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COMPARISON OF ESTIMATES OF SNOW INPUT WITH A SMALL ALPINE CATCHMENT

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ABSTRACT

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We have used five methods to estimate the snow water equivalent input to the Glacier Lakes Ecosystem Experiments Site (GLEES) in south-central Wyoming during the winter 1987–1988 and to obtain an estimate of the errors. The methods are: (1) the Martinec and Rango degree-day method; (2) Wooldridge et al. method of determining the average yearly snowfall from tree morphology; (3) precipitation gage measurements from the Wyoming Water Research Center Snowy Range Observatory; (4) NADP collector data; (5) an independent estimate from snow core data from a small catchment in the GLEES. Estimated water input ranged from a low of 65 cm H_2O (liquid water equivalent) for the precipitation gage to a high of 85 cm H_2O for the Martinec and Rango method. An evaluation of the biases in the methods indicate that the true value may be nearer the high end of this range.

INTRODUCTION

We have been studying atmospheric deposition to a 200-ha ecosystem in the Snowy Range of the Medicine Bow Mountains in south-central Wyoming. The Glacier Lakes Ecosystem Experiments Site (GLEES) was chosen because it is an alpine–subalpine area with lakes and streams that are sensitive to acidic inputs from the atmosphere. The major input to this catchment is water in the form of snow. Accurate estimation of the quantity of snow input to the catchment each year is basic to understanding the processes by which atmospheric chemicals affect these ecosystems. This paper compares five methods of snow input estimation for the 1987–1988 snow accumulation season.

The accurate estimation of the snow input to a catchment is a difficult task. Snow courses and precipitation gage data can be correlated with streamflow but do not give reliable estimates of the total input. Snow core surveys are very labor intensive and must be carefully designed for acceptable accuracy (Elder et al., 1988; Dozier and Bales, 1990).

Furthermore, in the absence of reliable absolute measurements of the amount of snow in the catchment, it is impossible to determine the absolute accuracy of any estimate. Evaluations of the accuracy of different methods of estimating the snow quantity in mountainous watersheds are difficult to obtain and are necessarily largely judgmental until an absolute calibration can be performed. In the absence of an absolute measure, an intercomparison of different methods of estimation can narrow the range of possible values and provide estimates of the potential errors. The five methods discussed in this paper for determining the amount of snow input to the GLEES are: (1) the Martinec and Rango (1986) degree-day method; (2) a method of determining the average yearly snowfall from tree morphology (Wooldridge et al., 1990); (3) point measurements from precipitation gages operated by the Wyoming Water Research Center Snowy Range Observatory at the GLEES; (4) an Alter-shielded precipitation collector at the NADP site in the GLEES; (5) an estimate from snow core data from a small subcatchment that is a part of GLEES.

Essentially, we perform a preliminary evaluation of the Martinec and Rango (1986) method for possible use in monitoring the yearly snow accumulation in our catchment. Their Snowmelt Runoff Model was developed to estimate

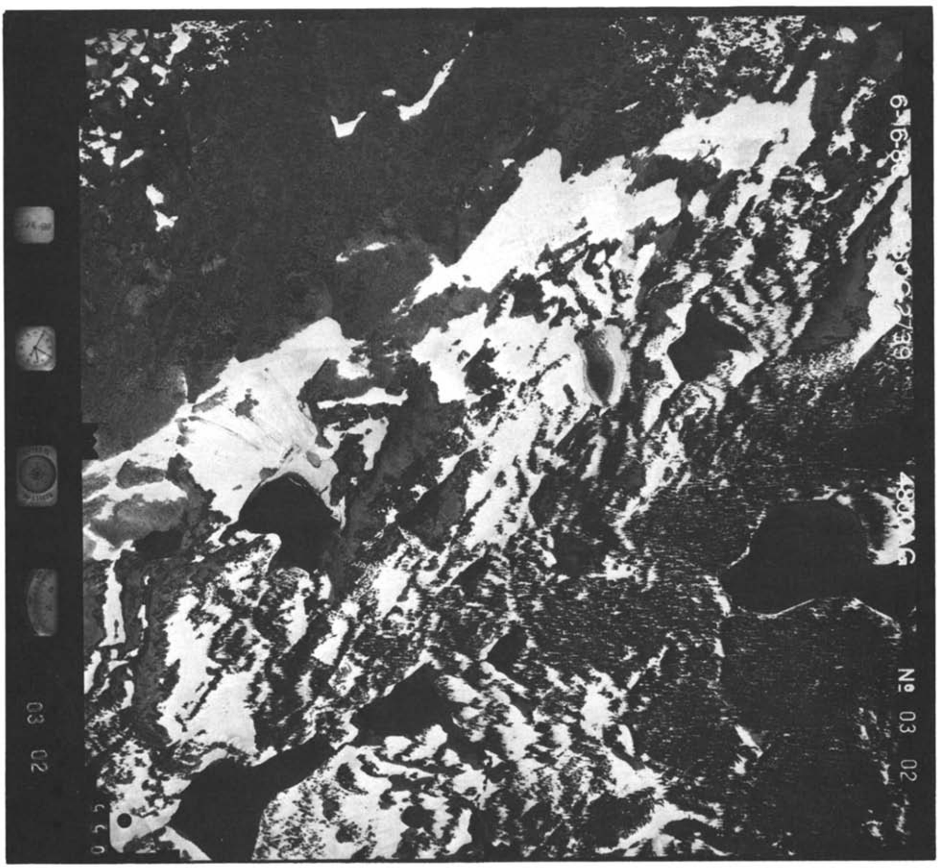

Fig. 1. Aerial photo of GLEES, 16 June 1988.

snowmelt runoff on mountain basins. By using the snowmelt approach in combination with remote sensing observations of the snow disappearance, the initial snow water equivalent can be estimated (Martinec and Rango, 1981). The snowmelt approach depends on a simple degree-day index to estimate the amount of vertical melt on each day. The horizontal extent of melt is determined from a limited set of snow area fraction measurements that can be made conveniently from aerial photos. The attraction of this method is its simplicity and the limited amount of data that are necessary for its application. Also, the measurement of the relative snow-covered area is straightforward and can be performed to a high degree of accuracy if necessary. The major disadvantage of the Martinec–Rango method is that the degree-day index of vertical melt lumps together a number of very complex processes that actually determine the amount of snow melt. Until the method is tested in a variety of conditions and locations, its applicability to a particular catchment, such as the GLEES, is open to question. As discussed above, the calibration of an estimation method is a complex and expensive undertaking. This analysis examines whether the method gives estimates that seem reliable enough to warrant further testing.

MARTINEC–RANGO DEGREE DAYS

Martinec and Rango have developed a model that uses degree days above 0°C, and snow-covered area depletion curves as inputs to estimate snowmelt runoff (e.g. Martinec and Rango, 1986). Abstracting from their eqn. (1), the snowmelt for Day n is:

$$Q_n = \alpha T_n A_n \tag{1}$$

where α is the degree-day factor (cm °C^{-1} day^{-1}), T is the number of degree days above 0°C, and A is the snow-covered area. T_n can be computed for GLEES using temperature data from a weather monitoring station on the site. A_n can be determined from aerial photos (Fig. 1).

Snow area recession

Aerial photos were taken of the GLEES on 13 May, 16 June and 21 July 1988 at a scale of 1 in: 800 ft. They were digitized on a microcomputer-based image analysis system consisting of a Compaq 386 computer, a Dage-MTI camera, a Data Translation DT2850 image capture board, and DT 2858 image processor. The Dage-MTI camera allows a wide range of control over black level, brightness level, and the brightness vs. output-level curve. This control allowed us to discriminate between the snow-covered and all other elements of the image. The resulting image was forced to be either black or white by choosing an appropriate brightness discrimination range. The relative areas were determined by a pixel count.

Figure 2 shows the recession curve in percent of area vs. degree days. It is adequately fit by a cubic spline function which was used for interpolation in

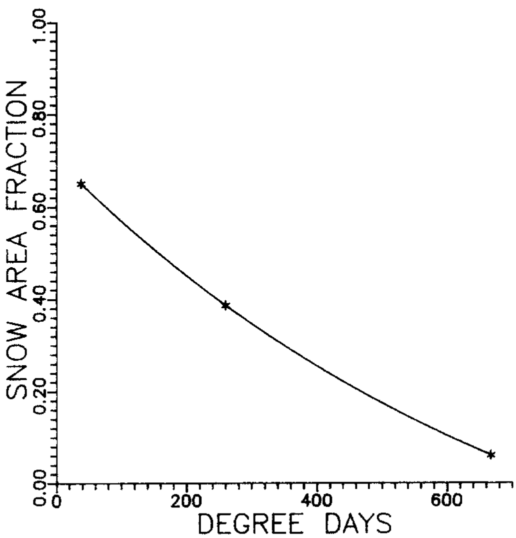

Fig. 2. Degree days vs. snow-covered area fraction. The three points were determined from aerial photos for the GLEES.

the calculations below. The use of three points have been shown to be adequate to determine this curve to an accuracy consistent with the accuracy of this method (Martinec and Rango, 1987; Rango, 1989).

Degree-day factor

The degree-day factor has been empirically determined by Martinec (Martinec and Rango, 1986) to be:

$$\alpha = 1.1\ \rho_s \tag{2}$$

where ρ_s is the snow density. Martinec and Rango (1986) show a plot of snow density vs. date that gives a range of densities from $\sim 400\,\mathrm{kg\,m^{-3}}$ on 10 May to

TABLE 1

GLEES mean snow densities

Pit date	Density (kg m^{-3})
26 Feb.	330
17 Mar.	373
30 Mar.–20 Apr. (61 cores)	410[1]
22 Apr.	392
17 May	386
1 June	485

[1]Corrected for + 9% systematic error (Work et al., 1965).

$500\,kg\,m^{-3}$ on 20 July. From snow pit (Bales et al., 1990) and snow core data (F.A. Vertucci, personal communication) at the GLEES, the mean snow densities are given in Table 1. For these calculations we used a density of $435\,kg\,m^{-3}$, the average of 17 May and 1 June. The reason for choosing this density is discussed in the next paragraph. Because we had no data for July we used a value of $500\,kg\,m^{-3}$ from Martinec and Rango (1986).

Starting date

It is not clear what should be used as a starting date for this application of the Martinec–Rango method. Before the snowpack contains liquid water throughout its whole depth, it has a cold content that must be satisfied to bring it to the freezing point and then to satisfy its irreducible water content. We chose the date at which water flow initiated at the base of the snowpack, 21 May (Bales et al., 1990). Our reasoning is that prior to that date, the liquid water percolating into the snowpack contributed to its densification. By using the snow density at that date as our starting density for the melt season average, we could account for the melt up to that point. In fact the limited density date that were available necessitated that we estimate the density. This is discussed further in the section on errors.

Martinec–Rango result

Accumulating the product of degree-days, relative area and the degree day factor gives:

$$Q_{tot} = 85\,\mathrm{cm\,H_2O} \tag{3}$$

for the total snowfall for the winter period to maximum accumulation in May.

TREE MORPHOLOGY SURVEY

Methodology

Wooldridge et al. (1990) used tree morphology to estimate the long term averge snow depth in GLEES. The area was divided into 100 m grid units for snow depth sampling (Fig. 3). Individual trees were identified at 100 m grid point intersections, if possible, and measured for height of snow accumulation as indicated by protection from abrasion damage or the height of fungus damage (Fig. 4).

Isopleths of 0.5 m depth intervals were constructed from the point measurements. These were overlaid on the 16 June aerial photos. Adjustments were made to the contours such that they matched the snow areas in the photos as closely as possible without violating the point depth measurements. Close correspondence was achieved, increasing confidence that the isopleths were representative of the actual snow cover.

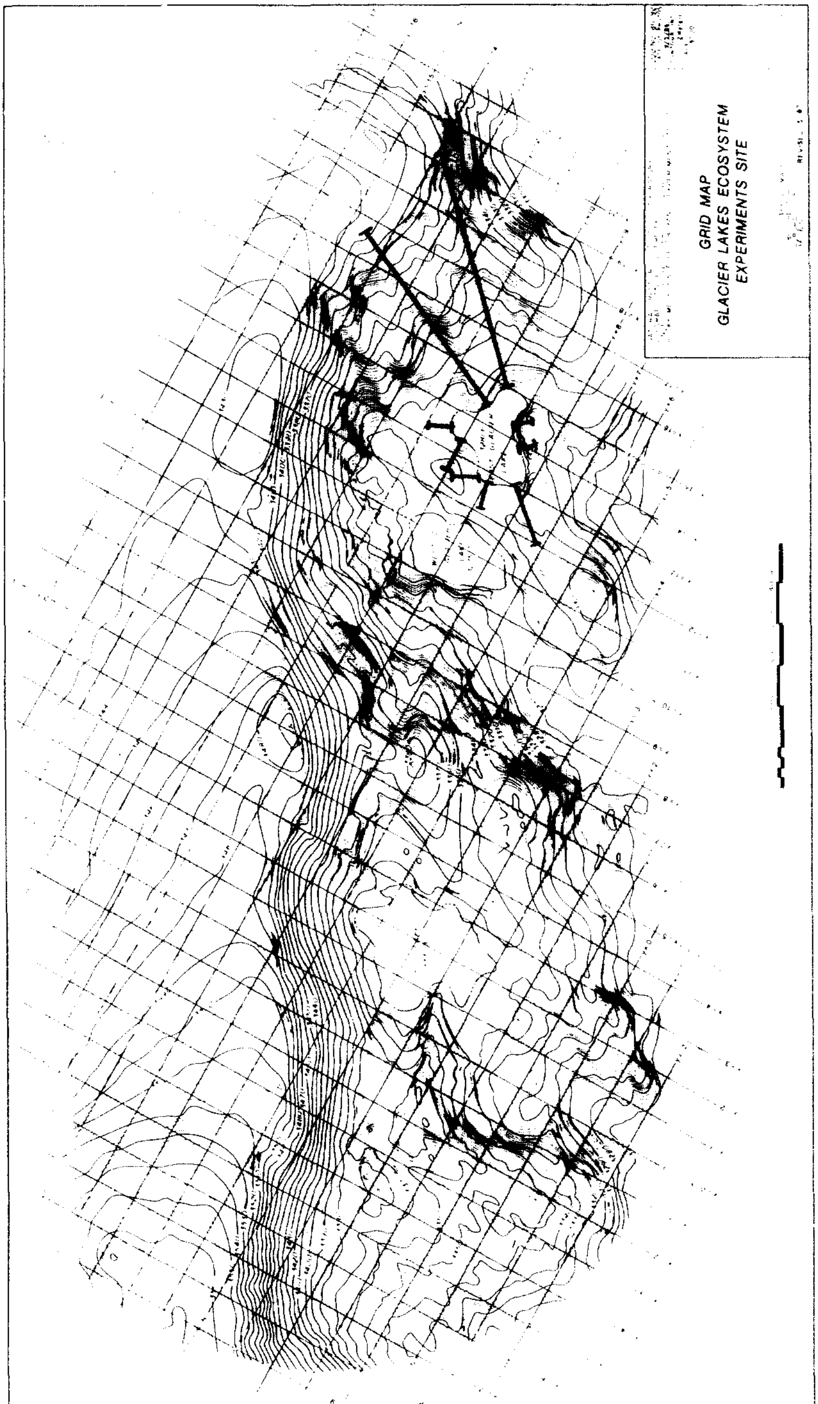

Fig. 3. GLEES map showing the grid and the East Glacier Lake snow core transects.

Fig. 4. Examples of tree damage by snow used in estimating long term average snow height.

Tree survey results

The average snow depth over the GLEES at all grid points that had trees associated with them was 200 cm. Using this average depth and our measured density of 401 kg m^{-3} near maximum accumulation gives a total average accumulation of 80 cm H_2O. The density was arrived at by averaging the 22 April pit (392 kg m^{-3}) and snow core densities (410 kg m^{-3}). We used the density of snow at maximum accumulation, rather than that estimated for the start of water flow at the base of the snowpack, because the tree survey gives an estimate of average yearly accumulation depth. The snow depth at the start of water flow was less than at maximum accumulation because of settling.

SNOWY RANGE OBSERVATORY PRECIPITATION GAGE

The Wyoming Water Research Center has maintained a Wyoming-shielded precipitation gage on the windswept knoll between East and West Glacier Lakes at the GLEES for a number of years as part of the Snowy Range Observatory (SRO). Data for November through April, 1979 – 1988 are shown in Table 2. For 1987–1988, the SRO gage measured 65 cm H_2O, 82% of the amount we determined using the Martinec–Rango method. Sixty-five cm H_2O is 89% of the average winter precipitation gage measurements for 1979–1988.

The Soil Conservation Service Brooklyn Lake snow course data for 1987–1988 showed an accumulation that was 89% of normal (T. Gilbert, SCS,

TABLE 2

Snowy Range Observatory precipitation date for November–April

Year	Precipitation (cm)
1979–1980	90.5
1980–1981	46.2
1981–1982	106.2
1982–1983	91.7
1983–1984	62.5
1984–1985	73.2
1985–1986	91.4
1986–1987	28.2
1987–1988	64.9
Average	72.8

personal communication, 1989). This site is < 2 km from the GLEES, and the result is consistent with the SRO precipitation gage measurements in that both data sets show that the 1987–1988 snow accumulation was 89% of the long-term average.

NADP PRECIPITATION COLLECTOR

An Alter-shielded precipitation gage is located at the Snowy Range NADP site at the GLEES. The site is a flat, open area ~ 15 m from the southeast shore of West Glacier Lake, downwind another 15 m from a grove of trees at the bottom of an east facing slope. The site is thus somewhat protected compared with the SRO site. Wind speeds are estimated to average 6 m s^{-1}, one-third less than the 9 m s^{-1} at the SRO site (Wooldridge et al., 1990). The NADP and Snowy Range Observatory sites are < 200 m from each other. Data from the NADP site indicate 82 cm H_2O from November 1987 through April 1988. No significant snow fell after April 1988.

SUBCATCHMENT SNOW CORE SURVEY

A non-random snow core survey (F.A. Vertucci, personal communication, 1989) was conducted in a small 31-ha subcatchment of the GLEES using several line transects taken from the top of the basin to the lake into which it flows. An attempt to compensate for the lack of randomness was performed as follows. Sample point information on snow depth, as determined along the sample transects, a series of aerial photographs taken during peak accumulation through melt, and a ground check during snow melt were used to derive a map of snow depths for the watershed. Iso-depths were drawn based on point depth measurements interpolated for regions of similar apparent depth as evidenced from the aerial photographs and direct inspection of subcatchment snow accumulation patterns. The area of the subcatchment represented by each 1-m interval depth class was computed. The density of snow, determined from the 61 cores, corrected for a systematic error (Work et al., 1965), and stratified into depth classes, was multiplied by the area of the subcatchment having that depth and density. These products were summed and divided by the total subcatchment area to provide an estimate of the cm H_2O on the subcatchment held in the snow-pack. The result from this method was 73 cm H_2O, and the 95% confidence range was 62–84 cm H_2O.

DISCUSSION OF ERRORS

Martinec–Rango degree days

Snow area recession curve

Visual comparison between the original and the black and white images of the aerial photos allowed the discrimination level to be chosen so that the snow was discriminated from the terrain to an estimated accuracy of better than $\pm 2\%$. This accuracy estimate was derived by taking values for the discrimination level that were at the limits of what would be acceptable and comparing the resultant areas. The tree-covered area was about 10% of the total as determined by setting a discrimination level that selected between the trees and the rocks and comparing the resultant areas. The settings of the discrimination levels in both these cases did not result in perfect discrimination and a considerable amount of judgement is involved in estimating the errors. No attempt was made to compensate for the trees that remained in the image after the discrimination level was chosen. However, they are sparse and short, so that much of the snow could be seen in the gaps. The snow in the trees has a lower density, so that it adds less to the total water. All snow was assigned the density measured in an open area. The error from assigning the snow next to the trees this higher density tends to compensate for not counting the snow under the trees. In addition, the scale of digitization was such that many of the trees did not cover an entire pixel. Often the presence of a tree decreased the pixel brightness but not below the discrimination level so that the pixel was counted as snow. We estimate that the error caused by ignoring the trees was smaller than -4%, again with the reservation that this error estimate is primarily based on judgment. Also note that this error is a systematic error tending to make the estimate too low. The only other source of systematic error that we have identified results from our choice of starting date. Some of the snow at higher elevations in the GLEES may have started to melt before our starting date. This systematic error would also tend to make the Martinec–Rango estimate too low.

Snow density

Errors in the density measurement translate directly into errors in the water equivalent estimate. Martinec and Rango (1986) give a range of average values of snow density for three alpine watersheds of about 345 kg m^{-3} on 1 April to

500 kg m^{-3} on 30 July with a range of $\pm$ 10% on each date. Bartos (1972) found that the major part of the variance in measured water equivalence is the result of variance in snow depth so that an adequate snow accumulation measurement only requires one density determination for every five depth determinations. The snow pit measurements and the corrected snow core measurements agree with each other and with the Martinec and Rango (1986) averages to $\sim$ $\pm$ 10%. We judge the true average snow density for the GLEES to be within the range of these estimates. The estimates we used for the Martinec–Rango calculations and for the tree survey estimate are also within these ranges. However, we recognize that a more careful density survey might contribute significantly to the accuracy of accumulation estimates.

Tree morphology

The Wooldridge et al. (1990) method is likely to be subject to systematic errors that could result in a low estimate. It is physically impossible for this method to overestimate the snow depth locally because it indicates years when the maximum annual snow depth is low. It is usual for a snowpack to attain more than 90% of its depth by the end of the first third of the winter. Additional snow then causes enough settling so that the depth changes only a little for the next two-thirds of the winter. We believe that over time, the tree damage would occur at the minimum snow depth at each site, keeping in mind that some snow-free branches may survive some mild-winter, low snow-depth years. The tree morphology estimate of average depth is 94% of the Martinec–Rango estimate for 1987–1988. The Soil Conservation Service snow course data and the Snowy Range Observatory data indicate that the 1987–1988 accumulation was 89% of normal. The tree morphology estimate is likely to indicate low snow years, since foliage above this height will be killed during the winter. Therefore, the tree morphology may be a more accurate indicator of an average low snow year than of an average snow year. These various results are consistent if the tree morphology method gives an average that is about 17% low (6% to agree with the M–R estimate, and 11% to compensate for this particular year). We feel safe in concluding that the Wooldridge et al. (1990) method gives a lower bound to the average snow depth.

SRO precipitation gage

Precipitation gages are very prone to undermeasure snow accumulation in windy regions (e.g. Martinec, 1986; Sturges, 1986). The SRO precipitation gage is a point measurement at the GLEES and is located on a windswept ridge which accumulates little snow. The precipitation gage measurement at this site for 1987–1988 is 65 cm H_2O, which is 19% lower than the average yearly accumulation determined from tree morphology. As discussed above, the tree morphology indicates the depth for low snowfall years such as 1987–1988. The location of this collector at a site with less accumulation, combined with the inefficiencies of the Wyoming shield in cold, windy conditions (Sturges, 1986), leads us to believe that this measurement underestimates snow accumulation at the GLEES. The SRO average collection is 72.8 cm H_2O. In comparison, the tree survey average, which is thought to be low, gives about 80 cm H_2O. The SRO precipitation gage measurement likely forms a lower bound of possible water equivalent values for 1987–1988.

NADP collector

As a result of the inherent collection inefficiencies of rain gages, this estimated water input to the watershed is considered low. However, research has shown that the Wyoming shield used at the SRO site is considerably less effective in collection efficiency than is the Alter shield used at the NADP site (Sturges, 1986). Thus, the underestimate is likely to be less than that for the SRO collector. Both the NADP collector and the SRO precipitation gage measurements are point measurements and may not be representative of the whole catchment. The snow depth isopleths derived from the tree morphology survey indicate that the SRO gage is in an area of abnormally low deposition while the deposition in the area of the NADP collector is about average. While both gages are above the major part of the drifting snow, the aerodynamic effects of the topography on the efficiency of either gage are not known.

Snow core survey

The snow survey was designed as a systematic survey, with all transects running from the top to the bottom of the watershed. As previously discussed, an attempt was made to adjust for this non-randomness. In addition, this small subcatchment may not be representative of the entire GLEES watershed. A low bias in the snow core survey is indicated by the Wooldridge et al. (1990) isopleths which show higher than average accumulation in the subcatchment while the core survey result is the second lowest for 1987–1988.

CONCLUSIONS

Table 3 shows the results of the different estimates of snow water equivalent input to the GLEES in winter 1987–1988. As discussed above, the Wooldridge et al. estimate is lower than the long term average but not as low for 1987–1988.

From these different estimates we conclude that the Martinec–Rango analysis gives a reasonable value, 85 cm H_2O, for the snow water equivalent input to the GLEES in winter 1987–1988. We are confident that the accuracy is within $\pm 20\%$, a range of 68–102 cm H_2O. Although the Martinec–Rango estimate is higher than other estimates, the biases inherent in the other measures indicate that they all likely underestimate actual snow accumulation. The only systematic errors we have been able to identify in the Martinec–Rango method would also tend to make this estimate low. However, the errors

TABLE III

Estimates of snow water input to the GLEES, 1987–1988

Method	Water equivalent ($cm\ H_2O$)	Bias[1]
Martinec and Rango (1986)	85	Unknown
Wooldridge et al. (1990)	80	Low average
SRO precipitation gage	65	Very low
NADP precipitation gage	82	Low
Subcatchement core survey	73	Low

[1]See text.

involved in lumping the physical processes that cause snow melt into the simple degree-day factor, α, and the direction of the errors involved in estimating the snow densities are not known. Thus, it is impossible to say whether the Martinec–Rango method is likely to be high or low.

The other estimators limit the possible range above 85 cm H_2O. For example, if the true value is 94 cm H_2O (+10%), that would imply that the Wooldridge et al. (1990) estimate is 26% (15 + 11) too low and the SRO precipitation gage is 31% low. If the true value is 102 cm H_2O (+20%), the tree survey estimate average would be 33% low and the SRO data would average 36% low, which would be near an extreme of each method's possible range. We conclude from the discussion of errors above that the accuracy is more realistically estimated as $\pm$10%, or a range of 77–94 cm H_2O. In general, we think this method of constraining the possible ranges of snow accumulation measurements by using several estimation methods can be applied to improve the reliability of such estimates.

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