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STATUS REPORT ON SNOWMELT MODELLING STUDIES

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PREFACE

Very limited research work has been carried out in India in the field of snow and glacier hydrology. Realising the importance of such studies in our country, a review has been made to assess the status of the research work in this field with particular emphasis on snow melt modelling studies. All types of models have been reviewed and presented in this report. This report also includes salient features of some important snow melt runoff forecasting models such as SSAAR, SRM, UBC, which are being used in other countries. The snowmelt process and different approaches, energy budget approach, degree-day approach are also discussed in this report.

This status report has been prepared by Dr Pratap Singh, Scientist B Western Himalayan Regional Centre, National Institute of Hydrology, Jammu, before undergoing for training under UNDP Project (IND/90/003-A/01/99) in the field of snowmelt modelling studies.

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ABSTRACT

A good percentage of annual run-off of rivers in the Northern region of India, originating from Himalayas is derived from snow and glaciers. Reliable estimates of volume of water contained in snowpack and its rate of release are very much required for efficient management of water resources which includes flood forecasting, reservoir operation and design of hydraulic structures etc. Investigations to understand the snowmelt processes and snowmelt forecasting techniques are required for proper utilization of abundant water resources available in the Himalayan region.

The status of studies carried out in the Indian Himalayan basins indicates that studies on snowmelt forecasting are very limited. Efforts are being made for seasonal snowmelt forecasting in Satluj catchment using regression technique based on remote sensing information on snowcover. Some investigations on snowmelt processes and snowmelt simulations and that too in one or two basins have been made.

Recognizing the need for proper planning and management of water resources ,it is found imperative to develop simple snowmelt run-off forecasting model based on generally available data of precipitation, temperature, and snowcover area for Himalayan snowbound catchments. The snowmelt models may be evolved either based on regression approach or simulation approach. The existing snowmelt simulation models based on degree-day approach such as SSARR, UBC and SRM etc. have been developed and are being used for operational purposes such as

flood forecasting ,reservoir operation and hydroelectric production,irrigation and water supply in countries like U S A , Canada , Switzerland , Sweden . Their application for snowmelt simulation on the rivers of the Himalayan region is yet to be tested. In order to achieve success with such models concerted action is required to strengthen the network of observatories in the Himalayas.

1.0 INTRODUCTION

Snow is the solid form of water and occurs as a part of the nature's hydrologic cycle. About 80% of fresh water on our globe is locked up in the form of snow ,ice, and glaciers (Bahadur,1983). It is an important source of run-off in many parts of world. In most of the high mountainous areas snow is the major source of water yield. Snow contributes significantly to soil moisture recharge of soil profile. With low rate of melt nearly all of it enters the soil whereas the rapid melt rate may cause higher run-off which may result in serious floods.

In the Northern part of our country , spring and summer run-off, comprising mostly of snowmelt ,is the source of water for irrigation ,hydroelectric power and drinking water supply. The major river systems of Northern India, namely the Indus, the Ganga,the Brahamputra and their tributaries which contribute a large part of water resources of the country, originate in the Himalaya. It is estimated that 30-50% of the total annual water yield of almost all the major rivers of upper India is provided by the snow and glacier melt run-off (Agarwal et al ,1983).Major contribution of the snowmelt to the river system from the non or marginally glacierized basins, however, remains restricted to peak summer months, through monsoon season. An accurate estimate of the volume of water contained in snowpack and the rate of its release are, therefore, needed for efficient management of water resources, which includes river and flood forecasting,reservoir operation, design of hydrologic and hydraulic structures etc. The plans for micro hyroelectric projects on Himalayan rivers in the country further emphasizes the needs for reliable estimate of snowmelt run-off.

With the advent of high speed digital computers ,streamflow simulation models came to be used widely in the recent years. Methods have been developed for studying various parts of snow cycle such as accumulation ,distribution, surface energy exchange, metamorphism, water movement and snow soil interactions etc. and combining them in streamflow simulation models for snowcovered watersheds. In this status report, snowmelt studies carried out in different countries including salient features of some important operational snowmelt models have been presented. Particular emphasis has been laid on the studies carried out on rivers of Himalayan region.

2.0 DISTRIBUTION AND PERIOD OF SNOWFALL OVER HIMALAYA

Winter precipitation in the hilly areas of Jammu & Kashmir ,Himachal Pradesh and Utter Pradesh occurs mainly under the influence of weather system known as Western disturbances. These are extra tropical disturbances which originate from Caspian and Mediterranean and move from west to east in the winter months across Iran, Afghan and Southern USSR.

The Himalaya, radiating eastward in three distinct ranges and extending about 2500 km in length and 250 to 400 km in width, influence the Western disturbances and the distribution of snowfall in the mountainous terrain of our country. Orography also plays an important role. The amount of precipitation over the Himalayan catchments depends on the location, elevation and meteorological factors. It is seen that whenever surface low associated with upper air cyclonic circulation is formed to North of 30°N latitude, Great Himalayan ranges receive maximum snow precipitation.The Pirpanjal ranges however do not receive same amount of precipitation.The snow precipitation decreases and seasonal snow line moves to higher altitudes in the eastern Himalayas in comparison to the western part of it.

Rao (1983) reported that several Western disturbances in a winter season move over Northern India with an average frequency of 5 to 6 disturbances in a month . The snowfall occurs in spells or storms spread out generally over a period of four months (Nov-Feb) on an average 13 spell occur in a winter. The duration of individual storms varies from 1 to 7 days, generally being 2 days. Precipitation of 3 to 4 cms/hour are quite common. The Great Himalayan range and Pirpanjal range constitute two distinct barriers from the point of influencing precipitation pattern associated With western disturbances.

The density of fresh dry snow not affected by wind is quite low, the value ranging between 0.04 to 0.14 gm/cc in the Himalayan regions. Maximum packing and settlement , takes place within the first 24 hours resulting in loose snow attaining densities in the range of 0.15 gm/cc. The wind affected fresh snow in some parts of Himalayas was observed to have densities in the range of 0.15 to 0.25 gm/cc (Rao,1983).

The normal snowfall season in India is generally considered to be from December to April. However occasions are not uncommon when snowfall occurs in the month of November. Generally first snowfall in the country occurs at altitudes below 3000 m in the month of December due to which the presence of snowcover is noticed. At higher altitudes the build up of snow commences from November itself. Because of clear weather for longer duration between snow spells, in the beginning of the season, the snowcover does not remain for long in the lower altitudes. It is only by about January standing snow is found at many locations. Rao (1983) reported that the highest snowfall amount occurs in the month of February for Greater Himalayan range whereas it occurs in the month of January for Pirpanjal. In all other winter months except January the snow precipitation is more for Greater Himalayas. Gulati (1972)has given the distribution of seasonal precipitation as percentage of annual precipitation for the Himalayan basins for different parts of Himalaya (Table 1).

Table 1 : Percentage of Annual Seasonal Precipitation

S No.	Name of Himalayan section	Snow Accu-mulation season (Dec-Feb)	Snowmelt season (Mar-May)	Monsoon season (Jun-Sep)	Ground water season (Oct-Nov)
1.	Kashmir Himalaya	22.1	22.0	53.6	2.3
2.	Punjab Himalaya	11.0	8.1	78.4	1.6
3.	Garhwal Himalaya	6.0	3.6	87.8	2.6
4.	Nepal Himalaya (western & central)	3.9	2.9	88.0	5.2
5.	Nepal Himalaya (eastern)	2.9	6.8	85.0	5.3
6.	Sikkim Himalaya	2.0	16.5	74.6	6.9
7.	Assam Himalaya	2.4	25.7	65.8	6.1

Source : Gulati (1972)

Attempts have been made for annual value of precipitation amount in the various parts of Himalayas, but it is not found in literature.

Upadhyay and Bahadur (1982) have dealt with hydrometeorological aspects of precipitation in the Western Himalayas and brought out salient features of orographic precipitation such as windward/leeward effects and altitudinal variations for the elevation range of 400-3200 m. The range of increase of precipitation has been reported as 3 to 200 mm/100 m.

3.0 ACCUMULATION AND DEPLETION OF SNOW

Each spell of snowfall deposits a new layer of loose snow. As the snowpack builds up progressively it is made up of dissimilar strata and individual layers correspond to different spells of snowfalls. The snow cover depth exhibits considerable fluctuations with passage of time. The major factors influencing the fluctuations in snowcover are deposition, settlement and loss of mass due to melting, evaporation , erosion and sublimation. All these processes are influenced by weather conditions, topography and vegetation. Winds cause snow drift leading to the depletion of snow cover entirely from some areas and accumulation at other locations. While the snow cover is more or less of uniform thickness in a plain country ,it is quite uneven in a mountainous terrain. The drift of snowcover is a major problem in our country because of high wind speeds prevailing in the Himalayas during winter.

Freshly deposited snow is a highly porous permeable aggregate of ice grains. The grains can be predominantly single crystal or close grouping of several crystals . The pores are filled with air and water vapour. The characteristics of snow at a given time and place depend on the shape, size and fabric, for their magnitude and time rate of change.

The transformations are of greater significance as far as seasonal snowcover is concerned since they influence the characteristic behaviour pattern of snow pack. Investigations have shown that the transformation of fresh snow into fine granular snow continues for nearly two weeks. The growth of the snow grains (2-3 mm) goes on for another month. By this time discrete crystals of regular form begin to appear in the snow and in another month these crystals are found in considerable number,giving the snow a coarse granular texture. Thus the three stages of recrystallization process take 2-3 months (Anisimov,1958).

In the process of melting , the distribution of snowcover undergoes modification. Snow at low elevations melts more rapidly than at high elevations. Snow on south facing slopes melts faster than on north facing slopes in the Northern Hemisphere. Snow in open areas melts at a different rate than snow in the forest. All these differential rates of melting produce a continual change in the distribution of water equivalent of snowcover.

The effect of snowmelt run-off is normally observed in the rivers in laterhalf of March or beginning of April, the maximum being generally in the month of June. The quantity of run-off varies with magnitude of snowfall in watershed . In the beginning of melt , the free water content in snowpack increases at a fast rate with consequent lubricating of the bonds and weakening of the structure leading to its ultimate disappearance at least at lower altitudes by the end of April. As the summer season advances the melt rate of snow increases and the snow line goes up. For example, it is observed in the Beas basin that snowline which is at about 4500 m at the beginning of winter has come down to 2000 to 2200 m by February . By mid of May the snow line receded to 3500 m and by mid of June,it was observed that most of the seasonal snow cover has disappeared (Rao,1983). Thus in July-August snow is expected to remain only on high mountain ranges near and above the "permanent snowline". Because of the higher altitudes of Greater Himalaya some of the peaks remain snowbound throughout summer season and act as the store house for the ice sheets and glaciers in the Himalayan region.

A study by Srinivasan and Raman (1972) on the snow hydrology of the Himalayan region, utilizing the API satellite photographs for one winter season of 1969-70, lends support to the above description in general. It was reported that by the peak of the summer the total snowcover in the Himalayan region reduced to about 1/10 of the maximum snowcover during the peak of winter season. The seasonal flow of rivers in Indus basin as percentage of annual flow on three monthly basis (Bagchi, 1981) given in Table 2.

Gupta (1983) reported that monsoon is generally feeble at elevations above 3000 m and does not contribute to the accumulation of snow and glaciers. Due to strong incoming insolation during July-August when the monsoon is very active in the region covered by the Himalayan foothills, sufficient amount of water is released by snow which aggravates the monsoon floods.

Table 2 : Percentage of Annual Seasonal Flow of Indus Systems

S No	Name of river	Apr-Jun	Jul-Sep	Oct-Nov	Jan-Mar
1.	Indus	31	54	8	7
2.	Chenab	28	56	7	9
3.	Jhelum	44	36	8	12
4.	Ravi	30	51	8	11
5.	Beas	15	67	10	8
6.	Satluj	23	62	9	6
7.	All rivers together	30	54	8	8

4.0 REMOTE SENSING APPLICATION IN SNOWMELT STUDIES :

In most of the snowmelt models, snow cover area and snow water equivalent are very important variables required for snowmelt simulation and forecasting. The use of remotely sensed data in snow related studies has shown encouraging results. Among the advanced technologies, satellite remote sensing has emerged as a powerful and promising tool for assessment and monitoring of snowcover. The repetitive coverage of satellite provides scope to monitor snowline movement and other characteristics of snow. Problems like location, recognition and measurement in targeting snow and ice are minimum and as such monitoring of snow covered area by remote sensing could be made automatic and operational. However, satellite imagery could not provide any information about either snow depth or about water equivalent of snow. Amongst the sensors available in different satellites, Landsat MSS 5 (0.6-0.7 um) and NOAA visible data are most suitable for snow cover area determination. Digital analysis has been found very useful in assessment of snowcover area from the satellite data. This method analyses individual resolution elements instead of average of several pixels as in the case of optical/visual interpretation. The supervised classification of digital analysis of remotely sensed data is found very effective in identifying snow and ground cover mixture.

The relation between mean snowline altitude and snow cover area has always been useful information for the hydrologists. From a number of imagery , the mean snowline altitude is derived with the help of a transparent overlay with topographic contours. The mean snowline altitude so determined is converted to equivalent of snow cover area using area- altitude curve for the watershed (Meier and Evans,1975). Zhidikov et al (1976) have made computation of snowmelt and melt water yield from open and wooded areas separately. The snowcover area in the open and wooded areas has been used as functions of cumulative melt. Aul and Ffolliott (1975) carried study on correlation between snow cover area and snow melt run-off and reported that a good correlation exists between these two variables. Seasonal run-off forecasts using snow cover area from satellite imagery have been made by Rango et al (1977). Baumgartner et al (1986) developed a method for determining the change of snow cover during snowmelt period using Landsat data. This information was used for snow melt modelling studies in the Swiss Alps. The application of remote sensing techniques for snow cover delineation has been also made in the snowmelt studies carried out in the Himalayan catchments. (Dhanju,1978; Thapa, 1980 ;Bagchi, 1981; Jeyram and Bagchi,1982; Dey et al, 1983; Dey and Goswami, 1984; Ramamoorthi, 1984,1986). The details of these studies has been reported in this report elsewhere.

Because of similar spectral similarity, separation of clouds from snow has always been a formidable problem (Conover, 1964; Meier, 1973; Barnes and Bowley ,1973). However other characteristics in data such as pattern, configuration etc. can often be used to advantage to separate cloud cover. As such separation of snow and cloud is relatively easy with analog techniques. A separation based on spectral signature from Landsat has become possible after recent developments.It has been observed that the integration of a band 1.55 to 1,75 um in satellite is very useful from snow and cloud separation. Accordingly two channels in middle infrared of spectrum 1.55 to 1.75 um and 2.09 to 2.35 um have been provided in Landsat 4 & 5 satellites. The band 5 (1.55 to 1.75 um) of Thematic Mapper(TM5) data from Landsat 4 & 5 with 30 m resolution are very useful for separation of snow from clouds.

Passive and active microwave sensors have shown potential to snowmelt studies. Microwaves have capability to penetrate the snow and respond to variation in subsurface properties of hydrologic importance. Moreover, microwave sensors permit observations during almost all weather condition. The large contrast between the dielectric constants of water (1.0), snow (2.0), ice (3.2) are enough to show a strong microwave response (Schmugge, 1980). But because of complexities in the microwave interaction significantly, more ground information is needed for microwave snow studies than for comparable visible, near infrared studies (Rango, 1985). Josberber and Beauvillian (1989) made a comparative study of the passive microwave images from NIMBUS Scanning Multichannel Microwave Radiometer (SMMR) and visual images from Defense Meteorological Satellite Program of Upper Colorado river basin and found that passive microwave imagery may be used to determine the extent of snow cover.

Airborne gamma ray spectrometry is a remote sensing technique for measuring snow water equivalent. This technique has been used successfully in USA (Carroll and Vadnais, 1980), Canada (Carroll et al, 1983), Sweden (Bergstorm and Brandtt, 1985) and the Soviet Union (Vershinina , 1985). When Satellite images and gamma ray spectrometry measurements are used together , snowmelt processes are monitored. Satellite imagery provides good information on snowline while gamma ray technique offers information on snow water equivalent. Kuittinen (1986) determined the spring time areal snow water equivalent values over Northern Finland using enhanced digital NOAA images and aircraft gamma ray spectrometry.

5.0 PHYSICS OF SNOWMELT AND METHODS OF SNOWMELT ESTIMATION

The physics of snowmelting is an important aspect of snow hydrology subject which is relatively a new branch of science.The various phases of sciences such as meteorology, thermodynamics, geology, and soil mechanics etc. are included in this subject. It really began with the pioneering work of Dr. J E Church who is considered as a father of snow surveys. In the early 1900's he began to study the factors which affect deposition and melt and to develop system of predicting water run-off . The first few snow laboratories were established in the Western United States with the objective of developing predictive equations with which severity of floods could be determined in advance.

Snowmelt is an over all result of the different heat transfer processes to the snowpack. The relative importance of the various heat transfer processes involved in the melting of snowpack varies with time and locale. There are two basic approaches generally adopted for estimation of snowmelt from a watershed. The first approach is known as energy budget or energy balance and second is the temperature index or degree-day approach.

5.1 Energy Budget Approach :

The energy budget approach is a theoretically sound heat transfer method and requires lot of data on various energy sources to the snowpack. The prccess of snowmelt is governed by the energy balance at the upper (snow-air) surface and lower (snow-ground) surface of the snowpack. An energy balance equation for a snowpack is written as:

$$Q_m = Q_n + Q_h + Q_e + Q_p + Q_g - du/dt \qquad (5.1.1)$$

where,

Q_n = Net all wave radiation flux

Q_h = Convective or sensible heat flux

Q_e = Flux of latent heat (evaporation, sublimation, condensation) at the snow air interface.

Q_g = Flux of heat from the snow-ground interface by conduction

Q_p = Flux of heat from rain

Q_m = Energy associated with the flux of melt water

du/dt = Rate of change of internal energy(or stored energy) per unit area of snow cover

The following brief description of different energy sources illustrate the relative importance of various snowmelt energy sources including simplified methods for snowmelt estimation by individual source of energy.

5.1.1 Radiation Flux :

Radiation is the largest energy source for melting of the snowpack. However, the importance of the energy source for snowmelt varies with both diurnally and seasonally and between open and forested areas. Many individual processes are involved in the interaction of solar radiation with atmosphere and subsequent reflection and absorption processes at the snow surface. Because of scattering of radiant energy, at the surface not only direct solar radiation but diffuse radiation is also observed.

The diffuse radiation reaching the snow surface from all directions has generally been ignored in energy balance estimates and direct beam component is taken as an index of total short-wave radiation. However, it has been reported that diffuse radiation consists of 10% of short-wave radiation on clear bright days. The significance of diffuse radiation in snowmelt processes increases appreciably on partially cloudy or overcast days. The diffuse radiation is made up of 50-100 % of short-wave radiation under such condition.

The melt by net shortwave radiation is given by

$$M_{SN} = I_s (1 - C) (1 - \alpha) \quad \text{mm/day} \qquad (5.1.2)$$

where C is the cloud cover and α is albedo of the snowpack and I_s is the incident radiation. The albedo and cloud cover reduce the potential melt through net solar radiation.

Determination of the radiation characteristics of the underlying snow surface is important for taking into account the radiation interaction atmosphere and snow surface. The predictive equations could be used for making basin wide estimates of net radiation over non-uniform terrain from albedo information, if available. The albedo for dry and fresh snow is approximately 0.99, whereas for old and moist snow it reduces to around 0.40 (U S Army Corps Engineers, 1956). Very little research has been done on the areal variability of snow albedo and methods for estimating mean value over an area. Few studies correlating albedo with grain size and density of snow (Choudhary ,1979) has revealed several inconsistencies in the results and need more experiments to arrive at definite conclusions.

The longwave radiation exchange depends on black body radiation from the snow itself and from clouds and tree cover and gray body radiation from the overlying air mass . The absorbed energy increases the temperature of air which inturn enhances the amount of longwave radiation to the snow surface. Under clear sky conditions in an open and tree free area, the net longwave radiation , I_{LN} at the melting snowpack is the difference between the incoming gray body radiation from the clear sky and the black body radiation from the snowpack. Using Stefan's law expressed as a linear function of temperature above freezing , plus a small higher order terms, the net longwave radiation can be written as (Quick and Pipes, 1988):

$$I_{LN} = 7.51\ T_a - 161 - 9.92\ T_s \quad \text{langleys/day} \qquad (5.1.3)$$

where T_a and T_s are the mean air and snow surface temperatures (°C) respectively. The snowmelt equivalent, M_{LN} becomes

$$M_{LN} = 0.094\ T_a - 20.1 - 1.24\ T_s \quad \text{mm/day} \qquad (5.1.4)$$

It shows that for a melting snowpack ($T_s = 0$°C), M_{LN} does not become positive until T_a exceeds 21.4 °C. Under partly cloud conditions the equation (5.1.4) becomes

$$M_{LNC} = (0.94\,T_a - 20)(1-C)+1.24\,T_c\,C \qquad \text{mm/day} \qquad (5.1.5)$$

where T_c is cloud temperature and can often be approximated by the dew point ,or by minimum air temperature.

5.1.2 Convective Heat Flux :

Convective or sensible heat transfer is produced by turbulent heat exchange between the air mass immediately above the snow pack. This heat transfer is dependent on both the wind and air temperature and particularly on the stability of air mass above the snowpack and is approximated by

$$Q_h = 0.904\,(p/101)\,T_a\,V \qquad \text{langleys/day} \qquad (5.1.6)$$

The corresponding melt is given by

$$M_h = 0.113\,(p/101)\,T_a\,V \qquad \text{mm/day} \qquad (5.1.7)$$

in which p is the atmospheric pressure in kN/m^2 for elevation being considered and V is the wind speed in Km/hour.

A warm air mass above the cold snow surface tends to be stable, resisting any downward transport of heat to the snowpack, unless there is enough wind to produce turbulent mixing. As air temperature increases, and if the wind is only moderate, the stability can increase to the point where very little convective heat transfer can occur. Convective heat transfer is therefore self limiting and becomes quite small at higher temperatures, unless there is very strong wind.

5.1.3 Latent Heat Flux :

The latent heat transfer, often termed condensation melt, is caused by the transport of moisture to and from the snowpack . Occurrence of condensation, releasing latent heat to the snow, or occurrence of evaporation , cooling the pack, depends on relative vapour pressures of the air and the snow surface. Wind is once again an important factor and so is stability as for convective heat transport. Latent heat transport can therefore produce snowmelt but, like the convective heat transport, becomes limited at higher temperatures by the stability of the warm air mass. The latent heat transfer flux is approximated by

$$Q_e = 3.52\, T_d\, V \qquad \text{langleys/day} \qquad (5.1.8)$$

and corresponding melt is given by

$$M_e = 0.44\, T_d\, V \qquad \text{mm/day} \qquad (5.1.9)$$

where T_d is dew point temperature and can be approximated by the minimum air temperature.

The simulation of the turbulent exchange processes for convective and latent heat transport is not as advance as that of the radiation exchange. The accurate determination of sensible and latent heat is found crucial to snowmelt forecast even in well instrumented catchment. In the mountainous areas this process becomes more complex because of the local winds in such regions. It has been shown through field experiments that the bulk aerodynamic formulae are adequate for estimation of turbulent energy exchange and can be used in energy balance estimates for daily or short term snowmelt forecasting. The nature of the turbulent exchange processes immediately above the snow surface is still to be investigated. The whole energy balance of the snowpack in the forested areas, atleast for forecasting purposes, can be represented by the net radiation because the contribution of the turbulent exchange is small (Hendrie and Price ,1978).

5.1.4 Rain Heat Flux :

The occurrence of warm rain on snowpack at uniform freezing temperature also causes higher rate of snowmelt resulting peak floods quite disappropriat to the rainfall. The amount of heat given up to the snow by rain water is directly proportional to the quality of rain water and its temperature excess (above that of snowpack). Assuming that rain falls at mean temperature, the snowmelt from rain is given by

$$M_r = K * TM * RN \qquad \text{mm/day} \qquad (5.1.10)$$

where,

RN = rainfall (mm/day)
TM = mean temperature (°C)
K = a factor representing the heat content of rain

Some investigators have considered wet bulb temperature as a suitable estimate of the rain temperature (Anderson, 1976). The details of each energy components have been reviewed by Singh (1987).

5.1.5 Ground Heat Flux :

The ground heat flux has special hydrologic significance since it causes melting during the winter and early spring when melt at the snow surface is non-existent. The cumulative contribution of ground heat flux to the snowpack becomes more significant when complete melt season is considered . Generally in snowmelt calculations, average values of ground heat flux are used . Speers et al (1978) accounted for the snowmelt resulting from conduction heat from ground by specifying constant melt rates of the order of 0.1 to 1.0 mm/day as a function of month of year.

5.2 DEGREE-DAY APPROACH :

The specific kinds of data required for detailed thermal budget analysis are rarely available to the hydrologist concerned with snowmelt studies. The most generally available data are daily maximum and minimum temperatures, humidity measurements and surface wind speed; less frequently are found stations having continuous temperature, humidity, and wind records, and few are stations having solar radiation or even duration of sunshine records.

The temperature indexes are widely used in snowmelt estimation because it is generally considered to be the best index of the heat transfer processes associated with snowmelt. It reflects the extent of radiation and vapour pressure of air and it is also sensitive to air motion. Moreover, air temperature measurements are the only data available from which snowmelt can be computed. Air temperature expressed in degree days is used in snowmelt computations as an index of complex energy balance tending to snowmelt. A degree day, in its broad sense, is a unit expressing the amount of heat in terms of the persistence of a temperature for 24 hour period of one degree centigrade departure from a reference temperature. The simplest and the most common expression relating to snowmelt to the temperature index is:

$$M = D_f \, (T_i - T_b) \qquad (5.2.1)$$

where,

M = melt produced in cm of water in a unit time
D_f = degree day factor (cm/°C/unit time)
T_i = index air temperature (°C)
T_b = base temperature (usually 0 °C)

Daily mean temperature is the most commonly used index temperature for snowmelt. The mean temperature is computed by:

$$T_i = T_{mean} = \frac{T_{max} + T_{min}}{2} \qquad (5.2.2)$$

Many other temperature indexes have also been tried in a attempt to improve the index relationship (Synder,1939). Most of these are based on daily maximum and daily minimum temperature because of wide availability and simplicity of use of these measures. Daily maximum temperature has also been extensively used by some investigators as an index of snowmelt (Mc Callyster and Johnson, 1962). Maximum temperature has been found to be more accurate index than daily mean temperature (U.S Army Corps of Engineers,1956).

There are several methods for dealing with the index temperatures used in calculating the degree day value. When using the maximum - minimum approach, the most common way is to use the temperature as they are recorded and calculate the average daily temperature. Garstka et al (1958) reported that some times computations of degree days from daily mean temperature are found misleading. In many parts of the Western United States mountainous areas, the drop in minimum temperature are so great that daily mean temperature are below 0°C, indicating no degree days, whereas snowmelt conditions have prevailed during part of the day when air temperatures were much above the freezing point. 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. In order to counteract such problems, alternatives have been suggested in which unequal weight to maximum and minimum temperature are given. U.S Army Corps of Engineers (1956) reported the use of the following index temperature:

$$T_i = \frac{(2T_{max} + T_{min})}{3} \qquad (5.2.3)$$

Another approach is given by :

$$T_i = T_{max} + \frac{(T_{min} - T_{max})}{B} \qquad (5.2.4)$$

where , B is a coefficient less than 2.

The 0°C base temperature is most commonly used in snowmelt computations in degree days. This follows from the idea that most snowmelt results directly from the transfer of sensible heat from the air in excess of 0°C. Its use stems largely from the lack of knowledge concerning base and their variations. Nevertheless, 0°C base temperature is a good average value for mean temperature indexes. Where maximum temperature indexes are used, higher base and somewhat lower degree day factors have been indicated.

The degree days are extrapolated to an elevation zone by using a suitable lapse rate i.e.

$$DT = \delta \ (h_{st} - h) \qquad (5.2.5)$$

where,

- DT = temperature lapse rate correction factor in °C.
- δ = temperature lapse rate in °C per 100 m .
- h = Zonal hypsometric mean elevation in m.
- h_{st} = altitude of temperature station in m,

In basin with little seasonal variation, a lapse rate of 0.65°C/100 m has been found suitable according to literature cited .

According to a study by the U.S Army Corps of Engineers (1956), the daily spring time snowmelt may be estimated by the correlation equations as a function of mean daily temperature T_{mean}, the maximum daily temperature T_{max} and the relative forest cover as given below. Varshneya(1979) reproduced these equation in metric system as follows

(i) For open sites :

$$M = 0.03 (9 T_{mean} + 40) \qquad (5.2.6)$$

$$M = 0.02 (9 T_{max} + 25 \qquad (5.2.7)$$

(ii) For forest sites :

$$M = 0.025 (9 T_{mean}) \qquad (5.2.8)$$

$$M = 0.02 (9 T_{max} - 50) \qquad (5.2.9)$$

where,

M = daily snowmelt in cms
T_{mean} = mean daily temperature in °C
T_{max} = maximum daily temperature in °C

The coefficients are known as degree day factors. The equations are applicable for T_{mean} in the range of 1 to 20°C and for T_{max} in the range of 6.5 to 25°C. There is wide variation in melt rates between periods, because air temperature alone does not adequately represent the entire melt process, particularly for open sites, The above relations, however, represent the average relationship between air temperature and snowmelt during active spring snowmelt period, for extremes of the forest cover.

5.3 Generalized Snowmelt Equations :

U S Army Corps of Engineers (1956) has given some generalized snowmelt equations for basin wide snowmelt calculations. Varshneya (1979) has also reproduced these equations in metric system as follows :

(a) Basin snowmelt during clear weather periods :

(i) For open areas (< 10 % forest cover)

$$M = 0.0129\, K' I_i\, (1-a) + (1-N)(0.098 T'_a - 2.13 + 0.133\, N\, T'_c$$
$$+ 0.0241\, K\, V\, (0.22 T'_a + 0.78 T'_d) \qquad (5.3.1)$$

(ii) For partly forested area (10%-60% forest cover)

$$M = 0.01\, K' I_i\, (1-a)(1-F) + 0.0241\, KV\, (0.22\, T'_a + 0.78\, T'_d$$
$$+ 0.133\, F\, T'_a \qquad (5.3.2)$$

(iii) For Forested area (60%-80% forest cover)

$$M = 0.0241\, K\, V\, (0.22\, T'_a + 0.78\, T'_d) + 0.133\, T'_a \qquad (5.3.3)$$

(iv) For heavily forested areas (> 80% forest cover)

$$M = 0.34\ (.053\ T'_a + 0.47\ T'_d) \qquad (5.3.4)$$

where,

M = snowmelt in cms
V = wind velocity at 15 m level in km/hour
a = observed and estimated average snow surface albedo
F = forest cover expressed in decimal
N = cloud cover expressed in decimal
K'= basin short wave radiation factor (between 0.90 and 1.1) depending on average exposure of open area to short wave radiation in comparison with an unshielded horizontal surface
K = basin covection- condensation melt factor
I_i = observed or estimated solar radiation on horizontal surface in langleys
T'_a = difference between air temperature at 3 m level and snow surface in (°C)
T'_d = difference between dew point temperature at 3 m level and snow surface (°C)
T'_c = difference between cloud basetemperature and snow surface temperature, estimated from upper air temperature or by temperature lapse rate ,preferably on snow free sites

(b) Basin snowmelt during rain :

(i) For open areas (< 10% forest cover) or partly forested areas (10%-60% forest cover)

$$M = (\ 0.133 + 0.0242\ K\ V + 0.01225\ P_r\ )\ T_a + 0.228 \qquad (5.3.5)$$

(ii) For heavily forested areas (> 80% forested cover)

$$M = (\ 0.34 + 0.01265\ P_r\ )\ T_a + 0.127 \qquad (5.3.6)$$

where ,

M = snowmelt in cms
V = mean wind velocity at 15 m level in km/hour
K = basin constant ranging from 1.0 to unforested plain to 0.3 for very thick forest
T_a = mean temperature of saturated air at 3 m level in °C
P_r = rate of precipitation in open portion of basin in cm/day

A comparison of temperature index and energy budget method reveals that the first approach has the advantage that value of primary index , air temperature, is commonly observed and its areal variability can be satisfactorily estimated in most cases. Its disadvantage is that it is an index method and there can be significant scatter in relationship between the index variable and snowcover heat exchange. The advantage of the use of a theoretical heat transfer approach is that this scatter is decreased. The disadvantage of the theoretical heat transfer approach lies in data requirements. In most cases some of the variables are not observed. If the variables are not observed , they must be estimated. The estimation of solar radiation from percent sunshine or cloud cover ; the estimation of atmospheric long-wave radiation from air temperature, vapour pressure, and an index to sky cover; and estimation of albedo from days since last snow or accumulated temperature indices, all induce error. The adjustment of data for geographical factors again creates additional errors.

Rockwood (1972) reported that temperature index method is found most practical method to use for day to day streamflow forecasting , considering basin limitations in availability of basic data as well as the need for simplified inputs for operational use on a real time basis. However , the selection of method for a particular evaluation depends upon the type of appliction. For design flood determinations , the use of thermal budget approach preferred because of the need to establish on a theoretical basis the upper limits of various snowmelt functions. For streamflow forecasting temperature index method is used.

The relationships of air temperature and radiation with snowmelt run-off are found linear having the high values of coefficient of correlation of 0.985 and 0.966 respectively (Jolly ,1972). These results indicate that air temperature and radiation could explain most of snowmelt run-off for rainfree days. Anderson (1976) carried out a careful and detailed analysis of energy balance and reanalysed some of U S Corps data. It was shown that energy budget and degree-day methods can give similar results when tested against snow course and snowpillow data (Quick and Pipes,1988).

6.0 CLASSIFICATION OF SNOWMELT MODELS :

There are two basic approaches used in snowmelt models based on which the models are categorized as statistical or simulation models.

6.1 Statistical Snowmelt Models :

The statistical correlation models are generally used for seasonal yield forecasting. The basic method uses the correlation analysis to relate the current snowcover or the past precipitation, or combinations thereof, to observed seasonal run-off (U.S. Army Corps of Engineers, 1956 ; Linsley et al, 1949). Other variables have been added to the analysis in an attempt to improve the results.These include base flow, soil moisture, wind, high elevation and low elevation water equivalent or precipitation ratios and areal extent of snowcover (Anderson, 1972). Long-term forecasts are made using statistically developed equations relating volume of run-off to snowpack and antecedent conditions.The long term forecasts are generally concerned with seasonal volumes or, some times, seasonal maximums using current snow cover conditions plus probable future climatic patterns. In such forecasts the detailed time distribution of run-off is not concerned although monthly distribution may be attempted. Statistical methods are widely used and yet they are limited in accuracy by the theoretical constraints on the data as well as by their inability to describe the complex physical interaction of run-off processes. The general form of equations generally obtained through regression analysis is

$$Q = k_1 x_1 + k_2 x_2 \ldots\ldots\ldots\ldots + k_n x_n \qquad (6.1.1)$$

where, Q is daily discharge $k_1, k_2 \ldots\ldots k_n$ are constants and $x_1, x_2 \ldots\ldots\ldots x_n$ are the independent variables.

6.2 Simulation Snowmelt Models :

Several simulation models of snow accumulation and melting processes have been developed which attempt to mathematically represent each components of total processes. The difference between models are the mathematical relationships used for each component. The simulation models are relatively recent innovations and offer a great potential as both a forecasting tool and a means to improve understanding of snowmelt processes. Such models have advanced considerably over the past more than thirty years. The development of simulation models is due to the increased use,speed and capacity of digital computers.

Hydrologic simulation models that include snow are generally divided into three basic components: the snowcover, a precipitation run-off relationship, and a run-off distribution and routing procedure. Most simulation models simulate the entire snow accumulation and melt season. Some of these models only simulate the snowmelt processes (Anderson and Crawford, 1964 ; Anderson, 1968 ; Anderson and Rockwood, 1970 ; Eggleston et al 1971 ; Willen et al, 1971). Whereas others are used in conjunction with models of soil moisture accounting. A typical flowchart of snowmelt simulation model is given in Figure 1.

Basin models that simulate snowmelt run-off can be classified using a number of model characteristics. The classification used by WMO(1986) is one of the more comprehensive.It begins with the division of the basin model into its two major components namely a snowmelt model and a transformation model. The snowmelt model simulates the processes of snow accumulation and snowmelt, while the distribution and routing is accomplished by the transformation model which relates snowmelt, and any rainfall run-off volumes to discharge hydrographs. The most common transformations functions is the

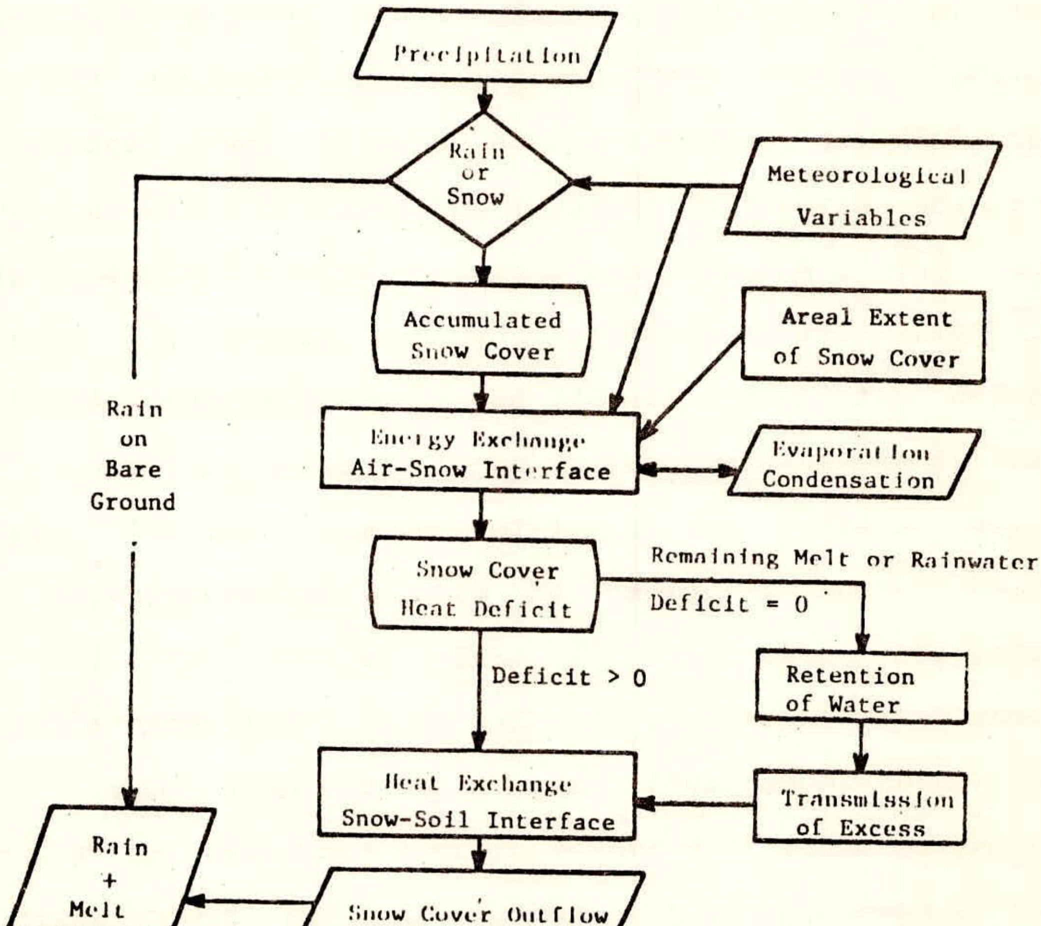


Figure 1. Flowchart of typical snow accumulation and ablation model.

unit hydrograph. For down stream channel reaches, routing may be accomplished by a number of hydrologic routing techniques.

The snowmelt and transformation models are classified further as being either lumped or distributed. Lumped models use one set of mean basin parameter values to define the physical and hydrological characteristics of a basin. Distributed models attempt to account for the spatial variability in these features by dividing a basin into sub areas. A separate set of parameter values is referred for each sub area. In practice,distributed model use one of these general approaches to divide a basin. (1) elevation zone (2) basin characteristics such as slopes, aspect, soil, vegetation, and elevation, and (3) a fixed or variable length, 2 - or - 3 - dimensional grid. Model complexity and data requirements increase substantially as one moves from the elevation zone to the 3-dimensional grid approach. Lumped and distributed snowmelt models are classified further by their use of an energy balance approach or a temperature index approach to simulate the snowmelt processes.

There are several snowmelt run-off simulation model used for operational short-term forecasting in the United States (Anderson and Rockwood,1970); Canada (Quick and Pipes,1977);and Soviet Union (Kuzmin,1969,Kovzel,1969, Kuchment et al,1986).

Anderson (1978) has also dealt in detail various aspects of model structure, including snow accumulation, surface energy exchange, water retention and movements, snowcover properties, snowcover distribution, snow-soil interaction etc.. Seasonal variation in melt factor has been considered essential primarily because of the variation in net available solar energy and the air temperature in general has been found as adequate index of snowcover energy exchange where the meteorological factors such as dew point , wind etc. do not deviate significantly from normal. The effects of transmission of melt water through snow, resulting in lag and attenuation has also been mentioned. Hannaford et al (1970), and Tarbel and Burnash (1971) have emphasized the need to use simulation models to isolate and evaluate those hydrologic relationships important to the forecasting of seasonal volume and it´s distribution. It has been suggested that simulation models may be useful in determining the the effect on runoff of any probable sequence of temperature and precipitation with respect to current hydrologic conditions. Thus it seems that future improvement in seasonal yield forecasting may come through the use of simulation models, plus incorporation of more advanced statistical methods, such as stochastic processes, which may allow more accurate statements as to the probability of future runoff volume and it´s distribution.

7.0 SALIENT FEATURES OF SOME IMPORTANT SNOWMELT MODELS OPERATIONAL FOR FLOW FORECASTING :

Some important snowmelt runoff simulation models which are operational for flow forecasting have been described including their data requirement, output, methods of computing snowmelt runoff etc. The other parameters required for the model habe been also discussed.

7.1 Streamflow Synthesis and Reservoir Regulations Model-(SSARR)

This model was developed progressively since 1956 to provide a generalized computer simulation techniques for analyzing and forecasting various types of hydrologic systems. Rockwood and his co- workers developed this model particularly to meet the demanding conditions of the mountainous Pacific Northwest with an objective for application to daily streamflow forecasting and reservoir regulation. The program description and user's manual for SSARR (1972) has been broughtout by U.S.Army Engineer Division, North Pacific, Portland Oregon. Rockwood (1981) has discussed the theory and practice of this model as related to analyzing and forecasting the response of hydrologic systems. The program provides the basic computer logic to operation on a variety of hydrologic conditions, including snowmelt as well as rainfall run-off. The calculations of snowmelt with SSARR model are accomplished by either (i) the temperature index approach or (ii) the use of generalized snowmelt equations for a partly forested areas. The temperature index method is commonly used for daily forecast operations, whereas the more detailed energy budget approach of generalized equation is found more appropriate for design flood derivations when adequate data is available.

Input Data Requirement :

(i) For Temperature Index Method :

- maximum and minimum air temperature
- precipitation
- area of watershed and its variation with elevation
- snowline elevation
- elevation of meteorological stations
- discharge data for comparison of results

(ii) For Energy Budget Method :

- maximum and minimum air temperature
- solar radiation
- srecipitation
- wind velocity
- dew point temperature above the snow surface and on the snow surface
- average snow surface albedo
- Area of watershed and its variation with elevation
- snowline elevation
- forest cover area
- streamflow data for comparison of results

Output Product :

- streamflow
- elevation of snowline
- soil moisture status

7.1.1 Temperature Index method:

Temperature data , usually mean daily or maximum daily, are input for one or more stations for each watershed as specified by the basin characteristic data. Each temperature station is weighted as with precipitation stations. The weighted temperatures for each computation period are averaged to obtain average temperature for the station.

$$T_a = \frac{T_1 * W_1 + \ldots\ldots + T_n * W_n}{n} \qquad (7.1.1)$$

where,

T_a = average temperature for the stations

$T_1 \ldots T_n$ = period temperature for each station

$W_1 \ldots W_n$ = respective station weighing factors

n = number of stations

The average elevation of temperature stations, E_s is also calculated. The station average temperature, T_a is then lapsed from E_s to the mean elevation of snowmelt zone. The zone of snowmelt is that portion of the snowpack below the elevation of the base temperature and above snowline. The elevation of base temperature, E_b, or the melting elevations, is calculated by :

$$E_b = E_s + (T_a - T_b) / LR \tag{7.1.2}$$

where,

T_b = base temperature, specified as a constant for a basin

LR = lapse rate specified as a constant in °F/1000 ft; model default value is 3.3 °F/1000 ft.

Snowmelt runoff (M) in inches for snow covered area is computed in the model for temperature index method as follows :

$$M = (T_A - T_b)\ R\ \{ PH/24 \} \tag{7.1.3}$$

where,

T_A = period temperature at the median elevation of the melting snowpack (°F)

R = melt rate, specified to the computer, or given as a function of accumulated runoff, in inches of water per degree day

PH = period length in hours.

Values for base temperature and melt rate are adjusted to conform daily temperature data. Melt rates are variable during snowmelt season and can be specified for each day of run. If desired , the melt rate may be specified only for beginning and ending days, or for critical intermediate periods depending upon the data availability. Linear interpolation by the computer provides values for the days not specified.

7.1.2 Generalized Snowmelt Equation :

The generalized snowmelt equation provides an energy budget approach to compute snowmelt for a catchment. The equation used by the model is:

$$M = k'(1-F)(0.0040\, I_i)(1-a) + k\,(0.0084\, V)(0.22\, T'_a + 0.78\, T'_d) + F\,(0.029\, T'_a) \qquad (7.1.4)$$

where,

T'_a = difference between the air temperature measured at 10 ft above the snow surface and the snow surface temperature (°F)
T'_d = difference between the dewpoint temperature measured at 10 ft above surface of snow and temperature of the snow surface
I_i = solar radiation on a horizontal surface ,in langleys
V = wind velocity at 50 ft above the snow, in miles/hour
a = average snow surface albedo, expressed as a decimal
k'= basin shortwave radiation melt factor,expressed as a decimal
F = average forest canopy cover , expressed as a decimal
k = convection-condensation melt factor, expressed as a decimal

The above equation is applicable only for partly forested areas.

7.1.3 Watershed Snowmelt Options :

In addition to the two methods of snowmelt computation, two options namely snowcover depletion and snow band options are available in SSARR model to evaluate the snowpack characteristics in a catchment.

7.1.3.1 Snowcover Depletion Option :

In this option, the snowcover depletion during the active snowmelt period is expressed as a function of accumulated generated run-off in percent of seasonal total. The general shapes of curves relating snowcover to generated run-off are found fairly uniform for different watersheds and from year to year. The general shape of these functions are expressed as the following equation:

$$SCA/100 = 1.0 - (\Sigma\ QGEN/100)^{n} \qquad (7.1.5)$$

where,

SCA = snowcovered area in percent of total watershed area
QGEN= accumulated generated runoff from snowmelt, in percent of seasonal total
n = parameter expressing characteristics of snowcover depletion for a basin.

Figure 2 illustrates a typical snowcover depletion curve as defined by the above equation.

Further, single watershed and split watershed approaches are available within the snowcover depletion option for analysis of run-off from rainfall as related to snowcover. In single watershed approach ,a single index of all watershed parameters is maintained for the entire watershed i.e. for both snowcovered and snowfree areas. Whereas in the split watershed approach , the snowcovered and snowfree areas treated as two separate watersheds ,each with its own characteristics and parameters. The snowfree area is the complement of the snowcovered area derived from the snowcover depletion function.

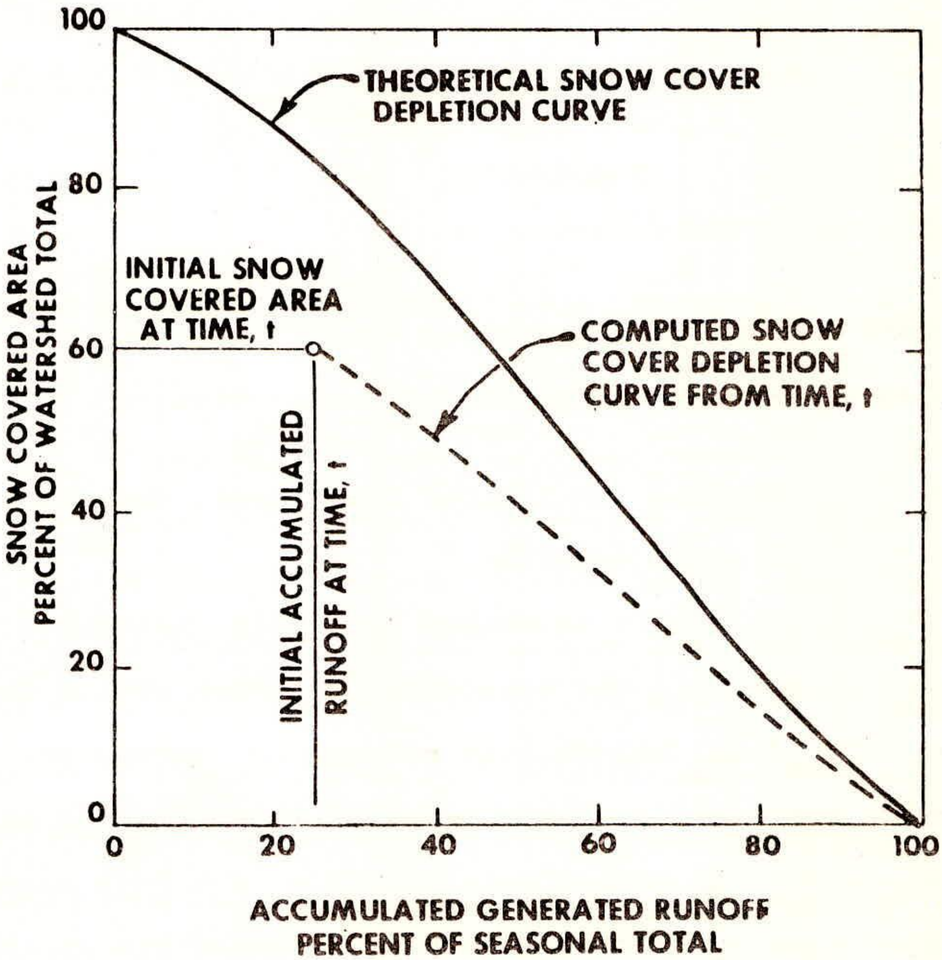


Figure 2 : Snow cover depletion curve used in SSARR model

The single watershed approach is useful in early melt season, when snowfree areas have a high soil water content and thus yield a high percentage of run-off from rainfall. As the snowfree area dries out, however, rainfall-run-off from the snowfree area is often less than from the snowcovered area. In such case, the split watershed approach is considered better so that a separate counting of soil moisture index is maintained for snowfree and snowcovered areas. The differentiation between snow covered and snowfree area is represented by a snow line which usually follows an elevation contours.

7.1.3.2 Snow Band Option :

The second option available with SSARR model, the snowband technique, provides the capability to subdivide a watershed into elevation bands or zones of relatively equal elevation. It allows more quantative appraisal of snowpack . An inventory of snow accumulation and melt is maintained on each band. This approach is particularly suited for mountainous watersheds where snowdepth increases with elevation. Each elevation band is treated as a separate watershed with its own characteristics and initial snow cover water equivalent. Accumulated snow is assumed to cover the entire band. Each band is considered either snowfree or 100 percent snowcovered. Speers et al (1978) have described further development in the operational snow band in SSARR model.

The snowpack conditioning routine accounts for the cold content and the liquid water deficiency of the snowpack is satisfied before liquid water can enter the soil system. Successive period quantities of cold content are accumulated while air temperature is below 0°C and depleted by the presence of liquid water from snowmelt or rainfall. Liquid water deficiency is accounted for specifying percent liquid water holding capacity, taken as a constant of the order of 2-5 % of the water equivalent of the snowpack. The capacity has to be exceeded before rain or snowmelt is available for input to the soil.

Run-off is computed in the same manner for snowmelt as for rainfall except that snowmelt and rainfall are summed to yield a total moisture input to the watershed on a period or daily basis. The continuity of run-off is maintained individually on each band or subbasin area when watershed is divided into elevation bands or snowcovered and snowfree areas. This provides the capability to evaluate differences in soil moisture and run-off condition on a zonal basis. A provision has been made in the program to compute daily snowmelt and to distribute the daily values of melt into 3, 6, or 12 hour periods according to the specified distribution. The flow chart of SSARR model has been given in Figure 3 & 4

The SSARR model has been used routinely for operational forecasts in the predominantly snowfed Columbia river basin. It has been utilized for studies of various river systems, the principal example being the studies of Mekong river in South east Asia (Rockwood,1968).

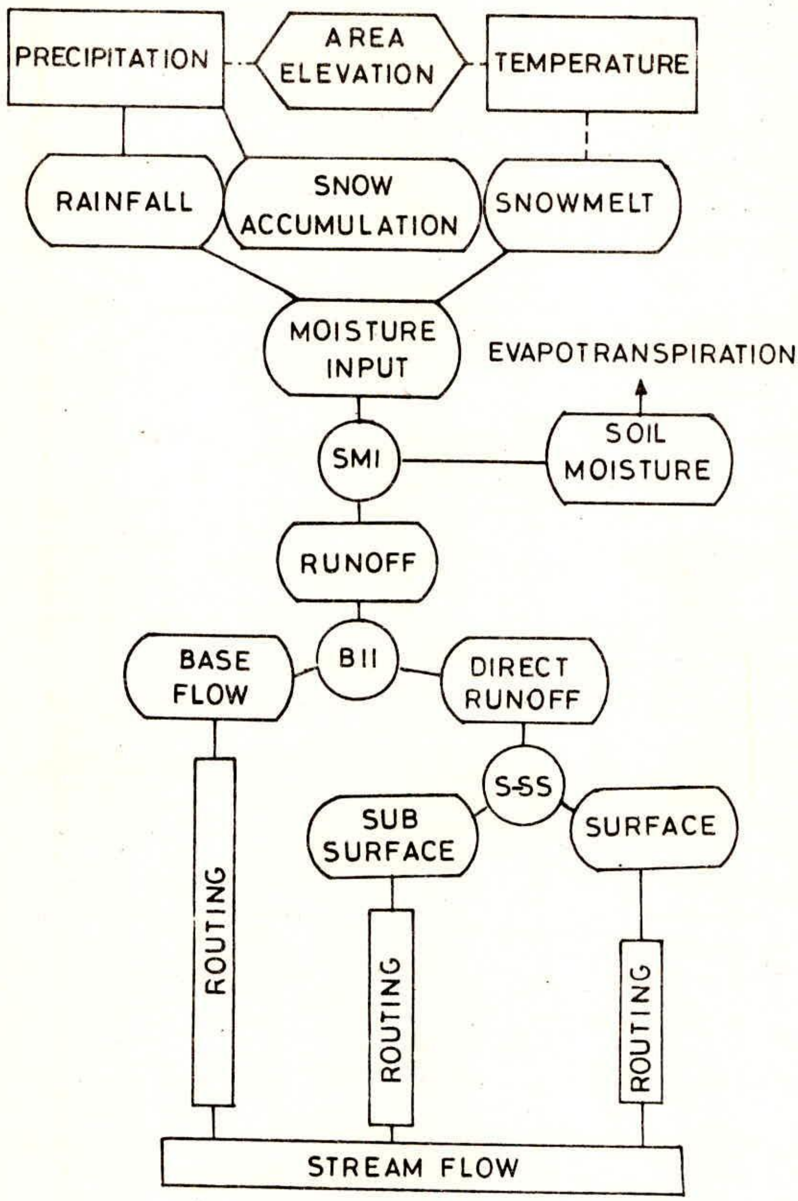


Fig.3. SSARR Watershed Model

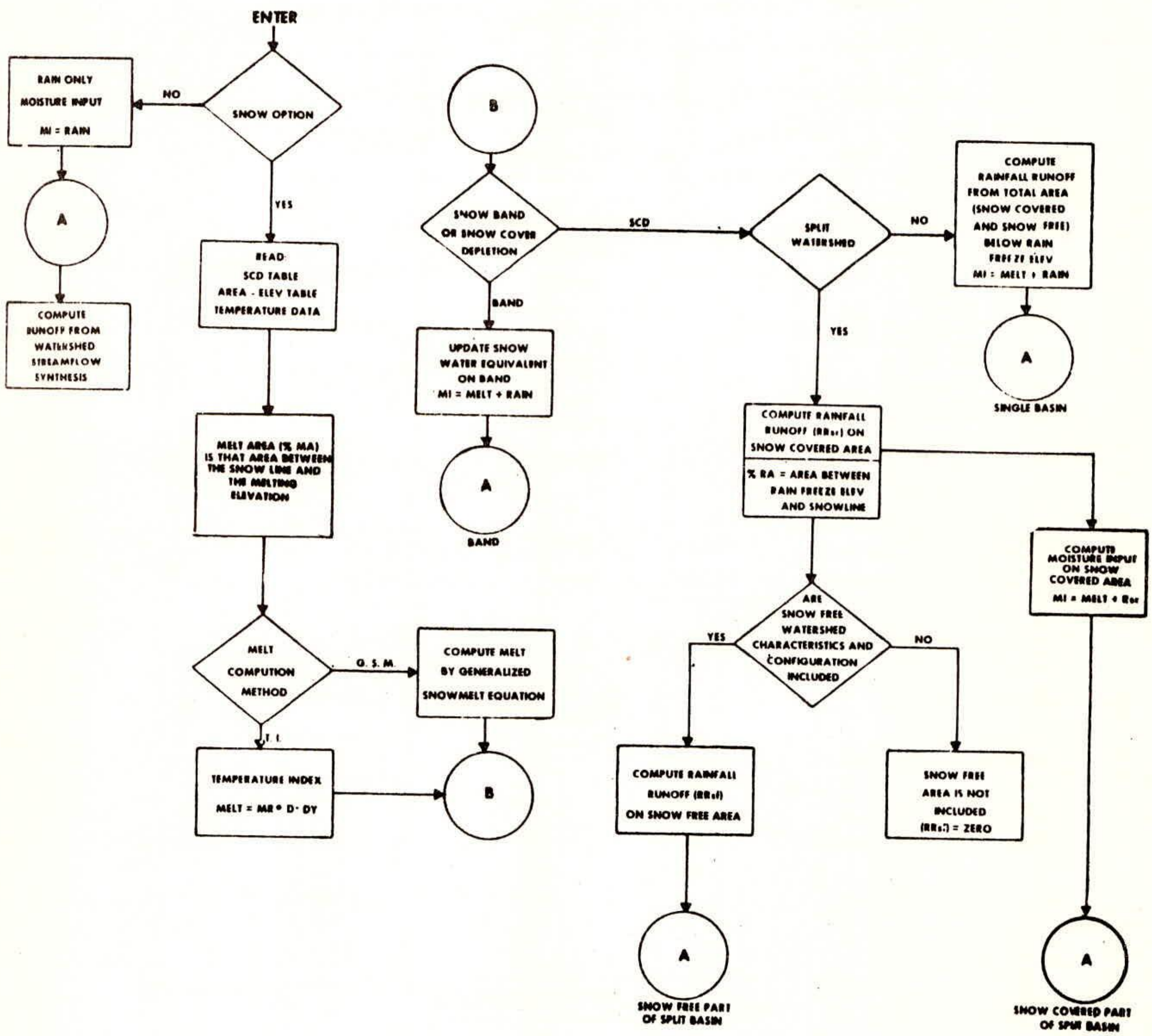


Figure 4 : Schematic diagram of snowmelt options in SSARR model

7.2 Snowmelt Runoff Model - (SRM) :-

The snowmelt run-off model, also referred to in literature as Martinec Model or Martinec-Rango Model is designed to simulate and forecast daily streamflow in mountain basins where snowmelt is a major run-off factor. This model was developed by Martinec (1975) in small European basins. With the breakthrough achieved in estimating snowcover data through satellites in 1970s , it has become possible to test SRM in larger basins. The area of the basin is divided in to various elevation zones and various model variables and parameters are applied to each zone for calculation of snowmelt run-off. The input data requirement for SRM are as follows:

Input Data Requirement :

- maximum and minimum air temperature
- precipitation
- snowcovered area in each elevation zone
- area of watershed and its variation with elevation
- elevation of meteorological stations
- streamflow data for comparison of results

Output Product :

- streamflow

Each day during the snowmelt season, the water produced from snowmelt and from rainfall is computed, superimposed on calculated recession flow, and transformed into daily discharge from the basin according to the following equation :

$$Q_{n+1} = c_n [a_n (T_n + DT) S_n + P_n](A \; 0.01/86400)(1-k_{n+1}) + Q_n k_{n+1} \qquad (7.2.1)$$

where,

Q = average daily discharge ,in cumecs
c = runoff coefficient
a = degree day factor (cm/°C/day)
T = number of degree days
DT = temperature lapse rate correction factor (°C)
S = ratio of snowcovered area to total area
P = precipitation contribution to runoff (cm)
A = area of basin or zone (sq m)
k = recession coefficient
n = sequences of days

0.01/86400 = conversion factor ($cm.\ m^2$/day to m^3/s)

Air temperature expressed in degree-days is used in SRM as an index of the complex energy balance leading to snowmelt. At stations where hourly temperature data is available, the number of degree-days for each 24 hour period is determined by summing the hourly temperatures and dividing by 24, using 0°C as the base temperature. Where only maximum and minimum temperatures are available, the number of degree days are determined by taking mean of maximum and minimum temperatures. The degree days are extrapolated to an elevation zone by using suitable lapse rate, δ as described by equation (5.2.5).The zonal mean hypsometric elevation , h, can be determined from an area elevation curve balancing the areas above and below the mean elevation as shown in Figure 5 . When applying SRM it is advisable to conduct a regional analysis of monthly lapse rates to determine the seasonal variation (Marinec et al, 1983).

The form of precipitation (rain or snow) is determined by using a critical temperature ,T_{CRIT} . The critical temperature is usually selected to be slightly above the freezing point and may

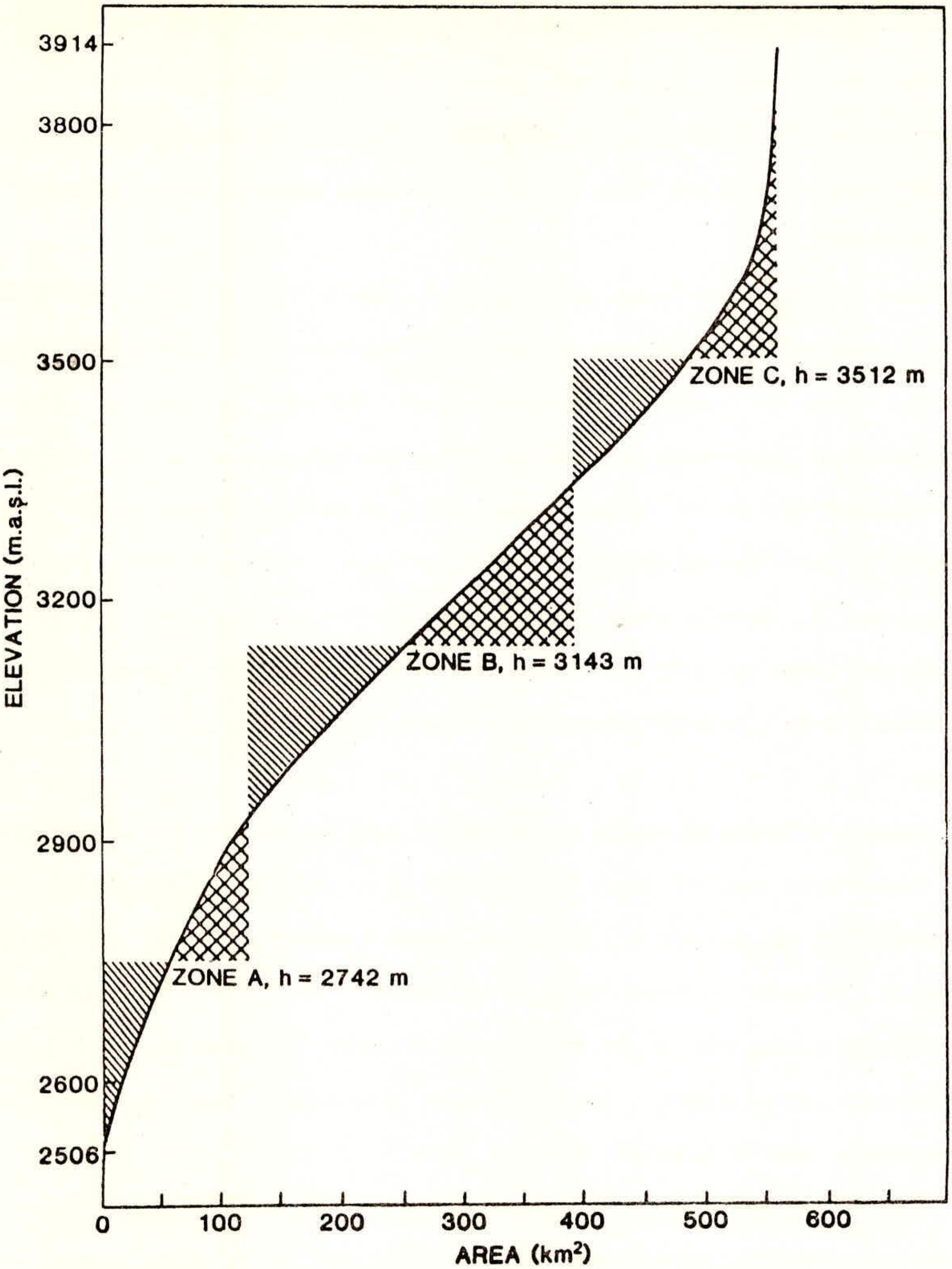


Figure 5 : Determination of zonal mean hypsometric (h) elevation using an area elevation for South fork of Rio Grande basin

vary from basin to basin. The distinction between rain and snow is considered important because rain contribution to run-off is on the same day that the rain occurs, whereas the snow contribution to run-off is delayed. The precipitation is stored in SRM and melted as soon as a sufficient number of degree-days has occurred.

The snowcovered area of a zone or basin is usually obtained from a depletion curve. A variety of sources of snowcover data may be used to compile the depletion curves including ground observations, aircraft photography and satellite imagery. If data are available, it is recommended that satellite imagery be used since it is the easiest to analyze and also quite accurate depending on basin size. The depletion curves are prepared by periodical monitoring of snowcovered area and for snowmelt run-off simulation , daily snowcover values are taken from depletion curves.

Daily discharge data from basin are required to determine the recession coefficient,and otherwise, only to determine the accuracy of simulation. The discharge preceding the start of snowmelt season (winter base flow) is to be known or estimated for initializing the model. Past continuous discharge records, if available, are useful to determine the time lag between the temperature and discharge cycles.

The other parameters used in model are run-off coefficient, recession coefficient, degree-day factor and time lag. The average value of run-off coefficient for a basin is given by the ratio of annual run-off/ annual precipitation. Because the run-off coefficient is likely to vary throughout the year as a result of changing vegetation and soil moisture conditions, the SRM program permits changes in run-off coefficient every 15 days. Usually run-off coefficient is higher for snowmelt than for rain. The degree day factor is used to convert degree days to snowmelt expressed in depth of water. The degree day factor is variable throughout the melt period because the changing properties of snow influence the melting process. Daily values of degree day factor are extremely variable so it is recommended to use a minimum of 3-5 days for averaging degree day factor.

For determination of recession coefficient daily discharge values for snowmelt season or the whole year are used. The discharge on a given day, Q_n is always plotted against the value of on the following day, Q_{n+1}. An envelope is drawn to enclose most of the points . The lower envelope line represents the extreme discharge decline i.e., the recession without any partial delay by possible precipitation or snowmelt.

The large basins with multiple elevation zones, the time lag changes during snowmelt season as result of changing spatial distribution of the snowcover with respect to the basin outlet. The time lag correction factors from n and n+1 days are changed accordingly. Such changes not only account for different time lags for each zone, but also for how they change from beginning to the end of snowmelt season. In order to obtain a more accurate split of run-off from days n and n+1, the planimetering of the areas under actual daily hydrographs has been recommended to come up with the appropriate time lag correction factor (Shafer et al 1981).

In the simulation mode , SRM produces daily discharge values from the start until the end of snowmelt period (usually 1-6 months using the actual sequence of temperatures and the depletion curves of snowcoverage obtained from snowcover monitoring. Simulations can serve not only for model testing but also serve to establish discharge series in ungaused basins. Seasonal volume simulations are obtained by summing daily flows over period of interest. Outside the snowmelt period , SRM can be operated but careful attention is required for run-off coefficients in which are included effects of evapotranspiration ,soil moisture etc which are not as important during snowmelt season. In the forecast mode of SRM ,periodical updating with actual temperatures and discharges and with recent snowcover information is desirable. The model can also be used for seasonal forecasts of the expected run-off volumes ranging from several weeks to the total duration of the snowmelt season. Such forecasts are based on medium range prediction techniques, climatological records or on statistically determined sequences of temperature and precipitation. An extrapolation snowcover depletion curves taking into account the forecasted temperatures is also required. To provide a means for forecasting run-off , it is suggested that depletion curves that normally relate the areal extent of snowcover to elapsed time be modified to relate the snowçoverage to accumulated degree days (Rango and Martinec, 1982). The functional flowchart of SRM computer program is Shown in Figure 6

The results of run-off simulation from various basins using Martinec snowmelt run-off model, in order to predict the accuracy

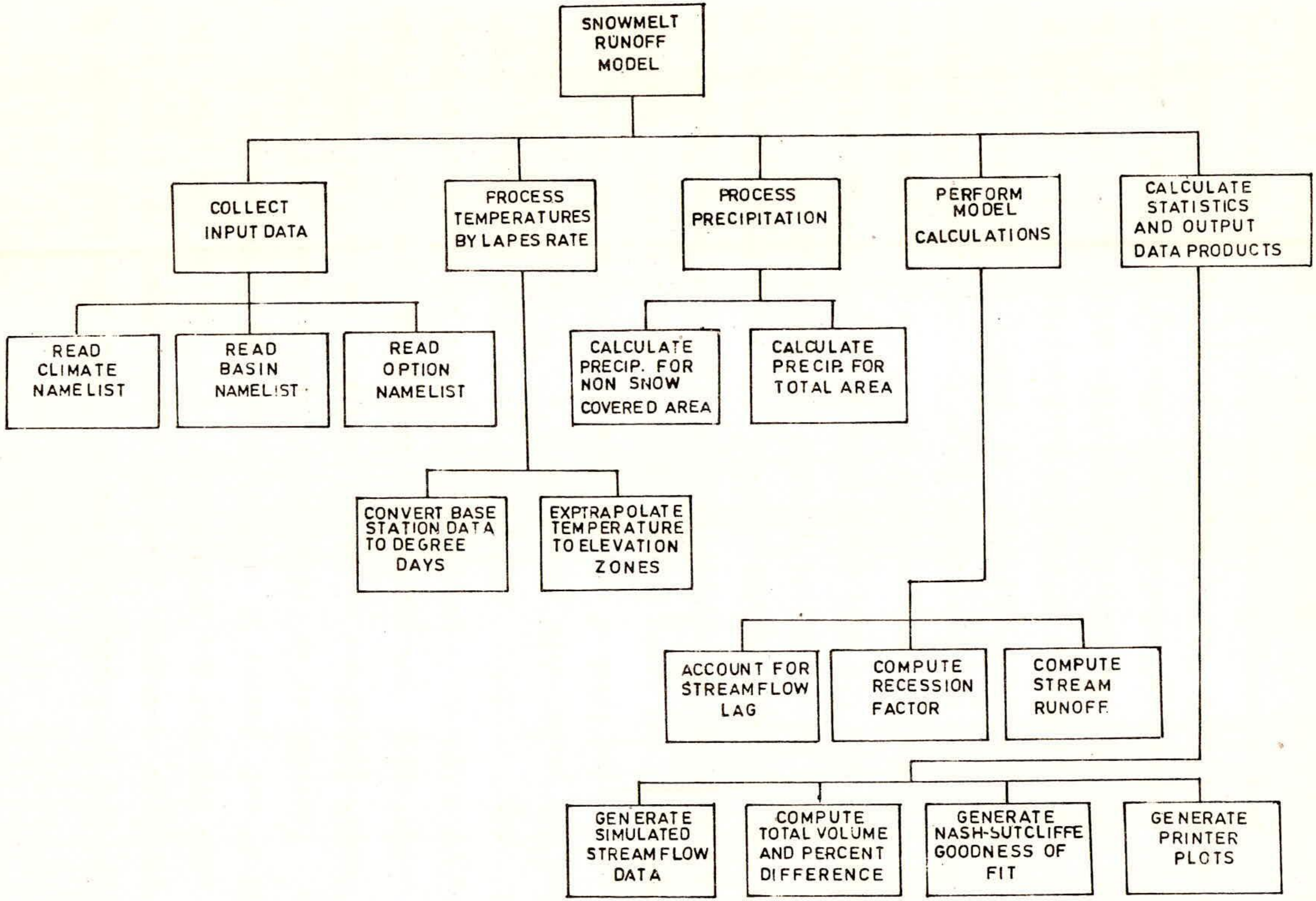


Fig. 6. Functional flow chart of the snowmelt runoff model computer program.

of simulation in future applications of the model have been reported by Martinec and Rango (1981). It has been shown that simulation accuracy depends on the quality of input data and most accurate simulation would result when (i) temperature and precipitation data are recorded at basin mean elevation,(ii) snowcover observations are available once per week,(iii) several climate stations are available for large basins and (iv) a few years of run-off record exists for determination of the recession coefficient. The availability of satellite observations of snowcover extent has been shown to be an important information for applications of model to large basins. With wide applications of satellite data for monitoring snow accumulation, the model has been refined over a period of time for applications on large basins-some in inaccessible terrain for simulation periods of only a few days to the entire year (Martinec et al 1983).

Martinec and Rango (1986) reported that the model performs well for basins ranging in size from 0.77 to 4000 Sq.km and the accuracy of prediction is generally not limited by the basin relief and climatic characteristics. However, the model simulation accuracy is observed to decrease with increasing basin size (Rango,1980). The simulations tend to be less accurate when there are significant amount of rainfall during the snowmelt period. Howley et al (1980) also concluded that SRM is most effective for basins of smaller rather than large size. It is very much suitable for mountainous basins having area less than 4000 Km with seasonal snowcover. This model is found more accurate for forecasts of 1 day or 2 days than for periods of

3,5,10, or 15 days. Delays in data collection of more than 1 day may significantly reduce the accuracy of SRM.

The model has been applied at various basins in Rocky mountains such as Dinwoody Creek, Bull Lake Creek, South Fork, Conejos river, Rio Grande and Kings river at Sierra Nevada. To asses the accuracy of model simulation in a subtropical environment and its performance on a daily basis for the snowmelt season, SRM has been applied in the Kabul basin in the Western Himalayas (63,657 Sq.km)(Dey at al (1989).

7.3 University of British Columbia Watershed Model - (UBC) : -

The UBC watershed model has been developed by Quick and Pipes (1977) at University of British Columbia, Canada. The model is designed primarily for mountainous watersheds and calculates the total contribution from both snowmelt and rainfall run-off. A separate calculation can also be made of run-off occurring from glacier covered areas. The model is designed to use sparse data networks which, generally, is found in mountainous regions. The basic structure of model depends on a division of the watershed into a number of elevation bands. The elevation increment for each band is the same and an area for each band is specified.The following data is required for UBC Watershed Model.

Input Data Requirement :

- maximum and minimum air temperature
- precipitation
- snow water equivalent at the starting date of analysis
- area of watershed and its variation with elevation
- glacier area, if any
- reservoir or lake area, if any
- impermeable area in watershed
- elevation of meteorological stations
- streamflow data for calibration and verification of computed results

Output Product :

- streamflow
- snow water equivalent in each zone
- current moisture status of the watershed including soil moisture, evaporation and quantities of water in various run-off storages.

The UBC Watershed Model contains two options for calculation of snowmelt.

1. An energy budget approach which is simplified for use when there is only temperature data available.The method is also easily usable if more detailed radiation, albedo and wind data are available. The details of simplified equations have been described in Section 5.1
2. A degree-day approach in which the resulting snowmelt formulation is specified for forested areas and for open areas.

The following equations are used for snowmelt computations using degree day method.

7.3.1 Forested Melt Formulation :

$$BM = \{ A + HM2MOF * (TD/PODIUF) + TN \} * A0MDDF \qquad (7.3.1)$$

where,

BM = band melt for a particular band mid elevation
TD = daily temperature range (TX-TN)
TX = band maximum daily temperature
TN = band minimum daily temperature
A = mean daily temperature of the band
A0MDDF= point melt factor
PODIUF= radiant energy factor
HM2MOF= energy partition multiplier

$$HM2MOF = \frac{TN + TD/PODIUF}{PODEWP + TD/PODIUF} < A0MODX$$

PODEWP= reference dew point temperature for snowmelt
AOMODX= a constant

The TD/PODIUF portion of the multiplier permits some radiation melt on days when TN is a little below freezing but maximum temperature goes somewhat above freezing.

7.3.2 Open Area Melt Formulation :

In the open areas , the radiation component of melt is much more significant, so that melt is more dependent on the daily maximum temperature than on the mean temperature. This formulation is specified above the tree-line and in open areas.

$$BMO = AOMDDF * \{ TX + HM2MOO (TN) \} * MMF \qquad (7.3.2)$$

where, MMF is specified as a monthly factor.

The watershed model accumulates precipitation falling as snow and then depletes these snowpack according to the calculated melt rate. The snow accumulation and depletion is carried out separately in each area- elevation band. There are two ways namely block budgeting and wedge budgeting in UBC in which this budgeting is carried out. In the block budget option snow is accumulated as if it was falling at the mid-elevation of each band. Snowmelt calculations are also made at this mid-elevation. The wedge budget option recognizes the final depletion of snow as a gradual recession of snowline from the bottom to the top of a band. There is smoother transition as snow is depleted completely from a band in the wedge budget.

Temperature and precipitation data are available as point values at given elevations in a watershed. The model is designed to handle data from upto maximum of three stations. Before watershed response calculations are made , these meteorological data at a point is distributed to mid elevation points of elevation bands. Two temperature lapse rates have been specified in the model, one for the maximum temperature and other for the minimum temperature. The lapse rate is calculated each day using the daily temperature range (diurnal range) as an index.

(i) Maximum Temperature Lapse Rate - TXLAPS :

$$TXLAPS = TZLAPS + (TLXM - TZLAPS) * TD/TERM \quad (7.3.3)$$

(ii) Minimum Temperature Lapse Rate- TNLAPS :

$$TNLAPS = TZLAPS - (TZLAPS - TLNM) * TD/TERM \quad (7.3.4)$$

where

$$TZLAPS = TZ - (PP/PPM) * (TZ - TZP) \quad (7.3.5)$$

for the above,

PP = daily precipitation
TXLM = 10 °C
TNLM = 0.5 °C
TZ = 6.4 °C/ km, the reference lapse rate for rain free conditions
TZP = 3.2 °C/km, the reference lapse rate when PP > PPM
PPM = 10 mm/day

The calibration parameter TERM equals to the maximum temperature range under open sky conditions. Figure 7 shows the daily temperature lapse rates as a function of daily temperature range.

The algorithms which describe the variation of precipitation with elevation are subdivided into two aspects to give the final orographic effect. The first algorithm describes the basic enhancement of precipitation with elevation barrier height if temperature is 0°C . The precipitation in any elevation band is calculated from the precipitation in the band immediately above

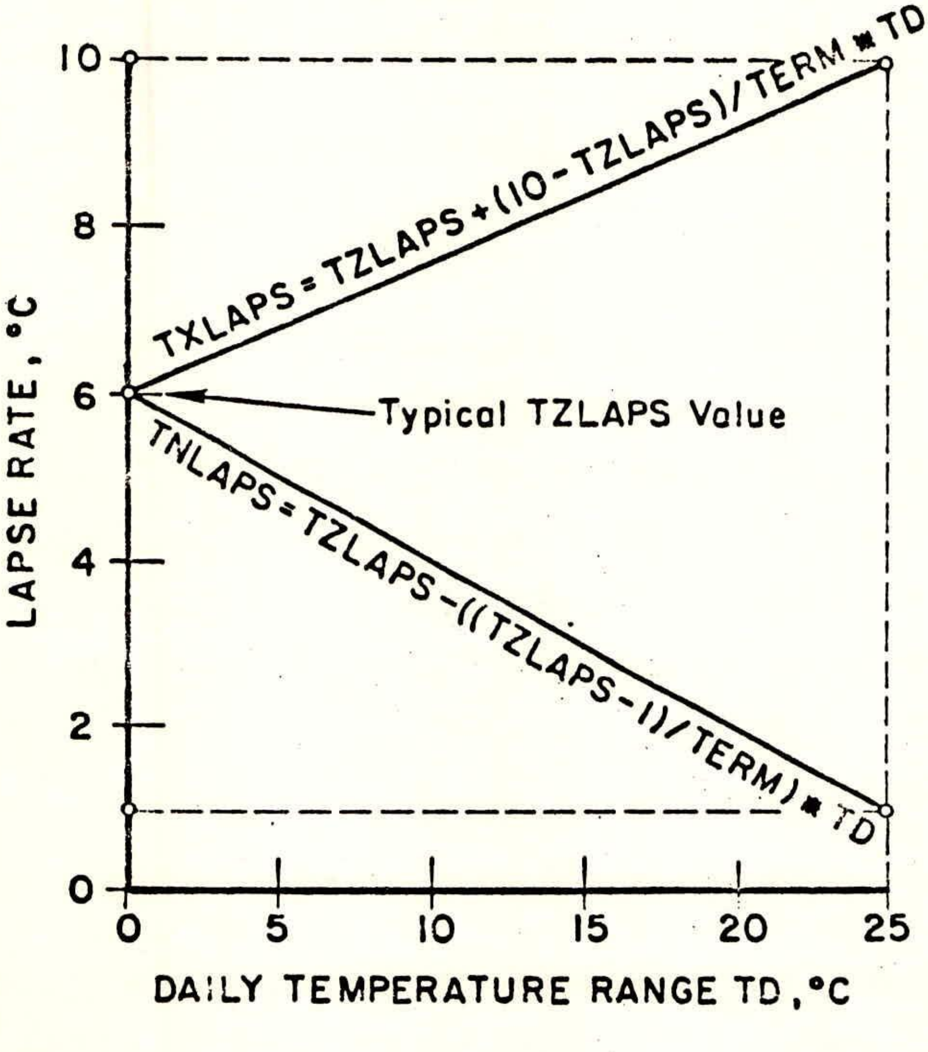


FIGURE 7: Daily Lapse Rates as a Function of

Daily Temperature Range.

(ASSUMES TERM=25°,TLXM=10°,TLNM=1°).

or below using the following equation ;

$$P_{I,J,L+1} = P_{I,J,L} * (1+ \alpha) \qquad (7.3.6)$$

where,

$P_{I,J,L}$ = precipitation from the meteorological station I for day J and elevation band L

α = precipitation enhancement factor

The second algorithm modifies α for temperature other than a base elevation band temperature of 0°C as follows :

$$\alpha = \alpha - S * (T) \qquad (7.3.7)$$

where S is the model parameter representing stability of air mass.

The form of precipitation is controlled by three logical statement and the temperature T , used in these statements is normally the mean daily temperature in each band but it can be specified to be the maximum or the minimum daily temperature in each band.

If T < 0°C all precipitation is snow

If T > A0FORM all precipitation is rain

A0FORM is specified in the parameter deck. If it is not specified or is set below 0°C , the default value of 2 °C is taken by the computer program. Between 0°C and A0FORM a proportion of precipitation will be specified by FORMPP where,

$$FORMPP = T/A0FORM \qquad (7.3.8)$$

Then rain, RN = PP * FORMPP (7.3.9)

and Snow, SN = PP * (1- FORMPP) (7.3.10)

Evaporation estimation are computed in three steps. In the first step, estimates are made of daily potential evapotranspiration for the reference meteorological station in the watershed. In the second step the potential evapotranspiration values are distributed to each elevation mid-band level. In the third step, the computed values on mid-band level , are used in conjunction with the calculated soil moisture deficit values to yield an actual evapotranspiration.

The response of watershed model to snowmelt and rainfall is controlled by a soil moisture model. The soil moisture status of each area-elevation band controls the sub division of the total snowmelt and rain input into the various components of watershed run-off response (Figure 8). The total snowmelt and rain input to each watershed band is subdivided on a priority basis. Each component of run-off undergoes delay before reaching the outflow point of the watershed.

Part of each elevation band is specified to be impermeable, so that any input of water to this area will enter as fast run-off component. Such run-off are thought of in terms of surface run-off or very superficial percolation through coarse sediments. The impermeable percentage of watershed can be varied with soil moisture deficit. The fast run-off is given the first priority in the model. Before any further run-off could occur ,other than fast run-off, soil moisture deficit must be satisfied. The soil moisture and actual actual evapotranspiration are considered in

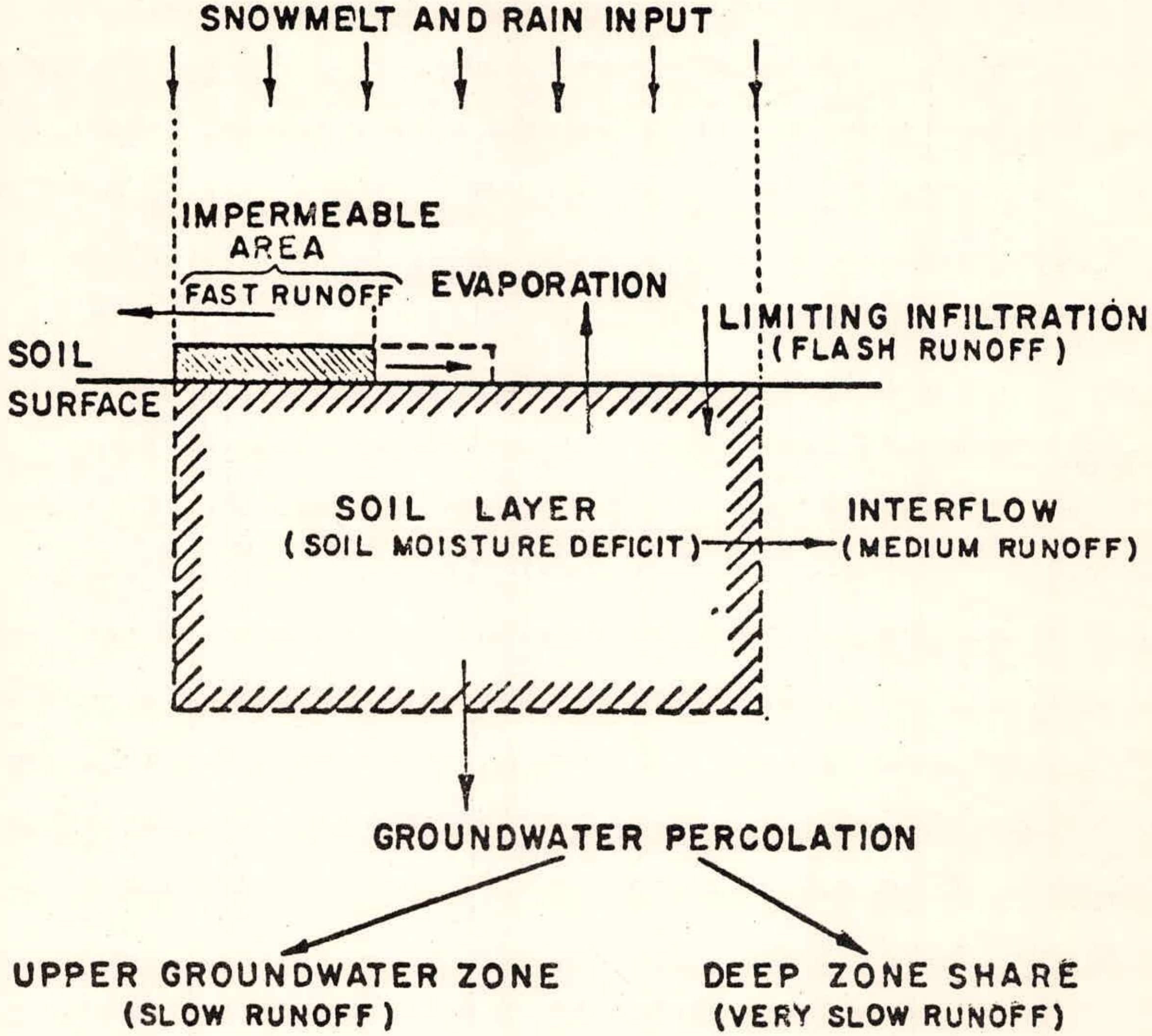


FIGURE 8 Model of Soil Layer and Subdivision of Runoff Components

second priority. The third priority is given for ground water percolation. Water which is assumed to be divided into specificiable fractions which go to the ground water components, the upper ground water and deep zone ground water components. Ground water percolation accepts any water excess upto a fixed limit. Any excess above this limit goes to the fourth priority, medium run-off or inter flow. Although this is the lowest priority, but it is also frequently the most significant run-off component during active snowmelt and rainstorms. In the model, inflow is considered to be large reservoir which receives inflows day by day during active snowmelt and rain. These inflows are the excesses remaining after satisfying soil moisture and ground water abstractions. This reservoir releases a certain factor each day, but the volume of water released do not immediately appear in the down stream channel system. Instead, this released water undergoes a convolution very similar to the fast component of unit hydrograph. This release from an interflow storage reservoir and convolution before reaching the channel outflow point, produces a much more sluggish response for this medium run-off component.

The water allocated to each of the components of run-off, are subjected to a routing procedure which produces a time distribution of run-off. The routing for each component is based on the concept of the linear storage reservoir. The fast and medium components of run-off are subjected to a cascade reservoirs which is essentially identical to unit hydrograph convolution. The slower components of run-off simply use a single reservoir. A flow diagram of UBC is given in Figure 9.

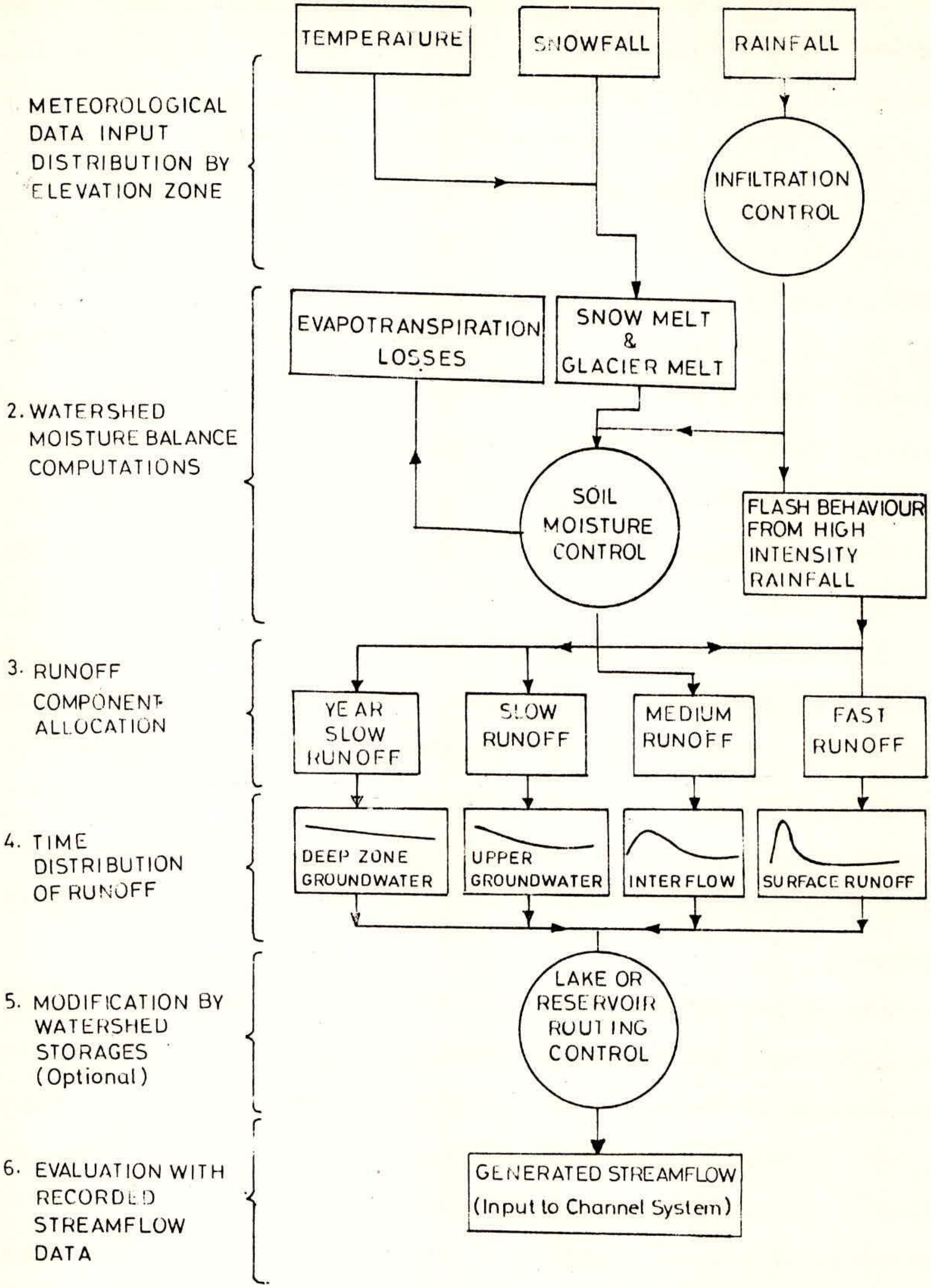


Fig. 9 . U B C Watershed Model – General Flow Chart.

The model has been used for catchment ranging from a few hundred sq.km upto areas of several thousand sq.km in both mountainous and plateau regions. The application of model includes Fraser River (215,000 sq. km), Columbia River (26500 sq. km) , Peace river (98000 sq. km), Okanagan Lake basin (6000 sq. km), South Saskatchewan river (9000 sq. km), and Canation Creek Research watershed (10 sq. km) etc..

7.4 Hydrologic Engineering Center Model- (HEC-1) :

The HEC-1 model was originally developed by Leo R Beard and other members of the Hydrologic Engineering Center staff under the US Army Corps of Engineers in 1967. This model has been designed to simulate the surface runoff response of a river basin to precipitation by representing the basin as an interconnected system of hyrologic and hydraulic components. Each component models an aspect of precipitation - runoff process within portion of basin, commonly referred as a subbasin. A component may represent a surface runoff entity, a stream channel, or a reservoir. Representation of a component requires a set of parameters which specify the particular characteristics of the component and mathematical relations which describe the physical processes. The result of the modelling process is the computation of streamflow hydrographs at desired locations in the river basin.

Input Data Requirement :

(i) For Degree Day Method :

- air temperature
- precipitation
- snow water equivalent
- area of catchment and it´s variation with elevation
- streamflow data for comparison of the results

(ii) For Energy Budget Method :

- solar radiation
- precipitation
- snow water equivalent
- wind velocity
- average albedo of snow surface
- dew point temperature above the snow surface and on the snow surface
- forest cover area
- streamflow data for comparison of results

Output Product :

- streamflow

For the snowbound catchments , there is a provision for separate computation of snowmelt in up to ten elevation zones within the subbasin. These zones are usually considered to be in elevation increment of 1000 ft., but any equal increments of elevation can be used as long as lapse rate (TLAPS) corresponds to the change in elevation with zones. The input temperature data are those corresponding to the bottom of the lowest elevation zone. Temperatures are reduced by the lapse rate in degrees per increment of elevation zone. The base temperature (FRZTP) at which melt will occur , is specified because variation from 32 °F or 0°C might be warranted considering both the spatial and temporal fluctuations of temperature within the zone. Precipitation is considered to fall as snow if the zone temperature (TMPR) is less than the base temperature (FRZTP).

plus 2 degrees. Melt occurs when the temperature (TMPR) is equal to or greater than the base temperature. Snowfall melt is substracted and snowfall is added to the snowpack in each zone. The computation of snowmelt is made either by degree day method or by the energy budget method. The basic equations for snowmelt computations are taken from US Army Corps(1956). The energy budget equations have been simplified for use in this model.

7.4.1 Degree Day Method :

The snowmelt by this method is computed using the following equation

$$SNWMT = COEF\ (TMPR - FRZTP) \qquad (7.4.1)$$

where,

SNWMT = snowmelt in inches (mm) /day in the elevation zone
TMPR = air temperature in °F or °C lapsed to the mid point of the elevation zone
FRZTP = temperature in °F or °C at which snow melts

COEF = melt coefficient in inches (mm)/degree/day

7.4.2 Energy Budget Method :

The computation of snowmelt is made using simplified snowmelt equations of energy budget. For heavily forested areas during rain the snowmelt is estimated the by the following equation :

$$SNWMT = COEF\ (\ 0.09 + (\ 0.029 + 0.00504\ WIND + 0.007\ RAIN\ )\ (TMPR - FRZTP) \qquad (7.4.2)$$

For melt during rainfree periods in partly forested areas (50 % forest cover) the following equation is used.

$$\text{SNWMT} = \text{COEF} \; (0.002 \; \text{SOL} \; (1 - \text{ALBDO}) + (0.0011 \; \text{WIND} + 0.0145) \; (\text{TMPR} - \text{FRZTP}) + 0.0039 \; \text{WIND} \; (\text{DEWPT} - \text{FRZTP}) \tag{7.4.3}$$

where,

SNWMT = melt in inches/day
TMPR = temperature of mid elevation zone in °F
FRZTP = base temperature in °F
DEWPT = dew point temperature in °F lapsed at a rate 0.20 TLAPS to the mid point of elevation zone.
COEF = a dimesionless coefficient to account for variation from general snowmelt equation
RAIN = rainfall in inches/day
SOL = solar radiation in langleys/day
WIND = wind speed at 50 ft above the snow surface in miles/hour
ALBDO = albedo of snow, $0.75/D^{0.2}$, constrained above 0.4, where D is the days since the last snowfall

It is reported HEC-1 (1981) the program has similar equations for the metric system which use the same variables with coefficients relevant to metric units. The average weighted precipitation is used in this model. Similar equations for energy budget approach have been used in SSARR model.

The computation of loss is made by the following empirical equations which relate loss rate of rainfall intensity / snowmelt and accumulated losses. Accumulated losses are representative of soil moisture storage.

$$\text{ALOSS} = (\text{AK} + \text{DLTK}) \; \text{PRCP}^{\text{ERAIN}} \tag{7.4.4}$$

$$\text{DLTK} = 0.2 \; \text{DLTKR} \; (1 - (\text{CUML}/\text{DLTKR}))^2 \quad \text{for CUML} \le \text{DLTKR} \tag{7.4.5}$$

$$\text{AK} = \text{STRKR} \; / \; (\text{RTIOL})^{0.1 \; \text{CUML}} \tag{7.4.6}$$

where,

ALOSS = potential loss rate in inches (mm) per hour during the time interval
AK = loss rate coefficient at the beginning of time interval
DLTK = incremental increase in AK during the first DLTKR

The cumulated loss is determined by summing the actual losses computed for each time interval.

The SCS unit hydrograph technique is used in subbasin runoff component to transform rainfall / snowmelt excess to subbasin outflow. A unit hydrograph can be directly input to the program or a synthetic unit hydrograph can be computed from user supplied parameters. HEC-1 sets automatically the duration of unit excess equal to computation interval selected for catchment simulation.

7.5 National Weather Service Snow Accumulation and Ablation Model (NWSRFS):

National Weather Service Snow Accumulation and Ablation Model (NWSRFS) has been developed by National Weather Service, USA for flood forecasting, water supply and lowflow forecasting and was used first time in 1973. This model simulates accumulation and ablation of snowcover using energy exchange at air-snow interface indexed by air temperature. The snowmelt is computed differently for rain and non-rain period. Areal extent of snowcover is related to areal water equivalent.

Catchment runoff can be modelled on a lumped basis or can be subdivided. Subdivision can be by elevation zones or can be based on physiographic factors such as forest cover , soil type etc. Generally , catchments are modelled on a lumped basis when elevation range is less than about 1500 m . The maximum number of elevation zones ever used in this model were three. The use of a snowcover area depletion curve for each sub area eliminates the need for a large number of elevation zones.The following are the data requirement for this model.

Input Data Requirement :

- point or mean areal air temperature
- point or mean areal precipitation
- snow cover area
- snow water equivalent (optional)
- potential evapotranspiration (optional)
- drainage area and its variation with elevation
- vegetation coverage in the watershed

Output Product :

- areal percentage of snowcover area
- water equivalent of snow
- mean areal snowcover area
- streamflow

Temperature lapse rates are determined from available data. Generally, mean monthly maximum and minimum temperatures for all stations are plotted against elevation. The monthly maximum and minimum temperature lapse rates determined from these plots are used in computing mean areal temperatures for each catchment or elevation zone. The mean areal temperatures are computed prior to model calibration.The mean areal precipitation is also computed from station data prior to model calibration. In flat areas, Theissen weights are used. In mountainous areas, station weights and the distribution of precipitation with elevation are generally based on a isohyetal analysis. In many areas a seasonal isohyetal analysis is used as the distribution of precipitation, especially with elevation , varies significantly from winter to summer. A correction factor is used to adjust snowfall for gauge catch deficiencies on an areal basis. This correction factor also implicitly includes losses or gains resulting interception of snow, sublimation and redistribution across catchment boundaries. Observed water equivalent can be used to update computed values when available.

The form of precipitation can be specified for each time interval if known or a critical temperature can be used to estimate the form of of precipitation if it is not known. When air temperature is greater than critical temperature, the precipitation is assumed to be rain and when air temperature is less than critical temperature , the precipitation is treated as snow. Critical temperature is constant for a given catchment.

An areal depletion curve is used to estimate the areal extent of the snowcover. The depletion curve is a plot of the areal extent versus a ratio indicating how much of the original snow cover remains. The ratio of the current mean areal water equivalent to an areal index, is used. The areal index is the smaller of (i) the maximum water equivalent that has occurred during the accumulation period ,or, (ii) a preset maximum. The preset maximum, a model parameter, specifies the water equivalent above which 100 percent areal cover always exists. If new snow occurs when area is partially bare, the area reverts to 100 percent cover for some period and then return to the depletion curve. The areal depletion curve not only accounts for areal extent of snowcover, but implicitly accounts for changes in melt factor as the areal coverage of snow decreases. Observed areal snowcover can be used to update computed values when available. Cold content and freezing of liquid water in snow are lumped together in the heat deficit. A model parameter species the maximum liquid water content holding capacity.

Percolation of the liquid water is based on empirical relationships which first lag and attenuate the excess surface melt plus rain. The transmission rate is a function of the amount of excess surface water and the water equivalent of snow cover. The Sacramento soil-moisture accounting model has been used to convert snowmelt plus rain into run-off. A unit hydrograph is used to translate the run-off to the gauging stations. The NWSRFS contains flexibility to use the same subdivision for transformation model as used for the snow model or to change the subdivisions. An automatic parameter optimization option is also available in the program.

There is no limit on size of drainage area, range of elevation or snowcover conditions to apply NWSRFS Model. It has been tested in 20 basins in USA representing a wide variety of climate and snowcover conditions.

7.6 Snowmelt and Rainfall Runoff Model - A USSR Model :

For a number of years work has been carried out at the Water Problems Institute of USSR Academy of Sciences aimed at developing a system of physically based hydrological models allowing to predict components of the water balance for river basins (primarily river runoff) on the basis of meteorological data basin parameters. These investigations have resulted in the construction and verification of models. A physically based model of the formation of snowmelt and rainfall runoff developed by Kuchment et al (1986) deals with snow cover formation, snow melting including snow accumulation, changes of its physical characteristics, heat and moisture transfer in snow, melt water formation and its vertical movement in the snowcover. It also considers heat and moisture transfer during soil freezing, water infiltration into frozen soil, vertical moisture transfer in non frozen soil and soil moisture evaporation, overland flow, subsurface flow, ground water movement and the interaction of surface and ground water on the river slope and in river channel and unsteady flow in the river channel system. The empirical formulae are used to make appropriate calculations of the hydro and thermo physical characteristics.

Input Data Requirement :

- air temperature, humidity, cloudiness, wind velocity, solid and liquid phase precipitation measured every 6 hour
- snow water equivalent measured in the end of snow season
- soil temperature and moisture in a 1 m layer measured in the end of snow season
- streamflow data for comparison

Output Product :

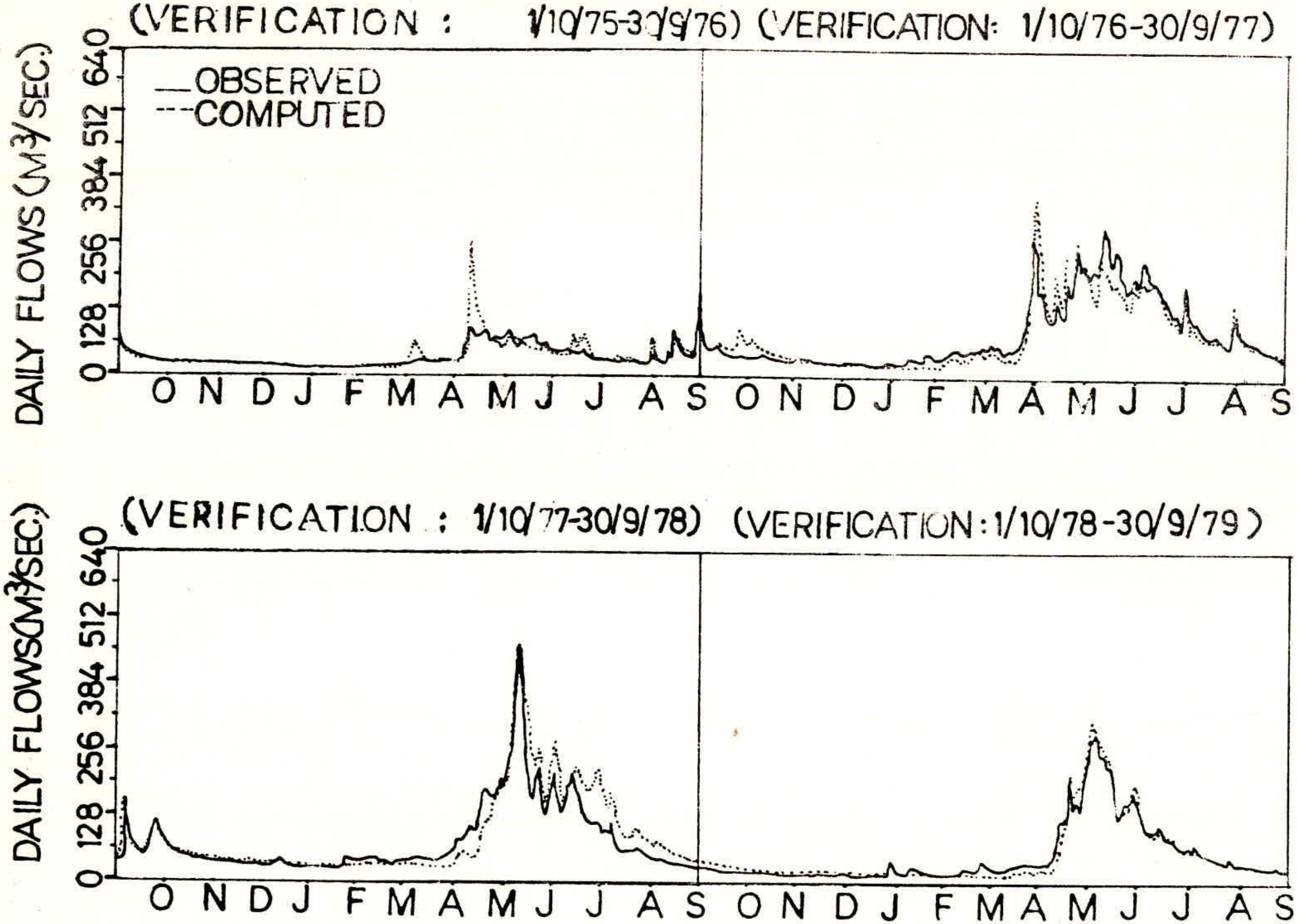
- streamflow

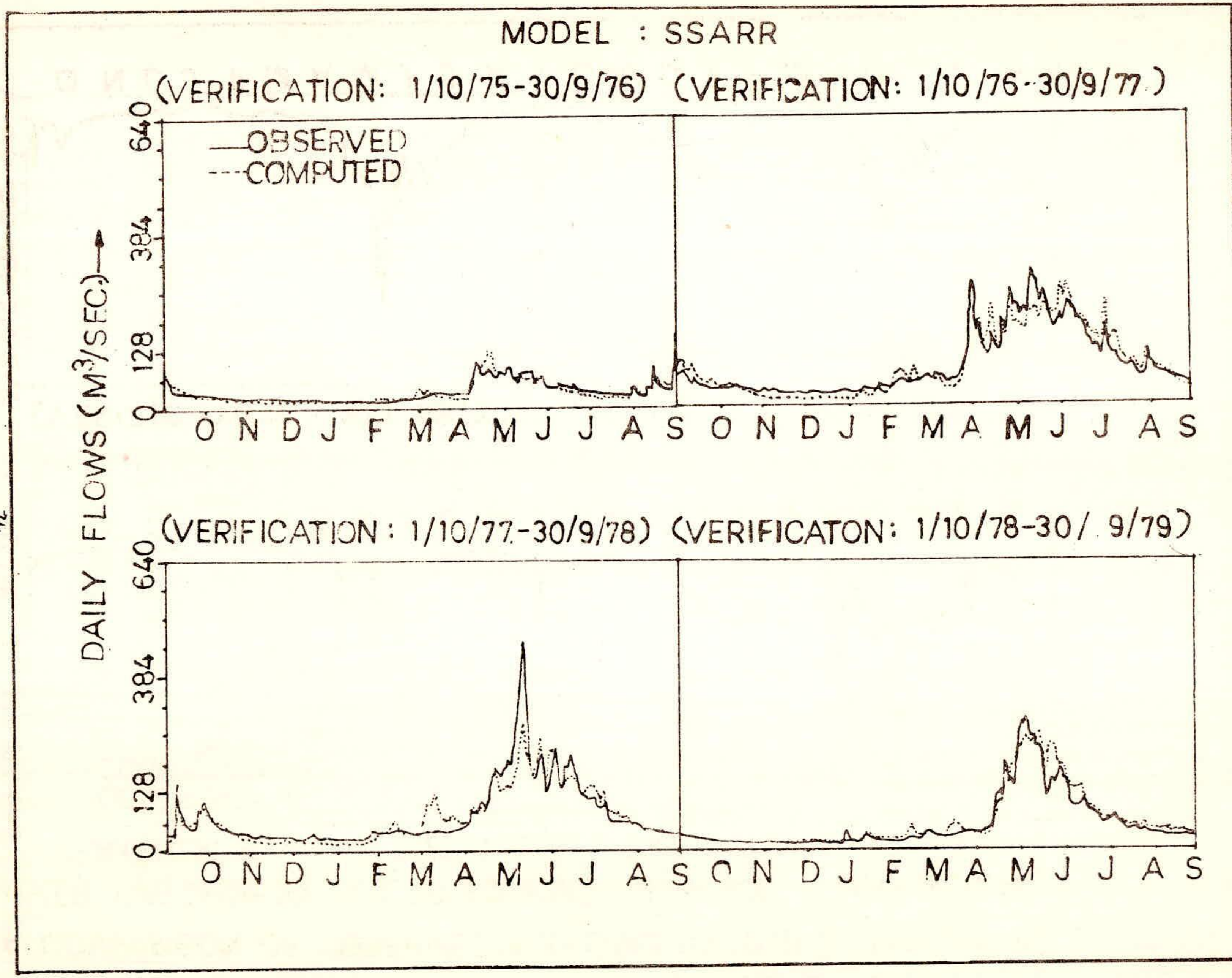
The computation of snowmelt is made dividing the watershed into various grids and dividing into various zones according to the type cf soils. The experience in calculating the processes of snowmelt and rainfall runoff formation shows that the meso structure of relief and small scale variations in soil characteristics on a subgrid scale play an important role in runoff formation. Relatively small depressions on the hillslope , not depicted on the maps, can form zones of water accumulation and cause considerable losses in the runoff. Sub-grid scale effects are of much importance in the description of snowmelt runoff formation. Relief depressions,when are not taken into account, where snow can accumulate and zones flow can not occur, can entail significant errors both in volume and form of the hydrograph. These factors are taken into account in this model. The distribution of snow water equivalent and depth of freezing are used to define the areas of noflow to calculate snowmelt runoff.

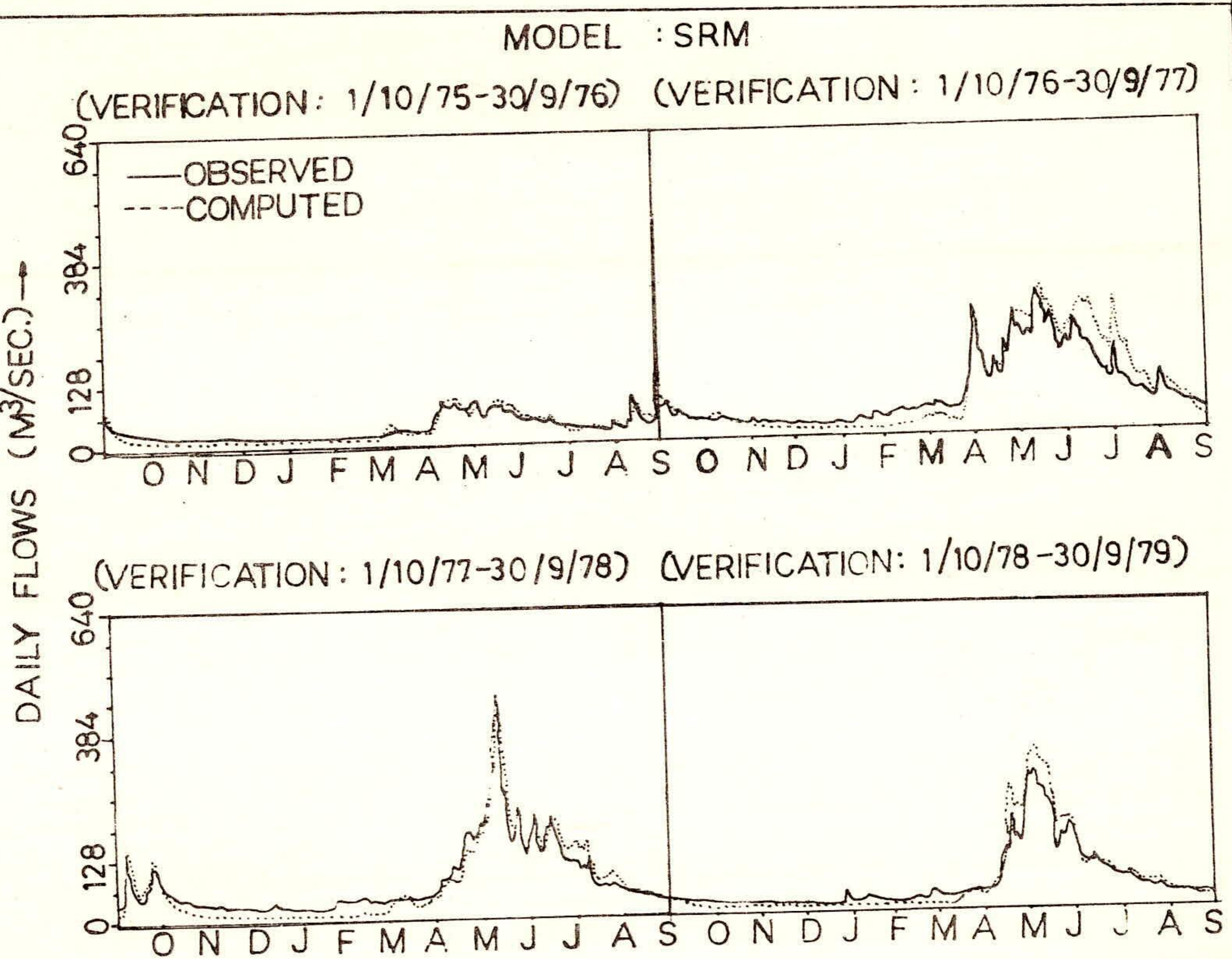
The range of hydrological phenomena that can be described using above model is wide and as a rule it proves expedient to use certain modifications of the model system for the solution of particular problems. These versions allow to simulate some of these processes, taking into account the type of initial data and required accuracy and detail of the calculations. The model has been tested in the Sosna river basin (16300 sq.km)

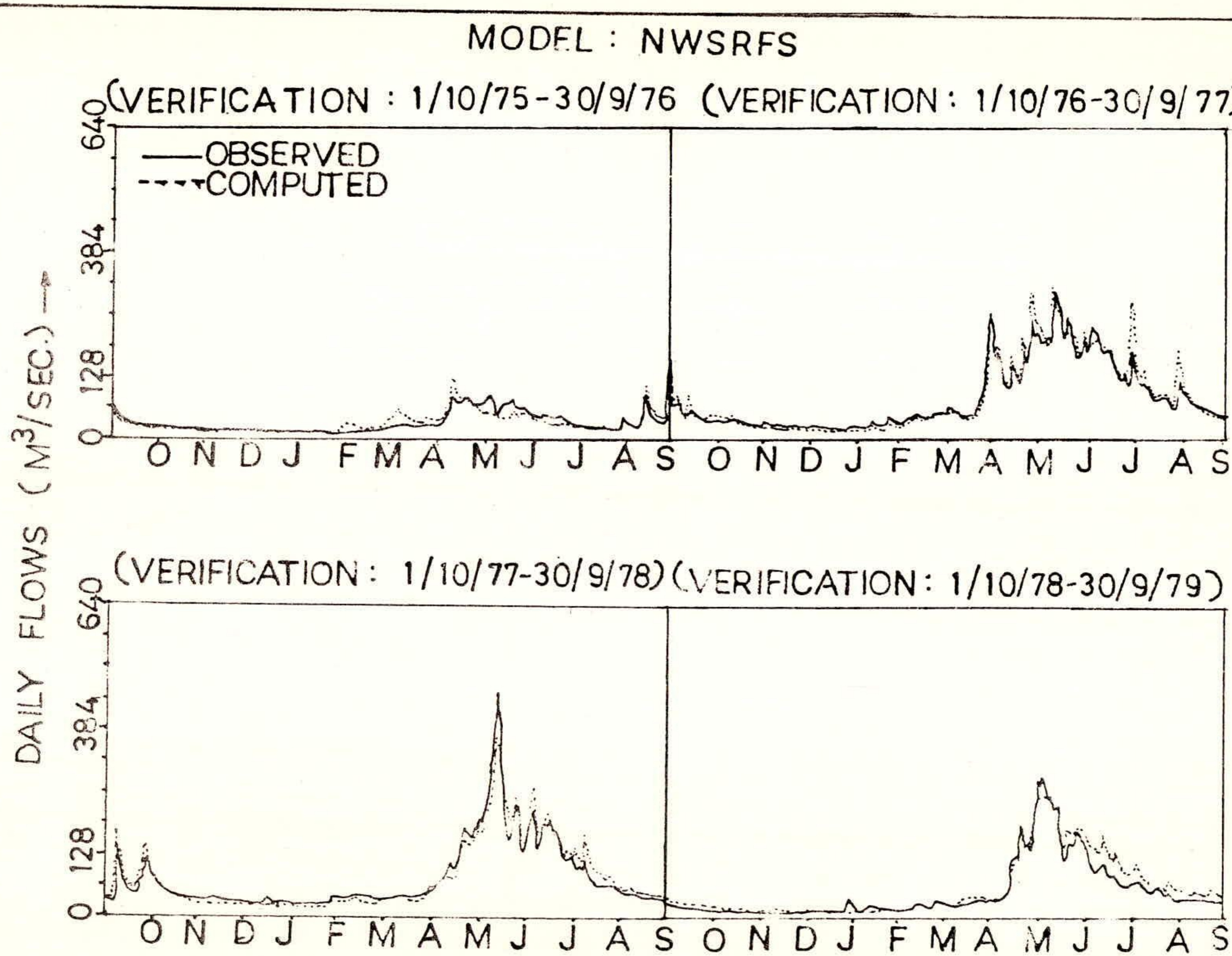
An intercomparison of eleven snowmelt models including UBC, SRM, SSARR and NWSRFS has been made by WMO(1986). These models were tested for six river catchments from climatologically and geographically varied conditions. These catchments were Durance river basin (2170 sq.km, France), W3 watershed (8.4 sq.km USA), Dunajec river basin (680 sq. km, Poland), Dischma basin (43.3 sq. km, Switzerland), Illecillwaet river basin (1100 sq km, Canada) and Kultsjon basin (1110 sq km, Sweden). Each data set consisted of two distinct periods : a calibration period (six years) and a verification period (4 years) following immediately thereafter. The simulated discharges produced by the tested models for both calibration and verification periods in each data set were evaluated and compared using graphical and numerical verification criteria.However on the basis of available information the model were not ranked or classified in order of performance. A comparison of simulated and observed discharges for few models in the Durance river has been shown in Figure 10.

FIG. 10 A COMPARISON OF OBSERVED AND SIMULATED STREAMFLOW IN DURANCE RIVER CATCHMENT (2170 SQ. KM.) FRANCE MODEL: UBC




MODEL : SSARR
(VERIFICATION: 1/10/75-30/9/76)
(VERIFICATION: 1/10/76-30/9/77)
OBSERVED
COMPUTED
DAILY FLOWS (M³/SEC.)
640
384
128
0
O N D J F M A M J J A S O N D J F M A M J J A S
(VERIFICATION : 1/10/77-30/9/78)
(VERIFICATON: 1/10/78-30/ 9/79)
640
384
128
0
O N D J F M A M J J A S O N D J F M A M J J A S


MODEL : SRM
(VERIFICATION : 1/10/75-30/9/76)
(VERIFICATION : 1/10/76-30/9/77)
(VERIFICATION : 1/10/77-30/9/78)
(VERIFICATION : 1/10/78-30/9/79)
——OBSERVED
----COMPUTED
DAILY FLOWS (M3/SEC.)→
0
128
384
640
O N D J F M A M J J A S O N D J F M A M J J A S


MODEL : NWSRFS
(VERIFICATION : 1/10/75-30/9/76
(VERIFICATION : 1/10/76-30/9/77)
(VERIFICATION : 1/10/77-30/9/78)
(VERIFICATION : 1/10/78-30/9/79)
OBSERVED
COMPUTED
DAILY FLOWS (M3/SEC.)
0
128
384
640
O N D J F M A M J J A S O N D J F M A M J J A S

8.0 SNOWMELT STUDIES FOR HIMALAYAN RIVERS :

The importance of snowmelt studies was felt first time in our country at the Annual Research Meeting of the Central Board of Irrigation (CBI) in August,1945, where late Shri Kanwar Sain presented a paper entitled ´ The role of glaciers and snow in hydrology of Punjab rivers´. In this paper ,the necessity for undertaking snow and glacier studies for forecasting run-off in the Himalayan rivers in the critical summer months of April, May, June, was stressed. As a result of discussions on this paper, it was decided to invite Dr. J E Church, then President of International Commission on Snow and Glaciers and an authority on snow surveying, to initiate snow surveys in the Himalayas for assessing the need of snowmelt forecasting in the summer months. From the experience gained from the snow surveys conducted in the eastern Himalaya during the spring seasons of the years 1947,1948, and 1949, it was seen that snow surveying has little utility in this region for snowmelt forecasting. It was reported that up to altitudes as high as 16000 ft. (asl) whatever snowfall takes place in the winter months does not accumulate but melts away rapidly due to prevailing high temperatures in this part of Himalayas.

The snowmelt studies carried out in the Himalayan region are broadly categorized as studies related with regression analysis , empirical relationships and application of snowmelt simulation models.

8.1 Development of Regression Relationships :

In most of the studies in this category, regression analysis has been made to correlate snowcover area and runoff. Efforts have also been made to correlate winter snowfall and snowmelt runoff. Generally snow cover area has been assessed from satellite imageries. The low resolution meteorological satellite data and photo interpretation techniques were used by Rango et al (1977) to map snow covered areas during early April over the Indus river and Kabul river basins in Pakistan using data of 1969-1973. The early spring snowcovered area was significantly related to April through July 31 , stream flow in regression analysis for each watershed. Predictions of 1974 seasonal streamflow using the regression equations were found within 7% of the actual flow.

The relationship between snowcover area and run-off of the Beas basin has been studied by Gupta et al (1982). Snowcover area was mapped from Landsat images for a number of years in various subcatchments. The snowcover area and subsequent run-off in different sub-basins was found to be well correlated. It has been interpreted that there have been years of uniformly heavier and lighter snowfall all over the basin and snowmelt discharges have consequently systematically varied. For a particular sub-catchment, the relationship between snowcover area and snowmelt run-off seems to be independent of geographic factors like solar radiation, catchment orientation and relative location. On the other hand, it appears to depend on geomorphological factors such as size of sub-catchment, permanent snowcover area, average altitude, lithology and stream order.

Jeyram and Bagchi (1982) estimated snowline altitude and snow cover using Landsat imagery for Tons basin in Himalayan region. A relationship amongst snowline of Beas ,Ravi, and Tons has been observed. Also, using Landsat imagery a relationship between the snowcovered area and time from beginning of snowmelt season (1st April) has also been established for Tons basin.

Lean season discharge of Sainj river, a tributary of river Beas has been studied by Krishna (1983) in reference to winter snowfall and discharge to establish the relationship between two variables. The studies have revealed that both these parameters have a fairly high correlation coefficient of 0.906. Based on this study, a simple linear regression model was evolved to forecast three to four months in advance, the lean season discharge of Sainj river solely on the basis of winter snowfall. To verify the validity of model, developed on the basis of data for the years 1967-68 to 1978-79, expected lean season discharge for the years 1979-80 to 1981-82 were computed against the observed discharge and were found to exceed 5%, 9% and 23% respectively. The studies have also brought to light that on an average 31% of annual flow of river Sainj is derived from seasonal snowcover. It was also concluded that in the absence of snow course data, total winter snowfall at a place is good indicator of lean season discharge of the nearby rivers.

Dey and Goswami (1983) have presented results of studies involving utilization of satellite snowcover observations for seasonal streamflow estimates in Western Himalayas. A regression model relating seasonal flow from april through July, 1974 to early April snowcover explained 73% and 82% of variance,respectively , of measured flows in Indus and Kabul rivers. It has been shown that remotely sensed snowcover area data provides the best available input in empirical snowmelt prediction techniques for the remote Himalayan basins. The study has also indicated high correlation of concurrent flows in adjoining Himalayan basins like Indus and Kabul.

A regression model using percentage of snowcovered area of Satluj basin above Bhakra and seasonal snowmelt run-off (April-June) for years of 1975-1978 was developed by Ramamoorthi (1983,1984,1986) at National Remote Sensing Agency. The delineation of snowcovered area for basin under consideration was made from NOAA imageries of the month of April. This model was used in 1980 to predict the seasonal snowmelt run-off of Satluj. For the forecast the percentage snowcovered area on the days of very heavy snowfall during January to March 1980 were extracted from judiciously selected cloud free NOAA imageries. At the end of June 1980 it was found that the difference between the forecast quantity and observed flows was 9%. This model was subsequently modified while giving seasonal snowmelt run-off for the years of 1981 and 1982.The forecasts as per the revised model were found to be within 10% than the actual observed flows. The seasonal flow forecasts were made for other years also. BBMB (1988) reported that forecasts made by NRSA and the actual flows during period April-June for 5 years period from 1980 to 1984 varied from 18% to 49%. in different years. In the year 1988 NRSA forecast of snowmelt run-off was 8.6% less than the observed flows. Similar snowmelt run-off forecasting models have been developed using NOAA satellite data and used in predicting the snowmelt run-off of Ganga at Devaprayag in the years of 1981 and 1982 which were within 10% difference from actual.

A model of snowcover area versus run-off against a concurrent flow correlation model in the Western Himalayas has been evaluated by Dey and Goswami(1984), using data of Satluj,Indus, Kabul and Chenab rivers. It was found that the concurrent flow correlation model explains more than 90% of the variability of flows of these rivers, while the snowcover model explains somewhat less of the variability in flows. It is mentioned in the study that these rivers carry significant amount of snowmelt run-off, which on an average,account for more than 55% of the mean annual flows. The mean seasonal snowmelt run-off (April to June) in Indus, Kabul,Satluj and Chenab rivers are given as 4027, 851735 , and 1508 cumecs respectively for catchment areas of 162100,88600,38000 and 26155 sq.km. The following relationship between snow cover area and seasonal run-off have been established.

(i) $Y = 0.06493\ X - 0.363325$ for Satluj river (8.1.1)

(ii) $Y = 0.472\ X + 4.73895$ for Indus river (8.1.2)

(iii) $Y = 0.54337\ X - 5.24243$ for Kabul river (8.1.3)

Where,

Y = Seasonal run-off (April-July) in $10^9\ m^3$

X = Average percent of snowcover of the basin

The importance of permanent snowcovered area in any study of snowmelt in Himalayan basins was brought out by Furguson (1985). A study was carried out for glacierized mountains (upper Indus in Pakistan) and a model was developed for annual variation of run-off and its forecasting . The approach is based on identification of a number of glaciological and climatological factors other than snowcovered area. Neglecting the rainfall run-off and ground water discharge , and also losses ,the total melt water run-off has been assumed to be sum of three components : (i) complete melting of a glacier snowpack, (ii) complete melting of glacier ablation snow cover, and (ii) glacier ice melt from a contributing fraction of area. The useful information is provided about characteristics of high mountain basins in Himalayan region based upon 1975-1978 data. The details of the results are given in Table 3.

Table 3: The Snowcover and Run-off Characteristics

S No.	River basin	Area sq. km	Run-off mm(mean) Apr-Aug.	Snowcover (%)	Icecover mean estimates (%)
1.	Hunza	13000	763	88	38
2.	Gilgit	26000	578	86	27
3.	Indus	160000	303	83	11
4.	Shyok	33000	292	93	9
5.	Jhelum	25000	644	74	2

Roohani (1986) related the information about extent of snowcover obtained from the Landsat MSS images for months of March to June with the snowmelt run-off assumed as total flow minus base flow for different sub-basins of Chenab basin. A general linear relationship has been obtained . It has been found that as the catchment size increases the regression lines , fitted for snowcovered area and subsequent premonsoon cumulative run-off are successively right shifted along the X-axis. There is also a systematic variation in slope of the regression line for different sub-basins. This is related to interplay of several factors like catchment area, permanent snowcover area , average altitude of the sub-basin , relief and channel slope,all of which can be generally considered together in terms of a single parameter i.e. stream order . The relationship obtained after the analysis of available information about snowcover from satellite images is useful in predicting subsequent snowmelt run-off in sub-basins of Chenab basin cumulated up to June 30, if the snowcover area is known at any stage after the end of snow accumulation . In this study, the analytic approach of SSARR model was found appropriate for expressing the relationship between snowcover depletion of any sub-basin and corresponding snowmelt run-off. The simple model structure based upon split watershed approach, subdividing it into permanent snowcovered, temporary snowcovered and snow free areas, and also dividing into melt and nonmelt areas using daily data of temperature and lapsed at a rate of 6.5 °C /km, change in elevation, has given a good performance in simulating daily flows. Base temperature of 0°C and rain freeze temperature of 0.56 °C have been found appropriate . The melt of snow has been computed using simple degree-day method, assuming initial melt rate on 1st March as 0.38 mm/°C/day and varying (increasing) it linearly using appropriate rate for sub-basin.

Mohile et al (1988) carried out a study to develop a regression relationship between temperature of Kaza and snowmelt runoff collected at a proposed dam across Spiti river 4 km upstream of Kaza at an elevation of about 3639 m in the Satluj basin. The extension of discharge data at the proposed dam site was made based on this relationship. Efforts were also made to develop a relationship between discharges of Kaza and Namgia in the same basin.

The correlation between snow temperatures and air temperatures has been established by Upadhyay et al (1981). However, these relations are valid for the months of January and February when air is found normally to be colder than snow surface. The study on heat transfer in seasonal snowpack has indicated that heat transfer from one part to the other part by conductions is restricted only to a few centimeters because of poor thermal conductivity of snow crystals. The bottom layer remains always near 0°C owing to ground heat whereas the upper layers of snow cover are influenced by temperature of atmosphere and exhibit diurnal variation.

8.2 Application of Empirical Relationships :

Several snowmelt studies using empirical relationships between temperature and snowmelt runoff have been made. Such empirical relationships are generally a function of degree day factor and snow cover area. Thapa (1980) estimated snowmelt by considering melt due to the influence of temperature and rainfall in the snowcovered area for Beas catchment up to Larji. A technique for estimation of snowmelt run-off during premonsoon period was studied. An attempt has been made to study the relationship between snowcover,acquired with the help of satellite imageries, and the cumulative discharge of the months of March,April and May of the year 1973,1975,1976 and 1977. An exponential trend has been observed. Efforts have also been made to identify and delineate the vegetal cover and land use features using visual interpretation technique. Due to availability of the limited meteorological and hydrological data , the study for estimation of snowmelt run-off has been confined to the sub-basin upstream of Manali. The sub-basin has been divided into permanent and temporary snow covered zones. The degree-day method and the melt due to rainfall on snow have been used to estimate to snowmelt run-off. With several trials the degree day factor of 0.0018 cm/°C/day for March and April and 0.00315 cm/°C/day for May have been assigned for the years 1977, 1978 and 1979. The snowmelt is then routed from various zones to the outlet of Manali catchment taking into account the recession coefficient as 0.90. The excess rainfall run-off from temporary and permanent snow covered zones and excess rainfall from the non snow covered zones also included. The run-off coefficients of o.595 and 0.275 have been calculated for the rainfall on the snowcovered and non snowcovered areas respectively. Since the time of concentration in the catchment is less than one day the run-off due to rainfall on the non snowcovered area is superimposed on the same day on the calculated hydrograph obtained from the snowmelt and the run-off from the rain falling on the snow covered area.

Bagchi (1981) carried out a study of snowmelt run-off in Beas basin using satellite imageries. For determination of snowmelt temperature index method was used assuming a lapse rate of 6.5°c/km. The value of degree day factor was considered as 2.1mm /°C/day . For finding effective precipitation at different altitudes in the Beas catchment, a coefficient as an orographic increase factor was used. It was shown that this coefficient increased from unity to 3.25 with a change of altitude from 1900m to 4000 m and then decrease to 0.9 at 5900 m. The percentage of snow in total precipitation has been assumed as a function of minimum daily temperature of the station.

An empirical model for prediction of snowmelt run-off in Satluj basin has been used by Upadhyay et al (1983). The empirical model for the computation of snowmelt run-off has been presented as a function of degree day factor. A techniques for computing total discharge with the input of rainfall, area of homogeneous state with a given rainfall, soil moisture deficiency, ambient air temperature ,wet bulb temperature, wind, altitude of freezing level, glaciated area and albedo has been evolved. Analyzing past 50 years data of rainfall and discharge of the basin, snowmelt and rainfall components have been presented as two distinct series. Upadhyay et al(1983) also analyzed the various components of energy input to a snowcover and monthly budget for net energy available for snowmelt have been worked out for a number of stations in Himalaya. Estimations of short-wave radiation, long-wave radiation, convective transfer and latent heat of condensation have been made by indirect approach using meteorological data on temperature, vapour pressure, wind and cloudiness. Chatterji and Chopra (1976) have studied snowmelt contribution in Satluj catchment for the purpose of flood and low flow forecasting of Bhakra reservoir. The average value of degree day factor was assumed as 0.05 inch per °C per day. 60% of rainfall was considered as contributory to run-off while 90% of snowmelt was considered as contributing to run-off.

The snowmelt run-off generation for a sub-catchment of Beas basin was made by Agarwal et al (1983), using point energy and mass balance approach. The contribution of various energy sources in different conditions was also worked out. Melt run-off has also been estimated using degree-day method. Melts thus arrived at have been compared with the observed run-off. Study was based on the ground data for the year 1981-92 collected from a snow courses located in the sub catchment. Results indicate that although net radiation balance remains the dominated source of melt energy, yet sensible and latent heat contribute in the range of 40% to 60% total energy for the altitudes below 3000 m in the open areas during clear and partly cloudy days in the active snowmelt period. The effect,however, becomes insignificant during cloudy days and longwave radiations. The positive longwave radiations take over the process of melt. The influence of radiations on cloudy days ranges from 20% to 34%. It has been demonstrated that flow generated through energy balance method generally agrees with the observed flow. However, the melt run-off determined through degree-day method shows variation up to + 226%. The results clearly demonstrate the relative merits of energy and mass balance method over the conventional degree-day method. It is also reported that low status of the generated flow using energy balance method and its variation with respect to the observed one has many contributory factors such as inaccurate delineation of physical parameters due to non availability of large scale maps, inaccurate knowledge of the extent of forest canopy density, lack of knowledge of the behaviour of reflected longwave radiations during partly cloudy days, non use of the principles of liquid water transmission through the snowpack, omission of routing techniques and the inaccuracies in the observed flow.

Jeyram et al (1983) have made study on snowmelt run-off computation for Beas river catchment up to Manali. Using the areal snowcover information obtained from Landsat imageries for the period April to July ,1971-1976. The basin area was divided into twenty altitude zones having each 200 m elevation difference. The daily snowline altitudes were obtained by linear interpolation of depletion curves . Every tenth day snowline elevation values are plotted against time from April-July from 1971-1976. The close examination of these curves indicated that snowcover depletes slowly in the month of April to mid of May and as soon as a small portion of the cover melted ,snowcover depletes very fast in the month of May and June. The comparison of depletion curves over a period of six years showed the similarity in the shape but shifted in time. This time shift is function of the volume of snow stored in the watershed. The study of relative displacement is useful in providing a rough estimate of volume of water produced when it melts. For computing daily melt, maximum daily temperature has been used alongwith a value of degree-day factor of 2.1 mm/°C/day. The lapse rate of temperature has been assumed to be 6.5 °C/km for obtaining temperature at mid altitudes of different elevation zones from temperature at the gauging station. The recession coefficient has been taken as an nonlinear function of discharge. The value of R^2 for performance of model varied from 0.5 to 0.80.

Using nomograms based on energy balance approach a computation for snowmelt run-off was made by Daoo and Shirvaiker (1983). Input data required for using the nomograms were insolation , cloud cover, air temperature, relative humidity ,and wind speed which are routinely measured at most of the weather stations. Estimates obtained from nomograms compare well with observed data from Beas catchment for clear or partly covered sky conditions. The computation was made for the months of April, May and June of 1969. However, for overcast or near overcast conditions ,(> 0.70) ,the nomograms underestimated the melt rates by a factor of about three. This is because of bulk parameterization formula for net long-wave radiation flux always overestimates the actual flux for overcast conditions (Kondratyev,1969), underestimating the net radiation flux and resulting in underestimation of snowmelt rates. Delay period for snowmelt water to reach the gauging point has also been approximately established.

An approach for estimation of maximum water equivalent of snowcover and time distribution of snowmelt from April to June using features like area - elevation relationship, freezing level, surface temperature, was presented by Abbi et al (1983). Monthly snowmelt has been computed using degree-day concept and compared with actual observed discharge at Kullu ,at the confluence of Beas and Parvati river with a view to evolve a relationship for forecasting discharge on the basis of snowmelt. Considering the variation in elevation, the watershed has been divided into two district zones namely where the precipitation occurs as snow during winter and where the precipitation occurs in the form of rainfall.

Upadhyay et al (1985) have shown for Beas and Satluj basin that snowmelt caused by incoming solar radiation is predominant over other physical processes such as longwave energy transfer at the snow air interface,the convective heat exchange and latent heat released by condensation. It has also been shown that the degree-day approach for snowmelt computation does not exhibit significant difference from the results obtained by thermal quality approach. The degree-day approach was recommended for operational use because of the relative ease in computation.

8.3 Application of Snowmelt Simulation Models :

Only limited studies have been made either to develop or testing of existing snowmelt simulation models in the Himalayas. A snowmelt run-off model was developed and verified using

1977,1978 and 1979 years data of Beas river catchments up to Manali by Seth (1983). The model uses the information regarding the areal extent of permanent and temporary snowcover obtained by comparison of satellite imageries, observed data of precipitation for November to May and daily temperature for premonsoon season. The model considers altitudinal effect on temperatures, orographic effect on precipitation, melt water effect of rain falling on snowcovered area. Simple routing relationship has been used for obtaining daily streamflow at catchments outlet. This study also deals with three different values of number of elevation zones viz. four, eight, and sixteen. The eight parameters representing degree-day factor for two parts of the season, losses from the snowmelt, rain on snow covered areas and rain on nonsnow covered area; lapse rate, melt due to rain and routing (recession) factor have been estimated for different alternative number of elevation zones, by pattern search optimization techniques using least squares objective criteria. The cross correlation analysis and sensitivity analysis have also been used for examining reasons for good or bad reproduction of observed flow. It was concluded that inspite of limitation of data and simple approach, for all the three years there is reasonably good reproduction of observed direct run-off.

It was found that the results generally improved with increase the number of elevation zones in comparison to those for case of 4 elevation zones as reported by Seth (1981). It has been suggested that as the area for elevation zones have been obtained by interpolation, the results are expected to improve if better information on area-elevation becomes available. The availability of satellite imageries for different times during melt period will also improve the performance of the model.

Singh (1989) has tested the snowmelt run-off model (SRM) developed by Martinec and Rango for Beas basin up to Manali in the Himalayan region for a limited period. The parameters used in this modelling study have been established for the basin. The result of model computation are verified by comparing observed and computed daily discharge for the years of 1978 and 1979. A good consistency has been found between the simulated and measured discharge. The goodness of fit measure has been computed to be 0.83 and 0.61 for the years of 1978 and 1979 , respectively. The higher value of goodness-of-fit measure for the year of 1978 attribute to the good information available regarding snowcover area of the basin.

Dey et al (1989) employed SRM to simulate run-off during the snowmelt season from the large ,data sparse Kabul river basin in the Himalayas. It has been reported that certain factors can lead to simulation problems on such basins such as climate stations located vertically and horizontally at a great distance from the snowmelt contributing area,unrepresentative lapse rates and other SRM parameter values,too large a basin area with very few elevation zones , and inadequate snowcover data. The analysis of initial SRM simulation for the Kabul basin led to the implementation of several potential modifications which drastically improved the simulation. More representative parameters , including lapse rate, were used, and temperature was extrapolated to a mean elevation representative of the tributary supplying the vast majority of snowmelt run-off. These new values produced a much improved simulation that was comparable to prior results of the region. Little could be done to solve problem of unrepresentative climate data.

9.0 PROPOSED SNOWMELT MODELLING STUDIES FOR SATLUJ CATCHMENT

A project of snowmelt modelling study has been conceived by the Snow and Ice Panel of Indian National Committee on Hydrology (INCOH) in Satluj catchment up to Rampur. In this river significant amounts of snowmelt runoff are known to contribute to the river flows during snowmelt season. Forecasts of the spring seasonal flows are required for regulation of release in the lean season from Bhakra reservoir by Bhakra Beas Management Board (BBMB).

The total catchment area of Satluj is about 56874 sq. km of which 37047 sq. km lies in Tibet and remaining 19826 sq. km in India. There is an adequate network of snow gauges stations of BBMB in the basin. 21 number of snow gauges is maintained by BBMB above Rampur . Snow stakes were also installed near the snow gauges to measure snow depth. Temperature data is available for 5 observatories namely Kaza, Rakchham, Namgia ,Kalpa and Rampur for last 6-7 years. The snow cover area would be delineated from the satellite imageries at different periods in the snowmelt season. The Landsat and IRS imageries will be used for this purpose.

It is proposed to carry out snowmelt study using temperature index method because of limited availability of precipitation and temperature data in the basin. The snowmelt models such as

UBC, SRM,SSARR and HEC-1 are proposed to be tested for the basin for snowmelt simulation over a period of three years data.

To carry out the study, the catchment would be divided into various elevation zones and temperature would be interpolated extrapolated using the derived computed lapse rate for the basin. The snow cover depletion curves would be established with time and degree-days as well. The other hydrological parameters of the basin to be used in snowmelt modelling would be established from available data of streamflow and characteristics of the basin. If required , The catchment will be divided into sub-catchments and study would be carried out for each sub-catchment separately and the flows would be routed to the outlet. The model's capability for operational use for forecasting snowmelt contribution would be tested.

10.0 CONCLUDING REMARKS :

Inspite of the fact that first snow surveys to assess the utility of snowmelt forecast in the Eastern Himalayan region were conducted way back in 1947 under the guidance of U S Expert Dr. J E Church, the studies on daily or seasonal snowmelt run-off forecasting have not been carried out to the desired level ,while these are of prime importance for the management of the water resources of our country. The present status of snowmelt studies shows that only regression relationship between snow cover area and runoff has been established for Satluj catchment. Very limited studies on snowmelt processes and simulation of snowmelt run-off are carried out. Further, these studies are confined to one or two watersheds and are inconsistent .This may be because of poor network in the Himalayan catchments to observe required data for snowmelt forecasts. In the recent years , the condition of the network in several Indian snowbound catchments has been improved.

The monitoring of snowfall in Western Himalayas at present is being done by India Meteorological Department (IMD),Central Water Commission (CWC), Bhakra Beas Management Board (BBMB),and Snow and Avalanche Study Establishment (SASE). Snowfall is measured by IMD using snow gauge and snow depth is measured by snow stakes. The CWC has a dense network of 33 stations in Chenab basin using snow gauges and 9 stations in Yamuna basin with a snowpillow installed at Jubbal. The BBMB has a network of 21 snow gauge stations in Satluj basin above Rampur where snow gauges and snow stakes are used for monitoring of Snow. The SASE has good network of snow gauges stations at very high elevations in the Western Himalayas extending from Kashmir to hills of Uttar Pradesh. At some of the locations weighing type precipitation gauges or snow stakes are used and at few locations snowpillows are being used. This enables to carry out the investigations in the field of snowmelt forecasting.

Mostly precipitation and temperature data is being collected in the Himalayan snowbound catchment. Keeping in view the rugged topography, type of data collection by the existing network, high altitude based large basins of Himalayas which are mostly inaccessible, hazardous, it is evident that only simple model structures with reasonable physical base , limited data requirements and capability of using remotely sensed data for snowcover area seem to be appropriate.

Attempts are required for regression analysis between available hydrological and meteorological variables for various basins. The snowmelt simulation models based on temperature , precipitation and snow cover data obtained through the remote sensing technique and having option for dividing the basin into elevation zones may suit to Himalayan catchments. Subdivision of river catchment into elevation zones is considered desirable because of the strong elevation dependent gradients of temperature and precipitation in the mountainous areas. Extrapolation of temperature and precipitation data to various zones would also compensate the limited data availability at different altitudes in the basin. Co-ordinated efforts by different organizations are needed to develop and test the regression or snowmelt simulations for short-term and long-term forecasting in an operational mode. In order to get reliable results, the refinements in the models may be done from time to time. Such investigations would offer a great potential as a forecasting tool and a means to improve understanding of snowmelt processes.

A number of advances in snowmelt run-off forecasting have been made in the past few decades. These advances resulted from an imprcved understanding of the physical processes of snowmelt and basin run-off and the development of new technologies in the area of data collection and computer technology. This has contributed to the development of simple as well as complex snowmelt models which are used for operational purpose in developed countries. The snowmelt models such as SSARR, SRM, UBC, NWSRFS require data which is generally available in Himalayan watersheds. Concerted efforts are also required to test these models in the conditions for Indian snowbound catchments.

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