

A
Dissertation Report
on

Thermodynamical Changes of the Debris-Covered Himalayan Glaciers

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

**Master of Science
In
Physics**

By

Arushi Sharma

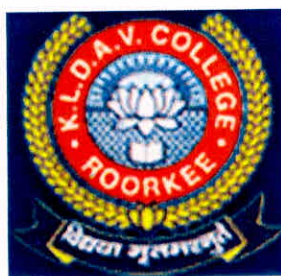
(Enrollment No. G11206817 Roll No. 15316343002)

Under the joint supervision of

Dr. Manohar Arora
Scientist – D
National Institute of Hydrology
Roorkee, Uttarakhand

Prof. Ajay Gairola
Department of Civil Engineering
Indian Institute of Technology Roorkee
Roorkee, Uttarakhand

Submitted to



Department of Physics

Kanahiya Lal D.A.V. (P.G.) College

(Affiliated to HNB Garhwal University Srinagar, Uttarakhand)

Roorkee, Uttarakhand – 247 667, INDIA

July 2017

Dedicated to
My Eternal Inspiration...
My Beloved Mother

DECLARATION

I hereby declare that the submission entitled - '**Thermodynamical Changes of the Debris-Covered Himalayan Glaciers**' for the award of the degree of **Master of Science in Physics** at **Hemwati Nandan Bahuguna Garhwal University, Uttarakhand** is my own work except where due acknowledgement has been made in the text, which has not been submitted for the award of any degree or diploma of the university or other institute of higher learning. This work is carried out under the joint supervision of Dr. Manohar Arora (Scientist-D, National Institute of Hydrology (NIH), Roorkee, Uttarakhand), and Dr. Ajay Gairola (Professor, Department of Civil Engineering, Indian Institute Technology of Roorkee (IITR), Uttarakhand) and Dr. Kapil Kesarwani (Postdoctoral Fellow, Department of Civil Engineering, IITR) at National Institute of Hydrology and Indian Institute of Technology Roorkee during 20 February to 24 July, 2017.

(Arushi Sharma)

Date : 24 July 2017

Place : Roorkee, Uttarakhand

Roll No. 15316343002

Kanahiya Lal D.A.V. (P.G.) College, Roorkee
(Affiliated to HNB Garhwal University Srinagar, Uttarakhand)

E-mail: arushisharma1395@gmail.com

DISSERTATION COMPLETION CERTIFICATE

This to certify that the dissertation report entitled - 'Thermodynamical Changes of the Debris-Covered Himalayan Glaciers' submitted by Ms. Arushi Sharma for the award of the degree of Master of Science in Physics (Kanahiya Lal D.A.V. (P.G.) College, Roorkee affiliated to Hemwati Nandan Bahuguna Garhwal University Srinagar, Uttarakhand) embodies the work of herself. She worked under our joint supervision for this report during 20 February to 24 July 2017 at National Institute of Hydrology (NIH) and Indian Institute of Technology Roorkee (IITR). This work and the contents, in full and parts have not been submitted to any other institute or university for the award of any other degree or diploma.

Supervisor



(Dr. Manohar Arora)

Scientist - 'D'

Manohar Arora, Ph. D.
National Institute of Hydrology
Scientist B
Surface Water Hydrology Division
Roorkee, Uttarakhand - 247 667
National Institute of Hydrology
Roorkee - 247 667

Co-supervisor (s)

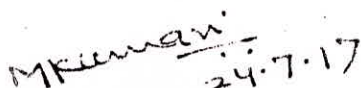


(Prof. Ajay Gairola)

Department of Civil Engineering
Indian Institute of Technology Roorkee
Roorkee, Uttarakhand - 247 667
Contact: +91-9897051525

Professor
Department of Civil Engineering,
Indian Institute of Technology Roorkee
ROORKEE - 247 667, India

Forwarded By

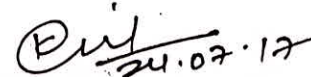


(Dr. Mithilesh Kumari)

Head of Department

Kanahiya Lal D.A.V. (P.G.) College,
Roorkee, Uttarakhand - 247 667

Contact: +91-1332-262268



(Dr. Kapil Kesarwani)

Postdoctoral Fellow

Department of Civil Engineering
Indian Institute of Technology Roorkee
Roorkee, Uttarakhand - 247 667

Contact: +91-8881377714

Professor
Department of Civil Engineering,
Indian Institute of Technology Roorkee
ROORKEE - 247 667, India

Date: 24 July, 2017

Place: Roorkee, Uttarakhand

विभागाध्यक्ष
भौतिक विज्ञान विभाग
के.एन.डी.ए.वी. (पी.जी.) कॉलेज
रूडकी

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(Arushi Sharma)

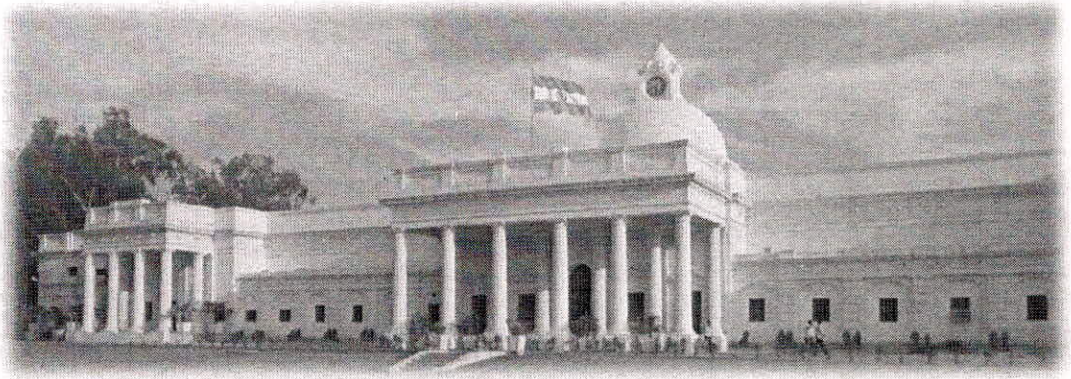
ABOUT THE INSTITUTES / ORGANIZATION

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 programme, 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.

Contents

Acknowledgements	i
About the Institute / Organization	ii
List of Figures	vii
List of Table	ix
List of Abbreviations	x
List of Symbols	xi
Chapter – 1 Introduction	1-17
1.1 Formation of glaciers	1
1.2 Classification of Glaciers	2
1.2.1 Continental Glacier	2
1.2.2 Valley (Mountain) Glacier	4
1.2.3 Glacier Classification on the basis of size and shape	4
1.2.4 Glacier Classification the basis of Thermal state	6
1.3 Glacial Geomorphology	6
1.3.1 Horn	6
1.3.2 Arete	8
1.3.3 Cirque	8
1.3.4 Bergschrund line	8
1.3.5 Accumulation zone	8
1.3.6 Equilibrium-Line-Altitude (ELA)	9
1.3.7 Ablation zone	10
1.3.8 Crevasses	10
1.3.9 Moulin	10
1.3.10 Supraglacial Lake	11
1.3.11 Snout	11
1.3.12 Moraine	13
1.3.13 Proglacial Lake	13
1.3.14 Supraglacial Debris	13
1.3.15 Supraglacial Channel	13
1.4 Research Motivation	15
1.5 Thesis Overview	16

Chapter – 2	Glacier Distribution	18-21
2.1	World Glacier Distribution	18
2.2	Himalayan Glacier System	19
Chapter – 3	Surface Energy Budget of Debris-covered Glacier	22-31
3.1	Introduction	22
3.2	Debris Energy-Balance (DEB) Model	23
3.3	Radiation Fluxes	25
3.3.1	Shortwave Radiation	25
3.3.2	Longwave Radiation	26
3.4	Turbulent Heat Flux	27
3.4.1	Sensible Heat Flux	28
3.4.2	Latent Heat Flux	29
3.5	Heat Flux due to Precipitation	30
3.5.1	Heat Flux due to Snow	30
3.5.2	Heat Flux due to Rain	30
3.6	Subsurface Heat Flux	30
Chapter – 4	Instrumentation and Data Acquisition	32-44
4.1	Automatic Weather Station	32
4.2	Air Temperature and Relative Humidity Probe (HC2S3)	34
4.3	Infrared Surface Temperature Sensor (SI-111)	37
4.4	Wind Monitor (05103)	37
4.5	Pyranometer (CMP3-L)	38
4.6	Sonic Ranging Sensor (SR50A)	39
4.7	Net Radiometer (NR-LITE)	40
4.8	Barometric Pressure Sensor (61302V)	41
4.9	Rain Gauge (TE 525M)	41
4.10	Data Logger (CR-1000)	42
Chapter – 5	Thermodynamical changes of Debris-covered Chorabari Glacier Surface	45-56
5.1	Chorabari Glacier	45

5.1.1 Climatic Condition	47
5.2.2 Characteristics of Debris-cover	48
5.2 Meteorological Observations	48
5.2.1 Air Temperature	48
5.2.2 Relative Humidity	50
5.2.3 Wind Speed	50
5.2.4 Radiative Fluxes	50
5.2.5 Atmospheric Pressure	51
5.2.6 Net Radiation	52
5.2.7 Precipitation	53
5.3 Results and Discussion	53
5.3.1 Estimation of Cloud Cover	53
5.3.2 Estimation of Longwave Radiation	54
5.3.3 Estimation of Energy Fluxes at the Chorabari Glacier	55
Chapter – 6 Synthesis	57-60
6.1 Summary and Conclusions	57
6.2 Future Scope	59
References	61-66

List of Figures

	Page No.
Chapter – 1 Introduction	
Fig. 1.1 Formation of glacier.	2
Fig. 1.2 a) Continental Glacier, b) Valley Glacier (Source: Kesarwani, 2015), c) debris-free (clean) glacier and d) debris-free (clean) glacier.	3
Fig. 1.3 Temperature mechanism of Polar and Temperate glaciers (Source: Kesarwani, 2015).	6
Fig. 1.4 a) Horn structure in mountain glacier, b) Arête structures in mountain glacier c) Cirque formation in mountain glacier d) Bergschrund line in Sperry Glacier, Montana, USA	7
Fig. 1.5 A typical diagram explaining the surface characteristics of a valley glacier.	9
Fig. 1.6 a) A view of longitudinal crevasses, b) An image showing Moulin structure, c) supra glacial lake, d) A view of snout of glacier.	12
Fig. 1.7 a) Glacial moraines above Lake Louise, Alberta, Canada b) Pro-glacial Lake, c) Supra-Glacial Debris, d) Supra Glacial Channel.	14
Chapter – 2 Glacier Distribution	
Fig. 2.1 Distribution of Glaciers around the world.	19
Fig. 2.2 Indian Himalayan Glacier system	20
Chapter – 3 Surface Energy Budget of Debris-covered Glacier	
Fig. 3.1 Schematic of the DEB model showing heat fluxes at the top and bottom of a debris layer of thickness (d). The debris temperature (T_d) is calculated for N layers of thickness h_d , with boundary conditions defined by the surface temperature (T_s) and the temperature of the debris/ice interface (T_i), which is assumed to stay at 273.15 K. The dash-dot curve is a temperature profile, where temperature increases towards the right. H_G is conductive heat flux at the debris surface and H_{Gi} is a conductive heat flux considered to reach the glacier ice (Source: Kesarwani, 2015).	24
Chapter – 4 Instrumentation and Data Acquisition	
Fig. 4.1 Complete setup of Automatic Weather Station installed at Base camp site (3820 m a.s.l.) of the Chorabari Glacier, Garhwal Himalaya (Source: Kesarwani, 2015).	33

Fig. 4.2	a) Air temperature and Relative Humidity Probe (HC2S3), b) Infrared Surface Temperature Sensor (SI-111) c) Wind Monitor (05103) d) Pyranometer (CMP3-L) e) Sonic Ranging Sensor (SR50A) f) Net Radiometer (NR-LITE).	36
Fig. 4.3	a) Barometric Pressure sensor (61302V), b) Rain Gauge (TE 525M) and c) Setup of CR-1000 Data logger.	43

Chapter – 5 Thermodynamical changes of Debris-covered Chorabari Glacier Surface

Fig. 5.1	Location of the Debris-covered Chorabari Glacier, Garhwal Himalaya, Uttarakhand. (Source: Kesarwani et al., 2015)	46
Fig. 5.2	Time series of daily mean meteorological variables recorded from 01 May 2012 to 31 October 2012. T_a , RH, and W_s are representing the air temperature, relative humidity and wind speed respectively.	49
Fig. 5.3	Daily mean recorded ($S_{\downarrow}$ and $S_{\uparrow}$) or computed (STOA and Snet) meteorological variables from 01 May 2012 to 31 October 2012. A data gap in $S_{\downarrow}$ from 17 August 2012 to 10 October 2012 was exists. Periods colored with grey are the linearly interpolated modelled data for this data gap. STOA, $S_{\downarrow}$, and $S_{\uparrow}$ are representing the extraterrestrial solar irradiance at top of atmosphere, incoming shortwave radiation, and reflected shortwave radiation, respectively.	51
Fig. 5.4	Daily observed atmospheric pressure (P_a) in the Chorabari Glacier catchment (Source: Kesarwani, 2015).	52
Fig. 5.5	Daily observed net radiation (R_n) in the Chorabari Glacier catchment. (Source: Kesarwani, 2015).	52
Fig. 5.6	Monthly observed rainfall records in the Chorabari Glacier catchment. (Source: Dobhal et al., 2013; Kesarwani, 2015).	53
Fig. 5.7	Daily estimated cloud cover during the ablation season of the Chorabari Glacier catchment (Source: Kesarwani et al., 2015).	53
Fig. 5.8	Daily estimated recorded ($L_{\downarrow}$ and $L_{\uparrow}$) meteorological variables from 01 May 2012 to 31 October 2012. $L_{\downarrow}$, and $L_{\uparrow}$ are representing the incoming longwave radiation and reflected longwave radiation (Source: Kesarwani, 2015).	55
Fig. 5.9	The average hourly energy fluxes calculated for the Chorabari Glacier catchment at the AWS site. S represents the net shortwave radiation, L represents the net longwave radiation, H_s represents the sensible heat flux, H_L represents the latent heat flux, H_G represents the conductive heat flux and P represents the Precipitation heat flux (Source: Kesarwani, 2015).	55

List of Tables

Page No.

Chapter – 2 Glacier Distribution

Table 2.1	Basin/Sub-basin wise distribution of glaciers in the Indian administrative part of the Himalaya	21
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Chapter – 4 Instrumentation and Data Acquisition

Table 4.1	Detailed specification of Temperature sensor	35
Table 4.2	Detailed specification of RH sensor	35
Table 4.3	Detailed specification of Wind monitor sensor	37
Table 4.4	Detailed specification of Pyranometer	38
Table 4.5	Detailed specification of Sonic ranging sensor	39
Table 4.6	Detailed specification of NR-LITE	40
Table 4.7	Detailed specification of Barometric pressure sensor	41
Table 4.8	Detailed specification of Rain gauge	42
Table 4.9	Specification of Data Logger	44

List of Abbreviations

AWS	Automatic Weather Station
Air_Temp	Air Temperature
DEB	Debris Energy Balance
e_hmp	Vapour Pressure
e_sat_vp	Saturated Vapour Pressure
ELA	Equilibrium-Line Altitude
IR	Infra-Red
SEB	Surface Energy Balance
TC	Thermal Conductivity
VPD	Dew Point Vapour Pressure
W_Dir	Wind Direction
W_Spd	Wind Speed

List of Symbols

Symbols	Meaning	Unit	Value
A	Empirical constant		0.021286
C_d	Specific heat capacity of debris	$J\ kg^{-1}\ K^{-1}$	2.09×10^3
C_F	Cloud factor		
C_i	Specific heat capacity of ice	$J\ kg^{-1}\ K^{-1}$	2.05×10^3
C_L	Bulk exchange Coefficient for latent heat flux		0.002
C_P	Specific heat capacity of dry air	$J\ kg^{-1}\ K^{-1}$	1005
C_s	Specific heat capacity of snow	$J\ kg^{-1}\ K^{-1}$	2.09×10^3
C_S	Bulk exchange coefficient for sensible heat flux		0.002
c_w	Specific heat capacity of water	$J\ kg^{-1}\ K^{-1}$	4.2×10^3
d	Total thickness of debris over glacier surface	m	
e	Vapour Pressure	h Pa	
e_a	Vapour Pressure of Air	h Pa	
e_s	Vapour Pressure of surface	h Pa	
H_G	Conductive heat flux	$W\ m^{-2}$	
H_{Gi}	Conductive heat flux at debris-ice interface	$W\ m^{-2}$	
H_L	Turbulent Latent heat flux	$W\ m^{-2}$	
H_P	Heat flux due to precipitation	$W\ m^{-2}$	
H_R	Heat flux provided by rain	$W\ m^{-2}$	
H_S	Turbulent sensible heat flux	$W\ m^{-2}$	
H_{snow}	Energy required melting or refreezing of snow	$W\ m^{-2}$	
h	Snow depth	m	
h_d	Debris thickness	m	
k	Thermal conductivity	$W\ m^{-1}\ K^{-1}$	
k_d	Thermal conductivity of debris	$W\ m^{-1}\ K^{-1}$	2.8
k_i	Thermal conductivity of ice	$W\ m^{-1}\ K^{-1}$	2.22
k_s	Thermal conductivity of snow	$W\ m^{-1}\ K^{-1}$	
$L_{\downarrow}$	Incoming longwave radiation	$W\ m^{-2}$	
$L_{\uparrow}$	Outgoing longwave radiation	$W\ m^{-2}$	
L_F	Latent heat of vaporisation	$J\ kg^{-1}$	2.514×10^6
L_n	Net longwave radiation	$W\ m^{-2}$	
L_w	Latent heat of fusion of water	$J\ kg^{-1}$	334.864×10^3
P	precipitation	mm w.e.	

P_a	Atmospheric Pressure	h Pa	
q_a	Specific humidity of air	$g\ kg^{-1}$	
q_s	Specific humidity of surface	$g\ kg^{-1}$	
R_g	Universal gas constant	$J\ K^{-1}\ mol^{-1}$	8.31
R_H	Relative humidity	%	
R_n	Net radiation	$W\ m^{-2}$	
R_o	Rainfall collected from ordinary rain gauge	mm	
$S_{\downarrow}$	Incoming shortwave radiation	$W\ m^{-2}$	
$S_{\uparrow}$	Outgoing shortwave radiation	$W\ m^{-2}$	
T_a	Air temperature	$^{\circ}C$	
T_d	Debris temperature	$^{\circ}C$	
T_D	Dew point temperature	$^{\circ}C$	
T_i	Temperature of debris/ice interface	$^{\circ}C$	
T_o	Melting temperature of ice	K	
T_R	Rain temperature	$^{\circ}C$	
W_D	Wind direction		
W_S	Wind speed	$m\ s^{-1}$	
α	Albedo	%	
α_d	Debris temperature	%	13
α_i	Ice albedo	%	34
α_s	Snow albedo	%	90
ϵ	Emissivity		0.622
ϵ_a	Atmospheric emissivity		
$\epsilon_a(\text{day})$	Atmospheric emissivity during day		
$\epsilon_a(\text{night})$	Atmospheric emissivity during night		
ϵ_d	Debris emissivity		0.94
ϵ_i	Ice emissivity		0.97
ϵ_s	Snow emissivity		1.0
ϵ_S	Surface emissivity		
ρ_a	Density of air	$Kg\ m^{-3}$	
ρ_d	Density of debris	$Kg\ m^{-3}$	1445
ρ_i	Density of ice	$Kg\ m^{-3}$	850
ρ_s	Density of fresh snow	$Kg\ m^{-3}$	85
ρ_{sp}	Density of fresh snow pack	$Kg\ m^{-3}$	610
ρ_w	Density of water	$Kg\ m^{-3}$	1000

Chapter – 1

INTRODUCTION

A Glacier is a perennial body of ice and snow, in which ice from a higher elevation is transported to a lower elevation. The movement of the ice is always from the upper part of the glacier towards the lower end. Glaciers are basically the ice mountain but unlike typical mountains these mountains can move.

Glaciers are made up of fallen snow that, over many years, compresses into large and thick ice masses. Glacier are formed when snow remain in one place before transforming into fern and then ice. Due to sheer mass, glaciers flow like very slow rivers. Presently, glaciers occupy about 10 % of the world's total land area with most located in Polar Regions like arctic and Antarctica.

1.1 Formation of Glacier

Glacier began to form when snow remains in the same area year-round, where enough snow collects to transform in to ice. 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 (Fig. 1.1). Gradually the grains grow larger and air trapped between the two start getting less in amount causing the snow to slowly to become compact in shape and thus increase its density. After about two winters, the

snow turns into firn –an intermediate stage between snow and glacier ice. Over time it becomes more compressed.

Glaciers are dynamic, and several elements contribute to glacier formation and growth. All glacier start with snowfall that accumulates faster than it melts. As the years pass, layers of snow pile up, compressing the layers below into dense ice. As the snow slowly accumulates and turns to ice and the glacier increases in weight due to this increase in weight the ice began to deform forcing the glacier to flow downhill towards lower altitude. This process continues until the mountain of ice becomes very heavy. It starts to move when the weight becomes too heavy to maintain the glacier shape.

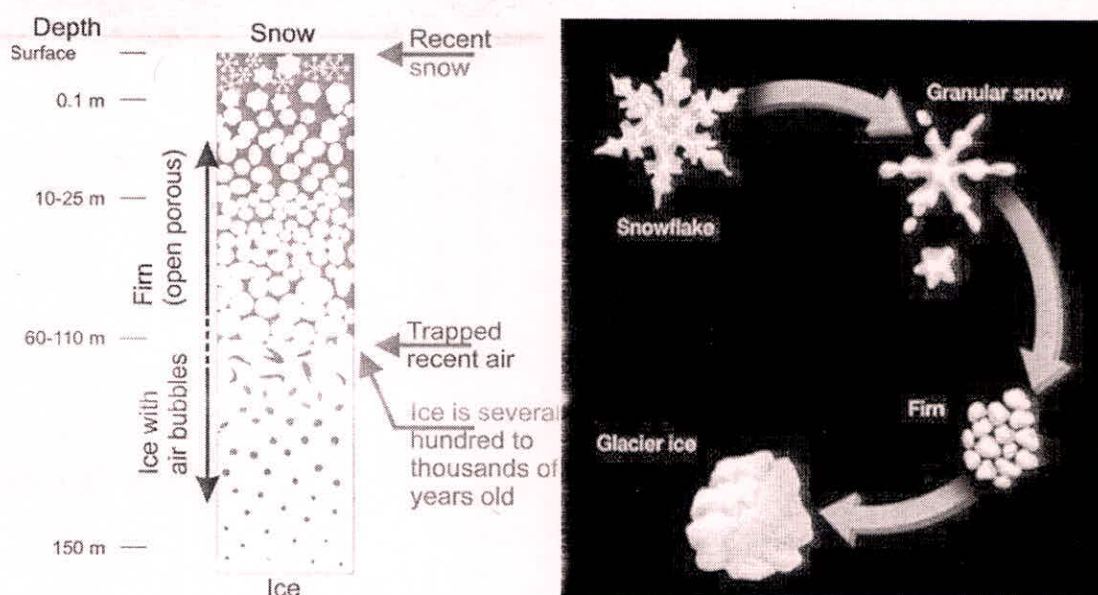


Fig. 1.1: Formation of glacier.

1.2 Classification of Glacier

A glacier in a way, is a river of ice, flowing from higher to lower levels, but unlike a river, the glacier movement is very slow and a glacier, under pressure at its back, can move upslope also.

1.2.1 Continental Glacier

The continental glacier is used for the glaciers that exist and originate out of vast ice caps and ice fields like that of Antarctica and Greenland (Fig. 1.2a). 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

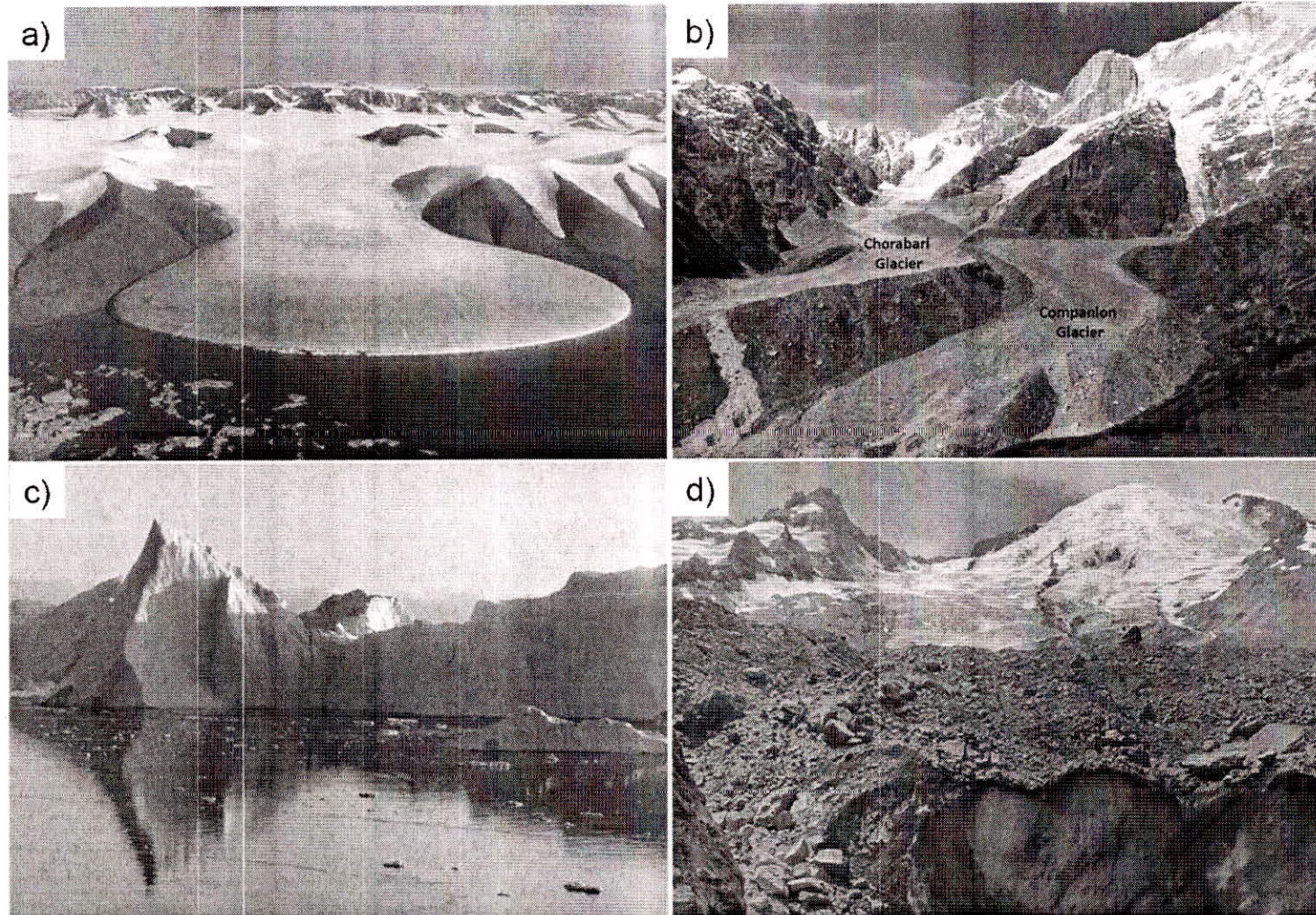


Fig. 1.2: a) Continental Glacier, b) Valley Glacier (Source: Kesarwani, 2015), c) debris-free (clean) glacier and d) debris-free (clean) glacier.

Hundreds of kilometres. The largest glaciers are continental ice sheets or icecaps, enormous masses (greater than 50,000 km²) of ice found only in Antarctica and Greenland. These sheets contain vast quantities of fresh water. If the Greenland ice sheet melted, it would cause sea levels to rise some 20 feet (6 meters) all around the world. If the Antarctic ice sheet melted, sea levels would rise up to about 200 feet (60 meters).

1.2.2 Valley (Mountain) Glacier

Glaciers that occur in the mountain regions of the world, like Himalayas, belong to the Alpine variety. They are, in reality Mountain glaciers that are often termed as the Valley glaciers. These glaciers develop in high mountainous regions in well-developed cirques, from where they flow in linear fashion along narrow troughs or valleys (Fig. 1.2b). The valley or the depression occupied by such glaciers may have been carved out by it, or it may have occupied an existing river valley that it later modified.

Valley glaciers are streams of flowing ice that are confined within steep walled valleys, often following the course of an ancient river valley. The downward erosive action of the ice carves the valley into a broad U shape, in contrast to the steeper V shape that is produced during the early stages of erosion by rivers. A U shape valley with a flat floor is good evidence of the past glaciation of an area. Often several corrie glaciers will combine to feed a single valley glacier. Valley glaciers may join and form larger glaciers with much greater erosional power than they had as smaller individuals. As the main glacier erodes its valley downwards more quickly than surrounding valleys, they are left behind, finally becoming hanging valleys. Rock debris falls from the steep valley walls and litters the glacier surface. This occurs particularly along the edges, where rock debris builds up and is known as lateral moraine. Where two glaciers meet and flow into each other, the two edges where the meeting occurs become the centre of the new glacier. The two lateral moraines combine to form a medial moraine.

1.2.3 Glacier on the basis of size and shape

Glaciers are also classified by their geographic location as ice sheets, ice shelves, ice caps, mountain glaciers, valley glaciers, piedmont glaciers, cirque glaciers, hanging glaciers, and tidewater glaciers. Each of these types of glaciers has unique characteristics. Ice sheets are the massively thick glaciers that covered much of the planet during the Ice Age. They occupy broad areas of gentle terrain, but sometimes grow so thick that they cover the underlying features. Ice sheets generally flow outwards

in all directions from a central accumulation zone. A modern example is the Antarctic ice sheet. Ice shelves are simply the floating sections of ice sheets – Greenland and Ellesmere Island have a number of such floating ice shelves. Smaller ice sheets which are restricted to mountain plateaus, are known as ice caps, such as the Penny Ice Cap on Baffin Island. These sheets, in turn, give rise to ice streams, or outlet glaciers, which flow down the valleys between the rocky peaks. From the air, these ice streams look very similar to river drainage systems. If an ice stream from a mountainous area dumps onto a wide, flat plain, rather than the ocean, a broad lobate form of glacier known as a piedmont develops. Cirque glaciers are found high on mountainsides and tend to occupy bowl-like hollows. Also called ice aprons, hanging glaciers cling to steep mountainsides. Like cirque glaciers, they are wider than they are long. Hanging glaciers are common in the Alps, but not in the Arctic, because mountain ranges are steeper. As their name implies, tidewater glaciers are valley glaciers that flow far enough to reach the sea. They are responsible for calving numerous small icebergs.

Further glacier can be divided on the basis of appearance of its surface. On the basis of surface, they can be divide in to two parts debris-free/clean (1.2c) and debris-covered glacier (1.2d). Mostly glacier are likely to be appear as spectacular glittering white and blue mass of ice landing its way down a mountain valley. Some of them can clean ice glacier but some of them can be covered with the sediment rock particles which we call debris. 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. 1.2d). These glaciers are known as debris covered glacier and they are particularly common in mountain of high Asia.

Debris cover changes the way the glacier behaves in response to climate condition. The dark surface of the debris means that higher proportion of incoming solar radiation is absorbed at the surface of the debris than at the surface of bright exposed glacier ice. When the debris cover is thick, the energy which is being absorbed by the surface is used to heat the rock material and much of this heat is re –released to the atmosphere at night, so the much less of the energy delivered to the surface reaches the underlying ice and consequently melting rates are reduced compared to exposed ice.

1.2.4 Classification on the basis of thermal state

On the basis of thermal state there are three types of glacier – Polar glacier and temperate glacier (Fig. 1.3). A glacier is called a temperate glacier if it is at melting point throughout the year, from its surface to its base. A glacier is called polar glacier if it is always below the freezing point from the surface to its base. In temperate glacier snow repeatedly freezes and thaws, changing into granular ice called firn. This granular ice fuses into denser firn. These layers of firn undergo compaction and thus form glacial ice.

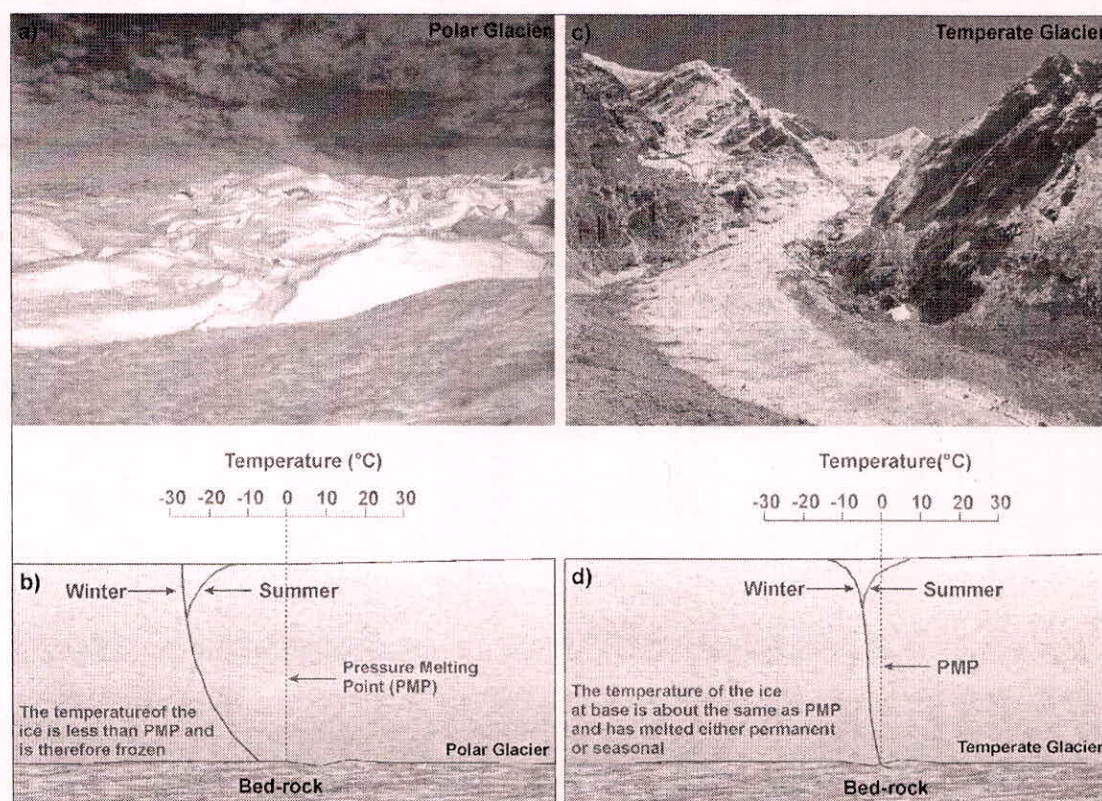


Fig. 1.3: Temperature mechanism of Polar and Temperate glaciers (Source: Kesarwani, 2015).

1.3 Glacial Geomorphology

1.3.1 Horn

Where three or more cirques of the adjacent watersheds intersect and encroach, frost action and ice abrasion leads to the conversion of the serrated ridges upwards into a pyramidal or triangular faceted peak. Such peaks arise above the general level and in glacial geomorphology are called 'Horns' (Fig. 1.4a).

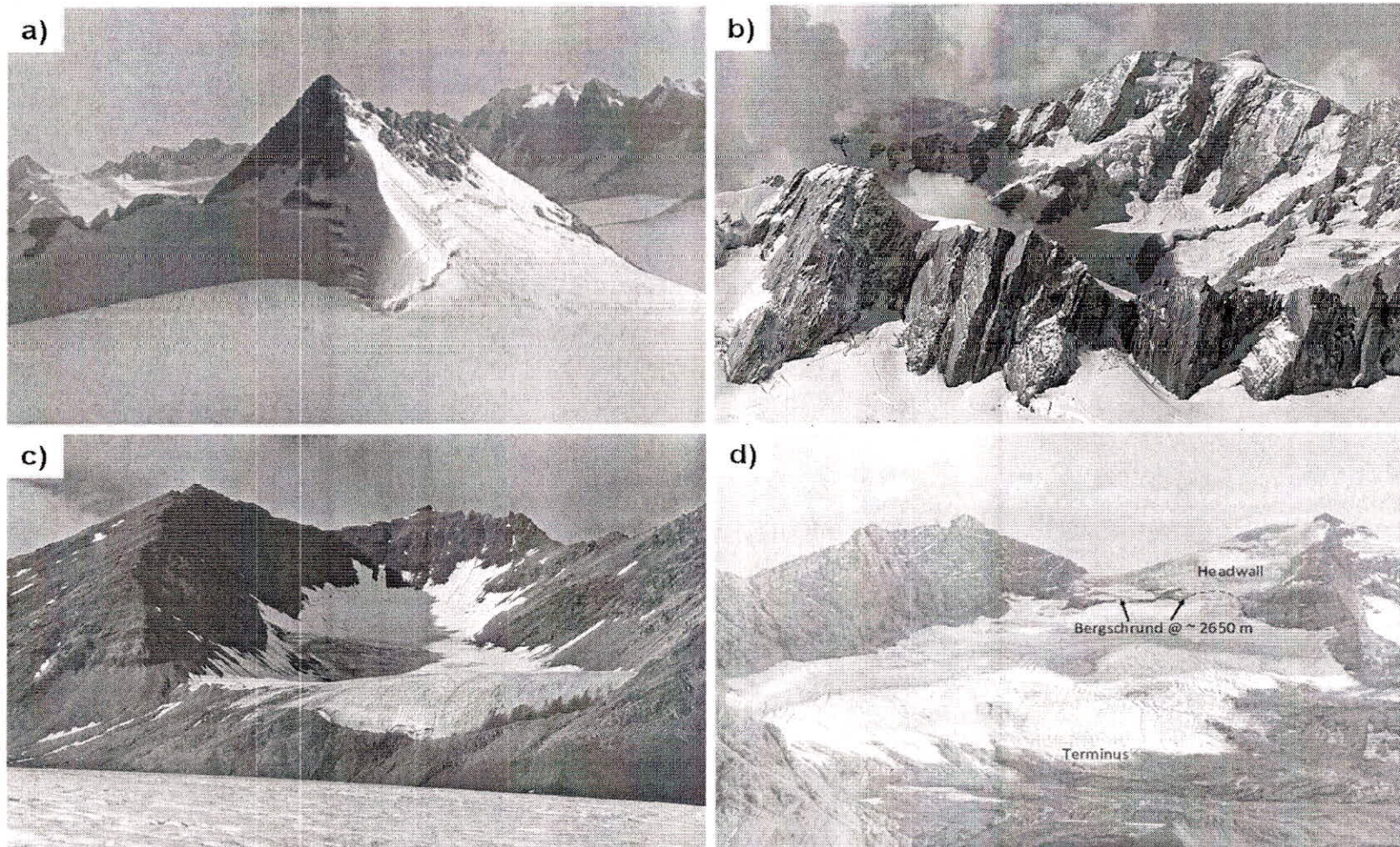


Fig. 1.4: a) Horn structure in mountain glacier, b) Arête structures in mountain glacier c) Cirque formation in mountain glacier d) Bergschrund line in Sperry Glacier, Montana, USA

1.3.2 Arête

Mountain ridges-water shed-separating the two adjacent cirques generally present a sharp edged-saw toothed or comb like- appearance. Such ridges are called Arêtes (Fig. 1.4b). An arête is a thin, crest of rock left after two adjacent glaciers have worn a steep ridge into the rock. A horn result when glaciers erode three or more arêtes, usually forming a sharp-edged peak. Cirques are concave, circular basins carved by the base of a glacier as it erodes the landscape.

1.3.3 Cirque

Cirque is amphitheatre-shaped basin with precipitous walls, at the head of a glacial valley (Fig. 1.4c). It generally results from erosion beneath the bergschrund of a glacier. A bergschrund is a large crevasse that lies a short distance from the exposed rock walls and separates the stationary from the moving ice; in early summer it opens, exposing the rock at its base to diurnal changes of temperature. Frost action then causes rapid disintegration of lower rock, which causes the upper rock to avalanche and produce an almost vertical head wall. Resulting rock material is embedded in the glacier and scours a concave floor, which may contain a small lake (tarn) if the glacier disappears. Expansion of neighbouring cirques produces sharp arêtes, cols, and horns. Because glaciers must originate above the snowline, a survey of the elevations of ancient cirques provides information on climatic change and on the former position of the snow line.

1.3.4 Bergschrund line

A bergschrund is a crevasse that separate stagnant ice or firn from moving glacier ice (Fig. 1.4d). Bergschrund line can extend to the bedrock and can have depth of 100 m. bergschrund 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. It often extends depth wise right to the cirque floor.

1.3.5 Accumulation zone

On a glacier, the accumulation zone is the area above the firn line, where snowfall accumulates and exceeds the losses from ablation, (melting, evaporation, and sublimation). The annual Glacier equilibrium line separates the accumulation and ablation zone annually. The accumulation zone is also defined as the part of a glacier's

surface, usually at higher elevations, on which there is net accumulation of snow, which subsequently turns into firn and then glacier ice (Fig. 1.5). Part of the glacier where snow builds up and turns to ice moves outward from there.

The glacier system receives snow and ice through processes of accumulation. Surface accumulation processes include snow and ice from direct precipitation, avalanches and windblown snow. The snow and ice is then transferred down slope as the glacier flows. Precipitation falling as rain is usually considered to be lost to the system. Internal accumulation may include rain and melt water percolating through the snowpack and then refreezing.

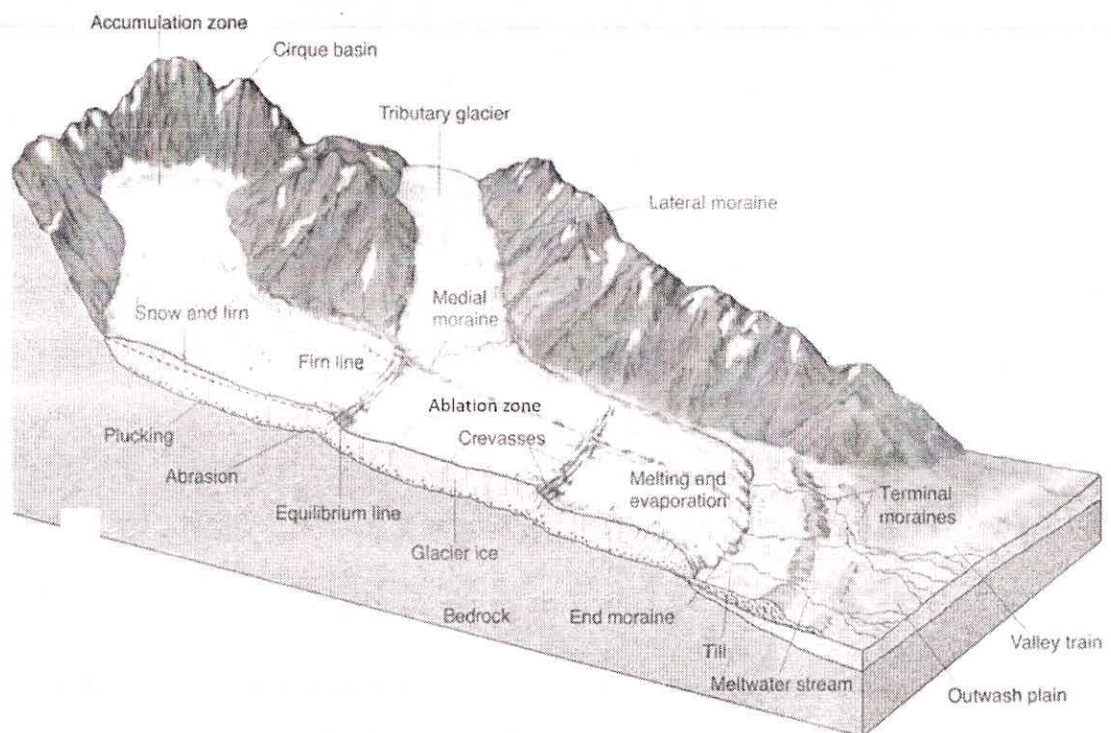


Fig. 1.5: A typical diagram explaining the surface characteristics of a valley glacier.

1.3.6 Equilibrium-Line-Altitude (ELA)

The equilibrium-line altitude (ELA) on glaciers is the average elevation of the zone where accumulation equals ablation over a 1-year period. The ELA can rarely be observed as a line at the same elevation across the entire width of the glacier due to local topographic and climatic variations in accumulation and ablation (Fig. 1.5). Thus, the ELA is the average altitude of the equilibrium line. The ELA is very closely related to the local climate, particularly winter precipitation and summer air temperature.

Variations in the ELA can therefore commonly be attributed to changes of these two variables. This can best be illustrated by its relationship to the net balance.

1.3.7 Ablation zone

Ablation zone or ablation area refers to the low-altitude area of a glacier or ice sheet below firn with a net loss in ice mass due to melting, sublimation, evaporation, ice calving, Aeolian processes like blowing snow, avalanche, and any other ablation (Fig. 1.5). The equilibrium line altitude (ELA) or snow line separates the ablation zone from the higher-altitude accumulation zone. The ablation zone often contains meltwater features such as supraglacial lakes, englacial streams, and subglacial lakes. Sediments dropped in the ablation zone forming small mounds or hillocks are called kames. Kame and kettle hole topography is useful in identifying an ablation zone of a glacier. The seasonally melting glacier deposits much sediment at its fringes in the ablation area. Ablation constitutes a key part of the glacier mass balance.

Ablation is the natural removal snow or ice from the surface of glacier or snowfield. This can occur through melting or sublimation and at or near the snout of glacier. Glaciers lose mass through processes of ablation. Surface ablation processes include surface melt, surface melt water runoff, sublimation, avalanching and windblown snow. Glaciers terminating in the sea or a lake will calve a photogenic iceberg. Other processes of ablation include subaqueous melting, and melting with in ice and at the ice bed.

1.3.8 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. 1.6a). 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 face. The length and width of a crevasse depends upon the size of the glacier and the gradient of the bedrock slope.

1.3.9 Moulin

During the summer season, glacier melt streams that flow over the glacier surface, often erode, vertically, the entire thickness of the glacier and give rise to deep and narrow pipes down which the melt stream flows. This pipe like feature is termed as Moulin (Fig.

1.6b). 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. Moulin are parts of the internal structure of glaciers, which carry melt water from the surface down to wherever it may go. 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. 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.

Water from Moulin may help lubricate the base of the glacier, affecting glacial motion. Given an appropriate relationship between an ice sheet and the terrain, the head of water in a Moulin can provide the power and medium with which a tunnel valley may be formed.

1.3.10 Supra-glacial Lake

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 warrant the name lake. A supra glacial lake is a water body of relatively substantial dimensions that at times can be seen even in an imagery. Such water body may continue to be present on the glacier surface for years to come till the water does not completely drain out (Fig. 1.6c).

1.3.11 Snout

Terminus of a glacier is called snout (Fig. 1.6d). It is a part of the ablation zone. Snout front of the glaciers in the Himalayas exhibit varying shapes and characters, depending upon the size of the glacier, nature of the valley and of course mass balance of the glacier. Large sized glaciers generally exhibit a broad-based snout front occupying the valley from one end to another with the melt water stream flowing out at either of the two flanks. Glacier front, highly weathered, is generally buried under the thick cover of surface moraines.

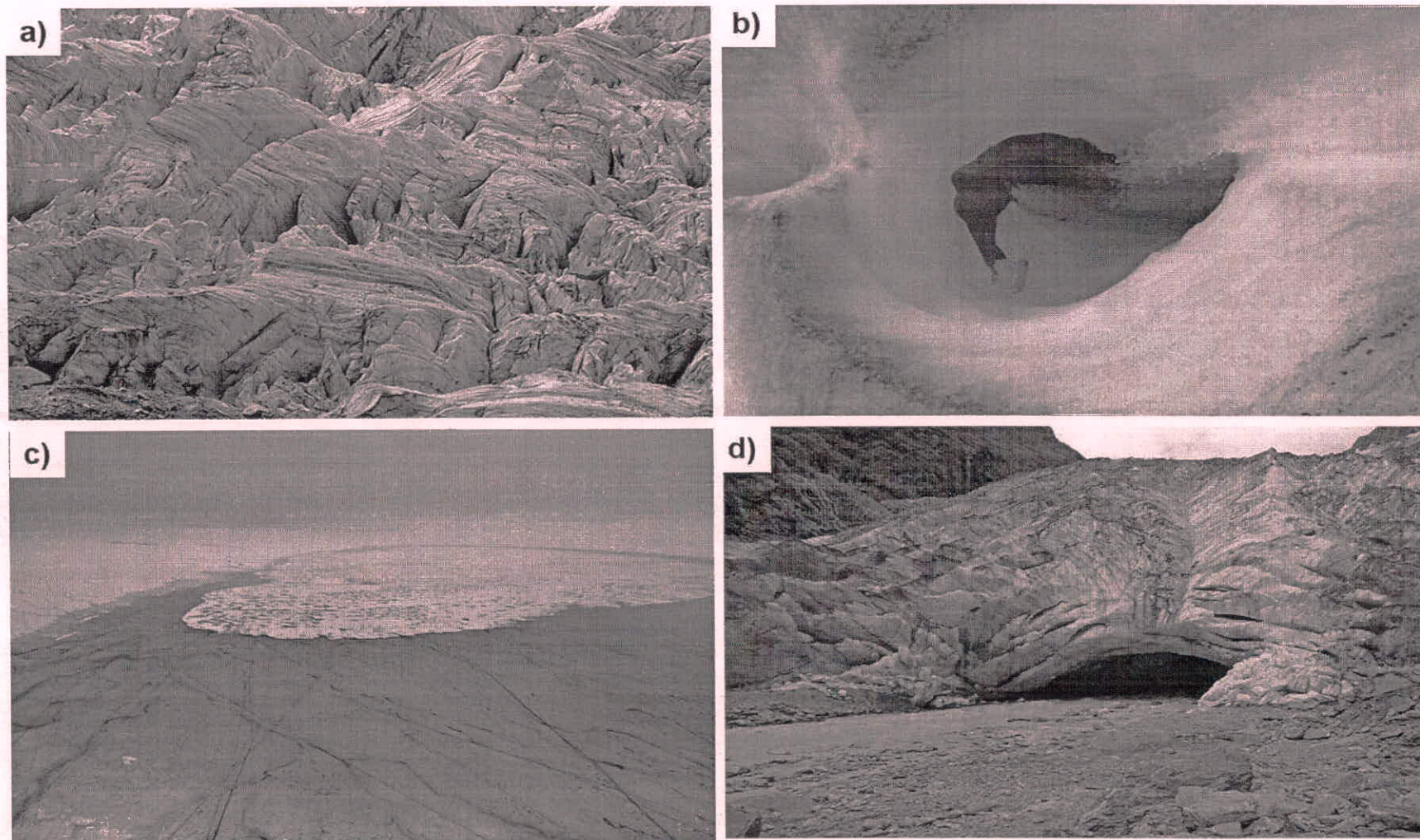


Fig. 1.6: a) A view of longitudinal crevasses, b) An image showing Moulin structure, c) supra glacial lake, d) A view of snout of glacier.

1.3.12 Moraine

Glacial moraines are formed by the deposition of material from a glacier and are exposed after the glacier has retreated (Fig. 1.7a). They usually appear as linear mounds of till, a non-sorted mixture of rock, gravel and boulders within a matrix of a fine powdery material. Terminal or end moraines are formed at the foot or terminal end of a glacier. Lateral moraines are formed on the sides of the glacier. Medial moraines are formed when two different glaciers merge and the lateral moraines of each coalesce to form a moraine in the middle of the combined glacier.

1.3.13 Proglacial Lake

A proglacial lake is a lake formed either by the damming action of a moraine or ice dam during the retreat of a melting glacier, or by meltwater trapped against an ice sheet due to isostatic depression of the crust around the ice (Fig. 1.7b). At the end of the last ice age approximately 10,000 years ago, large proglacial lakes were a widespread feature in the northern hemisphere.

1.3.14 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. More rounded material may also be present, raised from the glacier bed or reworked by surface streams (Fig. 1.7c).

1.3.15 Supra Glacial Channel

Supraglacial (surface) water on a glacier is formed by the ice melting during the summer. 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 (Fig. 1.7d). Often under the enormous pressure of the melt water so released englacial conduits and channels develop in the body of the glacier.

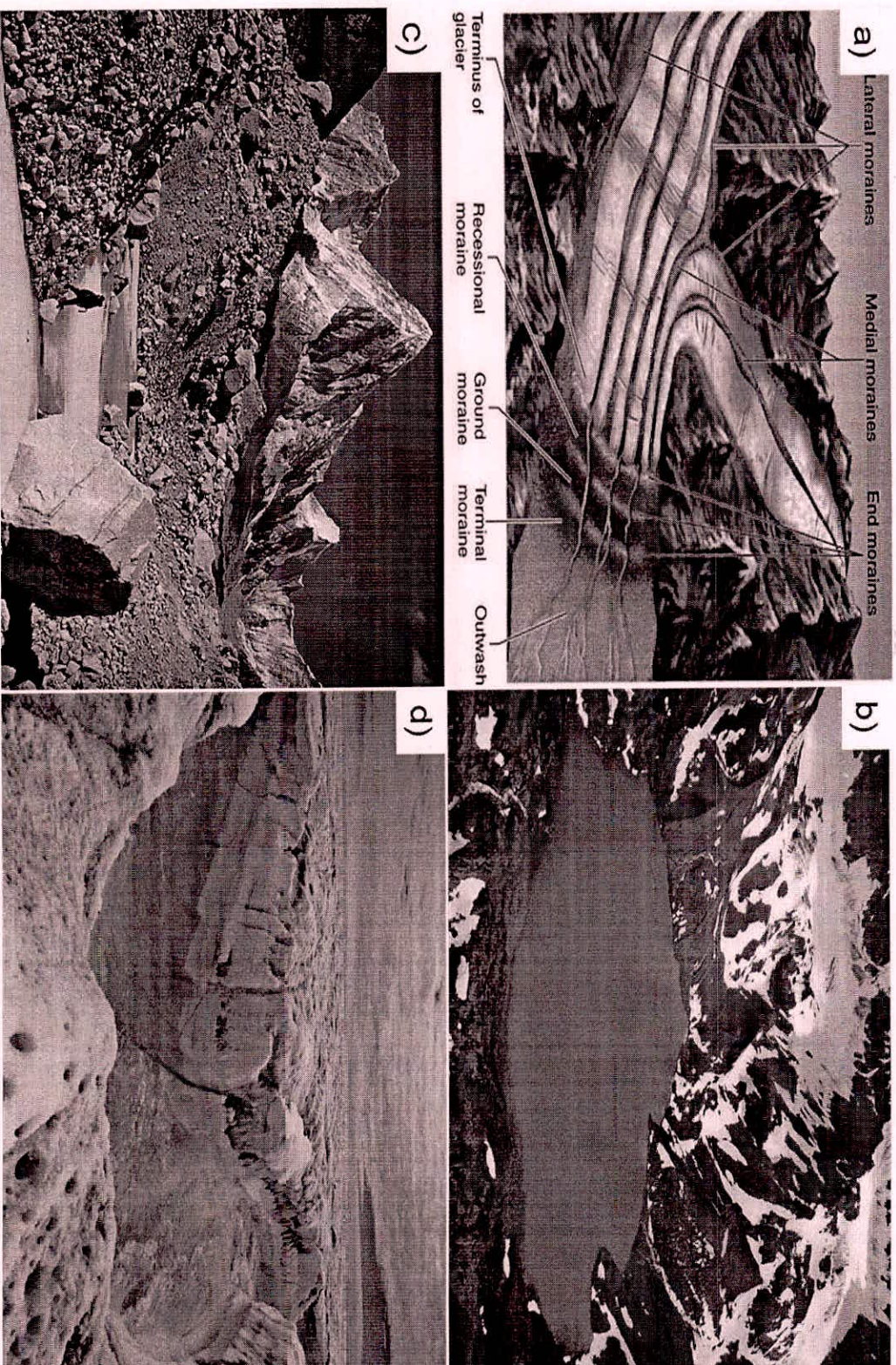


Fig. 1.7: a) Glacial moraines above Lake Louise, Alberta, Canada b) Pro-glacial Lake, c) Supra-Glacial Debris, d) Supra Glacial Channel.

1.4 Research Motivation

Glaciers are the most important source of the fresh water on the planet Earth. Glaciers are not only a perennial fresh water resource but considered as the highly sensitive indicator of the highly sensitive global climate change. Therefore, studies of glaciers are principally important for – i) Water Resources, ii) Climate Change and, iii) Environment (Kesarwani, 2015). It is a well-known fact that over the past decade due to anthropogenic activities such as burning of fossil fuels on an enormous scale have caused increase in the amount of carbon-dioxide in atmosphere resulting trapping of more heat (IPCC, 2007). This result increased globally air and surface temperature showing rapid retreat rate of glaciers. As global warming continues to increase the atmospheric temperature, it will lead to a continuous shift of zero temperature line (snow line) towards higher altitudes. Thus, glaciers will receive more liquid precipitation and less monsoonal snow precipitation. A shift in the snow line will result in lesser input to the glacier mass balance during summer periods. Therefore, higher atmospheric temperatures and more liquid precipitation at higher altitudes will lead to a rapid retreat of glaciers (Raina, 2009; Bolch et al., 2012). The melting fresh water from glaciers alters the ocean, by directly contributing to the global sea level rise. According to many scientific studies, sea level worldwide is increasing at a very rapid rate, due to which many people living in coastal areas are vulnerable to frequent floods (Chen et al., 2013; IPCC, 2013).

The way in which glaciers behave, whether they advance or retreats they will affect the environment directly or indirectly (Kesarwani et al., 2015). The behaviour of glaciers can be easily observed by defining the meteorological parameters which will lead to energy balance estimation. The energy balance of the glaciers is the key control of the interaction between glaciers and climate and a key step in the study of the mass balance of glaciers (Kesarwani, 2015). It provides the accurate information about the approximate melt runoff from the glacier surface which varies according to the conditions of climate. In addition, it also infers the possible mass changes over the glacier surface. 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 energy balance or thermodynamical changes of the 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. Therefore, to adopt the 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.

Here is an attempt has been made to study the basics of glaciology, how Physics works in natural system and how can we learn to implement the science in the field of glaciology where there is no ideal condition and everything keeps changing continuously. Therefore, main objective of this dissertation is to study the Thermodynamical changes of the debris-covered Himalayan Glaciers.

1.5 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 *Surface Energy Budget of Debris-covered Glacier*

This chapter focuses on the study of Debris Energy Balance model. It provides comprehensive information of the turbulent sensible and latent heat fluxes, heat flux supplied by rain, energy for melting or refreezing of Snow and subsurface heat flux using heat conduction through debris layers.

Chapter 4 *Instrumentation and Data Acquisition*

This chapter describes the instruments used for the study of meteorological parameters.

Chapter 5 *Thermodynamical changes of Debris-covered Chorabari Glacier Surface*

This chapter describes the climatic conditions of Debris-cover Chorabari Glacier, Meteorological observations and the analysis of collected meteorological data. It provides information about cloud cover, longwave radiation and budget estimation for DEB model.

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 focusses 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

Glacier Distribution

2.1 World Glacier Distribution

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. The Randolph Glacier Inventory (RGI) is a compilation of all the World's glaciers. It is the first global catalogue of glaciers, and it was developed to help IPCC scientists improve estimates of sea level rise). There are 1,98,000 glaciers in the World (RGI, 2014). However, this is a slightly arbitrary quantity, as it depends on the quality of the digital elevation model used, mapping resolution, and the minimum-area threshold used. Most analysts use a minimum area threshold of 0.1 km²; they will not map anything smaller than this due to difficulties in distinguishing between glaciers and snow packs. If these small glacierets are including, the number of glaciers in the World could be up to 400,000, but they would still only account for 1.4 % of the World's glacierised area. Together, these glaciers cover 7,26,000 km². The region with the most ice is the Antarctic and Sub-antarctic, with 1,32,900 km², closely followed by Arctic Canada North (1,04,900 km²).

At the other end of the scale, New Zealand has only 1160 km² of ice. In total, 44 % of the World's glacierised area is in the Arctic regions, and 18 % is in the Antarctic and Sub-antarctic. Glaciers cover 0.5% of the Earth's land surface.

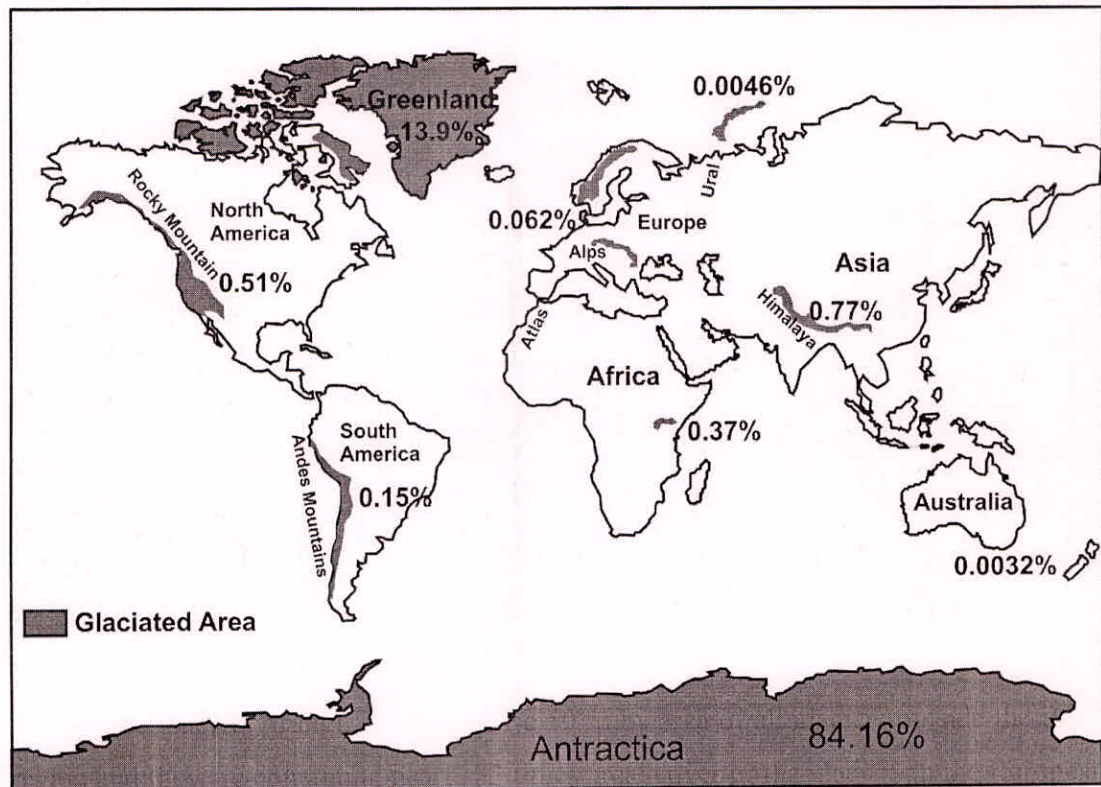


Fig. 2.1: Distribution of Glaciers around the world.

2.2 Himalayan Glacier System

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). 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).

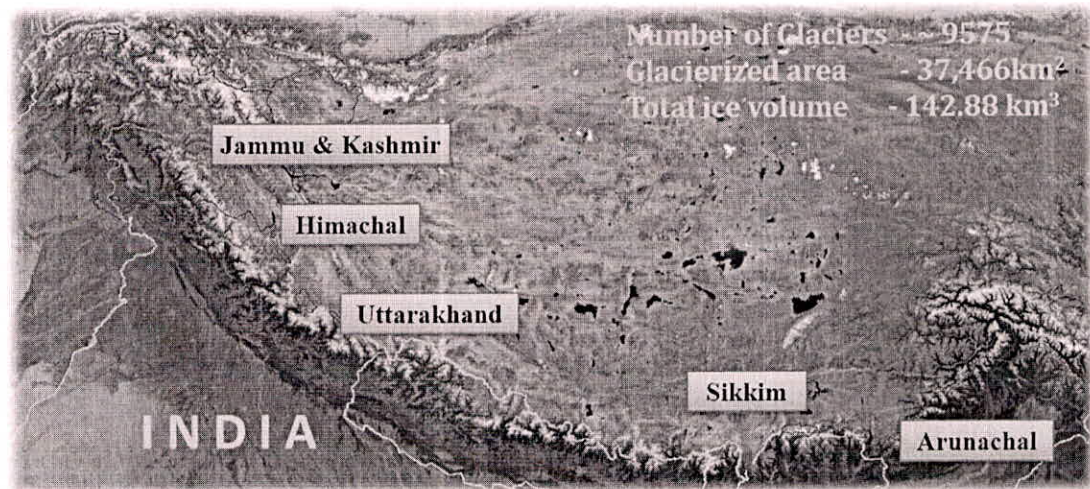


Fig. 2.2: Indian Himalayan Glacier system.

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>			<i>5344</i>	<i>29,213</i>	<i>889.07</i>
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>			<i>2653</i>	<i>4466</i>	<i>286.22</i>
<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>			<i>968</i>	<i>2,857</i>	<i>209.37</i>
<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>			<i>610</i>	<i>929</i>	<i>49.61</i>
Total (Indus + Ganga + Brahmaputra)			9,575	37,465	1,434.38

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

Chapter – 3

Surface Energy Budget of Debris-covered Glacier

3.1 Introduction

In the complex relationship between glaciers and climate, one of the key processes is melting of snow and ice at the glacier surface (Cuffey and Paterson, 2010). Especially on the mountain Glaciers, melting followed by runoff accounts for the most of the ablation. The total melt over a course of a year generally increases in warmer climatic conditions particularly during the summer, resulting the surface mass balance turns strongly negative. In warmer climate, an increase melt drives glacier retreat but air temperature alone does not determine the surface melt. There are some other parameters like radiative fluxes, surface temperature, wind speed, humidity, precipitation, clouds, air pressure and surface characteristics that play a noteworthy role. All of these parameters not only influence the melt directly but also influence the sensitivity of glacier melt to warming (Cuffey and Paterson, 2010). To understand the cause of glacier surface ablation by melting and sublimation, we need to examine the ways in which the glacier surface gains and losses heat under different weather conditions (Kesarwani, 2016).

The surface Energy Balance (SEB) of a glacier describes a very close physical relationship between snow/ice ablation and climate forcing (Haidong et. al., 2006; Pellicciotti et al., 2008). Melting at the glacier surface is governed by the SEB. It infers

the potential mass change over the glacier surface and provides the information about the potential mass change over the glacier surface and provides the information about fairly accurate discharge from surface melting which varies according to the conditions of climate.

3.2 Debris Energy-Balance (DEB) Model

The energy balance at the debris of Chorabari Glacier is determined by the sum of fluxes at the atmospheric/glacier boundary using the following relation (Nicholson and Benn, 2006; Brock *et al.*, 2010; Ried and Brock, 2010)-

$$S \downarrow + S \uparrow + L \downarrow + L \uparrow + H_s + H_L + H_p + H_G = 0 \quad (3.1)$$

where, $S \downarrow$ represent the incoming shortwave or global radiation, $S \uparrow$ is outgoing (reflected) shortwave radiation, $L \downarrow$ is incoming longwave radiation, $L \uparrow$ is outgoing longwave radiation, H_s and H_L are the net sensible heat flux and the net latent heat flux due to turbulent mixing of air adjacent to the surface, H_p is net heat flux due to precipitation and H_G is net conductive heat flux beneath the surface. Energy fluxes which are directed towards the glacier surface defined as positive and negative when directed away from surface except outgoing or reflected radiations (both shortwave and longwave) which were measured positive but assigned negative as they were always directed away from the surface the surface (Fig. 3.1). The cumulative flux of four radiations component (incoming and outgoing shortwave and longwave) was measured directly using net radiometer which measures the algebraic sum of incoming and outgoing all-wave radiation (i.e. both shortwave and longwave components). Since, the cumulative flux reduces errors in radiation measurement which is confirmed in subsequent studies (e.g. Long *et al.*, 2000; Mölg and Hardy, 2004; Sicart *et al.*, 2010), thus Eqn (3.1) can be expressed as-

$$\sum_{i=a}^b (R_n + H_s + H_p + H_G) = 0 \quad (3.2)$$

where, a and b are the number of days and R_n is the net radiation. In Eqn (3.2), the net radiation is expressed as-

$$R_n = S_n + L_n \quad (3.3)$$

where, S_n is the net shortwave radiation (sum of incoming and outgoing shortwave radiation) and L_n is the net longwave radiation (sum of incoming and outgoing longwave radiation).

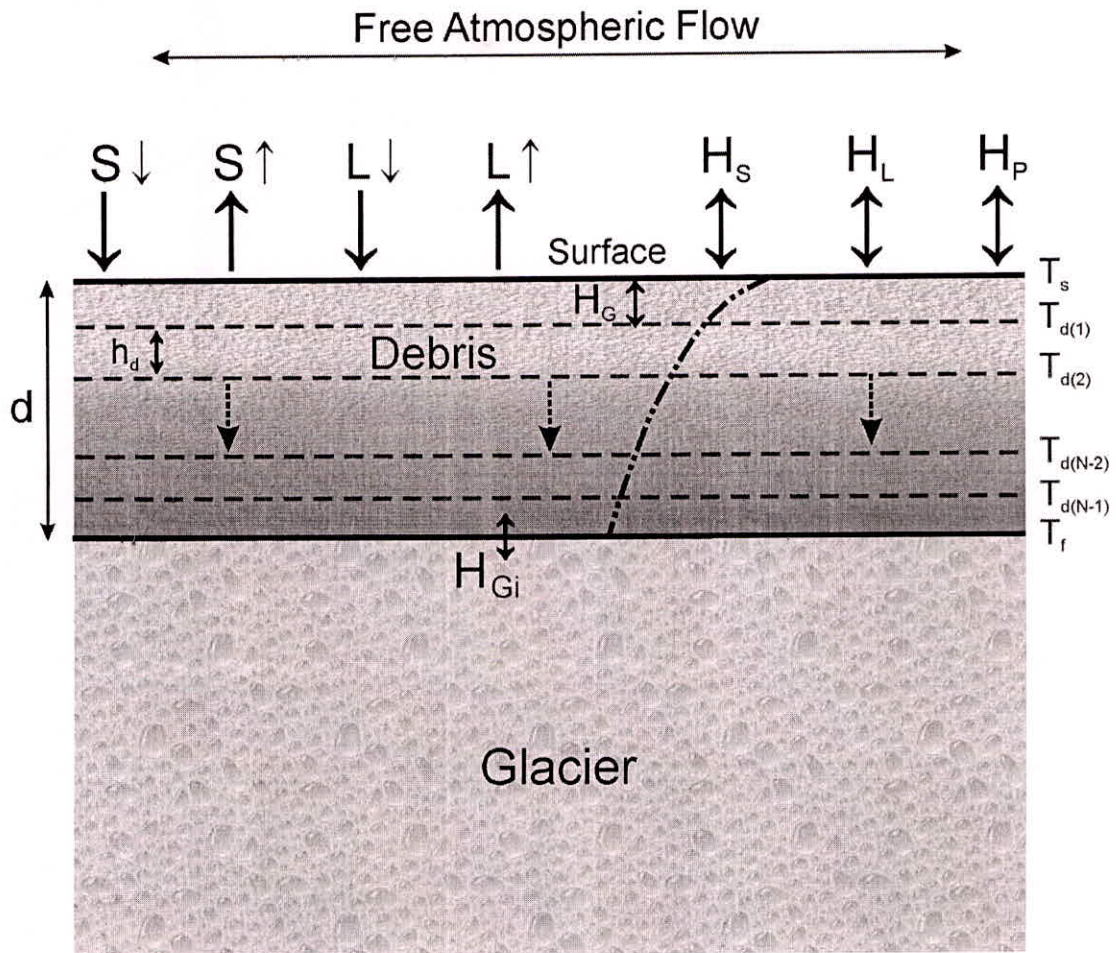


Fig. 3.1: Schematic of the DEB model showing heat fluxes at the top and bottom of a debris layer of thickness (d). The debris temperature (T_d) is calculated for N layers of thickness h_d , with boundary conditions defined by the surface temperature (T_s) and the temperature of the debris/ice interface (T_f), which is assumed to stay at 273.15 K. The dash-dot curve is a temperature profile, where temperature increases towards the right. H_G is conductive heat flux at the debris surface and H_{Gi} is a conductive heat flux considered to reach the glacier ice (Source: Kesarwani, 2015).

S_n and L_n are expressed as-

$$S_n = S \downarrow + S \uparrow \quad (3.4)$$

$$L_n = L \downarrow + L \uparrow \quad (3.5)$$

In Eqn (3.2), H_p is calculated by integrating the H_R (heat flux supplied by the rain) and H_{snow} (Energy required to melt or refreeze a volume of snow).

$$H_p = H_R + H_{\text{snow}} \quad (3.6)$$

All energy fluxes used in Eqn (3.1 to 3.6) are measured to estimate in W m^{-2} .

3.3 Radiative Fluxes

3.3.1 Shortwave Radiation Flux

Solar radiation is the key component of the energy budget at the earth's surface and also at glacier atmospheric interface. The amount of energy originating from the sun falling at normal incidence outside the earth's atmosphere, at the average distance of the earth from the Sun, is called Solar Centre (WRC) for it is 1367 W m^{-2} . A value of 1353 W m^{-2} has been accepted by the National Aeronautics and Space Administration (NASA) as standard. The Earth's surface always receives less direct solar radiation than the solar constant because of the process of absorption, reflection and transmission through the atmospheric path between the surface and top of the atmosphere due to atmospheric components (particles, water vapour, gases and clouds). Indeed, the attenuation, absorption, transmittance and scattering, either forward to the ground, back to the space or multiple scattering, are very complicated processes. However, the backscatter or outgoing shortwave radiation is highly influenced by the surface characteristic (e.g. snow cover, properties supraglacial debris, clean ice etc.) of glacier. Value of outgoing longwave radiation varies with wavelength and can be described as simple fraction of incoming shortwave radiation.

$$E_s^\uparrow = \alpha_s E_s^\downarrow \quad (3.7)$$

where α_s is the surface albedo and its value for snow and ice surfaces range from 0.2 to 0.85. The net shortwave radiation is strongly controlled by the surface albedo, as it controls the amount of global radiation that is absorbed by the surface, and thus is available for conversion to melt energy at the snow or ice surface. Since surface energy

balances during the ablation season are commonly dominated by the solar radiation, glacier surface melts are highly sensitive to albedo variation (Cutler and Munro, 1996).

3.3.2 Longwave Radiation Flux

The incoming longwave radiation (4.0-100 μm) is also a key parameter of the surface energy budget (Niemelä *et al.*, 2001). The incoming longwave radiation is a function of the vertical profiles of atmospheric temperature and greenhouse gases, including water vapour, of the distribution of clouds and their properties, and of temperature and emissivity of slopes that cover the upper hemisphere basin (Gruella *et al.*, 1997). This is justified to some extent by the fact that much of the longwave radiation received at the surface is emitted by the surface (Gruella and Genthon, 2004). Konzelmenn *et al.*, (1994) showed that about 30% of the longwave radiation reaching the surface is emitted by the lowest 10 m, about 60% by the lowest 100 m and about 90% by the lowest 1000m. Due to generally low air temperatures at the glaciers surfaces, the net flux of longwave radiation tends to be negative, except when the cloud cover is high.

Incoming Longwave Radiation -

Contribution of incoming longwave radiation ($L\downarrow$) to the SEB is calculated with standard meteorological variable using the Stefan-Boltzmann relation-

$$L\downarrow = \sigma \times \epsilon_a \times T_a^4 \quad (3.8)$$

where σ is the Stefan-Boltzmann constant ($5.67 \times 10^{-8} \text{ W m}^{-2} \text{ K}^{-4}$), ϵ_a is the atmospheric emissivity and T_a is the air temperature (K). While estimating $L\downarrow$, a true parametrization of atmospheric emissivity plays an important role (Crawford and Duchon, 1999). Therefore, ϵ_a is calculated using a physically rigorous formula developed by Butsaert (1975) during the day time (0700-1700 h; IST) which express the relationship between ϵ_a and meteorological parameters as-

$$\epsilon_{a(\text{day})} = (1 - C_F) + C_F \times 1.24 \times \left(\frac{e}{T_a} \right)^{\frac{1}{7}} \quad (3.9)$$

where C_F is cloud factor (non-dimensional) and e is vapour pressure (P_a). Estimation of C_F is given.

$$C_F = 1.3 - 1.4 \times \left(\frac{S \downarrow}{S_{TOA}} \right)$$

Where, S_{TOA} is Extraterrestrial Radiation. In Eqn (3.9), the vapour pressure (e) is calculated using the relation-

$$e = 6.11 \times 10^{\left[\frac{7.5 \times T_D}{237.7 + T_D} \right]} \quad (3.10)$$

where, T_D is dew point temperature at 2m above the surface and is evaluated using the relation –

$$T_D = \left[\left(\frac{R_H}{100} \right)^{\frac{1}{8}} \times \{112 + (0.9 \times T_a)\} \right] + (0.1 \times T_a) - 112 \quad (3.11)$$

The method presented in Eqn (3.11) is restricted for calculation of ε_a only during day time when $S \downarrow$ is present on the glacier surface. Therefore, for better accuracy, ε_a during the night time (1800-0600 h; IST) is also calculated using T_D according to the equation (Chen et al., 1995) -

$$\varepsilon_{a(night)} = 0.736 + 0.0057 \times T_D \quad (3.12)$$

where $\varepsilon_{a(night)}$ is the air emissivity during the night time.

Outgoing Longwave radiation

The outgoing longwave radiation ($L \downarrow$) is calculated by the following equation-

$$L \uparrow = R_n - S_n - L \downarrow \quad (3.13)$$

where R_n is the net radiation, S_n is the net shortwave radiation and $L \downarrow$ is the incoming longwave radiation.

3.4 Turbulent Heat Fluxes

There are two more processes that also transfer the heat between the Earth's surface and the atmosphere, and which are relevant for glacier melting, namely the turbulent fluxes of sensible and latent heat. In the first case, eddies conduct heat to the surface if this is colder than the air immediately above, whereas the surface loses heat if it is warmer than the air. In the second case, the surface gains latent heat when moisture evaporates from it

(Paterson, 1994). The amount of heat transferred by the processes, although controlled respectively by the temperature and humidity gradient, increases with the degree of turbulence in the atmosphere, hence the name of turbulent fluxes.

The simplest approach to calculate the turbulent sensible and latent heat flux is the 'Bulk Aerodynamics' method. In this method, the mean gradient of horizontal wind speed, mean air temperature and mean specific humidity are assumed to be equal to the finite difference between the measurement level and the surface (Klok and Oerlemans, 2002; Zhang *et al.*, 2013). In this case, the surface roughness and wind profiles are unknown over the snow and ice, constant bulk coefficients ($C_s=C_L=0.002$; Kondo and Yamazawa, 1986) can be used to calculate the turbulent heat flux as recommended by the Fujita and Ageta (2000) and Sakai *et al.*, (2009). Here, C_s and C_L are the bulk exchange coefficients for latent heat fluxes, respectively.

The turbulent fluxes are important component of the glacier surface energy balance. Measurement made on various glaciers have shown that in fact they contribute 20-40% of the total energy balance (e.g. Mölg and Hardy, 2004; Reid and Brock *et al.*, 2010; 2013), and the sensible fluxes are often the second main energy sources after the solar radiation (Bhutiyani, 1989; Pellicciotti *et al.*, 2008; Giesen *et al.*, 2009; Azam *et al.*, 2014).

3.4.1 Sensible heat flux

The turbulent sensible heat flux (H_s) is calculated using bulk aerodynamic method. Considering the gradients of mean of T_a , horizontal W_s to be equal to the finite difference between the measurement level and the surface, H_s is evaluated using the relation-

$$H_s = \rho_a \times C_p \times W_s \times C_s \times (T_a - T_s) \quad (3.14)$$

where ρ_a is air density (kg m^{-3}), C_p is specific heat capacity of air ($1005 \text{ J kg}^{-1} \text{ K}^{-1}$), q_a is specific humidity of air (kg kg^{-1}), C_s is the bulk exchange coefficients for H_s and T_s is the air/debris interface surface temperature. In this study, the surface roughness and wind profiles are unknown over the snow and debris. Therefore, $C_s=0.002$ (Kondo and Yamazawa, 1986) is used as suggested by Fujita and Ageta (2000) and Sakai *et al.* (2009)

Air density

The air density (ρ_a) is determined according to gas law-

$$\rho_a = \left(\frac{P_a \times M_a}{R_g \times T_a} \right) \quad (3.15)$$

where, P_a is measured atmospheric pressure (P_a) at 2m above the surface, M_a is molar mass of dry air ($0.002896 \text{ kg mol}^{-1}$) and R_g is universal gas constant ($8.31 \text{ J K}^{-1} \text{ mol}^{-1}$).

3.4.2 Latent Heat Flux

The turbulent latent heat flux (H_L) is calculated using bulk aerodynamic method. Considering the gradients of means of horizontal W_s and specific humidity (q) to be equal to the finite difference between the measurement level and the surface, H_L is evaluated using the relation-

$$H_L = \rho_a \times L_F \times W_s \times C_L \times (q_a - q_s) \quad (3.16)$$

where ρ_a is air density (kg m^{-3}), L_F is either latent heat of vaporization ($2.514 \times 10^6 \text{ J kg}^{-1}$) or sublimation ($2.89 \times 10^6 \text{ J kg}^{-1}$) used here based on the surface temperature, C_L is the bulk exchange coefficients for H_L , q_a is specific humidity of air (kg kg^{-1}) and q_s is specific humidity of surface (kg kg^{-1}). In this study, the surface roughness and wind profiles are unknown over the snow and debris. Therefore, $C_L=0.002$ (Kondo and Yamazawa, 1986) is used as suggested by Fujita and Ageta (2000) and Sakai et al., (2009).

Specific Humidity

The specific humidity of air or surface ($q_{a/s}$) is calculated using the relation (Peixoto and Oort, 1996; Ross and Elliott, 1996)-

$$q_{a/s} = 1000 \times \left[\frac{\varepsilon \times e_{a/s}}{P_a - (1 - \varepsilon) \times e_{a/s}} \right] \quad (3.17)$$

where, ε is 0.622 and $e_{a/s}$ is vapour pressure (h Pa) of air or surface calculated using 5.9 and $P_{a/s}$ is the atmospheric pressure of air or surface. As the atmospheric pressure is

measured at 2m above the surface, therefore, for the near-surface atmospheric pressure it assumed same.

3.5 Heat Flux due to Precipitation

The heat flux due to precipitation is calculated by integrating the heat flux supplied by the rain (H_R) to the glacier surface and heat flux required to melt or refreeze a volume of snow (H_{snow}). These fluxes are calculated as-

3.5.1 Heat Flux due to Snow

Snow accumulation plays a significant role in changing the heat budget of a glacier, energy required for melting or refreezing a volume of snow (H_{snow}) during the winter and summer is calculated for enhancing the accuracy of model using the relation-

$$H_{snow} = h_s \times \rho_s \times L_w \quad (3.18)$$

where, h_s is the amount of snow accumulation (m), ρ_s is density of fresh snow (kg m^{-3}) and L_w is the latent heat of fusion of water ($334.864 \times 10^3 \text{ j kg}^{-1}$).

3.5.2 Heat flux due to Rain

The heat flux supplied by rain (H_R) is determine using the relation (Singh, 2001; Reid and Brock, 2010)

$$H_R = \rho_w \times c_w \times \left(\frac{R_o}{1000} \right) \times (T_R - T_s) \quad (3.19)$$

where, ρ_w is density of water (1000 kg m^{-3}), c_w is specific heat capacity of water ($4.2 \times 10^3 \text{ j kg}^{-1} \text{ } ^\circ\text{C}^{-1}$), R_o is the rainfall (mm) and T_R is the rain temperature ($^\circ\text{C}$). Here, T_R was considered to be equal to corresponding measured T_a during rainfall event.

3.6 Sub-surface Heat Flux

In the alpine environments, debris-covered glaciers are common such as the Himalaya, the Southern Alps of New Zealand and the Peruvian Andes (Benn *et al.*, 2012). The glacier surface ablation is strongly influenced by the presence of supraglacial debris

(Østrem, 1959; Nakawo and Young, 1982; Nicholson and Benn, 2006), so, in the case of debris-covered glaciers, predicting run-off and response to climate change requires different treatment to that of clean glaciers. The empirical relationships between the ice-melt rates and supraglacial debris thickness were first introduced by Østrem (1959). The study indicates that under the thin debris (< 2 cm), ablation rates were higher than for clean ice, and under thicker debris ablation rates gradually turn down; a pattern that has been confirmed in several subsequent studies (e.g. Loomis, 1970; Mattson *et al.*, 1993; Kayastha *et al.*, 2000; Nicholson and Benn, 2006).

Conductive heat flux

The conductive heat flux (H_G) at the debris surface is calculated based on the temperature gradient at the top of the debris layer (Reid and Brock, 2010)-

$$H_G = k_d \left(\frac{dT_d}{dz} \right)_{\text{surface}} \approx k_d \frac{T_d(1) - T_s}{h_d} \quad (3.20)$$

where, k_d is thermal conductivity of debris ($\text{W m}^{-1} \text{K}^{-1}$), $T_d(1)$ is debris temperature at the first calculation layer for the debris and h_d is the thickness of each layer (usually set to 1 cm). Thermal conductivity of debris, $k_d = 0.94 \text{ W m}^{-1} \text{K}^{-1}$ was applied when snow was not present on surface and thermal conductivity of snow, $k_s = 0.16 \text{ W m}^{-1} \text{K}^{-1}$ was applied when snow was present.

Debris surface Temperature

The air/debris interface surface temperature, T_s is a key variable for the SEB of glaciers particularly for debris-covered glaciers (Reid and Brock, 2010). To obey laws of conservation of energy, T_s is considered to change on every time-step to a temperature which will cause the sum of heat fluxes at the debris surface to be zero. Since in Eqn (3.1) some of the fluxes are functions of T_s while some are not, debris surface temperature is calculated by adopting the methodology suggested by Reid and Brock (2010)-

$$S \downarrow + S \uparrow + L \downarrow + L \uparrow(T_s) + H_s(T_s) + H_L(T_s) + H_P(T_s) + H_G(T_s) = 0 \quad (3.21)$$

or simply:

$$F(T_s) = 0 \quad (3.22)$$

Where, $F(T_s)$ is the total surface flux. In the DEB model, we use an iterative Newton – Raphson method according to-

$$T_s(n+1) = T_s(n) - \frac{F(T_s(n))}{F'(T_s(n))} \quad (3.23)$$

Where $F'(T_s(n))$ is the derivative of the total surface flux with respect to T_s which is calculated numerically by the central difference method.

Debris Internal Temperature

If there is a nonzero gradient of temperature with respect to depth, or when some solar shortwave flux penetrates in to snow/ice, the flux due to debris internal temperature is significant. This conduction can be calculated from a heat conservation equation based on Fourier's law, taking in to account the partial derivatives of temperature, $T_{i,d,s}$ with respect to time (t) and depth (z)-

$$\rho_{i,d,s} C_{i,d,s} \left(\frac{\partial T_{i,d,s}}{\partial t} \right) = k_{i,d,s} \times \left(\frac{\partial^2 T_{i,d,s}}{\partial z^2} \right) + \frac{\partial Q}{\partial z} \quad (3.24)$$

Where, $\rho_{i,d,s}$, $C_{i,d,s}$, $T_{i,d,s}$ and $k_{i,d,s}$ are density(kg m^{-3}), specific heat capacity ($\text{J kg}^{-1} \text{K}^{-1}$), temperature (K) and thermal conductivity ($\text{W m}^{-1} \text{K}^{-1}$), respectively of glacier ice (i), supraglacial debris (d) and snow (S) that are used in DEB model based on the surface/Layer condition, and $\partial T_{i,d,s} / \partial t$ is the average rate of temperature change (K s^{-1}) where $T_{i,d,s}$ is the mean temperature of ice/snow/ debris layers.

Here, the measured density, standard specific heat capacity and thermal conductivity of ice ($\rho_i = 850 \text{ kg m}^{-3}$; $C_i = 2.05 \times 10^3 \text{ J kg}^{-1} \text{K}^{-1}$; $k_i = 2.22 \text{ W m}^{-1} \text{K}^{-1}$) and snow ($\rho_s = 85 \text{ kg m}^{-3}$; $C_s = 2.09 \times 10^3 \text{ J kg}^{-1} \text{K}^{-1}$; $k_s = 0.16 \text{ W m}^{-1} \text{K}^{-1}$) are used. On the right hand side of the Eqn (3.24), first term represents, subsurface heat conduction whereas, second term represents radiative fluxes (Q), penetrating in to ice, snow or debris layer from above. Eqn (3.24) is solved by dividing the debris layer into several calculation layers and applying a numerical algorithm with boundary conditions defined by surface temperature (T_s) and the temperature of the debris-ice interface, which is in contact with the glacier ice and is assumed to remain at the melting point of ice, during ablation season.

Chapter-4

Instrumentation and Data Acquisition

4.1 Automatic Weather Station

The long and continuous meteorological records can be easily obtained by using AWS. An automatic weather station (AWS) is an automated version of the traditional weather station, either to save human labour or to enable measurements from remote areas. An AWS will typically consist of a weather-proof enclosure containing the data logger, rechargeable battery, telemetry (optional) and the meteorological sensors with an attached solar panel or wind turbine and mounted upon a mast (Fig. 4.1). The specific configuration may vary due to the purpose of the system. The system may report in near real time via the Argos System and the Global Telecommunications System, or save the data for later recovery. In the past, automatic weather stations were often placed where electricity and communication lines were available. Nowadays, the solar panel, wind turbine and mobile phone technology have made it possible to have wireless stations that are not connected to the electrical grid or hard line telecommunications network



Fig. 4.1: Complete setup of Automatic Weather Station installed at Base camp site (3820 m a.s.l.) of the Chorabari Glacier, Garhwal Himalaya (Source: Kesarwani, 2015).

4.2 Air Temperature and Relative Humidity Probe (HC2S3)

The Air Temperature and Relative Humidity Probe (HC2S3) measures the air temperature and relative humidity (Fig. 4.2a). The probe contains two sensors- (i) 100 Ω PRT sensor for air temperature measurements and (ii) Rotronic's IN1 capacitive sensor to measure the relative humidity. The sensor measures the temperature in the range of -40°C to +60°C and relative humidity in the range of 0-100%. It is a digital probe with 0 to 1 V linear output signals for temperature and humidity. Relative humidity is defined by the equation below:

$$RH = \frac{e}{e_s} * 100 \quad (4.1)$$

where RH is the relative humidity, e is the vapour pressure in kPa, and e_s is the saturation vapour pressure in kPa. The vapour pressure, e , is an absolute measure of the amount of water vapour in the air and is related to the dew point temperature. The saturation vapour pressure is the maximum amount of water vapour that air can hold at a given air temperature. The relationship between dew point and vapour pressure, and air temperature and saturation vapour pressure are given by Goff and Gratch (1946), Lowe (1977), and Weiss (1977). Relative Humidity is relative to saturation above water, even below freezing point. This is why these sensors should not measure 100% RH below 0°C when the air temperature increases, so does the saturation vapour pressure. Conversely, a decrease in air temperature causes a corresponding decrease in saturation vapour pressure. It follows then from Eq. (4.1) that a change in air temperature will change the relative humidity, without causing a change absolute humidity.

For example, for an air temperature of 20°C and a vapour pressure of 1.17 kPa, the saturation vapour pressure is 2.34 kPa and the relative humidity is 50% (Table 4.1 & 4.2). If the air temperature is increased by 5°C and no moisture is added or removed from the air, the saturation vapour pressure increases to 3.17 kPa and the relative humidity decreases to 36.9%. After the increase in air temperature, there is more energy to vaporize the water. However, the actual amount of water vapour in the air has not changed. Thus, the amount of water vapour in the air, relative to saturation, has decreased. Because of the inverse relationship between relative humidity and air temperature, finding the mean relative humidity is often not useful. A more useful quantity is the mean vapour pressure

Table 4.1: Detailed specification of Temperature sensor

Specification	Range
Air temperature Measurement Range	-50 to +50°C
Thermistor Absolute Accuracy and interchangeability	± 0.1°C
Linearity	± 0.09°C
Maximum Uncertainty due to ±0.1% fixed resistors	±0.15°C @ -50°C ±0.9°C@0°C ±0.03°C @ 50°C
Response Time	10 sec in still air

Table 4.2: Detailed specification of RH sensor

Specification	Range
Probe length	25.4 cm (10 in.)
Probe Body Diameter	2.5 cm (1 in.)
Filter	0.2 µm Teflon membrane
Filter diameter	1.9 cm (0.75 in.)
Power consumption	<4 mA
Supply Voltage (via CSI switching circuit)	7 to 17 VDC
Settling Time	0.15 seconds
Relative Humidity Measurement Range	0 to 100% non-condensing
RH Output Signal Range	0.008 to 1 VDC
Accuracy at 20°C	±2% RH (0 to 90% Relative Humidity) ±3% RH (90 to 100% Relative Humidity)
Temperature Dependence of Relative Humidity Measurement	±0.05% RH/°C
Typical Long-Term Stability	Better than 1% RH per year
Response Time (at 20°C, 90% response)	15 seconds with membrane filter

or simply:

$$F(T_s) = 0 \quad (3.22)$$

Where, $F(T_s)$ is the total surface flux. In the DEB model, we use an iterative Newton – Raphson method according to-

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Where, $\rho_{i,d,s}$, $C_{i,d,s}$, $T_{i,d,s}$ and $k_{i,d,s}$ are density(kg m^{-3}), specific heat capacity ($\text{J kg}^{-1} \text{K}^{-1}$), temperature (K) and thermal conductivity ($\text{W m}^{-1} \text{K}^{-1}$), respectively of glacier ice (i), supraglacial debris (d) and snow (S) that are used in DEB model based on the surface/Layer condition, and $\partial T_{i,d,s} / \partial t$ is the average rate of temperature change (K s^{-1}) where $T_{i,d,s}$ is the mean temperature of ice/snow/ debris layers.

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Response Time	10 sec in still air

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Specification	Range
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Supply Voltage (via CSI switching circuit)	7 to 17 VDC
Settling Time	0.15 seconds
Relative Humidity Measurement Range	0 to 100% non-condensing
RH Output Signal Range	0.008 to 1 VDC
Accuracy at 20°C	±2% RH (0 to 90% Relative Humidity) ±3% RH (90 to 100% Relative Humidity)
Temperature Dependence of Relative Humidity Measurement	±0.05% RH/°C
Typical Long-Term Stability	Better than 1% RH per year
Response Time (at 20°C, 90% response)	15 seconds with membrane filter

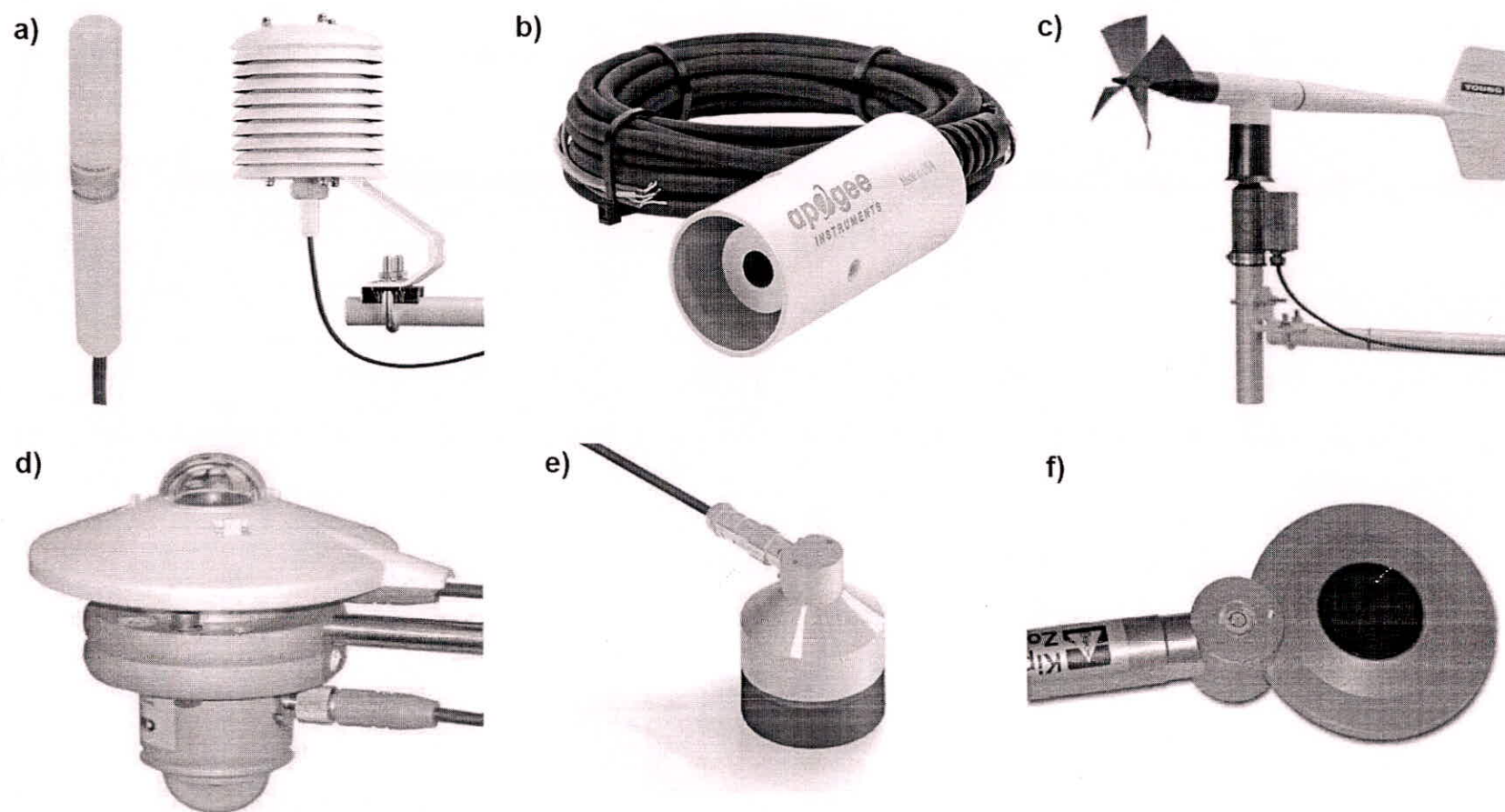


Fig. 4.2: a) Air temperature and Relative Humidity Probe (HC2S3), b) Infrared Surface Temperature Sensor (SI-111) c) Wind Monitor (05103) d) Pyranometer (CMP3-L) e) Sonic Ranging Sensor (SR50A) f) Net Radiometer (NR-LITE).

4.3 Infrared Surface Temperature sensor (SI-111)

The SI-111 is a precision infrared radiometer that determines the surface temperature of an object without physical contact (Fig. 4.2b). It measures both the subject's surface temperature and the sensor-body temperature. The SI-111 is an infrared temperature Sensor that provides a non-contact means of measuring the surface temperature of an object. It senses the infrared radiation being emitted by the target. SI-111 includes a thermopile for measuring a millivolt output dependent on the target to sensor body temperature difference. A thermistor measures the temperature of the sensor body. The sensor body temperature is used to reference the target temperature.

The SI-111 can be widely used for measurements of leaf, canopy, and average surface temperature. With contact sensors, it is difficult to avoid influencing the temperature, maintain thermal contact, and provide a spatial average. By mounting the infrared sensor at an appropriate distance from the target, it can be used to measure an individual leaf, a canopy, or any surface of interest.

4.4 Wind Monitor (05103)

The wind speed sensor is a helicoid-shaped, four blade propellers (Fig. 4.2c). Rotation of the propeller produces an AC-sine wave that has a frequency directly proportional to wind speed. The AC (alternating current)-signal is induced in a transducer coil by a six-pole magnet mounted on the propeller shaft. The coil resides on the non-rotating central portion of the main mounting assembly, eliminating the need for slip rings and brushes. Propeller is used for measuring the horizontal wind speed and direction (Table 4.3).

Table 4.3: Detailed specification of Wind monitor sensor.

Specification	Range
Wind speed sensor	
Range	0-100 m/s (0-224 mph)
Accuracy	± 0.3 m/s (± 0.6 mph)
Starting threshold	1.0 m/s (2.2 mph)
Distant constant (63% recovery)	2.7 m (8.9 ft)
Output	A/C Voltage (3 pulse per revolution) 1800 rpm 90 Hz=8.8 m/s (19.7 mph)

Wind direction sensor	
Range	0-360° Mechanical, 0-355° Electrical (5° open)
Accuracy	± 3°
Starting threshold at 10° displacement	1.1 m/s (2.2 mph)
Delay distance (50% recovery)	1.3 m (4.3 ft)
Damping ratio	0.25
Damped natural wavelength	7.2 m (23.6ft)
output	Analog D/C voltage from 10 kohm

The wind monitors use a potentiometer (10 k Ω) to measure wind direction. With a precision excitation voltage applied, the output voltage is proportional to wind direction.

4.5 Pyranometer (CMP3-L)

The CMP3-L measures global solar radiation and it has a high quality blackened thermopile protected by a dome to protect it from external influence (Fig. 4.2d). The blackened thermopile provides a flat spectral response for the full solar spectral range, which allows the CMP-3 to be used under plant canopies or lamps, when the sky is cloudy and for reflected radiation measurements (Table 4.4).

Table 4.4: Detailed specification of Pyranometer.

Specifications	Range
ISO Classification	Second class
Response time (95%)	18s
Zero offset	
(a) Thermal radiation (200 W/m ²)	± 15 W/m ²
(b) Temperature change (5 K/hr)	± 5 W/m ²
Non-stability (change/year)	± 1%
Non-linearity (0 to 1000 W/m ²)	± 2.5 %
Directional error	
(0 to 80° with 1000 W/m ² beam)	± 20 W/m ²
Temperature dependence of sensitivity	± 5 % (-10 to +40° C)
Tilt error (at 1000 W/m ²)	± 3%

The CMP3 produces a millivolt signal that is measured directly by datalogger. The upward pyranometer is used to measure the incoming shortwave radiation whereas, the downward pyranometer is used to measure the outgoing shortwave radiation.

4.6 Sonic Ranging Sensor (SR50A)

The SR50A Sonic Ranging Sensor measures the distance from the sensor to a target. The most common applications are measuring snow depths and water levels. The sensor is based on a 50 kHz (Ultrasonic) electrostatic transducer (Fig. 4.2e). The SR50A determines the distance to a target by sending out ultrasonic pulses and listening for the returning echoes that are reflected from the target. The time from transmissions to return of an echo is the basis for obtaining the distance measurement. The SR50A is capable of picking up small targets or targets that are highly absorptive to sound, such as low-density snow. The SR50A makes use of a unique echo processing algorithm to help ensure measurement reliability. If desired, the SR50A can also output a data value indicative of measurement quality. The SR50A was designed to meet the stringent requirements of snow depth measurement which makes it well suited for a variety of other applications. The rugged aluminium housing is built to withstand harsh environments and offers several mounting options (Table 4.5).

Table 4.5: Detailed specification of Sonic ranging sensor.

Specification	Range
Power Requirements	9-18 Volts D.C.
Power Consumption	Quiescent: SDI-12 mode <1.0mA RS-232/RS485 mode <2.0mA Active: 250 mA typical
Measurement Time	Less than 1.0 second
Measurement Range	0.5 to 10 meters (1.6 to 32.8 ft)
Depth Accuracy	±1 cm (.4 in) or 0.4% of distance to Sensor target (whichever is greater) Requires external temperature compensation or SR50AT version of the sensor (with temperature probe)
Resolution	0.25mm (.01 in)
Operating Temperature	-45°C to +50°C (-49°F to +122°F)

4.7 Net Radiometer (NR-LITE)

The NR-LITE is used for measuring solar and far infrared radiation balance (Fig. 4.2f). This balance is known as the net (total) radiation. Its upwards facing sensor measures the solar energy and far infra-red energy that is received from the entire hemisphere (180° field of view). Its downwards facing sensor measures the energy received from the surface of the soil. The two readings are automatically subtracted and the result converted to a single output signal. This output represents the net radiation, (which can be interpreted as meaning the radiative energy that is seen at the surface) and is expressed in Watts per square meter (W m^{-2}).

The NR-LITE is designed for continuous outside use. The sensor surfaces are coated with Teflon. This ensures good sensor stability, very long life and easy maintenance compared to the more usual radiometers fitted with plastic domes. However, it does have some disadvantages, particularly a higher sensitivity to wind speed with a subsequent lessening of accuracy (Table 4.6). It is, though, possible to correct for the wind speed sensitivity if the sensor is installed in a system where wind speed is also being measured. Although net radiometers are usually used in meteorology to measure radiation balance, the NR-LITE can also be used to measure indoor climate radiative stress.

Table 4.6: Detailed specification of NR-LITE

Specifications	Range
Sensor	Blackened thermopile
Spectral response	0 to 100 μm
Response time (e-1)	20 seconds (nominal)
Sensitivity	10 $\mu\text{V W}^{-1} \text{m}^2$ (nominal)
Output range	$\pm 25 \text{ mV}$
Range	$\pm 2000 \text{ W m}^{-2}$
Temperature range	-30° to 70°C
Weight	23 oz. (635 g)
Dimension	Sensor 3.1" (8.0 cm) diameter Support arm 0.6" (1.6cm) diameter x 31.5" (80 cm) L
Datalogger requirements	One differential or single-ended analog channel

4.8 Barometric Pressure Sensor (61302V)

Atmospheric pressure of the area is measured using Barometric Pressure Sensor (61302V). It is a versatile electronic barometer featuring high accuracy, low power, wide operating temperature range, and RS-232 serial data I/O or a calibrated voltage output (Fig. 4.3a). Low power consumption and wide temperature range make the 61302V ideal for remote applications using battery or solar power. The sensor measures the atmospheric pressure in the range of 500 to 1100 hPa (Table 4.7).

Table 4.7: Detailed specification of Barometric pressure sensor

Specification	Range
Pressure	500 to 1100 hPa standard range
Operating Temperature	-50 to +60 °C
Accuracy:	at 25°C ± 0.2 hPa *
Total Accuracy	± 0.3 hPa
Update Rate	3Hz
Serial Output:	Full duplex RS-232 9600 baud Polled or continuous ASCII text output
Analog Outputs:	0 to 5000 mV
Standard range:	500 to 1100 hPa Other ranges user settable
Resolution	Analog 0.1 hPa typical Digital 0.01 hPa typical
Power:	10 to 30 VDC at 3mA
Case:	Thermoplastic
Weight:	Net 0.2 lb

4.9 Rain Gauge (TE 525M)

The TE525 series is a tipping rain gauges (Fig. 4.3b). Rainfall measurements were carried out by using a Tipping Bucket Rain Gauge. These gauges funnel precipitation into a bucket mechanism that tips when filled to a calibrated level. The sensor measures rainfall in 0.1 mm increments (Table 4.8).

Table 4.8: Detailed specification of Rain gauge

Specifications	Range
Sensor type	Tipping bucket/magnetic reed switch
Material:	Anodized aluminum
Temperature:	0° to +50°C
Resolution:	1 tip
Cable:	2-conductor shielded cable
Rainfall per tip	0.004" (0.1 mm)
Orifice diameter	9.66" (24.5 cm)
Height:	11.5" (29.21 cm)
Weight:	2.7 lbs. (1.2 kg)
Accuracy:	
Up to 10 mm/hr:	±1%
10 to 20 mm/hr:	+0, -3%
20 to 30 mm/hr:	+0, -5%

4.10 Data Logger (CR-1000)

The CR1000 data logger is widely used to collect the data in long-term meteorological research, and routine weather measurement applications (Fig. 4.3c). The CR-1000 measures electrical signals and converts the measurement to engineering units, performs calculations and reduces data to statistical values. Most applications do not require that every measurement be stored but rather combined with other measurements in statistical or computational summaries. The CR1000 will store data in memory to await transfer to the PC with an external storage devices or telecommunications. CR1000 electronics are protected in a sealed stainless steel shell. This design makes the CR1000 economical, small, and very rugged (Fig. 4.2c). The CR1000 is operable with power from 9.6 to 16 V DC applied at the POWERIN terminals of the green connector on the face of the wiring panel. Datalogger support software are PC or Linux software available from Campbell Scientific that facilitate communication between the computer and the CR1000 (Table 4.9). A CR Basic program must be loaded into the CR1000 to enable it to make measurements, read sensors, and store data.

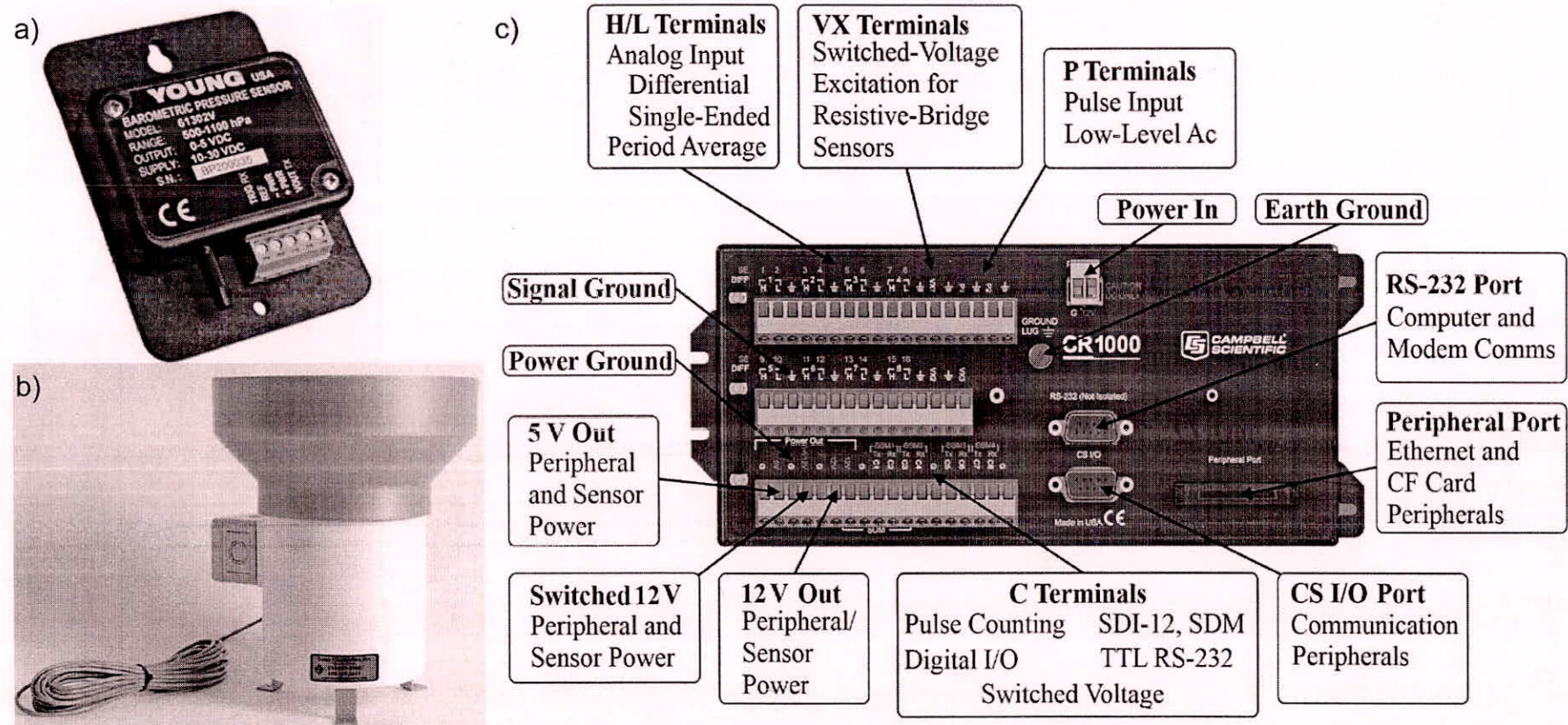


Fig. 4.3: a) Barometric Pressure sensor (61302V), b) Rain Gauge (TE 525M) and c) Setup of CR-1000 Data logger.

Table 4.9: Specification of Data Logger

Specification	Range
Analog inputs (SE1-SE16 or DIFF1-DIFF8)	8 differential (DF) or 16 single-ended (SE) individually configured
Accuracy	$\pm (0.125 \text{ of reading} + \text{offset})$, -25° to 50°C
Current sourcing	$\pm 25\text{mA}$
Maximum counts per scan	16.7×10^6
Maximum Input Frequency	250kHz
Maximum Input Voltage	$\pm 20 \text{ V}$

4.2 Data Acquisition

The meteorological parameters were collected using Automatic Weather Station (AWS) which was installed near the snout of Chorabari Glacier (3800 m a.s.l.). To collect the meteorological data at minute, hourly and daily timestamp, a standard coded programme was used for CR-1000 data logger. For getting the optimum results, all sensors are recalibrated annually as per the standards.

Air Temperature and Relative Humidity were measured with 1-minute sampling frequency. Errors were prevented by using radiation shield. Surface Temperature was measured with 1-minute sampling frequency. Wind speed and Direction were measured with 10 minute sampling frequency. To measure Radiative fluxes the pyranometer and net radiometer were installed on the southern side of the glacier valley to measure global radiation. Air Pressure was measured by a sensor which is fixed inside the Data logger. Total rainfall is measured by using the tipping bucket rain gauge at hourly scale which was mounted horizontally and 1m above ground level.

Chapter- 5

Thermodynamical changes of Debris-covered Chorabari Glacier Surface

5.1 Chorabari Glacier

Chorabari (7.5 km), a benchmark glacier of the Central Himalaya is located at latitude $30^{\circ}46'20.58''\text{N}$ and longitude $79^{\circ}2'59.38''\text{E}$ in Mandakini basin of the Alaknanda catchment (Fig.5.1). It is a medium-sized (6.66 km^2) compound valley type glacier flowing from North to South between 6420 and 3895 m a.s.l with an average slope of 20° (Kesarwani et al., 2015). The accumulation area (2.99 km^2) is formed by two cirque tributary glaciers below from Bhartkhunta (6578 m a.s.l.) and Kedar peak (6940 m a.s.l). A number of longitudinal and transverse crevasses are observed in the accumulation zone (6420-5100 m a.s.l) and near the Equilibrium-Line-Altitude (ELA; 5075 m a.s.l) as a consequence of steep slope (30°). However, the ablation area (3.67 km^2) is broad (0.1-0.4 km) with a gentle slope (10°) and covered with substantial amount of debris ranging from ~2 to 180 cm. The debris-covered area accounts for ~53% of total glacier area and debris thickness increases along the glacier flow. Several temporary small supraglacial lakes have been marked below 4200 m a.s.l in the ablation zone (5070-3895 m a.s.l). The glacier terminus (3895 m a.s.l) is wide and broad (<200 m) with bulky amount (>2.0 m) of debris in the middle portion. This glacier is the resource of the Mandakini River which ultimately joins the Alaknanda River near Rudraprayag (Uttarakhand).

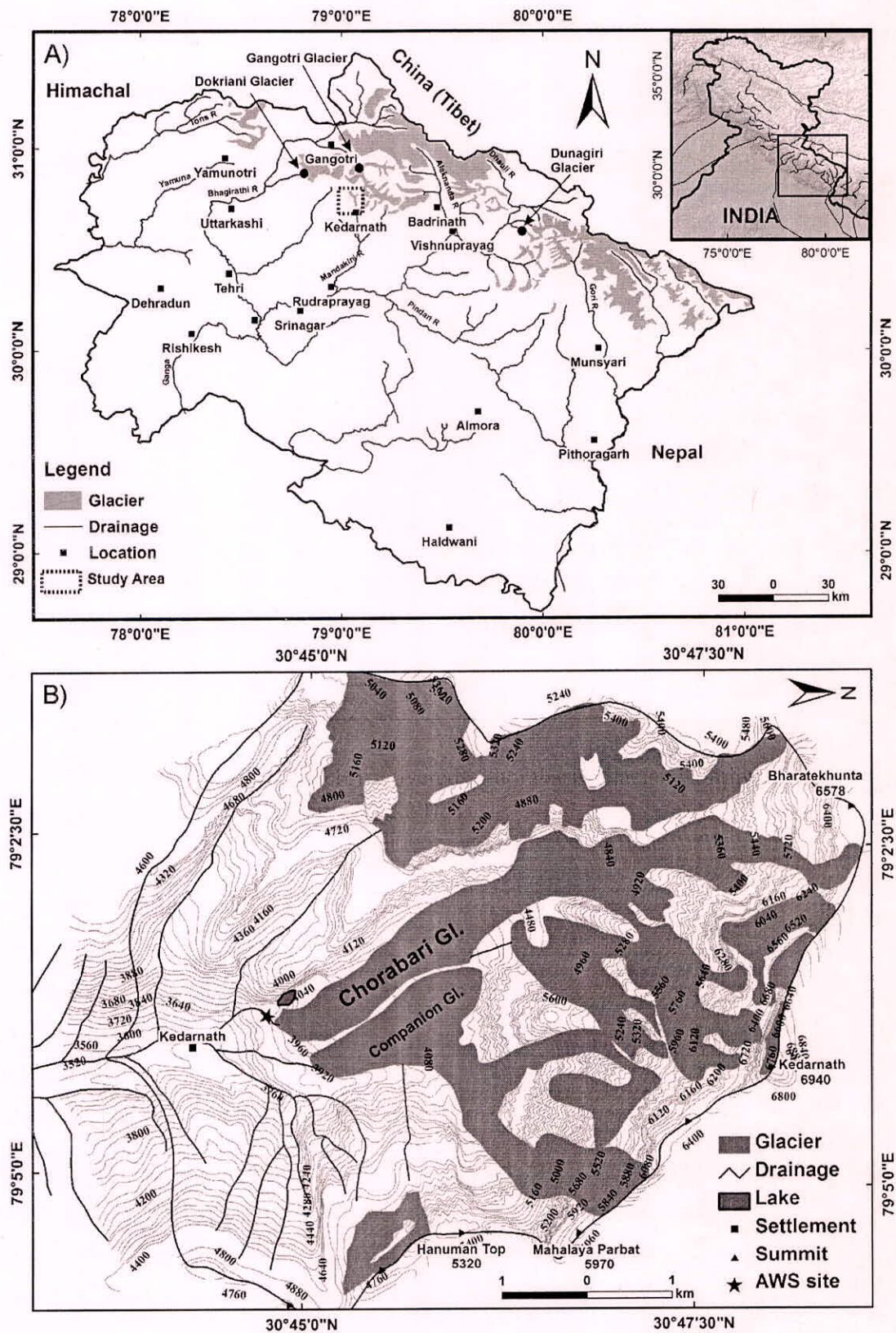


Fig. 5.1: Location of the Debris-covered Chorabari Glacier, Garhwal Himalaya, Uttarakhand. (Source: Kesarwani et al., 2015)

There is another glacier named the Companion Glacier (Length: 4.5 km; Area: 3.5 km²) flows parallel to the ablation zone of Chorabari Glacier ranging from 3810 to 4250 m a.s.l. Morphologically, it appears that originally Chorabari and Companion glaciers were a single glacier that got separated due to the shrinking. At present these glaciers are separated by medial moraine. Likewise, probably all the glaciers in the Western, Central and Eastern Himalaya, Chorabari Glacier has been losing mass over the past century and probably at an increasing rate with time. Since 1962 to 2010, Chorabari Glacier has retreated about 327 m with an average recession rate of 6.8 m a⁻¹ (Mehta *et al.* 2014). In addition, the average specific balance of Chorabari Glacier was -0.73 m w.e. a⁻¹ from 2003 to 2010 with fluctuating ELA between 5055 and 5075 m a.s.l. (Dobhal *et al.* 2013). Geologically, the area is situated north of the Pindari Thrust comprises of augen and granitic gneisses, calc-silicate, schist and granite pegmatite apatite veins belonging to the Pindari Formation (Valdiya *et al.* 1999; Mehta *et al.* 2012).

5.1.1 Climatic Condition

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. 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).

The Chorabari Glacier valley is nourished with Western-Disturbance (winter) and South-West monsoon (summer). Generally, climate of the area is dry cold in winter (November-April) and humid-temperate in summer (May-October). Based on the seasons and monsoon system, the glaciological year for the Dokriani Glacier is considered from 01 November to 31 October of the following year (accumulation period – November to April; ablation period – May to October) (Dobhal *et al.*, 2008; Pratap *et al.*, 2015).

5.1.1 Characteristics of Debris-cover

One of the most important characteristic features of the Chorabari Glacier is that its ablation zone is covered with thick layers of supraglacial debris. The supraglacial debris layer is found to be highly heterogeneous, from silt size to big boulders exceeding several meters. The supraglacial debris layer is composed of with different rock types that have different reflective (albedo) and thermal properties (thermal conductivity, thermal resistance, thermal inertia, emissivity etc.) consequently variable impacts on the surface energy balance of a glacier. Therefore, the knowledge and mapping of lithological composition of the supraglacial debris is of great importance as composition of debris will have direct bearing on energy interactions within the debris layer which in turn influence the surface velocity and ablation rates of the ice beneath the debris layer. Petrographic analysis on major rock type supraglacial debris collected from ablation zone carried out to identify the supraglacial debris characteristics. Results suggested that the supraglacial debris comprises predominantly of gneissic rock which includes augen gneiss, banded gneiss, felsic gneiss and fine-grained gneiss with some mixture of calc-silicate schist, meta-sedimentary, amphibolites, leucogranite and marble rocks.

5.2 Meteorological Observations

The meteorological data collected during the ablation season (01 May-31 October) of 2011-12 glaciological year used in the present study was reported by Kesarwani et al. (2015) for the Chorabari Glacier catchment. To collect the meteorological parameters an AWS was installed near the snout of Chorabari Glacier at an altitude of 3820 m a.s.l. (Fig. 5.1). Detailed description of the analysis of meteorological parameters is given below -

5.2.1 Air Temperature

Air temperature (T_a) tends to play a dominant role in glacier retreat. There is a strong statistical correlation between air temperature and glacier fluctuations over the large distances (Greene 2005). During the ablation season, generally, the trend of change in T_a followed an increasing trend till July and then started continuously decreasing. Likewise, at the time of precipitation, sudden drop in T_a was also marked (Fig. 5.2). The maximum T_a was observed 16.0, 16.7, 16.3, 15.2, 15.1 and 13.4°C for the months of May, June, July, August, September and October respectively, whereas minimum for these months

were -5.5 , 0.8 , 5.6 , 5.2 , 1.4 and -5.4°C . Similarly, the subsequent daily mean T_a for these months were 4.7 , 8.4 , 9.8 , 9.4 , 8.1 and 1.9°C respectively indicating that July was the warmest month while October being the coldest. The frequency distribution analysis of diurnal T_a specifies that maximum T_a occurred between 1200 and 1400 h (Local Time - LT). For the study period, the average daily T_a was calculated 7.0°C (Table 2).

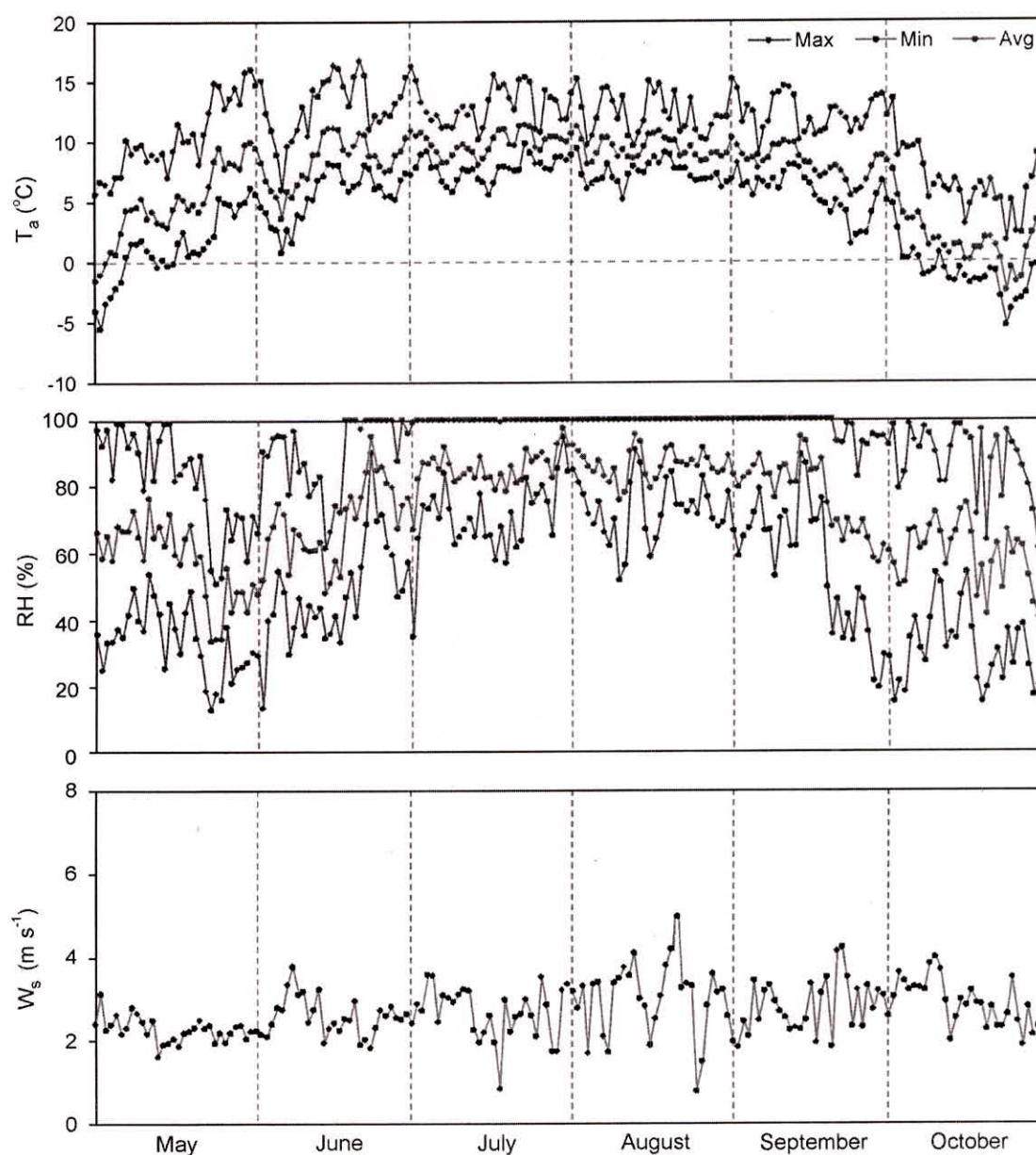


Fig. 5.2: Time series of daily mean meteorological variables recorded from 01 May 2012 to 31 October 2012. T_a , RH, and W_s are representing the air temperature, relative humidity and wind speed respectively.

5.2.2 Relative Humidity

The relative humidity (RH) has an inverse relationship with temperature. As a consequence of variations in air temperature, RH tends to be lower during the day and higher at the night. Daily values of RH ranged between 12.5-100% and average RH was computed to be 72.3% during the summer season (Fig. 5.2). The average daily maximum RH for the months of May, June, July, August, September and October were 82, 89, 100, 100, 98 and 87% respectively, whereas the average minimum RH for these months were 34, 47, 71, 73, 57 and 31%. Similarly, the average RH for these months was 57.7, 68.3, 85.4, 86.6, 77.2 and 58.7% indicating that August was the most humid month due to the influence of Indian Summer Monsoon. However, RH was observed relatively higher and quite stable during the summer monsoon (Fig 5.2).

5.2.3 Wind Speed

Wind plays a fundamental role in local climatology. In the high-altitude regions, wind is one of the major significant factors affecting the transportation of moisture, formation of clouds, occurrence of precipitation and melting of glaciers (Singh et al., 2007). Diurnal analysis of wind speed (W_s) pattern is observed dissimilar for day (0600-1800 LT) and night (1900-0500 hr LT), and varies with change in season. During the day time the W_s is observed to be higher around 1300 to 1400 h (LT) due to stronger temperature deficit (temperature difference between near-surface and ambient atmosphere) while minimum in the morning around 0600 to 0800 h (LT). Over the observation period, the mean W_s was recorded 2.4 ms^{-1} (Table 2). The frequency distribution analysis of hourly W_s indicates that during the monsoonal months (July, August and September), the area experienced majorly light W_s ($<2.0 \text{ m s}^{-1}$) whereas during the months of May, June and October frequency of high moderate W_s ($4.0\text{-}6.0 \text{ m s}^{-1}$) was higher (Fig. 5.2). The average daily W_s for the months of May, June, July, August, September and October were 3.4, 2.7, 1.7, 1.6, 2.0 and 2.7 indicating that highest average W_s was in the month of May.

5.2.4 Radiative Fluxes

The measured daily mean incoming solar radiation ($S_{\downarrow}$) showed large inter-daily as well as monthly variations caused by cloud conditions. Daily mean $S_{\downarrow}$ varied between 8 and 341 W m^{-2} while outgoing solar radiation ($S_{\uparrow}$) varies considerably between 01 and 182

W m^{-2} (Fig 4). The average $S_{\downarrow}$ for the months of May, June, July, August, September and October were 341, 294, 281, 241, 261 and 217 W m^{-2} respectively, whereas $S_{\uparrow}$ for these months were 253, 204, 140, 128, 172 and 163 W m^{-2} . During the ablation season the mean $S_{\downarrow}$ was 176 W m^{-2} whereas $S_{\uparrow}$ was 34 W m^{-2} . However, the net shortwave radiation ($S_{\text{net}} = S_{\downarrow} - S_{\uparrow}$) calculated for these months were 88, 90, 141, 113, 89 and 54 W m^{-2} which indicates that from May to September, the amount of absorbed solar radiation was quite high, particularly in the month of July resulting in more surface melting in the glacierized area (Fig. 5.3).

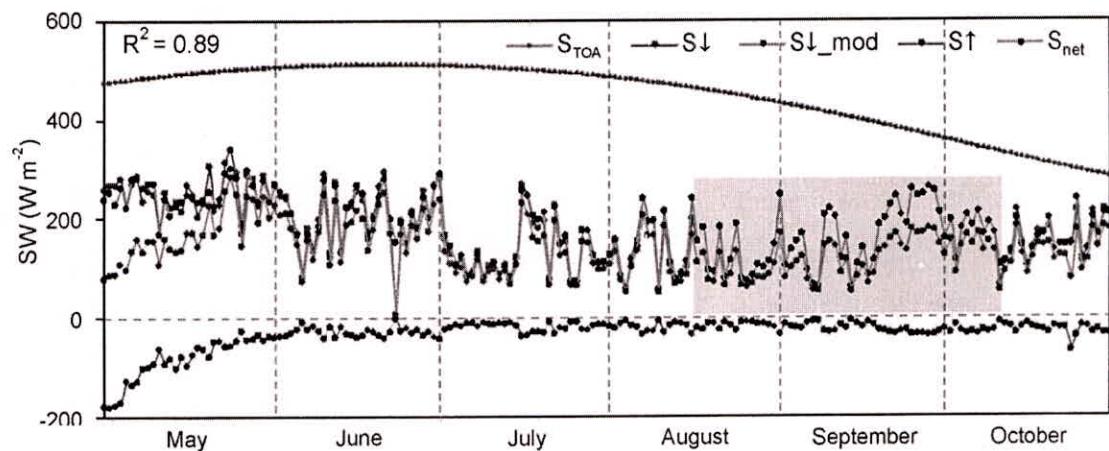


Fig. 5.3: Daily mean recorded ($S_{\downarrow}$ and $S_{\uparrow}$) or computed (S_{TOA} and S_{net}) meteorological variables from 01 May 2012 to 31 October 2012. A data gap in $S_{\downarrow}$ from 17 August 2012 to 10 October 2012 exists. Periods colored with grey are the linearly interpolated modelled data for this data gap. S_{TOA} , $S_{\downarrow}$, and $S_{\uparrow}$ are representing the extraterrestrial solar irradiance at top of atmosphere, incoming shortwave radiation, and reflected shortwave radiation, respectively.

5.2.5 Atmospheric Pressure

Atmospheric pressure (P_a), sometimes also called barometric pressure, is the pressure within the atmosphere of Earth (or that of another planet). In most circumstances, atmospheric pressure is closely approximated by the hydrostatic pressure caused by the weight of air above the measurement point. As elevation increases, there is less overlying atmospheric mass, so that atmospheric pressure decreases with increasing elevation. The maximum pressure was recorded during the study period was 646 hPa and minimum was observed 638 hPa. Average pressure over the study area was observed 641 hPa (Fig. 5.4).

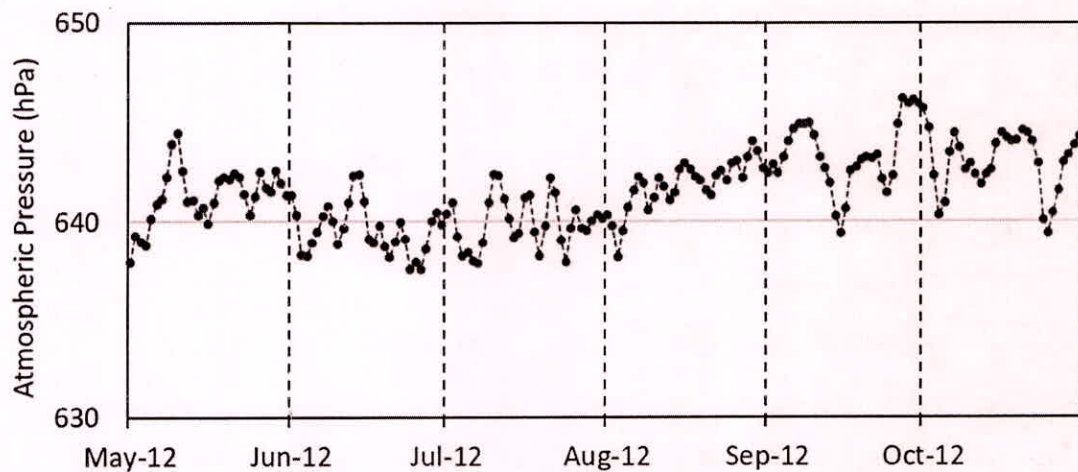


Fig. 5.4: Daily observed atmospheric pressure (P_a) in the Chorabari Glacier catchment (Source: Kesarwani, 2015).

5.2.6 Net Radiation

The net radiation (R_n) is the balance of the incident and reflected short-wave radiations and the incoming and outgoing long-wave radiations. NR-LITE, a rugged net radiometer manufactured by Kipp & Zonen is used to measure the net radiation. Net radiometers measure the energy balance between incoming short-wave and longwave radiation relative to surface reflected short-wave and outgoing longwave IR radiation. The maximum R_n was recorded during the study period was 158 Wm^{-2} and minimum was observed -8 Wm^{-2} . Average net radiation over the study area was observed 70 Wm^{-2} (Fig. 5.4).

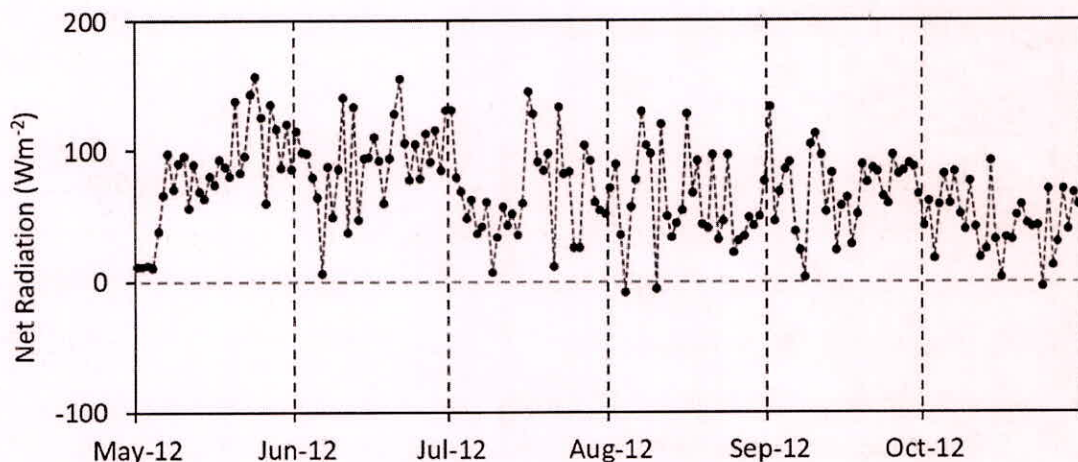


Fig. 5.5: Daily observed net radiation (R_n) in the Chorabari Glacier catchment. (Source: Kesarwani, 2015).

5.2.7 Precipitation

Rainfall represents the most important parameters of meteorology for any particular region and can be used as a climatic indicator. The maximum rainfall in the Chorabari Glacier catchment was 42 mm (Fig. 5.6). Based on the available *in-situ* measurements, results showed that August 2012 experienced highest rainfall followed by July 2012.

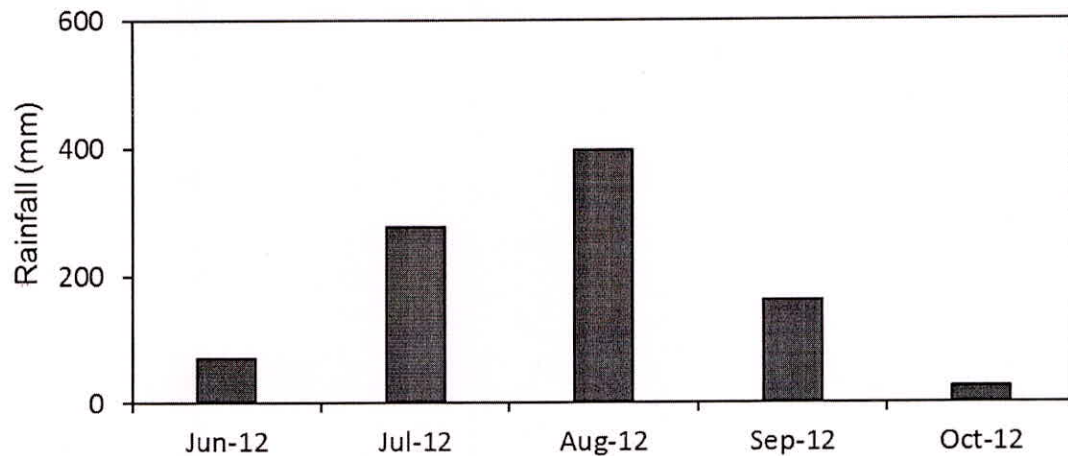


Fig. 5.6: Monthly observed rainfall records in the Chorabari Glacier catchment. (Source: Dobhal et al., 2013; Kesarwani, 2015).

5.3 Results and Discussions

5.3.1 Estimation of Cloud Cover

Cloud plays multiple critical roles in the climate system. In the study area, the anabatic wind mostly accelerates the formation of cumulus or stratocumulus clouds, which results in the afternoon showers on warm and humid days.

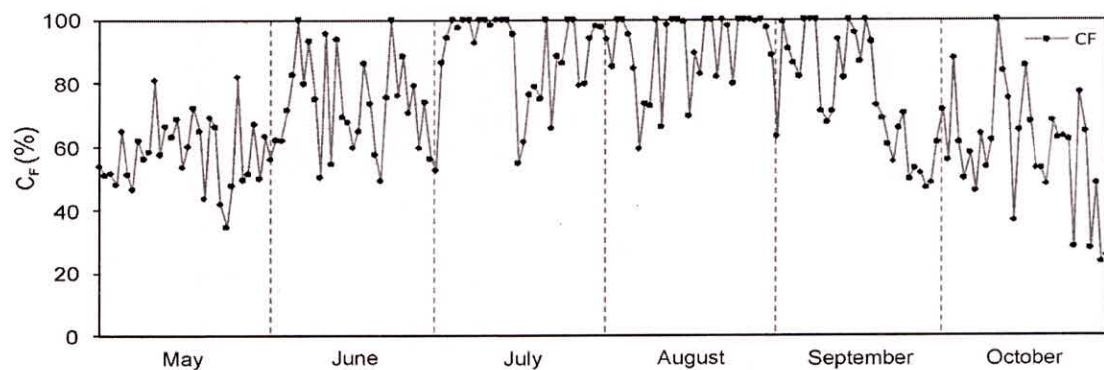


Fig. 5.7: Daily estimated cloud cover during the ablation season of the Chorabari Glacier catchment (Source: Kesarwani et al., 2015).

For the study of cloud cover (C_F), the cloud conditions of the area were divided into four sectors i.e. clear (0-5%), partly cloudy (5-50%), mostly cloudy (50-95%) and overcast (95-100%). The average daily C_F for the months of May, June, July, August, September and October were 58%, 73%, 89%, 91%, 76% and 59% indicating that August (the highest cloudy months) was under the range of overcast and rest of the months are in the mostly cloudy sector. However, computed average C_F was 74% representing that the area experienced mostly cloudy days during the ablation season (Fig. 5.7). C_F for the monsoonal months (July, August and September) was high that resulted in severe drop off in $S_{\downarrow}$ contribution and was partly compensated by the increased incoming long-wave radiation.

5.3.2 Estimation of Longwave Radiation

In the Earth's climate system, longwave radiation involves processes of absorption, scattering, and emissions from atmospheric gases, aerosols, clouds and the surface. Longwave radiation typically refers to radiation in the spectral region from 3 to 100 microns. Radiative cooling by outgoing longwave radiation is the primary way the Earth System loses energy. The balance between this loss and the energy gained by radiative heating from incoming solar shortwave radiation determines global heating or cooling of the Earth system (Energy budget of Earth's climate). The flux of energy transported by incoming and outgoing longwave radiation is measured in W/m^2 . Over 99% of outgoing longwave radiation has wavelengths between 4 μm and 100 μm , in the thermal infrared part of the electromagnetic spectrum. The incoming and outgoing longwave radiations are majorly affected by clouds and dust in the atmosphere, which tend to reduce it to below clear sky values.

The daily values of incoming and outgoing longwave radiations were calculated. On average, the average incoming longwave radiation was calculated $268.4 W m^{-2}$ during the ablation season (01 May 2012 to 31 October 2012). Similarly, the average outgoing longwave radiation was calculated $355.3 W m^{-2}$ (Fig.5.8).

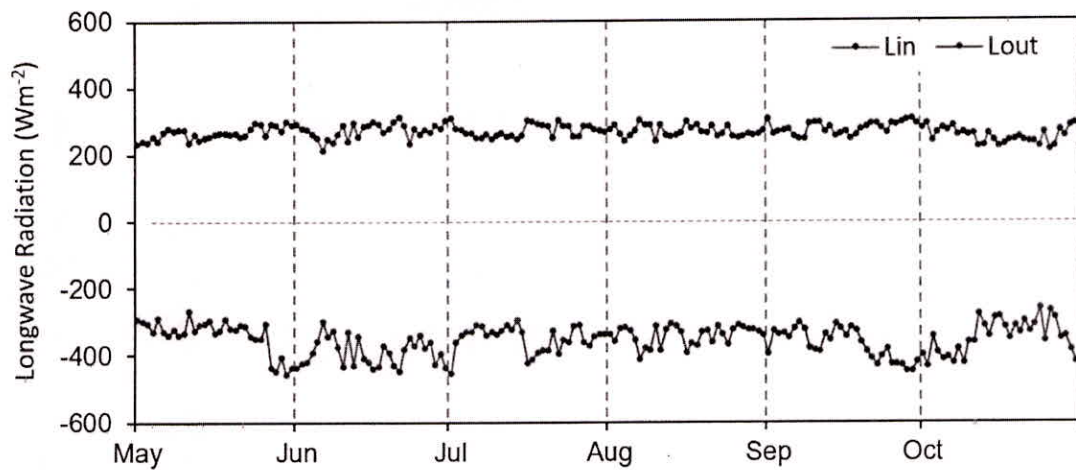


Fig. 5.8: Daily estimated recorded ($L_{\downarrow}$ and $L_{\uparrow}$) meteorological variables from 01 May 2012 to 31 October 2012. $L_{\downarrow}$ and $L_{\uparrow}$ represent the incoming longwave radiation and reflected longwave radiation (Source: Kesarwani, 2015).

5.3.3 Estimation of Energy Fluxes at the Chorabari Glacier

Based on the Eqn (3.1), each energy fluxes were calculated for the Chorabari Glacier catchment at the AWS site (Fig. 5.9).

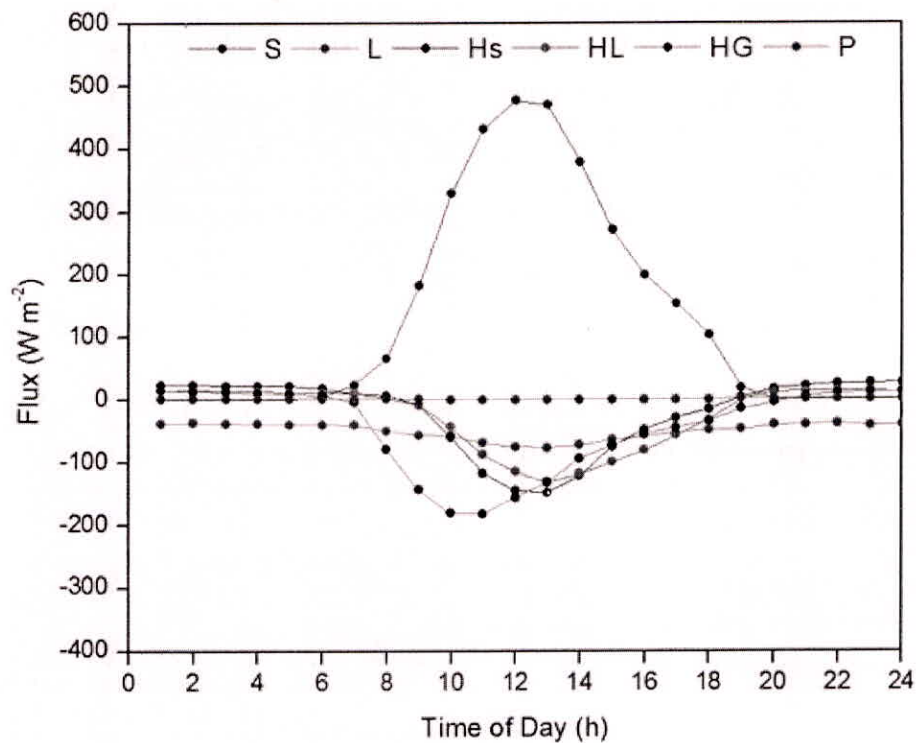


Fig. 5.9: The average hourly energy fluxes calculated for the Chorabari Glacier catchment at the AWS site. S represents the net shortwave radiation, L represents the net longwave radiation, H_s represents the sensible heat flux, H_L represents the latent heat flux, H_G represents the conductive heat flux and P represents the Precipitation heat flux (Source: Kesarwani, 2015).

S represents the net shortwave radiation, L represents the net longwave radiation, H_s represents the sensible heat flux, H_L represents the latent heat flux, H_G represents the conductive heat flux and P represents the Precipitation heat flux (Source: Kesarwani, 2015). The results suggest that fluctuations in mean seasonal net shortwave radiation is high and vary significantly. Besides the seasonal changes in Sun inclination, the main reason for seasonal fluctuation in net shortwave radiation was due to highly variable albedo. During this period, lower mean value of albedo (26%) indicates low outgoing shortwave radiation (50 W m^{-2}) and consequently high net shortwave radiation (127 W m^{-2}) that resulted from surface melting of the glacier. Average seasonal variations in L_n were high in the summer. The mean value of net longwave radiation was -87 W m^{-2} . The obtained variations are directly related to the seasonal fluctuations in incoming and outgoing longwave radiations. Values of incoming longwave radiations depend on the atmospheric air temperature, cloud conditions and humidity that were relatively high (260 W m^{-2}). However, outgoing longwave radiation is governed by the surface temperature and conditions, which have high value (347 W m^{-2}). Based on the recorded meteorological parameters given in, it was observed that Net radiation was very high with a mean value of 48 W m^{-2} . Significant seasonal variations in Sensible heat flux and Latent heat flux were marked. During this period, the temperature gradient between surface and air together with low wind speed was little, resulted in negative values of Sensible heat flux (average = -6 W m^{-2}). As a result, because of high air temperature and relative humidity (leads to a reversal of the humidity gradient) a significant shift from negative to positive in Latent heat flux (average = 2 W m^{-2}) was observed which results an enhancement in the surface melting (Kesarwani, 2015). During the summer monsoon, the heat flux used for melting the snow accumulation (H_{snow}) was 13 W m^{-2} that resulted inadequacies in the surface melting on the upper ablation and accumulation area of the Chorabari Glacier. Correspondingly, heat flux supplied by the monsoonal rainfall ($H_R = 1 \text{ W m}^{-2}$) was very small (Fig. 5.9). Conductive heat flux was negative ($H_G = -32 \text{ W m}^{-2}$) which indicates that during summer the temperature gradient was negative and consequently the subsurface was gaining the energy from atmosphere and transferring this towards the upper layers of surface which enhanced the surface melting.

An interpretation of the facts indicates that Net radiation (150 %; with dominated net shortwave radiation) was the largest contributor to the total surface energy flux followed by the turbulent Sensible heat flux (19 %; negative), Latent heat flux (6 %; positive) and Heat flux due to precipitation (37 %; negative) [Heat flux due to rain = 3 % (positive) and Heat flux due to snow = 40 % (negative)].

Chapter – 6

Synthesis

6.1 Summary and Conclusions

The biggest problem that we are facing today is the water scarcity and climate change and we have the purest source of fresh water called Glaciers that's why the study of glaciers is really important as they are the most important source of water and in addition they are the best system to detect any climate change in our environment. In this Dissertation work, an attempt has been made to study the basics of glaciology, how Physics works in natural system and with the use Debris Energy balance model we learn to implement the science in the field of glaciology to observe the thermodynamical changes of debris-covered Chorabari Glacier. The Chorabari is a south facing glacier (7.5 km) having well defined accumulation and ablation area and the Mandakini River, a major tributary of the Alaknanda River originates from this glacier. The ablation area of the glacier is characterized by thick debris-covered (80 % of total ablation area). In the literature review we studied the global distribution of glaciers and Himalayan distribution of glaciers because without the accurate knowledge about the glacier distribution it would be difficult to study their impact on our climate and environment. After studying about the distribution of glaciers an approach has been made to study the Debris effect on glaciers for that Debris energy balance is studied to know more

precisely about how debris-covered glaciers work in our nature system and how they react towards the sun exposure. The physically based DEB model (Kesarwani et al., 2015) which have been studied with the help of each of the relevant energy fluxes. The previously developed DEB model studied in this report was developed using the near-surface meteorological parameters (e.g. air & surface temperature, relative humidity, wind speed & direction, incoming & outgoing shortwave radiation, net all wavelength radiation, rainfall and snowfall, and atmospheric pressure). The previously published data used for these meteorological parameters were collected through a network of three Automatic Weather Stations (AWS) installed at different elevation. In DEB model, we have studied specific subsurface heat conduction more precisely, over debris-covered glaciers. Therefore, in the present study we have concluded that the minimum thickness of debris over a glacier should greater than or equal to 5cm, only then the debris over the glacier will act as blanket to protect it from melting otherwise it will increase the melting. Moreover, we have studied and learned about all the instruments used to collect data in the nature system and the basic principal of all the instruments.

Finally, the energy balance study gives us results which were based on previously published data (conducted at middle of the glacier ablation zone for both the glaciological budget years 2011-2012 and 2012-2013) and they indicate that during summer the net shortwave radiation was the major source of energy responsible for melting of snow/ice followed by the turbulent latent and rain heat fluxes. These findings are in well agreement with the results of many energy balance studies conducted in the Himalaya as well as in the other mountain regions of the world (e.g. Bhutiyani *et al.*, 1989; Hong *et al.*, 1992; Van de Wal and Oerlemans; 1994; Greuell and Smeets, 2001; Favier *et al.*, 2004; Haidong *et al.*, 2006; Nicholson and Benn; 2006; Giesen *et al.*, 2009; Brock *et al.*, 2010, Reid and Brock, 2010; Li *et al.*, 2011; Sicart *et al.*, 2011; Wheler and Flowers, 2011; Zhang *et al.*, 2013; Azam *et al.*, 2014; Sun *et al.*, 2014).

During the day time the incoming shortwave radiation is a good predictor of cloud cover while during the night time that depends on the temperature and humidity range (). Indeed, a key result of this work was to highlight the Importance of glacier study as they are closely related to climate and small climate change will affect them greatly. It should be stressed that this issue is equally important for both the energy balance and temperature index models. From all the study which is conducted during this dissertation report we have concluded that the DEB model can be useful at any location

on the glacier where input data (i.e. meteorological measurements) and surface characteristics are available. Since, the DEB model includes heat conduction through supraglacial debris layers; therefore, this model is especially beneficial for the melt-modelling study of debris-covered mountain glaciers of the world. Suggested that all important processes were moderately included in the DEB.

In summary, this dissertation report has been based on the study of glacier, their meteorological parameters and by using the previously developed DEB model we have studied the way in which glaciers behave, whether they advance or retreats and how they will affect the environment directly or indirectly (Kesarwani et al., 2015). The behavior of glaciers can be easily observed by defining the meteorological parameters in the energy balance of the glaciers. The model could be useful tool for the assessment of climatic forcing factors on glacierized basins for future melting towards ongoing climatic scenario but need long series of data set. The study is also useful for the Himalayan glaciers as many people are dependent on the Himalayan glacier and very less has been known about how they behave. To study the effect of climate variations on the water resources originating from the glacierized basins is an increasing demand in the current climate change scenario that has become an increasingly important. So, this dissertation work was carried out to study the glaciers as water resource and how they cause climate change. The results give us an insight to both the important factors of Research motivation.

6.2 Future Scope

The energy balance of the glaciers which we have studied in this work is the key control of the interaction between glaciers and climate and a key step in the study of the mass balance of glaciers (Kesarwani, 2015). In future studies, it can help us to provide the accurate information about the approximate melt runoff from the glacier surface which varies according to the conditions of climate. In addition, it can also help to study the possible mass changes over the glacier surface. 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 energy balance or thermodynamical changes of the 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. It is well established fact that the debris-covered over the glacier surface plays a vital role in controlling the surface melt rate that depends on its concentration (thickness) and wide spread distribution (coverage). Therefore, to adopt the 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 effect the environment. Further research can be carried out to study the melt water generated by glaciers so we will know how to use this water resource efficiently.

References

- Azam, M.F., Wagnon P., Vincent, C., Ramanathan, A.L., Favier, V., Mandal, A. and Pottakkal, J.G. (2014). Processes governing the mass balance of Chhota Shigri Glacier (western Himalaya, India) assessed by point-scale surface energy balance measurements. *The Cryosphere*, 8, 2195-2217.
- Bajracharya, S.R. and Shrestha, B. (2011). The status of glaciers in the Hindu Kush-Himalayan region. International Centre for Integrated Mountain Development, Kathmandu, Nepal, ISBN: 9789291152179, 140 pp.
- Benn, D.I., Bolch, T., Dennis, K., Gulley, J., Luckman, A., Nicholson, L.I., Quincey, D., Thompson, S., Toumi, R. and Wiseman, S. (2012). Response of debris-covered glaciers in the Mount Everest region to recent warming, and implications for outburst flood hazards. *Earth-Sci. Rev.*, 114 (1), 156-174.
- Bhambri, R., Bolch, T. and Chaujar, R.K. (2012). Frontal recession of Gangotri Glacier, Garhwal Himalayas from 1965 to 2006, measured through high-resolution remote sensing data. *Curr. Sci.*, 102 (3), 489-494.
- Bhutiyan, M.R. (1989). Modelling of glacier melt during ablation period from energy balance studies on Chhota Shigri Glacier, (HP), India. In: Second National Symposium on Hydrology, Jammu (J&K), India, A116-A126.
- Bolch, T., Kulkarni, A., Kääb, A., Huggel, C., Paul, F., Cogley, J.G., Frey, H., Kargel, J.S., Fujita, K., Scheel, M., Bajracharya, S. and Stoffel, M. (2012). The state and fate of Himalayan glaciers. *Science*, 336 (6079), 310-314
- Brutsaert, W. (1975). On a derivable formula for long-wave radiation from clear skies. *Water Resour. Res.*, 11 (5), 742-744.
- Chen, J.L., Wilson, C.R. and Tapley, B.D. (2013). Contribution of ice sheet and mountain glacier melt to recent sea level rise. *Nat. Geosci.*, 6, 549-552.
- Crawford, T.M. and Duchon C.E. (1999). An improved parameterization for estimating effective atmospheric emissivity for use in calculating daytime downwelling longwave radiation. *J. Appl. Meteorol.*, 38 (4), 474-480.
- Cuffey, K.M. and Paterson, W.S.B. (2010). The physics of glaciers - 4th Edition, Butterworth-Heinemann, USA. 704 pp.
- Cutler, P. and Munro, D.S. (1996). Visible and near-infrared reflectivity during the ablation period on Peyto Glacier, Alberta, Canada. *J. Glaciol.*, 42 (141), 333-340
- Dobhal, D.P., Mehta M. and Srivastava, D. (2013a). Influence of debris cover on

- terminus retreat and mass changes of Chorabari Glacier, Garhwal region, Central Himalaya, India. *J. Glaciol.*, 59 (217), 961–971.
- Dobhal, D.P., Gupta, A.K., Mehta, M. and Khandelwal, D.D. (2013b). Kedarnath disaster: facts and plausible causes. *Curr. Sci.*, 105 (2), 171–174.
- Dyhrenfurth, G.O. (1955). To the Third Pole: The history of the High Himalaya - 1st Edition, London: Werner Laurie, 233 pp.
- Eriksson, M., Xu, J., Shrestha, A.B., Vaidya, R.A., Nepal, S. and Sandstrom, K. (2009). The changing Himalayas: Impact of climate change on water resources and livelihoods in the greater Himalayas, ICIMOD, Kathmandu, Nepal, 22 pp.
- Fujita, K. and Ageta, Y. (2000). Effect of summer accumulation on glacier mass balance on the Tibetan Plateau revealed by mass-balance model. *J. Glaciol.*, 46, 244–252.
- Gadgil, S. (2003). The Indian monsoon and variability. *Annu. Rev. Earth Planet. Sci.*, 31, 429–467.
- Giesen, R.H., Andreassen, L.M., Van Den Broeke, M.R. and Oerlemans, J. (2009). The Cryosphere Comparison of the meteorology and surface energy balance at Storbreven and Midtdalsbreen, two glaciers in southern Norway. *The Cryosphere*, 3, 57–74.
- Greuell, W., Knap, W.H. and Smeets, P.C. (1997). Elevational changes in meteorological variables along a midlatitude glacier during summer. *J. Geophys. Res.*, 102 (D22), 25941–25954.
- Greuell, W. and Genthon, C. (2004). Modelling land ice surface mass balance. In Mass balance of the Cryosphere: Observations and modelling of contemporary and future changes (eds. J.L. Bamber and A.J. Payne), Cambridge University Press, 117–168.
- Haidong, H.A.N., Yongjing D. and Shiyin, L.I.U. (2006). A simple model to estimate ice ablation under a thick debris layer. *J. Glaciol.*, 52 (179), 528–536.
- Immerzeel, W.W., Van Beek, L.P.H. and Bierkens M.F.P. (2010). Climate change will affect the Asian water towers. *Science*, 328, 1382–1385.
- IPCC (2007). Climate Change 2007: Impacts, Adaptation and Vulnerability. Contribution of Working Group II to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (eds. M.L. Parry, O.F. Canziani, J.P. Palutikof, P.J. van der Linden, and C.E. Hanson), Cambridge University Press, Cambridge, UK, 976 pp.
- IPCC (2013). Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change. (eds. T.F. Stocker, D. Qin, G.K. Lattner, M. Tignor, S.K. Allen,

J.Boschung, A.Nauels, Y.Xia, V.Bex, and P.M. Midgley), Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, 1535 pp

- Kayastha, R.B., Takeuchi, Y., Nakawo M. and Ageta. Y. (2000). Practical prediction of ice melting beneath various thickness of debris cover on Khumbu Glacier, Nepal using a positive degree day factor. *IAHS Publ.*, 264 (Symposium at Seattle 2000 – Debris-Covered Glaciers), 71–81.
- Kesarwani, K. (2015): Energy Balance and Mass Fluctuation of the Himalayan Glaciers: A case study of Chorabari Glacier, Garhwal Himalaya. Ph.D. thesis submitted to Kumaun University, Nainital, Uttarakhand, pp. 236.
- Kesarwani, K., Dobhal, D.P., Durgapal, A. and Mehta, M. (2015). High Altitude Meteorology and Cloud Cover Conditions: A study from Chorabari Glacier, Central Himalaya, India. *Himal. Geol.*, 36 (2), 134-142.
- Klok, E.J. and Oerlemans, J. (2002). Model study of the spatial distribution of the energy and mass balance of Morteratschgletscher, Switzerland. *J. Glaciol.*, 48, 505–518
- Kondo, J. and Yamazawa, H. (1986). Bulk transfer coefficient over a snow surface. *Bound.-Layer Meteorol.*, 34 (1–2), 123–135.
- Konzelmann, T., van de Wal, R.S.W., Greuell W., Bintanja, R., Henneken, E.A.C. and Abe-Ouchi, A. (1994). Parameterization of global and longwave incoming radiations for the Greenland ice sheets. *Glob. Planet. Chang.*, 9, 143-164
- Long, C.N. and Ackerman, T.P. (2000). Identification of clear skies from broadband pyranometer measurements and calculation of downwelling shortwave cloud effects. *J. Geophys. Res.*, 105 (D12), 15,609-15,626
- Loomis, S.R. (1970). Morphology and ablation processes on glacier ice. *Assoc. Am. Geogr. Proc.*, 2, 88–92.
- Mattson, L.E., Gardner J.S. and Young. G.J. (1993). Ablation on debris covered glaciers: an example from the Rakhiot Glacier, Punjab, Himalaya. *IAHS Publ.*, 218 (Kathmandu Symposium 1992 - Snow and Glacier Hydrology), 218, 289–296.
- Mehta, M., Majeed, Z., Dobhal, D.P. and Srivastava, P. (2012). Geomorphological evidences of post-LGM glacial advancements in the Himalaya: A study from Chorabari Glacier, Garhwal Himalaya, India. *J. Earth Sys. Sci.*, 121 (1), 149-163.
- Mehta, M., Dobhal, D.P., Kesarwani, K., Pratap, B., Kumar A. and Verma, A. (2014). Monitoring of glacier changes and response time in Chorabari Glacier, Central Himalaya, Garhwal, India. *Curr. Sci.*, 107 (2), 281-289.
- Mölg, T. and Hardy, D.R. (2004). Ablation and associated energy balance of a horizontal

- glacier surface on Kilimanjaro. *J. Geophys. Res.*, 109 (D16), doi:10.1029/2003JD004338.
- Nakawo, M. and Young, G.J. (1982). Estimate of glacier ablation under a debris layer from surface temperature and meteorological variables. *J. Glaciol.*, 28 (98), 29-34.
- Nicholson, L. and Benn, D.I. (2006). Calculating ice melt beneath a debris layer using meteorological data. *J. Glaciol.*, 52 (178), 463-470.
- Niemelä, S., Räisänen, P. and Savijärvi, H. (2001). Comparison of surface radiative flux parameterizations. Part I: Longwave radiations. *Atmos. Res.*, 58, 1-18.
- Østrem, G. (1959). Ice melting under a thin layer of moraine, and the existence of ice cores in moraine ridges. *Geogr. Ann.*, 41(4), 228-230.
- Paterson, W.S.B. (1994). *The Physics of Glaciers* -3rd Edition, Pergamon Press, New York, 496 pp.
- Peixoto, J. and Oort, A. (1996). The climatology of the relative humidity in the atmosphere. *J. Climate*, 9, 3443-3463.
- Pellicciotti F., Helbing, J., Favier, V. and Corripio, J. (2008). A study of the energy balance and melt regime on JuncalNorte Glacier, semi-arid Andes of central Chile, using melt models of different complexity. *Hydrol. Process.*, 22, 3980-3997
- Pratap, B., Dobhal, D.P., Bhambri, R., Mehta, M. and Tewari, V.C. (2015a). Four decades of glacier mass balance observations in the Indian Himalaya. *Reg. Environ. Change*, doi:10.1007/s10113-015-0791-4.
- Raina, V.K. and Srivastava, D. (2008). *Glacier atlas of India*. Geological Society of India, Bangalore, 315 pp.
- Raina, V.K. (2009). *Himalayan Glaciers-A State-of-Art Review of Glacial Studies, Glacial Retreat and Climate Change*. Discussion Paper, MoEF, Government of India, 60 pp.
- Reid, T.D. and Brock, B.W. (2010). An energy-balance model for debris-covered glaciers including heat conduction through the debris layer. *J. Glaciol.*, 56 (199), 903-916.
- Ross, R.J. and Elliott, W.P. (1996). Tropospheric Water Vapor Climatology and Trends over North America: 1973-93. *J. Climate*, 9, 3561-3574
- Sakai, A., Fujita, K., Nakawo, M. and Yao, T. (2009). Simplification of heat balance calculation and its application to the glacier runoff from the July 1st Glacier in northwest China since the 1930s. *Hydrol. Process.*, 23 (4), 585-596.

- Sangewar, C.V. and Shukla, S.P. (2009). Inventory of the Himalayan Glaciers: A Contribution to the International Hydrological Programme Spl. Publ., 34, Geological Survey of India: An Updated Edition, 594 pp.
- Sicart, J.E., Hock, R., Ribstein, P. and Chazarin, J.P. (2010). Sky longwave radiation on tropical Andean glaciers: parameterization and sensitivity to atmospheric variables. *J. Glaciol.*, 56 (199), 854- 860.
- Singh, P. (2001). Hydrological modelling of snow and glacier covered basins. *Proc. Nat. Acad. Sci. India*, 71 (A), 195-222.
- Singh, P., Haritashya, U.K. and Kumar, N. (2007). Meteorological study for Gangotri Glacier and its comparison with other high altitude meteorological stations in central Himalayan region. *Nord. Hydrol.*, 38 (1), 59-77
- Valdiya, K.S., Paul, S.K. and Chandra, T. (1999). Tectonic and lithological characterization of Himadri (Great Himalaya) between Kali and Yamuna rivers, Central Himalaya. *Himal. Geol.*, 20, 1-17
- WGMS (1989). World glacier inventory - Status 1988. (eds. W. Haeberli, H. Bösch, K. Scherler, G. Østrem, and C.C. Wallén.), IAHS (ICSII)-UNEP-UNESCO, World Glacier Monitoring Service, Zurich, Switzerland, 458 pp.
- Yancheva, G., Nowaczyk, N.R. and Mingram, J. (2007). Influence of the intertropical convergence zone on the East Asian monsoon. *Nature*, 445 (7123), 74-77.
- Zhang, G., Kang, S., Fujita, K., Huintjes, E., Xu, J., Yamazaki, T., Haginoya, S., Wei, Y., Scherer, D., Schneider, C. and Yao, T. (2013). Energy and mass balance of the Zhadang Glacier surface, central Tibetan Plateau. *J. Glaciol.*, 213, 137-148