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STATUS AND PROBLEMS RELATED WITH MOUNTAIN HYDROLOGY

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In spite of the significant role of mountains in providing water resources, the information and knowledge on hydrology of mountain regions are incomplete and fragmentary. Extremely difficult terrain, the harsh environment, sparse settlements, and the need for a special design of instruments are major problems in the understanding of mountain hydrology. The understanding of the basic hydrological variables like, precipitation, temperature and evaporation is very poor for the mountainous basins. Moreover, their spatial and temporal distribution is not properly understood, particularly for the Himalayas. Perhaps non-availability of hydrometeorological data at higher altitude for long duration is the main reason for lack of understanding of distribution of these variables. Strengthening of the network in the mountainous regions is badly needed to improve the hydrological database required for the hydrological modelling studies. For a better understanding of different hydrological variables, a long-term operation of a few multidisciplinary experimental research basins is required in different regions of mountain areas.

This report deals with identifying the hydrological problems related to mountainous areas, both low altitude and high altitude regions. Hydrological characteristics of mountainous catchments have been discussed in detail. Application of remote sensing in the mountain areas, particularly for snow and glacier studies has been reviewed and status of such studies is presented. Norms of hydrological network in the mountains is reported. Effect of climate change on the hydrological behaviour of the mountainous river is reviewed. A literature review on the assessment of water resources, sediment and erosion, hydrochemistry is carried out.

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Abstract

This report discusses about the status and progress of studies of various hydrological problems related to mountainous areas experiencing rain or snow or both. The principal issues related to mountain hydrology on the local and regional scales are discussed. And highlighted. The status of the hydrological studies carried out for the mountainous basins shows that understanding of basic hydrological variables like, precipitation, temperature and evaporation is poor for the mountainous basins. Spatial and temporal distribution of these hydrological variables is not well understood. Perhaps non-availability of hydrometeorological data at higher altitute for long duarion is the main reason for lacking in understanding of distribution of variables. There is urgent need to improve network design in the mountainous area in particular for high altitude region. For a better understanding of different hydrological variables, a long-term operation of a few multidisciplinary experimental research basins is required in different regions of mountain areas. Distribution of precipitation and temperature with altitude in a high altitute basin becomes important to compute melt runoff from snow and ice.

Application of remote sensing techniques has a tremendous potential to determine extent of snow and glacier covered areas on the basin scale. Year to year variation in snow covered area should be studied. Snow and glacier contribution in annual flows of Himalayan rivers is to be estimated. There is need to develop a hydrological model which can take into account the melting of snow and ice from high altitude part and contribution of rain from the lower part of the Himalayan basins. Impact of climate change on different hydrological parameters is to be studied for the snow and glacier fed rivers. High altitude lakes in the Himalayan region are also lacking for hydrological study. An accurate assessment of sediment transported by the Mountainous river is required for the planning of water resources in such regions.

1.0 INTRODUCTION

A large portion of the earth's surface is covered by mountains. As the mean altitude of the land area of the Globe is 875 m and about 28% of the land area rises above 1,000 m (Kundzewicz and Kraemar, 1996). Most of the Earth's mountainous regions are very rich in water resources and high mountain areas are providing reliable water supply for a greater proportion of society. Sources of all major rivers in the world are located in mountains. Mountains remain the dominating control factor in the flow of most major rivers for thousands of kilometres. In the Asian continent alone, seven of the world's largest rivers originate from the Tibetan Plateau. Mountain regions are areas of great physical and cultural diversity where rapid environment changes are often coupled with severe hydrological changes along with socio-economics consequences. These high altitude land features are very sensitive to any kind of minor to very-minor man-made or natural changes. The mountain environment is highly energised, dynamic, and extremely vulnerable to irreversible changes.

The mountains are the abodes of snow and ice, and therefore, they are also very big sources of fresh water for sustaining the life in all forms in the mountains, as well as beyond. Snow and glaciers are the reservoirs with vast storage of fresh water. About 80% of fresh water on our globe is locked up in the form of snow and ice. Although only 3% of this permanent snow and ice is distributed over mountains in various continents outside Polar region (Flint, 1971). This small amount is source for sustenance of major part of population of the world. Out of the mountain glaciers, central Asian mountains contain about 50% of the glacier, a large portion of which drain into the land mass of Indian sub-continent. The present estimate of the glaciated area is 14.9 million km^2 which is about 10% of the land area of the globe (IAHS, 1993). The hydrology of the mountain areas is, therefore, an important study area, complexities and diversities of mountain topography and weather conditions make it difficult as compared to plain areas.

In the context of India, about 35% of the geographical area in India is mountainous and 58% of this is accounted for by the mighty Himalayas. Indian mountains can be described in following series of hills and mountains viz. Himalayan series, Satpura range, Vindhya range, western Ghats and eastern Ghats (Figure 1). In India, no mountain ranges, except Himalayas, experience snow. Therefore, glaciers also exist only in the Himalayas. The Himalayan mountain system is the tallest water tower on our planet and contains enormous renewable water reservoirs of perennial snow and ice at the highest elevations. The perpetual snow zone of these mountains has given rise to a number of rivers. The snowmelt and glacier melt discharges were maximal from June to August in the upper reaches of the river (above 2,400 m). Further, the role of the Himalayas in controlling the climatology is well pronounced in the Indian subcontinent. Ongoing demand along with the scarcity of power, has necessitated exploration of potential sites for hydroelectric schemes. According to the Central Electricity Authority of India (CEA 1985), of an estimated hydropower potential of 64,000 MW, a major portion is available in the north-west and middle Himalayan and north-eastern regions of the country.

A high energy turnover in generation of atmospheric precipitation and in evaporative processes at the land surface in mountain areas have consequences on not only for hydrological investigation and water resources assessments but also for the accuracy in simulation and modelling of the general circulation of climate globally. Mountain areas are very sensitive to changes in the environment and are good indicators of the changing scenario. An accurate

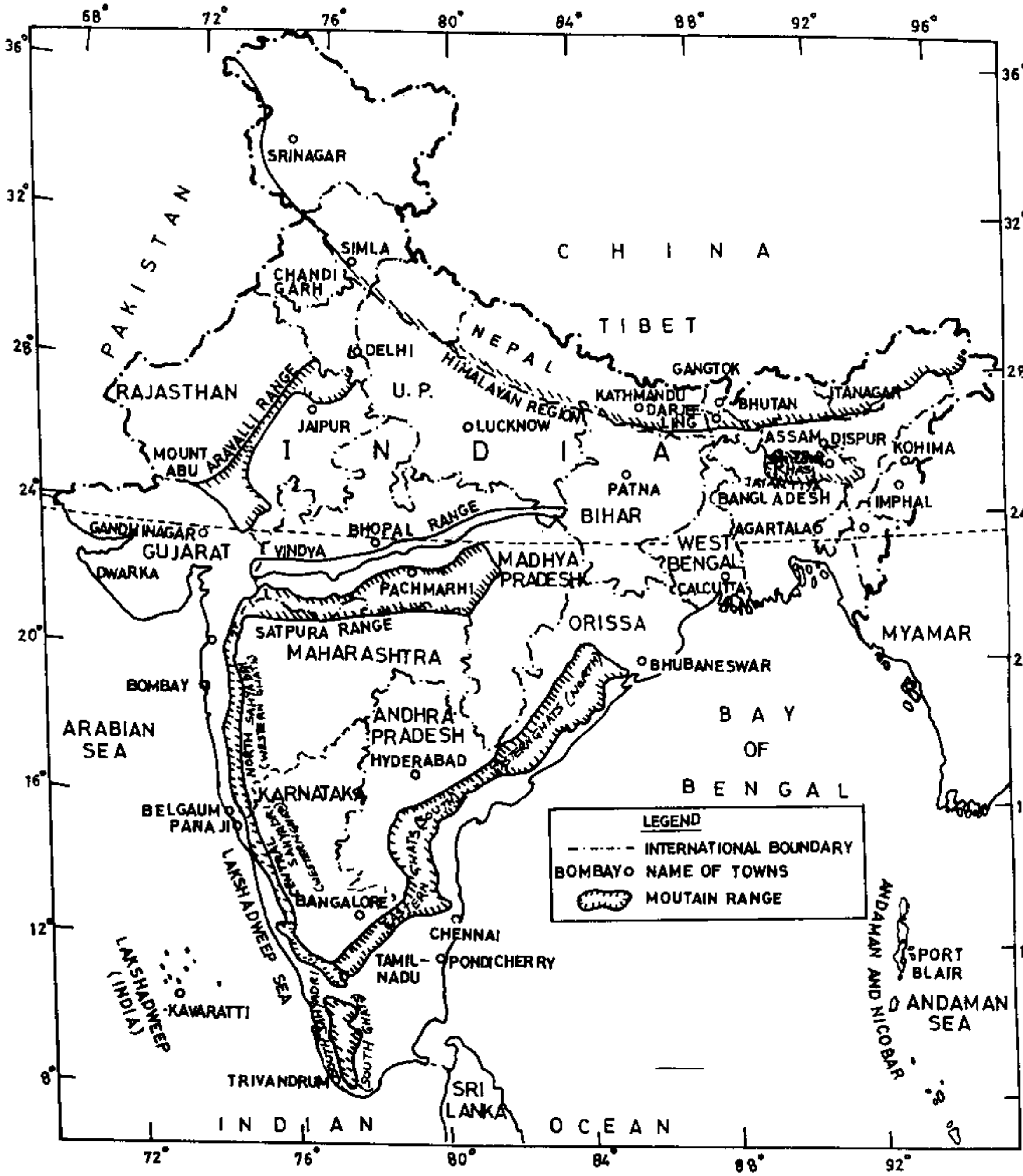

FIG.1: MOUNTAIN RANGES IN INDIA

prediction of global changes in response to increasing greenhouse gas concentration, or majorchanges in land use, and the elucidation of soil degradation on the regional and local scales would depend on a better understanding of the hydrological cycle and associated bioecochemical cycles in coupled hydrological/ atmospheric models. Greater efforts are required to improve prediction capabilities and the interpretation of their regional consequences, especially in the mountain region where the predicted change is maximum.

The processes that govern the hydrology along with ecology and interrelationship between these two is yet to be properly investigated and understood. It is a well known fact that rapid population growth and an indiscriminate race for accelerated economic development have caused enormous pressure on mountain environments which created various kinds of problems in and surrounding watershed areas. Until recently, the hydrology-related problems were considered to be emerging due to human activities in the mountains. Currently, the possible impact of climate change, which are of uncertain nature, have further added to the difficulties in improving our knowledge and understanding of the hydrology of the mountain. These above referenced man-made changes exerted considerable pressure on the land resource and vegetation of mountain areas in the world. Such changes in land use and land cover have brought about modifications in water-flows, nutrients, sediments and pollutants as well as loss of biodiversity almost everywhere in the mountains.

In spite of the significant role of mountains in providing water resources, the information and knowledge on hydrology of mountain regions are incomplete and fragmentary. Extremely difficult terrain, the harsh environment, sparse settlements, and the need for a special design of instruments are major problems in the understanding of mountain hydrology properly and closely. Although the developments of hydropower and irrigation schemes in some high mountain areas has promoted hydrological networks and investigations, but natural hydrological processes, impacts of human interferences and climate change on the availability of water, highland-lowland linkages and sustainable use of water are not understood properly. Therefore, there is an urgent need for a better understanding of the problems of mountain areas because of the vulnerability of the land-water system to human activities and climate change impacts in mountains, which are very fragile hydrologically and susceptible to irreversible changes. There is a consensus on the lack of knowledge of the hydrology of mountainous areas. As noted by Rodda (1994) in his paper 'Mountains- A Hydrological Paradox or Paradise', although mountains are the source of the bulk of the water resources of the world, knowledge of mountain water resources is 'generally much less extensive, reliable and precise than for other physiographic regions'. Klemes (1990) has made a similar statement: 'Hydrology of mountainous areas is lagging behind many other areas of hydrological inquiry in proportion to its greater difficulty'. Yet mountain hydrology is a challenge to researchers wishing to improve understanding of this hydrological paradox on which a plethora of challenging and rewarding hydrological studies can be undertaken. This backs Rodda's (1994) concept of a hydrological paradise. The main objective of this report is to discuss about the problems of mountain hydrology in general and with reference to India in particular.

2.0 SPECIAL HYDROLOGICAL CHARACTERISTICS OF MOUNTAINOUS CATCHMENTS

The particular structure and form of high mountains provide the following important attributes.

i) Delayed runoff: There is a temporary storage of water in the form of snow and ice which provides delayed runoff. Snow cover building up in the higher mountains in winter produces snowmelt in spring and contributes a large volume of water to the discharges of rivers.

ii) High hydropower potential: Huge amounts of potential energy are available that can be used to generate hydroelectricity. There are many countries in which much of the national energy demand is covered by hydropower from mountainous areas.

iii) Large volumes of natural and man-made storage: Man-made reservoirs in mountainous areas may be constructed for different purposes, such as flood protection (holding surplus water before it reaches the lowlands where it might otherwise cause catastrophic flooding), water supply for irrigation and hydro power, maintaining water quality (e.g., augmentation of low flows downstream), and recreation.

iv) Sedimentation and erosion problems of the mountainous areas: Because of steep slopes , deforestation and improper use of land, as a result of the increasing pressure of population and unscientific developmental activities, the flow behaviour of water and sediment had been changing and threatening the ecological equilibrium.

3.0 HYDROLOGICAL NETWORK IN THE MOUNTAINS

WMO's Guide to Hydrological Practices (1994) recommends that observation networks in mountainous regions should be denser than in most other areas. This is because of the heterogeneity of the mountain environment which causes a large spatial and altitudinal variation in the hydrological variables. However, as shown by the result of WMO's Basic Hydrological Network Assessment Project (BNAP), in practice, hydrological networks in mountainous regions are less dense than elsewhere. Table 1 gives a comparison of the proportion of area rising above particular altitudes versus the proportions of the global number of non-recording and recording rain gauges located above these altitudes. It can be noted that at all altitudes over 500 m, precipitation networks worldwide are less dense than from the simple prorating of the area. The only exception is the proportion of non-recording gauges over the elevation of 2500 m. The adequacy of hydrological networks in the mountainous areas in different regions along with percent of mountainous area is shown in Table 2. It is noted that in all the regions, density of hydrological networks in mountainous areas is lower than the average for all physiographic regions.

Perks and Winkler (1992) has shown that in a number of areas, deficiencies exist in the observational networks of essential hydrological variables. A comparison of recommended and actual densities of hydrological networks in the mountainous regions is given in Table 3. It can be noted that actual densities of two most important hydrological variables, namely, precipitation and discharge are significantly lower than the numbers of recommended by WMO (1994). In fact, the networks of these two indispensable hydrological variables should be adequate because they are the basic hydrological variables for the assessment of water resources and for planning their development and management. In addition, the operation of existing networks has been affected by budgetary constraints which led to the abandonment of some of the hydrological stations. Despite the recommendation of the World Meteorological Organisation (1994), it is seen the hydrological networks in mountainous regions are less dense than elsewhere.

Table 1: The global number of non-recording and recording gauges at different altitudes (WMO, 1994)

Elevation range (m)	Percent area	Percent non-recording gauges	Percent recording gauges
< 500	52.7	76.85	76.56
500 – 1000	19.5	15.42	15.93
1000 – 2000	15.4	5.93	6.75
> 2000	12.3	1.75	0.6

Table 2: Proportions of mountainous areas (in %) and the proportion of the total number of stage observations (in %) located in mountainous areas for particular WMO regions (WMO, 1994).

Region	Percent mountainous areas	Percent mountainous gauges
1. Africa	16	13
2. Asia	39	34
3. South America	23	8
4. North and Central America	21	10
5. South- West Pacific	26	0.3
6. Europe	22	8
World	22	10

Table 3: Comparison of recommended and actual densities of hydrological networks in the mountain environment (Perks and Winkler, 1992)

	Non recording precipitation	Recording precipitation	Evaporation	Discharge	Sediment	Water quality
Recommended mean area per station (km^2)	250	2,050	50,000	1,000	6,700	20,000
Actual mean area per station (km^2)	3,879	4,548	11,406	3,261	5,567	5,839
Adequacy of the observational net work	Grossly inadequate	Inadequate	Adequate	Inadequate	Adequate	Adequate but.....

In general the network is inadequate or the available stations are unevenly distributed. There is no information available on the altitudinal distribution of hydrometeorological stations in the Himalayan region or in the other mountainous region of India. As such, the network is very poor because of various reasons. There are a number of central and state organisations which maintain the network in the mountainous areas like Central Water Commission (CWC), Bhakra Beas Management Board (BBMB), India Meteorological Department (IMD), Snow and Avalanche Study Establishment (SASE), Central for Water Resources Development and Management. (CWRDM) and other academic and research organisations. There is a need to compile the information on whole network in the mountain areas of India. There are a few mountainous basins like Chenab, Satluj and Beas basins, which have relatively better network than other basins. In these basins also most of the stations are located in the valley and do not represent the true precipitation falling on the slopes and high altitude area of the basin. The network of snow gauges at the high altitude region in all the Himalayan basins is very poor.

4.0 UNDERSTANDING OF HYDROMETEOROLOGICAL VARIABLES

4.1 Precipitation

In large mountainous basins weather systems interact with topography and result in highly non-uniform precipitation. A unique feature of the mountain environment is the substantial cloud moisture input, augmenting the precipitation in the form of rain and snow and thus mountains have a strong impact on precipitation distribution. The changes in precipitation with elevation is known as the orographic effect on precipitation. The orographic effect on precipitation can be caused by a number of mechanisms, and their relative importance depends on topography and storm type. Uplift of moisture laden air currents striking against a mountain barrier, provides good rainfall on the windward side. Gradients in amount and intensity of precipitation depend upon several factors such as topography, strength of moisture bearing winds, its moisture content and orientation of the mountain range with respect to the prevailing wind direction. Information on precipitation distribution helps to provide realistic assessment of water resources, estimation of probable maximum precipitation and hydrological modelling for mountainous areas including planning and managing water resources, simulation of runoff and to prepare precipitation maps of the basin/region. Precipitation distribution for the Himalayas is poorly known as compared with many other mountains of the world. A summary of important precipitation distribution studies carried out in different parts of the world is given in Table 4. In general, there is no single universal relationship between precipitation and elevation, as a number of additional factors play a significant role. Very important is the location of the site considered with regard to the prevailing wind directions (leeward versus windward sides) and the sources of moisture, but topography, season, and type of storm have also to be taken into consideration. Mountains usually have increased precipitation compared to the adjacent lowlands. Depending upon the relief of a mountain, there may not be continuous increase in precipitation with altitude: above a particular altitude, it may begin to decrease. It is known that precipitation increases up to some elevation and beyond that decreases. A great spatial and temporal variability of hydrological variables is noted because of the large heterogeneity of mountainous environments. Significant changes in the hydrological variables may occur in slope angle, exposure, geology, and vegetation over a very local scale. Therefore, one has to carefully plan observations of processes and characteristics in a number of locations in order to assess the spatial averages accurately. Due to very poor instrumentation networks in the high altitude regions, the information on the spatial and temporal distributions of hydrological and meteorological variables representing processes such as evaporation, solar radiation, soil moisture, overland flow, and erosion is not available.

Himalayan studies on orography effect

A review of studies on precipitation distribution over mountainous areas of India, shows only a few studies were carried out while such studies are very important for this region. In the context to India, detailed studies to assess the orographic effect on precipitation in the Himalayan region were not done. The main reason for the limited number of studies has been the lack of information on precipitation at the higher elevations because of inaccessibility and other difficulties. Recently, Singh et al. (1995), Singh and Kumar (1997) have reviewed the status of orographic studies and carried out a detailed study on the precipitation distribution with altitude in the western Himalayan region. They studied seasonal and annual precipitation distribution in the Satluj and Beas basins for each Himalayan range separately. For this study, a year has been

Table 4: A summary of some important world wide precipitation distribution studies (Singh and Kumar, 1997)

Authors/reference	Study Area/basin	Type of Precipitation.	Relationship with altitude/gradients	Other specific conclusion
Rumley (1965)	Andes Mountains in Ecuador	Rain	-	Two zones of maximum rainfall observed. Along western slope at 1000m and along eastern slope at 1400m
Golding (1968)	Rocky Mountains	Snow	87mm/100m	Linearly increases with altitude
Engman and Hershfield (1969)	North-eastern Vermount, USA	Rain, snow	-	Average number of precipitation days and hours increases with altitude
Hamon (1971)	South-western Idaho	Snow	-	winter precipitation at 2100m was 4 times higher than at 1200m
Duckstein et at. (1972)	Santa Catalina Mou-ntains near Tucson	Rain	quadratic polynomial	-
Storr and Ferguson (1972)	Rocky Mountains	Precipitation	64mm/100m for ann-ual precipitation and 10mm/100m for sum-mer rain	-
Caine (1975)	San Jaun Mountains near Colorado	Snow	65mm/100m	Linearly increases with altitude
Sudgen (1977)	Antarctica and North Greenland	Snow	-	Increases to about 1500-1600m and thereafter decreases
	South-east Greenland	Snow	-	Maximum snow occurs at about 700m
Hendrick et a. (1978)	Mansfield, Vermount, USA	Precip.	-	A three fold increase in the hours of precipitation between 400 and 1200m
Dhar and Rakhecha (1981)	Central Himalayas (Nepal Himalayas)	Rain	polynomial of 4rth degree	Maximum rainfall occurs at 2000-2400m
Higuchi et al. (1982)	Nepal Himalayas	Rain	-	Rainfall decreased with altitude in the range from 2800-4500m.
Witmer et al. (1986)	Alps	Snow	80mm/100m on S-E slopes, and 730mm/ 100m on northern slopes	-
Niemczynowicz (1989)	Jamtland area, Swedish Mountains	Rain	9.5% per 100m	The highest rainfall observed behind the crest of mountains on leeward side
Ohmura (1991)	Western Greenland	Snow	-	Maximum precipitation zone around 2500m at 69EN and decending northward to about 1500m at 76EN

Table 4 (continued)

Authors/reference	Study area/basin	Type of Precip.	Relationship with altitude/ gradients	Other specific conclusions
Barry (1992)	Alps	Precip.	-	Precipitation increased with altitude to the highest level of 3000-3500m
Singh et al. (1994,95)	Chenab basin, Western Himalayas	Rain, snow	Second order polynomial for annual rainfall on windward of outer Himalayas, windward and leeward of middle Himalayas. Linear increase on leeward of outer Himalayas and exponential decrease in the greater Himalayas. Linear increase in snow with altitude.	Spill over effect was noticed in the outer Himalayas. Maximum rainfall was in the outer Himalayas
Loukas and Quick (1993, 96)	British Columbia	Rain, snow		Increases up to mid elevation of basin, then decreases and/or levels off or may increase. Hourly rainfall intensity decreases with altitude
Singh and Kumar (1997)	Beas and Satluj basin, western Himalayas	Rain, snow	Rainfall increases linearly in outer Himalayas. In the middle Himalayas second order polynomial trends for winter and pre-monsoon, and linear increase for post-monsoon and monsoon season. Rainfall exponentially decreases in greater Himalayas. Snowfall linearly increase in greater Himalayas.	

divided into four quarters; October-December, January-March, April-June and July-September which form postmonsoon, winter, premonsoon and monsoon seasons, respectively. Depending upon the availability of precipitation data, studies are extended for outer, middle and greater Himalayan ranges. The following conclusions are drawn from this study:

(a) Rainfall distribution

1. On the leeward side of outer Himalayas, for all seasons, rainfall increases linearly with elevation for both Satluj and Beas basins. In the satluj basin, both more rainy days and higher rainfall intensity are responsible for increasing rainfall with altitude in the outer Himalayan range. In the Beas basin, rainfall intensity plays a major role in increasing rainfall with elevation. Rainfall on the windward side is higher than that on the leeward side in the outer Himalayan range of the Satluj basin.

2. Rainfall varies linearly with altitude for the postmonsoon and monsoon seasons in the middle Himalayan range. Annual rainfall follows a similar distribution. But, for winter and premonsoon seasons, rainfall first increases and then decreases after a certain elevation. The rainfall distribution was fitted reasonably well by a second-order polynomial for these two seasons. A sudden rise in altitude of the middle Himalayan range in the Beas basin, behaved as a giant mountain barrier and increased rainfall very significantly on the windward side of this range. Due to this orographic effect, average annual rainfall on the leeward side is less than half on the windward side. Gradients in annual rainfall on the windward and leeward sides are about 106mm/100m and 13mm/100m, respectively. More rainy days contribute to increasing rainfall on the windward side, whereas intensity is higher on the leeward side. Dharamshala experiences exceptionally heavy rainfall (average 1972mm) during the monsoon season. Standard errors of the mean rainfall are always higher on the windward side.

3. There is little rain in the greater Himalayan range of Satluj basin. Most of the moisture is precipitated over the outer and middle Himalayan ranges. Rainfall decreases approximately exponentially with elevation in the postmonsoon, premonsoon seasons. However, rainfall in the monsoon season has no trend. The annual rainfall distribution also follows an exponentially decreasing trend with altitude. Winter rainfall decreases linearly with elevation in this range. Negligible rainfall is observed above 3000m elevation in winter. The reduction in rainfall at higher elevation is due to fewer rainy days at those elevations. Standard errors of the mean rainfall decrease with altitude in the greater Himalayas.

4. The orographic effect on rainfall has led to maximum rainfall in middle Himalayan range in the Beas basin and in the outer Himalayan range in the Satluj basin. Average annual rainfall decreases considerably from the outer Himalayas to greater Himalayas in the Satluj basin. Average annual rainfall in the greater Himalayas is about one-sixth of the outer Himalayas rainfall.

5. Monsoon rainfall contributed most to the annual rainfall, 45%-71% in the Satluj basin and 39%-71% in the Beas basin, over all Himalayan ranges. Minimum rainfall occurs in the postmonsoon season in the outer and middle Himalayas because of the lower moisture content in this season. But, in the greater Himalayan range, minimum rainfall is in winter because most of precipitation falls as snow over this range. Contribution of premonsoon rainfall in annual rainfall increases from outer Himalayas to greater Himalayas and becomes significant in annual

rainfall in the greater Himalayan range. Winter rainfall is also significant in the middle Himalayan range for both basins.
6. There is a higher spatial correlation of rainfall in the outer Himalayan range which may be because of lesser relief in the outer Himalayan range than in other ranges.

(b) Snowfall distribution

7. Snow increases linearly with elevation in the Spiti and Baspa sub-basins. In the upper Satluj sub-basin, it first increases and then decreases. Trend of snowfall variation in the upper Satluj sub-basin is significantly influenced by the magnitude of snowfall of one station, otherwise it also suggests a linear increase with altitude. Snow gradient in the Spiti sub-basin (43mm/100m) is determined to be more than four times than in the Baspa sub-basin (10mm/100m). Standard errors of the mean snow water equivalent increase with altitude in both Spiti and Baspa sub-basins with a higher variability in the Spiti sub-basin.

8. Ratio of snowfall to annual precipitation varies linearly with altitude. All the stations recorded more than 60% snow contribution in annual precipitation. An extrapolation of this linear relationship suggests an equal contribution of rainfall and snowfall in annual precipitation at about 2000m. Moreover, above 5000m altitude all the precipitation falls as snow.

Spatial correlation function for Satluj and Beas basins

Spatial correlation for annual rainfall and snowfall was studied for both Satluj and Beas basins. Correlation coefficient (r) for rainfall and snowfall series were determined for different ranges. Spatial correlation functions were developed using correlation coefficients and following equation;

$$r(d) = r(0)\exp(-d/d_o) \qquad (1)$$

where, r(d) is the spatial correlation coefficient, d is the distance between two stations and d_o is

Table 5: Spatial correlation function for different Himalayan ranges of Satluj and Beas basins (Singh and Kumar, 1997).

Basin	Himalayan Range	Type of precip.	Spatial correlation function	Distance for r = 0.75
Satluj	Outer (Leeward)	Rain	$r(d)= 0.984\exp(-0.0038\,d)$	71 km
Spiti sub-basin	Greater (Leeward)	Snow	$r(d)= 1.200\exp(-0.0277\,d)$	17 km
Baspa sub-basin	Greater (Leeward)	Snow	$r(d)= 1.039\exp(-0.0280\,d)$	12 km
Beas	Outer (Leeward)	Rain	$r(d)= 0.877\exp(-0.0035\,d)$	43 km
	Middle (Windward)	Rain	$r(d)= 1.231\exp(-0.0441\,d)$	5 km
	Middle (Leeward)	Rain	$r(d)= 0.851\exp(-0.0037\,d)$	34 km

the correlation radius (the distance at which the correlation reduces e times) and r(0) is the correlation function at zero distance). Spatial correlation functions obtained for different ranges and type of precipitation, and distance for which spatial correlation coefficient is greater than 0.75 are given in Table 5. For other regions either data were not sufficient to develop the spatial correlation coefficient or the spatial correlation coefficient was very poor.

Spatial correlation of rainfall appears to be higher in the outer Himalayas than other ranges. It may be because of lesser relief in the outer Himalayan range as compared with middle and greater Himalayan range. Spatial correlation of rainfall is better on the leeward than the windward side in the middle Himalayan range. Recently, Loukas and Quick (1996) made similar studies for storm precipitation in the southwestern British Columbia, Canada. They reported that all types of storms indicated correlation coefficients larger than 0.75 for distances smaller than 32km. Rainfall spatial correlation over the British Columbia mountain watershed matches that on the leeward side of the middle Himalayan range of the Beas basin.

4.2 Temperature

Air temperature is considered to be the best index of heat transfer processes associated with melting of seasonal snow and glacier. Estimation of evaporation is also made using temperature data. There are numerous snowmelt models based on the temperature index method of snowmelt computation. In the snowmelt models having option for snowmelt computation using energy balance approach based on simplified or generalized equations, air temperature is required. The form of precipitation, rain or snow is also determined by the air temperature prevailing during the precipitation period. Once the form of precipitation is determined, the models take into account the different pattern of contribution in the run-off for the snowfed and rainfed area. In computing snowmelt because of rainfall on the snowpack, air temperature plays an important role. The temperature and precipitation are the only meteorological data which are available at few stations in the snowbound catchments.

Generally, most of snowmelt simulation models have been designed with the concept of partitioning a basin into elevation zones or bands according to the relief of the basin. Partitioning provides the ability to account for spatial and temporal variation of the physical and hydrologic characteristics, climate variables and system response. The status of the network in the mountainous and snowbound catchments, generally, is not adequate because of inaccessibility and hazardous terrains. Only a few basins may have a network to meet the requirement of snowmelt models. Consequently, the temperature data is distributed to various elevation zones in the basin on the basis of temperature lapse rate (TLR) defined as rate of decrease of temperature with elevation. The TLR changes from region to region and varies with time. Most of the models use a fixed temperature lapse rate, some use a seasonally varying lapse rate and some models use a lapse rate formulation derived statistically from daily observations. However, daily observations have been used on a statistical basis, lapse rates calculated each day from temperature data at different elevations have not yet proved to be useful (WMO, 1986). Some models use monthly data average over a number of years to arrive at values of separate maximum and minimum temperature lapse rates for each month of year.

In our country, the distribution of temperature at various elevation zones in a basin has been made using a fixed value of TLR (Thapa, 1980; Bagchi, 1981; Jeyram et al 1983; Agarwal et al, 1983; Seth, 1983; Roohani, 1986; Singh, 1989). Devi (1987) has made an attempt to analyse the typical variation of temperature in some places in the Uttar Pradesh Himalayas. It was observed that TLR of surface temperature actually observed, both in valley bottoms as well as along mountain slopes, is less than 6.5 °C/km., the mean environmental lapse rate. It was reported that temperature generally decreases up to elevation of 1700 m. Above this elevation temperature first shows an increasing trend and then runs in zig-zag manner from 1800 to 2300m. Bhutiyani (1989) has reported a variation of TLR from 3.8 to 9.6 °C/km in the month of August on Chhota Shigri Glacier in Himachal Pradesh. The results presented by Bhutiyani (1989) are found based on very limited period, about a month's data only. Upadhyay et al. (1989) have analyzed the monthly normals of surface temperature for various stations situated at different altitudes in the western Himalayas and reported that monthly lapse rates of temperature are fairly constant. They have considered data from the stations scattered over a wide range of mountainous part of Himalaya. Presence of several valleys and high crests between the considered stations affect the distribution of temperature. It occurs because of numerous cold and warm air pockets in different valleys at different altitudes. Therefore, TLR values determined from the stations located in different valleys does not represent the value which can be used for temperature interpolation for snowmelt estimation in a basin. Singh (1991) studied the behaviour of TLR in the Satluj basin (Figure 2 and 3). The maximum and minimum temperature data is observed at 5 observations namely Rampur, Kalpa, Namgia, Rakchham and Kaza in the catchment area where snowfall occurs.

The effect of temperature lapse rate value on snowmelt runoff computation for a small sub-basin of Beas has been illustrated in Figure 4 (Singh, 1991). To show the effect of lapse rate value on snowmelt run-off computation, the snowmelt run-off model (SRM) developed by Martinec and Rango (1983), was run with values of TLR 5.5, 6.5, and 7.5 °C/km for the months of April and May. The results showed that even a minor change in the value of TLR contributes very much into the snowmelt computation. It was noticed that if the TLR is increased or decreased by 1°C/km, keeping other parameters same, the total snowmelt runoff volume varied in the range of 28-37% for two months period. Evidently this variation may be higher if the computation is made for longer durations and basin is having a larger area. The results indicate that consideration of TLR for snowmelt studies is one of the most important aspect, which should be handled very carefully.

The mean air temperature lapse rate (environmental lapse rate) in mountains is generally considered to be around 6.5 °C/km. There are seasonal variations and also a high varying surface characteristics and local climate effects. This study shows that a constant value of TLR for all the basins should not be used and efforts should be made to use the representative value of TLR in accordance with temperature data used (maximum, minimum or mean temperature). The computation of TLR for the proposed basin for snowmelt studies should be made at least on the monthly basis for the snowmelt season (March-June). For higher accuracy, a separate value of TLR may also be used for each month in snowmelt season. The use of derived TLR value would help in accurate estimation of snowmelt run off and the other parameters used in the modelling studies would also be calibrated properly.

4.3 Evaporation

The water balance of a catchment is basic to any hydrological study, but it is only usable if all the acting components are determined. Besides precipitation and runoff, the analysed complex system encompasses the areal evapotranspiration and soil water storage in the catchment. The

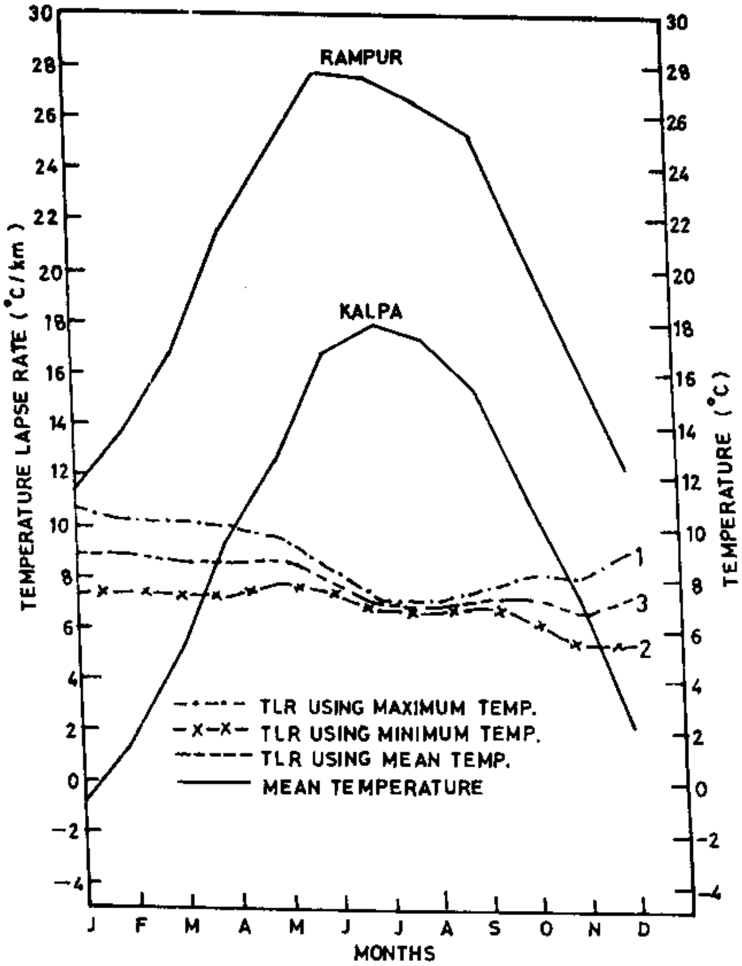

FIG.2 MEAN MONTHLY TEMPERATURE AT RAMPUR AND KALPA AND TLR BETWEEN THESE TWO STATIONS

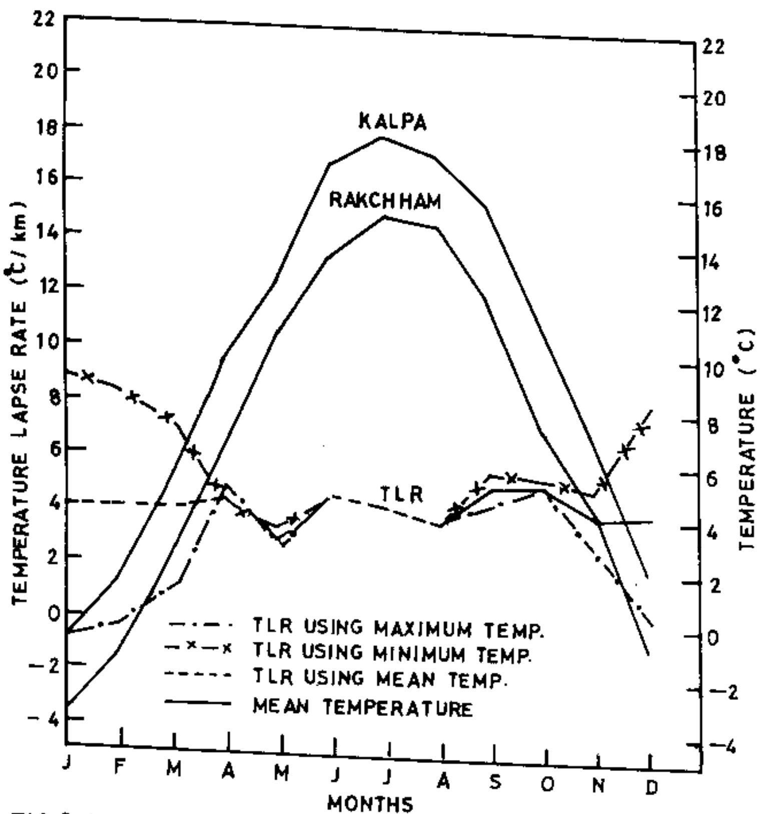

FIG.3: MEAN MONTHLY TEMPERATURE AT KALPA AND RAKCHHAM AND TLR BETWEEN THESE TWO STATIONS

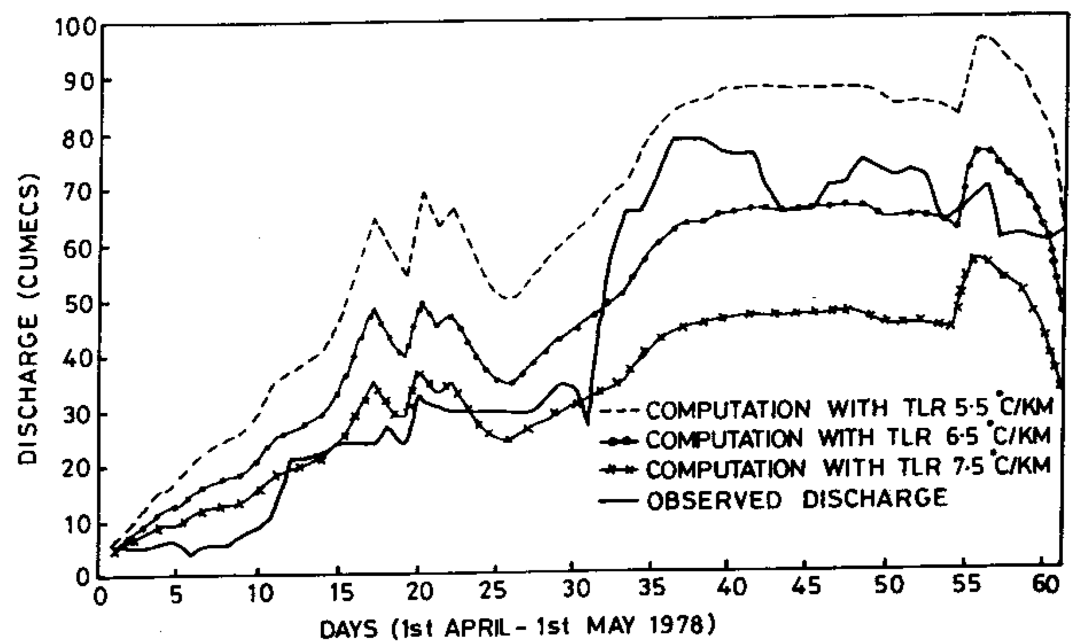

FIG.4: COMPUTED SNOWMELT RUNOFF IN BEAS BASIN WITH TLR VALUES 5·5, 6·5 AND 7·5 °C/KM

evapotranspiration represents the evaporation from various surfaces and transpiration by vegetation. In most cases, the actual evapotranspiration is calculated from meteorological data only and therefore the actual transpiration, which plays so decisive a role in the environment, is just estimated or not taken into account at all. The spatial and temporal variations of evaporation in the high mountains is not much known. The measurement of evaporation is included in the routines of some meteorological network stations, but typically these data are observed only in a few full scale climatological stations. Directly measured transpiration as an element of actual evapotranspiration is the biological component in the hydrological cycle and hence affected by anthropogenic influences and possible climate changes. Direct measurements of transpiration are carried out in experimental and research stations only (Molnar and Meszaros, 1990). However, the role of transpiration depends in turn on the type of vegetation cover, and so it is highly related to all human activities in the very fragile mountainous ecosystems. Naturally, this same component will be among the first influenced by climate changes. Therefore, the availability of evaporation and/or evapotranspiration data is very limited, and other approaches of estimation are used. Due to the many problems associated with direct evaporation measurement and the heterogeneity of topography, soil types, vegetation, and climatological gradients in such catchments, very little regional information is obtainable on the subject Sometimes, mean monthly values of evaporation are used where these data are not available (Mendel and Pekarova, 1995). Mean annual values of evaporation are relatively easier to obtain because these are determined from maps.

The importance of evapotranspiration in water balance studies in special cases has been shown by Miklanek (1992). A decrease in evapotranspiration with altitude, as expected, has been confirmed by actual computation of evapotranspiration for Tatra mountains in Central Europe (Table 6 and Figure 5). In Tatra mountains the elevation of stations used in the study varied from 115 m to 2635 m. The evapotranspiration coefficient showed that in the lowlands the actual evapotranspiration represents upto 80% of the precipitation, in main valleys it is about 70%, in the middle slopes varies between 60% to 40% and on the tops it decreases to 20% of the precipitation. It exhibits that evapotranspiration for the Tatra watersheds forms about one third of the water balance and this should be taken in account in different water management studies in this region.

Jong and Ergenzinger (1996) studied the spatial variations of evaporation for Dischma valley, Switzerland. The period for the study was from 15 June to 15 September which included summer season/snowmelt season and late autumn. It was found that evaporation is very heterogeneous for the different sites in the Dischma valley. This is due to very diverse radiation and wind conditions, both substantially influenced by the diversity in topography. However, calculated evaporation does not vary as much due to the minimal influence that some of the main factors have on the Penman formula. It was suggested that evaporation has to be considered in terms of the vegetation community, aspect, windiness and exposure to radiation. In addition, the rapidly increased response of plants to snowmelt and the decreased response of plants after extensive rainfall are both factors not adequately weighted in the formula. Actual evaporation is highest immediately after snowmelt and during the summer peak, but not during other similarly warm periods. New approaches and new adaptation of the traditional formulae are therefore required for high altitude catchments.

Table 6: Mean potential evapotranspiration (PET), actual evapotranspirations (ET), and ratio to precipitation (P) for 1956-1980 in the Tatra Watershed in Central Europe (Miklanek ,1992).

Stations	Elevation (m)	PET (mm/yr)	ET (mm/yr)	ET/P (%)
Hurbanovo	115	859	431	79
Lipt. Mikulas	576	662	482	70
Lipt. Hradok	643	622	466	69
Poprad	703	632	444	74
Tatr. Lomnica	850	565	473	59
Podbanske	972	588	492	53
Strbske Lake	1330	563	475	49
Skainate Lake	1778	496	466	36
Chopok Peak	2004	337	319	28
Lomnicky Peak	2635	297	288	22

There is a strong need to study the point and areal evapotranspiration in the mountainous regions. There are very limited studies are carried out on the spatial or altitudinal variation of evaporation or evapotranspiration for mountainous area of India, including Himalayas. Such studies are needed for the accurate assessment of water resources as well as modelling of rainfall-runoff, snow and glacier melt runoff.

Energy transport and exchange

Because of the dominant influence of energy fluxes on snowpack evolution and meltwater generation, efforts have been made to better understand spatial and temporal patterns in fluxes and assess their sensitivity to physical characteristics. A comprehensive set of data from the Emerald Lake basin in California revealed the changes in relative magnitudes of snow surface energy fluxes throughout the accumulation and snowmelt season (Marks and Dozier, 1992). Radiation accounted for 66-90% of the energy available for melt; sensible and latent heat fluxes were roughly of the same magnitude, but of opposite sign and therefore cancelled. About 20% (50 cm) of the SWE was lost to sublimation and evaporation over the course of the snow season. The seasonal contribution of soil heat conduction to the energy budget is negligible, but may significantly contribute to midwinter melt from the base of the pack. A 16-year time series of snow depths, air temperatures, and soil temperatures from a mid-continental site indicated that a snow depth of 42.5 cm was required to maintain steady soil temperatures (Sharratt et al., 1992).

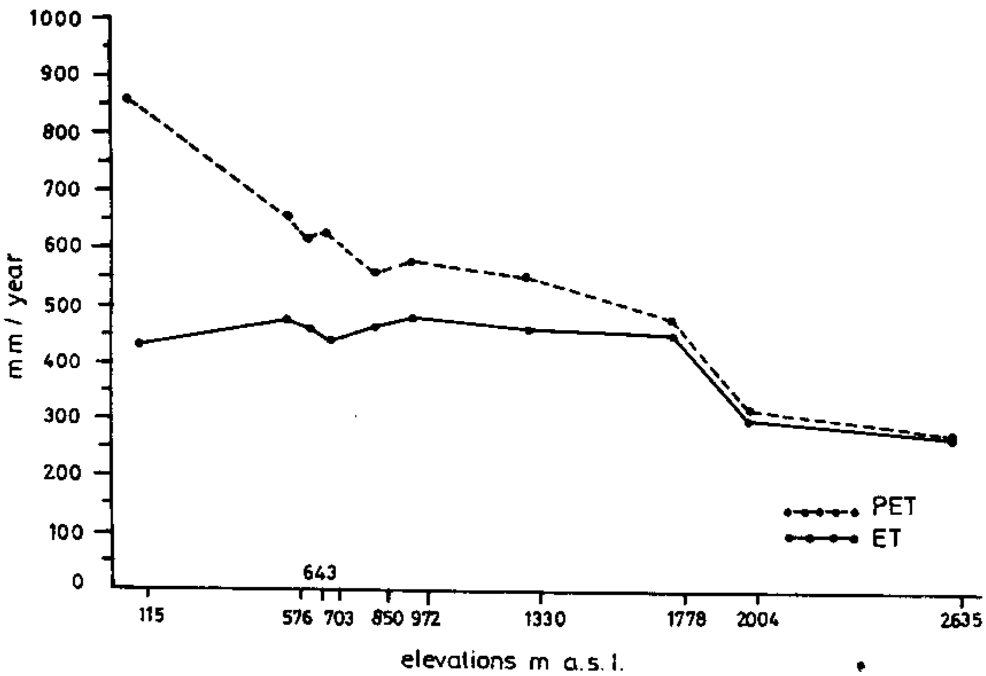

FIG.5 THE DECREASE OF POTENTIAL EVAPOTRANSPIRATION (PET) AND ACTUAL EVAPOTRANSPIRATION (ET) WITH ELEVATIONS IN THE TATRA WATERSHED IN THE CENTRAL EUROPE (MIKLANEK,1992)

Colbeck (1991) reviewed how layering can have profound effects on the physical properties of snow that are important in both thermal and hydrologic phenomena. Because of the importance of water-vapour diffusion in snow metamorphism, the water-vapour diffusion coefficient is of great interest. Changes in pore geometry during experiments to determine porous-medium parameters of snow make such determinations highly uncertain, and different experiments often lead to contradictory measurements (Colbeck, 1993). Sommerfeld et al. (1991) preconditioned the air flow prior to making air-permeability measurements on dry snow by passing the air stream through a column of snow similar to the one being measured, equilibrating the air with the conditions in the snow, thus reducing erosion of the test snow during the experiment. Their results suggested that a simple snow density-permeability relationship fit their data better than dimensionally correct specific-surface-area methods. Hardy and Albert (1993) made concurrent measurements of the permeability and physical properties of snowpack strata, finding that physical properties correlated poorly with permeability measurements and that stereological parameters correlated well for dry snow, but poorly for old snow and ice layer.Vapour transport does not contribute to thermal transport within the snowpack as much as conduction and advection by dry air. The effect of ventilation on the snow-pack temperature profile is most likely to be observed when the temperature gradient in the snowpack is not great or when air flow within the pack is relatively high (Albert and McGilvary, 1992). Meltwater movement, heat and solute transport, and snow metamorphism are affected by the advance of wetting fronts through the pack. As the meltwater penetrates the pack, the thermal characteristics of the pack change; therefore, the formation of roughly columnar zones, or flow fingers, of wet snow that contain moving meltwater penetrating the snowpack ahead of the main melting front can be observed insitu by monitoring the thermal changes within the pack (Sturm et al., 1993; Conway and Benedict, 1994).

The penetration of visible light into the snowpack has been suggested as a cause for the presence of a temperature maximum about 10 cm below the surface. This explanation appears less viable when radiative-transfer calculations are carried out with high spectral resolution (Brandt, 1993). Wavelengths that penetrate the snowpack (visible) are scattered back to the surface, and wavelengths that are absorbed are absorbed in the first few millimetres, indicating that subsurface absorption is minimal. A significant heating of the interior of the snow occurs when the scattering coefficient is low (e.g. blue ice) and when the thermal conductivity is low (e.g. near-surface depth hoar).

Heat flow was affected by the presence of trees, because of interception of snow by the collage resulting in irregular snow depths. For example, heat flow from tree wells can be more than twice that of undisturbed snow (Sturm, 1992). Snow ablation in forest openings depends on the size of the opening. Ablation from openings of diameter H (where H is the average height of adjacent uncut trees) was less than from openings of 0, 3 and 5 H because net radiation was least in the 1H diameter openings. Evaporation in the canopy-covered case (0H) was slightly greater than the 1H case because of night time radiative transfer from the canopy to the snow, which kept the canopy-covered snow warmer than in the openings, and because the canopy prevented nighttime condensation. Furthermore, lateral advective energy transfer between openings and forest canopy appears to be significant (Berry and Rothwell, 1992; Bernier and Swanson, 1993).

5.0 HYDROLOGICAL MODELLING STUDIES

River basins in mountain regions are characterized by strong areal variations of land surface features, including soil and vegetation. The driving forces behind these variations are the combined effects of geomorphological and meteorological climate processes. The corresponding hydrological system of such regions reflects this complexity in the great areal variation of its specific processes. Consequently, the modelling efforts directed towards the hydrological components of a mountain river basin need to take account of few particular characteristics of mountain systems. Great care has to be exercised for (i) vertical and horizontal structures when developing the model components, (ii) selection of an appropriate spatial resolution, and (iii) in the required temporal resolution of the model simulations, which much depends on the specific purpose of the model.

Seasonal snowpacks in the Earth's mountain ranges account for the major source of the runoff for streamflow over wide areas of the mid-latitudes. In most alpine regions, snowmelt runoff is responsible for the annual maximum instantaneous discharge and most of the annual flow (Bales and Harrington, 1995). There is big scope in India for implementation of new hydropower projects and the middle Himalayan region together with the north east region is identified for such developments. A general feature of the Himalayan region is that areas below an elevation of 4200 m experiences seasonal variation of snow cover. As a result, during hot weather periods, additional flows take place. Hence, reliable estimation of snowmelt is necessary for successful planning and design of structures across such streams. It is therefore important to understand the processes which are responsible for controlling the snowmelt runoff. Progress in snow hydrology over the past quadrennium has been heavily influenced by the availability and anticipated future availability of remote sensing data at times and spatial scales appropriate for addressing local, regional and global hydrologic issues (Dozier, 1992). Recent research has examined topics contributing to better estimation of snowpack properties, snow distribution, and melting of snow.

Understanding the linkages between the processes controlling snow accumulation and melt is critical to developing a predictive ability to describe the response of a basin to changes in inputs of water, energy. Due to their steep, variable topography, alpine catchments are characterized by a large degree of heterogeneity in the important properties controlling snow accumulation, snowmelt and meltwater runoff. The diurnal pattern in snowmelt results in a phase shifted and modulated pattern in streamflow with respect to snowmelt. Snowpack depth and snow-covered area decrease as the melt season progress, decreasing the time lag between peaks in meltwater generation and streamflow. This decrease in time lag was used to obtain a measure of catchment wide snowpack hydraulic conductivity in the 40-80 cm/hr range (Caine, 1992). In a recent National Research Council report (National Research Council, 1993), it was emphasized that "one of the main obstacles in understanding surficial processes is the high spatial variability of surface features and hydrologic variables". The hydrology of snow-covered areas was one of the "critical and emerging" areas identified in the report for future research. List of some snowmelt runoff model used for intercomparison by World Meteorological Organisation (WMO, 1986) is given Table 7. Most of these models take care of rainfall runoff processes for the snow free area of the basin. The drifting of snow in the mountain areas becomes important for both hydrology and avalanche studies because they depend on the distribution and thickness of snow in a basin. But, usually wind drifting of snow is not taken into account in such studies.

Table 7: List of some snowmelt runoff model used for intercomparison by World Meteorological Organisation (WMO, 1986).

S. No.	Model Name	Location/organisation	Year	Approach used
1	UBC Watershed model	University of British Columbia, Canada	1974	Degree - day index equation
2	CEQUEAU Model	University Of Quebec, Canada	1971	Degree - day index modulated by a solar radiation factor
3	ERM (Emperical Regression Model)	Bratislava, Czechoslovakia	1978	Degree - day index equation
4	HBV Model	Swedish Meteorological and Hydrological Institute	1976	Degree - day index equation
5	Tank Model	National Research Centre for Disaster Prevention, Japan	1963	Degree - day index equation
6	SRM	Federal Institute for Snow and Avalanche Research, Switzerland, and NASA, USA	1973	Degree - day index equation
7	SSARR Model (Streamflow Synthesis and Reservoir Regulation	Corps of Engineers, Portland, Oregon, USA	1967	Degree - day index equation
8	PRMS Model (Precipitation Runoff Modelling System	US Geological Survey, USA	1973	Energy balance approach
9	IHDM (Institute of Hydrology Distributed Model)	U.K.,Denmark, France	1979	Energy balance approach

Areal non-uniformity of snow cover apparently causes obstacles in simulating runoff in snow covered basins. In high mountain areas like the Himalayas, the problem is most severe. The natural vertical redistribution of snow deposits and impact of drift and frequent avalanches on the distribution is not known. Consequently, vertical gradients of snow depth and water equivalent are not known under such circumstances. Even areal approaches using modern monitoring techniques, such as remote sensing and distribution models, can only provide approximate pictures of reality. In areas where changes in land use influence the rainfall runoff process, it is necessary to take into account any trend in the affecting phenomena. For the time being, similar situations are developing in many areas of the Himalayas. Analyses and trials have been carried out during the last years which had as their main goal to explain different aspects of the runoff process in these hilly regions (Braun et al., 1993; Fukushima et al., 1991) or in other areas (Braun and Aellen, 1990; Rango and Katwijk, 1990). The results of the hydrological experiments were presented to clarify the sensitivity of runoff towards actually occurring and/or assumed environmental changes due to climatic warming. However, since snow deposits are usually affected also by land-use changes, it would be desirable to evaluate both factors for their impact on hydrological regimes simultaneously.

Both conceptual and physical approaches have been employed in snowmelt-runoff modelling. Conceptual models propose a mathematical relationship between snowmelt and measured quantities; thus melt can be calculated without treating in detail all the physical processes and parameters that affect snowmelt. Conceptual models have the benefit of requiring less input, but suffer from the uncertainty that parameters estimated under one set of model conditions are applicable to other conditions. Conceptual models based on temperature index methods have been used to illustrate the sensitivity of snow-covered basins to climate change (van Katwijk et al., 1993), and such efforts will improve with the development of models that more directly incorporate the process of radiative exchange of heat into the calculation .

One of the main obstacles to physically based modelling is the accumulation of the necessary meteorological and snow-cover data to run, calibrate, and validate such models. For example, basin discharge has frequently been used as the sole physical criterion of model calibration and performance assessment for conceptual snowmelt models. But as it is an integrated response to melt and runoff, basin discharge is not sufficient to discriminate between the effects of the multiplicity of data inputs driving physical models and that distributed snow-cover data are required to assess model performance (Bloschl et al., 1991a; Bloschl et al, 1991b). A comparison of a detailed research-oriented point-snowmelt model and a simpler operational model revealed that in the absence of detailed measurements of snowpack variables, a detailed physical model of the snowpack is needed to reduce the need for parameter calibration (Bloschl and Kirnbauer, 1991). Several simplifying assumptions are necessarily made in order to solve physically based point models, thus limiting their validity in many field situations (Illangasekare et al., 1990). Nevertheless, physically based models are being tested and applied in alpine catchments (Ranzi and Rosso, 1991).

Estimation of distributed snow-water equivalence (SWE) is challenging because of the many factors that affect its distribution, and the small correlation length of the SWE spatial distribution. Further, difficulties associated with accurately determining the time of maximum accumulation present a problem for snowmelt-runoff forecasters. The simplicity of regression models makes them an attractive means of estimating SWE because of the large amount of work required to directly measure SWE on the catchment scale. Using only meteorological variables, SWE estimates with variations in the 60-81% range were obtained for sites in New England (Samelson and Wilks, 1993), but such methods can be confounded by the highly variable topography found in mountainous areas. Photogrammetry shows promise as a technique to measure snow depth in alpine catchments by differenciating snow-free digital elevation surfaces from snow-covered digital elevation surfaces (Cline, 1993). Redistribution of snow by wind was found to affect snowmelt runoff in an Arctic watershed in Northern Alaska (Kane et al., 1991). Thus, additional accumulation of snow due to wind becomes an important consideration in the timing of snowmelt runoff (Hinzman and Kane, 1991). Cline (1992) used a geographical information system based analysis of slope an terrain variability to model the redistribution of snow by wind in alpine terrain.

An approach to modelling spatially distributed snowmelt in steep, alpine basins was proposed using net potential radiation, distributed across the basin using a digital elevation model, as the main factor determining relative snowmelt (Elder et al., 1991). Such an approach enables using a detailed, physically based snowmelt model for each physically different sub-region of the basin at the scale of interest. Testing this approach on the 1.2 km^2 Emerald Lake basin in California's Sierra Nevada suggests that little information is lost in going from a 5-m to 25-m grid, but that use of a 100-m grid may result in significant inaccuracies (Bales et al., 1992).

Snowmelt models that work well at the catchment scale may still be inadequate for some applications, such as integrated chemical modelling (Bales et al., 1992) or erosion (Tarboton et al., 1991). For erosion, one needs to capture the spatial distribution of snowmelt delivery at each point on a hillside or small basin, and use time steps on the order of one to a few hours. Similarly, integrated chemical models require detailed spatial distributions of snowmelt to properly route meltwater through soil parcels.

In India rainfall runoff modelling studies are also limited both in the Himalayan region and mountainous region in the middle and southern part of India. Various models have been tested and their performance is evaluated in different river basins. Jain and Ramasatri (1992) attempted to derive the representative unit hydrograph for Hemavati sub-basin up to Sakleshpur, an area of 600 km^2, using HEC-1 flood hydrograph package based on Clark's approach. The rainfall-runoff data for four events are considered for caliberation, The performance of the model was validated by reproducing the four independent events. The model was found capable of reproducing the flood peak magnitudes, time to peak and overall flood hydrograph reasonably well. Kandasamy et al. (1992) tested four different rainfall-runoff models to two types of basins namely Chaliyar basin comprising four sub basins and Kuttiyadi basin of Kerala in southern part of India. The models tested were linear, linear perturbation, constrained linear system and the TANK model. They concluded that Linear Perturbation Model be preferred for application to small, mountainous catchments subjected to monsoon rainfall. James et al..(1992) also reported similar results about applicability of linear perturbation model for run-off estimation from five catchments of the western Ghat region. The model efficiency are very good for mountainous catchments having area gretaer than about 1000 km^2, model efficiency reduced as the size of the basin decreased.

6.0 REMOTE SENSING APPLICATIONS

The problems of snow cover estimation in mountainous areas are still unsatisfactorily solved. The main cause is the acute shortage of up-to-date information on the distribution of precipitation and snow cover in mountains. From the standard observations it is impossible to determine the snow conditions in most of the mountain areas by earlier methods. The methods of remote sensing that are successfully used for snow cover estimation on plains still do not give satisfactory results for mountainous regions. A major theme of progress in snow hydrology over the past quadrennium has been the expanded use of remote sensing for determining snow properties, which then are used to estimate snow distributions and snowmelt runoff. There has also been a move towards development of physically based snowmelt models to be used with this emerging data, particularly for alpine areas. The coupling of remote sensing and physically based approaches will enable making not only more-accurate basin-scale forecasts, but will also provide spatially distributed estimates of snowmelt. A good review of the remote sensing applications in snow studies is given by Bales and Harrington (1995).

Because of the difficulty of making field measurements in snow-covered mountainous regions, remote sensing is attractive as a means of measuring snow-cover properties. There is always a big scope of using remote sensing information for the modelling of snow melt runoff and to determine the snowpack properties. A major breakthrough in remote sensing application is the deciphering of information on the snow covered area and its variation with time in the inaccessible mountainous areas. Several researchers have reported on efforts to incorporate remote-sensing data into snowmelt-runoff modelling. Rango (1993) reviewed the progress that has been made incorporating remote-sensing data into regional hydrologic models of snowmelt runoff. The National Oceanic and Atmospheric Administration's Advanced Very High Resolution Radiometer (AVHRR) sensor provides daily views over large areas (1000 km swath) and snow-cover maps are produced operationally. Estimates of snow-covered area based on remote sensing data can significantly improve the performance of even simple snowmelt models in alpine terrain (Kite, 1991; Armstrong and Hardman, 1991). For operational purposes, empirical approaches using combining remote sensing data to estimate snow-covered area, and snow-depth networks to estimate SWE are continuing to improve (Martinec and Rango, 1991; Martinec et al., 1991). Mcmanaman et al. (1993) have combined airborne and ground based measurements to produce gridded SWE estimates for the upper Colorado River region.

For hydrological applications, the primary differences between remote sensing instruments are in their spatial, temporal, spectral, and radiometric resolutions. Optimizing one type of resolution generally involves some sacrifice in other types of resolution, e.g., the Landsat Thematic Maper (TM) (30 m pixel size) has a much better spatial resolution than the AVHRR (1 km pixel size). However, the AVHRR can provide daily coverage of a given point, whereas the TM can only provide bi-weekly coverage. The possibilities for detecting snowpack properties are largely determined by the wavelength being recorded by the remote sensing instrument. Visible and near infrared wavelengths, because they do not penetrate far into the snowpack, mainly provide information about the surface of the snowpack. However, microwave wavelengths can penetrate the snowpack, thereby providing the opportunity to collect volume integrated data.

Both visible and near-infrared, and passive microwave remote sensing data are being used to develop estimates of snow-covered area for operational forecasting of snowmelt, much of which is currently done using conceptual models. Development of accurate snow-cover information for areas with steep, variable topography requires higher resolution data than are currently available from operational remote sensing instruments. Models using the higher-resolution satellite data are expected to give good results as compared to the data acquired from aircraft platforms. Determination of other snowpack properties in alpine areas, such as grain size and albedo (from visible/infrared) and snow-water equivalence (from active microwave) are topics of continuing research. Progress in both algorithm development and testing with field data sets shows that obtaining these properties is achievable. The volume integrating capability of microwave remote sensing has received much attention, because it offers the possibility of determination of snowpack properties. However, the lack of multipolarizing SAR will limit the ability to reliably estimate SWE in alpine areas.

Operational hydrology still depends largely on ground-based methods to develop estimates of SWE. Since remote sites are often not telemetered, it is desirable to develop correlations between remote, high-elevation sensors and those at lower elevations. Limited success toward this end was reported in a California study involving ten pairs of high and low elevation sensors (McGurk et al., 1993). In a 4000-m elevation basin in China's Tien Shan, comparison between an intensive snow survey and SWE estimates based on a stake network suggest that while the stake network adequately estimated mass balance, the spatial details of the SWE distribution were not well reproduced (Elder et al., 1992). Runoff from the surrounding basin underwent delays of 5-15 days before reaching the stream, confounding problems with snowmelt modelling based primarily on energy balance (Kattelmann and Yang, 1992).

6.1 Visible and near-infrared measurements

By exploiting the sensitivity of near-infrared snow reflectance to snow grain size, hyperspectral data were used to recover grain size using regression of Airborne Visible and Infrared Imaging Spectrometer (AVIRIS) observations and modeled spectra (Nolin and Dozier, 1993). Comparisons of satellite and ground measurements of snow reflectance highlight the variability in satellite-derived reflectance due to grain size and surface contamination (Winther, 1992).

AVHRR data have been routinely used for binary classification of snow covered vs snow-free area (Xu et al., 1993). Because of the low (1 km pixel size) resolution of AVHRR data, the estimation of snow-covered area would benefit from adoption of mixed-pixel image processing methods. Such a method was implemented for AVIRIS imagery by Nolin et al. (1993), and for TM imagery by Davis et al. (1993a).

Measurement of snow albedo is useful because of the importance of the snowpack energy budget in snowmelt calculations. Efforts to validate reflectance derived from TM imagery using ground-acquired spectrometry were carried out in the Canadian Arctic and the importance of solar zenith angle in determining reflectance was highlighted (Hall et al., 1992; Hall et al., 1993). For off-nadir look angles in the forward-scattering directions, reflectance consistently exceeded 1.0 due to enhanced forward scattering. Because of the importance of bidirectional effects in reflectance of snow, spectral-reflectance curves and the nature of the directional anisotropy were found to be more useful in discriminating different snow properties. Hall et al. (1990) reported improvements in estimating the reflectance of snow-covered surfaces, which will be quite important in the global albedo estimates that should result from the planed NASA Earth Observing System (EOS) Moderate Resolution Imaging Spectrometer (MODIS). Spatial resolution and terrain both affect the calculation of surface energy budgets. TM imagery has been used in conjunction with-digital-elevation data to derive surface albedo and brightness temperature. In mountainous terrain, irradiance calculations must account for reflection of nearby terrain features; calculations thus carried out yield irradiance within about 10% of pyranometer-measured irradiance (Gratton et al., 1993). Use of TM imagery to determine albedo in improved surface energy balance modelling was reported by Dugvay (1993). Improvements in radiative-transfer calculations were reported for the microwave (Kuga et al., 1991) and for the visible and infrared, using both ground-based and air-borne instruments (Nolin et al., 1990). The spectral reflectance of snow was modelled by Davis et al. (1993 b) and compared favourably with field observed reflectance.

6.2 Microwave measurements

Microwave remote sensing is promising because of its ability to penetrate the snowpack, and its capability to acquire data in cloudy or night time conditions. Microwave signal from the snowpack is a volume-integrated response to wetness, density, surface and subsurface roughness. Since longer wavelengths penetrate the snowpack to greater depths, comparison of different bands can aid in discriminating between the effects of surface roughness, stratigraphy, and snow wetness (Rott and Davis, 1993). Active microwave sensors such as being deployed on the First European Remote Sensing Satellite (ERS-1) and Canadian RADARSAT offer the possibility to observe seasonal snowcover characteristics in detail over the entire snow-cover characteristics in detail over the entire snow-cover season (Way et al., 1993). In one simulation of RADARSAT data, snow-cover classification accuracy was 80%, comparable to aircraft Synthetic Aperture Radar (SAR) (Donald et al., 1993). Comparing a classification of snow-covered area based on SAR with that done using TM suggests that a SAR based classification is sufficiently accurate to substitute for visible-and-near-IR based estimates when such data are not available, for example due to cloudiness (Shi and Dozier, 1993). The sensitivity of SAR to wet snow offers the possibility to determine the onset of snowmelt (Bernier and Fortin, 1991). Shi et al. (1991) reported the capability to discriminate between snow, glacier, and rock regions in an alpine basin using C-band SAR imagery when topographic information is not available.

A method of estimating snow depth based on the ratio of and difference between HH (horizontal transmit; horizontal receive) and VV (vertical transmit; vertical receive) polarizations from multifrequency radar was presented (Shi et al., 1990); unfirortunately, these multiple polarizations are not available on current satellites. Multiple frequencies and polarizations can also help solve problems such as mixed pixels (snow and vegetation) encountered in a test in Rio Grande basin (3419 km^2) in southwestern Colorado (Chang et al., 1991a).

Measurements of the change in microwave brightness temperature for ice overlain by an evolving snowpack were made and the change in brightness temperature due to increased volume scattering as snow grains grow, were demonstrated (Lothanick, 1993). Sturm et al. (1993) observed that the reduced microwave emissivity of snow cover versus soil was entirely due to a coarse-grained depth-hoar layer; therefore, other stratigraphic features of the snowpack could not be distinguished by the microwave emissivity alone, and Walker and Goodison (1993) report difficulty classifying wet snow in the presence of a vegetation canopy using passive microwave measurements. Modelling has shown the importance of using a distribution of grain sizes when modelling scattering of microwaves in snowpacks (West and Tsang, 1993; Boyarskij et al., 1993; Shi et al., 1993). Kuga et al. (1991) present a radiative-transfer model for three frequencies, 35, 95 and 140 GHz and show that the back scattering coefficient is sensitive to liquid-water content at all three frequencies, with 35 GHz being the most sensitive.

Passive microwave signals are also sensitive to the liquid-water content snow and therefore are useful to estimate the wetness of snow wetness. The sensitivity of passive microwave signals to snow wetness aids in determining the onset of spring melt and the occurrence of multiple melt events during the winter (Goodison and Walker, 1993). Wang et al. (1992) found good agreement between aircraft and microwave depth estimates for an Alaskan snowpack; but they also noted that the radiometric correction for the effect of atmospheric absorption is important at all wavelengths used for a reliable estimation of snow depth. Scanning multichannel microwave radiometer (SMMR) microwave brightness temperatures and ground-based snow-depth measurements were used with a regression routine to estimate snow depth over heterogeneous terrain in the Canadian high plains (Chang and Chiu, 1991) and western China (Chang and Tsang, 1992), and have the potential to be applied to mountainous areas such as the Colorado River basin in the U.S. (Chang et al., 1991b; Josberger et al., 1993).

The difficulties of integrating field, aircraft, and satellite measurements were illustrated by measurements made on Arctic snow covers by Hall and Sturm (1991). Good correlations were observed between satellite-borne and aircraft-borne microwave brightness temperatures, but the cause of a regional minimum in brightness temperature observable in both aircraft and satellite data could not be seen in ground based measurements. This may be attributable to the difficulty in capturing the variability of snow stratigraphy in the field using snow pits. Wind affects the microwave backscattering properties of snowpacks by promoting evaporative loss, decreasing snow wetness (Koh, 1992).

7.0 CLIMATE CHANGE EFFECT

The atmospheric concentration of CO_2 and other trace gases has increased substantially over the last century and double concentration of CO_2 is expected by the middle or latter part of next century, if no control measures are adopted (NAS, 1979; Pearman, 1980). This steady increase in the concentration of greenhouse gases has resulted in global warming. The global mean surface air temperature has increased by 0.3 to 0.6 EC over last 100 years (Jones et al., 1990). Further, average global surface temperature will rise by 0.2 to 0.5 EC per decade during next few decades, if human activities which cause greenhouse gas emissions continue unabated (IPCC, 1990). The striking feature, however, is that inter-annual variability of global temperature is much larger than the trend. Under the double CO_2 concentration scenario, precipitation may increase or decrease by as much as 15% (IPCC, 1990).

Several studies of climate variability on both short and long time scales have been carried out to establish climate changes over India (Jagannathan and Parthasarathy, 1972; Hingane et al., 1985; Sarker and Thapliyal, 1988; Thapliyal and Kulshrestha, 1991). It is observed that a warming of the Indian sub-continent by 0.4°C has taken place over the period 1901-1982 (Hingane et al., 1985). This warming since 1900 is broadly consistent with observed global warming over the last century. Thapaliyal and Kulshreshtha (1991) examined the trend of annual rainfall over India and reported that the five years running mean has fluctuated from normal rainfall within " one standard deviation. Based upon the results from high resolution General Circulation Models (GCMs), the IPCC (1990) reported that for the Indian sub-continent by 2030 on 'Business-as-Usual' scenarios (if few or no steps are taken to limit greenhouse gas emissions), the warming varies from 1 - 2 °C throughout the year. Precipitation changes little in winter and generally increases throughout the region by 5 - 15% in summer. Lal et al. (1992) studied the impact of increasing greenhouse gas concentrations on the climate of Indian sub-continent and its variability by analysing the GCM output data of the Hamburg global coupled atmosphere-ocean circulation model. The model results obtained from the greenhouse warming experiment suggested an increase of over 2 °C over the monsoon region in the next 100 years. The mean annual increase in surface runoff over the Indian subcontinent simulated by the model for the year 2080 is estimated to be about 25% (Lal and Chander, 1993). Singh and Kumar (1997) studied the effect of climate change scenarios on the streamflow characteristics of a high altitude Himalayan river. Their findings are discussed in this report elsewhere.

The warming of the earth-atmosphere system is likely to change temperature and precipitation which may affect the quantity and quality of the freshwater resources. One of the most important impacts to society of future climatic changes is expected on the regional water availability, specifically the timing and magnitude of surface runoff and soil moisture fluctuations (Gleick, 1986; WMO, 1987). Existing global models suggest that climatic changes will have dramatic impacts on water resources leading to major alterations of regional water systems. For example, a study based on GCM indicated that streamflow from the rivers in western US will be reduced by 40 to 75% (NRC,1983). Rind and Lebedeff (1984) used a GCM to assess the effect of doubling CO_2 on hydrological variables and concluded that precipitation would increase by about 11% and evaporation would increase proportionally, while snowpack would decrease by 20% due to higher temperatures. Because quantitative estimates of the effects of climate change on the hydrology of different regions are essential for understanding, planning and management of future water resources systems, therefore, the problems of global warming and its impact on water resources has received considerable attention in the recent years. There have been several co-ordinated efforts by World Meteorological Organisation (WMO), Inter-governmental Panel of Climate Change (IPCC), United Nations Environmental Programme (UNEP), and International Council of Scientific Union (ICSU) to bring together experts involved in projects concerned with climate variability and change, their impact on hydrology and water resources, and to identify the problems in this area. Further, WMO et al. (1991) suggested that possible effects of climate change in the design and management of water resources systems should also be examined.

Applications of hydrological models to study hydrological response of a basin under changed climatic scenarios

The prospect of climatic change is particularly critical to regions with seasonal snowcover, because increased frequency of rain-on-snow events, changes in precipitation volumes, changes in timing of accumulation and ablation seasons, and changes in location of the snow line all affect snowmelt runoff occurrence and the ecosystemic availability of water (Lettenmaier and Gan, 1990). It cannot be overemphasized that the results of such modelling efforts depend on what the altered climate is assumed to be, and such assumptions are themselves highly uncertain. Calculations to determine the countervailing effects of increased evapotranspiration and increased precipitation are confounded by uncertainty in such factors as the effect of increased atmospheric CO_2 on transpiration and precipitation volume (van Katwijk et al., 1993). Karl et al. (1993) found that 78% of the variance of regional snowcover can be explained by anomalies in monthly mean maximum daily temperature. Cooler, seasonally snow-covered areas have a longer memory of the past season's precipitation than do warm low-elevation basins. Since synoptic-scale weather patterns drive the catchment scale accumulation of snow, large-scale averages of temperature and precipitation perform nearly as well as local measures when assessing the annual variability of individual streams or watersheds (Cayan et al., 1993). In an analysis of station records for 1951-85, snow accumulation anomalies in the Rocky Mountain states fell into three types: (i) years where snow accumulation deviates throughout the whole region, (ii) years with a distinct north-south gradient, and (iii) average years (Changnon et al., 1993). Increased winter and early spring streamflow for two streams in the Sierra Nevada of California during the period 1965-1990 versus 1939-1964 were attributed to small increases in temperature, which increased the rain-to-snow ratio at lower altitudes and caused the snowpack to melt earlier in the season at higher altitudes (Pupacko, 1993). Reductions in forest density and area as well as climate warming would be expected to increase the severity of peak spring floods (Kattelmann, 1991). Rango and van Katwijk (1990) reported that increasing temperatures could markedly advance the onset of snowmelt runoff, increasing spring runoff and decreasing runoff later in the year. Using satellite-derived snow cover maps for sites in Alaska, Canada, Scandinavia, and Siberia, it was determined that there has been a tendency towards earlier snowmelt during the 1980's (Foster et al., 1991).

The studies carried out to demonstrate impacts of climate changes on various components of hydrologic cycle may be classified broadly into two categories; (i) studies using GCMs to predict impact of climate change scenarios (U. S. Department of Energy, 1980; Gleick, 1987; Cohen 1986;, IPCC, 1990, Sausen et al., 1994, McCabe, 1994, Loaiciga et al., 1996) and, (ii) studies using hydrological models with assumed hypothetical climatic inputs (Nemec and Schaake, 1982; Nemec, 1989; McCabe and Ayers, 1989; Sanderson and Smith, 1990; Thomsen, 1990; Rango, 1992, Cayan and Riddle, 1993; Burn, 1994; Rango and Martinec, 1994, Chiew et., 1995). While the GCMs are invaluable tools for identifying climatic sensitivities and changes in global climate characteristics, but their grid system is generally too large to assess the impact on major hydrological parameters, especially, on the regional scales. A single grid may encompass hundreds of square kilometres, including mountainous and desert terrain, oceans and land areas. Despite recent improvements in modelling of the climate dynamics with complex and large scale models, we are still seriously limited in evaluating regional details of climatic changes or details of the effects of such changes on hydrologic processes and water availability. Loaiciga et al. (1996) have presented a detailed review on the interaction of GCMs and hydrological cycle. Until current GCMs improve both their spatial resolution and their hydrologic parameterization, information on hydrologic effects of global climatic changes can best be studied using regional hydrological models only.

The advantages of using hydrological models for assessing the impacts of climatic change have been discussed by several investigators. Such models are considered suitable for assessing the regional hydrologic consequences due to changes in temperature and precipitation and other climatic variables. The ability of hydrologic models to incorporate projected variations in climatic variables, snowfall and snowmelt algorithms, ground water fluctuations and soil moisture characteristics makes them especially attractive for water resources studies of climatic changes. Moreover, such models can be combined with plausible hypothetical climate change scenarios to generate information on water resources implications of future climatic changes (Gleick, 1986). Various hydrologic models have been used to study the impacts of climate change scenarios, depending on the purpose of study and model availability. Gleick (1987) used a water balance model to estimate the impact of climate on monthly water availability. Detailed studies using a deterministic model in mountain basins (National Weather Service River Forecasting System (NWSRFS) Model) are carried out (Lettenmaier and Gan, 1990; Cooley, 1990; Nash and Gleick 1991; and Panagoulia 1991). Rango (1993) used snow melt runoff model (SRM) for Rio Grande and Kings river basins to study the changes in snowmelt runoff under warmer climate scenarios. Recently Rango and Martinec (1994) examined the influence of changes in temperature and precipitation on the snow cover using SRM and their results are discussed below.

It is worth mentioning, for water resources systems dominated by snow and glacier melt runoff, vulnerability to changes in global climatic conditions can be understood better using the conceptual hydrological models which have algorithms to develop and deplete snowpack using meteorological data. This is especially true with respect to changes in snowfall and snowmelt, because climate changes will also affect the magnitude and distribution of the snowfall occurring during the preceding winters. The models with only snowmelt runoff simulation approach, but without the ability to accumulate the snowpack, may not be suitable to assess the effect on both snow water equivalent and snowmelt runoff. For example, Yeh et al. (1983) found that suddenly removing the snow cover on 15 March would bring about a significant reduction of zonal mean soil moisture for the following spring and summer seasons. They did not, however, model the effects of changes in climate on the development of the snowpack. Similarly, Rango (1989, 1993) modelled the changes in snowmelt runoff caused by temperature increase during the snowmelt period without considering the effect of climate change on snow water equivalent over the basin. Their results indicated that the warmer temperature produced an earlier hydrograph peak, but essentially the same seasonal volume since they started with the same snowpack. Rango and Martinec (1994) reported that temperatures changes +2 °C and +4 °C both had a much more important effect on snow cover than doubling the precipitation occurring during the snowmelt period. Cayan and Riddle (1993) examined the influence of climate parameters on seasonal streamflow in watersheds over a range of elevations and found that temperature sensitivity of seasonal streamflow is greater in Spring and early summer. It was reported that temperature effect upon runoff in early summer is partially counteracted by the opposite effect in earlier spring, but perhaps not totally.

Climate study for the Himalayan basins

The vulnerability of the Indian sub-continent to the impact of changing climate is of vital importance because major impact of climate change in this continent would be on the hydrology, affecting water resources and agricultural economy. The major river systems of the Indian sub-continent namely Brahmaputra, Ganga and Indus which originate in the Himalayas are expected to be more vulnerable to climate change because of substantial contribution from snow and glaciers into these river systems. Recently, Singh and Kumar (1997) studied the effect of climate change on snow water equivalent, snowmelt runoff, glacier melt runoff and total streamflow and their distribution was examined for the Spiti river. The basic objective of this study is to assess the impacts of various climate scenarios on the hydrological response of the high altitude Spiti river in the Himalayas. The climatic scenarios were constructed on the basis of simulations of the Hamburg coupled atmosphere-ocean climate model for the study region. The influence of these scenarios on the snow water equivalent, snowmelt runoff, glacier melt runoff, total streamflow and their distribution have been studied. The adopted changes in temperature and precipitation ranged from 1 to 3EC and -10 to +10 %, respectively. The University of British Columbia (UBC) Watershed Model was used to simulate the hydrological response of the basin under changed climatic scenarios. The following conclusions are drawn from this study:

1. Snow water equivalent over the study basin reduces with an increase in air temperature (T+1, T+2, T+3 °C; P+0%). However, no significant reduction in annual SWE is observed for these projected increases in air temperature over the basin. It seems the high altitude and low temperature regime of the basin limited the reduction in SWE. An increase of 2 °C in air temperature reduced annual SWE in the range of 1 to 7% (T+2 °C, P+0%). The changes in SWE are found proportional to the changes in precipitation. Maximum reduction in annual SWE (13-18%) is found under a T+3 °C, P-10% scenario.

2. Under warmer climate scenarios (T+1, T+2, T+3 °C; P+0%), snowmelt runoff, glacier melt runoff and total streamflow indicate early response along with change in their runoff distribution over time. All these hydrological components linearly increase with an increase in temperature from 1 - 3 °C. The most prominent effect of temperature increase has been noticed on glacier melt runoff as compared to snowmelt runoff and total streamflow. An increase of 2 °C in air temperature has increased annual snowmelt runoff, glacier melt runoff and total streamflow ranging from 4-18%, 33-38% and 6-12%, respectively. Maximum increase in snowmelt runoff (19-41%), glacier melt runoff (53-64%) and total streamflow (16-24%) are observed corresponding to (T+3 °C, P+10%), (T+3EC, P-10%), (T+3 °C, P+10%) scenarios, respectively.

3. The snowmelt runoff and total streamflow increases linearly with changes in precipitation, but glacier melt runoff is inversely related to changes in precipitation (P-10 to P+10%) for different temperature scenarios (T+1, T+2, T+3 °C). It is found that snowmelt runoff is more sensitive to changes in precipitation than glacier melt runoff. A general long-term effect of temperature and precipitation changes on the glaciers can be addressed on the basis of present results. An increase in air temperature or decrease in precipitation for a longer time will reduce the size of

glaciers due to higher melt runoff from them. They may retreat because of their faster depletion under warmer climate. The study suggests that a combination of increase in temperature and decrease in precipitation will provide a compound effect in reducing their size. However, under higher precipitation scenarios, the glaciers might grow in size and result in their advancement.

4. The seasonal analysis of total streamflow indicates that increase in temperature, (T+1, T+2,T+3 °C; P+0%), can produce a large increase in the pre-monsoon season streamflow followed by an increase in the monsoon season. Post-monsoon and winter streamflow are not affected significantly by increase in temperature. No significant changes in the winter streamflow are found for this river.

5. Similar studies should be carried out on basins located in different geographic and climatic regions to investigate potential impacts of projected climate warming on hydrology and water resources in India.

The water resources of mountains are important in climate change studies. Information on the advancement or retreat of mountainous glaciers is one of the basic indicators of climate change. Observing the natural forest boundaries and the distribution of ecotones, which can be driven by water-related factors, also plays a valuable role in climate change impact studies. In the assessment of possible impacts of climatic variations on the water flow and on water resources, a model has to be employed which takes account of these spatial variabilities. Ribstein et al. (1995) analysed a research basin in the Cordillera Real of Bolivia and found that there was a recent and considerable retreat of the glacier because precipitation does not compensate for loss from melting. It is expected that high runoff may be associated with warm, dry periods during which much water is derived by melting of glaciers. Kotlyakov and Lebedeva (1996) presented the information on the possible changes in glacierized areas as a result of global warming. The results are presented in Table 8. It appears that precipitation in the Hindu Kush will increase by 15% in the northern part and by 10% in the southern part, while in the Himalayas its amount will even decline by five per cent. The average summer air temperature will change similarly. It will drop by 0.5 °C in the Hindu Kush, Hindu Raj, Nanga Parbat, and even in the Western Himalayas, but will remain unchanged in the Central Himalayas, whereas, in the High and Eastern Himalayas, it will be 0.5 °C higher than at present. As a result, in the northwestern part of the region, the equilibrium line will decrease by 50-200 m, while in the southeastern part it will rise approximately by just as much. The result will be a changing pattern of glacierisation: glacier covered areas in the Hindu Kush, Hindu Raj, and Nanga Parbat massifs will expand; in the Himalayas, especially in their eastern part, glaciers will begin to shrink.

Table 8: Changing glacierization conditions early in 21st century (Kotlyakov and Lebedeva, 1996).

Mountain Country	Change by early 21st century X,: t(EC) %,: summer : winter	Equilibrium line altitude, M ELA a : b	Ablation A, mm/yr a : b	ELA m	A mm/yr
Hindu Kush	10 - 15 : -0.5 : -0.5	3500 - 5200 : 3280 - 5050	8000 - 1600 : 9200 – 1750	- 220 -150	1200 150
Hindu Raj	10 :-0.5 : 0.0	3800 - 4800 : 3580 - 4600	6450 - 2600 : 7100 – 2900	-220 -200	650 300
Nanga Parvat	5 : -0.5 : 0.0	4200 - 5200 : 4000 - 5100	7400 - 3200 : 7800 – 3340	-200 -100	400 140
Western Himalayas	5 : -0.5 : 0.5	4500 - 5000 : 4450 - 4950	5100 - 3400 : 4850 – 3230	-50 -50	-250 -170
Central Himalayas	-5 : 0.0 : 1.0	5000 - 5500 : 5070 - 5550	3100 - 1700 : 2950 – 1600	70 50	-150 -100
High Himalayas	-5 : 0.5 :1.0	4500 - 5800 : 4720 - 5920	4400 - 750 : 4180 - 650	220 120	-220 -100
Eastern Himalayas	-5 : 0.5 : 0.75	4500 - 5200 : 4720 - 5350	4400 - 2200 : 4180 - 1900	220 150	-220 -300

a- current value b- anticipated value early in the 21st century

8.0 ASSESSMENT OF WATER RESOURCES

Because the availability of data in the mountainous areas is poor due to various reasons as described above, accurate assessment of water resources is generally, not possible. An additional problem in the assessment of mountain water resources is connected with the boundaries between countries. Sometimes data from neighbouring countries might be incompatible, as they might have been collected with different instruments and methodologies. Assessing and understanding mountain water resources on the global scale are imperative for sustainable development and managing fragile and vulnerable mountain ecosystems. Remote sensing offers a considerable potential for studying the hydrology of mountainous areas. Because of difficult access and the expensive operation of hydrological stations, radar or satellite data are particularly appropriate but of course ground truth data are indispensable in the calibration and verification of remotely sensed data. For research and operational needs, a systematic data collection network for measuring precipitation in mountainous areas is urgently required. Improvement of hydrological monitoring of streams and rivers within mountains is necessary, in addition to the monitoring of same river in the low land, which is done with far greater accuracy. It is of paramount importance to try to launch a GIS supported assessment of fresh water resources in mountain areas and to estimate the contribution of these areas to global water resources.

9.0 SEDIMENT AND EROSION

Himalayan rivers transport the maximum sediment load in the world, resulting in problems such as the siltation of reservoirs, blockage of river channels, quality of water supplies, transport of chemical pollutants etc. The major sources of sediment are considered to be glacial debris, landslides, and intensively cultivated hill slopes, but little is known about the characteristics of the sediments from these sources, the movements into the river system, or the transport down the rivers. The development of a methodology of classifying sediments in the Himalayan rivers, including the full range of sources and methods of transportation will provide an essential information source for a future project which will aid to quantify the sediment loads of the rivers. Standardised methods of surveying sediments from first order rivers to the Piedmont zone are being developed by field sampling and statistical analysis. In addition, a GIS frameworks is being developed from two of the study river basins, the Upper Ganges, in north eastern India, and the Trisuli, in Central Nepal. The establishment of this framework will be fundamental in the quantification of both sediment supply from the major sources and the transportable load within the river channel (Johnson and Collins,1996).

The sediment loads of Asian rivers are reported to be the highest in the world, delivering an estimated 80%t of the total sediment input to the oceans (Holeman, 1968). These large sediment loads are due to the exposure of the geologically young rocks forming the Himalayan mountain chain, which has the world's greatest range of relief and extremes of climate. This results in high erosion rates, estimated to be 1.7 mm/yr for the Central Himalayas (Valdiya and Bartarya 1989) and rapid rates of transport down the rivers. The loads are sometimes said to be the most significant water quality problem in this region of Asia. Problems range from the rapid infilling of reservoirs, reduced quality of potable water, siltation of river channels, transport of chemical pollutants, and the loss of aquatic habitats.

Some data on sediment yields from Himalayan rivers exist, but they are extremely sparse considering the scale of the sediment problem. In addition, few details exist on the type of sediment in the system and the processes of release and transport down the river. For water resources planning, the size and shape of the particles, are important in determining the rate of down river transport, the impact on the quality of potable water supplies, and the likely distribution of material deposited in reservoirs or in approaches to barrages and bridges. For the transport of chemical pollutants, the main problem is in the erosion of soil from agricultural areas where nutrient wash out is a problem. As far as aquatic ecosystems are concerned, the sediments create many of the habitats for breeding and feeding, but changes in the sediment characteristics cause impacts such as the loss of habitat from increased mobility or the infilling of interstices by fine particles.

The description and quantification of sediment loads in Himalayan rivers presents a major challenge to the fluvial geomorphologist. Techniques exist for measuring erosion rates on plot scales and for measuring the accumulation of sediment in lakes or reservoirs, but there is no widely accepted method of quantifying the total sediment load being transported within a river system. A new method is currently being developed by the authors to quantify the sediment loads of Himalayan rivers using a classification of the sediments within the channel and relevant spatial data sets for whole river basins.

9.1 Cloudburst and erosion

A large extent of Himalayas is most vulnerable to hazard due to intense thunderstorms or cloudbursts; high rainfall, structurally crushed; folded, and faulted rock strata; and highly erodible poor soil subjected to faulty land use practices. In some areas, intensive grazing over the years alongwith removal and degradation of forest products, has resulted in rudimentary vegetation cover not capable of holding the soil. Therefore, land use which allows greater water retention in the fragile catchments of Himalayan mountains should be encouraged for soil and water conservation in this region. In the Himalayan mountains, as a consequence of loss of forest cover coupled with the influence of the monsoon pattern of rainfall, the fragile catchments have become prone to low water retention and high soil loss associated with runoff (Valdiya 1985, Rawat and Rawat 1994, Joshi and Negi, 1995). Therefore, understanding the hydrometeorology of the Himalayan mountains needs to be developed before any mitigating plans can be formulated.

The highly rugged Himalayan terrain often faces extreme hydro-meteorological conditions resulting in flash floods. The temporal and the spatial variations in such extreme rainfall constitute the chief determinant factors of the floods. A cloudburst may last for a few minutes or a few hours. The impact of such events is more severe if the area involved is a small catchment characterised by steep hill slopes and high riverbed gradients. The most important adverse effect of a cloudburst is the triggering of large scale mass movements, which introduce enormous amounts of sediment into the drainage system (Carson, 1985). The consequences of large scale erosion due to a cloudburst in small catchments is two fold - i) the excessive sediment load may cause aggradation conditions of the riverbed further downstream, thereby increasing the water level in general and flood hazards in particular (Pal and Bagchi, 1975), and ii) the debris, including big boulders resulting from the sudden and large scale erosion, may temporarily dam the river. Subsequent breaching of the dam may cause a surge of water, leading to excessive mass movements along its course and causing a devastating flood and widespread damage to life and property. Anbalagan (1996) has discussed the disastrous events triggered by a cloudburst in July 1983 in Karmi village of Almora Kumaun.

9.2 Glacier-dammed lake outburst floods (GLOF)

Glacial lakes are found in the high elevation portion of many river basins throughout the Nepal Himalayas as well as other glacierized areas of he world (Ives 1986). They are formed when water is impounded by glacial ice and/or moraines. There is a wide variety of these lakes, ranging from water ponds on the surface of glaciers to large lakes in side valleys dammed by a glacier in the main valley. With the general retreat of glaciers in this century, a topographic depression is often formed by the moraine at the maximum extent of the glacier. If this enclosure is watertight, melt waters will accumulate in the basin until seepage or overflow limits the lake level. Such moraine-dammed lakes appear to be the most common type of glacial lakes now found in Nepal (Yamada and Sharma, 1993).

In some cases, the impoundment of the lake may be unstable, leading to a sudden release of large quantities of stored water. Failure of these ice or moraine dams as very destructive events has been documented throughout the world (Ives, 1986; Vuichard and Zimmermann, 1987). Outburst floods can be several times greater than floods produced by even extreme rainfall (Dhital et al., 1993). Perhaps the most common trigger of an outburst flood involves surge waves generated by snow, ice, or rock avalanches into the lake that then overtop the dam and rapidly erode the outflow channel, ultimately collapsing the moraine in the vicinity of the channel. Other failure mechanisms include slow melting of an ice core within the moraine, seepage and piping through the moraine, progressive thinning of the moraine by landslides, and earthquakes.

As glacial lake outbursts have threatened more people and property in the Himalayas, they have received increasing attention in the past two decades. Damage to bridges, roads, trails, and hydroelectric facilities has generated much of the current interest in these floods. Historical analyses have documented more than a dozen outburst events in Nepal between 1964 and 1991 (Yamada and Sharma, 1993). Modelling of glacier outbursts from the ice-dammed lakes and from moraine lakes is not done.

10.0 HYDROCHEMISTRY

The chemical composition of snowfall and snow on the ground attribute to the chemical constituents of melt runoff from the basin. Still there is much to be explored in the chemistry of both falling snow and snow on ground. Depending upon the location of study area, a big variation is found in the concentration of chemical constituents in the snow. The major difference in the chemical composition of fresh snow and old snow at the ground surface results due to additional dry fall out from the atmosphere on the ground, which changes the constituents of the snow surface lying on the ground surface. In order to improve the understanding of chemistry of snow, extensive world-wide observational network are to be established.

The sustainable use of land, water, and agricultural resources is an essential component in the future development of Asian, especially those regions bordering the Himalayas, where population growth is rapid and predicted to continue. This population pressure results in an intensification of agriculture through the adoption of multiple annual crop rotations supported by large applications of inorganic and organic fertilizers and the expansion of cultivated land on to steeper hill slopes and previously forested areas continued and excessive application of mineral and organic fertilizers to increase crop yield may, however, lead to detrimental effects in the form of soil and water acidification as well as downstream eutrophication (Schreier et al., 1994). The problem may be exacerbated as a consequence of deposition of acidic oxides from fossil fuel burning and, in this respect, the rapid industrialisation of many Asian countries is predicted to increase pollutant emissions in the future (Arndt and Carmichael, 1995).

Population growth in the Himalayan region has led to an increased demand for food, which has been met by increased use of fertilizers and expansion of agricultural land. These changes modify the quality and quantity of river flows downstream from the affected areas and therefore have a regional as well as a local impact. In the Himalayas, nitrate and sulphate concentrations of similar magnitude to those observed in NW Europe have been reported in snow samples (Mayewski et al.,1986). The deposition of acidic oxides is predicted to increase in the future in Asia in line with an expected increase in emissions as industrial development continues (Galloway 1989). Jenkins (1996) has presented the data from the first wide-scale surveys of the chemistry of medium to high altitude first-order streams, draining relatively undisturbed and apparently unpolluted catchments, in the Himalayas of Nepal. In general, the water from these catchments have high background concentrations of sulphate, calcium, and bicarbonate, all derived from bedrock weathering sources. High p^H indicates that a large buffering capacity exists within the system and that further additions of anthropogenic sulphur oxides would be readily buffered. Many streams, however, already demonstrate significant concentrations of nitrate. This is probably the result of low biological requirements, given the poor vegetation and cold temperatures at high altitude. Clearly, a future increase in atmospheric nitrogen deposition might exacerbate the situation and promote yet higher concentrations, whilst this is unlikely to cause acidification of these upland-most streams, the biological implications of increased nitrate concentrations are not yet known and downstream eutrophication problems may occur. The data are used to determine the major controls on surface water chemistry in this largely unpolluted region and to determine the areas most susceptible to acidification from future increased deposition of acidic oxides.

Future surveys of water chemistry are needed in Himalayan region to provide the necessary database to assess the impact of land use on stream chemistry. Such investigations require good quality hydrological data on the appropriate spatial scale and temporal resolution.

11.0 CONCLUSIONS

This report discusses about the status and progress of studies of various hydrological problems related to mountain area. The principal issues related to mountain hydrology on the local and regional scales are as follows.

1. Variation in hydrological parameters, especially hydrological variables like precipitation and temperature on the basin and regional scale are to be studied. GIS supported modelling of areal rainfall in the mountainous region is required.

2. Network design, instrumentation, data collection and processing methodologies are to be improved. There is an urgent need for development and improvement in measurement techniques and development of automated instruments for the harsh and extreme conditions in the maintains

3. For a better understanding of different hydrological variables, a long-term operation of a few, carefully selected, multidisciplinary experimental research basins is required in different regions of mountains areas.

4. Impact of climate change on different hydrological parameters is to be studied for the snow and glacier fed rivers.

5. Variability in snow covered area and its relationship with temperature is to be studied for snow covered mountainous basins. Extent of glacierized area in each mountain basin must be determined. There is need to make use of such information for the management of water resources in the mountainous regions.

6. Modelling of snow and glacier melt runoff for the mountainous basins is to be carried out. Snow and glacier contribution in annual flows of Himalayan rivers is to be estimated. Modelling of rainfall-runoff processes in the rain-fed mountainous basins is also equally important and need attention.

7. Application of remote sensing techniques are needed for the modelling of snow melt runoff for the mountain basins to determine extent of snow and glacier covered areas.

8. The cloudburst is one of the major natural hazards in the Himalayas, occurring amid extreme hydro-meteorological conditions. Dynamics and hazards of erosion and sedimentation, ecosystems of high mountain areas, and landscape processes is to be studied.

9. Modelling the interrelations between climate, hydrology, and ecosystem as well as the effect of climate change and other impacts on natural resources on different spatial and temporal scales.

10. Lake studies in the high altitude regions of the Himalayas

11. The heterogeneity of the mountain region provides an opportunity for comparing patterns and processes ranging from those in the high mountains to those in the flat alluvial plains. It is necessary to promote cooperation at both regional and global levels by making a state of the art inventory of existing knowledge, information, and data and making it accessible to researchers willing to undertake scientific investigations and studies.

12. It is recommended that the temperatures recorded at the same time for a set of stations situated at different elevations may help in higher accuracy of temperature lapse rate. The temperature lapse rate values may be determined for each Himalayan basin which need strengthening of existing network in the Himalayas. A micro-scale study is required to study the local and regional factors influencing the temperature variation at the high altitude basins.

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