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Modelling of Snow and Glacier Melt Runoff

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INTRODUCTION

Snow is an important part of the hydrologic cycle and considered as a dominant source of streamflow in many parts of the World. In the context of India, Himalayas are the reservoir of snow and glaciated ice and regulate the annual water distribution. A substantial amount of melt water is received in the major river systems of India, namely the Indus, the Ganges and the Brahmaputra. Studies carried out in the western Himalayan region confirm a substantial contribution of snow and glacier melt runoff into annual streamflow of these rivers (Singh et al. (1994a, 1994b, 1996c). Accurate estimates of the volume of water stored in the basin in the form of snow in winter and its rate of release due to melting in summer, are needed for many purposes. These include streamflow and flood forecasting, reservoir operation, watershed management, water supply, and the design of hydrologic and hydraulic structures. The planning of new multi-purpose projects in the Himalayan region further emphasizes the need for reliable estimates from rain, snow and glacier runoff. During the monsoon season, there is good rain in the outer and middle Himalayan ranges and therefore, modelling of rainfall-runoff is also required when total streamflow is simulated/ forecasted (Singh et al. (1995), Singh and Kumar (1996b)). A thorough understanding of the relationships between meteorological variables and the snow and glacier melt processes is needed for seasonal and short-term water yield forecasting.

The present status of the snow and glacier melt studies in India show that daily and seasonal snowmelt runoff forecasting have not been carried out and yet they are of prime importance for management of water resources. Only some efforts are made to develop regression relationship between snow cover area and runoff. This may be because only limited meteorological and streamflow data are available for Himalayan catchments, less than required for snowmelt forecasts. Inaccessibility due to rough and rugged terrain, and harsh weather conditions are the basic reasons for lacking in data collection. However, in the recent years the condition of the data network has been improved for several snowbound catchments and systematic and continuous efforts are being made to collect the required data. Forecasting of snow and glacier melt runoff are of immediate need for all the Himalayan rivers. Detailed information on melt water yield from glaciers and its distribution with time is lacking for all the glaciers in the Himalayan region.

Presently various Institutes/Organisations are associated with research and development in the area of snow and glacier hydrology. National Institute of Hydrology (NIH) is carrying out snow melt modelling studies in few Himalayan basins. For understanding of glacier hydrology, glacier runoff from Dokriani glacier is being monitored with other meteorological parameters. Attempts are being made for the modelling of the glacier melt runoff. Precipitation and temperature distribution in the

Himalayan regions has been investigated. The following two projects are completed/under progress in this area of research. An estimation of snow and glacier contribution in the Chenab and Ganga rivers has been made by the Institute.

MODEL TYPES AND THEIR USE FOR VARIOUS HYDROLOGIC APPLICATIONS

In general, snow melt models can be divided into two types of models, namely index models and energy balance models. Broadly, energy balance models require the information on radiant energy, sensible and latent heat, energy transferred through the rainfall over the snow and heat conduction from ground to the snowpack. Several meteorological parameters are to be monitored to obtain this information over the snowpack. A thorough understanding of the basic energy transfer processes and their role in melting of snowpack, helps in improving the performance of the operational snow melt models.

Index models use one or more variables in an empirical expression to estimate snow cover energy exchange. Air temperature is the most commonly used index, but other variables such as net radiation, wind speed, vapour pressure and solar radiation are also used. Zuzel and Cox (1975) studied the relative importance of meteorological variables in snowmelt and found that if only one meteorological variable is available for snowmelt prediction, average temperature is the best predictor. The degree-day method is more popular because temperature represents reasonably the energy flux and at the same time, it is relatively a easy parameter to measure, extrapolate and probably to forecast. However, snowmelt prediction can be significantly improved by using vapour pressure, net radiation and wind rather than the temperature variable alone.

Need for development of a snow and glacier melt runoff model for the Himalayan basins

Most of the Himalayan basins experience runoff from the rain, snow and glacier. The contribution from the lower part of the basin ($\sim < 2000$m) comes from the rain, from the middle part between 2000m to 4000m comes from the combination of rain and snowmelt and from the high altitude region >4000m arrives from the glacier melt. The contribution from snow and glacier is controlled by the climatic conditions and therefore, varies year to year.

It shows that Himalayan basins are complex in nature in terms of input to the basin and contribution from all three sources to be estimated. For these type of basins, situation becomes more complex because contribution from each component is not known separately. The observed flow consists of the contribution from all these three sources in addition to the base flow/ground flow contribution. Keeping in view the Himalayan basins, the first and most important factor influencing the development of model and the approach to be adopted, is the availability of data. There is very sparse network in the high altitude region of the Himalayas. Data collected at those stations consist of mostly temperature and precipitation only. Data required for the application of energy balance approach are hardly available. Therefore, obviously development of a conceptual model with an index approach for calculating the snow and glacier melt runoff would be the right start in this direction for the Himalayan basins. The proposed model will include the simulation of snow accumulation, snowmelt and glacier melt runoff and rainfall-runoff. Algorithms for modelling of snowpack cold content is needed for the seasonal conditioning of the snowpack before generating the melt water. The provisions to handle the evapotranspiration, soil moisture deficit, base flow infiltration and routing of surface and sub-surface runoff should be incorporated in the model.

The successful application of the model is dependent upon derivations of the various parameters and relationships specific to a particular basin or river system. Some of the relationships are general and, therefore, are applicable to many sub-basins/zones within a major drainage basin. Others can be specifically derived for a particular watershed. Some are relationships which can be observed or derived, while others must be considered to be model parameters which only have qualitative physical significance. Watershed runoff characteristics are primarily determined by trial-and-error solutions with the computer program to obtain the best fit of historic streamflow data. This procedure is repeated until adequate verification of observed flow is obtained. The characteristics are then tested with independent data. The efficiency of adjustment procedures is dependent upon the judgement and skill of the user who must evaluate the interactions of the various parameters. The knowledge gained in verification studies can add significantly to the overall understanding of basin hydrologic characteristics.

Division of basin into elevation bands

In the mountainous watersheds where temperature and snow depth vary with elevation, the drainage area is divided into some convenient number of the bands (usually elevation-contour bands) and each elevation band is treated as a separate watershed with its own characteristics and initial snow water equivalent. The areas can be derived planimetrically from topographic map. The number of bands in a basin will depend upon the topographic relief of the basin. However, there is no specified range of altitude for slicing the basin in the bands, but an altitude difference of about 500 m or so is considered appropriate for dividing the basin into elevation bands. Moisture input for each band is the sum of snowmelt and rainfall. Runoff for each band is computed from watershed runoff characteristics developed for that particular band. Streamflow for the whole basin is found by summing the runoff synthesized for all bands.

HANDLING OF METEOROLOGICAL DATA

Distribution of temperature with altitude

Temperature data are available as point values at different elevations in a basin. These point values are extrapolated or interpolated to the mid elevation of each elevation zone using a predefined temperature lapse rate in the model. Lapse rates are known to be quite variable, ranging from high values of about the dry adiabatic lapse rate to low values representing inversion conditions. For example, during continuous rainstorm conditions the lapse rate will approximate the saturated adiabatic rate, whereas under clear sky, dry weather conditions, the lapse rate during the warm part of the day will tend to the dry adiabatic rate. During the night, under clear sky conditions, radiation cooling will cause the temperatures to fall to the dew point temperature, and this is particularly true for a moist air mass. As a result, night time lapse rates under clear skies will tend to be quite low, and at times even zero lapse rates will occur. The trends of variation of temperature show that a complete and detailed representation of the variability of temperature lapse rate is not possible, but in case temperature data are available at least at two stations, one at the bottom of the basin and one near the top, an actual temperature lapse rate could be computed and used for extrapolation/interpolation in the various zones. When one station is used, a lapse rate has to be assumed in order to extrapolate temperature from base station to the appropriate mean hypsometric elevation.

Generally temperature is lapsed, at either 6.5 °C/km or at a specified rate, from mean elevation of the index stations to the median elevation of the melt area of each band. The temperature at which snowmelt begins is the base temperature specified in the model. Temperature for each time period is lapsed from the index station to the rain-freeze and base temperature elevations. In fact in order to minimize the errors in extrapolation / interpolation of meteorological variables, at least one station for each elevation band is required. This will minimize the vertical distance for generating the information on meteorological variables.

Distribution of precipitation with altitude

A common problem with rainfall is the correct evaluation of point rainfall/precipitation in terms of areal distribution. Distribution of precipitation with altitude is much more difficult than the distribution of temperature. Charbonneau et al. (1981) mentioned that determination of the spatial distribution of precipitation is more important than the selection of a modelling approach. The distribution of precipitation is very much influenced by the topography of the basin in which height of the mountain barrier and direction of the air currents are the dominant factors.

The algorithms which describe the variation of precipitation with elevation can be described through the basic enhancement of precipitation with elevation barrier height. Because snow and rain indicate different patterns of variation with altitude, therefore, separate relationships should be provided. Previous studies conducted to understand the trends of precipitation variation with altitude in the study basin or in the nearby basin are very helpful to generate the precipitation at the different elevations bands of the basin, otherwise user's knowledge of the study area is to be used. Studies on precipitation distribution for few Himalayan basins has been carried out by Singh et al.(1995), and Singh and Kumar (1996b).

ESTIMATION OF VARIABLES ASSOCIATED WITH HYDROLOGICAL PROCESSES

Computation of degree-days

Air temperature expressed as degree-days is used in snow melt computations as an index of complex energy balance tending to snow melt. A degree-day, in its broad sense, is a unit expressing the amount of heat in terms of the persistence of temperature for 24 hour period of 1°C departure from a reference temperature. The simplest and most common expression relating snow melt to temperature index is,

$(\sigma T^4 - \sigma T_2^4)$

$$M = D_f (T_i - T_b)$$

$= m s \Delta t$

where,

M = depth of melt water (mm) produced in a unit time
D_f = degree-day factor (mm °C^{-1} day^{-1})
T_i = index air temperature (°C)
T_b = base temperature (usually, 0°C)

Although temperature and other hydrologic conditions vary continuously throughout the day, daily mean temperature is the most commonly used index of temperature for snow melt. Where only maximum (T_{max}) and minimum temperatures (T_{min}) are available, the number of degree-days is computed as

$$T_i = T_{mean} = (T_{max} + T_{min})/2$$

At stations where hourly readings are made, the number of degree-days for each 24 hour period is determined by summing the hourly temperatures and dividing by 24. Further 0°C base temperature is generally used in computation of degree-days. This follows from the idea that most snowmelt results directly from the transfer of heat from the air in excess of 0°C. Some investigators have also used only T_{max} for computing the degree-days. In that case a lower value of D_f is used along with a lower value of base temperature.

Degree-day factor

The degree-day factor (D_f) is an important parameter and converts the degree-days to snow melt expressed in depth of water. D_f is influenced by the physical properties of snowpack and because these properties change with time, therefore, this factor also changes with time. The seasonal variation in melt factor is well illustrated by the results obtained from the study reported by Anderson (1973). The lower value being in the beginning of melt season and higher towards the end melt season. A wide range of a values has been reported in the literature with a generally increase as the snowpack ripens. For example, Garstaka (1964) reported extreme values of D_f as low as 0.7 mm °C^{-1} day^{-1} and as high as 9.2 mm °C^{-1} day^{-1}. Yoshida (1962) reported the values of D_f to be between 4.0-8.0 mm °C^{-1} day^{-1}, depending on the location, time of year and meteorological conditions. Singh and Kumar (1996a) determined the D_f factor by monitoring a known snow surface area of the snow block within the snowpack at an altitude of about 4000m in the western Himalayan region in the summer. The mean daily value of the D_f was computed to be 5.94 mm $°C^{-1}$ day^{-1}, while for a dusted block it increased to 6.62 mm $°C^{-1}$ day^{-1}. In glacierized basins, the degree-day factor usually exceeds 6 mm $°C^{-1}$ day^{-1} towards the end of summer when ice becomes exposed (Kotlyakov and Krenke, 1992). As discussed above that D_f changes with season, therefore, when using degree-day approach changes in D_f with season should be taken into account. For rain periods, D_f can be adjusted as follows:

$$D_{f,rain} = D_f + 0.0126 * P_r$$

where P_r is rainfall in mm. It is to be noted that this adjustment is required only when melt caused by rain is not computed separately. If this effect is taken into account with additional calculations, then unadjusted D_f is to be used throughout the snow melt calculations.

Form of precipitation

The distinction between rain and snow for each elevation band is very important for all the snow melt models because precipitation falling in the form of rain and snow behaves differently in terms of contribution to the streamflow. The contribution of rain to the streamflow is faster than that of snow because snow is stored in the basin until melted whereas rain is immediately processed. Precipitation in each period is assumed to fall as snow above the freezing elevation and rain below that elevation. On any one of the bands, precipitation might fall as rain, as snow, or a combination of the two. Rain on a band or part of a band is added directly to moisture input. Snow is added to the previous accumulation, if any. Accumulated snow is assumed to cover the entire band. Each band must be either snowfree or 100% snow covered. Further, the response of snow fallen over the accumulated snow or snow-free area, are different. The new snow that falls over the previously snow-covered area becomes the part of the previous seasonal snowpack and its effect on runoff depends on the condition of the snowpack. For example, rain falling over the cold snowpack, like in the early melt season, will be frozen in the snowpack and it is not immediately available for runoff. It melts when favourable atmospheric and snowpack conditions are available. But rain falling over the ripe snowpack is transferred through the snow layer and becomes available to contribute to runoff. The new snow falling over the snow-free area is considered as precipitation to be added to snowmelt, though with this effect delayed until the next warm day.

For this purpose a critical temperature is specified in the model to determine whether measured precipitation is rain or snow. This is of particular importance for models which simulate the build-up of snow cover from precipitation data. From direct observations, T_{crit} is generally higher than 0 °C as found by Charbonneau et al. (1981). Therefore, value of T_{crit} is usually selected slightly above the freezing point and may vary from basin to basin. A temperature 2 °C is proposed to be considered in this model. It indicates that

if $T_m \geq 2°C$, all precipitation is rain
if $T_m \leq 0°C$, all precipitation is snow
if $T_m \leq 2°C$ and $\geq 0°C$, precipitation will be considered as a mixture of rain and snow and their proportion will be worked out in the following way:
$Rain=(T_m / T_{crit})*P$
$Snow=P - Rain$

where P is the total observed precipitation.

Snow coverage and glacier covered area

Because snow accumulation and depletion is handled by the model, zonewise distribution of snow cover is computed by the model itself. In other words, extension and depletion of snow cover area is given by the model. However, if possible, it is recommended to update the computed snow covered area by the observed snow covered area for the dates at which observed snow covered area is available. For this purpose, data available through remote sensing technique can be used. The satellite (imagery or digital data) data on good resolution may be preferred and standard methods can be adopted for the analysis. Snowfall occurs during the snowmelt period might modify the depletion of snowpack by increasing the snow covered area for some time. Further, to obtain an insight in to the pattern of depletion of snowpack, it is suggested that the depletion curves that normally relate the areal extent of the snow cover to elapsed time, may also be correlated with accumulated degree-days. The distribution of glacier area in the basin is needed for estimating contribution from the glacier.

The melt contributing area can be estimated using altitude of base temperature and status of basin covered with snow. The base temperature altitude can be determined by extrapolating/ interpolating the temperature data. However, there may be two ways to obtain the area of snow covered basin. First, direct information on snow covered area from the satellite data and second compution of the snow covered area from the model itself.

Rain Melt

Rain-on-snow is a common feature in various alpine parts of the world and plays a significant role in generating high streamflows in some cases. Most of the largest floods in British Columbia, Washington, Oregon and California have been associated with rain-on-snow (Kattelmann 1987; Brunengo, 1990). Archer et al. (1994) reported that in Britain more frequently flooding results from combination of melting snow and rainfall. A good rainfall occurs in the high altitude regions of Himalayas during the active melting period (Singh et al., 1995; Singh and Kumar, 1996b). Further, rain-on-snow events are also considered a major cause in the release of avalanches. Introduction of liquid water into snow weakens the bond between grains and alters the snow texture which results in reduced mechanical strength of the snowpack. Various studies have been carried out to understand the role of rain in triggering avalanches in maritime climates (Conway et. al, 1988, Heywood, 1988; Conway and Raymond, 1993).

When rain falls on the snowpack it is cooled to the temperature of snow. The heat transferred to the snow by rain water is the difference between its energy content before falling on the snow and its energy content on reaching thermal equilibrium within the snowpack. For snowpacks isothermal at 0°C, the release of heat results in snowmelt, while for the colder snowpack this heat tends to raise the snowpack temperature to 0°C. In case the snowpack is isothermal at 0°C, the melt occurring due to rain is computed by

$$Q_p = \rho_w C_p(T_r - T_s)P_r / 1000$$

where,

Q_p = energy supplied to the pack by rain (kJ/(m^2.day))
ρ_w = density of water (1000 kg/m^3)
C_p = specific heat of water (4.20 kJ/(kg.°C))
T_r = temperature of rain (°C)
T_s = temperature of snow pack (0°C)
P_r = depth of rain (mm/day)

Substituting the values of various parameters in the above equation, it reduces to

$$Q_p = 4.2\ T_r\ P_r$$

Usually, rain temperature is considered equal to the air temperature on that day.
The melt caused by this energy is computed as

$$m_r = Q_p/(\rho h_f B)$$

Here,

m_r = melt caused by the energy supplied by rain (mm/day)
h_f = latent heat of fusion of water (~335 kJ/kg)
B = Thermal quality of snow (0.95-0.97)

In the other case, when snowpack is at sub-zero temperature, the freezing of the rain water exerts a considerable impact on the thermal regime of the snowpack. The latent heat of fusion of water plays an important role in increasing the temperature of the snowpack under such conditions. Increase in temperature (°C) is computed as

$$\Delta T = \rho_w P_r h_f / (1000.\rho_s d_s C_s)$$

Where,

ρ_s = density of snow (kg/m^3)
C_s = specific heat of snow (2.09 kJ/(kg.°C))
d_s = depth of snow (m)

Routing of surface and sub-surface runoff

On large mountainous basins, the routing time for water travelling from the source (snowpack) to the basin outlet is often more than one day. Therefore, of the water produced by snowmelt on day n, one portion can be expected to appear as runoff on day n, while the remaining water will appear as runoff on subsequent days. As discussed above, moisture input is divided into surface flow (fast runoff) and sub-surface flow (slow runoff) and water allocated to each components of runoff is routed to produce the time distribution of runoff. The routing will be done using the concept of linear storage reservoir. According to this concept, the resulting outflow from the nth reservoir at time t from a unit instantaneous inflow can be written as

$$u(t) = \frac{1}{k^n} \cdot \frac{t^{n-1}}{\Gamma n} \cdot e^{-t/k}$$

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where,
k is the linear storage constant for each of reservoirs in the cascade,
n is the number of linear reservoirs in the cascade
t is the time after occurrence of input

For the routing of surface runoff (fast runoff), a cascade of linear reservoirs ($n > 1$) will be considered. The sub-surface flow accumulated in a linear storage reservoir ($n=1$) and is released slowly as long-term recession flow. To produce the total runoff at a particular time, the routed flow from each components will be added. Consideration of flow into two components (surface and sub-surface) will result is less number of parameters to be optimized. The minimization of parameters is an important aspect of the model which should be addressed and is important if the model is not to become too cumbersome. It has been shown by the environmental isotope tracer studies (Martinec, 1985) that overland flow is not a major part of the snowmelt runoff as previously believed. There is increasing evidence (Kobayashi, 1985) that a major part of the melt water infiltrates and quickly simulates a corresponding outflow from the groundwater reservoir.

The calibration may be accomplished by any of three methods: analytical, numerical, or subjective optimization. With subjective optimization, initial values of model parameters are selected using tester's knowledge of the system and then trial- and-error is used to adjust the initial values in order to fit the goodness-of-fit statistics. Analytical method is more precise and reproducible method of calibration. The criterion function is defined to be the sum of squares of the prediction error. To obtain the accurate estimate of the parameters, the model should be calibrated for by operating continuously for a number of years of data at a time.

CONCLUSIONS

1. Inspite of substantial contribution from the Himalayan rivers, at present no hydrological models is available in our country which can be applied to for forecast runoff for the Himalayan rivers which consists of the contribution from rain, snow and glaciers. A conceptual model/ models based on simple approach and minimum data requirement should be developed for the Himalayan basins.

2. Estimation of melt water yield from the representative glaciers and its distribution during the melt period should be made. The modelling of glacier melt runoff should be attempted separately and then integrated with snow melt runoff model.

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