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Hydrological Modelling of Streamflow in the Glacierized catchment of Central Himalaya (Uttarakhand, India)

A

Dissertation Report

**Submitted for partial fulfillment of the requirement for the award of
the Degree of**

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In

Physics

By

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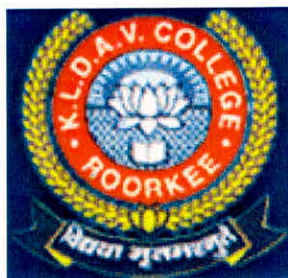
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Roorkee, Uttarakhand – 247 667, INDIA

July, 2017

Dedicated to
MY INSPIRATION...
MY BELOVED MOTHER

DECLARATION

I hereby declare that the submission entitled- '**Hydrological Modeling of Streamflow in the Glacierized catchment of Central Himalaya (Uttarakhand, India)**' for the award of the degree of **Master of Science (M.Sc.) in Physics** at Kanahiya Lal D.A.V. (P.G.) College, Roorkee (affiliated to HNB Garhwal University, Uttarakhand) is my own work which has not been submitted for the award of any degree or diploma of the university or other institute of higher learning, except where due acknowledgement has been made in the text. This work is carried out under joint supervision of Dr. Manohar Arora (Scientist – D, National Institute of Hydrology, Roorkee, Uttarakhand), Prof. Ajay Gairola and Dr. Kapil Kesarwani (Department of Civil Engineering, Indian Institute Technology of Roorkee, Roorkee, Uttarakhand) during February 20, 2017 to July 24, 2017.

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DISSERTATION COMPLETION CERTIFICATE

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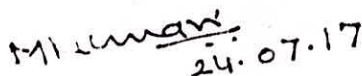
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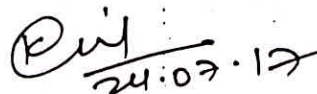
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(Darshan Hans)

ABOUT THE INSTITUTES / ORGANIZATION

1. National Institute of Hydrology, Roorkee (Uttarakhand)



The National Institute of Hydrology (NIH), Roorkee (Uttarakhand) is a premier organization involved in Research and Development. The Institute has completed a number of consultancy and sponsored research projects to solve real life field problems. The Institute has developed software for hydrological analysis and design. The Institute has established Water Quality, Remote Sensing & GIS, Nuclear Hydrology, Soil & Water Laboratories and Groundwater Modeling Unit with state-of-art equipment and software. Under technology transfer programmed, the Institute has organized a number of training courses all over the country on different topics relevant to hydrology and water resources. The Institute has also organized number of international, National and regional activities. Through its strategic plan, NIH has acquired a leadership role in the area of hydrology. The basic and result-oriented research and development programs are integrated and executed using very innovative, dynamic and focused team of scientists. The future prospects of NIH are tremendous. We are establishing collaborations with National and International Organizations and Institutes. NIH is growing in an era of unparalleled scientific and technological advancement.

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The Department of Civil Engineering at the Indian Institute of Technology, Roorkee is the oldest and the largest in the country. The department has produced several eminent engineers who have made significant contributions in the planning and execution of Civil Engineering projects in India as well as abroad.

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List of Abbreviations

Avg	Average
D	Degree-day factor
EVU	Elevation Vegetation Unit
GCA	Glacier covered area
HBV	Hydrologiska-Bryans-Vattenbalansavdelning
IITR	Indian Institute of Technology Roorkee
Max	Maximum
Min	Minimum
NIH	National Institute of Hydrology Roorkee
SCA	Snow covered area
SFA	Snow free area
SMI	Soil moisture index,
SNOWMOD	Snow Model
SRM	Snowmelt runoff model

List of Symbols

Symbols	Meaning	Unit	Value
AE	Actual evaporation	mm	
AET	Actual evapotranspiration	mm	
AG	Calibration parameter	(mm ⁻¹)	
Alpha	Non linearity coefficient		[0,inf)
BETA	parameter that determines the relative contribution to runoff from rain or snowmelt		(0,inf)
C _{ET}	Potential evaporation correction factor	1/°C	[0,1]
CFGlacier	Glacier correction factor	-	(0,inf)
CFMAX	Degree-day factor	mm/Δt °C	[0,inf)
CFR	Refreezing coefficient		[0,inf)
CFSlope	Slope correction factor	-	(0,inf)
CWH	Water holding capacity		[0,inf)
D	Degree-day factor	mm °C ⁻¹ d ⁻¹	
DELAY	Time period over which recharge is evenly distributed	Δt	[0,inf)
dKG	Maximum minus minimum outflow coefficient	t ⁻¹	
Epot	potential evaporation at day t	mm d ⁻¹	
FC	maximum soil moisture storage Soil		(0, inf)
h _j	The zonal hypsometric mean elevation	m	
h _{base}	Elevation of the base station (m),		
KGmin	Minimum outflow	t ⁻¹	
KSI	Snow to Ice conversion factor	t ⁻¹	
K0	Recession coefficient	Δt ⁻¹	[0,1)
K1	Recession coefficient	Δt ⁻¹	[0,1)
K2	Recession coefficient	Δt ⁻¹	[0,1)
LP	soil moisture value above which AET reaches PET		[0,1]
M _{g,i,j}	Glacier melt in terms of depth of water	mm	[0,inf]

P	Total observed precipitation	mm	
P _{area}	areal average precipitation	mm Δt^{-1}	
PART	portion of the recharge which is added to groundwater box 1		[0,1]
PCALT	change of precipitation with elevation	%/100m	(-inf,inf)
Pelev	elevation of precipitation data in the PTQ file	m	(-inf,inf)
PERC	threshold parameter	mm Δt^{-1}	[0,inf)
Q	Outflow/ Discharge	mm d ⁻¹	
Q _{obs}	Observed discharge	mm d ⁻¹	
Q _{sim}	Simulated discharge	mm d ⁻¹	
SFCF	Snowfall correction factor		[0,inf)
SLZ	Storage in lower zone	mm	[0,inf)
SM	Water content of soil box	mm	[0,inf)
SWE	Water equivalent of the snowpack on top of the glacier		
SUZ	Storage in upper zone	mm	[0,inf)
T _b	Base temperature	°C	
T _c	Critical temperature	°C	
TCALT	change of temperature with elevation	°C/100m	(-inf,inf)
Telev	elevation of temperature data in the PTQ file	m	(-inf,inf)
T _i	Index air temperature	°C	
T _{max}	Maximum temperature	°C	
T _{mean}	Mean temperature	°C	
T _{min}	Minimum temperature	°C	
TT	Threshold temperature	°C	
UZL	threshold parameter	mm	[0,inf)
W	Change in glacier water content	mm	(-inf,inf)

Chapter - 1

Introduction

A glacier is a persistent body of dense ice that is constantly moving under its own weight. It forms where the accumulation of snow exceeds its ablation (melting and sublimation) over many years, often centuries (Cuffy and Paterson, 2010). Glaciers slowly deform and flow due to stresses induced by their weight, creating crevasses, cirques, and other distinguishing features. The movement may be as little as few centimeters a day to much larger or faster.

1.1 Glacier Formation and Structure

1.1.1 Formation of Glacier

Glaciers begin to form when snow remains in the same area year-round, where enough snow accumulates to transform into ice (Fig.1.1). Each year, new layers of snow bury and compress the previous layers. This compression forces the snow to re-crystallize, forming grains similar in size and shape to grains of sugar. Gradually the grains grow larger and the air pockets between the grains get smaller, causing the snow to slowly compact and increase in density. After about two winters, the snow turns into firn - an intermediate state between snow and glacier ice. At this point, it is about two-thirds as dense as water. Over time, larger ice crystals become so compressed that any air pockets between them are very tiny. In very old glacier ice, crystals can reach several inches in

length. For most glaciers, this process takes more than a hundred years (Cuffy and Paterson, 2010).

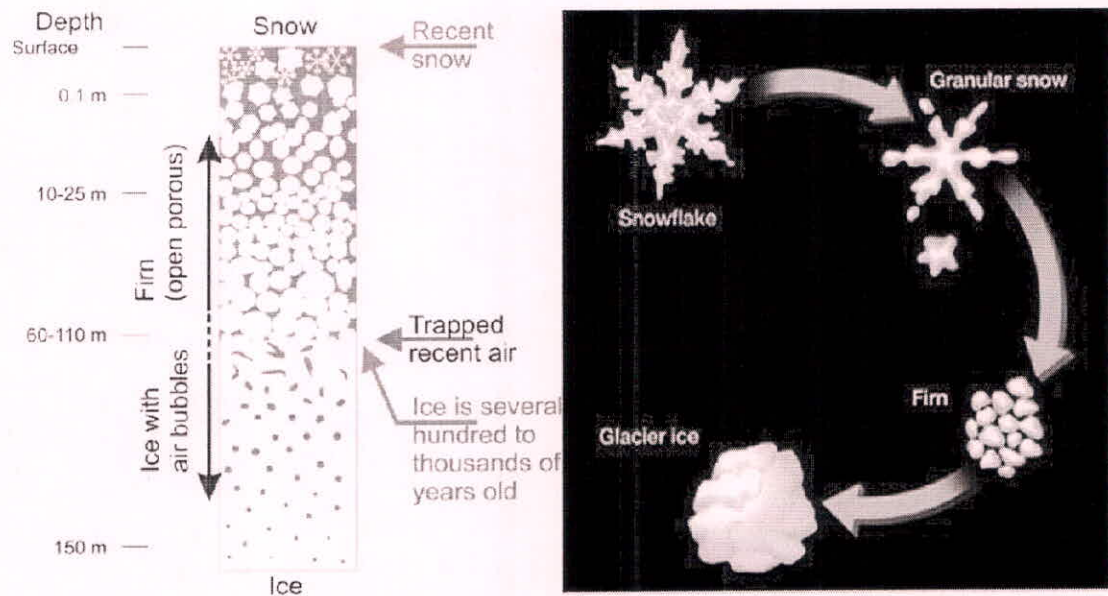


Fig. 1.1: Formation of a glacier.

1.1.2 Structure of Glacier

A glacier originates at a location called its glacier head and terminates at its glacier foot, snout or terminus (Fig. 1.2). Glaciers are broken into zones based on surface snowpack and melt conditions.

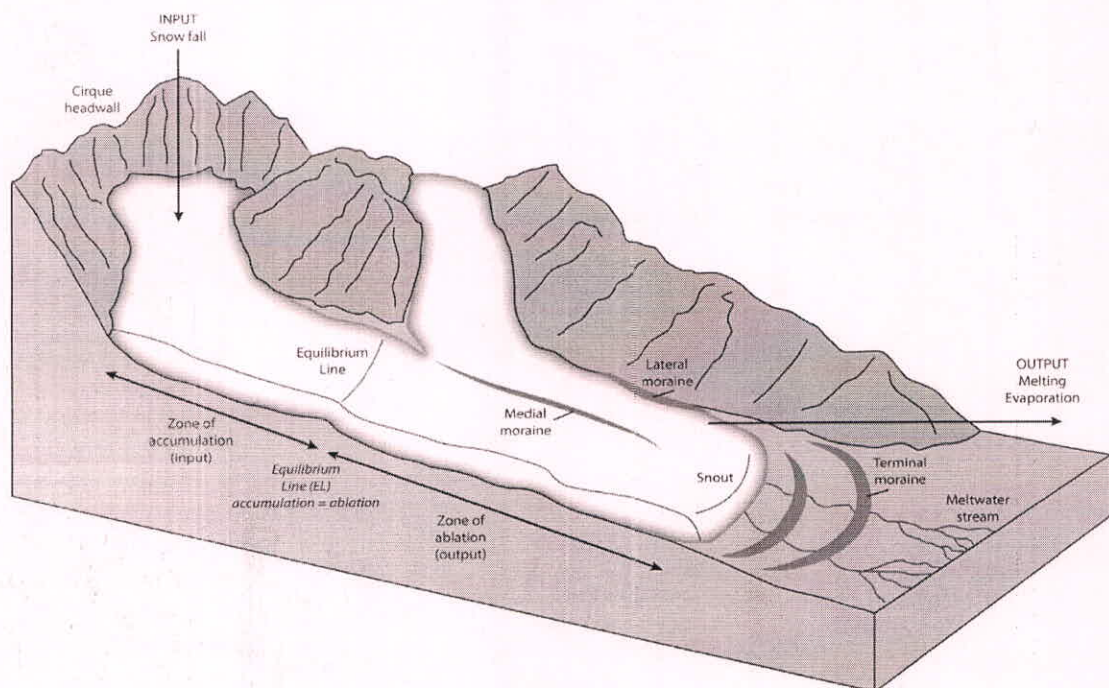


Fig. 1.2: Typical Structure of a Glacier.

The ablation zone is the region where there is a net loss in glacier mass. The equilibrium line separates the ablation zone and the accumulation zone; it is the altitude where the amount of new snow gained by accumulation is equal to the amount of ice lost through ablation. The upper part of a glacier, where accumulation exceeds ablation, is called the accumulation zone. In general, the accumulation zone accounts for 60–70 % of the glacier's surface area.

1.2 Classification of Glaciers

1.2.1 Classification based on temperature

Cold Glacier

It is the terminology used for the glaciers that are presumably cold enough, so that the ice forming the glacier exhibits a temperature gradient that is well below the pressure-melting point (Fig. 1.3a,b). Glacier ice, in the case of such glaciers, is supposedly bound to bedrock underneath; so all motion within the glacier occurs only by the deformation of the ice (Raina and Srivastava, 2008). As the glacier is believed to have a dry base, its motion is very slow and hence it does not cause the characteristic feature like striations on the underlying rock floor. The term is generally restricted to the glaciers that exists in the Polar Regions.

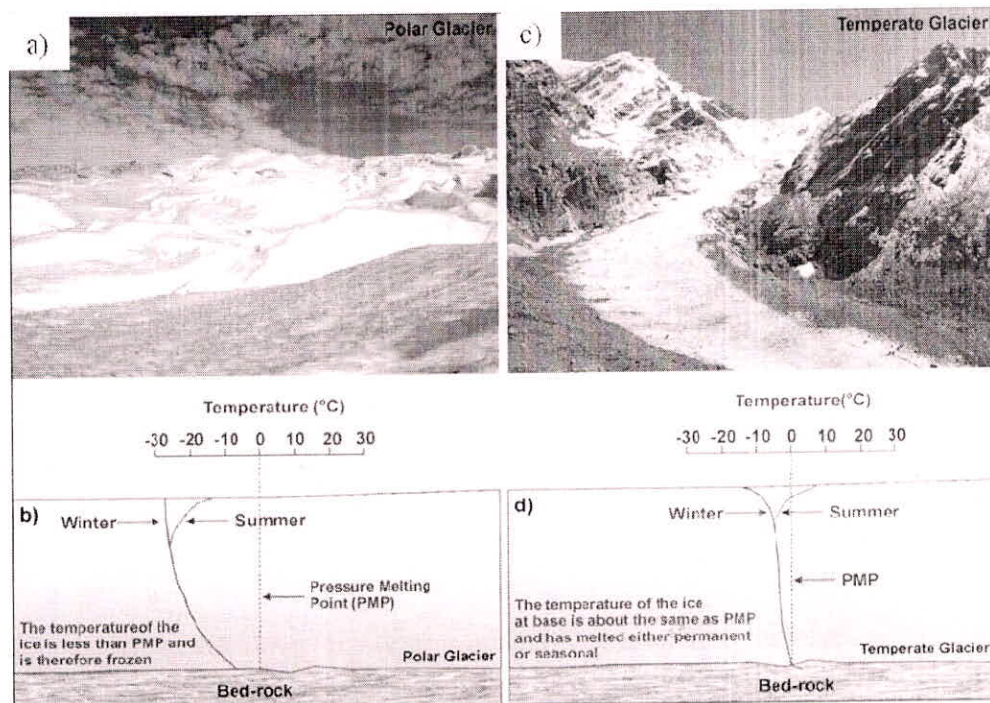


Fig. 1.3: (a) Cold Glacier near the Bharti station (Indian), Larseman Hills, Antarctica, (b) Temperature profile of the Cold Glacier, (c) Temperate Glacier - Dokriani Glacier, Garhwal Himalaya (d) and Temperature profile of the Temperate glacier.(Keserwani, 2015)

Temperate / Warm Glacier

The term used to denote the glacier that occurs in the Himalayas and Alps. Ice forming such glaciers is supposed to be stable at what is termed as the pressure-melting point temperature, and hence small changes can produce melting and slippage (Fig. 1.3c,d). Glaciers of this group, are supposed to have a wet base and motion is much more rapid (Cuffy and Paterson, 2010). Such glaciers consequently are greater erosion agents than the cold glaciers.

1.2.2 Classification based on the size and nature

Continental Glacier

Continental glaciers are continuous masses of ice that are much larger than alpine glaciers (Raina and Srivastava, 2008). Small continental glaciers are called ice fields. Big continental glaciers are called ice sheets. Greenland and Antarctica are almost entirely covered with ice sheets that are up to 3500 m (11,500 ft) thick. They generally exhibit a piedmont lobe front where they debouch out of the ice cap and where they occupy linear depressions, these glaciers exhibit exceptionally vast dimensions often extending over hundreds of kilometers (Fig. 1.4a).

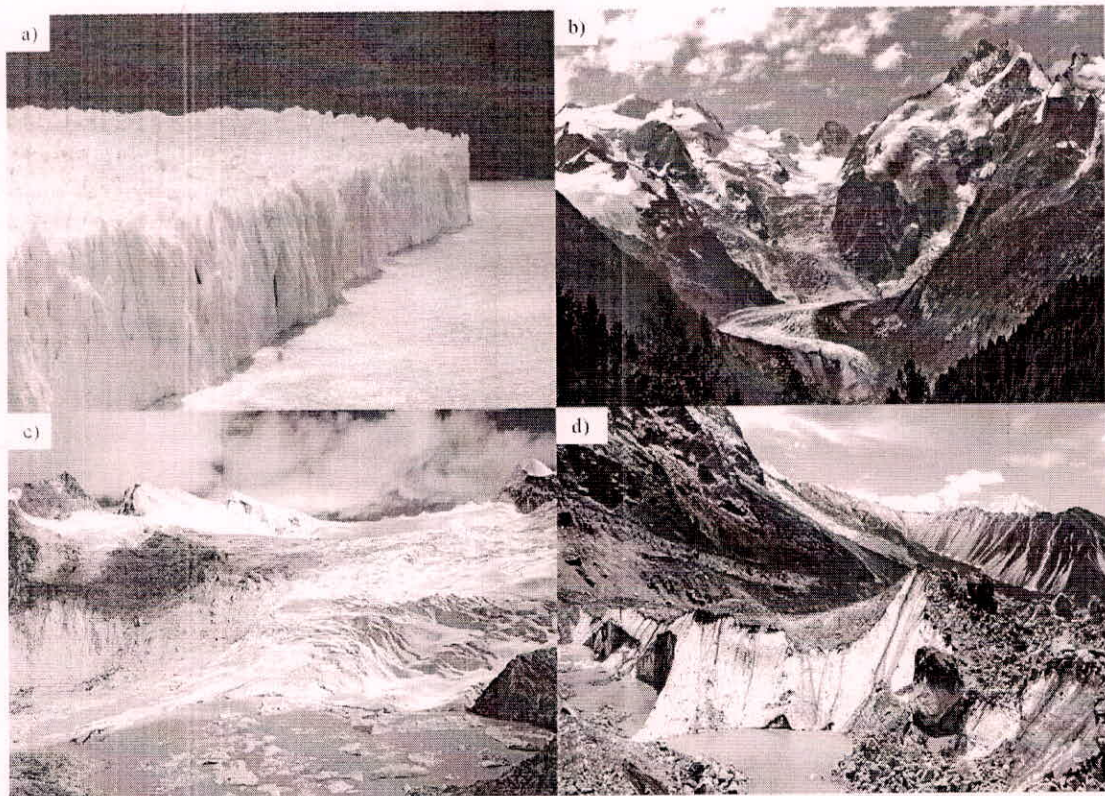


Fig. 1.4: Synoptic view of a) Continental glacier b) Mountain glacier c) Debris free glacier and d) Debris covered glacier.

Mountain/Valley/Alpine Glacier

Glaciers that occur in the mountain regions of the world, like the Himalayas, belong to the Alpine variety (Raina and Srivastava, 2008). Alpine glaciers begin high up in the mountains in bowl-shaped hollows called cirques (Fig. 1.4b). As the glacier grows, the ice slowly flows out of the cirque and into a valley. Several cirque glaciers can join together to form a single valley glacier. When valley glaciers flow out of the mountains, they spread out and join to form a piedmont glacier.

1.2.3 Classification based on the surface characteristics

Clean or Debris free glaciers

The term clean glacier has been used for the glaciers that, for whatever the reason, are, practically devoid of any debris cover over the entire glacier surface including the ablation zone (Fig. 1.4c).

Debris-covered Glacier

Debris cover is a blanket of glacier which protects them from melting if it has a proper thickness. In many mountain ranges, the lowest reaches of big glacier are covered by a continuous layer of sand gravel and rock (Fig. 1.4d). These glaciers are known as debris covered glacier and they are particularly common in mountain of high Asia.

1.3 Glacier Geomorphology

1.3.1 Horn

A pyramidal peak, sometimes in its most extreme form called a glacial horn, is an angular, sharply pointed mountain peak which results from the cirque erosion due to multiple glaciers diverging from a central point (Fig. 1.5a). Horns are created when three or more glaciers act on an arete to form a mountain peak.

1.3.2 Arete

An arête is a narrow ridge of rock which separates two valleys. The word "arête" is actually a French word meaning 'edge' or 'ridge'. It is typically formed when two glaciers erode parallel U-shaped valleys (Fig. 1.5b). Arêtes can also form when two glacial cirques erode headwards towards one another, although frequently this results in a saddle-shaped pass, called a col. The edge is then sharpened by freeze-thaw weathering,

and the slope on either side of the arete steepened through mass wasting events and the erosion of exposed, unstable rock.

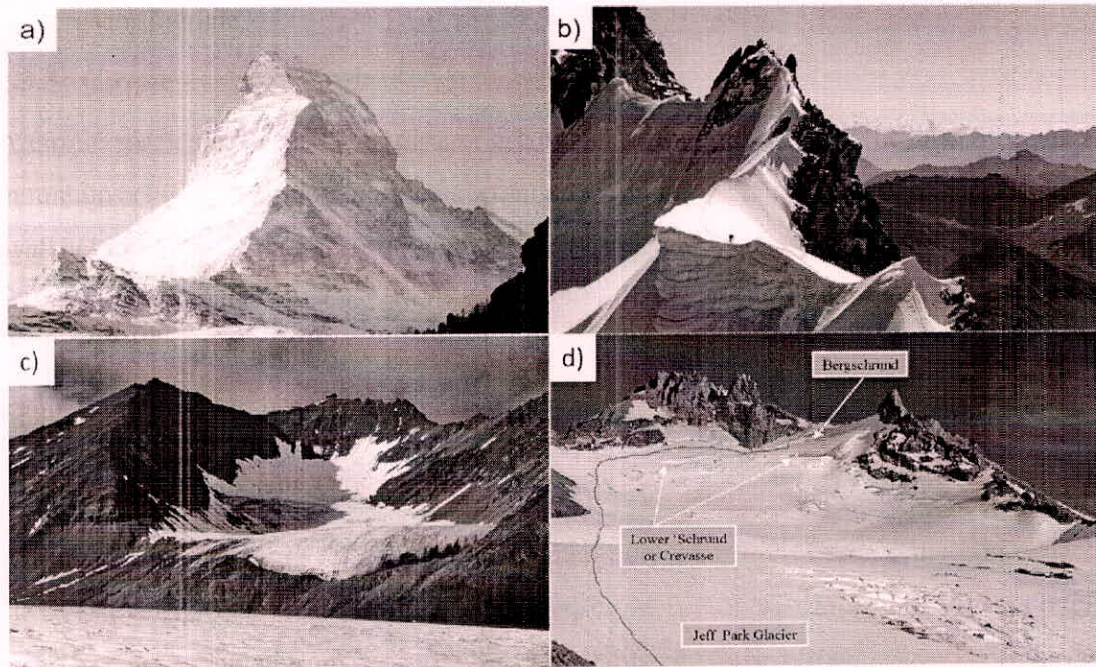


Fig. 1.5: a) Structure of Horn, b) Arete c) Structure of Cirque in a mountain glacier d) Bergschrund line.

1.3.3 Cirque

A cirque glacier is formed in a cirque, a bowl-shaped depression on the side of or near mountains (Fig.1.5c). Snow and ice accumulation in cirque often occurs as the result of avalanching from higher surrounding slopes. If a cirque glacier advances far enough, it may become a valley glacier. Additionally, a valley glacier retreats enough that it is within the cirque, it is a cirque glacier again. In these depressions, snow persists through summer months, and becomes glacier ice.

1.3.4 Bergschrund line (Starting Point)

A Bergschrund (German term meaning mountain crack) is a crevasse that forms where moving glacier ice separate from the stagnant ice or firn above (Fig. 1.5d). It separates the cirque of a glacier from the valley wall at the head of the glacier. Bergschrund line can extend to the bedrock and can have depth of 100 m. It can be filled by snow in winters because of avalanches but in summer, due to melting it remains as open line. It is a German term meaning mountain crack that is used to identify the crevasse that occurs rather separates the cirques of a glacier from the valley wall at the head of the glacier.

1.3.5 Accumulation Zone

The zone over the glacier surface, where the accumulation takes place relative to the previous year's surface, is termed as the accumulation zone (Fig. 1.6). On a glacier, the accumulation zone is the area above the ELA (Equilibrium Line Altitude) line, where snowfall accumulates and exceeds the losses from ablation (melting, evaporation & sublimation). Accumulation is the addition of ice or snow to the glacier surface by snowfall, hail and rain that freeze (Fig. 1.7a).

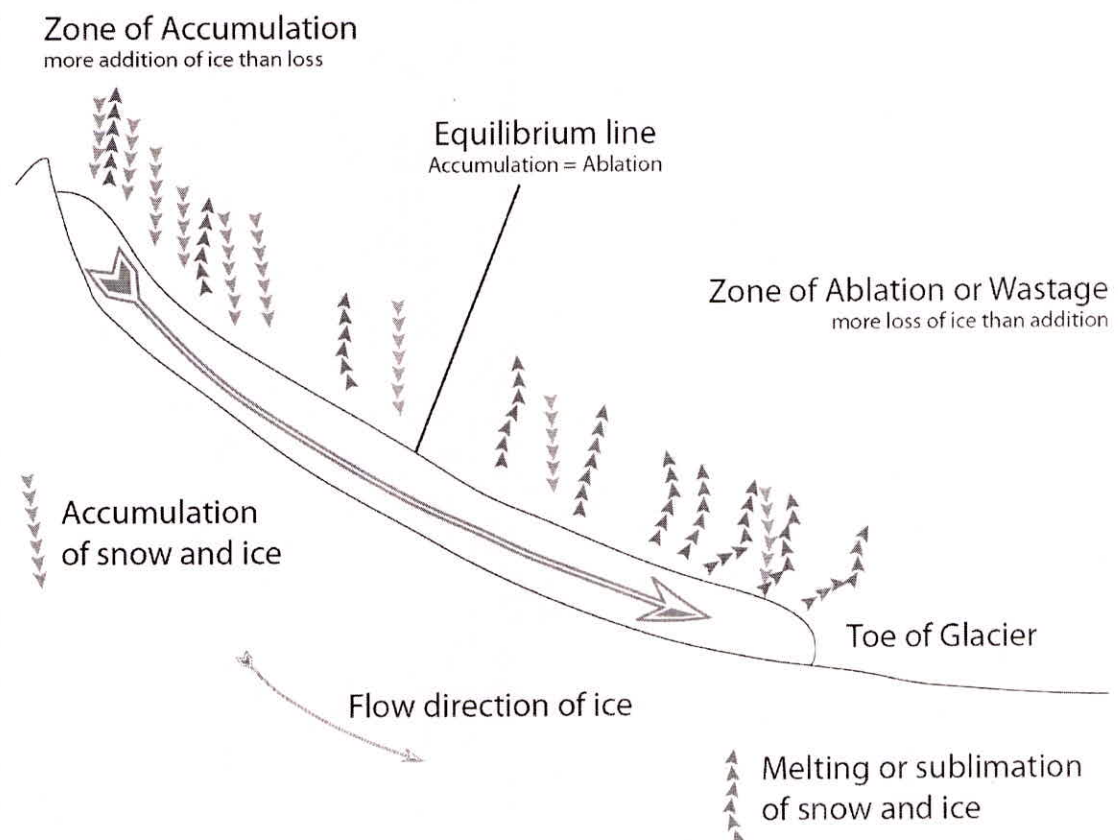


Fig. 1.6: Different zones of a typical valley glacier.

1.3.6 Ablation Zone

The zone, over the glacier surface, within which the loss of ice takes place as compared to the previous year's surface, is called the ablation zone (Fig. 1.6 and Fig. 1.7b). Ablation zone or Ablation area refers to the low altitude area of a glacier or ice sheet below the firn with net loss in ice mass due to melting, sublimation, evaporation, ice calving, aeolian processes like blowing snow avalanche, and any other ablation (Cuffy and Paterson, 2010). The ablation zone often contains meltwater features such as supraglacial lakes, englacial streams, and subglacial lakes (Raina and Srivastava,

2008). Sediments dropped in the ablation zone forming small mounds or hillocks are called kames. Ablation constitutes a key part of the glacier mass balance (Fig. 1.7b)

1.3.7 Snout

A glacier terminus or toe is called a snout where meltwater outflow (Fig. 1.6 and Fig. 1.7c). It is a part of ablation zone. Although, glaciers seem motionless to the observer, in reality glaciers are in endless motion and the glacier terminus is always either advancing or retreating.



Fig. 1.7: Synoptic view of a) Accumulation zone b) Ablation zone and c) Snout of Glacier

The location of the terminus is often directly related to glacier mass balance, which is based on the amount of snowfall which occurs in the accumulation zone of a glacier, as compared to the amount that is melted in the ablation zone.

1.3.8 Equilibrium-Line-Altitude (ELA)

The Equilibrium-Line-Altitude (ELA) or snow line is that line that theoretically separates the ablation zone from the higher-altitude accumulation zone (Fig. 1.6). Basically, the position of line is governed by the mass balance of the glacier. Positive balance brings

down the position of this line on the glacier surface and the negative balance leads to the recession of this line on the glacier surface.

1.3.9 Moraine

Moraine is material transported by a glacier and then deposited (Fig. 1.8). A glacier carries rock debris on its surface, within the body as englacial, and along the base as subglacial. This rock debris, which is obtained from the valley walls and or the bed rock due to frost action and creep in the case of the former and by scrapping and scooping in the case of latter, is termed as moraines. Moraines are formed from debris previously carried along by a glacier and normally consist of somewhat rounded particles ranging in size from large boulders to minute glacial flour.

Lateral Moraine

Lateral moraines are parallel ridges of debris deposited along the each side of a valley glacier (Fig. 1.8a). It is a ridge of glacial load by the side of a glacier or lying along the side of a valley which was formerly occupied by a glacier.

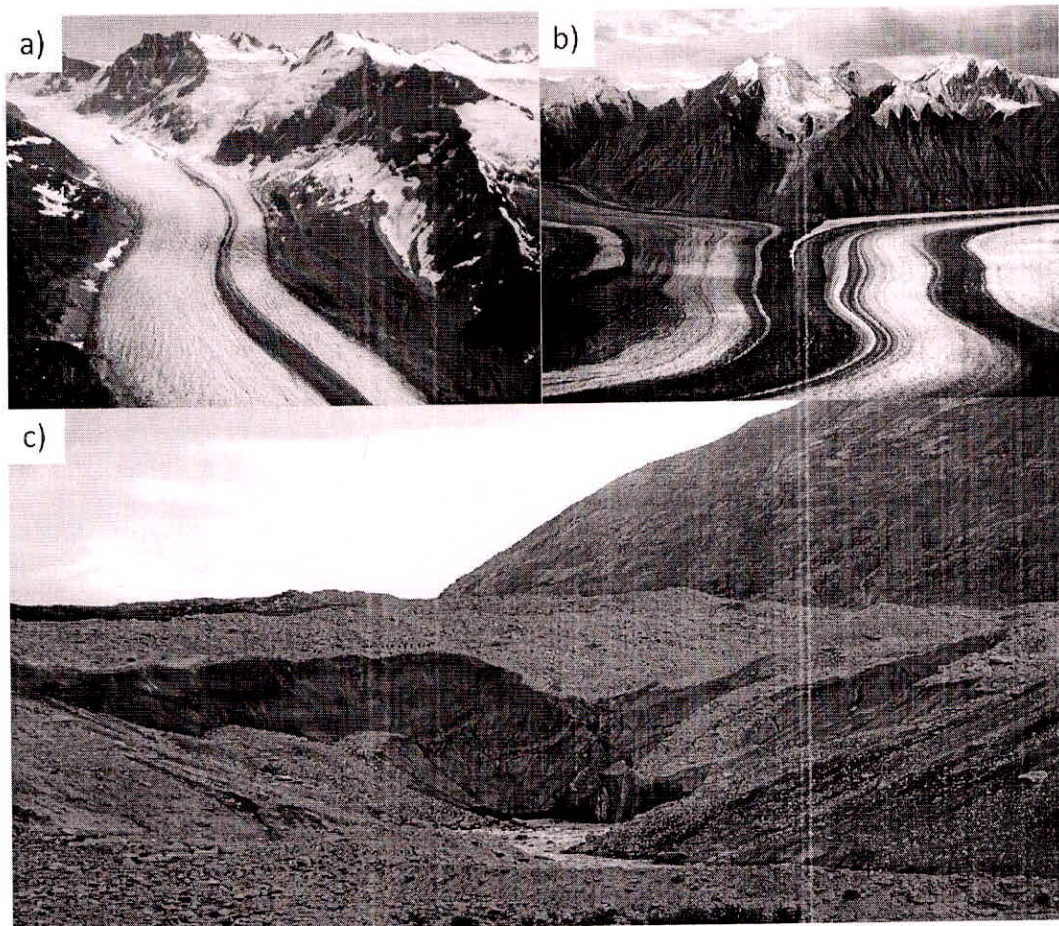


Fig. 1.8: (a) Lateral moraines along each side of a glacier. (b) Medial Moraine (c) Terminal Moraine (Baltoro Glacier, Pakistan)

Medial Moraines

When two glaciers flow side by side, they do not unite. The result is that two lateral moraines of these glaciers mingle together, giving rise to a medial moraine (Fig. 1.8b).

Terminal Moraines

A terminal moraine, also called end moraine, is a type of moraine that forms at the snout (edge) of a glacier, marking its maximum advance. At this point, debris that has accumulated by plucking and abrasion, and has been pushed by the front edge of the ice, is driven no further and instead is dumped in a heap (Fig. 1.8c). Because the glacier acts very much like a conveyor belt, the longer it stays in one place, the greater the amount of material that will be deposited. The moraine is left as the marking point of the terminal extent of the ice

1.3.10 Supraglacial Lakes

Glaciers in the Himalayas generally show the presence of large water bodies over the glacier surface within the ablation zone. Size of these water bodies can vary and some of them are large enough and deep enough to be called as lake. A supraglacial lake is any pond of liquid water on the top of a glacier (Fig. 1.9a). Although, these pools are ephemeral, they may reach kilometers in diameter and be several meters deep. They may last for months or even decades at a time, but can empty in the course of hours.

1.3.11 Supraglacial channels

Supraglacial channels of meltwater on a glacier are formed by the ice melting during the summer (Fig. 1.9b). It flows off the glacier, incising a number of cracks similar to an ordinary river system. These channels are often sinuous, and the water can flow at rates of up to several metres per second. During the summer months, when the quantity of the water stored, in supra glacial lakes, exceeds its capacity and combined with the fact that the lake waters become relatively warmer, melt cavities or melt channels develop along the base or in the walls of these lakes. It leads to the lake getting partially drained along a drainage channel over the glacier surface. Often under the enormous pressure of the melt water so released englacial conduits and channels develop in the body of the glacier.

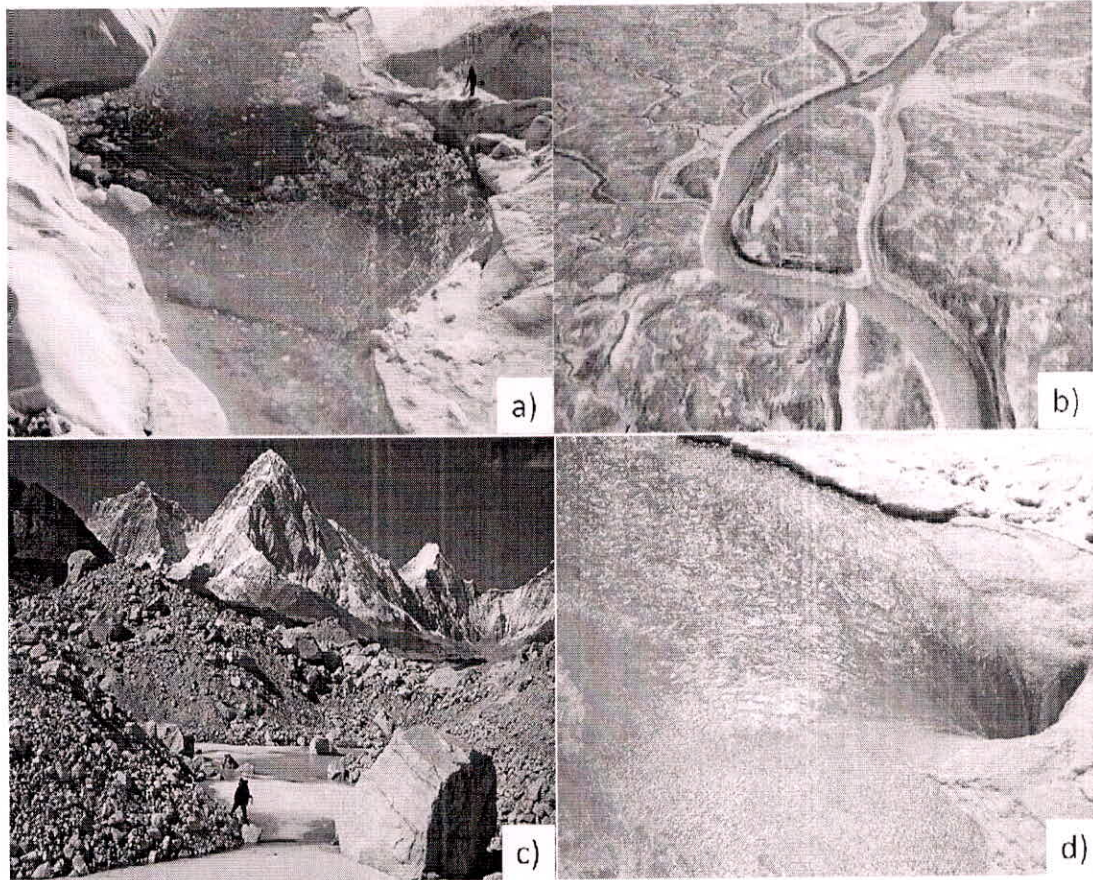


Fig. 1.9: a) Supraglacial lake b) Surface meltwater flowing through Supraglacial channel on glacier c) Supraglacial debris and d) Moulin structure

1.3.12 Supraglacial Debris

Debris that is carried on the surface of a glacier. It is normally derived from rock falls and tends to be angular in character (Fig. 1.9c). More rounded material may also be present, raised from the glacier bed or reworked by surface streams.

1.3.13 Moulin

A moulin or glacier mill is a roughly circular, vertical to nearly vertical well-like shaft within a glacier or ice sheet which water enters from the surface (Fig. 1.9d). They can be up to 10 meters wide and are typically found on ice sheets and flat areas of a glacier in a region of transverse crevasses. Moulin can reach the bottom of the glacier, hundreds of meters deep, or may only reach the depth of common crevasse formation (about 10–40 m) where the stream flows englacially. They are the most typical cause for the formation of a glacier cave. Moulin are parts of the internal structure of glaciers, that carry meltwater from the surface down to wherever it may go (Raina and Srivastava, 2008). Water from a moulin often exits the glacier at base level, sometimes into the sea, and

occasionally the lower end of a moulin may be exposed in the face of a glacier or at the edge of a stagnant block of ice. Water from moulins may help lubricate the base of the glacier, affecting glacial motion.

1.3.14 Crevasses

A crevasse is a deep crack, or fracture, found in an ice sheet or glacier, as opposed to a crevice that forms in rock (Fig. 1.10). Crevasses form as a result of the movement and resulting stress associated with the shear stress generated when two semi-rigid pieces above a plastic substrate have different rates of movement. The resulting intensity of the shear stress causes a breakage along the faces.

Transverse crevasse

Transverse crevasses are the most common crevasse type. They form in a zone of longitudinal extension where the principal stresses are parallel to the direction of glacier flow, creating extensional tensile stress (Fig. 1.10a). These crevasses stretch across the glacier transverse to the flow direction, or cross-glacier. They generally form where a valley becomes steeper.

Longitudinal crevasses

Longitudinal crevasses form parallel to flow where the glacier width is expanding (Fig. 1.10b). They develop in areas of tensile stress, such as where a valley widens or bends. They are typically concave down and form an angle greater than 45° with the margin.

Splaying Crevasses

Splashing crevasses result from shear stress from the margin of the glacier and longitudinal compressing stress from lateral extension (Fig. 1.10c). They extend from the glacier's margin and are concave up with respect to glacier flow, making an angle less than 45° with the margin. At the centre line of the glacier, there is zero pure shear from the margins, so this area is typically crevasse-free.

Marginal crevasses

A crevasse pointing obliquely up-valley that develops on either side of some valley glaciers. This term is used for the crevasses that develop along the lateral margins of the valley glaciers (Fig. 1.10d). Marginal crevasses develop when the central area of the glacier moves considerably faster than the outer edges.

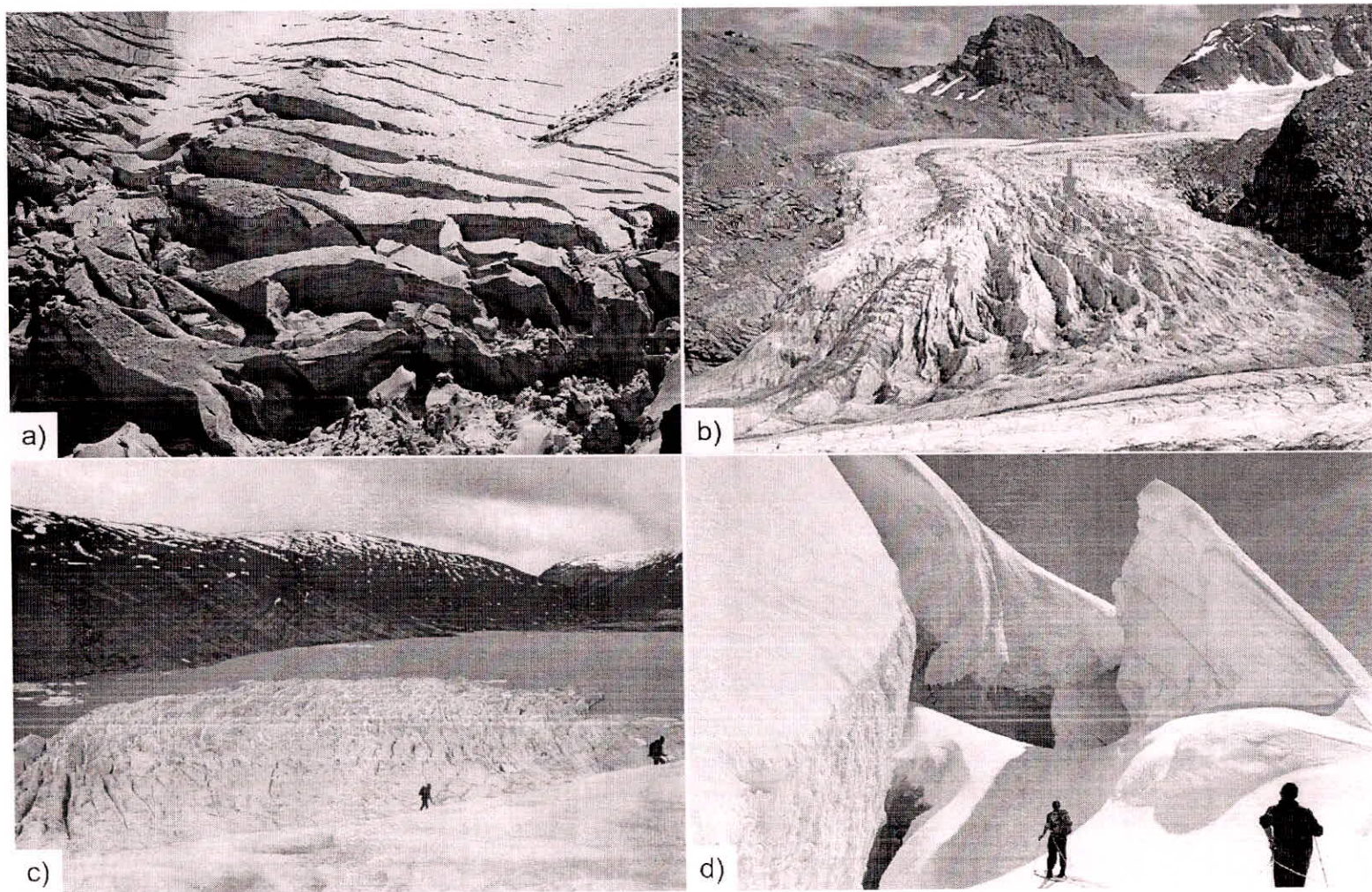


Fig. 1.10: a) Transverse crevasses, b) Longitudinal crevasses, c) Splaying Crevasses and d) Marginal crevasses.

1.4 Research Motivation

Glaciers, a perennial source of fresh water are the important component of the Earth's hydrological cycle. However, it is worldly confirmed by several studies that in many mountainous ranges of the world including the Arctic and the Antarctica; the glaciers melting have been accelerated over the past three decades and now it is occurring at an unprecedented rate (Bauder *et al.*, 2007; Ohmura *et al.*, 2007).

Current concerns about global warming have produced widespread scientific interest in the behaviour of glaciers. The impact of changing climate and melting of glaciers due to rising temperature are now creating a devastating effect. For example, the melting of Himalayan glaciers will affect the drinking supplies of the Asia where the highest population of the world live who rely only on the meltwater rivers originating from these glaciers.

The ongoing glacier retreat influences the hydrological conditions in many alpine watersheds (Stahl *et al.*, 2008). The water supply from snow and glacier melt is very sensitive to a warming climate (Barnett *et al.*, 2005; Bradley *et al.* 2006). In glacierized watersheds, the water sources that contribute to discharge shift as temperatures increase. The currently negative glacier mass balance results in an increased discharge, with large contributions from ice melt, during a transitional phase. In future, after additional glacier loss, runoff will likely decline, however, with an increasing contribution from seasonal snow melt and rainfall rather than ice melt (Hannah *et al.*, 2007). Thus, in glacierized watersheds, the hydrological conditions have changed much during the past hundred years due to glacier retreat, and the water availability in the future is also likely to differ much from the conditions observed today. The ability to reconstruct and predict changes in hydrological conditions is among other reasons important because the water leaving those watersheds is often used for production of hydropower energy (Schaeffli *et al.*, 2007). Therefore, future variations in discharge may have a large economic impact and proper management of this resource is necessary, in particular long - term planning of reservoir capacity. A sound understanding of the interaction between meteorological forcing, snow cover and glaciers on the scale of whole catchments is necessary for developing hydrological models yielding, for example, reliable discharge estimations. With such models, the hydrology of glacierized watersheds may be reconstructed using long-term meteorological recordings or can be predicted based on climate change scenarios.

Runoff from the glacier is naturally regulated and the water generated through the melting processes of Himalayan glaciers is extremely beneficial for the human beings especially for the irrigation, ingestion and generation of electricity through hydropower (Haritashya *et al.*, 2010; Kumar *et al.*, 2014; Srivastava *et al.*, 2014). Previous studies have shown that 70-80 % of Himalayan glaciers are debris covered (Shroder *et al.*, 2000), which can significantly impact on the surface ablation rate and the meltwater production. With the fast growing need for fresh water and awareness for its sustainable use, the snow, ice and glacier studies have acquired considerable importance these days. The vision “*Uttarakhand: An Urja Pradesh*” of Uttarakhand Government to make the state as Energy-State has inspired to embark upon an ambitious plan of having more than 180 big and small hydroelectric projects to generate more than 21,200 MW electricity.

Therefore, an attempt has been made in this dissertation report to understand the influence of changes in climate and snow/glacier cover to streamflow over the Himalayan region assessed by semi-distributed conceptual hydrological model.

1.5 Aims and Objectives

To study all the glaciers in the Himalaya is tough due to rugged terrain and harsh weather conditions. Therefore, in this scientific approach Gangotri Glacier has been selected for the study of hydro-meteorological conditions and Bhagirathi streamflow. The objectives are -

- a) To study the glaciers, glacier system of Himalaya and hydrological models.
- b) To study the hydro-meteorological conditions of Gangotri Glacier.
- c) To estimate the stream flow in the Glacierized Catchment using Semi-distributed Conceptual Hydrological model.

1.6 Thesis Overview

The dissertation report is structured in to 6 chapters. The details are-

Chapter 1 Introduction

The chapter gives knowledge about the basics of glaciers, different types of glaciers and how glaciers are formed. Moreover, the chapter Glacial Geomorphology and highlights the motivation of research.

Chapter 2 Glacier Distribution

This chapter provides the knowledge about the world glacier distribution and Himalayan Glacier system.

Chapter 3 Hydrological Modelling

This chapter focuses on the study of Hydrological runoff model. It provides comprehensive information about types of modeling used to estimate streamflow of Glacierized basin.

Chapter 4 Hydro-Meteorological Conditions of Gangotri Glacier

This chapter describes the climatic conditions of Gangotri Glacier system, for summer season (May to September) of 2000 and 2001. Description of meteorological observations and the analysis of collected meteorological data.

Chapter 5 Hydrological Modelling of Streamflow

All the interpolated and simulated hydro-meteorological parameters along with Goodness of fit of Gangotri Glacier system, for summer season (May to September) of 2000 and 2001 are discussed in this chapter.

Chapter 6 Synthesis

The findings of the research are concluded in this chapter by discussing the main results and synthesis the conclusion of all chapters to present the overall result and goal of study. Further, the chapter focuses on the future implication of this study especially for the glacier-climate interaction.

Reference All the references cited in this dissertation are listed in this section

Chapter -2

Distribution of Himalayan Glaciers and River System

2.1 Indian Himalayan Glacier System

Glaciers, ice sheets and ice caps cover approximately 10 % of the Earth's surface and out of the Earth's total ice reserves that exist in the form of ice fields and glaciers; almost 98 % (15 million km²) of the glaciers is confined to the Antarctica and the Greenland (Cuffey and Paterson, 2010). Remaining 2 % of the glacier cover exists in the mountains of Asia, Europe and America (WGMS, 1989; Paterson, 1994; Raina and Srivastava, 2008). Glaciers store about 75 % of the world's fresh water.

In Asia, the Hindu-Kush Himalayan (HKH) region comprises the biggest mountain range (4 million km²) on the Earth and considered to be the largest store house of snow and glaciers outside the polar region; subsequently known as the '*Third Pole*' (Dyhrenfurth 1955, Bajracharya and Shrestha, 2011) (Fig. 2.1). There is 0.77% glacierized area (approximately 1,19,126 km²) covered in the HKH region out of which nearly 37,465 km² in India, 16,933 km² in Pakistan, 5,322 km² in Nepal and 59,406 km² in China (Raina and Srivastava, 2008; Bajracharya and Shrestha, 2011). More than 10 major rivers originate from here including the Indus, the Ganga and the Brahmaputra (Eriksson et al., 2009).

The glaciers of high Asia comprise 50 % by area of all glaciers outside the Polar region and they are containing nearly 33 times areal cover of the glaciers in the European Alps. The Himalaya, include Tibetan plateau contains one of the largest concentrations of the glaciers and permanent snowfield. About 17 % of the Himalaya is permanently covered with glaciers and additional nearly 30-40 % area is snow covered (Vohra, 1978). The Great Himalaya, the highest mountain range in the world, extends along the northern frontiers of Pakistan, India, Nepal, China, Bhutan and Bangladesh broadly classified into Western, Central and Eastern Himalaya (Table 2.1).

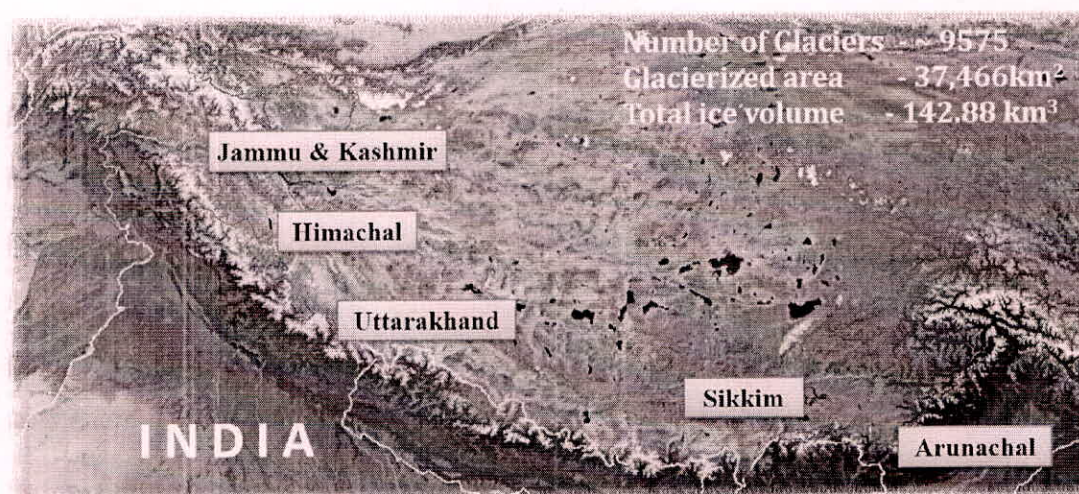


Fig. 2.1: Indian Himalayan Glacier system.

The Himalayan Mountain System is home for 14 highest peaks including the Mount Everest. The Himalaya is a spectacle of awesome dimensions, ranges along ranges, tiers of rock, sharp sky piercing peaks and canyons, deep beyond measure. The Himalaya supports nearly half of humanity of the Indian sub-continent. In its 2,600 km. length, Himalaya presents endless variations in terms of climate variability and topography. From the sub-tropical jungles of Arunachal to the cold desert of the Ladakh; primulas and rare orchids, frozen waterfalls and verdant forest, bare rocks and glacial wastes make Himalaya resourceful. The Western Himalaya, from the Himachal Pradesh onwards, has a much greater depth or width, than the Eastern Himalaya. The Eastern Himalaya is also climatically very different. High rainfall and gentler conditions make the Eastern Himalaya a recognized haven for biological diversity.

Table 2.1. Basin/Sub-basin wise distribution of glaciers in the Indian administrative part of the Himalaya.

State	Basin/ Sub-Basin	Identification Code	Glaciers (No.)	Glacierized Area (km ²)	Ice Volume (km ³)
<i>Indus Basin (Western Himalaya)</i>					
Jammu & Kashmir	Gilgit	5Q 11	535	8,240	--
	Shyok	5Q 131	2454	10,810	601.71
	Nubra	5Q 13105	204	1,536	204.55
	Indus	5Q 132	1,796	8,370	73.58
	Kishanganga	5Q 2111	222	163	5.93
	Jhelum	5Q 214	133	94	3.30
<i>Total</i>			5344	29,213	889.07
Himachal Pradesh	Ravi	5Q 2102	172	193	8.04
	Chenab	5Q 212	1,278	3,059	206.30
	Beas	5Q 221	277	579	36.93
	Satluj	5Q 222	926	635	34.95
<i>Total</i>			2653	4466	286.22
<i>Ganga Basin (Central Himalaya)</i>					
Uttarakhand	Yamuna	5O 112	52	144	12.20
	Bhagirathi	5O 131	238	755	67.02
	Alaknanda	5O 132	407	1,229	86.38
	Kaliganga/ Ghagra	5O 13	271	729	43.77
<i>Total</i>			968	2,857	209.37
<i>Brahmaputra Basin (Eastern Himalaya)</i>					
Arunachal Pradesh	Brahmaputra	5O 200	161	223	10.0
Sikkim	Tista	5O 201	449	706	39.61
<i>Total</i>			610	929	49.61
Total (Indus + Ganga + Brahmaputra)			9,575	37,465	1,434.38

(Source: Raina and Srivastava, 2008; Sangewar and Shukla, 2009)

2.2 Himalayan River System

Himalayan River system that is sustained, at the head, by glacier bearing basins, had been initially classified into three first order basins, 5Q (Indus basin), 5O (Ganga basin) and 5P (Brahmaputra basin) (Fig. 2.2). Subsequently the persons associated with the glacier inventory project, felt that as Brahmaputra flowed into the Ganga, the two basins could be classified as one basin. Brahmaputra basin would then become as second order basin of the Ganga basin: 5-O-2. Indus and Ganga (Ganges) thus became the two main first order rivers systems, originating from the Himalayas, and have been designated the numbers 5Q and 5O in the International coding system, (Fig. 2.3).

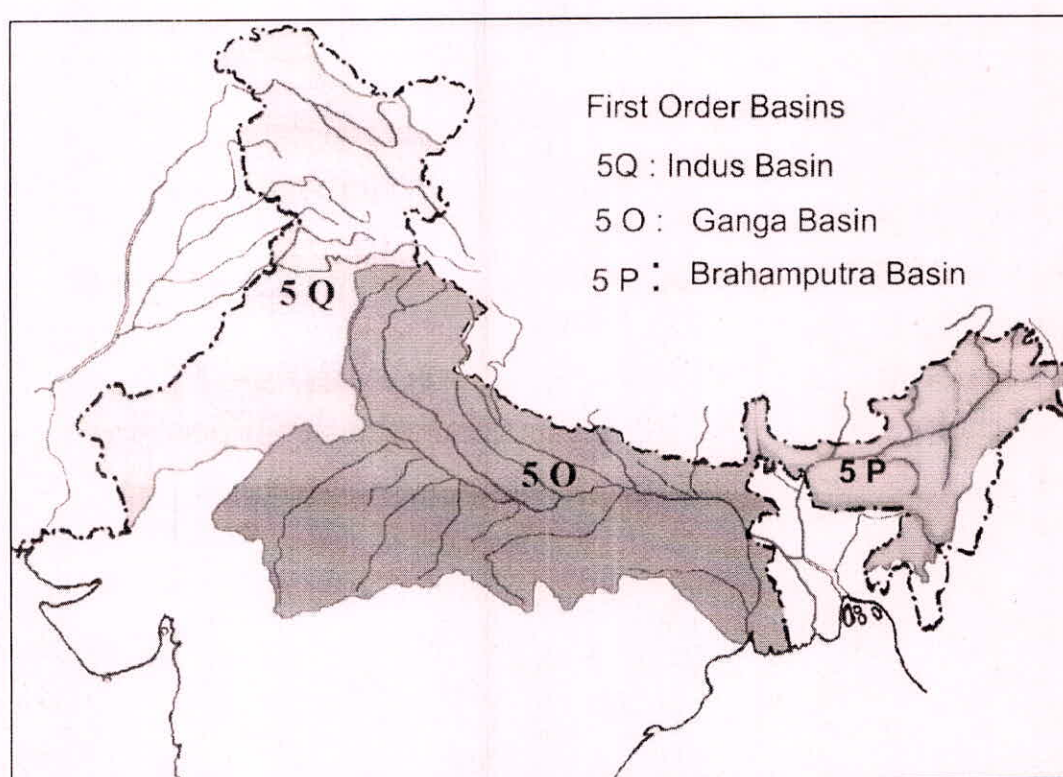


Fig. 2.2: First order basins of the Himalayas.

2.2.1 Fifth order basins of Indus (5 Q)

The fifth order Indus basin (5 Q) falling in the Indian territory, supports two second order basins, Indus (5Q 1) and Satluj (5Q 2) (Fig. 2.2). The former has three third order basins, Zhob, (5Q 11), Kabul (5Q 12) and Indus (5Q 13). The first two flow in Pakistan and thus have not been classified further for glacier inventory purpose, while the third one supports two fourth order basins, Shyok (5Q 131) and Indus (5Q 132), having eight

fifth order basins, each. There are also seven fifth order basins which join Indus (5Q 13) directly without any fourth order basin in between.

2.2.2 Fifth order basins of Ganga (5 O)

Fifth order basins of Ganga (5 O) Ganga basin (5O) supports two second order basins viz., Ganga (5O 1) and Brahmaputra (5O 2), which are further sub-divided as under:- Ganga (5O 1) has three third order basins, viz. Kosi (5O 11), Yamuna (5O 12) and Ganga (5O 13) (Fig. 2.3). Glacier covered portion of Kosi lies in Nepal, and hence has not been further classified, whereas Yamuna and Ganga have six and two fourth order basins, respectively. The fourth order basins viz. Gandak, Son and Ghagra join directly the second order Ganga (5O 1), without the intervening third order. Again, Damodar (5O 001), another fourth order basin, joins first order Ganga (5O), without the intervening second and third order basins.

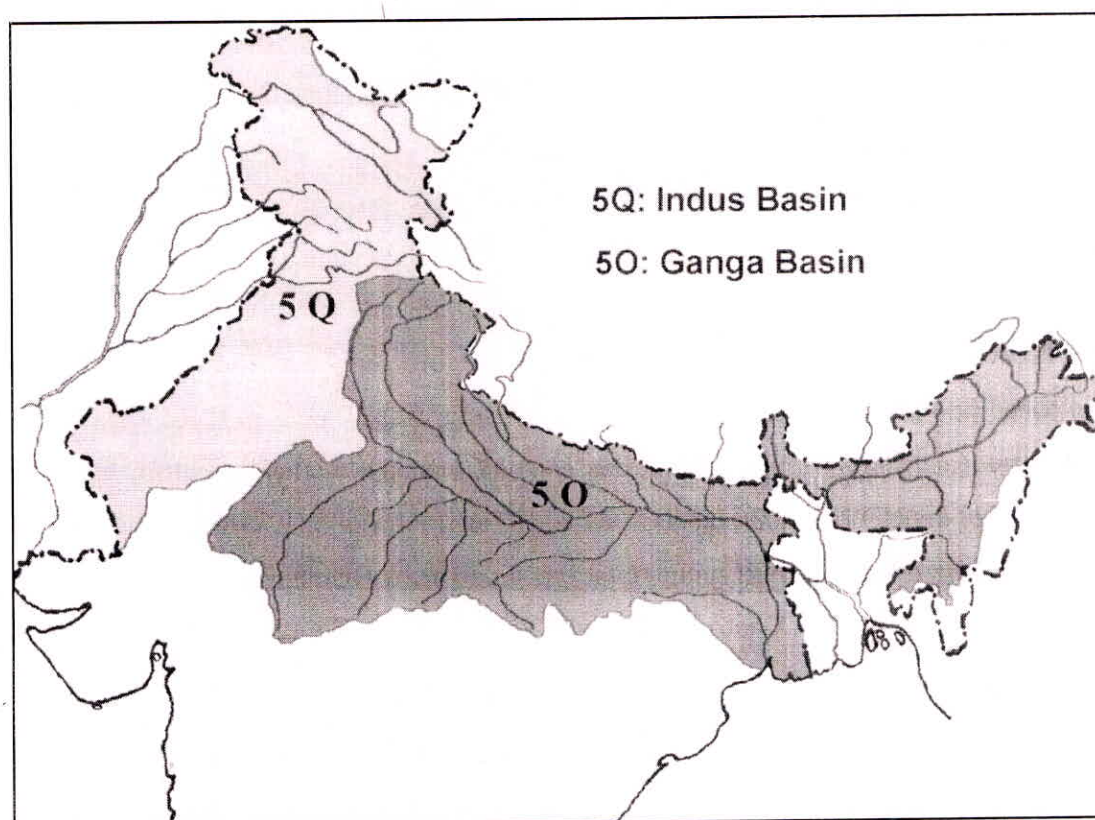


Fig. 2.3: Revised First order basins of the Himalaya.

In all there are 22 fifth order basins distributed in the four glacier bearing fourth order basins, of Ghagra (5O 103), Yamuna (5O 123), Bhagirathi (5O 131) and Alaknanda (5O 132). The Damodar (5O 001), Gandak (5O 101), Son (5O 102), Ken (5O 121), Betwa. (5O 122), Banas (5O 124), Parvati (5O 125) and Chambal (5O 126), which

are all fourth order basins are not glacier bearing and are mostly flowing in the Gangetic plain, and hence no attempt has been made to demarcate the fifth order basins of these rivers. The basins have been shown Fig.

The second order Brahmaputra (5O 2) has two third order basins viz., Dihang (5O 21) and Dibang (5O 22). The former has three fourth order basins, out of which two, i.e., 5O212 and 5O213 are in Tibet, whereas the third one Dihang (5O211) supports three fifth order basins; the latter has two fourth order basins, viz. Dibang (5O221) and Lohit (5O 222), supporting four and two fifth order basins, respectively. Another fourth order, Subansiri (5O 202) and Tista (5O 201) joins directly the second order Brahmaputra (5O 2), without an intervening third order, whereas Meghna (5O 002) joins the first order Ganga (5O), directly. Out of the three fifth order basins joining Brahmaputra (5O 2) directly, Manas (5O 200 02) and Kameng (5O 200 03) lie in the Indian Territory, while the third one 5O 200 01 lies in the Bhutan.

2.3 Climatic Conditions and Cryosphere Status of Uttarakhand

In Central Himalayan, the regional climate is largely controlled by the two major weather systems i.e. (i) the mid-latitude Western Disturbance (WD) and (ii) the Indian Summer Monsoon (ISM; Owen *et al.*, 1996). The mid-latitude WD brings moisture from the Mediterranean, the Black and the Caspian seas, and delivers over the North-West and Central Himalaya in the winter (Immerzeel *et al.*, 2010; Bolch *et al.*, 2012). On the other hand, the ISM brings moisture from the Indian Ocean and Bay of Bengal and delivers substantial amount of precipitation during the summer (Chao and Chen, 2001; Gadgil, 2003; Yancheva *et al.*, 2007). Based on the relative dominance of these two weather systems, the Uttarakhand state receives solid precipitation both due to the WD in winter and the ISM in summer (Fig. 2.4). The average temperature of the state varies from below 0 to 30°C and average total rainfall varies between 500 and 3000 mm (Government of Uttarakhand, 2014).

Uttarakhand is most vulnerable to climate mediated risks. Mountainous regions are vulnerable to climate change and have shown “above average warming” in the 20th century. Studies conducted by MoEF have shown an increase in annual temperature in the Himalayan region, a net increase in rainfall in the 2030s with respect to the simulated rainfall of the 1970s in the Himalayan region (Government of Uttarakhand, 2014).

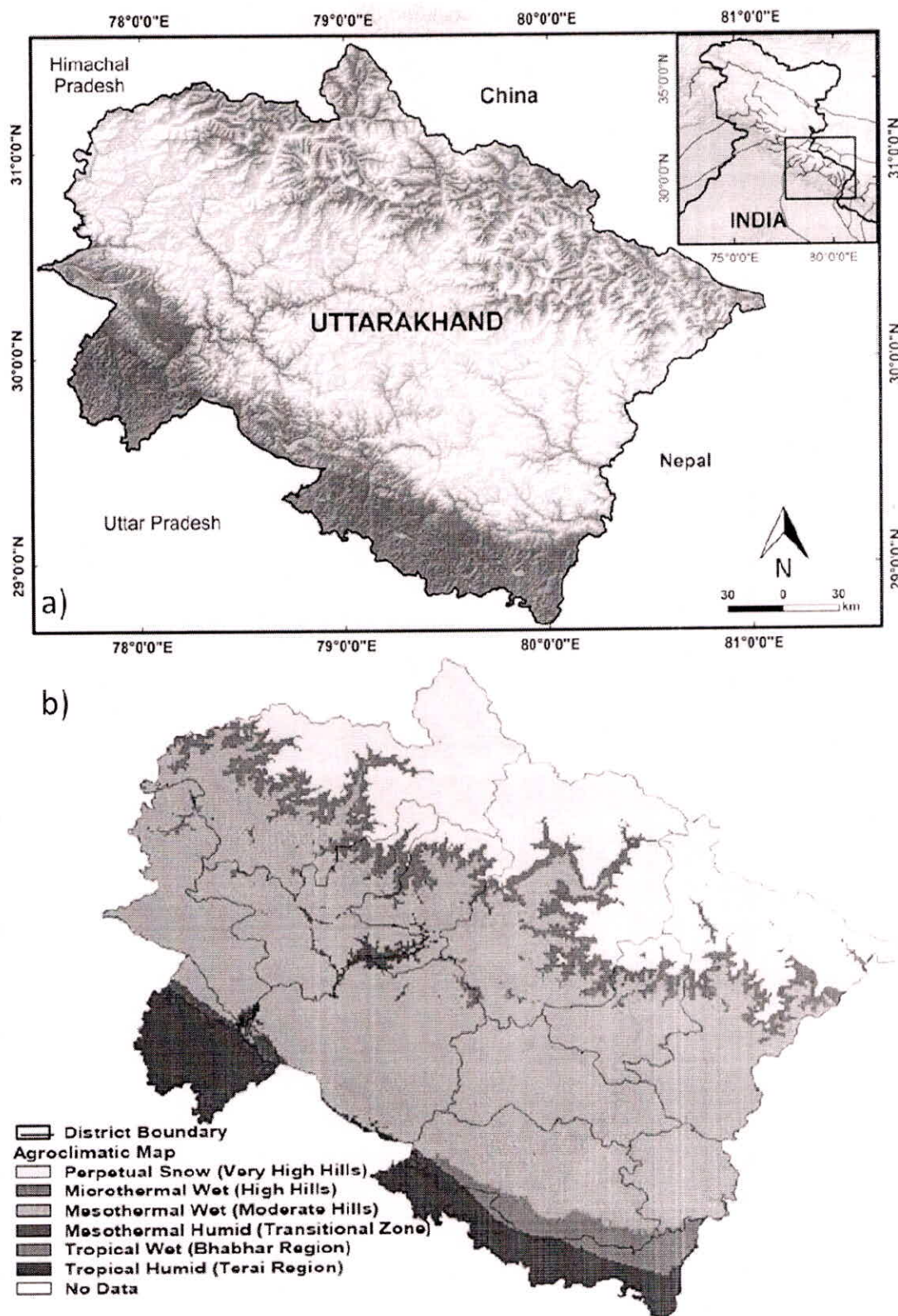


Fig. 2.4: a) Location map and b) Overall climatic condition of Uttarakhand state (source: Government of Uttarakhand, 2014).

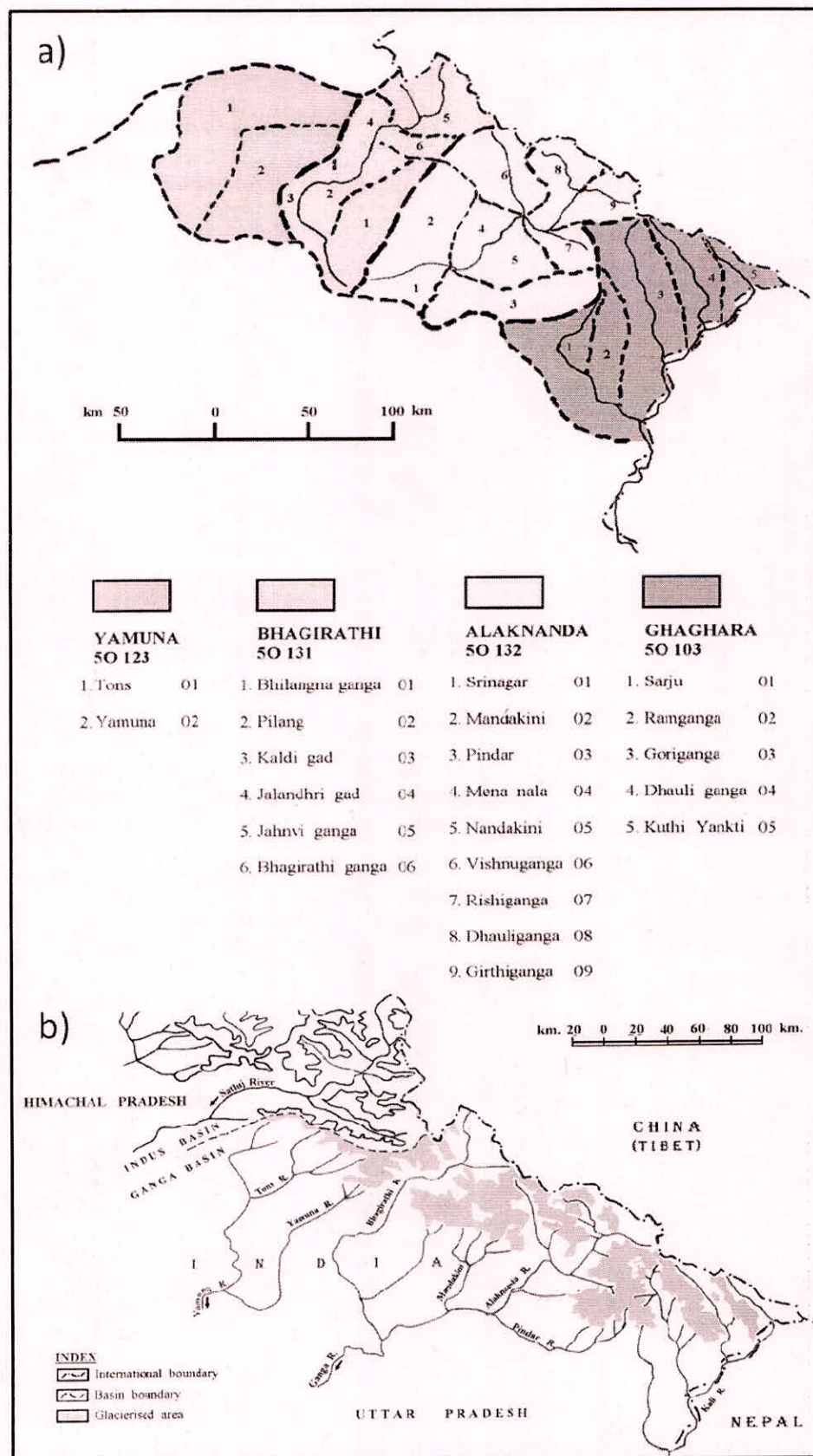


Fig. 2.5: a) 5th Order basin of upper reaches of Ganga catchment in Uttarakhand State and b) Major River System and glacierized basin of Uttarakhand (source: Raina and Srivastava, 2008)

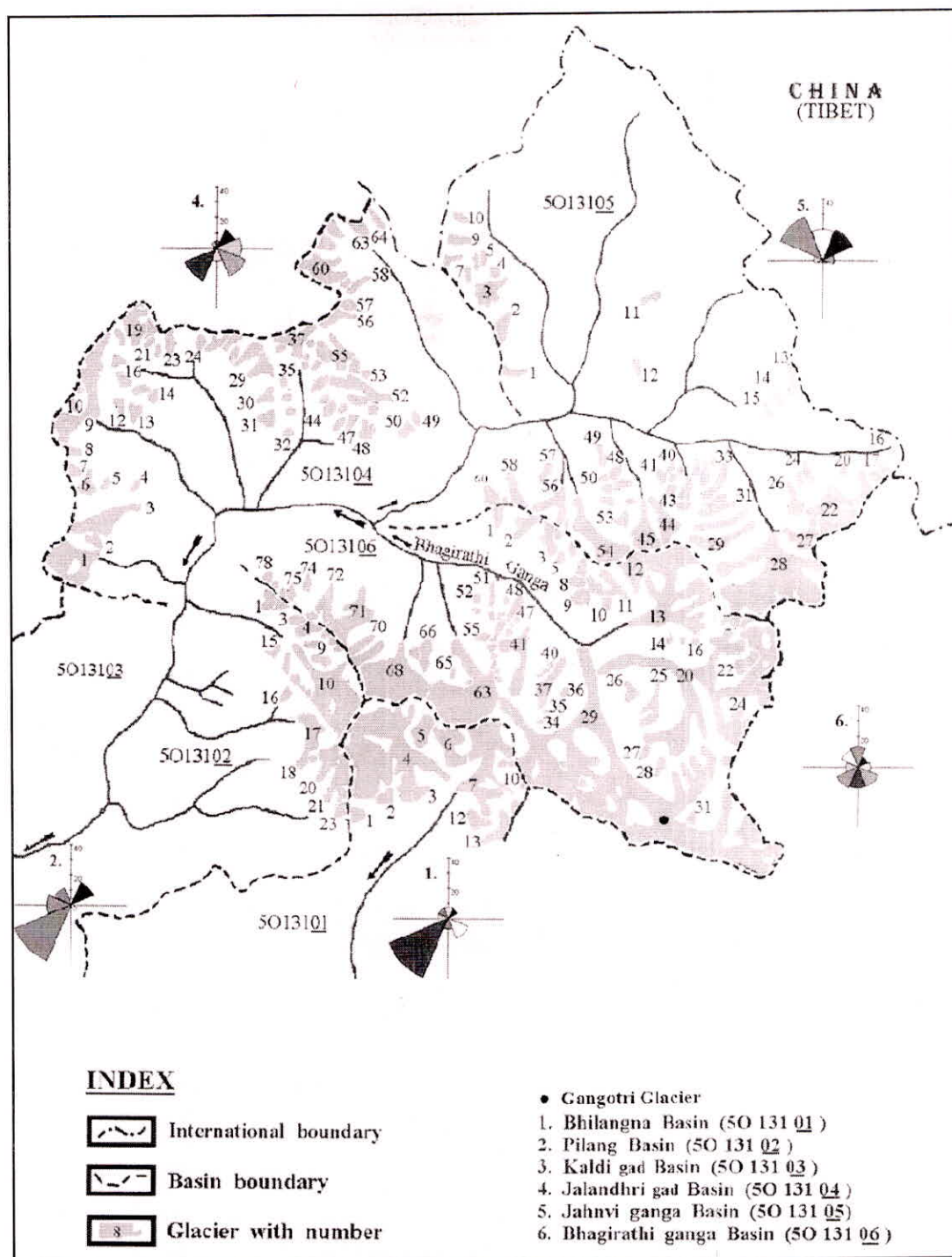


Fig. 2.6: Glacier map of Bhagirathi basin, Uttarakhand (source: Raina and Srivastava, 2008).

Impacts are expected to range from reduced genetic diversity of species to erratic rainfall leading to flash floods to glacial melt in the Himalayas leading to increased flooding that will affect water resources within the next few decades. Some of the reported climate-change-induced changes in the Uttarakhand Himalaya include receding glaciers and an upwardly moving snowline, depleting natural resources, erratic rainfall,

irregular winter rains, advancing cropping seasons, fluctuations in the flowering behaviour of plants, shifting of cultivation zones of crops, reduction in snow in winter, a rise in temperature, an increasing intensity and frequency of flash floods and drying up of perennial streams.

In Uttarakhand region, the Ganga Basin 5 O covers an extensive area between the Tons River basin in West and Kali Ganga in East (Fig. 2.5a). The Uttarakhand Himalaya covers Yamuna, Bhagirathi, Alaknanda and Kali Ganga / Ghagra basins (Fig. 2.5b). The state has 968 glaciers with 2,850 km² glacier covered area which is less than 8% to the total basin area (Raina and Srivastava, 2008). This shows ~13.8% glacier area of the entire Indian Himalaya exists in Uttarakhand. The largest number of the glaciers (total-407) has been marked in the Alaknanda basin (1,229 km² glaciated area) followed by the Kali Ganga / Ghagra (total-271; 729 km² glaciated area), Bhagirathi (total-238; 755 km² glaciated area) and Yamuna (total-52; 144 km² glaciated area). The Yamuna basin is further subdivided into two sub basins (Tons and Yamuna), Bhagirathi into six sub basins (Bhilangna Ganga, Pilang, Kaldi Gad, Jalandhri Gad, Jahnavi Ganga, Bhagirathi Ganga), Alaknanda into nine sub basins (Srinagar, Mandakini, Pindar, Mena Nala, Vishnu Gad, Rishi Ganga, Dhauri Ganga, Girthi Ganga) and Ghagra into five sub basins (Sarju, Ram Ganga, Gori Ganga, Dhauri Ganga, Kuthi Yankti). The Gangotri (length: 30 km; surface area: 144 km²) - a well-documented glacier Bhagirathi basin, is the largest glacier of the state (Fig. 2.6). The Gangotri glacier and river Ganga has important status in the Indian culture. Scriptures and faith of the population consider it a sacred river. Perhaps two billion of the world's people depend on these rivers for their water and food supply (Srivastava, 2012). Lot of population is dependent on it for irrigation and drinking water purposes.

Chapter – 3

Hydrological Modelling

3.1 Introduction

The term hydrology can be treated as an important subject for the people and their environment. It treats water of the earth, their occurrence, circulation and distribution, their chemical and physical properties and their reaction with the environment including their relation to living things (Ray, 1975). It also deals with the relationship of water with the environment within each phase of hydrologic cycle. Due to rapid urbanization and industrialization including deforestation, land cover change, irrigation, various changes have been occurred in hydrologic systems. Along with climate change, soil heterogeneity has also got a direct impact on the discharges of many rivers in and around the world.

Different hydrologic phenomena and hydrologic cycle are to be thoroughly studied in order to find out these variations. Now days, various hydrological models have been developed across the world to find out the impact of climate and soil properties on hydrology and water resources. Each model has got its own unique characteristics. The inputs used by different models are rainfall, air temperature, soil characteristics, topography, vegetation, hydrogeology and other physical parameters. All these models can be applied in very complex and large basins.

3.2 Types of Hydrological Modeling

According to Sorooshian et al. (2008), a model is a simplified representation of real world system. The best model is the one which give results close to reality with the use of least parameters and model complexity. Models are mainly used for predicting system behavior and understanding various hydrological processes. A model consists of various parameters that define the characteristics of the model. A runoff model can be defined as a set of equations that helps in the estimation of runoff as a function of various parameters used for describing watershed characteristics. The two important inputs required for all models are rainfall data and drainage area. Along with these, water shed characteristics like soil properties, vegetation cover, watershed topography, soil moisture content, characteristics of ground water aquifer are also considered.

Hydrologic models are simplified, conceptual representations of a part of the hydrologic cycle. They are primarily used for hydrologic prediction and for understanding hydrologic processes. These are now a day considered as an important and necessary tool for water and environment resource management. Three major types of hydrologic models can be distinguished:

3.2.1 Empirical models (Metric model)

These are observation oriented models which take only the information from the existing data without considering the features and processes of hydrological system and hence these models are also called data driven models (Devi G. K., et. al.,2015) . It involves mathematical equations derived from concurrent input and output time series and not from the physical processes of the catchment. These models are valid only within the boundaries. Unit hydrograph is an example of this method. Statistically based methods use regression and correlation models and are used to find the functional relationship between inputs and outputs.

3.2.2 Conceptual methods (Parametric models)

This model describes all of the component hydrological processes. It consists of a number of interconnected reservoirs which represents the physical elements in a catchment in which they are recharged by rainfall, infiltration and percolation and are emptied by evaporation, runoff, drainage etc. Semi empirical equations are used in this method and the model parameters are assessed not only from field data but also through

calibration. Large number of meteorological and hydrological records is required for calibration. The calibration involves curve fitting which makes the interpretation difficult and hence the effect of land use change cannot be predicted with much confidence. Many conceptual models have been developed with varying degree of complexity. Stanford Watershed Model IV (SWM) is the first major conceptual model developed by Crawford and Linsley in 1966 with 16 to 20 parameters.

3.2.3 Physically based model

This is a mathematically idealized representation of the real phenomenon. These are also called mechanistic models that include the principles of physical processes. It uses state variables which are measurable and are functions of both time and space. The hydrological processes of water movement are represented by finite difference equations. It does not require extensive hydrological and meteorological data for their calibration but the evaluation of large number of parameters describing the physical characteristics of the catchment are required (Abbott et al., 1986 a). In this method huge amount of data such as soil moisture content, initial water depth, topography, topology, dimensions of river network etc. are required. Physical model can overcome many defects of the other two models because of the use of parameters having physical interpretation. It can provide large amount of information even outside the boundary and can be applied for a wide range of situations. SHE/ MIKE SHE model is an example of the physically based model (Abbott et al., 1986 a, b).

3.3 SNOWMOD

3.3.1 Introduction

There is substantial contribution from snowmelt runoff to the annual stream flows of the Himalayan Rivers (Singh *et al.*, 1997a; Singh & Jain, 2002). The water yield from a high Himalayan basin is roughly twice as high as that from an equivalent basin located in the peninsular part of India. A higher water yield from the Himalayan basins is mainly due to the large inputs from the snowmelt and glaciers. Depending upon the climatic conditions, the snowpack depletes either fully or partially during the forthcoming summer season. Because of variation in climatic conditions and changes in the areal extents of snow-covered area (SCA) and snow-free area (SFA) with time, the contributions from the rain and snow to the streamflow vary with season. The contribution from rain dominates in the lower part of the basins (about <2000 m).

The middle and upper parts of the basins (about >2000 m) receive contributions from both rain and snowmelt and this changes with altitude. As the elevation of the basin increases, the rain contribution to streamflow reduces and the snowmelt contribution increases. Runoff is dominated by the snowmelt runoff above 3000 m altitude. Modelling of streamflow involves transformation of precipitation received in the basin to the outgoing streamflow by considering losses to the atmosphere, temporary storage, lag and attenuation etc. Most hydrological runoff models, which account for snow need snowfall data as input for snowmelt estimation (Singh & Singh, 2001). For the high altitude regions in the Himalayan basins, snowfall data are not available. Poor accessibility to the high altitude sites due to rugged terrain makes it difficult to install, monitor and maintain the meteorological instruments, so the observational network is poor and application of models which need snowfall data as input is hampered. The snowmelt runoff model (SRM; Martinec et al., 1994) uses snow-covered area as input instead of snowfall data, but it does not simulate the baseflow component of runoff. In other words, SRM does not consider the contribution to the groundwater reservoir from snowmelt or rainfall, nor its delayed contribution to the streamflow in the form of baseflow, which can be an important component of runoff in the Himalayan rivers, and plays an important role in making these rivers perennial. Almost all the streamflow during winter, when no rainfall or snowmelt occurs, is generated from the baseflow.

3.3.2 Structure of the model and its application

The snowmelt model (SNOWMOD) is designed to simulate daily streamflow for mountainous basins having contribution from both snowmelt and rainfall. The process of generation of streamflow from such basins involves primarily the determination of the input derived from snowmelt and rain, and its transformation into runoff. For simulating the streamflow, the basin is divided into a number of elevation zones and various hydrological processes relevant to snowmelt and rainfall runoff are evaluated for each zone. The model deals with snowmelt and rainfall runoff by performing the following three operations at each time step: (a) available meteorological data are extrapolated to the different elevation zones, (b) rates of snowmelt and/or rainfall are calculated at different points, and (c) snowmelt runoff from SCA and rainfall runoff from SFA are integrated, and these components are routed separately with proper accounting of baseflow to the outlet of the basin.

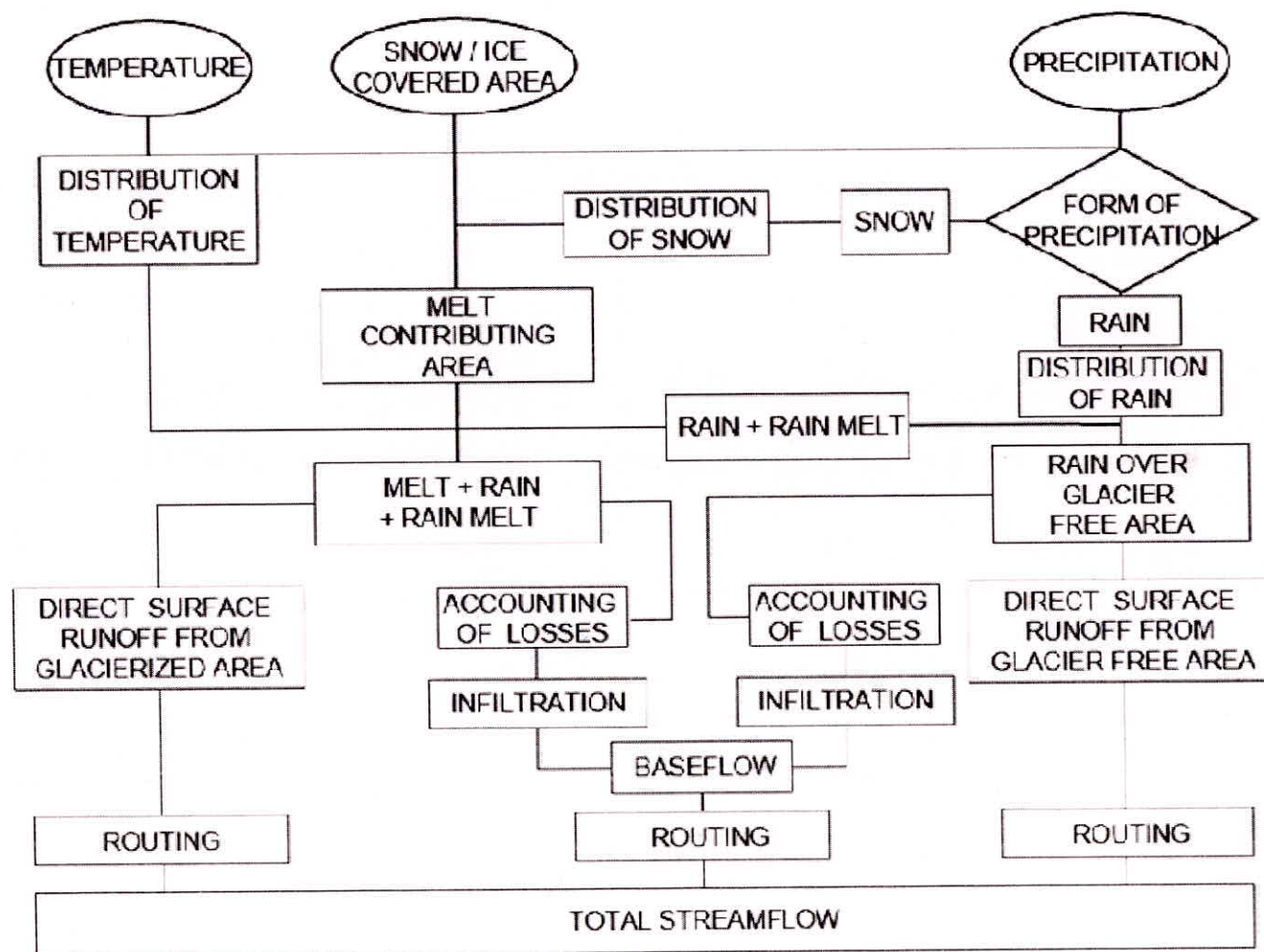


Fig. 3.1: Structure of the snowmelt model (SNOWMOD).

3.3.3 MODELLING OF STREAMFLOW

Temperature and rainfall are the main input to the model. In addition, information on the physical features of the basin, which includes glacierized area, elevation zones and their areas, total area of the basin, altitude of meteorological station, and other watershed characteristics affecting runoff are used in the model. In a mountainous basin, temperature and precipitation vary with elevation. The temperature generally decreases with elevation and is influenced by aspect and topographic shading; precipitation generally increases with elevation, as does the proportion that falls as snow, and is influenced mostly by aspect, slope position and other exposure indices.

Precipitation The distinction between rain and snow for each elevation zone is important for all the melt runoff models because precipitation falling in the form of rain and snow behaves differently in terms of contribution to the streamflow. The contribution of rain to the streamflow is faster than that of snow because snow is stored in the basin until it melts, whereas rain contributes to streamflow almost immediately. The temperature in a particular elevation zone determines the form of precipitation and the model handles it accordingly. A critical temperature, T_c , is specified in the model to determine whether the measured precipitation was rain or snow. Therefore, the value of T_c is usually selected slightly above the freezing point ($T_c = 2^\circ\text{C}$ in the present study). The model uses the following concept to determine the form of precipitation:

- i) if $T_m \geq T_c$, all precipitation is considered as rain
- ii) if $T_m \leq 0^\circ\text{C}$, all precipitation is considered as snow,

where T_m is the daily mean temperature. In the case where $T_m \geq 0^\circ\text{C}$ and $T_m \leq T_c$, the precipitation is considered as a mixture of rain and snow and their proportion is determined as follows:

$$\text{Rain} = \frac{T_m}{T_c} \times P \quad (3.1)$$

$$\text{Snow} = P - \text{Rain} \quad (3.2)$$

where P is total observed precipitation.

3.3.4 Temperature index and distribution of temperature

As discussed above, due to the limitations of data availability, a temperature index or degree-day method was considered to be the most suitable method for melt computation.

Air temperature expressed in degree-days is used in melt computations as an index of the complex energy balance. The simplest and most common expression relating daily melt to the temperature index is:

$$M = D(T_i - T_b) \quad (3.3)$$

where M is the depth of meltwater (mm d^{-1}), D is a degree-day factor ($\text{mm } ^\circ\text{C}^{-1} \text{ d}^{-1}$), T_i is the index air temperature ($^\circ\text{C}$) and T_b is the base temperature (usually 0°C).

In general, daily mean temperature is the most commonly used index of temperature for melt computation. Mean temperatures, T_{mean} , or the number of degree-days are computed using maximum, T_{max} , and minimum temperatures, T_{min} , as:

$$T_i = T_{\text{mean}} = \frac{(T_{\text{max}} + T_{\text{min}})}{2} \quad (3.4)$$

There are several methods of dealing with the index temperatures used in calculating the degree-day value. As reported by Garstka et al. (1958), sometimes the degree-days from the daily mean temperature do not represent the actual thermal regime over the basin. Such cases have been reported for mountainous areas in many parts of the western USA. Thus, inclusion of minimum temperature at an equal weight with maximum temperature gives higher emphasis to surface heat deficit effect, while maximum temperature alone excludes this effect.

Moreover, air temperature varies with altitude and melt runoff computations need air temperature data for all the elevation bands of a basin to provide a better representation of temperature distribution. However, temperature data is generally recorded for only one station located near the terminus of the glacier. Therefore, temperatures were extrapolated or interpolated to the mid-elevation of each elevation zone using a predefined temperature lapse rate in the model, given by:

$$T_{ij} = T_{i,\text{base}} - \delta(h_j - h_{\text{base}}) \quad (3.5)$$

where, T_{ij} is the daily mean temperature on the i th day in the j th zone ($^\circ\text{C}$); $T_{i,\text{base}}$ is the daily mean temperature ($^\circ\text{C}$) on the i th day at the base station, h_j is the zonal hypsometric mean elevation (m), h_{base} is the elevation of the base station (m), and δ is the temperature lapse rate ($^\circ\text{C}/100 \text{ m}$). For the Himalayan basins, temperatures were lapsed at $0.60^\circ\text{C}/100$

m to the mean hypsometric elevation of different elevation zones for melt computation (Singh & Jain, 2002)

3.3.5 Degree-day factor

The degree-day factor (D) is used to convert the degree-days into melt expressed in depth of water. This factor is influenced by the physical properties of the snow/ice and, because these properties change with time, this factor also changes with time. However, this factor still has practical importance for accurate ablation estimation using limited data (Singh & Singh, 2001; Hock, 2003). It is understood that, while using the temperature index model, a seasonally changing degree-day factor should also be taken into account.

3.3.6 Estimation of Runoff Components

Each component of runoff was computed for each elevation zone separately and then output from all the zones was integrated to provide the total runoff from the basin. Details of the methodology adopted for estimating different components of streamflow are discussed below.

3.3.7 Surface runoff from Glacierized area

The surface runoff generated from the glacierized part of the basin can be categorized in three parts, namely: (a) melt caused due to prevailing air temperatures; (b) under rainy conditions, melt due to heat transferred to the ice surface from rain; and (c) runoff generated from the rain falling over the glacierized area. Melt runoff for each elevation zone of the basin was computed using the degree-day approach and the extent of glacier-covered area (GCA) in that zone.

$$M_{g,i,j} = C_{g,i,j} D_{i,j} T_{i,j} G_{c,i,j} \quad (3.6)$$

where $M_{g,i,j}$ represents glacier melt in terms of depth of water (mm d^{-1}), $C_{g,i,j}$ is the runoff coefficient for glacier melt, $D_{i,j}$ is the degree-day factor ($\text{mm } ^\circ\text{C}^{-1} \text{ d}^{-1}$), $T_{i,j}$ is the index temperature ($^\circ\text{C}$) and $G_{c,i,j}$ represents the ratio of GCA to the total area of the elevation band. The suffixes i and j denote day and zone, respectively.

3.3.8 Runoff depth due to melt from rain falling on the glacier

The depth of melt caused by rain in an elevation zone is given by:

$$M_{r,i,j} = \left(\frac{4.2 T_{i,j} P_{i,j} G_{c,i,j}}{325} \right) \quad (3.7)$$

where $M_{r,i,j}$ is the melt in terms of depth of water due to rain on glacier (mm d^{-1}), $T_{i,j}$ is the temperature of the rain ($^{\circ}\text{C}$), $P_{i,j}$ is the depth of the rain (mm d^{-1}), and $G_{i,j}$ is the depth of rainfall on the glacier (mm d^{-1}). Rainfall events occurring at higher temperatures would cause melting due to rain; otherwise this component would not be so significant.

The runoff depth from rain falling over the GCA, $R_{g,i,j}$, for each zone is given by:

$$R_{g,i,j} = C_{g,i,j} P_{i,j} G_{c,i,j} \quad (3.8)$$

It is to be noted that, for the computation of runoff from rain, the coefficient C_g is used (not C_r , rainfall runoff coefficient), because the runoff from the rainfall falling on the GCA responds in the same way as the runoff from the melting of the glacier.

The daily total discharge from the GCA, Q_{GCA} , is computed by adding contributions from each elevation zone. Thus, discharge from the GCA for all the zones is given by

$$Q_{GCA} = \alpha \sum_{j=1}^n (M_{g,i,j} + M_{r,i,j} + R_{g,i,j}) A_{GCA,i,j} \quad (3.9)$$

where n represents the total number of zones, $A_{GCA,i,j}$ is the glacier-covered area (km^2), and α is a conversion factor ($1000/86400$) used to convert runoff depth (mm d^{-1}) into discharge ($\text{m}^3 \text{s}^{-1}$).

3.3.9 Surface runoff from the glacier-free area

The source of surface runoff from the glacier-free area (GFA) is only rainfall. As for the melt runoff computations, runoff from the GFA, $R_{f,i,j}$ was also computed for each zone using:

$$R_{f,i,j} = C_{r,i,j} P_{i,j} G_{f,i,j} \quad (3.10)$$

Where $C_{r,i,j}$ is the coefficient of runoff for rain and $G_{f,i,j}$ is ratio of GFA to the total area of j th zone on the i th day. Because GCA and GFA are complementary, $G_{f,i,j}$ can be directly calculated as $1 - G_{c,i,j}$. The total runoff from the GFA for all the zones is thus given by:

$$Q_{GFA} = \alpha \sum_{j=1}^n (R_{f,i,j} A_{GFA,i,j}) \quad (3.11)$$

where, $A_{GFA,i,j}$ is the glacier-free area in the j th zone on the i th day.

3.3.10 Estimation of subsurface runoff

The subsurface flow or the baseflow represents the runoff from the saturated zone (subsurface) of the basin to the streamflow. The direct surface runoff having been accounted for from the melt and rainfall, the remaining water contributes to the groundwater storage through infiltration and appears at the outlet of the basin after much delay as subsurface flow or baseflow. The model deals with the hydrological processes separately for the GCA and GFA. No evaporation losses were considered from the GCA, whereas such losses were only considered from the GFA where some moisture is retained when rainfall occurs in this area. The depletion of soil water storage takes place due to evapotranspiration and percolation of water to the deep groundwater zone. It is assumed that 50% of the water retained in the soil percolates down to shallow groundwater and contributes to baseflow, while the remaining 50% accounts for loss from the basin in the form of evapotranspiration and percolation to the deep groundwater aquifer, which may appear further downstream or becomes part of deep inactive groundwater storage. The depth of runoff contributing to baseflow from each zone, $R_{b,i,j}$, is given by:

$$R_{b,i,j} = \beta \left[(1 - C_{r,i,j}) R_{f,i,j} + (1 - C_{g,i,j}) M_{t,i,j} \right] \quad (3.12)$$

Where $M_{t,i,j} = M_{g,i,j} + M_{r,i,j} + R_{g,i,j}$ represents the total input to a particular zone from different sources and β is a coefficient ($=0.50$).

The subsurface runoff was computed by multiplying the depth with conversion factor α and area, and given as follows

$$Q_b = \alpha \sum_{j=1}^n (R_{b,i,j} A_{i,j}) \quad (3.13)$$

where $A_{i,j}$ is total area of zone j on the i th day and represents the sum of $A_{GCA,i,j}$ and $A_{GFA,i,j}$. This component is also routed separately before being added to the other components of discharge from the melt and rainfall.

3.3.11 Routing of Different Components of Runoff

Routing of runoff from the GCA and GFA was done separately because the hydrological responses from these areas are different. The linear cascade reservoir approach was used

for routing. Considering n reservoirs in a series, the outflow from the second reservoir becomes the inflow to the third reservoir, etc. The outflow from the n th reservoir represents the response of the basin in, terms of outflow. As mentioned, routing of different components of streamflow has been done separately and then the total outflow from the basin was computed by summing the different routed components of runoff. The model routes the subsurface runoff similarly to surface runoff with a given value of subsurface storage coefficient, k_b , which is determined using streamflow records of the recession period. In order to determine the storage coefficient for the baseflow, the streamflow of the recession period was plotted against time on semi-log paper and a straight line was fitted. (Haritashya et al., 2006b).

3.3.12 Total streamflow

The daily total streamflow emerging from the basin is calculated by adding the different routed components of discharge for each day:

$$Q = Q_{GCA} + Q_{GFA} + Q_b \quad (3.14)$$

The direct surface runoff results from the overland or near surface flow, while the baseflow is regarded as the contribution from the water stored in the groundwater reservoir to the streamflow. In order to consider the soil moisture deficit, soil moisture index, SMI, which represents the soil moisture deficit in the basin at the beginning of simulation, was considered. The contribution to baseflow only starts after saturation of the topsoil.

3.4 HBV-Light Model

The HBV (Hydrologiska-Bryans-Vattenbalansavdelning) model, (Bergström, 1992) is a conceptual hydrological runoff model developed by the SMHI (Swedish Meteorological and Hydrological Institute) simulates daily discharge using daily rainfall and temperature, and monthly estimates of potential evaporation as input. The model consists of different routines, e.g. the snowmelt is computed by a degree-day method, the groundwater recharge and the actual evaporation are functions of actual water storage in a soil box, runoff formation is represented by three linear reservoir equations and channel routing by a triangular weighting function (Seibert, 1998). All model parameters are described briefly in Fig.3.2 and Table 3.1. During the last 20 years, the HBV-model

(Bergström 1976) has become widely used for runoff simulations in Sweden (Bergström, 1990; 1992). This model is an example of semi distributed conceptual model.

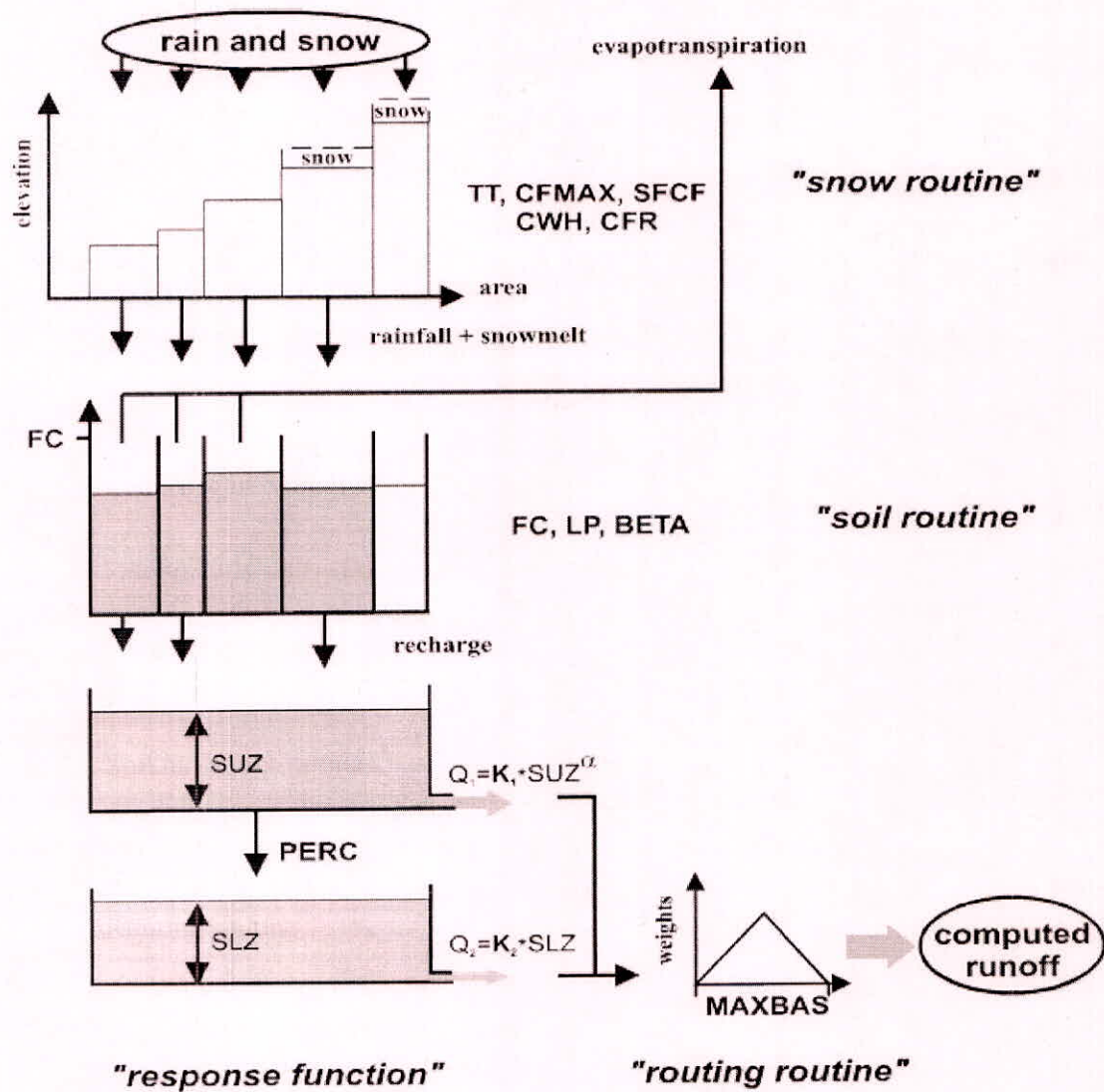


Fig. 3.2: General structure of HBV Model.

3.4.1 Structure of Model Routine

HBV Model mainly consist of five basic routines :

1. Snow routine
2. Soil moisture routine
3. Response function
4. Routing routine

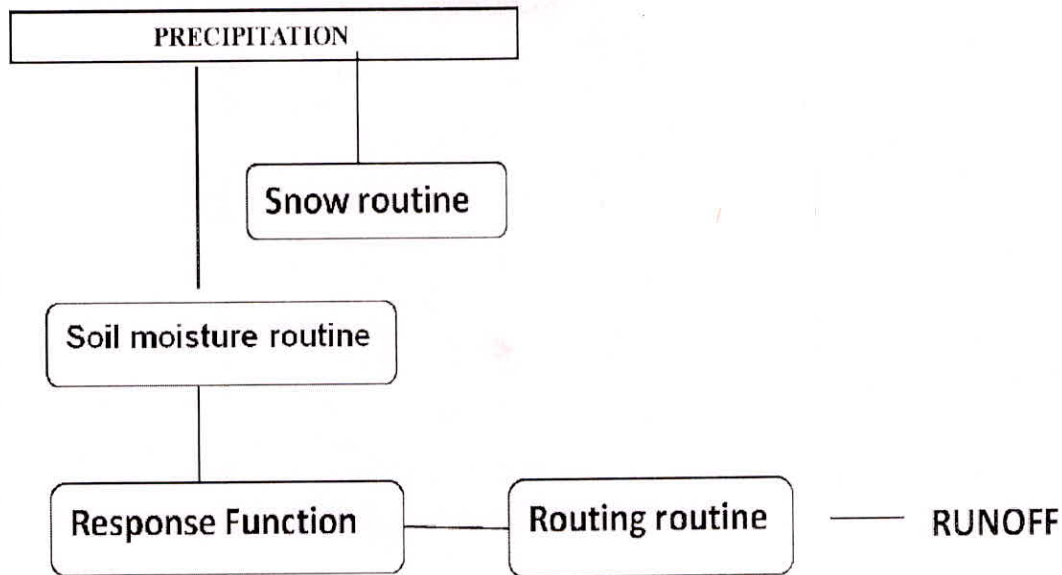


Fig. 3.3: Schematic structure of the HBV model

Table 3.1: Input and output for corresponding sub-models.

Sl. No.	Sub-model	Input data	Output data
1.	Snow routine	Precipitation, Temperature	Snow pack, snow-melt
2.	Soil routine	Potential evaporation, precipitation, snowmelt	Actual evaporation, 'soil moisture', groundwater recharge
3.	Glacier Routine	Temperature, Snowmelt, Elevation Range, Catchment Area	Glacier mass balance, Water Content of the glacier, Glacier outflow
4.	Response function	Groundwater recharge,(pot.ET)	Runoff 'groundwater level'
5.	Routing routine	Runoff	Simulated runoff

3.4.2 Snow Routine

Precipitation is simulated to be either snow or rain depending on whether the temperature is above or below a threshold temperature, TT [$^{\circ}\text{C}$]. All precipitation simulated to be snow, i.e. falling when the temperature is below TT , is multiplied by a snowfall correction factor, SFCF [-]. Melt of snow starts if temperatures are above T_T calculated with a simple degree-day method (Equation 3.15)

$$\text{meltwater} = CFMAX (T - TT) \text{ (mm day}^{-1}\text{)} \quad (3.15)$$

$CFMAX$ varies normally between 2.5–9.0 $\text{mm } ^{\circ}\text{C}^{-1} \text{ d}^{-1}$ for Gangotri Glacier region (Singh et al., 2008). The snow pack retains melt water until the amount exceeds a certain portion (CWH , usually 0.1) of the water equivalent of the snow pack. When

temperatures decrease below TT this melt water refreezes again and given by equation (3.16)

$$\text{Refreezing meltwater} = \text{CFR} * \text{CFMAX} (TT - T) \quad (3.16)$$

Table 3.2: Description of Parameters used in Snow Routine.

Symbol	Description	Unit	Default value	Range
TT	Threshold temperature	°C	0	(- inf, inf)
CFMAX	Degree-day factor	mm/Δt °C	3	[0,inf)
SFCF	Snowfall correction factor		1	[0,inf)
CFR	Refreezing coefficient		.05	[0,inf)
CWH	Water holding capacity		.1	[0,inf)

If different elevation zones are used the changes precipitation and temperature with elevation are calculated using the two parameters *PCALT* [%/100 m] and *TCALT* [°C / 100 m] (Equation 3.17 and 3.18).

$$P(h) = P_0 \left(1 + \frac{PCALT(h - h_0)}{10000} \right) \quad (3.17)$$

$$T(h) = T_0 - \frac{TCALT(h - h_0)}{100} \quad (3.18)$$

Where *PCALT* and *TCALT* are change of precipitation and temperature with elevation respectively. To interpolate the meteorological parameters the reference height (h_0) is considered as 3800m

3.4.3 Soil Moisture Routine

Table 3.3: Description of Parameters used in Soil Moisture Routine.

Symbol	Description	Unit	Default value	Range
FC	maximum soil moisture storage Soil	(mm)	200	(0, inf)
LP	soil moisture value above which AET reaches PET	(mm)	1	[0, 1]
BETA	parameter that determines the relative contribution to runoff from rain or snowmelt	-	1	(0, inf)

Rainfall and snowmelt (P) are divided into water filling the soil box and groundwater recharge depending on the relation between water content of the soil box (SM [mm]) and its largest value (FC [mm]) (Equation 3.19).

$$\frac{recharge}{P(t)} = \left(\frac{SM(t)}{FC} \right)^{BETA} \quad (3.19)$$

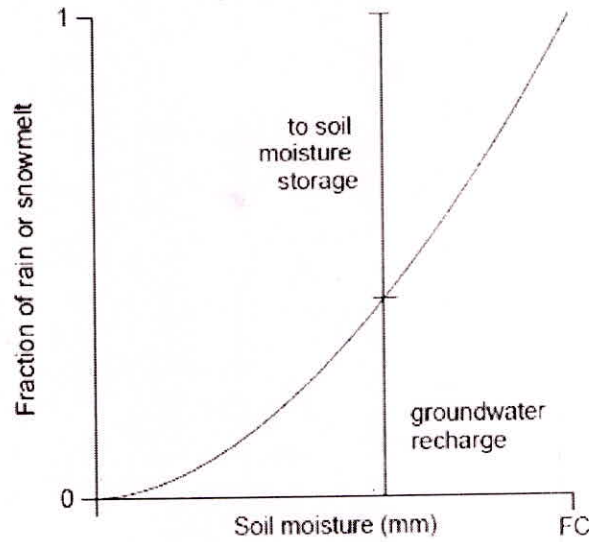


Fig. 3.4: Contributions from rainfall or snowmelt to the soil moisture storage and to the upper groundwater zone.

Actual evaporation from the soil box equals the potential evaporation if SM/FC is above LP [-] while a linear reduction is used when SM/FC is below LP (Equation 3.20).

$$E_{act} = E_{pot} \min \left(\frac{SM(E)}{FC.LP}, 1 \right) \quad (3.20)$$

The long-term mean of the potential evaporation, $E_{pot, M}$ for a certain day of the year can be corrected to its value at day t , $E_{pot}(t)$, by using the deviations of the temperature, $T(t)$, from its long-term mean, T_M , and a correction factor, C_{ET} [$^{\circ}C^{-1}$] (Equation 3.21)

$$E_{pot}(t) = (1 + C_{ET}(T(t) - T_M)) E_{pot, M} \quad (3.21)$$

$$\text{but } 0 \leq E_{pot}(t) \leq 2E_{pot, M}$$

FC is a model parameter and not necessarily equal to measured values of 'field capacity'.

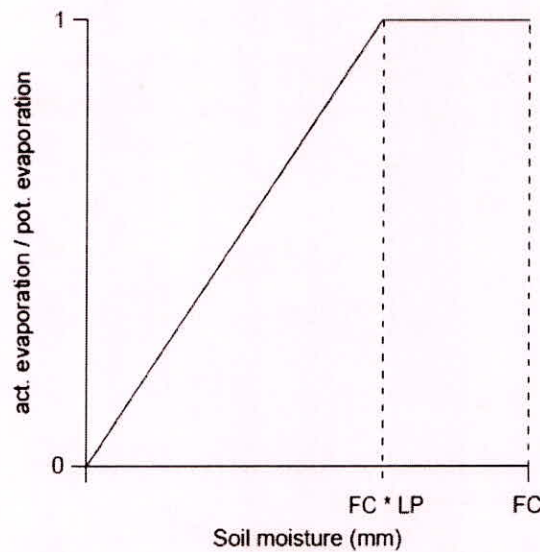


Fig. 3.5: Reduction of potential evaporation depending on the soil moisture storage.

3.4.4 Glacier Routine

Glacial melt is computed using the degree-day approach, and added to the water content of the glacier. A small fraction (KSI) of the simulated snow storage is simulated to become glacial ice each time step. The glacial water storage-outflow relationship varies over time to represent the seasonal development of the subglacial drainage system (Stahl et al., 2008)

$$Q_g = (KG_{\min} + dKG * e^{(-AG * SWE)}) * S \quad (3.22)$$

SWE is the water equivalent of the snowpack on top of the glacier, S the water content of the glacier.

Table 3.4: Description of Parameters used in Glacier Routine.

Symbol	Description	Unit	Default value	Range
CFGlacier	glacier correction factor	-	1	[0,inf)
CFSlope	slope correction factor	-	1	(0,inf)
KGmin	Minimum outflow	t ⁻¹		
dKG	Maximum minus minimum outflow coefficient	t ⁻¹		
AG	Calibration parameter	(mm ⁻¹)		
KSI	Snow to Ice conversion factor	(Δt ⁻¹)		

The degree day factor CFMAX is multiplied by a glacier correction factor (CFGlacier), and multiplied / divided by a slope correction factor (CFSlope) for southern / northern slopes, respectively. These calculations are carried out separately for each elevation and vegetation zone.

3.4.5 Response Function

Simple linear reservoir

The model of a single linear reservoir is a simple description of a catchment where the runoff $Q(t)$ at time t is supposed to be proportional to the water storage $S(t)$.

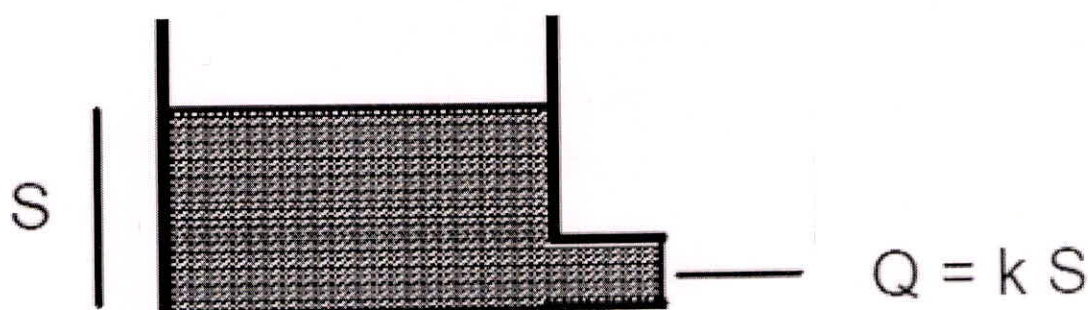


Fig. 3.6: Schematic diagram showing Simple linear reservoir

$$Q(t) = k \cdot S(t) \quad (*) \quad (3.23)$$

S = storage (mm)

Q = outflow (mm day⁻¹)

t = time (day)

k = storage (or recession) coefficient (day⁻¹)

The water balance of the catchment is -

$$P(t) = E(t) + Q(t) + \frac{dS(t)}{dt} \quad (3.24)$$

Ignoring precipitation and evapotranspiration gives -

$$0 = Q(t) + \frac{dS(t)}{dt} \quad (3.25)$$

and together with (*) the differential equation

$$0 = \frac{d}{dt} \left(\frac{Q(t)}{k} \right) + Q(t) \quad (3.26)$$

with the solution function

$$Q(t) = Q(t) \cdot e^{(t-t)k} \quad (3.27)$$

Table 3.5: Description of Parameters used in Response Function.

Symbol	Description	Unit	Default value	Range
Recharge	input from soil routine	mm day ⁻¹		
PERC	threshold parameter	mm Δt ⁻¹	1	[0,inf)
UZL	threshold parameter	mm	20	[0,inf)
SUZ	storage in upper zone	mm		
SLZ	storage in lower zone	mm		
K0	recession coefficient	Δt ⁻¹	0.2	[0,1)
K1	recession coefficient	Δt ⁻¹	0.1	[0,1)
K2	recession coefficient	Δt ⁻¹	0.05	[0,1)

Groundwater recharge is added to the upper groundwater box (SUZ [mm]). PERC [mm d⁻¹] defines the maximum percolation rate from the upper to the lower groundwater box (SLZ [mm]). Runoff from the groundwater boxes is computed as the sum of two or three linear outflow equations depending on whether SUZ is above a threshold value, UZL [mm], or not (Equation 3.18).

$$Q_{GW}(t) = K_2 SLZ + K_1 SUZ + K_0 \max(SUZ - UZL, 0) \quad (3.28)$$

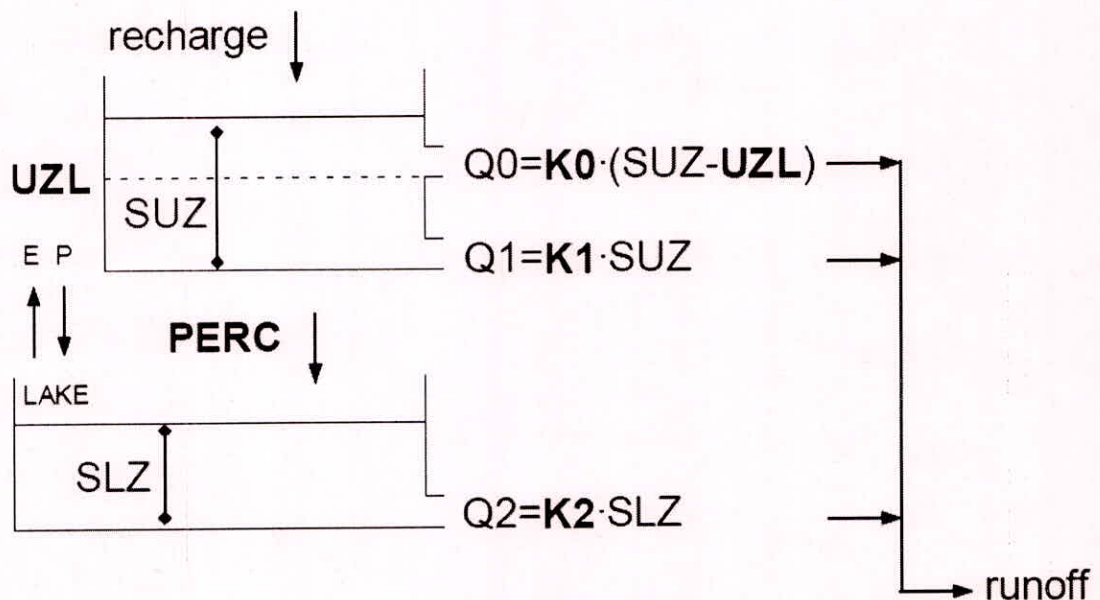


Fig. 3.7: Response function.

NOTE :

1. SUZ has no upper limit.
2. Q_2 can never exceed PERC.
3. SLZ can never exceed PERC/K2.

3.4.6 Routing Routine

Transformation function

This runoff is finally transformed by a triangular weighting function defined by the parameter MAXBAS which determines the base in a equilateral triangular weighting function (Equation 3.29), to give the simulated runoff [mm d-1].

$$Q_{sim}(t) = \sum_{i=1}^{MAXBAS} c(i) Q_{GW}(t-i+1) \quad (3.29)$$

where,

$$c(i) = \int_{i-1}^i \frac{2}{MAXBAS} \left| 4 - \frac{MAXBAS}{2} \right| \frac{4}{MAXBAS^2} du$$

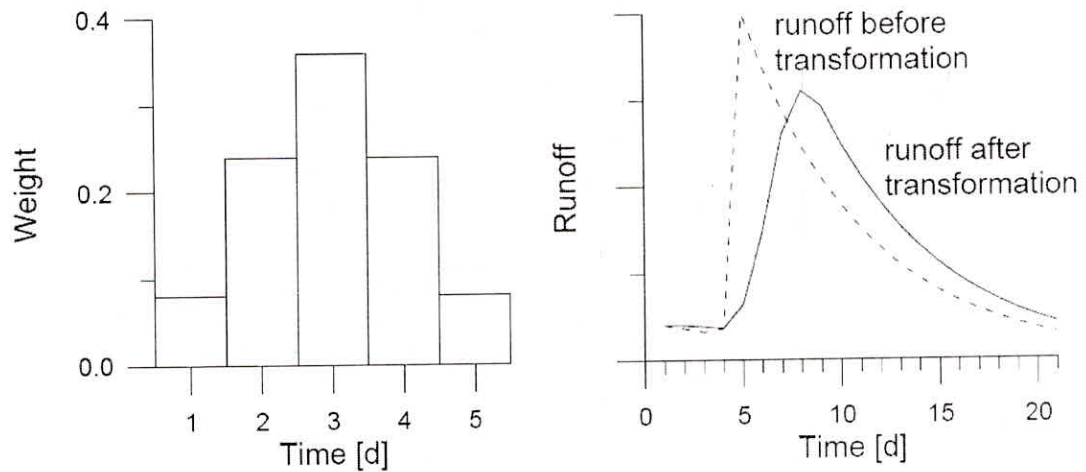


Fig. 3.8: Transformation function.

3.4.7 Model Variants

a. Basic Model

When the basic model variant is selected, the area of each individual Elevation Vegetation Unit (EVU) has to be specified. Each EVU is uniquely defined by its vegetation zone parameters and its elevation.

b. Aspect Model

When the aspect model variant is selected, the area of each individual Elevation Vegetation Unit (EVU) is subdivided into three parts:

- ☐ North orientated
- ☐ South orientated
- ☐ East / West orientated

For each vegetation zone, a slope correction factor (CFSlope) has to be specified. For the North slope, the degree- Δt factor CFMAX will be corrected by dividing its value by CFSlope. For the South slope, the degree- Δt factor CFMAX will be corrected by multiplying its value with CFSlope. For the East/West slope no correction will be applied to CFMAX.

c. Glacier Model

When the glacier model variant is selected, the area of each individual Elevation Vegetation Unit (EVU) is subdivided into three parts:

- i) North orientated
- ii) South orientated
- iii) East / West orientated

Furthermore, each EVU can be categorized as glacier or non-glacier area. For each vegetation zone, a slope correction factor (CFSlope) and a glacier correction factor (CFGlacier) have to be specified. For the North slope, the degree- Δt factor CFMAX will be corrected by dividing its value by CFSlope. For the South slope, the degree- Δt factor CFMAX will be corrected by multiplying its value with CFSlope. For the East / West slope no correction will be applied to CFMAX. In between the snow routine and the soil routine, the glacier routine will be applied to all EUV's categorized as glacier area. The glacier routine computes the melt of the glacier and the modification of snow into ice. The glacier routine uses the CFMAX value (Table 3.7) multiplied by CFGlacier as degree- Δt factor.

3.4.8 Model calibration

The calibration of the model is usually made by manual try and error technique (Bergström, 1992). Different criteria can be used to assess the fit of simulated runoff to observed runoff – i) visual inspection of plots with Q_{sim} and Q_{obs} ii) accumulated difference and, iii) statistical criteria

The coefficient of efficiency, R_{eff} , is normally used for assessment of simulations by the HBV model.

$$R_{eff} = 1 - \frac{\sum (Q_{Sim}(t) - Q_{Obs}(t))^2}{\sum (Q_{Obs}(t) - \bar{Q}_{Obs}(t))^2} \quad (3.30)$$

R_{eff} compares the prediction by the model with the simplest possible prediction, a constant value of the observed mean value over the entire period.

$R_{eff} = 1$ Perfect fit, $Q_{Sim}(t) = Q_{Obs}(t)$

$R_{eff} = 0$ Simulation as good (or poor) as the constant-value prediction

$R_{eff} < 0$ Very poor fit

3.4.9 Catchment Parameters

In the 'Catchment parameters' groupbox, lower and upper bound values can be specified for each of the catchment parameters. For each run during the Monte Carlo simulation, parameter values are randomly chosen within the given range. When the Gaussian random number option is selected, mean and standard deviation values can be specified as well. In case there is more than one subcatchment, one can specify in the column containing radio buttons if random values should be generated for each of the subcatchments separately (R), or if the parameter value should be equal for all subcatchments (=). Also, if there are multiple subcatchments, values can be specified per subcatchment by selecting the checkbox in the 'C' column. If the checkbox is selected, the 'Edit' button appears, which can be used to open the Monte Carlo Parameter form.

Table 3.6: Detailed description of catchment parameters used in the HBV model.

Name	Unit	Valid range	Default value	Description
PERC	mm/ Δt	[0,inf)	1	Threshold Parameter
Alpha	-	[0,inf)	0	non-linearity coefficient
UZL	mm	[0,inf)	20	threshold parameter
K0	1/ Δt	[0,1)	0.2	storage (or recession) coefficient 0
K1	1/ Δt	[0,1)	0.1	storage (or recession) coefficient 1
K2	1/ Δt	[0,1)	0.05	storage (or recession) coefficient 2
MAXBAS	Δt	[1,100]	1	length of triangular weighting function
Cet	1/ $^{\circ}\text{C}$	[0,1]	0	potential evaporation correction factor
PCALT	%/100m	(-inf,inf)	10	change of precipitation with elevation
TCALT	$^{\circ}\text{C}/100\text{m}$	(-inf,inf)	0.6	change of temperature with elevation
Pelev	m	(-inf,inf)	0	elevation of precipitation data in the PTQ file
Telev	m	(-inf,inf)	0	elevation of temperature data in the PTQ file
PART	-	[0,1]	0.5	portion of the recharge which is added to groundwater box 1
DELAY	Δt	[0,inf)	1	time period over which recharge is evenly distributed

3.4.10 Vegetation Zone Parameters

In the 'Vegetation zone parameters' group box, lower and upper bound values can be specified for each of the vegetation zone parameters. For each run during the Monte Carlo simulation, parameter values are randomly chosen within the given range. When the Gaussian random number option is selected, mean and standard deviation values can be specified as well. In case there is more than one sub catchment, it can be specified in the column containing two radio buttons per row if random values should be generated for each of the subcatchments separately (R), or if the parameter value should be equal for all subcatchments (=). In case there are multiple vegetation zones, it can be specified in the column containing four radio buttons per row, if a random value should be

generated for each vegetation zone (R), if the value should be equal for all vegetation zones (=), if random values should be in increasing order from the first till the last vegetation zone (<), or if the random values should be in decreasing order (>). If there are multiple sub catchments and/or vegetation zones, values can also be specified per sub catchment and/or vegetation zone by selecting the checkbox in the 'C' and 'V' columns respectively. If at least one of those checkboxes is selected, the 'Edit' button appears, which can be used to open the Monte Carlo Parameter form.

Table 3.7: Detailed description of Vegetation Zone parameters used in the HBV model.

Name	Unit	Valid range	Default value	Description
TT	°C	(-inf, inf)	0	threshold temperature
CFMAX	mm/Δt °C	[0,inf)	3	degree-Δt factor
SFCF	-	[0,inf)	1	snowfall correction factor
CFR	-	[0,inf)	0.05	refreezing coefficient
CWH	-	[0,inf)	0.1	water holding capacity
CFGlacier	-	[0,inf)	1	glacier correction factor
CFSlope	-	(0,inf)	1	slope correction factor
FC	mm	(0,inf)	200	maximum soil moisture storage
LP	-	[0,1]	1	soil moisture value above which AET reaches PET
BETA	-	(0,inf)	1	Soil parameter that determines the relative contribution to runoff from rain or snowmelt

3.4.11 Input data

The model simulates daily discharge using daily rainfall, temperature and potential evaporation as input.

Precipitation

The areal average precipitation p_{area} is calculated as weighted mean of precipitation stations in and around the catchment.

$$p_{area} = \sum c_i p_i \quad (3.31)$$

The catchment can be divided into different elevation zones. For each zone the precipitation will be corrected according to the its increase with elevation above sea level (usually 10-20% per 100 m, parameter PCALT).

$$P(h) = P \left(1 + \frac{PCALT(h-h_e)}{10000} \right) \quad (3.32)$$

Temperature

Temperature data is needed in catchments with snow and is calculated as weighted mean of stations in and around the catchment. When different elevation zones are used temperature will be corrected for elevation above sea level with usually -0.6°C per 100 m (parameter TCALT).

$$T(h) = T_e - \frac{TCALT(h-h_e)}{100} \quad (3.33)$$

Potential evaporation

Estimates of the potential evaporation may be provided by calculations using, for instance, the Penman formula or measurements by evaporimeters. Normally, monthly mean values are assumed to be sufficient. The long-term mean evaporation can be corrected by using the deviations of the temperature from its long-term mean (Equation 3.24)

$$E_{pot}(t) = (1 + C_{ET}(T(t) - T_M)) E_{pot,M} \quad (3.34)$$

$$\text{but } 0 \leq E_{pot}(t) \leq 2 E_{pot,M}$$

$E_{pot}(t)$ potential evaporation at day t [mm d⁻¹]

C_{ET} correction factor [°C⁻¹]

$T(t)$ temperature at day t [°C]

T_M long-term mean temperature for this day of the year [°C]

$E_{pot,M}$ long-term mean evaporation for this day of the year [mm d⁻¹]

Chapter – 4

Hydrological Modelling of Streamflow

4.1 Gangotri Group of Glaciers: Geographical Setting

It is estimated that the Himalayan glaciers provide around $8.6 \times 10^6 \text{ m}^3$ of water annually to 1.3 billion of people (Bolch et al., 2012). The three great rivers of India – the Indus, the Ganga and the Brahmaputra collectively provide close to 50% (320 km^3) of the total country's utilizable surface water resources (690 km^3) (Srivastava, 2012). These rivers are originated from the Cryosphere system of Himalaya. Therefore, it is essential to monitor the Cryosphere system of the Himalaya. To study all the glaciers in the Himalaya using ground survey method and instrumental techniques is not possible due to rough terrain and harsh climatic conditions. In the present study, streamflow of Gangotri group of glaciers (source of Bhagirathi River) is studied using semi-distributed conceptual rainfall-runoff model.

The Gangotri Glacier and the River Ganga has important status in the Indian culture. Scriptures and faith of the population consider it a sacred river. Perhaps billion of the Indian people depend on these rivers for their water and food supply. Lot of population is dependent on it for irrigation and drinking water purposes. Because of these reasons it has attained lot of attention. Climate change and global warming has added more concern to the health of the Gangotri glacier and Ganga River. It is one of

the largest glaciers in the country. It is a north-westerly flowing 'compound basin' type valley glacier and occupies longitudinal U-shaped valley. Gangotri catchment area falls in higher Himalayan region. The Gangotri Glacier has an identification number as 50 131 06 029 for glacier inventory purpose and it lies between latitude $30^{\circ}43'20''$ to $31^{\circ}01'07''$ N and longitude $78^{\circ}59'42''$ to $79^{\circ}17'10''$ E (Fig. 4.1).

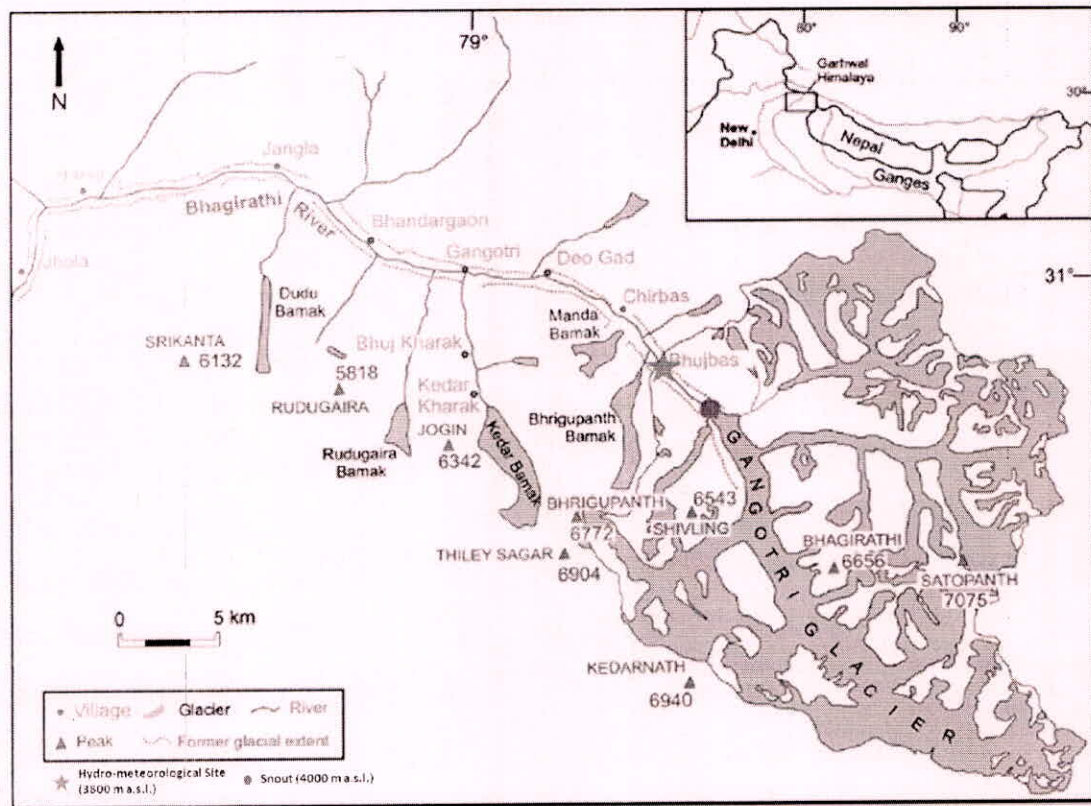


Fig. 4.1: Physiographic map of Gangotri Glacier (Dutta et al., 2004)

It is originating from Chaukhamba peaks (7138 m a.s.l.) is 30.2 km long with a glacierised area of about 143.58 km^2 (Fig. 4.1). The present day snout is at about 18 km from Gangotri Township. Gangotri glacier is one of the most important glaciers in Central Himalaya which is the source of the river Bhagirathi, the largest tributary of Ganges, emanating from the Gangotri group of glaciers (Kirti Bamak, Ghanohim Bamak, Swachhand Bamak, Maiandi Bamak) in Uttarkashi district of Uttarakhand. The other major tributary Alaknanda emanating from the Satopanth and Bhagirathi glaciers north of Badrinath in Chamoli district of Uttarakhand meet at Devprayag and give rise to Ganga river, which is the most important source of fresh water supporting the population of north Indian Himalaya and plains.

The total area of the Gangotri catchment is 549.25 sq km and the perimeter of the same is 222.53 km (Tangri et al., 2004) upto the established water discharge site (3 km downstream of Gaumukh). The Bhagirathi basin has 238 glaciers of different sizes which occupy about 12% of the basin area. Gangotri Glacier is the largest in the basin. The glacier originates from Chaukhamba group of peaks at the elevation of 7138 m and flows for 30.2 km in northern direction with an accumulation zone over a length of 4.2 km and ablation zone length of 26.0 km (Srivastava, 2012). Its valley is nearly 2.15 km wide in the accumulation zone and 1.35 km in ablation zone. The mean width of the glacier is about 1.5 km. The Bhagirathi river valley is nearly a km in width near the snout front. The gradient of the valley in the accumulation zone is 20° – 25° and 5° in the ablation zone. The lower half of the glacier is covered by the supraglacial moraines. The surface of the glacier in general, is undulating due to the presence of crevasses and heaps of morainic materials. The crevasses are either longitudinal, transverse or curve (Tewari, 1972). The glacier characteristics, as for as length, area and orientation wise are given in Table 4.1 (Kaul, 1999).

The mean elevations of accumulation and ablation zone are 6200 m and 4800 m a.s.l., respectively. The total area occupied by the Gangotri Glacier is 143.58 km² wherein 92.48 km² is in the ablation zone and remaining area is occupied by accumulation zone. The established volume of ice is 28.716 km³ considering average thickness of the glacier ice to be 200 m. Accumulation Area Ratio (AAR) is found to be 36%. The snout area in general and the ice cave in particular are popularly known as 'Gaumukh' from where the glacier melt water channel (Bhagirathi River) originates. The glacier has a very steep fall from its head at 7143 m a.s.l. down to an elevation of about 5380 m a.s.l. Thereafter, except for the last 3 km or so where the gradient is of the order of 10° the glacier has a general gradient of about 5° only. In the ablation zone the glacier ice is hardly visible under surface moraines except on the crevasse faces and at the snout. The snout situated at a height of 4000 m consists of a 40 m high ice wall and a large ice cave, Gaumukh (30 m x 20 m), from which the melt water, giving rise to the Bhagirathi River gushes out. Bhagirathi River has cut a deep, narrow, vertical walled gorge within the hard granitic rocks in Gangotri area. The gorge is 40-60 m deep as compared to the main valley and at places, is less than 2 m in width. Gorge is extending upto Jangla (Vohra, 1971).

Table 4.1: Inventory characteristics of glaciers of Bhagirathi basin

Morphological Parameters	Characteristic	Percentage of total
Basin Area (km ²)	6183.32	-
No. of glaciers	238	-
Glacierized area (km ²)	755.47	12.22%
Total Ice Volume (km ³)	67.02	-
Largest glacier	Gangotri Glacier	-
Length (km)	30.20	
Glacier area ((km ²)	143.58	
Glacier Volume (km ³)	28.72	
Orientation		
N	29	12 %
NE	34	14 %
E	23	10 %
SE	31	13 %
S	19	8 %
SW	47	20 %
W	18	8 %
NW	37	15 %
Length wise distribution (km)		
>5	32	14 %
>1	136	57 %
<1	70	29 %
Area wise distribution (km ³)		
>10	13	5 %
>5	17	7 %
>1	61	26 %
<1	147	62%

The glacier is dirty mainly due to ubiquitous supraglacial debris varying in thickness from 10 cm to 2 m, comprising loose sand and unsorted assemblage of small rock fragments to very large boulders. The main rock types within the till are the biotite granite and tourmaline granite with subordinate quantity of gneiss, mica schist and quartzite (Srivastava, 2012). There are numerous crevasses of various sizes all along the glacier.

4.1.1 Tributary Glaciers

The Gangotri glacier is fed by four major tributary glaciers - i) Kirti Bamak ii) Ghanohim Bamak, iii) Swachhand Bamak and iv) Maiandi Bamak. The length and elevations from where the tributary glaciers originate are given in table 4.2. Kirti Bamak and Ghanohim Bamak confluence at 11 and 15 km from its head on left bank whereas Swachhand Bamak and Maiandi Bamak confluence with Gangotri from right bank at a distance of 4.5 and 7.5 km (Table 4.2).

Table 4.2: Length and peaks of tributary glaciers of Gangotri Glacier

S. No.	Tributary Glacier	Length (km)	Peak (m)
1.	Kirti Bamak	7.00	6940 (Kedarnath Peak)
2.	Ghanohim Bamak	6.00	5970 (Mahalaya Peak)
3.	Swachhand Bamak	7.00	6190
4.	Maiandi Bamak	5.00	6380

4.1.2 Gangotri Glacier Area-Altitude Distribution

To interpolate the meteorological parameter at hypsometric scale from discharge site (3800 m a.s.l.) to upper accumulation zone (7000 m a.s.l.) of the Gangotri Glacier, total glacier area is divided into eight elevation zones (E1–E8) (Table 4.3).

Table 4.3: Glacierized and non- Glacierized area covered at different elevation bands (Singh et al., 2008).

Zone	Elevation range (m a.s.l.)	Glacierized area (km ² / %)	Non-glacierized area (km ² / %)	Total zone area (km ²)
E1	3800–4200	2.38 / 0.0	6.49 / 0.01	8.87
E2	4201–4600	14.39 / 0.03	17.66 / 0.03	32.05
E3	4601–5000	52.93 / 0.10	37.72 / 0.07	90.65
E4	5001–5400	65.17 / 0.12	57.55 / 0.10	122.72
E5	5401–5800	72.36 / 0.13	68.43 / 0.12	140.79
E6	5801–6200	53.00 / 0.10	56.35 / 0.10	109.35
E7	6201–6600	21.37 / 0.04	23.85 / 0.04	45.22
E8	6600–7000	4.39 / 0.01	1.79 / 0.0	6.17
Total	3800–7000	285.99 / 0.51	270.24 / 0.49	556.23

These elevation zones are referred as E1 (3800-4200 m a.s.l.), E2 (4200 -4600 m a.s.l.), E3 (4600-5000 m a.s.l.), E4 (5000-5400 m a.s.l.), E5 (5400-5800 m a.s.l.), E6 (5800-6200 m a.s.l.), E7 (6200-6600 m a.s.l.), E8 (6600-7000 m a.s.l.).

4.2 Meteorological Observatory and data collection

For the present study, a standard meteorological observatory (30 m x 30 m) was established at about 3800 m a.s.l. near the snout of the Gangotri Glacier and required data were collected (Fig. 4.2). This observatory was equipped with ordinary rain gauge, self-recording rain gauge, thermograph, maximum and minimum thermometers, dry and wet bulb thermometers, hygrograph, evaporimeter, anemometer, wind vane and sunshine recorder (Singh et al., 2005). Standard timings for data collection were according to the practice followed by the India Meteorological Department (IMD). The observatory is located in the valley floor on right bank of the Bhagirathi River, about 3 km downstream to the snout of the glacier and surrounded by steep sloping hills. The valley floor has been extensively reworked and resedimented by glacial and paraglacial activity. The valley shows NW–SE trend within the granitic terrain. The Indian birch (*Betula utilis*) is the only tree found in and around the observatory.

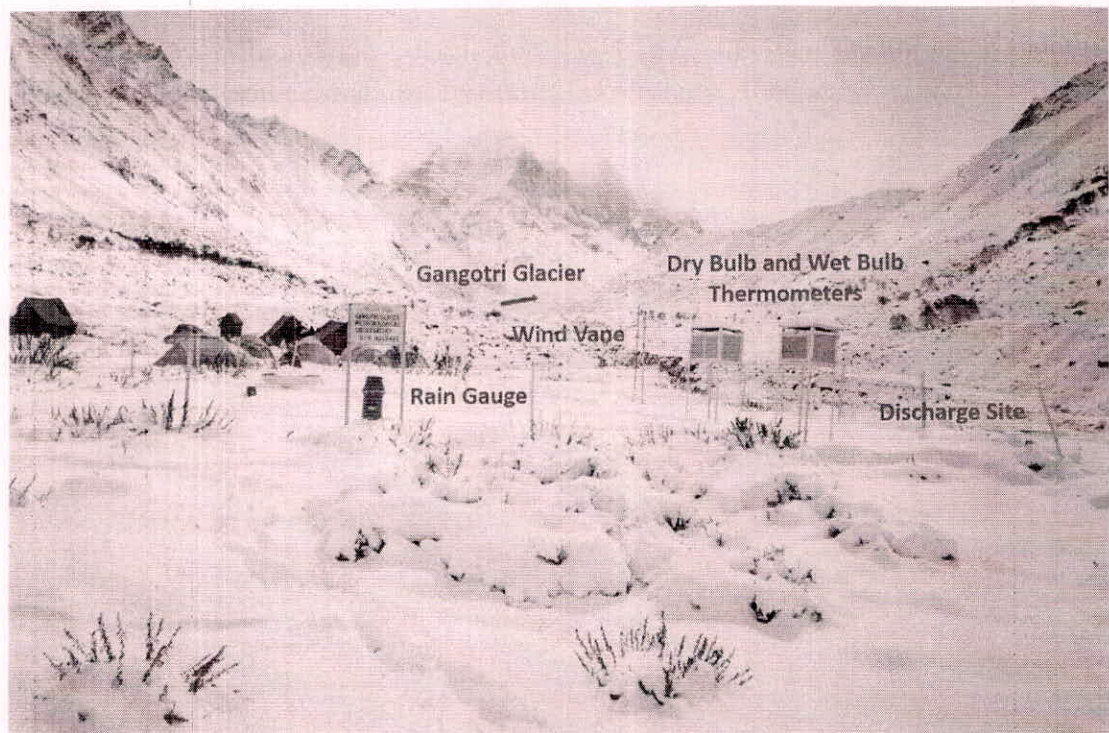


Fig. 4.2: Hydro-meteorological observatory (3800 m a.s.l.) established near the snout of Gangotri Glacier (source: National Institute of Hydrology, Roorkee).

4.3 Hydro-meteorological Conditions: 2000 and 2001

4.3.1 Air temperature

Temperature and Rainfall air represent the most important parameters of meteorology for any particular region (Singh et al., 2005; Kesarwani et al., 2015). They can be used as a climatic indicator of the region. Daily maximum and minimum air temperatures observed near the snout of Gangotri Glacier for different years are shown in graphs. Broadly, the trend of changes in temperature over the summer season is found to be almost similar for both the years, i.e. it follows increasing trend till July and then starts decreasing. It is observed that changes in minimum temperature are more significant than those in maximum temperature.

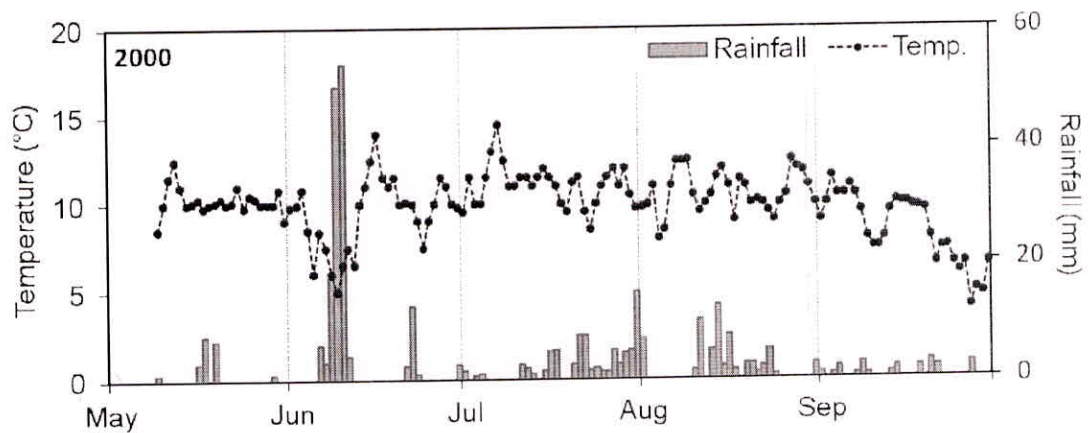


Fig. 4.3: Mean temperature (T) (°C), and rain (R) (mm) observed during various months (May to September) near the terminus of Gangotri Glacier for the melting season 2000. Here, line represents T and bar represents R.

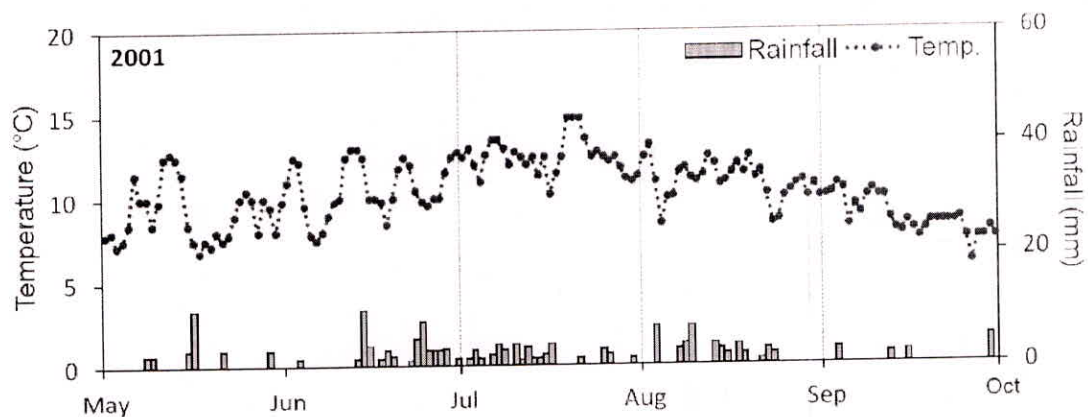


Fig. 4.4: Mean temperature (T) (°C), and rain (R) (mm) observed during various months (May to September) near the terminus of Gangotri Glacier for the melting season 2000. Here, line represents T and bar represents R.

Table 4.4: Analysis of observed hydro-meteorological records of the Gangotri Glacier for the summer season (May-September) of 2000 and 2001.

Parameters	Units	2000					2001				
		May	June	July	August	September	May	June	July	August	September
Max. Temperature	°C	12.5	14	14.5	12.5	11.5	12.8	13	14.8	13.2	10.8
Min. Temperature	°C	8.5	5	8.5	8	4	6.8	7.5	10.2	8.5	6
Avg. Temperature	°C	10.2	9.4	11	10.6	8.4	9.2	10.7	12.5	10.9	8.7
Max. Rainfall	mm day ⁻¹	7.5	54	75.6	12.5	3	10.1	10.2	4	7	5
Avg. Rainfall	mm day ⁻¹	0.8	5.1	2.4	2	0.7	0.7	1.7	1.3	1.3	0.4
Max. Discharge	m ³ s ⁻¹	65	125.5	175	137	100	48	98	175	170	110
Min Discharge	m ³ s ⁻¹	12.5	38	70	60	18	8	40	80	90	39
Avg. Discharge	m ³ s ⁻¹	43.1	72	117	85.7	58.2	24.9	68.4	118.8	121.1	69.5

Over the study period, average daily maximum and minimum temperatures were observed to be 14.7 and 4.1°C respectively, whereas average mean temperature was 9.4°C. As usual, sudden drops in air temperature were observed during rain or snowfall events. For example, during a six-day rainstorm in June 2000, the mean temperature dropped to about 6.4°C, which represents a deviation of -3.0°C from the mean temperature of June 2000 (9.45°C). The maximum, minimum and average temperatures for May to September were 12.5, 14, 14.5, 12.5 and 11.5°C and 8.5, 5.0, 8.5, 8.0, 4.0 °C and 10.2, 9.4, 11.0, 10.6, 8.4 °C for the year 2000 (Fig. 4.2 and Table 4.4) and 12.8, 13.0, 14.8, 13.2, 10.8, and 6.8, 7.5, 10.2, 8.5, 6.0°C and 9.6, 10.7, 12.4, 10.9, 8.7°C for the year 2001 (Fig. 4.3 and Table 4.4).

It reveals that the diurnal temperature range was highest in May and September, and lowest in August. Minimum diurnal temperature range in August is possible due to minimum sunshine hours (or presence of more clouds) in this month. The presence of clouds prevents the shortwave solar radiation in the daytime to heat the surface and longwave radiation in the night-time to emit the heat, resulting in lower temperature range (Kesarwani et al., 2015). The reverse is true for the months of May and September.

4.3.2 Rainfall

Records of daily rainfall observed for the period 2000 and 2001 near the snout of the Gangotri Glacier are shown in Figure 4.5. These results suggest that during the summer season daily rainfall hardly exceeds 15 mm in the study area, except some unusual heavy rainfall events (Singh et al., 2005). During the period of observation, one major storm occurred in the study area. The major storm occurred in June 2000. During the first major storm, rainfall occurred continuously for 6 days (6–11 June 2000) and a flash flood was generated. Rainfall observed at the site for 6–11 June was 6, 3, 17.5, 50, 54 and 4 mm for the year 2000. Rainfall was not much (6.5 mm) on the first two days (6 and 7 June), it was 17.5 on 8 June but it was increased (104 mm) for the subsequent 2 days (9–10 June 2000) providing total rainfall of 330 mm. The maximum rainfall (54 mm) was recorded on 10 June 2000 the day on which the flood occurred for 2000 and (10.2 mm) on 14 June 2001 for 2001. The number of days when there is no rainfall is 89 and 107 and average rainfall was 2.83 mm and 1.10 mm for 2000 and 2001, respectively. Maximum rainfall was 7.5, 5.4,

75.6, 12.5, 3.0 mm and 10.1, 10.2, 4.0, 7.0, 5.0 mm, respectively for 2000 and 2001 (Fig. 4.2, Fig. 4.3 and Table 4.4).

Rainfall observations clearly indicate that although the intensity was higher in May, the number of rainy events was much lower during this month. Whereas, in July when the monsoon approaches, the total number of rain events increased due to drizzle-type of rainfall throughout the day. It also shows that the timings of rainfall occurrence varied over the summer season. At the start of the melt season, the rainfall occurs mostly due to formation of convective clouds. As the melting season advances and monsoon season (July–September) approaches, the frequency of rainfall increased due to a combination of local convective activity and a partial influence of monsoon. It is to be pointed out that most of the moisture in the monsoon clouds precipitates before they reach the high-altitude regions beyond 4000 m a.s.l., like the present study area and, therefore, have only limited influence on rainfall.

Generally, orographic effect on precipitation is found in the high-altitude mountainous regions. Altitude controls both temperature and rainfall (Kesarwani et al., 2015). Mountains act as barriers and force the moisture-bearing winds to ascend, resulting in such effects on precipitation. Precipitation is concentrated on the windward slopes, and a rain shadow is produced on the leeward side (Singh et al., 2005). The summer monsoon in India produces heavy rainfall on the southern slopes of the Himalayas, whereas north Himalayan ranges are deprived of precipitation. Some studies show that rainfall increases with altitude up to about 2500 m a.s.l., and then starts decreasing. Studies related to rainfall characteristics during monsoon season in High Mountain areas of Nepal Himalayas suggest that rainfall decreases with altitude in the range from 2800 to 4500 m a.s.l. It was observed that in the Greater Himalayan range, rainfall decreased exponentially with elevation. Generally, during the summer period a little precipitation (100–300 mm) is observed at high altitudes in the Himalayan region. Rainfall in the Gangotri region was also in the same range. However, orographic effect on rainfall in the study region is yet to be investigated.

4.3.3 Actual Evaporation

Actual Evaporation (AE) is an important component of the hydrological cycle and plays a vital role in studying the hydrology of the basin/region, especially for water balance and modeling of streamflow studies (Singh et al., 2005).

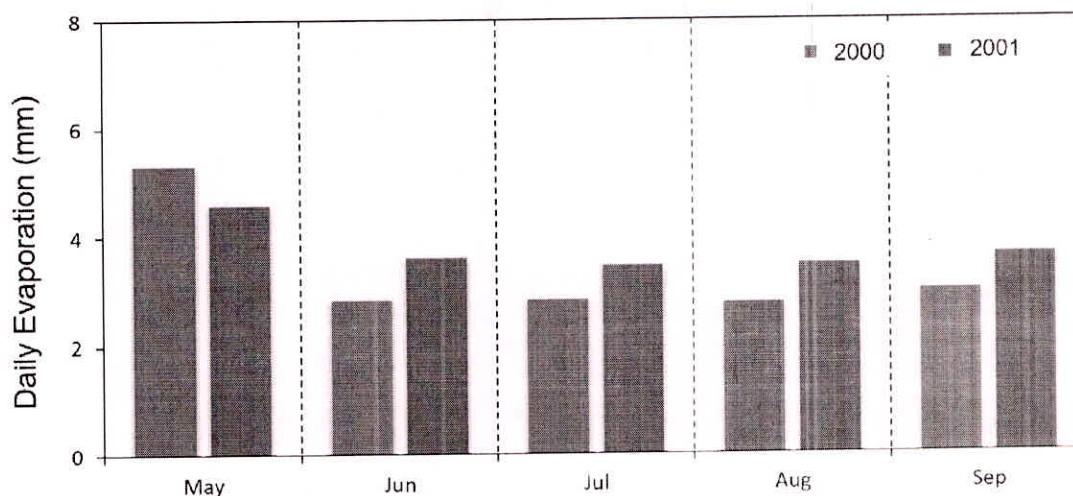


Fig. 4.5: Daily evaporation observed for different months near the snout of Gangotri Glacier (Bhojwasa, 3800 m) during summer seasons 2000–01.

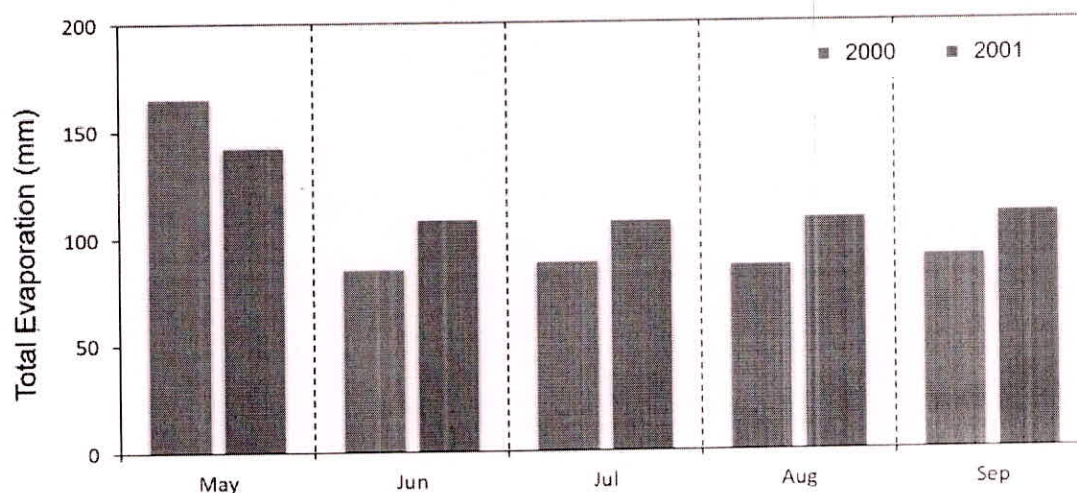


Fig. 4.6: Monthly pan evaporation observed for different months near the snout of Gangotri Glacier (Bhojwasa, 3800 m) during summer seasons (May–September) of 2000 and 2001.

Atmospheric temperature, wind velocity, wind direction and relative humidity are the main meteorological parameters that control evaporation from the basin. Mean daily pan evaporation observed is 5.3, 2.8, 2.8, 2.7 and 3.0 mm for May–September respectively for

the year 2000 and 4.5, 3.6, 3.4, 3.4 and 3.6 mm for corresponding year 2001 (Fig. 4.4, Fig. 4.5 and Table 4.5). Evaporation was maximum in May and minimum in August. Lower relative humidity, high sunshine hours and high wind speed are responsible for higher evaporation observed during May, while low evaporation in August is possible due to less number of sunshine hours and high relative humidity.

Table 4.5: Daily and monthly actual evaporation data at the snout of Gangotri Glacier.

Months	Monthly			Daily	
	2000	2001	2000-2001	2000	2001
	Evaporation (mm)	Evaporation (mm)	Mean Evaporation (mm)	Mean (mm)	Mean (mm)
May	165	142	153.5	5.32	4.58
June	85	108	96.5	2.83	3.6
July	88	107	97.5	2.84	3.45
August	86	108	97	2.77	3.48
September	90	110	100	3	3.67

The one-day maximum value of pan evaporation was observed to be 5.3 mm, while minimum was recorded to be zero. Mean monthly total evaporation during two ablation periods was 153.5, 96.5, 97.5, 97.0, and 100 mm for May–September, respectively (Table 4.4 and Fig. 4.5). Distribution of monthly pan evaporation is shown in Figure 4.5. The total pan evaporation during the summer seasons 2000–01 was 514 and 575 mm respectively. Based on total records, mean daily evaporation for the summer season is found to be 3.5 mm.

4.3.4 Discharge

In hydrology, discharge is the volume rate of water flow that is transported through a given cross-sectional area (Haritashya et al., 2010). Daily discharge observed near the snout of Gangotri Glacier for summer season May to September for two years 2000 and 2001 are shown in Fig. 4.6, Fig. 4.7 and Table 4.4. Broadly, the trend of changes in discharge over the summer season is found to be almost similar for both the years. The monthly maximum discharge for summer season May to September were observed to be 65, 125, 175, 137 and

100 $\text{m}^3 \text{s}^{-1}$ whereas minimum discharge of the streamflow were observed 12, 38, 70, 60 and 18 $\text{m}^3 \text{s}^{-1}$. The average discharge of the streamflow for these months was 43, 72, 117, 85 and 58 $\text{m}^3 \text{s}^{-1}$ for 2000 (Fig. 4.6). Similarly, the monthly maximum discharge for summer season May to September were observed to be 48, 98, 175, 170 and 110 $\text{m}^3 \text{s}^{-1}$ whereas, minimum discharge of the streamflow were observed 8, 40, 80, 90 and 39 $\text{m}^3 \text{s}^{-1}$ (Haritashya et al., 2010). The average discharges of the streamflow for these months were 25, 69, 119, 121 and 70 $\text{m}^3 \text{s}^{-1}$ for 2001 (Fig. 4.7). Results indicate that the study area experienced maximum discharge in the mid of July month.

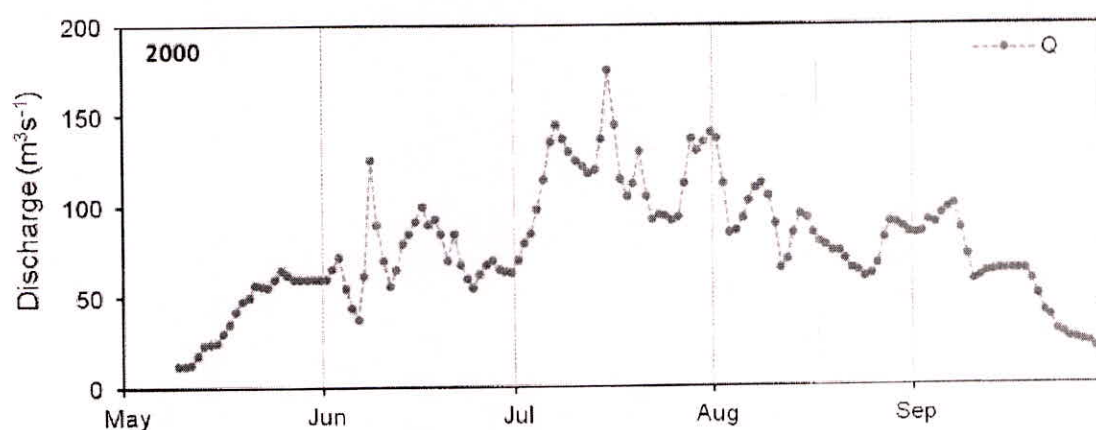


Fig. 4.7: Daily discharge (Q) (m^3/s), observed during various months (May to September) near the terminus of Gangotri Glacier for the melting season 2000.

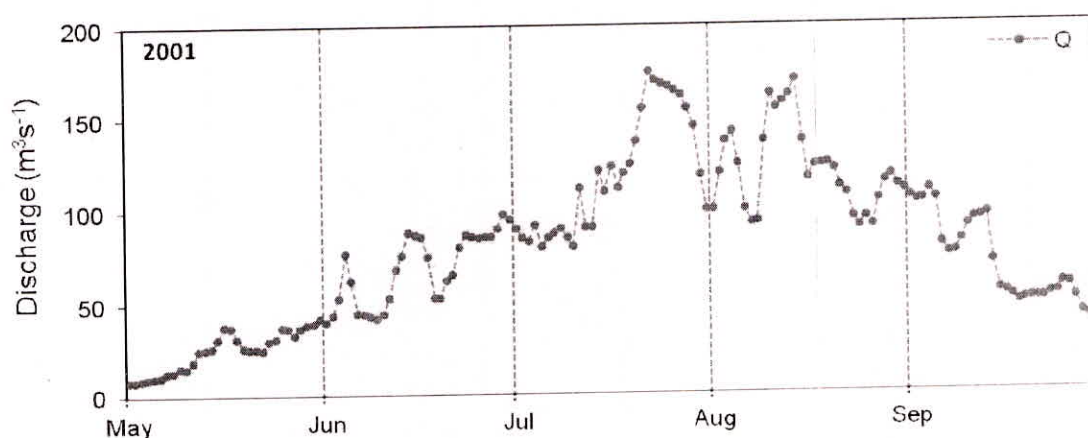


Fig. 4.8: Daily discharge (Q) (m^3/s), observed during various months (May to September) near the terminus of Gangotri Glacier for the melting season 2001.

Chapter - 5

Hydrological Modelling of Streamflow

5.1 Introduction

The lack of hydrological data, especially in mountainous regions, makes the extension of existing records an important task for hydrologists (Devi et al., 2015). One possibility is the use of conceptual rainfall-runoff models. Various runoff models are available in market to estimate the discharge from any basin or catchment (Seibert, 1998). Some examples are Stanford Watershed Model IV (SWM), SHE/ MIKE SHE model, HBV Model, SNOWMOD etc. (Devi et al., 2015). In present study, we are using HBV Model to estimate the discharge from Gangotri basin because it provides different functionalities, such as an automatic calibration using a genetic algorithm or a Monte Carlo approach, as well as the possibility to perform batch runs with predefined model parameters. The HBV model (Bergstrom 1976) has been applied in numerous studies, e.g., to compute hydrological forecasts, for the computation of design floods or for climate change studies (Bergstrom, 1992). The model consists of different routines e.g. the snowmelt is computed by a degree-day method, the groundwater recharge and the actual evaporation are functions of actual water storage in a soil box, runoff formation is represented by three linear reservoir equations and channel routing by a triangular weighting function. and simulates catchment discharge, usually on a daily time step, based on time series of precipitation and air temperature as well as estimates of monthly long-term

potential evaporation rates (Seibert, 1998). In addition, a routine for simulation of glaciers has been also implemented. Also, it gives result with very high efficiency with least requirement of number of meteorological parameters.

5.2 Hydrological Modeling of Bhagirathi River: 2000

5.2.1 Interpolated Meteorological Parameters

For different elevation zones, the changes precipitation and temperature with elevation are calculated using the two parameters PCALT [%/100 m] and TCALT [°C / 100 m] (Equation 5.1 and 5.2).

$$P(h) = P_0 \left(1 + \frac{PCALT(h-h_0)}{10000} \right) \quad (5.1)$$

$$T(h) = T_0 - \frac{TCALT(h-h_0)}{100} \quad (5.2)$$

Where PCALT and TCALT are change in precipitation and temperature with elevation respectively. To interpolate the meteorological parameters the reference height (h_0) is considered as 3800 m. For each elevation monthly average temperature, precipitation (in the form of rainfall and snow) are simulated.

Average temperature is simulated from E1 to E8 band for Summer season vary from 7.8 to -9.0°C, 7.0 to -9.8°C, 8.6 to -8.2°C, 8.2 to -8.6°C and 6.0 to -10.8°C from May to September, respectively (Table 5.1). Similarly, average rainfall is simulated from E1 to E8 band for Summer season vary from 1.2 to 2.0 mm, 1.5 to 9.2 mm, 1.1 to 5.5 mm, 0.2 to 4.5 mm and 0.6 to 1.5 mm May to September, respectively (Table 5.1).

5.2.2 Amount of water in soil box

The majority of models dealing with soil water distribution and transport can be said to belong to one of two different categories. The simplest approach is found in the models usually described as 'conceptual'. Soil water storage in those models is represented as one or more boxes (Singh, 1995; Wilby, 1997). The other common modeling strategy is to divide the soil into many layers and calculate water flow between the layers by solving for example mass and energy transfer equations for all layers. This type of models is often classified as 'physically based' since they are based on physical laws.

Amount of water in soil box (denoted by SM) is highly sensitive to snowfall in comparison to rainfall. For summer season May to September in 2000, within a month SM is nearly constant for all elevation zones as very less snowfall is simulated by model in different elevation E1-E8 (3800-7000) (Fig. 5.1 and Table 5.1).

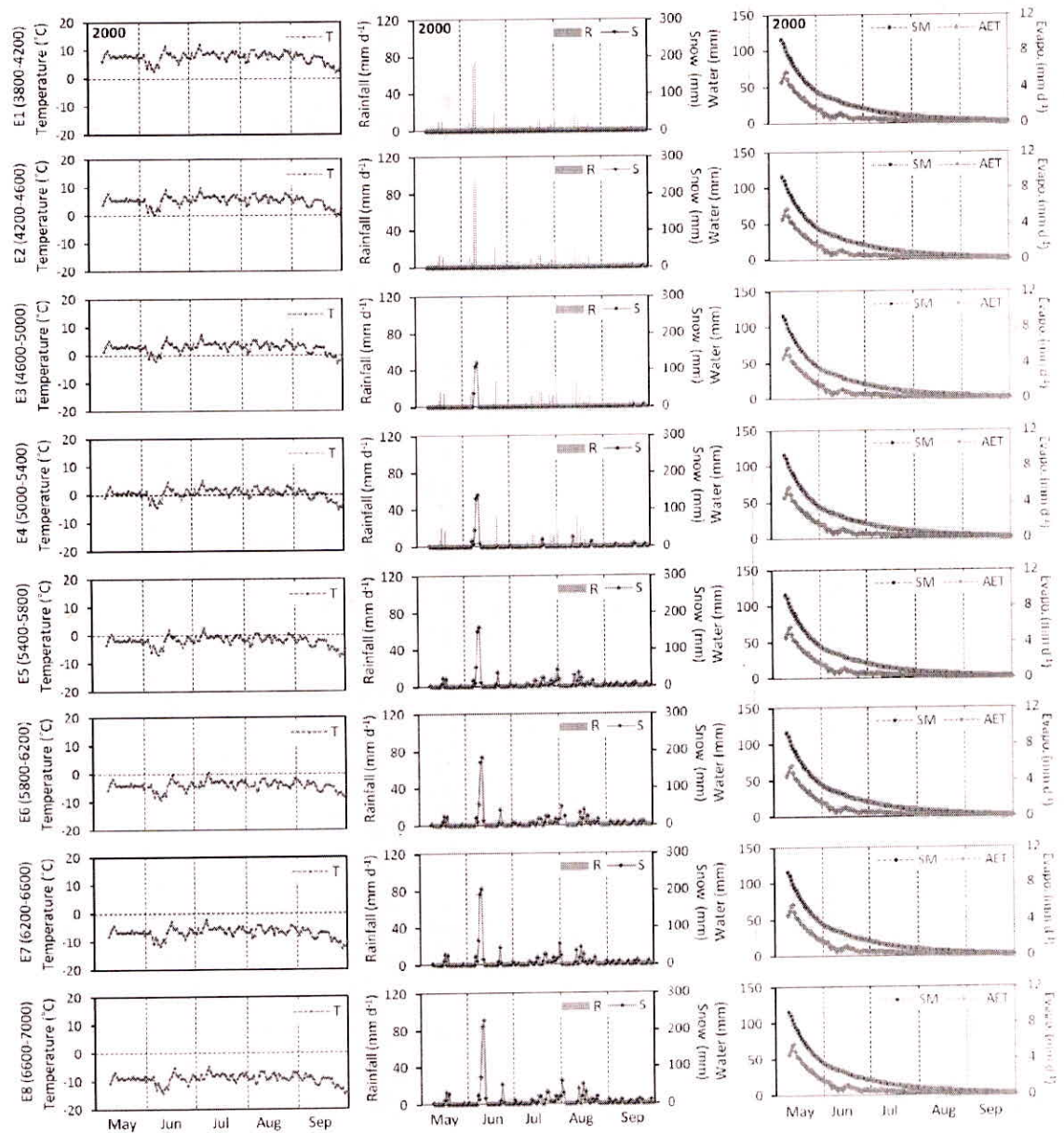


Fig. 5.1: Simulated temperature (T), rainfall (R), snowfall (S), water content of soil box (SM), and actual evapotranspiration (AET) at hypsometric scale from E1 (3800-4200 m a.s.l.) to E8 (6600-7000 m a.s.l.) for the Gangotri Glacier during May to September, 2000.

Table 5.1: Simulated meteorological parameters, water contents and evapotranspiration for each elevation band using HBV-Light model for the Gangotri Glacier during May, June, July, August and September of the year 2000.

Band	Elevation (m a.s.l.)	Temperature (°C)	Rain (mm d ⁻¹)	Snow (mm)	Snow melt (mm d ⁻¹)	W* (mm)	SM [‡] (mm)	AET [†] (mm d ⁻¹)
Gangotri Glacier 09 May – 31 May 2000								
E1	3800-4200	7.8	1.2	0.0	0.0	-5.2	72.8	3.3
E2	4200-4600	5.4	1.5	0.0	0.0	-21.5	72.8	3.3
E3	4600-5000	3.0	1.8	0.0	0.0	-33.7	72.8	3.3
E4	5000-5400	0.6	2.0	0.1	0.0	-10.2	72.8	3.3
E5	5400-5800	-1.8	0.0	2.5	0.0	-0.7	72.8	3.3
E6	5800-6200	-4.2	0.0	2.8	0.0	0.0	72.8	3.3
E7	6200-6600	-6.6	0.0	3.2	0.0	0.0	72.8	3.3
E8	6600-7000	-9.0	0.0	3.5	0.0	0.0	72.8	3.3
Gangotri Glacier 01 June – 30 June 2000								
E1	3800-4200	7.0	7.1	0.0	0.0	-19.9	30.1	0.8
E2	4200-4600	4.6	9.2	0.0	0.0	-79.4	30.1	0.8
E3	4600-5000	2.2	2.3	8.9	0.0	-123.7	30.1	0.8
E4	5000-5400	-0.2	1.5	11.7	0.0	-39.3	30.1	0.8
E5	5400-5800	-2.6	0.0	15.3	0.0	-2.8	30.1	0.8
E6	5800-6200	-5.0	0.0	17.3	0.0	0.0	30.1	0.8
E7	6200-6600	-7.4	0.0	19.4	0.0	0.0	30.1	0.8
E8	6600-7000	-9.8	0.0	21.4	0.0	0.0	30.1	0.8
Gangotri Glacier 01 July – 31 July 2000								
E1	3800-4200	8.6	3.4	0.0	0.0	-39.9	13.9	0.4
E2	4200-4600	6.2	4.4	0.0	0.0	-163.5	13.9	0.4
E3	4600-5000	3.8	5.4	0.0	0.0	-257.0	13.9	0.4
E4	5000-5400	1.4	5.5	0.9	0.0	-98.4	13.9	0.4
E5	5400-5800	-1.0	1.1	7.3	0.0	-11.4	13.9	0.4
E6	5800-6200	-3.4	0.0	8.3	0.0	-0.1	13.9	0.4
E7	6200-6600	-5.8	0.0	9.3	0.0	0.0	13.9	0.4
E8	6600-7000	-8.2	0.0	10.2	0.0	0.0	13.9	0.4
Gangotri Glacier 01 August – 31 August 2000								
E1	3800-4200	8.2	2.9	0.0	0.0	-61.0	6.5	0.2
E2	4200-4600	5.8	3.7	0.0	0.0	-253.7	6.5	0.2
E3	4600-5000	3.4	4.5	0.0	0.0	-408.6	6.5	0.2
E4	5000-5400	1.0	3.9	1.4	0.0	-167.0	6.5	0.2
E5	5400-5800	-1.4	0.2	6.1	0.0	-16.0	6.5	0.2
E6	5800-6200	-3.8	0.0	7.0	0.0	-0.2	6.5	0.2
E7	6200-6600	-6.2	0.0	7.8	0.0	0.0	6.5	0.2
E8	6600-7000	-8.6	0.0	8.6	0.0	0.0	6.5	0.2
Gangotri Glacier 01 September – 31 September 2000								
E1	3800-4200	6.0	0.9	0.0	0.0	-80.7	3.0	0.1
E2	4200-4600	3.6	1.0	0.2	0.0	-335.7	3.0	0.1
E3	4600-5000	1.2	1.5	0.3	0.0	-540.6	3.0	0.1
E4	5000-5400	-1.2	0.6	1.1	0.0	-221.4	3.0	0.1
E5	5400-5800	-3.6	0.0	2.0	0.0	-18.0	3.0	0.1
E6	5800-6200	-6.0	0.0	2.2	0.0	-0.2	3.0	0.1
E7	6200-6600	-8.4	0.0	2.5	0.0	0.0	3.0	0.1
E8	6600-7000	-10.8	0.0	2.8	0.0	0.0	3.0	0.1

*W – Change in glacier water content

[‡]SM - Water content of soil box

[†]AET – Actual Evapotranspiration

5.2.3 Actual Evapotranspiration

Actual evapotranspiration is the amount of water that is actually removed from a surface due to the processes of evaporation and transpiration. It is decreasing from May to September for all elevation from E1 (3800 m a.s.l.) to E8 (7000 m a.s.l.) as evaporation mainly depends on the fluctuation of temperature (Table 5.1).

5.2.4 Simulated Discharge

Glaciers delay melting and runoff from seasonal snow until late in the summer. Within a Glacierized catchment, they also supplement rainfall and seasonal snowmelt with melt water from glacial ice. Within a given glacier, there can be year to year variations in melt water storage and runoff. However, a few generalizations are possible. Here total discharge and elevation wise discharge is simulated by model for Gangotri Glacier for the summer season (May to September) 2000. Simulated and observed discharge are shown in Fig 5.2 and Table 5.2. Simulated discharge is maximum in the month of July. The monthly maximum discharge for the summer season of 2000 for the months of May to September were simulated to be 12.01, 14.75, 18.07, 17.89 and 12.96 mm d^{-1} whereas minimum discharge of the streamflow were simulated to be 2.7, 6.92, 11.07, 11.93 and 3.53 mm d^{-1} .

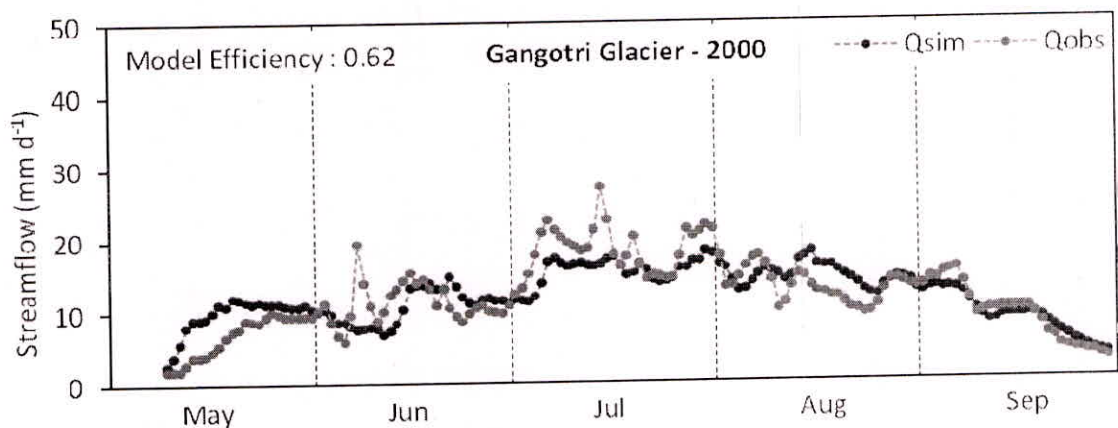


Fig 5.2: Simulated and observed discharge of Gangotri glacier for 2000.

The average discharge of the streamflow for these months was 9.81, 10.65, 15.25, 14.72 and 8.74 mm d^{-1} and monthly total discharge was 225.8, 319.9, 472.9, 456.4 and 262.2 mm for (Fig. 5.2 and Table 5.2).

Table 5.2: Simulated monthly streamflow of Gangotri Glacier during the summer seasons (May – September) of 2000 and 2001.

Particulars	Streamflow (mm d ⁻¹) - 2000					Streamflow (mm d ⁻¹) - 2001				
	May	Jun	Jul	Aug	Sep	May	Jun	Jul	Aug	Sep
Min. Discharge	2.7	6.92	11.07	11.93	3.53	1.66	7.95	16.39	11.02	4.18
Max. Discharge	12.01	14.75	18.07	17.89	12.96	12.1	17.24	25.04	19.21	11.19
Avg. Discharge	9.81	10.65	15.25	14.72	8.74	6.87	12.34	20.84	15.17	7.29
Total Discharge	225.8	319.9	472.9	456.4	262.2	213.1	370.3	646.1	470.3	218.9

5.3 Hydrological Modeling of Bhagirathi River: 2001

5.3.1 Interpolated Meteorological Parameters

For different elevation zones, the changes precipitation and temperature with elevation are calculated using the two parameters PCALT [% /100 m] and TCALT [°C / 100 m] (Equation 5.1 and 5.2). For each elevation monthly average temperature, precipitation (in the form of rainfall and snow) are simulated given in Table 5.2.

Average temperature is simulated from E1 to E8 band for Summer season vary from 6.8 to -9.0°C, 8.2 to -8.5°C, 10.1 to -6.7°C, 8.5 to -8.3°C and 6.3 to -10.5°C from May to September, respectively (Table 4.7 and Fig. 4.10). Similarly, average rainfall is simulated from E1 to E8 band for Summer season vary from 1.6 to 0.2 mm, 3.7 to 1.6 mm, 3.4 to 0.2 mm, 3.3 to 0.8 mm and 0.9 to 0.3 mm May to September, respectively (Table 5.3 and Fig. 5.3).

5.3.2 Amount of water in soil box

Amount of water in soil box (denoted by SM) is highly sensitive to snowfall in comparison to rainfall. For first three elevations (E1-E3), it is continuously decreasing from May to September as there is no snowfall is simulated by model. For next three elevations E4 and E5, it is not continuously decreasing but attains some constant value for few days within a month (Table 5.3 and Fig. 5.3). Further on increasing elevations E6-E8, snowfall increased and hence amount of water in soil box increased. Therefore, at

higher elevation it is nearly constant for all months throughout the summer season (Table 5.3 and Fig. 5.3).

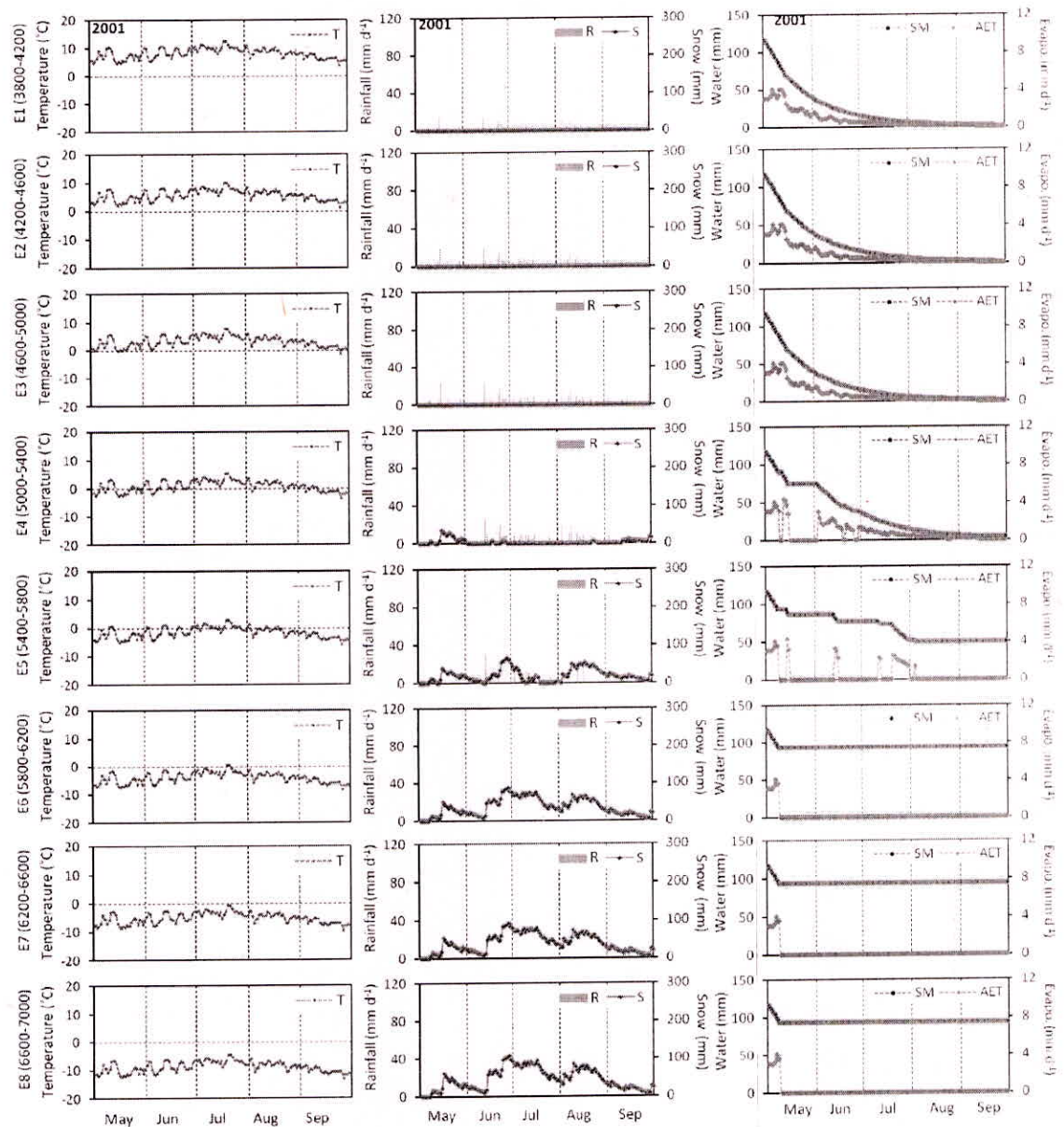


Fig. 5.3: Simulated temperature (T), rainfall (R), snowfall (S), water content of soil box (SM), and actual evapotranspiration (AET) at hypsometric scale from E1 (3800-4200 m a.s.l.) to E8 (6600-7000 m a.s.l.) for the Gangotri Glacier during May to September, 2001

Table 5.3: Simulated meteorological parameters, water contents and evapotranspiration for each elevation band using HBV-Light model for the Gangotri Glacier during May, June, July, August and September of the year 2001.

Band	Elevation (m a.s.l.)	Temperature (°C)	Rain (mm d ⁻¹)	Snow (mm)	Snow melt (mm d ⁻¹)	W* (mm)	SM [‡] (mm)	AET [†] (mm d ⁻¹)
Gangotri Glacier 01 May – 31 May 2001								
E1	3800-4200	6.8	1.0	0.0	0.0	-6.2	72.2	2.6
E2	4200-4600	4.4	1.3	0.0	0.0	-23.7	72.2	2.6
E3	4600-5000	2.0	1.6	0.0	0.0	-29.5	72.2	2.6
E4	5000-5400	-0.4	0.2	10.2	0.5	-8.4	84.9	1.5
E5	5400-5800	-2.8	0.0	14.3	0.3	3.7	92.1	1.1
E6	5800-6200	-5.2	0.0	18.9	0.0	3.2	96.4	0.9
E7	6200-6600	-6.4	0.0	20.0	0.0	1.3	96.4	0.9
E8	6600-7000	-10.0	0.0	23.4	0.0	0.2	96.4	0.9
Gangotri Glacier 01 June – 30 June 2001								
E1	3800-4200	8.2	2.4	0.0	0.0	-23.2	25.3	0.8
E2	4200-4600	5.9	3.1	0.0	0.0	-91.7	25.3	0.8
E3	4600-5000	3.5	3.7	0.0	0.0	-125.8	25.3	0.8
E4	5000-5400	1.1	4.2	1.0	0.7	-29.7	52.1	1.3
E5	5400-5800	-1.3	1.6	22.8	0.6	17.2	80.8	0.3
E6	5800-6200	-3.7	0.0	41.8	0.0	15.7	93.2	0.0
E7	6200-6600	-4.9	0.0	44.3	0.0	6.3	93.2	0.0
E8	6600-7000	-8.5	0.0	51.6	0.0	0.8	93.2	0.0
Gangotri Glacier 01 July – 31 July 2001								
E1	3800-4200	10.1	1.8	0.0	0.0	-45.1	9.9	0.3
E2	4200-4600	7.7	2.3	0.0	0.0	-187.9	9.9	0.3
E3	4600-5000	5.3	2.9	0.0	0.0	-298.8	9.9	0.3
E4	5000-5400	2.9	3.4	0.0	0.0	-124.8	23.1	0.7
E5	5400-5800	0.5	3.1	10.5	1.8	40.5	69.1	0.8
E6	5800-6200	-1.9	0.2	55.5	0.2	43.7	93.2	0.0
E7	6200-6600	-3.1	0.0	60.4	0.0	17.4	93.2	0.0
E8	6600-7000	-6.7	0.0	70.5	0.0	2.2	93.2	0.0
Gangotri Glacier 01 August – 31 August 2001								
E1	3800-4200	8.5	1.9	0.0	0.0	-68.8	3.9	0.1
E2	4200-4600	6.1	2.4	0.0	0.0	-294.4	3.9	0.1
E3	4600-5000	3.7	3.0	0.0	0.0	-499.6	3.9	0.1
E4	5000-5400	1.3	3.3	0.4	0.1	-255.1	9.3	0.3
E5	5400-5800	-1.1	0.8	34.5	0.3	34.0	50.3	0.0
E6	5800-6200	-3.5	0.0	47.1	0.0	68.8	93.2	0.0
E7	6200-6600	-4.7	0.0	50.5	0.0	27.5	93.2	0.0
E8	6600-7000	-8.3	0.0	59.0	0.0	3.4	93.2	0.0
Gangotri Glacier 01 September – 31 September 2001								
E1	3800-4200	6.3	0.6	0.0	0.0	-88.1	1.6	0.0
E2	4200-4600	3.9	0.8	0.0	0.0	-375.5	1.6	0.0
E3	4600-5000	1.5	0.9	0.0	0.0	-630.4	1.6	0.0
E4	5000-5400	-0.9	0.3	4.1	0.0	-308.1	4.5	0.1
E5	5400-5800	-3.3	0.0	13.5	0.0	54.5	50.2	0.0
E6	5800-6200	-5.7	0.0	15.6	0.0	86.6	93.2	0.0
E7	6200-6600	-6.9	0.0	16.5	0.0	34.6	93.2	0.0
E8	6600-7000	-10.5	0.0	19.3	0.0	4.3	93.2	0.0

*W – Change in glacier water content ‡SM – Water content of soil box †AET – Actual Evapotranspiration

5.3.3 Actual evapotranspiration

Actual evapotranspiration is the amount of water that is actually removed from a surface due to the processes of evaporation and transpiration. It is denoted by AET. It is inversely proportional to snowfall amount indicating higher the snowfall lower the evapotranspiration. It is decreasing from May to September for E1, E2 and E3. For elevation E4 and E5, a zig zag pattern is simulated (Table 5.3 and Fig. 5.3). Peak values are simulated where snowfall is zero. Further increasing elevation E6-E8, it is zero throughout the summer season except in the month of May. As snowfall increases its value decreases due to cloud cover at high elevations which reflect most of the incoming short wave radiations and hence decrease the temperature.

5.3.4 Simulated Discharge

Total discharge and elevation wise discharge is simulated for Gangotri Glacier system for the summer season (May to September) 2001. Simulated and observed discharge is shown in (Fig. 5.4).

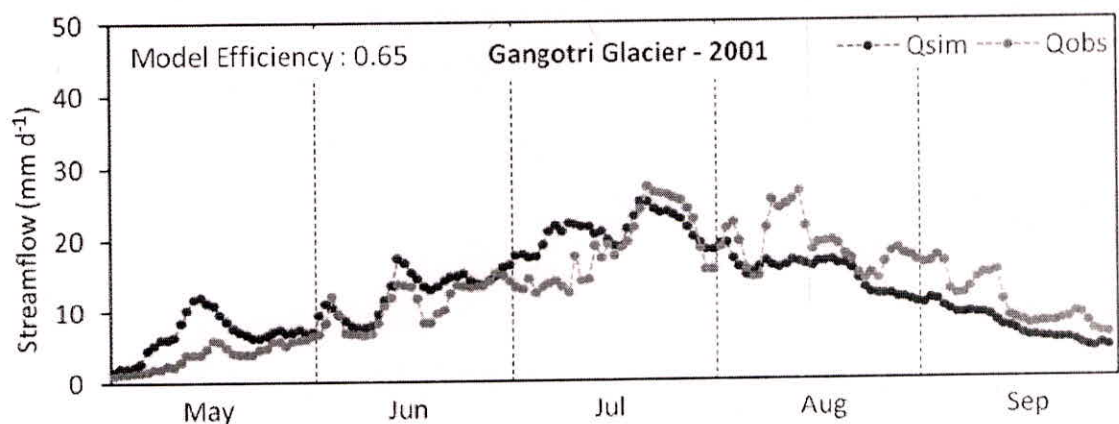


Fig. 5.4: Simulated and observed discharge of Gangotri glacier for 2000.

Simulated discharge is maximum in the month of July due to presence of good rainfall and snowfall, while minimum in the months of May and September. The monthly maximum discharge for the summer season of 2001 for the months of May to September were simulated to be 12.01, 17.24, 25.04, 19.21 and 11.19 mm d⁻¹ whereas minimum discharge of the streamflow were simulated to be 1.66, 7.95, 16.39, 11.02 and 4.18 mm d⁻¹. The average discharge of the streamflow for these months was 6.87, 12.34, 20.84, 15.17 and 7.29 mm d⁻¹ and monthly total discharge was 213.1, 370.3, 646.1, 470.3 and 218.9 mm for (Fig. 5.4 and Table 5.2).

5.4 Validation: Goodness of Fit (2000 and 2001)

Goodness of fit is a component of regression analysis, which is a statistical method used in finance and a variety of other fields to make predictions based on observed values. In other words, it is a measure of how correlated a group of actual observations are to a model's predictions.

5.4.1 Coefficient of determination

The coefficient of determination (denoted by R^2) is a key output of regression analysis. It is interpreted as the proportion of the variance in the dependent variable that is predictable from the independent variable. It is a statistic used in the context of statistical models whose main purpose is either the prediction of future outcomes or the testing of hypotheses, on the basis of other related information. It provides a measure of how well observed outcomes are replicated by the model, based on the proportion of total variation of outcomes explained by the model.

$$R^2 = \frac{\left(\sum (Q_{obs} - \overline{Q_{obs}})(Q_{sim} - \overline{Q_{sim}}) \right)^2}{\sum (Q_{obs} - \overline{Q_{obs}})^2 \sum (Q_{sim} - \overline{Q_{sim}})^2}$$

(5.3)

where, Q_{obs} is observed discharge and Q_{sim} is simulated discharge.

Coefficient of determination plot between observed and simulated streamflow is shown in Fig.5.5. It is observed that determination of coefficient is 0.62 for 2000 while, for summer season 2001, R^2 has a very good value of 0.65, and all most all the points fall along the line which reflects a fair correlation between observed and simulated streamflow (Table 5.4 and Fig. 5.5). R^2 for 2000 and 2001 shows that simulations are in good agreement with observations.

Table 5.4: Showing Goodness of Fit functions for 2000 and 2001 with computed values.

Goodness of Fit	Description	Simulated results	
		2000	2001
R^2 (Daily data)	Coefficient of determination	0.62	0.65
R_{eff}	Model Efficiency	0.62	0.65
Mean Diff	Mean difference	0.01	-0.35

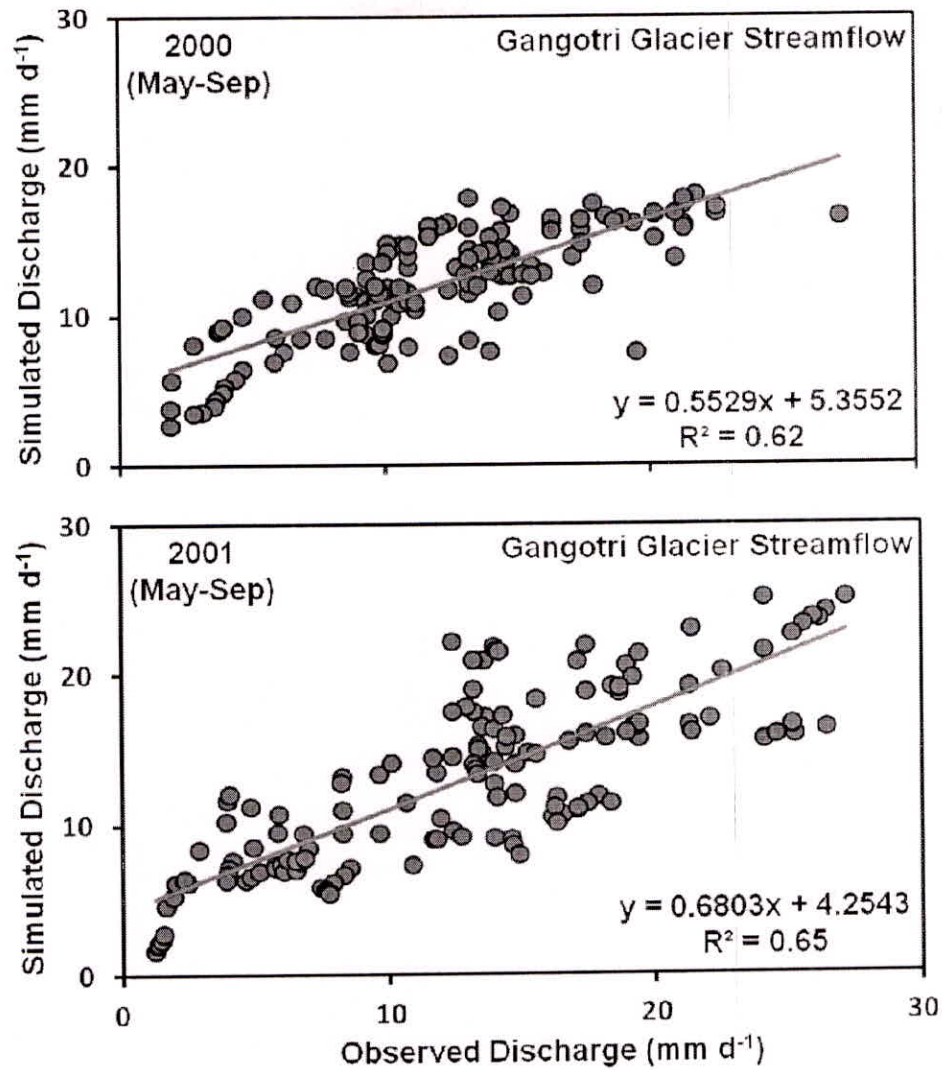


Fig. 5.5: Coefficient of determination plot between observed and simulated melt (mm d⁻¹) for the summer seasons (May-September) of 2000 and 2001.

5.4.2 Model Efficiency

The model efficiency (here denoted by R_{eff}) compares the prediction by the model with the simplest possible prediction, a constant value of the observed mean value over the entire period. It is normally used for assessment of simulations by the HBV model. Model efficiency for both years are calculated by using given equation.

$$R_{eff} = 1 - \frac{\sum (Q_{obs} - Q_{sim})^2}{\sum (Q_{obs} - \bar{Q}_{obs})^2} \quad (5.4)$$

Model efficiency was calculated 0.62 for 2000 and 0.65 for 2001 (Table 5.4). Results show that efficiency is positive for both years which indicate good performance of the model. In 2001, it is estimated that the model efficiency is 5% higher than 2000.

This is due to higher amount of summer snowfall and winter residual snow in 2001 as the model simulates snow ice surface melting using degree-day method. This method works well for the simulation of clean snow /ice melt.

5.4.3 Mean Difference

The mean difference may also refer as mean absolute difference (univariate) is a measure of statistical dispersion equal to the average absolute difference of two independent values drawn from a probability distribution.

$$MeanDiff = \frac{\sum (Q_{obs} - Q_{sim})}{n} \quad (5.5)$$

where, Q_{obs} is observed discharge, Q_{sim} is simulated discharge and n is the number of time step.

Mean difference between simulated and observed discharge for 2000 is calculated positive i.e. 0.0090 (Table 5.4) which indicates that the simulated discharge is underestimated and for 2001 it is negative i.e. -0.3534 (Table 5.4) which indicates that the simulated discharge is overestimated. This is due to higher amount of summer snowfall and winter residual snow in 2001 as the model simulates snow ice surface melting using degree-day method. This method works well for the simulation of clean snow / ice melt.

Chapter – 6

Synthesis

6.1 Summary and Conclusions

Presently climate change, global warming and water scarcity are some of the major problems we are facing. Glaciers store about 75 % of the world's fresh water and are important indicators of global warming and climate change in several ways. Melting ice sheets contribute to rising sea levels. As ice sheets in Antarctica and Greenland melt, they raise the level of the ocean. Tons of fresh water are added to the ocean every day. Also glaciers are important features in the hydrologic cycle and affect the volume, variability, and water quality of runoff in areas where they occur. Estimation of streamflow in any basin is important in regards of hydro power plants, irrigation, ingestion. Therefore, it is essential to study the glacier and glacial hydrology.

In this scientific approach, an attempt has been made to study the basics of glaciology, how Physics works in natural system and use of mathematical modeling in the field of glaciology to estimate the streamflow from glacierized catchment. Without the accurate knowledge about the glacier distribution it would be difficult to study their impact on our climate and environment. Thus, global distribution of

glaciers with special reference to the Himalaya is studied. Further, types of hydrological models are studied to estimate the hydrological modeling of streamflow of Gangotri Glacier.

Gangotri Glacier is one of the largest glaciers in Garhwal Himalayas. It lies in the central crystalline zone within the latitudes $30^{\circ}43'$ – $31^{\circ}01'N$ and longitudes $79^{\circ}0'$ – $79^{\circ}17'E$. The proglacial melt-water stream, known as Bhagirathi River, emerges from the snout of the Gangotri Glacier at an elevation of 4000 m a.s.l. The Gangotri Glacier system (most commonly known as Gangotri Glacier) is a cluster of many glaciers (Kirti Bamak, Ghanohim Bamak, swachhand bamak and Maiandi bamak) comprising the main Gangotri Glacier (length: 30.20 km; width: 0.20–2.35 km; area: 86.32 km^2) as the trunk part of the system. Total catchment area of the Gangotri Glacier system up to the meteorological site established downstream the snout is about 556 km^2 , out of which the glacierized area is about 286 km^2 .

The Gangotri Glacier and the River Ganga has important status in the Indian culture. Scriptures and faith of the population consider it a sacred river. Perhaps billion of the Indian people depend on these rivers for their water and food supply. Lot of population is dependent on it for irrigation and drinking water purposes. Because of these reasons it has attained lot of attention. Climate change and global warming has added more concern to the health of the Gangotri glacier and Ganga River.

Previously developed semi-distributed conceptual hydrological model (HBV-Light) is used to estimate the streamflow of Bhagirathi River using observed hydro-meteorological parameters viz. temperature, precipitation, rainfall and discharge. The used hydro-meteorological data was collected by Haritashya et al. (2010) at about 3800 m a.s.l. near the snout of the Gangotri Glacier at hydro-meteorological observatory site. The streamflow of Gangotri Glacier system is simulated using HBV-Light hydrological model during the two ablation seasons (May to September) of 2000 and 2001. To interpolate the meteorological parameters (air temperature, rainfall, snowfall and actual evaporation), the Gangotri Glacier area-altitude is divided into eight elevation zones viz. E1 (3800-4200), E2 (4200-4600), E3 (4600-5000), E4 (5000-5400), E5 (5400-5800), E6 (5800-6200), E7 (6200-6600) and E8 (6600-7000 m a.s.l.). For different elevation zones, model interpolate temperature and precipitation

using slope lapse rate approach and corrections factors. Whereas, the meltwater from snow/ice is simulated using degree-day method. The HBV Model is operated in standard mode including Snow routine, Soil Routine, Response function, Routing routine including glacier routine. The model calculates elevation wise temperature, rainfall, snowfall, actual evaporation, change in glacier water content, contribution of rain to the simulated runoff, contribution of snow to the simulated runoff, contribution of glacier to the simulated runoff, storage in the specific zone, water content in soil box and simulated runoff.

Results show that the air temperature in the Gangotri Glacier valley ranges between 8.6 °C to -15.2 °C for the ablation season of 2000, whereas it was 10.1 °C to -10.5 °C for the 2001. Temperature is maximum in the month of July for both year. Rainfall ranges between 9.2 mm to 0 mm for 2000, whereas it was 4.2 mm to 0 mm for 2001. Rainfall is maximum in the month of June for both year. Snowfall is maximum in the month of June and July for 2000 and 2001, respectively. Actual evapotranspiration is highly sensitive to snowfall. It is maximum when there is no snowfall i.e. for low elevation (E1 to E3), while it is minimum for higher elevation because this area is highly covered with clouds. It is possible due to minimum sunshine hours (or presence of more clouds) in these elevations. The presence of clouds prevents the shortwave solar radiation in the daytime to heat the surface and longwave radiation in the night-time to emit the heat, resulting in lowers the actual evapotranspiration. It is maximum in the month of May and minimum in the month of September for both year. The model efficiency is calculated 0.62 and 0.65 for 2000 and 2001, respectively. Results also suggests that the HBV model is highly sensitive to change in degree-day factor (CFMAX), precipitation calibration factor (PCALT) and temperature laps rate (TCALT) of the catchment.

Computer models, especially conceptual models, are frequently used for catchment hydrology studies. Different functionalities, such as an automatic calibration using a genetic algorithm or a Monte Carlo approach, as well as the possibility to perform batch runs with predefined model parameters make the HBV model interesting especially for estimating streamflow. HBV model also continue to be used as important tools in various kinds of hydrological applications. The HBV model is a semi-distributed model, which means that a catchment can be separated

into different elevation and vegetation zones as well as into different sub-catchments. The model consists of different routines and simulates catchment discharge, usually on a daily time step, based on time series of precipitation and air temperature as well as estimates of monthly long-term potential evaporation rates. In addition, a routine for simulation of glaciers has been also implemented. From all the study which is conducted during this dissertation report we have concluded that the HBV model can be useful at any location on the glacier where input data (i.e. meteorological measurements) are available. Since, the HBV model includes soil routine and glacier routine therefore, this model is especially beneficial for the estimation of melt-modelling streamflow of Glacierized catchment of the world. Suggested that all important processes were moderately included in the HBV model.

6.2 Future Scope

As 1.8 billion People are directly affected by the water resource of the Himalaya and are dependent on the melt water generated by the Himalayan glaciers (Bolch et al., 2012). However, very less attention has been given especially on the study of estimation of streamflow from Himalayan glaciers because of the extreme variations in altitude and aspect. Meanwhile, the water cycle in the remote area yet remains a mystery. We don't know that how glaciers change in a future will affect the river flow. In the perspective of water resource analysis, the runoff from glacierized basin needs to be evaluated precisely. It is well established fact that the discharge from the glacier melt gives major contribution to river. Therefore, to predict the glacier melt discharge in river and changing climate, we need to understand that how the Cryospheric system of Himalaya is working and which component is the most responsible for the fluctuation of glaciers and how this varies throughout the seasons. The more we know about the physics of glaciers more we will be able to make use of it in an efficient way. Therefore, more research is needed in Himalayan Glaciers which will make more us aware about the water availability in glaciers and how the change in glaciers effects the environment.

The estimation of streamflow for a glacierized catchment with the help of semi-distributed conceptual runoff model is the key tool to predict the hydrology in a particular area for making hydro-power dams, availability of fresh water for irrigation and ingestion. In present study, the streamflow (discharge) of a glacierized catchment

is simulated on daily scale. Further, in this study average slope lapse rate or temperature correction factor (TCALT) has been used for the entire summer season (May-September) due to unavailability of observed temperature records at altitudinal scale. Since, the slope lapse rate is continuously changing according to the condition of climate, this improvement can be done to get the more accuracy in simulated streamflow. However, in future to know the diurnal pattern of snow/ice surface melting and streamflow variations can be estimated by running the model on hourly scale. This needs collection of accurate meteorological data at hourly scale which is possible only by using advanced automatic instrumentation techniques such as Automatic Weather Station (AWS), Automatic Water Level Recorder, Total Precipitation Gauge, etc. It can help us to provide the accurate information about the approximate melt runoff from the snow/glacier surface which varies according to the conditions of climate. This could lead to a better understanding of streamflow variations at diurnal scale which will be helpful for future prediction of streamflow in a glacierized catchment more accurately. In addition, there are some forcing factors such as weather conditions (radiative fluxes, role of wind speed, change in atmospheric pressure, etc.) and surface parameters (debris-cover, ponds/lakes, etc.) which are not included in the HBV model. In future, this could be an opportunity to enhance this model by including the forcing factors which are responsible for snow/ice surface melt. Further research can be carried out to study the melt water generated by different sub-catchments in glacierized regions of the Western, Central and Eastern Himalaya using semi-distributed conceptual runoff model.

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