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Generation of the Snowline

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Abstract: *A method of determination of daily snowline altitude in mountainous terrain, which is applicable in areas where the snowline may be assumed to follow a contour, is presented. The method is based on a combined use of Landsat images and hydrometeorological data.*

A possible application area includes the generation of streamflow in snowfed rivers where the model needs data regarding daily snowcovered area. However, the method will prove useful only in models that assume the area above the snowline to be completely covered with snow, e.g., the Swedish HBV model. The methodology has been developed for a Himalayan basin with considerable rainfall during snowmelt. It is hoped that it is also applicable to other mountainous basins.

Introduction

An important problem in satellite snow hydrology is the mapping of snowcovered areas at a spatial and temporal scale appropriate to hydrologic analysis. Normally, the satellites with suitable temporal resolution (GOES, NOAA TIROS-N series) have too coarse a spatial resolution, while Landsat has fine spatial resolution but too long an interval between overpasses. This paper attempts to overcome the problem of long time intervals by generating the snowline by combining calculations for snow accumulation and snow ablation as functions of elevation.

The methodology has been developed for a Himalayan basin where the meteorological data are collected at the lowest point only, whereas the elevation range itself is considerable. This may be considered to be a typical basin in this part of the world.

Calculation of snow accumulation in such a basin requires development of techniques to estimate

- the form of precipitation and
- the increase (or decrease) of precipitation with altitude.

Similarly, snowmelt also has to be calculated from data as available at the lowest elevation. In this scheme Landsat images find application in determining the pattern of variation of precipitation with altitude. Also, snowline elevation measured on the available Landsat image serves as a check on the snowline elevation figures generated from meteorological data.

Experimental Basin

The Beas basin above Manali, situated in the western Himalayas, has been chosen as the experimental basin (Figure 1). The basin extends from 1900 m (elevation of Manali) to 6000 m and covers an area of 345 sq km. During winter almost the entire basin is covered with snow, while by September about 95 percent of the basin has become bare. The permanent snowline is at an altitude of about 5000 m, above which there is perennial snow. Only 3 percent of the basin lies above that permanent snowline; glaciation is therefore insignificant. Vegetation is fairly dense at Manali and extends upward to about 2700 m where it merges with shrubs. Beyond 3000 m the basin is almost without any vegetation.

The following hydrometeorological data are collected daily

at Manali:
daily rainfall,
daily snowfall (water equivalent),
daily maximum temperature T_{max}°C,

0099-1112/83/4912-1679$02.25/0

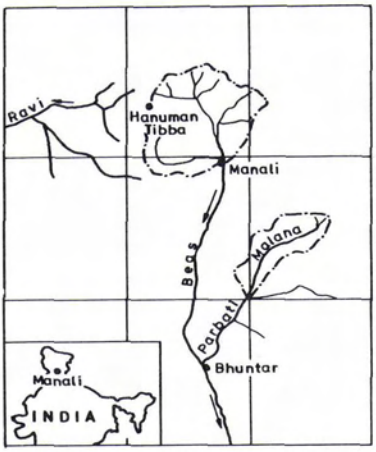

FIG. 1. The Beas River basin.

daily minimum temperature T_{min}°C, and
discharge in the Beas River.

These data, available since 1970, are collected by government agencies routinely and must not be taken as snow-laboratory data. Discharge is measured once a day at 11 AM. A continuous record of river flow, rain/snowfall, or temperature is not collected. There is no other data collection point, and also there is no snow course in the basin or anywhere in this part of the Himalayas. Information on snowcover has been available from Landsat images since 1972, though the number of usable images is very small.

The basin is divided into 20 elevation zones (j), each 200m high. The area of each zone (ΔA_j) is planimetered from the topographic map of that area at a scale of 1:50,000 with a contour interval of 40 m. The total area then is

$$\sum_{j=1}^{20} \Delta A_j = A = 345 \text{ sq km.}$$

SNOWLINE

The transient snowline appears at an elevation where the seasonal snowmelt equals the accumulated snow. At the latitude of the Himalayas there would not be any significant snowfall but for the elevation. The elevation effect is pronounced. With the advent of winter the snowline gradually descends along the slope, and ablation during summer results in upward movement of the snowline. In higher latitudes the general trend of snowline movement is horizontal, from higher to lower latitudes in winter and vice verse in summer. In the Alps perhaps both influences operate with equal significance. These general remarks are made to suggest that in the Himalayas the snowline may be expected to follow elevation contours whereas in the Alps the elevation effect may not be clear. There are places, as in Norway for example, where the concept of a transient snowline is not apparent (Andersen, 1982).

There are other effects also. Both snowfall and snowmelt are influenced by wind, vegetation, slope, and aspect of the basin. Thus, even if the elevation effect were predominant in the Himalayas, making the snowline follow elevation contours, there are likely to be many local departures because of these effects.

DETERMINATION OF SNOWLINE ALTITUDE FROM LANDSAT IMAGES

Snowline altitude can be determined from a Landsat image by projecting it to a topographic map. In this particular experiment a stereoplotting instrument, a Wild Autograph A8, was available and was used monocularly for this purpose. A Landsat positive transparancy, band 5, at a scale of 1:1,000,000 was used. The available magnification ratio was 1:10 whereas the topographic map was at a scale of 1:50,000. The map was therefore reduced to 1:100,000 scale. Planimetric registration was performed by matching recognizable details, invariably streams. The instrument reference mark was set on the snowline, and the elevation of the snowline at that point was then read from the map contours. One point per kilometre on the snowline was read wherever the snowline was clear enough. The mean of the elevations of all the points thus read was taken as the elevation of the snowline. A few points are worth mentioning.

- One of the advantages of the assumption that the snowline follows a contour line is that a partially cloud covered image can be used to determine the snowline altitude. Recognition of snow from cloud has not been found to be difficult. At the magnification used, the texture of snow can be differentiated from the "lumpiness" of cloud.
- Matching of the image to the map presents some problem. In a field of view where there is no relief effect (since because the image was viewed monoscopically), recognition of streams was somewhat difficult. It was therefore decided to repeat the observations two or three times, each time with independent settings.
- In a highly accidented terrain shadows present a major problem, and it was not always possible to discern snowline inside dark shadows. The problem is more acute in winter because of the low sun angle.
- In a precision instrument such as the wild A8, the planimetric inaccuracy of the Landsat hardcopy image shows up as a mis-match with the map details. Some compromise setting had to be done.

Table 1 shows some of the snowline elevations as read and also estimates of the standard deviation. An elementary statistical check was applied. The distribution of departures from the sample mean satisfies the requirement of a normal distribution. This is however certainly not a proof that the de-

TABLE 1. SNOW ALTITUDE MEASUREMENTS FROM LANDSAT IMAGES OF THE BEAS BASIN ABOVE MANALI

Date	Measurement No.	No. of points read	Mean snowline altitude (m)	Estimate of standard deviation, s (m)	Estimate of standard deviation of the mean, $\bar{s}$ (m)
1 April 76	1	14	2270	164	44
	2	9	2333	178	59
19 April 76	1	22	2569	142	30
	2	28	2655	239	45
9 March 77	1	21	2790	284	62
	2	27	3198	238	46
	3	25	3168	259	52
27 March 77	1	36	3054	205	34
	2	46	3246	273	33
9 April 78	1	17	2754	237	57
	2	11	2664	276	83
27 April 78	1	34	2979	218	37
	2	31	3056	254	46
	3	22	3109	274	58
15 May 78	1	33	3508	270	47
	2	41	3311	220	34
	3	28	3387	263	50
14 March 76	1	15	2240	175	45
	2	9	2169	107	36

partures of the snowline from the elevation contour are normally distributed, though it is believed to be so.

The daily snowline elevation can be calculated by linear interpolation between the values obtained above. Figures 2 (a), 2(b), 2(c), and 2(d) were constructed by plotting Landsat derived data (and one or two ground observed values to fill critical gaps). The distribution of points in Figure 2 is far from satisfactory and brings out the problem of insufficiency of the number of usable images. A great many images were lost because of cloud cover. The problem is more acute in the Himalayas because of the absence of a Landsat ground station in this part

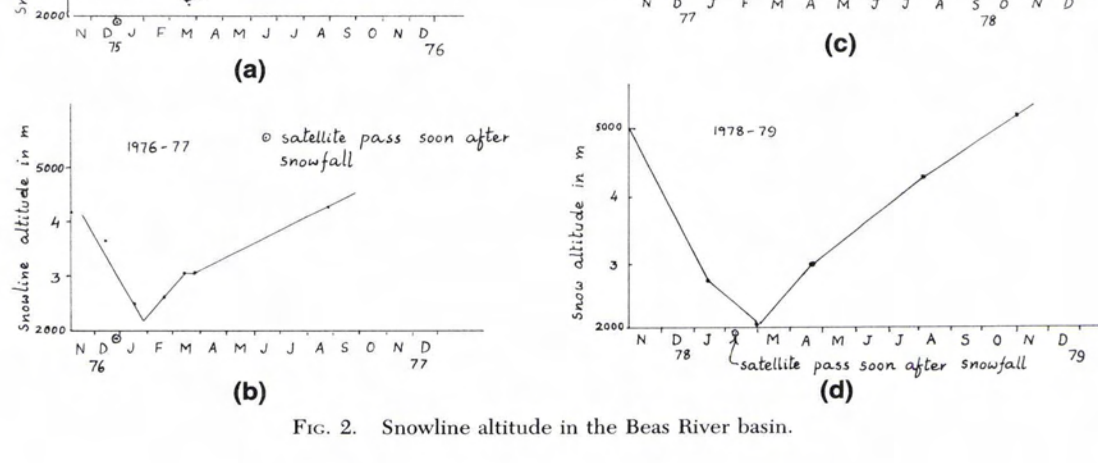

FIG. 2. Snowline altitude in the Beas River basin.

of the globe (one has since come up in 1981). The data on board the satellite had to be stored on magnetic tape, which has a capacity of half-an-hour's data. Because of the limited capacity, and also due to the anxiety of the administration to prolong the tape recorder's useful life, the sensors were switched on over the Indian territory only infrequently. This further reduced the number of available images over the Himalayas.

If there were a satellite pass immediately after a snowfall, then normally it would be of no use because the snowfall occurs in too widespread an area, and it melts within a couple of days (Garstka *et al.*, 1958). Images taken on 27 December 1976 and 9 February 1979 are examples of wasted data because of fresh snowfall immediately before the Landsat pass. Records show that there were 14 cm snowfall (14 mm water equivalent) at Manali on 26 December 1976 and 5 cm on 8 February 1979, making snow cover far in excess of that which can go in the streamflow model.

Generation of the Snowline

The linear interpolation between two consecutive Landsat passes separated by a long interval, at times months, is without logical basis and is only a dictate of practical consideration. Similarly, a model which accepts daily snowcovered area as an input should not reject any image simply because the overpass is immediately after snowfall. A method was therefore developed to generate the daily snowline altitude from meteorological data. As stated earlier, Landsat images find application in determining precipitation variation characteristics and also serve as checks on the generated snowline. Thus, once the orographic variation of precipitation with altitude is established for a basin, it is possible to generate the snowline from meteorological data only. One self-imposed constraint in this research is that the methodology should not call for any data that are not normally available in the Himalayan region.

Snow Accumulation

Snow accumulation at higher altitudes can be calculated from data recorded at the lowest point, provided a method can be found to transfer precipitation data vertically, in form and quantity.

The form of precipitation may be taken to be directly related to the ambient temperature (Murray, 1952). To determine the form of precipitation from ambient temperature one would need to have a continual record of temperature and the time of occurrence of precipitation. None of them are available in an average Himalayan basin; hence, the following method has been developed, based on analyzing the records of precipitation and daily minimum temperature T_{min} available at Manali (Bagchi *et al.*, 1981).

It has been found that in general the entire day's precipitation is in the form of rain if $T_{min} \geqslant 3.5°C$ and it is completely in the form of snow if $T_{min} \leqslant -7.5°C$. In between these extremes the following equation gives the percentage of snowfall (x) in the day's precipitation, 97 percent of the time within ± 15 percent:

$$x = 9(3.5 - T_{min}) \quad (1)$$

To estimate the percentage of snowfall (x_{ij}) at any higher altitude zone (j) in the i^{th} day's total precipitation, one can apply the above equation, calculating $(T_{min})_{ij}$ from the record of the base station by means of a suitable lapse rate. The implied assumption is that Equation 1 would hold good at that altitude. Studies reported by the U.S. Army Corps of Engineers (1965) suggest that this indeed may be the case, but caution must be exercised because the experiments were limited to a rather small elevation range. Another point to note is that the equation above has been developed from point data whereas it would be used subsequently to calculate an areal value. This is indeed a conceptual weakness.

In the absence of an actual study of the lapse rate in the Himalayas, it has been taken as 0.65°C per 100 m(Von Nostrand, no date) throughout. It is worth mentioning that Rango and Martinec (1979) accepted a seasonally variable lapse rate of 0.8 to 0.95°C per 100 m. Bergstrom (1976) used a different lapse rate in different basins while applying the Swedish HBV model, e.g., 0.5°C/100 m for the Kultsjom, Malgomaj, and Strom Vattudal basins and 0.6°C/100 m for the Filefjell basin.

Orographic Increase in Precipitation

To calculate the depth of zonal snowfall (as opposed to percentage of snowfall), a method for the transfer of precipitation data vertically has been developed (Bagchi, 1982). It is known that precipitation generally increases with altitude on the windward side of the mountain slope (Fairbridge, date?). In the Himalayas one must take into consideration this orographic increase because it is considerable, areal precipitation being 2.5 times the point precipitation recorded at Manali, as found out by correlating annual runoff in the Beas River to the annual precipitation, both at Manali. Point precipitation in mountainous regions varies considerably, being influenced by a great many meteorological and topographical factors. Determination of the pattern of variation of precipitation, and finding an answer to such questions as whether precipitation variation is correlated to altitude by making measurements at different points located at various altitudes, is therefore likely to be a frustrating exercise unless possibly one takes prolonged measurements at a great many points. The concept of collecting ground data at a large number of points has no relevance to the Himalayan region, soaring to as it does 6000-m altitude. Precipitation-runoff models actually require areal measures of precipitation, and in this context the important point is whether there is at all a stable areal value. Evidence suggests that there is indeed a stable relation between point value at the base and basin areal value if we use a suitable time scale. Table 2 gives the values of $\bar{\beta}$ (a dimensionless quantity defined as basin areal precipitation minus evapotranspiration divided by point precipitation at the base station Manali) calculated over several years. The moving average figures are stable enough. $\bar{\beta}$ *is a value lumped over the whole basin.* To determine the pattern of variation of precipitation with altitude, it will be assumed that zonal precipitation is correlated with altitude.

We define β_j as areal precipitation minus evapotranspiration in zone j divided by point precipitation at the base station (Manali, $j = 1$, $\beta_j = 1$). Then on the i^{th} day, depth of snowfall in zone j is $(x_{ij}\, \beta_j\, P_{i1})/100$ where P_{i1} is the precipitation in zone 1. Now from Figure 2 it is possible to calculate the number of days (n) during which a particular zone remained under snowcover. Thus, the altitude zone between 3700 m and 3900 m ($j = 10$) remained snow covered during the period 9 December 1978 to 16 June 1979.

Thus, the total effective snowfall (snowfall minus evapotranspiration) in zone j during the n days is equal to

$$\frac{1}{100} \sum_{i=1}^{n} x_{ij}\, \beta_j\, P_{i1} = \frac{1}{100}\, \beta_j \sum_{i=1}^{n} x_{ij}\, P_{i1} \qquad (2)$$

As mentioned earlier, the point values of β_j in any altitude zone are likely to show temporal and spatial variations. β_j has, however, been defined as a lumped zonal entity, and shifting it outside the summation implies that it is assumed to be invariant over time. This is indeed an untested assumption and is likely to remain so for quite sometime because of logistic difficulties in experimental verification in the high Himalayas.

The total effective snowfall is equal to the sumtotal of snowmelt during the period. Snowmelt M_{ij} in the j^{th} zone on the i^{th} day can be estimated by the temperature index method, which gives

$$M_{ij} = a(T_{max})_{ij} \qquad (3)$$

where $(T_{max})_{ij}$ is the daily maximum temperature in the j^{th} zone on the i^{th} day determined from $(T_{max})_{i1}$, measured at Manali, by means of the adopted lapse rate, and a is the degree-day factor defined as the melt in mm during 24 hours per degree by which T_{max} exceeds 0°C. The literature is replete with discussions about the limitations of this method, and also there are many variants of the temperature index method. That the method is inadequate to give the melt at a point on a given day is very clear from the experiments conducted by various authors. Martinec (1960) found the values of a varying as much as 0.8 to 15.8 mm per °C day. Similarly, the U.S. Army Corps of Engineers (1956) found considerable variation in point values. Similar doubts regarding the validity of the temperature index method to monitor snow ablation at a point has also been raised by Anderson (1976). In spite of these, the temperature index method continues to be used extensively (Martinec, 1975; Bergstrom, 1976; Rango and Martinec, 1979; Hawley *et al.*, 1980; Shafer *et al.*, 1982), and obviously the method does work adequately and reasonably. The reason is that the areal value of degree-day factor is quite a different entity from the point value. There appears to be a reasonably stable value of the degree-day factor lumped areally and temporally (Bagchi, 1983).

The total melt during the period during which the zone j was under snow is given by

$$\sum_{i=1}^{n} M_{ij} = a \sum_{i=1}^{n} (T_{max})_{ij} \qquad (4)$$

Equating the total effective snowfall to the total snowmelt,

$$\beta_j = \frac{a \sum_{i=1}^{n} (T_{max})_{ij}}{\frac{1}{100} \sum_{i=1}^{n} x_{ij}\, P_{i1}} \qquad (5)$$

The areal value of the degree-day factor has been calculated as 2.1 mm per °C per day (Bagchi, 1983). The degree-day factor so obtained gives the amount of runoff generated per °C day where evapotranspiration has already been taken care of.

It is now possible to calculate the values of β_j from the meteorological data recorded at Manali and the adopted lapse rate, the number of days of summation being obtained from Figure 2. These values, as obtained from 1978–1979 data, are plotted in Figure 3. β_j can be determined in this way for zones which remain under snowcover for a sufficiently long period of time. Thus, β_j for the lower few zones had to be determined by interpolation. Similarly,

TABLE 2. VALUES OF $\bar{\beta}$

Period considered (day/month/year)	$\bar{\beta}$	Moving average $\bar{\beta}$	Average $\bar{\beta}$	Estimate of standard deviation
10/01/70 to 9/01/71	2.13	2.13		
10/01/71 to 9/01/72	2.02	2.07		
10/01/72 to 9/11/72	2.24	2.13		
10/11/72 to 19/11/73	2.93	2.30		
20/11/73 to 19/11/74	3.52	2.57	2.5	0.5
20/11/74 to 30/12/75	2.77	2.60		
31/12/75 to 19/11/76	2.13	2.53		
20/11/76 to 19/11/77	1.97	2.46		
20/11/77 to 19/10/78	2.82	2.50		
20/10/78 to 31/10/79	2.32	2.49		

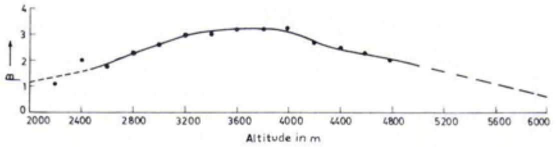

FIG. 3. Variation of β with altitude.

for the zones above the permanent snowline area, β_j has been calculated by extrapolation.

VERIFICATION

Actual field measurement of areal precipitation at different altitudes for the determination of β_j is quite impossible in the Himalayas at present; hence, some indirect verification may be resorted to. The average over the basin of the precipitation (minus evapotranspiration) factor $\overline{\beta}$ is now given by

$$\overline{\beta} = \frac{1}{A} \sum_{j=1}^{20} \beta_j \, \Delta A_j \qquad (6)$$

which calculates to 2.6. This compares favorably with the average value in Table 2.

It is now possible to calculate runoff that originates at any altitude zone from snowmelt and rainfall on any particular day by using the β factor and lapse rate, and meteorological data recorded at the base station. These runoff figures may in turn be used as inputs for generating discharge or runoff in the stream. Any error in the β_j values will affect the generated runoff in the stream, but the effect will be more pronounced in the rainy season (July to September). Discharge in the Beas River has been calculated for 90 days starting from the 180th day of the years 1976 to 1979 using a model which may be taken as a modified version of Martinec Rango model (Rango and Martinec, 1979). The generation is based on the discharge data of the 180th day and using the same β values based on 1978–1979 (Figure 3). The four mass curves appear in Figure 4. The agreement between the observed and generated mass curves looks satisfactory and may be taken as a second verification on the concepts of the β factor and the calculated values of the same. It should be mentioned here that in generating the mass curves an implied assumption is made that evapotranspiration is the same for rain or snow. This indeed may not be true, but such an assumption has still been made in the absence of precise knowledge in this regard.

SNOW DEPTH

Keeping an account of daily snowfall and snowmelt, it is now possible to calculate the depth of

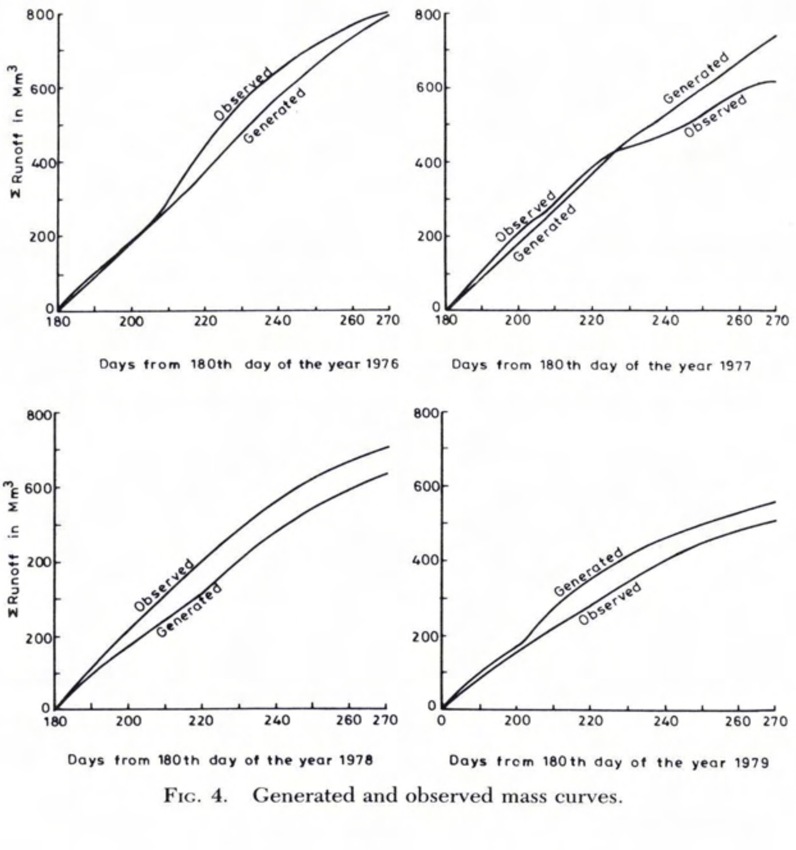

FIG. 4. Generated and observed mass curves.

standing snow (water equivalent) at any altitude zone on any day. The depths so obtained are the areal average values which would ultimately become runoff from the particular zone. Snow depths cannot be calculated by this method beyond the permanent snowline because of downward movement of the accumulated snow in the form of glaciers.

The calculated snow depths for 1977, 1978, and 1979 appear in Figure 5. These could not be verified by actual measurement. The only possible method of determination of areal average snowdepth, by natural of gamma ray technique, is inapplicable in the Himalayas because of the requirement for low flying (less than 200 m (Odegaard and Ostrem, 1977)) and because the response reaches the saturation value at about 0.6 m of standing snow (Tom Andersen, personal communication).

The standing snowdepth data can be utilized to generate future summer streamflow (snowmelt component only). This was done for 1978 and 1979, on 1 April of both years. The Martinec-Rango model was used; to calculate snowmelt it was necessary to forecast the monthly mean summer T_{max}. The forecast of runoff obviously did not include any contribution from precipitation subsequent to 1 April. The mass curves as forecast and as observed (minus the contribution subsequent to 1 April) appear in Figure 6. The agreement is fair and may be taken as a proof that the snowdepths as generated are reasonably correct.

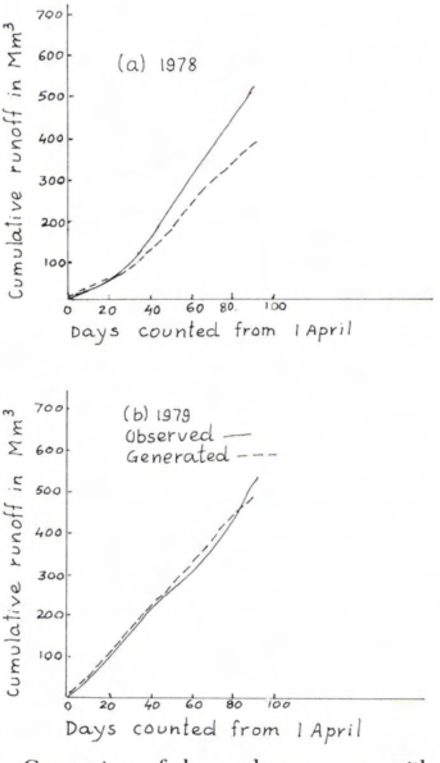

FIG. 6. Comparison of observed mass curves with those predicted on 1 April.

Snowline Generation

It is now easy to calculate the altitude at which the standing snow vanishes in depth. This is precisely the transient snowline altitude. A computer program was written to generate daily snowline altitudes. Figure 7 shows the generated figures for the years 1970–1971 to 1978–1979. The snowline altitudes, as read from Landsat images and a few ground observed points, appear in the figures. It can be seen that these data can be used as a check on the daily snowline altitudes generated using meteorological data by the technique described in this paper. Further, it is now possible to generate snowline altitudes without any Landsat images. The use of Landsat images in these cases remain restricted to the determination of β_j values from the year's data for which Landsat images are available. One assumption is that the β values remain fixed over the years. β values have since been calculated using 1975–1976 to 1978–1979 data and they satisfy the requirements of the χ^2 test. The daily discharges have been generated continuously since 1971 using the same value of β as calculated for the 1978–1979 data, and the generated data agree reasonably with the observed hydrograph. Nothing more about the fixed nature of the β values can be stated at the present moment.

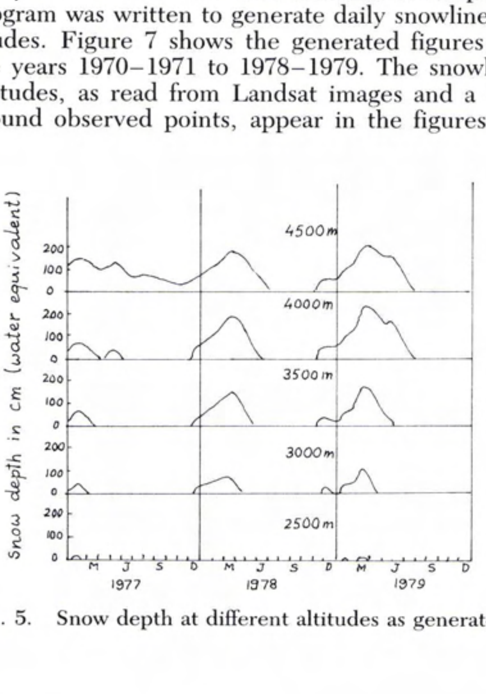

FIG. 5. Snow depth at different altitudes as generated.

Application of the Generated Snowline

The generated daily snowline altitude allows us to generate daily hypographs under the assumption that the area above the snowline is completely covered with snow. The generated hydrographs and mass curves based on the meteorologically generated snowline for the year 1979 appear in Figures 8 and 9. The same data, using satellite image-derived snowline altitudes, appear in Figures 10 and

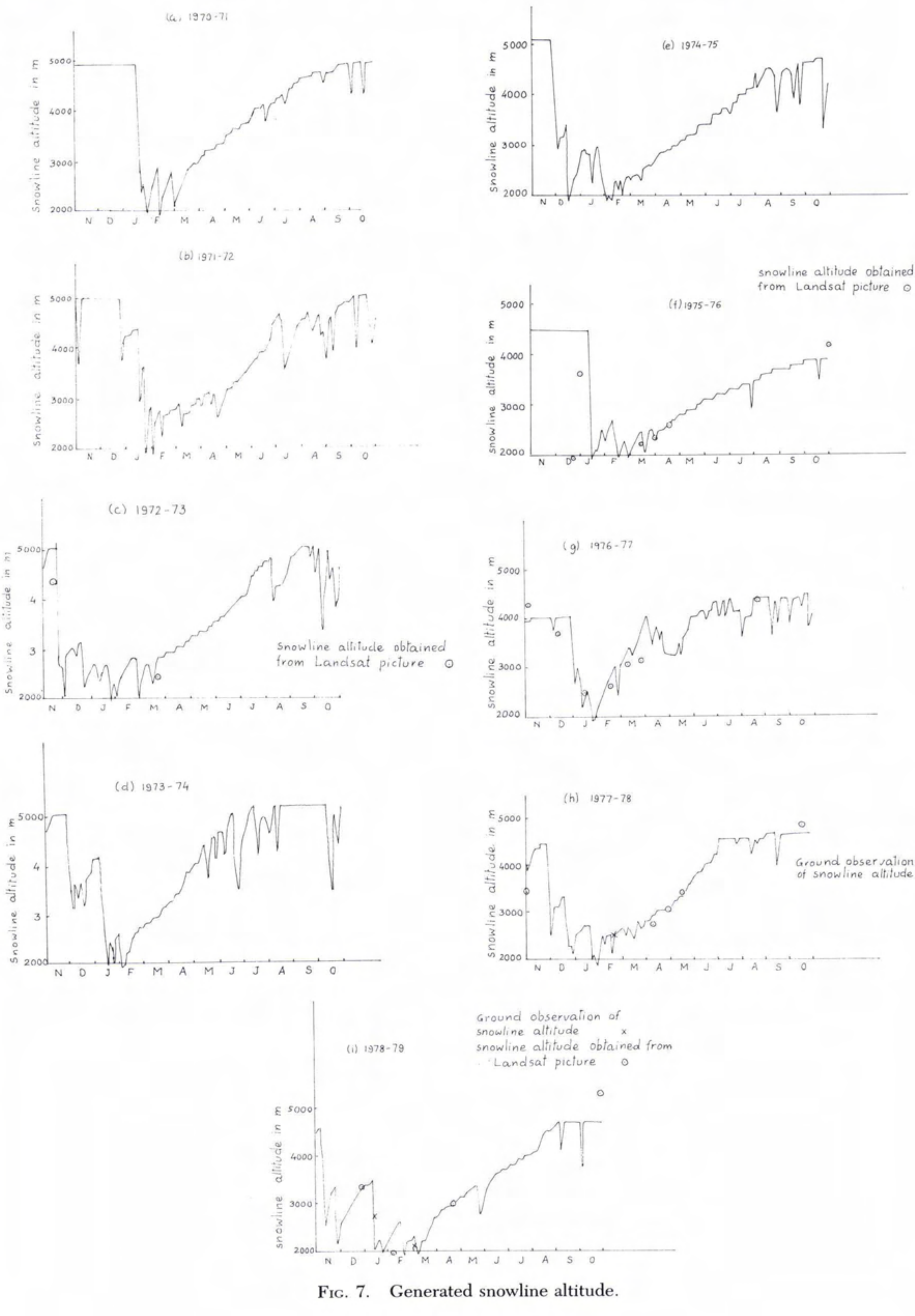

FIG. 7. Generated snowline altitude.

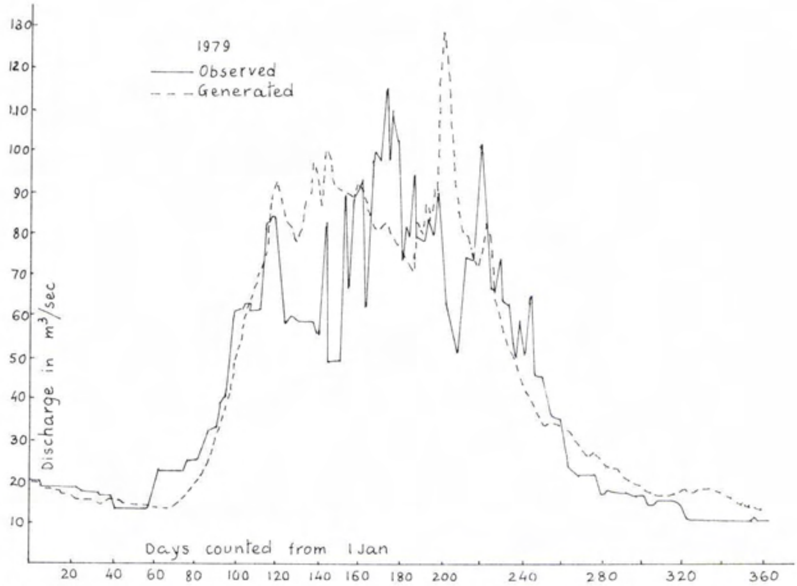

FIG. 8. Comparison of observed hydrograph with generated hydrograph (based on the generated snowline).

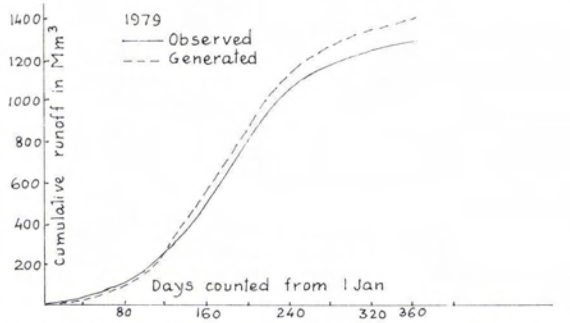

FIG. 9. Comparison of observed mass curve with the generated mass curve (based on the generated snowline).

11. The differences between the two sets do not clearly point to the superiority of one over the other. Table 3 gives a comparative study of the efficiency of generation of the hydrograph, using the Nash-Sutcliff criterion (Nash and Sutcliff, 1970), for the two methods of determination of snow covered areas. Again the difference between the two is not apparent. Figures 12 and 13 give the hydrograph and mass curves of 1971 obtained from generated snowline altitudes. This happened to be the year when there was no Landsat image available in the basin.

There may be other applications of the daily snowline altitude figures, perhaps for studies of high

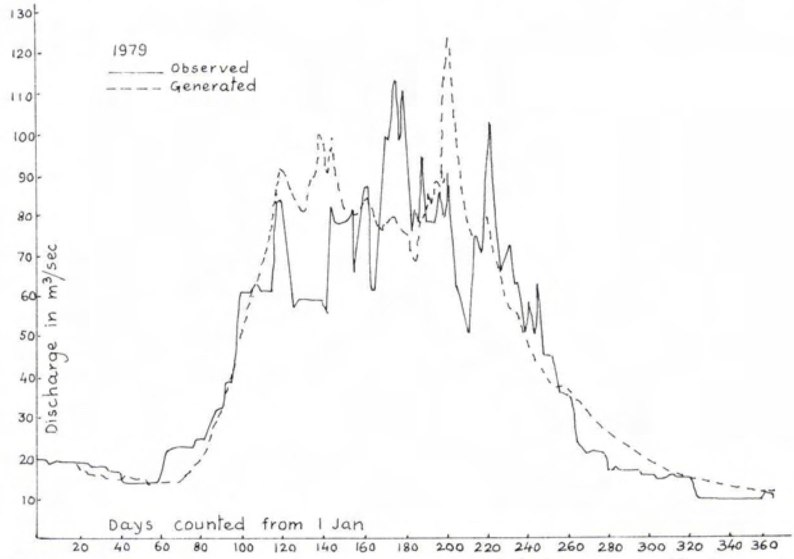

FIG. 10. Comparison of observed hydrograph with the generated hydrograph (based on the snowline observed on Landsat imagery).

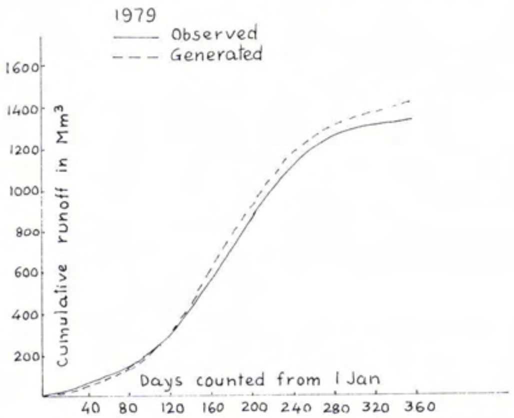

FIG. 11. Comparison of observed mass curve with the generated mass curve (based on the snowline observed on Landsat imagery).

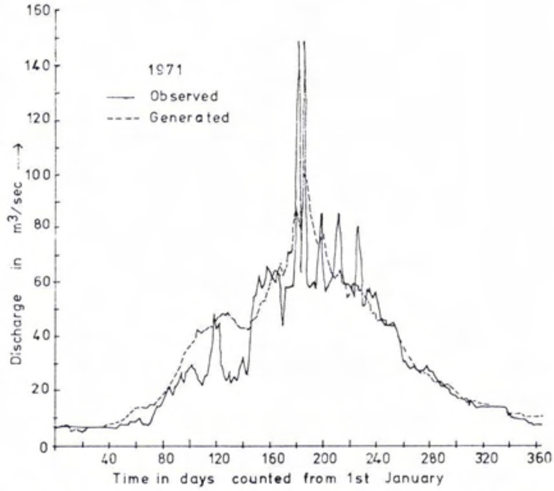

FIG. 12. Comparison between the observed and generated hydrograph (based on the generated snowline).

mountain climate phenomena. The snowline altitude as generated may in that case be used to interpolate between those measured on Landsat images. Figure 14 gives the result of fitting of those figures in between the observed data.

CONCLUSION

A method has been presented to generate snowline altitudes from meteorological data observed at the base station. These generated snowlines can be used to simulate the hydrograph, thus solving the problem of long intervals between satellite overpasses. The methodology has worked in the experimental basin, and it is anticipated that it will work in other mountain basins where the snowline follows elevation contours.

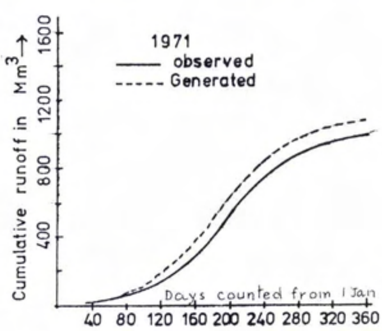

FIG. 13. Comparison of observed mass curve with the generated mass curve (based on the generated snowline).

TABLE 3. RESULTS OF STREAMFLOW GENERATION

	1976		1977		1978		1979	
	1	2	1	2	1	2	1	2
R^2	0.752	0.764	0.434	0.415	0.775	0.690	0.771	0.702
Cumulative error as percentage of observed runoff	+12	+8	+17	+12	−13	−15	+6	+10

1. Based on Landsat derived snowcover data.
2. Based on snowline altitude generated from meteorological data observed at Manali.
In both cases the entire area above the snowline has been taken as snow covered.

$$R^2 = \frac{\frac{1}{n}\sum_{i=1}^{n} (Q_i - \overline{Q})^2 - \frac{1}{n}\sum_{i=1}^{n} (Q_i - Q_i')^2}{\frac{1}{n}\sum_{i=1}^{n} (Q_i - \overline{Q})^2}$$

where
R^2 = measure of model efficiency
Q_i = observed discharge
Q_i' = generated discharge
$\overline{Q}$ = mean of observed discharge
n = number of discharge values

(Nash and Sutchliff, 1970)

FIG. 14. Adjusted transient snowline altitude.

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