-season snow patches in heavy forest, May 1953.
12. BERGER, P., Radiation in forest at Willamette Basin Snow Laboratory, June 1953.
13. HUMPHREY, H. N. and T. H. Pagenhart, Additional studies of the influence of terrain characteristics on snowpack water equivalent, June 1953.
14. MONDRILLO, G., Preliminary unit-graph studies, Mann Creek, Willamette Basin Snow Laboratory, June 1953.

15. MILLER, D. H., Thermal balances and snowmelt runoff associated with upper-air flow over the western United States in May 1949 and May 1950, September 1953.

16. MILLER, D. H., Snow-cover depletion and runoff, September 1953.

17. HILDEBRAND, C. E. and T. H. Pagenhart, Lysimeter studies of clear weather snowmelt at an unforested site, December 1953.

18. BOYER, P. B., Analysis of January 1953 rain on snow, observations at Central Sierra Snow Laboratory, Soda Springs, California, May 1954.

19. MONDRILLO, G. and C. E. Jencks, Clear weather snowmelt runoff in a densely forested area, Willamette Basin Snow Laboratory, May 1954.

(Supplement to Res. Note 19) MONDRILLO, G., Clear-weather snowmelt runoff in a densely forested area, North Santiam River Basin, with Appendix: Thermodynamics of transpiration in heavy forest during active snowmelt, May 1955.

20. McCLAIN, M. H., Precipitation, evapotranspiration, and runoff, Willamette Basin Snow Laboratory, July 1954.

21. HILDEBRAND, C. E. and T. H. Pagenhart, Determination of annual precipitation, Central Sierra Snow Laboratory, September 1954.

22. MILLER, S., Forecasting seasonal runoff by the water-balance method, September 1954.

23. ROCKWOOD, D. M., A coastal winter-flow index method of forecasting seasonal runoff for Columbia River near The Dalles, Oregon, September, 1954.

24. JENCKS, C. E., Analysis of February 1951 rain on snow in a densely forested area, April 1955.

25. HILDEBRAND, C. E. and T. H. Pagenhart, Lysimeter studies of snowmelt, March 1955.

TECHNICAL BULLETINS
Corps of Engineers, Civil Works Investigations
Project CW-171

1. Criteria for estimating runoff from snowmelt, (Project bulletin 1: objectives of project and administrative details.) May 1949.

2. SNYDER, F. F. Heat balance and amount available for melting snow. June 1949.

3. PARSONS, W. J. Use of snow laboratory data by Sacramento District. November 1949.

4. HULLINGHORST, D. W. Progress report on project CW-171, Criteria for estimating runoff from snow melt. April 1950.

5. MILLER, D. H. The depletion method of estimating solar radiation absorbed by the snow. April 1950.

6. MILLER, D. H. Albedo of the snow surface with reference to its age. April 1950.

7. HERING, W. S. Evaluation of outward long wave radiation from the snow surface. April 1950.

8. HULLINGHORST, D. W. and D. H. Miller. Interim report on a current study (Reconstitution of stream flow from meteorologic data). May 1950.

9. HULLINGHORST, D. W. and D. H. Miller. Refinement of flow estimates of Technical Bulletin No. 8. September 1950.

10. HAMILTON, R. M. Application of estimation procedures to independent data. September 1950.

11. MILLER, D. H. Micro-meteorological conditions over snow pack in open forest: preliminary report on factors influending convective heat-exchange. August 1950.

12. HIMMEL, J. M. Radiation heat exchange between the snowpack and its environment, Central Sierra Snow Laboratory, 27 April - 9 June 1950. September 1950.

13. HILDEBRAND, C. E. The general snowmelt equation. May 1951.

14. HILDEBRAND, C. E. A unit-hydrograph method of hydrograph synthesis for snow-covered areas. September 1952.

15. THOMS, M. E., Determination of areal snow cover by aerial reconnaissance in Kootenai and Flathead Basins, February 1954.

16. ALLISON, I. D., Melting of deep snow packs by conduction of heat from the ground, June 1954.

17. BOYER, P. B. and P. Merrill, Storage effect of snow on the flood potential from rain falling on snow, December 1954.

18. ROCKWOOD, D. M. and C. E. Hildebrand, An electronic analog for multiple-stage reservoir-type storage routing, March 1956.

APPENDIX I - Continued

MISCELLANEOUS REPORTS

1. GERDEL, R. W. - Evaluation of snow cover distribution from horizontal photographs, Cooperative Snow Investigations Progress Report, May 6, 1949. (Unpublished)

2. GERDEL, R. W. - A review of soil moisture measuring methods and apparatus, Cooperative Snow Investigations, Technical report, March, 1949. (Unpublished)

3. MILLER, D. H. - Rain-on-snow flood of 18-20 November 1950, CSSL: Preliminary Report and outline for investigation as of 24 November 1950. Office memo to technical director, CSI, 24 November 1950. (Mimeographed)

4. PATTON, C. P. - Five-year meteorologic summary, station 3, Central Sierra Snow Laboratory. Cooperative Snow Investigations: SIPRE Analytical Unit, 1 May 1952.

5. PATTON, C. P. - Meteorologic elements and snowpack characteristics at micrometeorological project, Central Sierra Snow Laboratory, 1950-51 season, Cooperative Snow Investigations: SIPRE Analytical Unit, 1 June 1952.

6. WALSH, K. J. - Wind-speed and air-temperature gradients for January-May 1951 at micrometeorological project, Central Sierra Snow Laboratory, Cooperative Snow Investigations: SIPRE Analytical Unit, 5 January 1953.

7. WALSH, K. J. - Variations in snowpack density, Central Sierra Snow Laboratory, Cooperative Snow Investigations: SIPRE Analytical Unit, 4 February 1953.

8. Synopsis of Snow Investigations and Plans for FY 1954, August 1953, North Pacific Division, Corps of Engineers, U. S. Army, Portland, Oregon.

APPENDIX II

COMPLETED TOURS OF DUTY, PROJECT CW-171

1) F. F. SNYDER (OCE) June 1949 - Initiation of Project Study of heat balance and amount available for melting snow, (Tech. Bull. 2).

2) C. PEDERSEN (Portland Dist.) June 1949 - UCSL studies; Relationship of radiation to various meteorological elements; seasonal variations of albedo, snow density, water equivalent; degree-day melt rate computations.

3) R. H. CONWAY (Walla Walla Dist.) June 1949 - Preliminary reconstitution 1948 flood, UCSL, to investigate criteria governing snowmelt; inquiries concerning degree-day vs. heat-balance methods.

4) N.J. MACDONALD (Seattle Dist.) June 1949 - Relationship studies; density of new snow vs. max. temp. at Summit, Mont., and Soda Springs, Calif.; normal annual precip. vs. topog., Columbia Basin (similar to USWB study of Colorado Basin); spillway design flood methods from unit hydrographs, UCSL.

5) E. W. McCLENDON (MRD) August,1949 - Study and review of snow hydrology problems; discussion of basic snow and frost problems; discussion of basic snow and frost problems in Missouri River Basin; mountain snowmelt vs. plains snowmelt.

6) M.E. THOMS (Seattle Dist.) Sept. 1949 - Snowmelt determinations published in "Report on Derivation of Standard Project Flood, Skagit River near Sedro Woolley, Washington."

7) W. S. HERING (Walla Walla Dist.) Sept. 1949 - Empirical evaluation of condensation and outward longwave radiation over snow (Tech. Bull. 7); study on upper air temp. as index of mean surface temp.

8) S. A. MILLER (Denver Dist.) Oct. 1949 - Study of temp. index, snow cover, and runoff relationships using concept of "active snowmelt line" (daily temp. trace through melt season of degrees required at index station to produce melt).

9) J. SUMMERSETT, JR., (Portland Dist.) Oct. 1949 - Study and review of snowmelt problems in Willamette Basin Snow Laboratory.

10) S. MILLER (Walla Walla Dist.) Nov. 1949 - Use of temperature data in determining incident radiation (formulas, correlations, results, presented). Discussion of snow hydrology problems, Lucky Peak Dam.

COMPLETED TOURS OF DUTY, PROJECT CW-171 - Continued

11) C.E. JENCKS (Portland Dist.) Feb. 1950 - Study and review; streamflow study of Blue River above Quentin Creek, WBSL.

12) F. C. MURPHY (Seattle Dist.) Feb. 1950 - Review of spillway design problems in Columbia Basin; specifically at Albeni Falls dam site.

13) H. LOBITZ, JR., (Walla Walla Dist.) June 1950 - Study of hydrograph reproduction by the degree-hour method using variable S-curves for distribution of the melt.

14) E. W. McCLENDON (MRD) Aug. 1950 (2nd visit) - Study and review of current methods of estimating streamflow from snowmelt (e.g. Tech. Bull. 8); hydrograph reconstitutions, 1948 and 1950, CSSL.

15) W.S. HERING (Walla Walla Dist.) Sept. 1950 (2nd visit) - Hydrograph reconstitution by thermal-budget method, applying S-curve principles, 1949, Boise River above Twin Springs, Idaho.

16) S. MILLER (Walla Walla Dist.) Nov. 1950 (2nd visit) - Hydrograph reconstitution by degree-day method using constant loss of 7,000 d.s.f., 1949, Boise River above Twin Springs, Idaho.

17) F.C. MURPHY (Seattle Dist.) Dec. 1950 (2nd visit) - Discussion and review; spillway design problems, Libby project.

18) R. ASCHENBRENNER (Walla Walla Dist.) Jan. 1950 - Various reconstitutions by degree-day and heat-balance methods, 1943 and 1949, Boise River above Twin Springs, Idaho.

19) N. J. MACDONALD (Seattle Dist.) Jan. 1951 (2nd visit) - Libby damsite spillway design study: 1947 hydrograph reconstitution by degree-day method, Kootenai River at Libby, Montana.

20) M. J. ORD (Walla Walla Dist.) Feb. 1951 - Discussion and review of Boise River studies and of general snow hydrology for application to District snowmelt runoff problems.

21) G. L. GAY (Portland Dist.) Feb. 1951 - Green Peter Dam Study.

22) R. H. CONWAY (Walla Walla Dist.) Mar. 1951 (2nd visit) - Reconstitution of '36, '43, '48, and '50 flood hydrographs by degree-day methods, Snake River at Heise, Idaho

23) M. E. THOMS (Seattle Dist.) Mar. 1951 (2nd visit) - Spillway design studies, Kootenai River at Libby, Montana: '42 and '48 flood reconstitutions by degree-day method, '47 and '48 reconstructions by heat-balance method.

24) S. MILLER (Walla Walla Dist.) Sept. 1951 (3d visit) - Runoff volume forecast study, Boise River above Lucky Peak, Idaho.

25) J. SUMMERSETT, JR. (Walla Walla Dist.) Oct. 1951 (2nd Visit) UCSL snow cover vs. heat-exchange study; discussion and study of reconstitution methods, degree-day vs. thermal budget.

26) D. E. PHILLIPS (Walla Walla Dist.) Feb. 1952 - Snow cover depletion vs. accumulated degree-days; flood reconstitutions using maximum temperatures as index.

27) D. M. ROCKWOOD (NPD) Jan. 1952 - Forecasting flood season runoff from early-season flows and temperatures for Columbia River at the Dalles.

28) M. J. ORD (Walla Walla Dist.) Feb. 1952 (2nd Visit) - Study and review of heat-balance factors; discussion of degree-day vs. heat-balance methods for basin application.

29) F.C. MURPHY (Seattle Dist.) Feb. 1952 (3rd visit) - Discussion and review of available procedures for runoff forecasting and reservoir regulation, Libby Dam.

30) M.E. THOMS (Seattle Dist.) Feb. 1952 (3d visit) - Flood reconstitutions, Kootenai River at Libby, Montana: degree-day and heat-balance methods.

31) R. H. CONWAY (Walla Walla Dist.) Mar. 1952 (3d visit) - Synthetic reconstitutions of Boise River floods, '43, '48, heat-balance method.

32) M. LARSON (Portland Dist.) Apr. 1952 - Study and review: Snowmelt studies CSSL, '49; work on project CW-170 (Radioisotope-radiotelemetering snow gage).

33) C. JENCKS (Portland Dist.) June 1952 (2nd visit) - Rain-on-snow studies, '52, WBSL: lapse rate study, WBSL.

34) M. E. THOMS (Seattle Dist.) Mar. 1953 (4th visit) - Studies preparatory to draft; "Forecasting inflows to Libby Reservoir."

35) N. J. MACDONALD (Seattle Dist.) Mar. 1953 (3d visit) - Seasonal forecast study, Albeni Falls Dam.

36) M. LARSON (Portland Dist.) Apr. 1953 (2d visit) - Forecasting and reservoir regulation procedures, Detroit Dam, (N. Santiam River, Ore.)

37) M.E. THOMS (Seattle Dist.) Oct. 1953 (5th visit) - Preparation of draft: "Determination of areal snow cover by aerial reconnaissance in Kootenai and Flathead Basins." (Tech. Bull. 15)

38) C.W. TIMBERMAN (MRD) Oct. 1953 - Study and analyses for draft: "Reconstitution of 1950 snow-melt flood on Cannonball River at New Leipzig, North Dakota;" also 1950 flood, Heart River Basin, North Dakota.

39) N. J. MACDONALD (Seattle Dist.) Dec. 1953 (4th visit) - Seasonal forecast procedure, Albeni Falls Dam.

40) H. D. WILDERMUTH (Los Angeles Dist.) Jan. 1954 - Design floods for Gila River basin above Painted Rock damsite; (sub-basins studied: San Francisco River at Clifton, Arizona, and Verde River at confluence with Salt River.)

41) H. N. HUMPHREY (SPD) Jan. 1954 - Same as 40 above.

42) C.A. BURGTORF (Garrison Dist.) Jan. 1954 - Reconstitutions of spring 1950 and 1952 snowmelt floods on Spring Creek above Zap, North Dakota.

43) G.E. GALLAGHER (Portland Dist.) Jan. 1954 - Criteria for forecasting seasonal runoff from snowmelt, Middle Fork Willamette River above Lookout Point Dam, Oregon.

44) K.A. JOHNSON (Omaha Dist.) Jan. 1954 - Reconstitution of spring snowmelt floods on Papillion Creek at Ft. Crook, Nebraska, 1948, and Spring Creek at Zap, North Dakota, 1952.

45) F.C. MURPHY (Seattle Dist.) Mar. 1954 (4th Visit) - Discussion and review, seasonal forecast procedures, Hungry Horse Dam.

46) J. W. HANSON (Portland Dist.) Apr. 1954 - Study and review, forecasting seasonal runoff, Columbia River at the Dalles.

47) S. NAIMARK (Portland Dist.) June 1954 - Daily operation schedule for Detroit Reservoir (N. Santiam River, Ore.)

48) K.W. WISE (Walla Walla Dist.) Sept. 1954 - Forecast procedure for Snake River above Moran, Wyoming.

49) N.J. MACDONALD (Seattle Dist.) Jan. 1955 (5th visit) - Forecast procedure for seasonal runoff into Hungry Horse Reservoir (So. Fork, Flathead River, Montana)

50) R. J. DEFANT (Portland Dist.) Jan. 1955 - Standard Project Flood for Cougar Dam (So. Fork, McKenzie River, Oregon).

51) N. J. MACDONALD (Seattle Dist.) Mar. 1955 (6th visit) - Seasonal runoff forecast, Hungry Horse Reservoir.

52) O.C. JOHNSON (Portland Dist.) Mar. 1956 - Seasonal runoff forecast, Lookout Point Reservoir (Mid-Fork, Willamette River, Ore.)

APPENDIX III

LIST OF SNOW HYDROLOGY SYMBOLS

Symbol	Concept
a	Albedo (reflectivity) of snow pack and/or ground ($a = I_r/I_i$)
A	Area (of snow cover, of drainage area, etc.)
b	Ablation (decrease in depth) of snow pack
B	Thermal quality of snow pack ($B = 1 - f_p/100$)
c_p	Specific heat (constant pressure)
C_r	Recession constant (ratio of current rate of flow to previous days rate of flow)($q_t = q_o C_r^{\,t}$)
d	Depletion (decrease in areal cover) of snow pack
D	Depth (of snow pack, etc.) Coefficient of determination
e	Vapor pressure a subscript denotes vapor pressure of air s subscript denotes saturated vapor pressure Base of Naperian logarithms Emissivity
f	Infiltration rate
f_p	Liquid-water content of the snowpack, in percent of W
f_p'	Liquid-water deficiency of the snowpack, in percent of W
f_p''	Liquid-water-holding capacity of the snowpack, in percent of W, ($f_p'' = f_p + f_p'$)
F	Forest cover

Symbol	Concept
G	Intensity of radiation (all wave) d subscript denotes radiation directed downward or toward the snow pack u subscript denotes radiation directed upward or from the snow pack
h	Rate of net heat transfer to snow pack from its environment ($h = h_c + h_e + h_g + h_p + h_{rl} + h_{rs}$) c subscript denotes convection (and conduction) from air ce subscript denotes convection-condensation from air ($h_{ce} = h_c + h_e$) e subscript denotes condensation (or evaporation) from air ($h_e = kq_e$) g subscript denotes conduction from ground p subscript denotes heat capacity of rain ($h_p = ki_rT$) r subscript denotes all-wave radiation ($h_r = h_{rl} + h_{rs} = G_d - G_u$) rl subscript denotes long-wave radiation ($h_{rl} = R_d - R_u$) rs subscript denotes short-wave radiation ($h_{rs} = I_i - I_r = (1 - a)\ I_i$)
H	Quantity of net heat transfer to snow pack from its environment ($H = H_c + H_e + H_g + H_p + H_{rl} + H_{rs}$)(subscripts as above for rate of net heat transfer)
i	Intensity of precipitation r subscript denotes rainfall s subscript denotes snowfall
I	Intensity of short-wave radiation i subscript denotes incident radiation o subscript denotes radiation at upper limit earth's atmosphere r subscript denotes reflected radiation ($I_r = aI_i$)
k	Coefficient, exponent, or conversion factor
k_c	Thermal conductivity

Symbol	Concept
K_i	Solar radiation transmission coefficient for forest (ratio of radiation incident on snow surface beneath forest to radiation incident in open)
K_r	Ratio of downward long-wave radiation, R_d, to that of a hypothetical black body at air temperature ($K_r = R_d/\sigma T_a^4$)
l	Loss rate _e_ subscript denotes loss by evaporation _g_ subscript denotes loss by deep percolation _t_ subscript denotes loss by transpiration _et_ subscript denotes loss by evapotranspiration
L	Loss (quantity) (subscripts as above for loss rate)
m	Rate of snow melt (subscripts as above for rate of net heat transfer)
M	Quantity of snow melt (subscripts as above for rate of net heat transfer)
n	Number of items
N	Cloud cover
p	Atmospheric pressure
P	Quantity of precipitation _r_ subscript denotes rainfall _s_ subscript denotes snowfall (water equivalent)
q	Rate of stream flow, runoff, or discharge (water transport) _e_ subscript denotes rate of condensation _g_ subscript denotes rate of ground water discharge _i_ subscript denotes rate of interflow _l_ subscript denotes loss rate _o_ subscript denotes initial rate

List of Snow Hydrology Symbols - Continued

Symbol	Concept
Q	Quantity of water (subscripts as above for rate of stream flow, etc.)
r	Correlation coefficient
R	Intensity of long-wave radiation (subscripts as above for intensity of radiation)
s_y	Standard deviation
s_{yx}	Standard error of estimate
S	Storage (= Inflow-Outflow)
t	Time c subscript denotes concentration time s subscript denotes storage time
T	Temperature a subscript denotes air temperature d subscript denotes dew-point temperature g subscript denotes ground temperature s subscript denotes snow temperature w subscript denotes wet-bulb temperature
U	Relative humdiity
v	Wind speed
V	Wind travel
w	Mixing ratio
W	Water equivalent of snow pack
W_f	Liquid water in snow pack $W_f = f_p W/100$
W_p	Precipitable water in atmosphere

List of Snow Hydrology Symbols - Continued

Symbol	Concept
z	Altitude, height
Z	Zenith angle of sun
λ	Wave Length (lambda)
ρ	Density, specific gravity ("density") of snow (rho)
σ	Stefan-Boltzmann constant (sigma)
ϕ	Latitude (phi)
β	(beta) Standard partial regression coefficient
μ	(mu) micron
π	(pi) 3.1416
Σ	(Sigma) sum of...
∞	infinity
$>$	greater than
$<$	less than
$\approx$	approximately equal to