Estimation of Spatially Distributed Values of Daily Precipitation in Mountainous Areas

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Introduction

Hydrologic modeling in mountainous areas is challenging from the outset due to difficulties in simply estimating the basic watershed inputs, such as precipitation and temperature, which are subject to orographic effects and high spatial variability. With the spatially distributed hydrologic models that have been developed in recent years (e.g., Wigmosta et al., 1994), there are now even more demanding requirements for estimation of watershed inputs; that is, spatial averages are no longer sufficient, but spatially distributed values of inputs are required.

With the availability of geographic information systems (GIS), digital elevation data, fast computer workstations, and high-elevation data collection sites such as those in the western United States (SNOTEL) and Canada, it is now possible to make estimates of watershed input quantities on a spatially distributed basis. Although new radar systems may become useful in estimating mountain precipitation, there are limitations caused by terrain blocking the radar's view and proper accounting for orography. In addition, radar is limited to measuring precipitation and does not measure other important hydrometeorological quantities. There remains a need for a spatial interpolation procedure based on conventional ground measurements. This paper describes such a procedure. Although the focus here is on precipitation, the procedure has also been applied to temperature and snow water equivalent data. Daily precipitation data have been used here because this is a standard reporting frequency, it is a time step often used in hydrologic modeling, and it is a time resolution that is feasible to interpolate.

Spatial Estimation Procedure

The procedure used here is that reported by Garen et al. (1994). In that paper, the work focused on mean areal precipitation, and the procedure was applied to a small research watershed. In the current paper, focus is on using the procedure to estimate spatial precipitation fields and to apply it to a larger watershed with a more typical data site density.

The algorithm is based on detrended kriging. Kriging is an optimal spatial interpolation procedure that calculates an estimate of a quantity at an unmeasured site as a weighted sum of nearby measurements. The weights are derived by solving a system of linear equations, the coefficients of which represent the distances among the data sites and the site to be estimated as well as the spatial correlation structure of the measured quantity. The weights are generally larger in magnitude for nearby measurements and smaller for more distant measurements, and, in this application, the weights are constrained to sum to unity. Detrending is required to account for nonstationarity of the field due to orography. This is accomplished by calculating linear precipitation-elevation relationships from the measured data and performing spatial interpolation on the residuals. Precipitation-elevation relationships are calculated separately for contiguous groupings of several days in length, to avoid instability that could be introduced by using individual days to calculate them. These relationships also represent the time-varying orographic effects due to differing storm types, intensities, and directions. The spatial correlation structure of the regression residuals is described by a linear semivariogram. Precipitation is estimated for each grid cell within a watershed, and these can either be used in spatially distributed form or arithmetically averaged to give the mean precipitation over the watershed or sub-areas thereof.

The calculation procedure is as follows:

(1) For each grid cell, calculate the kriging weights to be applied to each precipitation station. These weights will be used for those days when all precipitation stations have data (no missing data).

(2) Choose a temporal aggregation period for the precipitation-elevation relationships. Typical choices include 7, 14, and 28 days, or storm periods, if these can be adequately and conveniently defined.

(3) For each aggregation period, calculate average daily precipitation at each station, using only days for which at least one station has precipitation ("wet" days). Calculate the linear regression of

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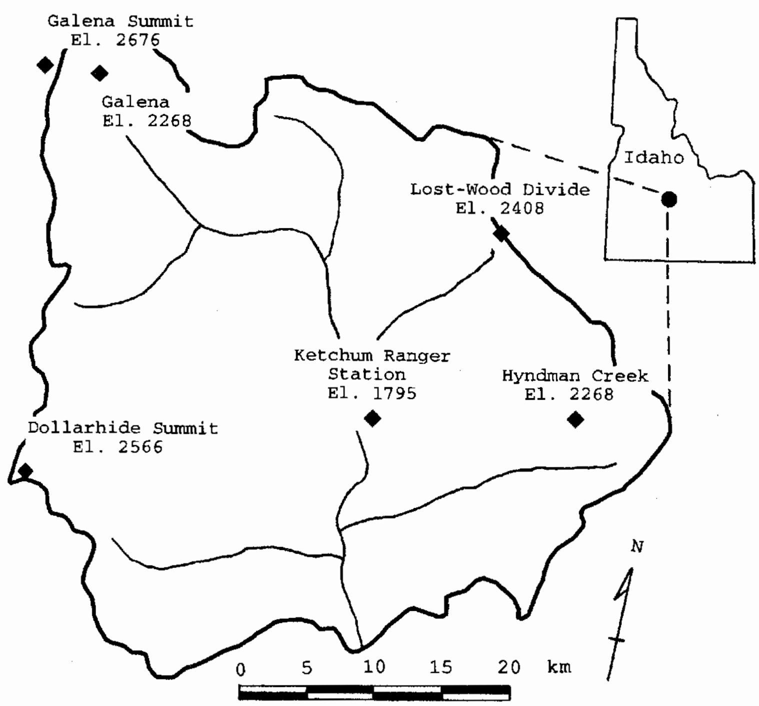

Fig. 1 Big Wood River Watershed and Precipitation Gauges (Elevation in metres)

average daily precipitation (dependent variable) versus station elevation (independent variable).

(4) For each wet day within the period, subtract the linear precipitation-elevation trend from the precipitation observations to obtain the residuals. For each day where all stations have zero precipitation ("dry" days), set the precipitation to zero for all grid cells, and do not process these days further.

(5) For each wet day within the period and for each grid cell, calculate the estimated grid cell residual by multiplying the precipitation station residuals by the kriging weights appropriate for the grid cell and summing. If one or more precipitation stations have missing data, the kriging weights must be recalculated to use only the stations that have data; otherwise, the weights calculated in step 1 can be used.

(6) For each wet day within the period and for each grid cell, add the linear precipitation- elevation trend to the grid cell residual, based on the elevation of the grid cell, to obtain the estimated grid cell precipitation.

The basic idea in this procedure, then, is to separate the sources of spatial variability into a component due to elevation (vertical) and a component due to distance (horizontal). These are modeled separately via detrending and kriging, respectively. Once the overall orographic effect is removed by detrending, the remaining variability is assumed to be a function of distance. So if a grid cell is near one of the precipitation stations, and that station has, say, a positive residual from the precipitation-elevation trend, the grid cell is also likely to have a positive residual. In this way, local effects are accounted for as they are represented by the behavior of the individual precipitation stations with respect to the overall precipitation- elevation trend and the proximity of the grid cells to the stations.

Application to Big Wood River

The procedure was applied to the Big Wood River in southcentral Idaho. The watershed above the streamgaging station at the town of Hailey has an area of 1660 km2, with elevations ranging between 1610 and 3660 m. The GRASS geographic information system (United States Army Corps of Engineers, 1993) was used to establish grid networks of two resolutions, 0.9 km (2008 grid cells) and 2.5 km (256 grid cells) by arithmetic averaging of the base digital elevation data, which has a resolution of 100 m. Daily precipitation data for the period 1983-1993 at six stations were used. One station, Ketchum Ranger Station, is from the National Weather Service cooperative network; the other five are from the Natural Resources Conservation Service (formerly Soil Conservation Service) SNOTEL network. The stations are shown on the map in Figure 1, and Figures 2 and 3 display the elevation at the two spatial resolutions.

The general nature of the precipitation-elevation relationship can be seen from the regression using mean annual precipitation (over the period 1983-1993) shown in Figure 4. Although the regression lines for individual periods, years, and temporal resolutions vary considerably (see Table 1 for an example), the relative positions of the stations around these lines is quite consistently the same as that shown in Figure 4. For example, Galena Summit apparently has some local influence making it drier than one might expect for its elevation, and similarly, Dollarhide Summit and Lost-Wood Divide are somewhat wetter than the regression would indicate. Note also that the positions above or below the line do not seem to be related to regions within the watershed; for example, Lost-Wood Divide and Hyndman Creek are both on the eastern side of the watershed, and Dollarhide Summit and Galena Summit are both on the western side. With only six stations and no demonstrable difference in the precipitation-elevation relationship among watershed sub-regions, it was felt that it was legitimate to use all stations together to estimate precipitation throughout the entire watershed.

Another point to note in Figure 4 is that it does not indicate a flattening of the relationship at the higher elevations, although it must be kept in mind that the highest data site is still almost 1000 m lower than the highest peaks in the watershed. This flattening has been suggested by many and was recently demonstrated in western British Columbia by Loukas and Quick (1994). Since a flattening cannot be proved with measurements in the Big Wood River, the linear trend was used throughout the elevation range. If this were a poor assumption, one would expect the procedure to overestimate the volume of precipitation. One indication of the reasonableness of the values is to note that in Figure 4, the mean areal precipitation (MAP) estimated by the procedure, when plotted at the mean watershed elevation, falls close to the regression line, well within an expected magnitude of deviation from the line.

Another way of evaluating the reasonableness of the estimated precipitation values is to compare them with runoff and evapotranspiration values. The water year sums of MAP were compared to runoff and potential evapotranspiration estimated by the temperature- based method of Hargreaves (Jensen et al., 1990). The temperature data used were watershed area average daily values obtained by applying this detrended kriging procedure to maximum, minimum, and average temperature observations at the six stations. Since the SNOTEL stations began their temperature records in 1989, this comparision was only possible for the 1989-1993 water years. These results are given in Table 2 along with calculated runoff coefficients and actual evapotranspiration ratios (precipitation minus runoff divided by potential evapotranspiration; this assumes that there is negligible change in watershed storage from year to year). The magnitudes of the values and the ratios indicate that the quantity of precipitation estimated by the procedure is reasonable.

To obtain an indication of the differences due to spatial resolution and the temporal aggregation of precipitation-elevation relationships, the daily spatial fields were averaged over the watershed and summed for each water year. For the two spatial resolutions (0.9 and 2.5 km) and for 7, 14, and 28 day temporal aggregations, there was virtually no difference in total water year precipitation among the six combinations. The daily values of areally averaged precipitation for water year 1993 were also compared among the six combinations, and again the differences were very small, usually less than a millimeter on any given day. These findings are similar to the results in the previous study reported by Garen et al. (1994) and indicate the robustness of the procedure.

Another issue requiring some attention is to verify the use of a linear semivariogram. After examining

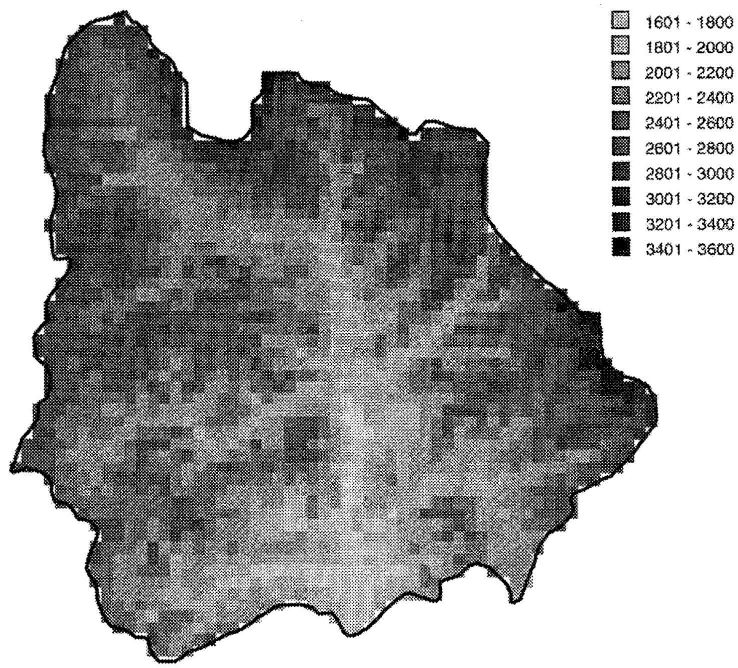

Fig. 2. Elevation Field (in meters) at 0.9 km Resolution

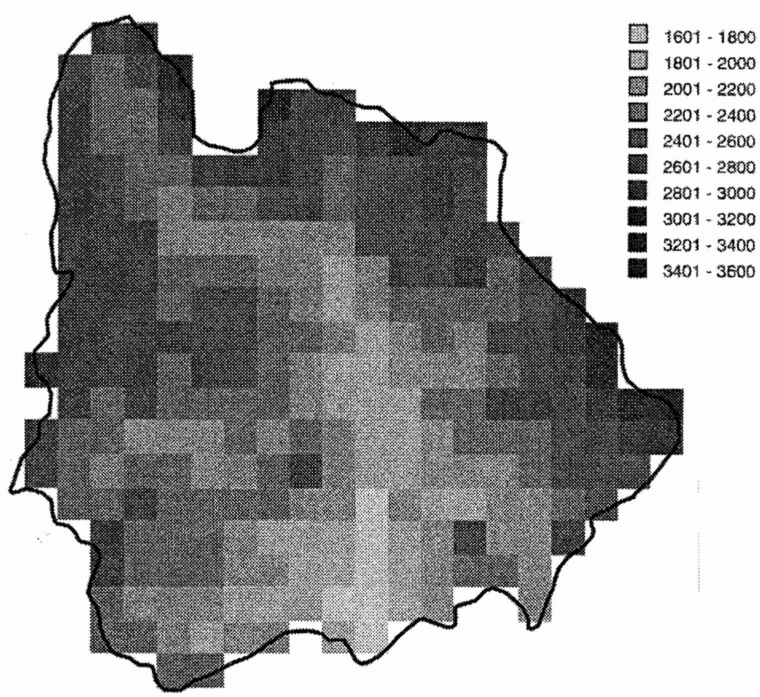

Fig. 3. Elevation Field (in meters) at 2.5 km Resolution

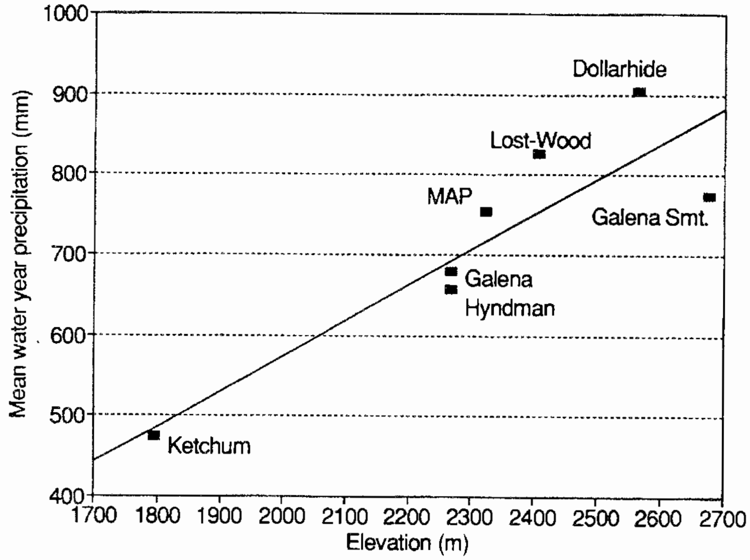

Fig. 4 Mean Water Year Precipitaion (1983-1993) vs. Elevation. Estimated mean areal precipitaion (MAP) is plotted at the mean watershed elevation.

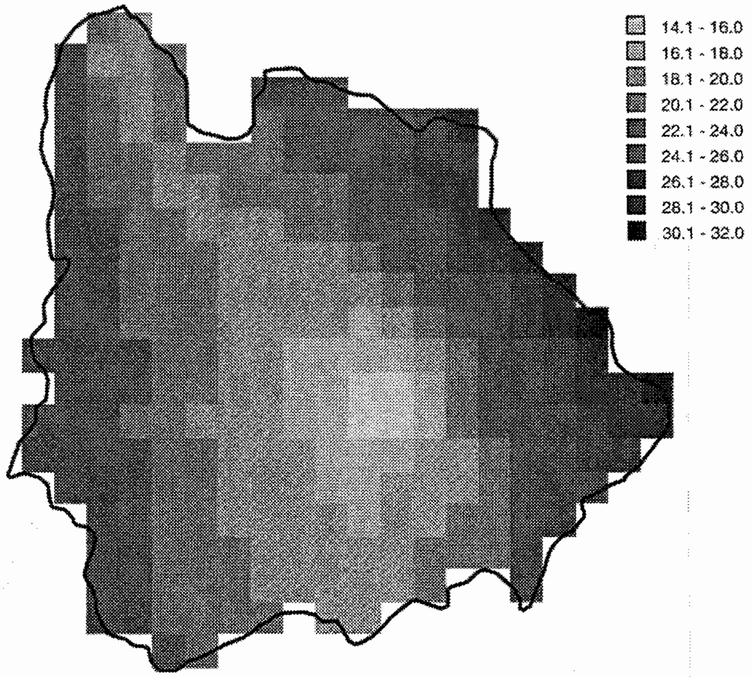

Figure 5. Precipitation Field (mm) for October 3, 1992, 2.5 km Resolution, 28-Day Aggregation Period. Observed values at precipitation stations: Ketchum 14.5, Hyndman 27.9, Galena 17.8, Lost-Wood 30.5, Dollarhide 25.4, Galena Summit 25.4.

many empirical semivariograms estimated from the detrended daily data, the linear semivariogram appears to be appropriate. Some days have stronger spatial relationships than others, but in general, the shape is simply a general upward trending scatter, with no noticeable flattening with distance. At a daily time scale, the range of influence apparently is greater than the maximum distance among the precipitation stations and grid cells in this watershed (approximately 50 km). As a comparison, Obled et al. (1994) observed that in the Mediterranean region, the range of influence is 15 km at a 0.5 hour time step and 20 km at a 1 hour time step. If one were to extrapolate these observations, a 50+ km range of influence at a daily time step seems reasonable.

An example of the spatial output is given in Figure 5, which displays the precipitation field for October 3, 1992, at the 2.5 km resolution using 28-day periods. Although a detailed examination of the precipitation fields at the two resolutions was not done, it is felt that the 0.9 km resolution is probably too fine given the data network and size of this watershed. The 2.5 km resolution seems more reasonable; for comparison, Daly et al. (1994) used 6 km and in more recent work 3-4 km, while the new Doppler radars estimate precipitation with a 4 km spatial resolution. The choice of spatial resolution must balance a faithful representation of the terrain with a reasonable level of detail that can be supported by the data network density. Unfortunately, it is not clear exactly how to make those tradeoffs, so the choice of a spatial resolution remains subjective at this point.

Conclusion

A procedure has been developed that estimates spatial fields of precipitation at a daily time scale in mountainous areas. Its application to the Big Wood River indicates that the procedure gives reasonable values. At minimum, this procedure provides the basis for improving the estimation of areal average watershed precipitation inputs in that it: (1) uses specific precipitation-elevation relationships for each time period rather than assuming that climatological average relationships always apply; (2) explicitly accounts for the spatial correlation and variability of the precipitation fields; and (3) determines station weights objectively. Obled et al. (1994) remarked that they felt the greatest value in describing the spatial variability of precipitation was to improve the estimate in the overall volume of precipitation input to the watershed. If this is so, then this detrended kriging procedure provides the basis for making such an improvement. Additionally, however, the procedure provides the basis for estimating the time series

TABLE 1. Ranges of Precipitation-Elevation Regression Line Statistics for Average Daily Precipitation, 2.5 km Grid, 28-Day Aggregation Period, 1983-1993

Period	Dates (non-leap year)	Intercept (mm)	Slope (mm / 10^3 m)	Correlation Coefficient
1	Oct 1 - Oct 28	-4.39 - 8.96	-3.24 - 3.69	-0.53 - 0.94
2	Oct 29 - Nov 25	-10.16 - 0.09	0.93 - 6.66	0.41 - 0.99
3	Nov 26 - Dec 23	-3.86 - -0.52	1.25 - 3.98	0.53 - 0.89
4	Dec 24 - Jan 20	-15.73 - 0.78	1.12 - 8.05	0.61 - 0.97
5	Jan 21 - Feb 17	-6.55 - 2.24	0.77 - 5.14	0.32 - 0.93
6	Feb 18 - Mar 17	-6.00 - 0.16	2.20 - 5.42	0.58 - 0.98
7	Mar 18 - Apr 14	-7.44 - -0.98	0.84 - 4.60	0.57 - 0.97
8	Apr 15 - May 12	-5.55 - 2.01	0.42 - 3.32	0.29 - 0.96
9	May 13 - Jun 9	-0.18 - 7.85	-0.91 - 2.41	-0.16 - 0.76
10	Jun 10 - Jul 7	-3.27 - 4.22	-0.79 - 1.78	-0.48 - 0.62
11	Jul 8 - Aug 4	-1.88 - 5.83	-1.77 - 1.33	-0.69 - 0.84
12	Aug 5 - Sep 1	-1.59 - 5.72	-1.61 - 1.74	-0.37 - 0.79
13	Sep 2 - Sep 30	-2.99 - 6.51	-0.79 - 2.56	-0.29 - 1.00

TABLE 2. Mean Areal Water Year Precipitation, Runoff, and Potential Evapotranspiration (mm) for 2.5 km Grid, 28-Day Aggregation Period

Water Year	Precip.	Runoff	Potential ET	Runoff Coeff.	Actual ET Ratio
1983	1123	454		0.40	
1984	902	375		0.42	
1985	731	215		0.29	
1986	964	371		0.38	
1987	541	153		0.28	
1988	528	131		0.25	
1989	740	180	782	0.24	0.72
1990	678	148	781	0.22	0.68
1991	673	148	748	0.22	0.70
1992	493	119	828	0.24	0.45
1993	920	283	707	0.31	0.90

Notes: (1) Runoff coefficient is runoff divided by precipitation.
(2) Actual evapotranspiration ratio is the difference between precipitation and runoff divided by potential evapotranspiration.

of spatial precipitation fields needed for spatially distributed hydrologic modeling.

Certainly, there are refinements that can be made. For example, other topographic or meteorological information may be useful in improving the detrending. The procedure has also been used for other quantities, such as temperature and snow water equivalent, but this could use further testing. Guidelines for choosing an appropriate spatial resolution given the data network available and the complexity of the terrain are needed. These things will come with further work on the procedure and by gaining experience in applying it and using its results in hydrologic modeling.

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