SOME HYDROLOGIC ASPECTS OF SNOWMELT RUNOFF UNDER SUMMER CONDITIONS, IN THE BARPU GLACIER BASIN, CENTRAL KARAKORAM, HIMALAYA, NORTHERN PAKISTAN

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Some Hydrologic Aspects of Snowmelt Runoff Under Summer Conditions, in the Barpu Glacier Basin, Central Karakoram, Himalaya, Northern Pakistan

By

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THESIS

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Abstract

Snow and ice in high mountains represent an important water resource in many parts of the world, especially the dry continental interior of Central Asia. In the Northern Areas of Pakistan, mountain ranges are the primary sources of annually renewed water supplies They give rise to rivers which are the only significant, sustainable source of fresh water The Indus basin is drained by the river Indus and its major tributaries, the Kabul, Jhelum, Chanab, Ravi and Sutlej Snowmelt contributes about 70 percent to the annual flow of these rivers, but is not timed to meet the requirements for crop production, hydroelectric power generation, and other multi-purpose objectives This situation has led to the development of an irrigation economy that requires effective management of the water resources in these drainage basins A basic understanding of snow distribution and its contribution to streamflow is needed for effective prediction of flow events

Snowfields between elevation of 2,500 m and 5,500 m constitute a small percentage of the area contributing to runoff in the Central Karakoram, Northern Pakistan However, they are considered to have a higher water content, and to produce runoff for longer periods than snowpacks at lower elevations The melt regime of a basin may be better understood by examining the snowpack recession and runoff hydrograph Identifying the time of daily peak flow, snow cover/runoff relation and its variation through the season may prove helpful to flow modelling

This study involved taking hydrological and meteorological observations in two small snow-fed basins having different aspects in the basin of Barpu Glacier in the Central Karakoram range of the Himalaya in Pakistan The observational network was designed to cover a range of elevations within the experimental basins. This type of network is essential to account for the effects of topography and microclimate on snow hydrology

Patterns of snowmelt runoff examined in two contrasting environments within the Barpu Glacier Basin suggest that topography influences the rate of spring snowmelt in several ways Aspect and degree of slope modify the winter and spring snowpack by causing unequal rates of ablation Relief creates an unequal distribution of snow which in turn causes areal variation in the volume of spring melt

Normal linear and curvilinear multiple regression analysis is an appropriate method for studying of hydrologic relationships Snow cover area and subsequent snowmelt runoff can be correlated to estimate streamflow For a particular catchment, the relationship between area of snow-cover and snowmelt runoff appears to depend on morphometrical factors such as elevation, aspect, slope, and drainage density However, for each basin a different empirical relation exists between snow cover and snow-melt runoff The logarithmic relationships between

snow-cover and snowmelt runoff indicates a substantial increase in snowdepth with increasing elevation The results also suggest that mean temperature is the best single indicator of runoff variation Meteorological observations over a range of elevations provide valuable information concerning the altitudinal gradient

The concluding chapter reviews some of the practical and technical implications of the work for hydrological investigations in the Upper Indus Basin It suggests what can (and cannot) be learned from this type of study in relation to macroscale water resource assessment and forecasting

Dedication

To my Mother

who prays every moment without knowing exactly
what I am doing but believes that it is best
for my present and for my future

Acknowledgements

During a project such as this Master's Thesis, where one has to work in remote areas, it can not be possible for one person to accomplish the work without the help of others Throughout the field work and writing many others have devoted their time and given assistance With this in mind, I must acknowledge the contribution of these people to this thesis research.

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I offer my expression of gratitude to those whose names do not come readily to mind as I write this If through carelessness I have forgotten any of those who have helped me, it is my loss, and so I offer thanks to you now Finally, continuing a

tradition essential in writing acknowledgements, I claim full responsibility for any errors of presentation or interpretation which clearer heads may find here, this effort was largely my own, and its faults are exclusively so

Ghazanfar Ali

Waterloo
April 1989

Table of Contents

Acknowledgements i
Abstract ii
List of Tables viii
List of Figures xi

1 INTRODUCTION

1 1 BACKGROUND 1
1 2 RESEARCH OBJECTIVES 8
1 3 LITRERATURE REVIEW 10
1 3 1 Variation in Precipitation 10
1 3 2 Ablation in Karakoram 13
1 3 3 Snow cover Area in Relation to Streamflow 17

2. CLIMATE AND GEOGRAPHY

2 1 OVERVIEW 20
2 2 GEOGRAPHY OF THE KARAKORAM 24
2 2 1 Regional Setting 24
2 2 2 Glacier of the Karakoram 26
2 3 BROAD CLIMATIC CONTROLS 26
2 4 ROLE OF MONSOONS IN THE WESTERN HIMALAYAS 29
2 5 LOCAL CLIMATE The effect of altitude and topography 30
2 5 1 Radiation 34
2 5 2 Temperature 34
2 5 3 Precipitation 37
2 5 4 Wind 38
2 6 HYDROLOGY 39
2 7 CONCLUSION 41

3. SNOW ACCUMULATION AND ABLATION IN MOUNTAINOUS AREAS

3 1 INTRODUCTION 44
3 2 FACTORS AFFECTING SNOW ACCUMULATION AND DISTRIBUTION 46
3 2 1 Wind 46
3 2 1 1 Transport of snow by wind 48
3 2 1 2 Modes of transport 49
3 2 1 3 Effects of wind on snow density 51
3 2 2 Topographic Factors 52

3 2 2 1 Elevation 52
3 2 2 2 Slope 55
3 2 2.3 Aspect 56
3 3 SNOWMELT IN THE MOUNTAIN ENVIRONMENT 57
3 3 1 Sources of Heat Energy 57
3 3 2 Thermal Quality of Snow Pack 58
3 3 3 The Radiation Balance at Snow Surface 59
3 3 3 1 Direct solar radiation 60
3 3 3 2 Direct radiation received on slopes 60
3 3 3 3 Variations of direct radiation with altitude 62
3 3 4 Diffuse Radiation 64
3 3 4 1 Diffuse radiation over snow surface 65
3 3 4 2 Variation of diffuse radiation with altitude 66
3 3 5 Reflected Radiation or Albedo 66
3 3 5 1 Changes in snow surface properties 68
3 3 6 Effective Long-wave Radiation 68
3 3 7 Turbulent Energy Transfer 71
3 3 7 1 Air-temperature 72
3 4 AVALANCHES 74
3 5 CONCLUSION 74

4 FIELD OBSERVATION PROGRAM AND PROCEDURE

4 1 INTRODUCTION 76
4 2 STUDY AREA 77
4 3 SELECTION OF EXPERIMENTAL BASINS 82
4 4 STREAMFLOW MEASUREMENTS 84
4 5 METEOROLOGICAL STATIONS Location and Description 86
4 5 1 Southwest Slopes 89
4 5 2 Northeast Slopes 92
4 5 3 Meteorological Station at Miar Glacier 92
4 6 PRECIPITATION AND CLOUD COVER 93
4 7 SNOWMELT PATTERNS 94
4 8 SNOWPIT STUDY 94
4 9 CONCLUSION 95

5 ANALYSES OF HYDROMETEOROLOGICAL DATA AND RESULTS

5 1 ALTITUDINAL GRADIENT OF TEMPERATURE 96
5 1 1 Introduction 96
5 1 2 Observation Sites and Data 97
5 1 3 Results and Discussion 98
5 1 3 1 Southwest Slopes 98
5 1 3 2 Northeast Slopes 104

5 1 4 Maximum Isotherm 106
5 1 5 Conclusions 106
5 2 VARIATIONS IN SNOWMELT RUNOFF 108
5 2 1 Overview 108
5 2 2 Snow Disappearance and Runoff 110
5 2 2 1 Assessment of the areal extent of snowcover 111
5 2 2 2 Relationship between snow cover depletion and runoff 113
5 2 2 3 Logarithmic relationship between snow cover area and streamflow 116
5 2 3 Description of Snowmelt Runoff Pattern 119
5 2 3 1 Seasonal runoff pattern 119
5 2 3 2 Diurnal runoff pattern 127
5 2 4 Conclusions 133
5 3 RELATIVE IMPORTANCE OF METEROLOGICAL VARIABLES IN SNOWMELT RUNOFF 136
5 3 1 The Statistical Approach 136
5 3 2 Correlation Computations 137
5 3 2 1 Effect of recognizing the recession flow 137
5 3 3 Results and Discussion 143
5 3 3 1 Southwest 143
5 3 3 2 Northeast 152
5 3 4 Comparison of Results in terms of Aspect 161
5 3 5 Comparison of Observed and Predicted Hydrograph 163
5 3 6 Discussion of Analytical Methods 163
5 3 7 Conclusion 165

6 SUMMARY AND CONCLUSIONS

6 1 INTRODUCTION 167
6 2 RESULTS SUMMARY 168
6 2 1 Variation in Runoff 168
6 2 1 1 Seasonal variation 169
6 2 1 2 Diurnal variation 170
6 2 2 Runoff Indexes 171
6 2 2 1 Snow-cover area vs runoff relation 172
6 2 2 2 Correlation 173
6 3 CONCLUSIONS 175
6 4 FURTHER STUDIES 178

REFRENCES 182

APPENDICES		192
APPENDIX A	GLACIER INVENTORY FOR BARPU/BUALTAR BASIN	A1
APPENDIX B	RATING CURVE FOR STUDY STREAMS	B
B1	RATING CURVE FOR PHAHI PHARI STREAM	B1
B2	RATING CURVE FOR MIAR STREAM	B2
APPENDIX C	BASIN MAPS (P PHARI) SHOWING SNOW COVER (%) IN EACH ELEVATION BAND	C
APPENDIX D	BASIN MAPS (MIAR) SHOWING SNOW COVER (%) IN EACH ELEVATION BAND	D
APPENDIX E	COMPUTER PROGRAM USED FOR THE CALCULATION OF INCOMING SHORT WAVE RADIATION ON TWO ASPECTS	E
APPENDIX F	LIST OF HYDROMETEOROLOGICAL DATA	F

List of Tables

1 1	Rough estimate of moisture input in the accumulation zone of Biafo Glacier (1985-1986)	14
1 2	Net annual accumulation of snow in Central Karakoram	14
2 1	Mean monthly temperatures selected Upper Indus Basin stations	35
2 2	Wind at 2m height. Selected stations in Upper Indus Basin	38
2 3	Basin Characteristics of the Upper Indus Basin Rivers	40
3 1	Densities of snow cover	51
3.2	Results of regression of snow properties according to elevation	54
3 3	Mean accumulation of snow (cm) under different vegetation cover conditions in the Rockies as related to aspect	56
4 1	Physical characteristics of the study basins	83
4 2	Summary of streamflow measurements (P.Phari Stream)	87
4 3	Summary of streamflow measurements (Miar Stream)	88
4 4	Summary of Barpu Basin meteorological stations (Period of record and instrumentation)	90
4.5	Stratigraphic characteristics of snowpit	95
5 1 1	Altitudinal gradients and standard deviations for the daily max., min and mean temperatures	101
5 1 2	Comparison of temperature gradients calculated for two different periods for SSW and NNE slopes	103
5 2 1	Snow cover (percent) in each elevation band for the study basins (southwest and northeast)	112
5 2 2a	Relation between change in snow covered area and basin water yield (P.Phari Stream)	117
5 2 2b	Relation between change in snow covered area and basin water yield (Miar Stream)	117

5 2 3	Dates of maximum mean daily figures for the two basins	125
5 2 4	Relative values of incident solar radiation on different aspects	129
5 2.5	Daily time range of maximum discharge, radiation and temperature	131
5 3 1	List of variables used in statistical correlation	138
5 3 2	Comparison of simple correlation results using daily figures of runoff (total, mean) with correlations using one day's snowmelt runoff (total, mean)	139
5 3 3	Comparison of multiple correlation results using daily figures of runoff (total, mean) with correlations using one day's snowmelt runoff (total, mean)	141
5 3 4	Comparison of linear and log linear regression results using P.Phari meteorological data vs runoff	142
5 3.5	Correlation coefficients between independent variables (P.Phari)	144
5 3 6	Multiple regression equations using P.Phari meteorological data	146
5 3 7	Correlation coefficients between independent variables (Rush Lake)	148
5 3 8	Multiple regression equations using Rush Lake data	149
5 3 9	Comparison of correlations using total one day runoff with correlations using mean daily runoff	153
5 3 10	Correlation coefficients between independent variables (Miar)	154
5 3 11	Multiple regression equations using Miar meteorological data	155
5 3 12	Correlation coefficients between independent variables (Miar Hill)	157
5 3 13	Multiple regression equations using Miar Hill data	158
5 3 14	Comparison of simple correlation results interms of aspect	162

List of Figures

1 1	Indus River and its Tributaries	2
1 2	Glaciers and drainage in northern Pakistan	3
1 3	Schematic diagram of Indus Basin System	5
1 4	Relation of mean daily ablation to till cover thickness on the Rakiot Glacier in Nanga Parbat Region 1986	16
2 1	Location map and physiographic and hydrologic units of the Indus River Basin	21
2 2	Upper Indus Basin above Tarbela Reservior and Wapda gauging stations	23
2 3	Physiography of the Northwestern Himalayas	25
2 4	Mean monthly temperature and precipitation for Gilgit and Skardu	33
2 5	Annual discharge variation in Upper Indus Basin Rivers	42
3 1	Approximate variation of horizontal mass flux of blowing snow with height	50
3 2	Sources of heat that generate snowmelt	57
3 3	Direct radiation on slopes	61
3 4	Sources of diffuse radiation	64
3.5	Albedo versus snow surface density	67
3 6	Diurnal variation of net radiation components	69
3 7	Deflection of large air masses over mountain barriers in the northern hemisphere	72
3 8	Isolines of the mean monthly lapse rate of daily temperature in the Zaravshan River valleys	73
4 1a	Phahi Phari - Yengutz Har Peak Ridges facing southwest in Barpu Glacier Basin Central Karakoram Mountains	78

4 1b	Sumaiyar Bar and Miar Ridges facing northeast in Barpu Glacier Basin Central Karakoram Mountains	79
4 2	Study basins and meteorological observation location in Barpu Glacier Basin Central Karakoram	80
4 3	Area/elevation curves for the Phahi Phari and Miar stream basins	84
5 1 1a	Graphic representation of air temperatures recorded at different elevations (Southwest facing slopes) Barpu Glacier Basin, Central Karakoram	99
5 1 1b	Graphic representation of air temperatures recorded at different elevations (Northeast facing slopes) Barpu Glacier, Basin, Central Karakoram	100
5 1 2a	Temperature inversion on slopes mainly due to downslope wind (katabatic)	105
5 1 2b	Development of downslope winds (katabatic) in the presence of snow	105
5 2 1	Some characteristic shapes of areal depletion curve of snow cover	113
5 2 2	Plot of snowfree area (%) vs subsequent runoff for the two study basins (P.Phari and Miar)	115
5 2 3	Relationship between snow cover area vs subsequent runoff for the two study basins	115
5 2 4	Composite charts consisting of hydrographs of P.Phari and Miar streams showing snow coverage in relation to streamflow	118
5 2 5a	Mean daily runoff (P.Phari stream) and related meteorological variables recorded at 3510m a.s.l on southwest facing slope	120
5 2.5b	Mean daily runoff (P.Phari stream) and related meteorological varaibles recorded at 4572m a.s.l on southwest facing slope	121
5 2 6	Mean daily runoff (Miar stream) and related meteorological variables recorded at 3550m and 4200m a.s.l on northeast facing slopes	126
5 2 7	Hourly fluctuations in runoff, radiation and air-temperature for the two study basins	130

5 3 1	Snowmelt hydrograph showing area under one day's snowmelt appearing in the first 24 hours period	140
5 3 2	Relation between minimum and mean air-temperature recorded at 3510 and 4572m a.s.l respectively on SSW facing slopes in Barpu Glacier Basin Central Karakoram	151
5 3 3	Relation between maximum and mean air-temperature recorded at 3550 and 4200 m a.s.l. respectively on NNE facing slopes in Barpu Glacier Basin Central Karakoram	160
5 3 4	Comparison of observed and predicted hydrographs for the two study streams	164

CHAPTER 1

Introduction

1 1 BACKGROUND

The surface water hydrology of Pakistan is dominated by the Indus River and its five major tributaries, Kabul, Jhelum, Chenab, Ravi and Sutlej (Fig 1 1) The Indus basin stretches from the highlands of Tibet (China) to the Arabian Sea. The main stem of the river is 3200 km long With the division of rivers of the Indus basin between India and Pakistan under the Indus Waters Treaty 1960, Pakistan is entitled to receive only the water from the western rivers of the Indus basin (including Kabul, Jhelum and Chenab)

The rivers of the Indus basin rise in mountains with elevations ranging from 4500 to 7500 meters above sea level (a.s.l) These mountains are covered with winter snow from at least January to March and in some regions from October to July The tremendous arc of the Karakoram Mountains, which extends over 350 km, holds the greatest concentration of snow and glacier ice on the Asian mountains (Hewitt 1986 Fig 1 2) Melting of this large resource of snow and ice provides a major portion of runoff during summer The melting starts in early March in some basins and in April in others and continues throughout the summer River flow

FIGURE 1 1

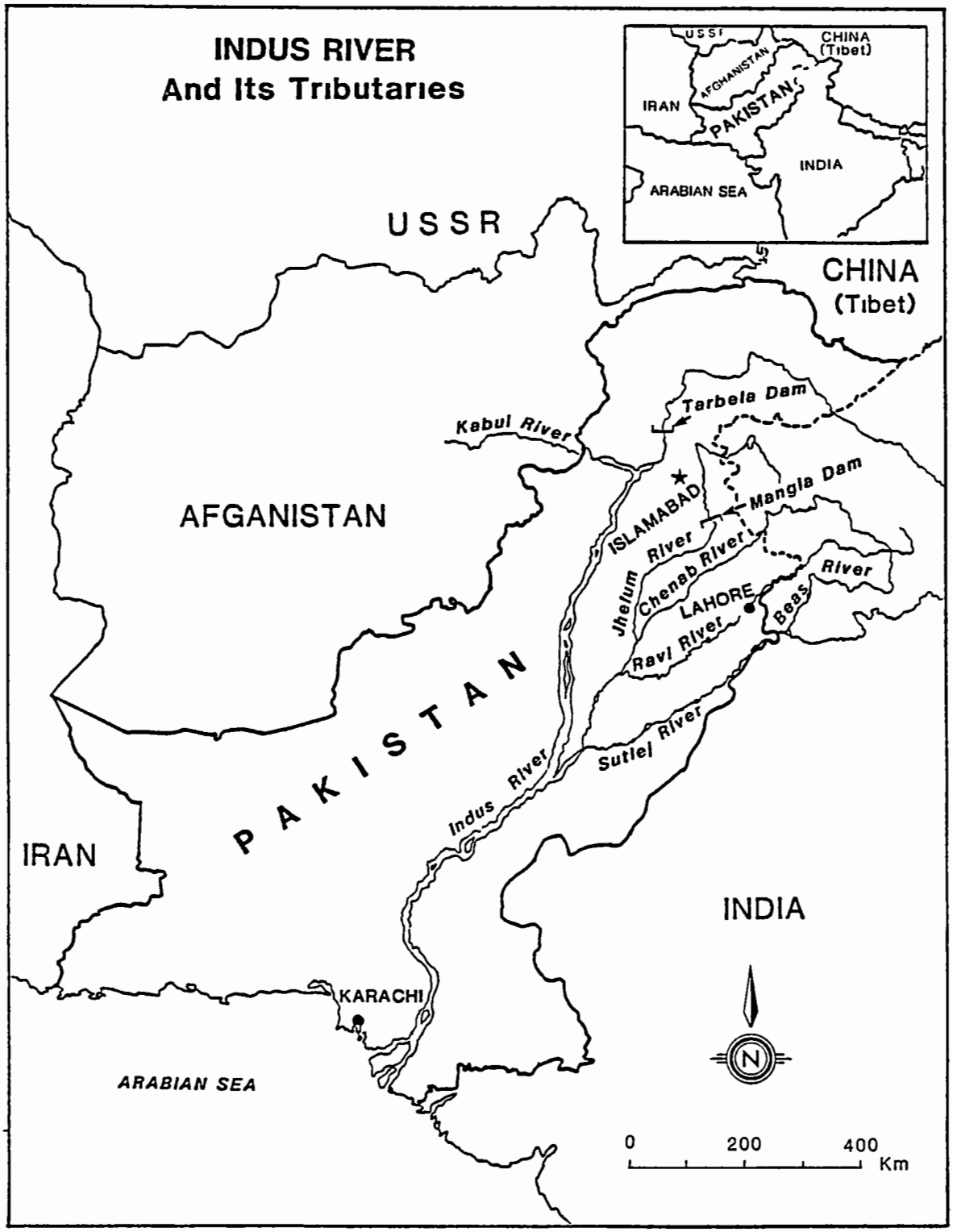
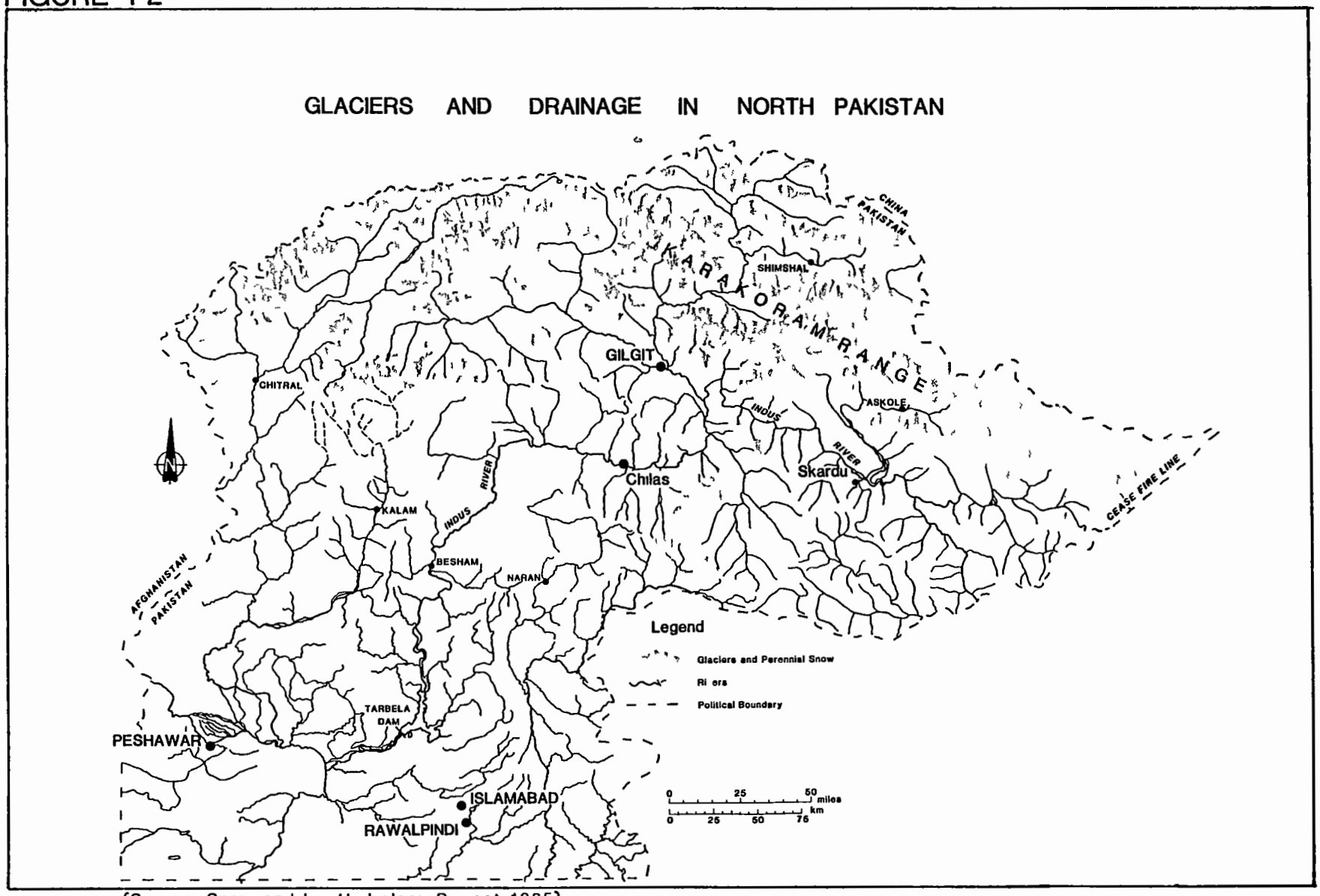

INDUS RIVER
And Its Tributaries
USSR
CHINA
(Tibet)
IRAN
AFGHANISTAN
PAKISTAN
INDIA
ARABIAN SEA
Kabul River
Tarbela Dam
ISLAMABAD
Mangla Dam
Jhelum River
Chenab River
LAHORE
Beas River
Ravi River
Sutlej River
Indus River
KARACHI
N
0
200
400
Km

FIGURE 12



(Source Snow and Ice Hydrology Project 1985)

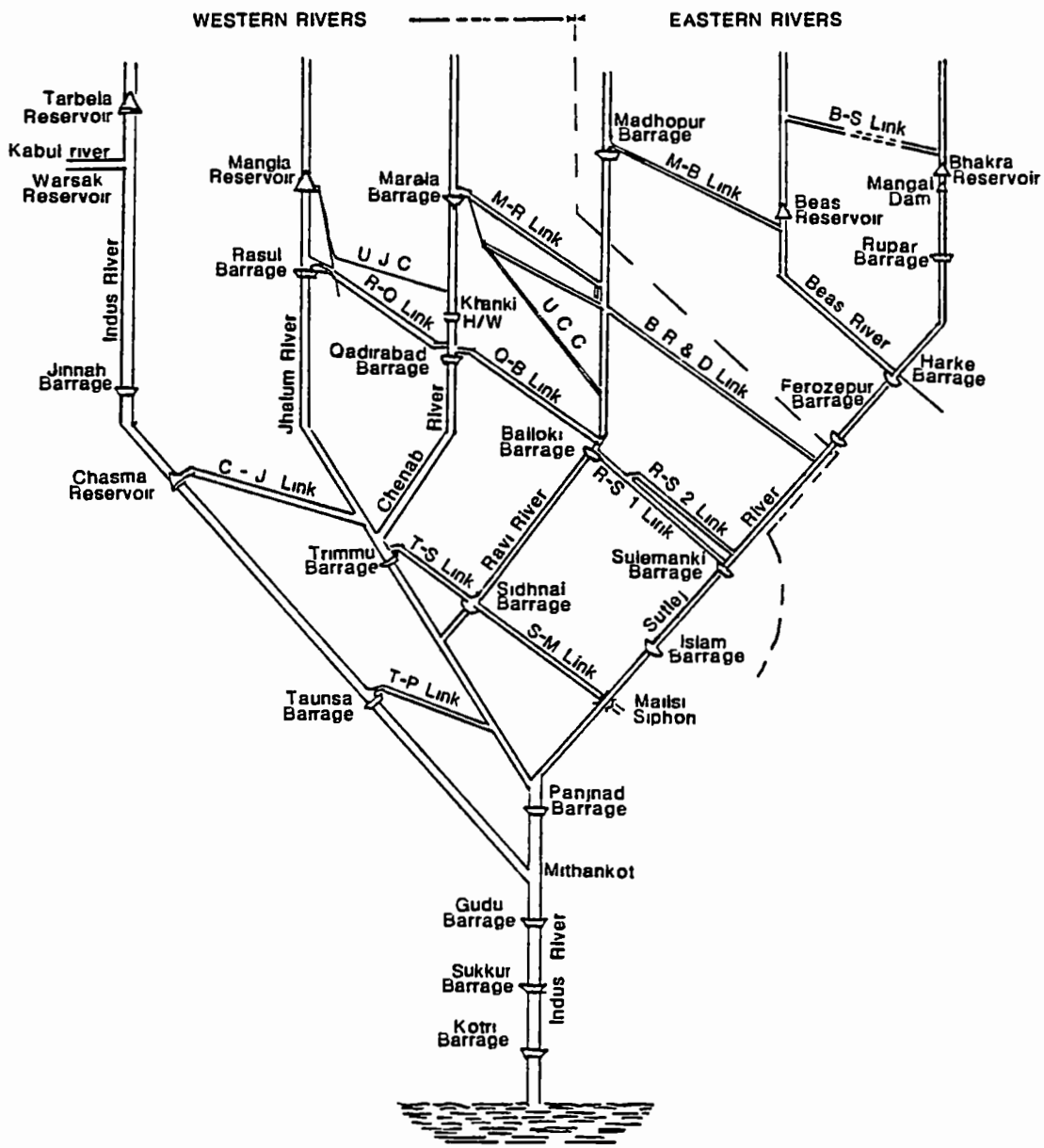
consists mostly of snowmelt until early July After that, glacier melt becomes a major factor under the influence of monsoonal air masses. However, in August heavy rainfall is usually much reduced as the air masses lose most of their moisture or rarely penetrate to the northwestern parts of the Himalayas.

In early spring the Upper Indus Basin rivers derive most of their water supply from snowmelt runoff This is a critical period when the reservoirs are empty and water is very much needed, but is not timed to meet the requirements for crop production, hydroelectric power generation and other multipurpose objectives such as commercial fishing and recreational activities. This situation has led to the development of a system based upon reservoir control and management of the water resources in snow and ice fed basins.

Pakistan is basically an agrarian country with a population of approximately 90 million people mostly dependent upon irrigated agriculture in the Indus plains. This is served through the world's largest contiguous irrigation system developed over the last 100 years in the Indus plains The system is fed through 16 diversion dams (barrages) and 580 km of inter-river link canals which connect the western rivers under Pakistan's control with the eastern rivers diverted upstream by India (WAPDA 1982) In addition, the system has three major reservoirs, (Mangla, Tarbela and Chasma at the upstream "rim" of the Indus plains) which regulate, as well as supplement, the water needed for agriculture, power generation and other purposes (Fig 1 3)

Tarar (1982) stated that roughly 70-80% of the total annual runoff from the Upper Indus Basin originates as snow and ice melt in the Himalaya, Hindukush and

FIGURE 1 3



SCHEMATIC DIAGRAM OF INDUS BASIN SYSTEM

(Source Water and Power Development Authority of Pakistan (WAPDA) 1982)

Karakoram Thus, for effective river management not only man-made but also natural storage has to be considered in the form of perennial ice and snow in these high mountains. The overwhelming role of these mountains in general, and the Karakoram in particular in the hydrology of the Upper Indus Basin draws attention to the meteorological and hydrological conditions prevailing in these mountain ranges.

To meet the food and fiber requirements of its growing population, Pakistan has given high priority to water resource planning, development and management including flood control In 1985 a project was established with the help of the Canadian Government under the title "Snow and Ice Hydrology Project" (S.I.H.P), to study the snow and ice conditions above 3000 m a.s.l in the Upper Indus Basin This is a collaborative project funded jointly by the Canadian International Development Research Centre (IDRC), the Water and Power Development Authority of Pakistan (WAPDA), and Wilfrid Laurier University, Waterloo, Canada The main purpose of this project is to provide necessary information regarding snow and ice conditions in the Upper Indus Basin with a view towards developing a monitoring and forecasting network so that Pakistan can better manage the water resources from the high mountains

The hydrological system of the Upper Indus Basin is complex It combines runoff from glaciers, snowmelt and rainfall These are further complicated by variable snowcover, in space and time and by the migration of melting temperatures with altitude Snowmelt is the larger fraction of water supply in the western Indus streams in most years A thorough understanding of the processes by which accumulated winter snow pack is converted to spring and summer streamflow, is fundamental to our research and operation programs.

In order to determine the proportion in which snowmelt water contributes to the total discharge of the Indus river system in the different seasons, one important parameter is the altitudinal oscillation of snowline during the year Obviously area snowmelt can occur above the transient snowline Avalanched snow may lie below that and contribute to runoff from lower altitudes later in the year

The basic concern, therefore, is to look at snowmelt as a fundamental component of hydrology of the Upper Indus Basin This provides the opportunity to study aspects of regional hydrology, to define some hydrologic and climatological parameters, and indicate how they are important for runoff forecasting It should be noted that seasonal snowmelt (as opposed to glacier ablation) occurs in two main forms, direct *area snowmelt* if the snowpack where it fell, and *avalanche* snowmelt (de Scally 1987) Both are involved in the present study, but the emphasis is upon the former

The meteorological records, presently available for the Upper Indus Basin, come from weather stations which lie in the main towns They are not only in rain shadow climates that make little contribution to runoff but are subject to the powerful topoclimatic effects of valley wind systems (Butz and Hewitt 1986) Most of the precipitation in the Karakoram Mountains occurs at elevations above 3000 m a.s.l It is this precipitation in the form of snow that along with the high Himalaya and Hindu Kush catchment areas, creates the large moisture surplus in the region and provides the bulk runoff in the Upper Indus Basin Rivers by subsequent melting

The present study is an attempt to isolate the snowmelt component in the Karakoram It is carried out by taking hydrological and climatological observations

in two small snow fed basins having opposite aspects in a glacierized basin (Barpu Glacier Basin) in the Central Karakoram. The observational network is designed to cover a range of elevations within the experimental basins This type of network is essential to take account of the topoclimatic and microclimate effects on hydrology The study attempts to define the relationship between certain hydrological and climatological parameters. It also provides unique hydrometeorological information about the highest, and least studied, glacierized area in the world (Mayewski et. al 1984)

1.2 RESEARCH OBJECTIVES

The exceptionally high and rugged mountain environment of the Karakoram controls snowfall occurrence Precipitation inputs vary greatly in quantity and regime with altitude, topography and aspect Similarly, redistribution of snow from wind and avalanches can have great affect on the subsequent runoff regime Two outstanding factors which significantly influence the whole hydrological system in the Karakoram are the great variation of snow pack water equivalence with altitude, and the altitudinal migration of melting temperatures over the hydrological year (Hewitt 1985) These two factors combine to ensure that only a fraction of the whole Upper Indus Basin - probably less than 30 percent - contributes perhaps more than 80% of the river's flow (Hewitt 1988) Most of the melt water contributing to streams originates from 3000 - 5500 m a.s.l and dominates the summer hydrograph till mid July (Hewitt 1985) Above 5500 m a.s.l there is often heavy snowfall accumulation but little melting due to freezing temperatures

The absolute and relative contributions of snow and ice to the Indus flow vary

enormously from year to year In most years they compensate each other For example, in a winter with above average precipitation, snow accumulation on the glacier will retard the onset of ice ablation as a result of the thickness of the snow to be melted before underlying ice is exposed Since snow has an albedo higher than that of ice the amount of melting under direct radiation will be considerably reduced If winter snowfall is light the glaciers tend to melt more, compensating for lower precipitation (Young 1981) However, the timing of snowmelt and glacier melt is different Snowmelt is critical early in the summer, glacier melt in the later part.

In the Karakoram mountains, energy inputs available for snow and ice melt in general are negligible in winter and intense in summer But, superimposed on this dominant pattern of variability in energy inputs with seasons is the smaller scale variability resulting from seasonal or short term weather and the variety of slope aspects within any one basin The difference on north and south facing slopes can be critical in influencing the *timing* of runoff In summary the purpose of this investigation is to define some of the hydrological and climatological parameters which control snowmelt runoff Although this paper is not directly concerned with the prediction of runoff, the relationships considered are among those required in the formulation of prediction models The specific subset of problems in this research concern an understanding of

1 the determination of environmental temperature lapse rate with elevation,

2 the spatial and temporal relationship between snow cover depletion and snow melt runoff volume,

3 the determination of seasonal and diurnal stream flow patterns in high mountain basins,

4 the relationship between snowmelt causatives such as air-temperature, radiation, wind speed, relative humidity, and cloud cover with runoff volume,

5 the effect of aspect on each of the above

1 3 LITERATURE REVIEW

1 3.1 Variation of Precipitation with Elevation

The name Karakoram means "ice mountains", a mass of rock and ice extending for 402 km from the Shyok to the Hunza. Besides the permanent snow in this region, snow accumulates over much larger areas in winter and melts in the subsequent summer However, observational data concerning snowfall are very meagre Therefore, only a qualitative description of the snowfall has been possible on the basis of findings of various expeditions Here, the findings of a few studies made in the Himalayas are presented to demonstrate the fact of increased precipitation with elevation. Rainfall data collected by Pakistan's Meteorology Department at Gilgit and other valley stations, and measurements below the snout of the Batura Glacier (Batura Investigation Group 1976) show annual totals of only 100-200 mm. That is less than the evapotranspiration calculated by Hewitt and Butz (1986) However, it is clear that precipitation exceeds potential evapotranspiration roughly from 3000m a.s.l in the Central Karakoram The following three facts suggest much higher precipitation input at higher elevations.

1 large stores of perennial snow and ice, in the form of valley glaciers comprise 50% of all glaciers outside of the polar regions and contain approximately 33 times the areal cover of the glaciers in the European Alps over a similar area (Wissman, 1960)

2 higher annual runoff of the rivers draining these mountainous basins – than the precipitation input recorded at the valley meteorological stations.

3 greater accumulation of snow, of the order of 700 - 1000mm above - 4500m a.s.l is reported by numerous expeditions (Hewitt 1968, Batura Glacier Group 1979, Yafeng and Wenying 1980)

This evidence indicates that precipitation must increase rapidly with elevation. It is also supported by the fact that high-altitude terrain receives more precipitation due to forced or orographic lifting of air masses as they cross the highlands. Also the decrease in air temperature with increasing altitude, helps in producing precipitation as snow, rather than as rain This topic is discussed in more detail in Chapter 3

The only comprehensive study to demonstrate the phenomenon of increased precipitation with elevation in Karakoram Mountains is that of the S.I.H.P Annual Reports (Hewitt 1986, 1987, Wake 1987) However, some glaciologists had earlier provided some quantitative information from their work in Himalayas Gilbert et. al (1969) during their work on a northwest facing cirque glacier in the Hindu Kush estimated the mean net accumulation of snow water equivalent 1300 mm annually at an elevation of 5809 m a.s.l This study was based on an 8 year record of snow measurements in the bergschrund where each yellow-brown ice layer was interpreted as the summer surface They suggest that most of the snowfall occurs during late winter or spring

In 1973 a Chinese group recorded a net winter accumulation 1030 - 1250 mm water equivalent at an elevation 4840 m a.s.l. near the equilibrium line of the Batura Glacier They mentioned that even though a considerable amount of the accumulated snow melts and evaporates in summer every year near the snowline, it is still ten times greater than the annual precipitation measured at 2680 m a.s.l in the valley bottom (Batura Investigation Group 1979) During summer 1974, Batura Glacier Group (1976) reported twice as much precipitation at 3400 m a.s.l. than at base camp 900 meter lower on Batura Glacier

The Gara Glacier, in the Western Himalaya was studied by Raina, Kaul and Singh (1977) They determined the net balance for the 1974-75 season over the entire glacier (4700-5600 m a.s.l) A positive balance was recorded above 5050 m level The mean net accumulation of 5 elevation bands from 5400 - 5600 m a.s.l was 2250 mm water equivalent They suggest that the basin receives very little precipitation during the monsoon period and the majority of snowfall occurs during the winter with as much as 6 m of snow accumulation

Yafeng and Wenying (1980) and Yafeng (1980) stated that on the Batura Glacier at the firn basin near the snowline, the annual net accumulation ranges between 1,000 and 13,00 mm But on the basis of measured ice thickness of annual layers in the firn basin and the annual discharge of the melt water of the glacier, they estimate that annual precipitation above the snowline may be over 2,000 mm

Wake, (1987) on the basis of chemical analysis and physical characteristics of fresh snow samples collected during 1985-86 from the snow accumulation area of Biafo Glacier, Central Karakoram, suggests that precipitation increases with

increasing elevations and the maximum zone of accumulation occurs in the elevation band from 4900-5100 m a.s.l and decreases thereafter The net annual accumulation calculated from a two year record was 1900 mm water equivalent. Table 1 1 shows the net increase of snow accumulation in different elevation bands on the Biafo Glacier and Table 1.2 shows both an overall view of snow accumulation on different gliciers in the Central Karakoram Mountains and zones of maximum accumulation These findings and water yield calculated from the gauging records of the rivers draining these catchment areas, show that precipitation over much of the Karakoram Mountains exceeds 10 to 20 times that recorded at valley weather stations. However, little is known about the actual shape of the precipitation gradient. Hewitt (1985) stated that the delay in the rising limb of the hydrograph of the rivers draining Karakoram Mountains as compared to those of Cis-Himalaya, reflects a much thinner snow cover below 3500 meters. It does not increase the river flow appreciably until melting reaches the middle zones (3500 - 5000 m a.s.l) where the snowpack is much thicker This reflects the gradient of snow cover but the presence of rugged topography and great relief makes the actual shape of precipitation gradients more complex.

1 3.2 Ablation in Karakoram

The Karakoram Mountains, a major contributor of runoff from the Upper Indus Basin, remains one of the least studied regions. This is especially true of the behavior of snow and ice melt with increasing altitude Few studies have attempted to measure the ablation of glacier ice and characteristics of snow melting in the accumulation areas of Karakoram glaciers.

Table - 11 Rough estimate of moisture input in the accumulation zone of the Biafo Glacier, Central Karakoram (1985-86)

Elevation band	Area (Km^2)	Net annual ACC (m w.e)	10^6 m^3 Water
4572 - 4877*	84 0	1 0	84 0
4877 - 5181*	109.5	1 8	197 1
5181 - 5486*	99 2	1.5	148 8
5486 - 5791*	61 8	1 0	61 8
5791 - 6096	27 6	0 8 (?)	22 1
6096 - 6401	8 6	0 6 (?)	6.2
6401 - 6706	1 3	—	——
		Total	519

*Elevation bands for which data exists. (Source Wake 1987)

Table - 12 Net annual accumulation of snow water equivalent in Central Karakoram

Relative Elevation	Snowpit	Accumulation m w e
High	Shark Col (5660 m)	0 71
	Hispar Dome (5450 m)	1 20
Middle	Approach Glacier (5100 m)	1 88
	Whaleback Glacier (4900 m)	1 79
Low	Hispar Glacier East (4850 m)	1 07
	Equilibrium Line (4650 m)*	0 90

* Value for annual accumulation only (Source Wake 1987)

Loewe (1959) recorded 7-8 cm of ablation per day on the Chungphar and Bazhin Glaciers (elevation range 3050 - 3630 m a.s.l) in the middle of September Loewe concluded that 55 - 60 percent melting of snow would be due to radiation and no change in the ablation rate was observed with the change in elevation of 610 m Gilbert et al (1969) concluded from the behavior of streamflow fed by a northwest facing cirque glacier in the Hindu Kush Mountains, that streamflow response radiation melting by showing minimum flow at 10 hours local time and maximum about 18 hours.

The Batura Glacier Group (1979) found that 89.2% of the total heat for ice ablation is derived directly from solar radiation and 10 8% from heat conduction of air and condensation heat of vapor Yafeng (1980) has stated that maximum depth of annual ablation on the bare ice of Batura Glacier at 2600 m a.s.l measured was 18 4 meters per annum, but this ablation rate decreased markedly to the level of 4 36 meters per annum on the lower 20 kilometers which is covered with thick debris.

In another study Yafeng and Xiangsong (1984) measured ablation of Batura Glacier by setting 66 stakes on 16 cross sections on the surface of the glacier during the period of intensive ablation. Four of the stations near the glacier terminus recorded whole year ablation. From these measurements they found 18 94 m of ablation at an elevation of 2644 m a.s.l Yafeng and Xiangsong concluded that mean daily temperature is the most suitable index of surface ablation on the Batura Glacier

Ablation measurements recorded during summer in 1985-86 by S.I.H.P on the Biafo Glacier show average ablation rates in relatively clean ice, of about 6 - 7 cm

per day between the elevation range 3885 to 4080 m a.s.l. These figures are subject to variation with aspect, elevation and the concentration of dust particles on the ice Results indicate that the ablation rates on a medial moraine between elevation range 3885 to 4080 m a.s.l. on Biafo Glacier under extensive debris cover are roughly half that of relatively clean ice, seasonal snow cover on the glacier surface takes longer to melt than relatively clean ice once exposed (Snow and Ice Hydrology Project Annual Report, 1985)

Surficial debris is a wide-spread phenomenon of the Karakoram Glaciers. The rate of ice melting beneath such debris cover depends upon their thickness. A study was carried out on the Rakhiot Glacier on the north slope of the Nanga Parbat during field season 1986 by Gardner (S.I H.P Annual Report 1986) Preliminary results of ablation measurements indicate that there is a threshold thickness of debris cover of about 1 0 cm above which ablation rates decrease with an increase in debris cover (Fig 1 4)

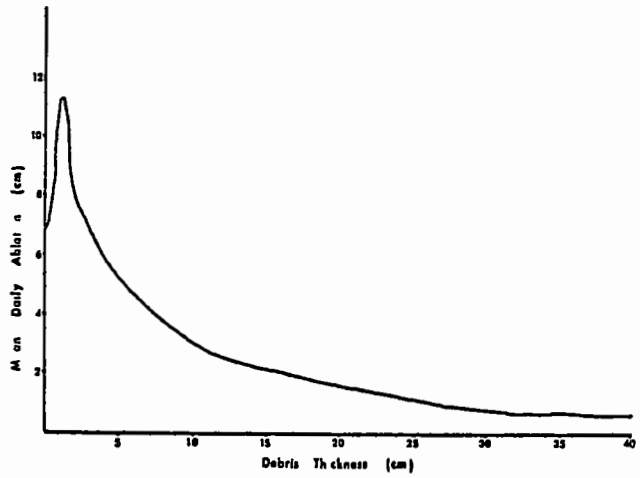


FIG 14 Relation of mean daily ablation to till cover thickness on the Rakhiot Glacier 1986 (Source Snow Ice Hydrology Project Annual Report 1986)

The threshold value of thickness is attributed to the phenomenon that a thin layer of debris, rather than insulating the underlying ice, absorbs shortwave and longwave radiation, which is quickly transmitted to the ice surface by conduction. This contributes to rapid ablation However the rate of ablation depends upon the size and physical characteristics of the debris cover and local topographic conditions (Gardner, 1986)

1 3.3 Snow-cover Area in Relation to Streamflow

Numerous studies have utilized satellite derived snowcover estimates to predict seasonal runoff from the Upper Indus Basin (UIB) Salomson and MacLeod (1972) mapped the areal extent of snowcover in the UIB for the years 1969 - 1970, using an image dissector camera on Nimbus 3 They related increasing runoff to decreasing snowcover area. Results indicated that it might be possible to predict seasonal runoff by monitoring seasonal snowlines and snowcover area using satellite imagery

Rango et. al (1977) using LANDSAT imagery measured snow covered area with a planimeter for the Upper Indus Basin above Besham Qila (162,000 km^2) and the Kabul River above Nowshera (88,600 km^2) They attempted to correlate this with seasonal runoff for the period 1967-73 using a regression equation The seasonal flow predicted for the year 1974 was within 7% and 2% of the observed seasonal discharge on the Indus and Kabul Rivers, respectively They conclude that the degree of confidence is not great for making particular flow predictions, but in the absence of appropriate hydrometeorological networks, snowcover estimates from satellite imagery may provide best estimates of seasonal flow

Tarar (1982), using LANDSAT-2 imagery calculated the snowcover area for different rivers of the Upper Indus Basin He developed a regression equation between snowcover area and subsequent runoff for the Shyok, Hunza, Gilgit, Indus at Besham and for other rivers of the Upper Indus Basin for the years 1975-78 The results indicate that regression equations are statistically significant The correlation coefficient between snowcover area and flow represented by r^2 varies from 0 889 to 0 996 On the basis of four years record, deviation between computed and observed runoff was within ±10% Tarar suggests that timely availability of Landsat data is essential to make use of this simple, inexpensive and less time-consuming method

Dey et. al. (1983) attempted to improve the relationship developed by Rango et. al. (1977) by extending the period of record from 1969 through to 1979 The techniques used are similar to those described by Rango et. al (1977) The results shown are little improved with 10% difference between the estimated and observed discharge values averaged over the 11 year study period They conclude that satellite derived snow-covered area is the best available input for snowmelt runoff estimation in remote, data sparse basins like the Indus and Kabul Rivers

Rango et. al (1983) compared LANDSAT derived data with conventional methods for use in hydrologic models in six watershed basins throughout the USA They suggest that the difference between the two data inputs with hydrologic models is minimal and, for basin larger than 10 square miles, the LANDSAT technique is more cost effective

However, Makhdoom and Solomon (1986) have critically examined the studies developed specifically for forecasting purposes in the Indus River Basin in Pakistan

They conclude that it is premature to assume that snow cover area in the basin at the time when the snow-melt begins is a sufficient index for operationally predicting flow volume during the snow-melt period It is possible that the errors of forecasting are related, to large extent, to the initial assumption that snow volume is directly proportional to snow cover area This assumption involves ignoring the aspects of snow depth and density and the effects of glacier-melt. They further mention that significant correlation coefficients between flow volume in spring - summer may be the result of random factors as the period of record on which these correlations are based is short. It certainly appears doubtful in a basin where i) most of the water yield from seasonal snowpack comes from less than half the area (that is from the elevation higher than 3000m a.s.l where precipitation is much greater), and ii) where glacier melt provides about half the annual runoff, and is likely to compensate the poor snowfall years and vice versa.

CHAPTER 2

Climate and Geography

2 1 OVERVIEW

From meteorological and hydrological view points, Pakistan is a country of extremes. Some areas experience extremely heavy precipitation in some periods and long dry spells in others. Even in some parts, very arid areas are located near extremely humid areas.

The variation of meteorological conditions in Pakistan generally, and in the Upper Indus Basin particularly, is due to its topography and geographical location The wide Indus valley is limited at its western side by the Soleiman Mountains, to the north and northeast by the Himalaya Mountains, and to the east by a low plateau separating it from the Ganges Plains (Fig 2 1) Air masses can move virtually unimpeded across the Indo-Gangetic Plains but are impeded northwards by the mountains. Due to this configuration, the meteorological conditions in Pakistan, particularly its northern portions, are influenced by weather systems developing not only in the adjacent Arabian sea, but also by those originating in the Gulf of Bengal (Monsoonal) and Mediterranean Sea (Atlantic)

Air masses originating from the Atlantic Ocean and Mediterranean Sea,

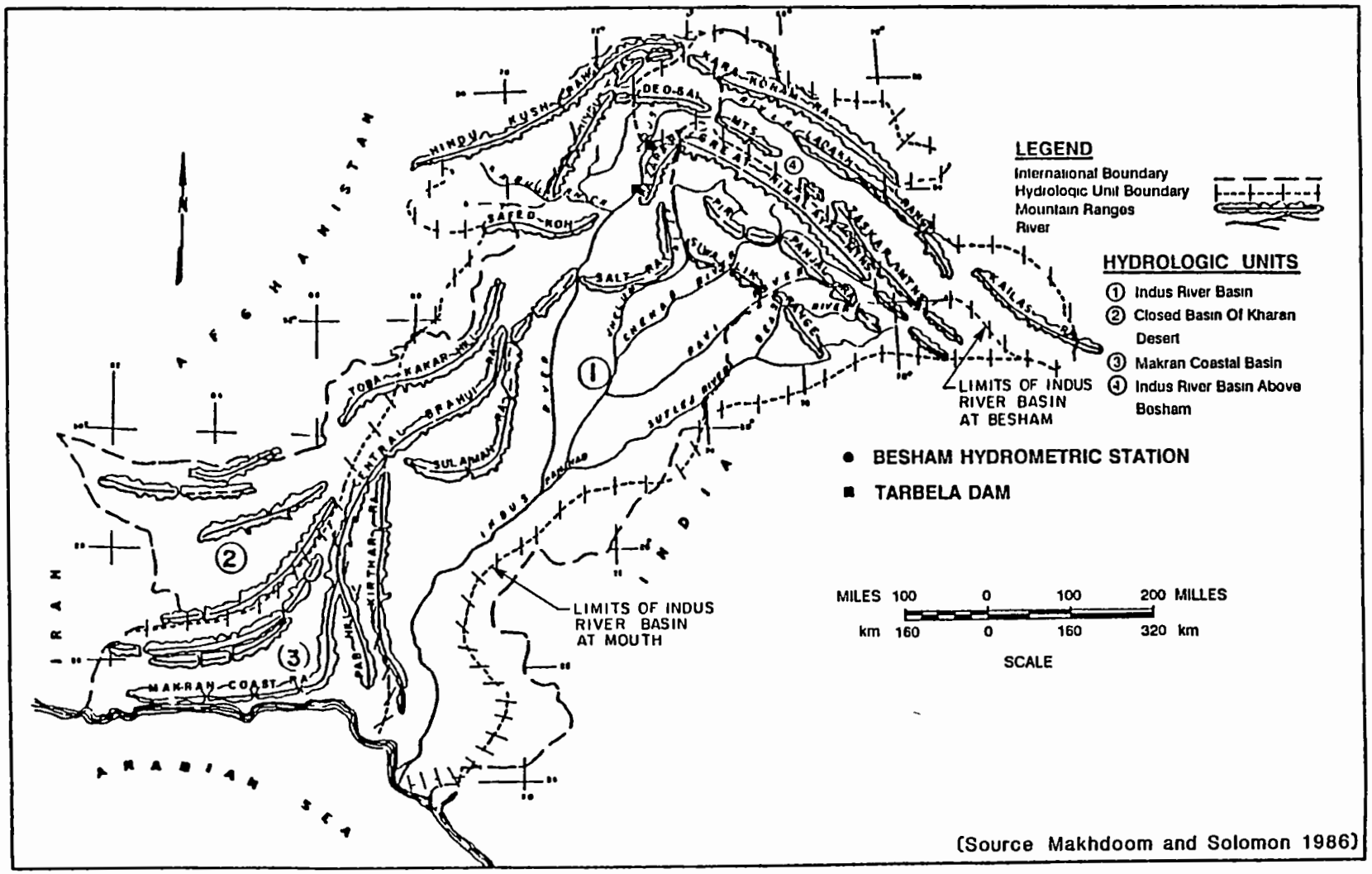


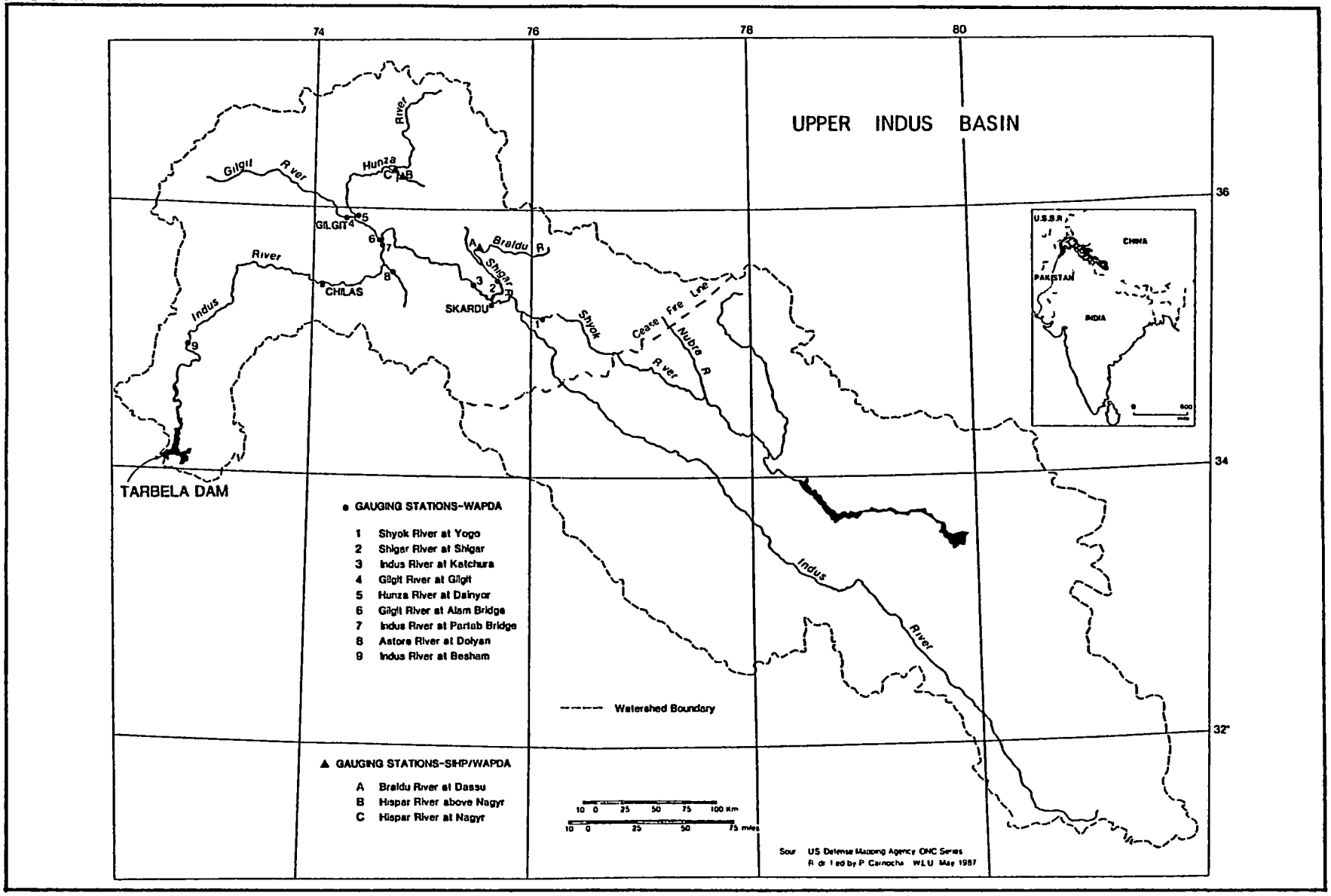
FIG 2 1 Pakistan location map, physiographic and hydrologic units

especially in the winter are a major moisture source for the northwestern part of the country Here the Karakoram Range stores a large quantity of the moisture input in the form of glaciers. Most of the Karakoram lies within the Upper Indus Basin watershed above the Tarbela reservoir and covers an area approximately 164,000 Km^2 (Fig 2 2) The average runoff contributed by the Upper Indus Basin rivers measured at Besham Qila above the Tarbela Reservoir is 2370 m^3/sec annually (Snow and Ice Hydrology Project Annual Report 1986)

The Indus Waters Treaty was a critical bench mark in the economic development of the Indus Basin The evolution of irrigation and construction of an irrigation-based infrastructure in the Indus Basin furnished a major field of research and management for hydrologists, economists and geographers Study of this system provides a case book example of how humans have been able to adapt to, and utilize water resources in a difficult environment

The Indus Water Treaty has increased Pakistan's dependence on water from the main Indus, which has consequently increased the dependence on snow and ice melt in the Himalayas, especially the Karakoram and surrounding Ranges The latter is the major source of snow and ice melt water during July and August Effective use of this meltwater is only possible through an understanding of snow and ice conditions and the terrain characteristics in the Karakoram In particular a knowledge of the hydrological and climatological variations which control melting during summer is essential This melting depends upon the surplus moisture present in the form of valley glaciers, seasonal snowpack and the thermal conditions. Such understanding will help Pakistan to predict runoff and manage the reservoirs and canal system more effectively

FIGURE 2 2


UPPER INDUS BASIN
74
76
78
80
36
34
32°
Gilgit River
Hunza River
GILGIT
CHILAS
SKARDU
Indus River
Braldu R
Shigar R
Shyok River
Nubra R
TARBELA DAM
U.S.S.R.
CHINA
PAKISTAN
INDIA
GAUGING STATIONS-WAPDA
1 Shyok River at Yogo
2 Shigar River at Shigar
3 Indus River at Katchura
4 Gilgit River at Gilgit
5 Hunza River at Dainyor
6 Gilgit River at Alam Bridge
7 Indus River at Partab Bridge
8 Astore River at Dolyan
9 Indus River at Besham
Watershed Boundary
GAUGING STATIONS-SIHP/WAPDA
A Braldu River at Dassu
B Hispar River above Nagyr
C Hispar River at Nagyr
10 0 25 50 75 100 Km
10 0 25 50 75 miles
Sour US Defense Mapping Agency ONC Series
R dr I ed by P Carnochs WLU May 1987

2 2 GEOGRAPHY OF THE KARAKORAM

2 2 1 Regional Setting

Immediately north of the Greater Himalaya lies a series of well outlined ESE-WNW trending mountains forming a range called the Karakoram. The name Karakoram means "black gravel", a mass of rock and ice extending for 402 km from the Shyok to the Hunza, with the greatest assemblage in the world of giant peaks - 33 over 7,315m a.s.l - culminating in the tremendous keeps of the three Gasherbrum summits, all over 7,925m a.s.l and finally K2 8489m a.s.l (Fig 2 3) These mountains are parallel to the main rock formations which extend over 2500 km from the Hindu Kush to the eastern Himalaya The Karakoram range is surrounded by the Greater Himalaya to the southeast, the Kunlun Shan to the northeast, the Pamirs to the northwest and the Hindu Kush to the southwest (Fig 2 3) These mountains not only remain permanently covered with snow throughout the year but some of the largest glaciers outside the polar regions such as the Batura, Biafo, Chogolungma, Baltoro, Hispar, Rimo and Siachen are also found here According to Wissman (1959), about 37% of the area is ice-covered The rainfall is very low and snow melt (including glaciers) is the main source of water contribution to the rivers in this region (WAPDA 1982) Most of the meltwater from the Karakoram flows into the Indus through its major tributaries such as Shyok, Shigar, Hunza and Gilgit Rivers (Fig 2 2) It is this water which is more precious than land in the Indus plains and sustains human settlement in the mountains as well as in the plain areas of Pakistan

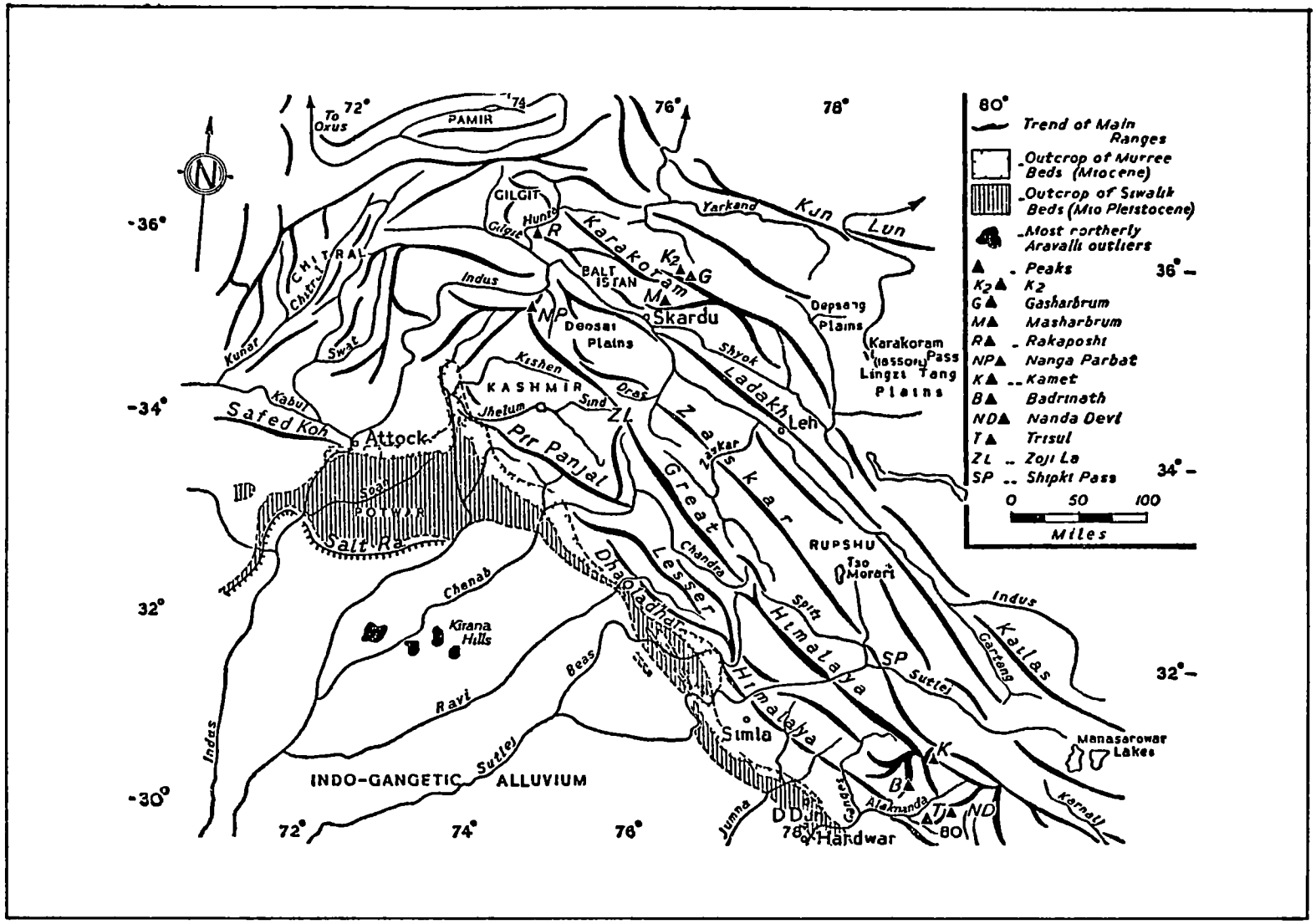


FIG 2 3 Physiography of the Northwest Himalayas (Source Spate and Learmonth, 1967)

2.2.2 Glaciers of the Karakoram

The Karakoram range has an extensive ice-cover, which extends from Batura Mustagh (7,795 m) in the west through K2 (8,611 m) to Sasir Kangri (7,672 m) in the east. The Karakoram Mountains are distinguished from surrounding ranges by greater ice cover (i e 37% of the total area), longer valley tracts subject to contemporary glacial action, and lower penetration of large glacier tongues (Hewitt 1968, Goudie et al. 1984)

The largest glaciers of the region are those of the southern face of the Karakoram They descend well into semi-arid areas All these glaciers discharge into the Indus - the Hispar and Batura, 58-61 km long - while the Biafo and Baltoro glaciers of the Shigar river, a tributary of the Indus, are about 60 km in length (Wadia 1968) Accumulation zones of the Karakoram glaciers extend from 4500 to 7000 m a.s.l This shows the high rate of precipitation with increased elevation which is attributed to an extreme orographic effect (Hewitt 1968) More than 60% of the Indus flank of the greater Karakoram is covered with glaciers. These hold great importance in the hydrological cycle as runoff from these highly glacierised basins is concentrated in the same period i e from June to mid-September (Hewitt 1968)

2.3 BROAD CLIMATIC CONTROLS

The Himalayas lie in the subtropical high-pressure belt where seasonal meridional migration of pressure and wind systems greatly alter the weather regime between the different months In winter, the middle latitude westerlies sweep over the ranges and precipitation comes from the "troughs of low pressure" in the

westerly circulation During the monsoonal months most part of the Himalayas are under the sway of easterlies, in contrast to winter

The Karakoram range of the Himalayas Mountains extend well into the troposphere and are directly affected by the upper air system (Hewitt 1968) Therefore, in winter the subtropical westerly jet stream in the upper troposphere, is steered by this system, and split into two currents, one to the north and one to the south of the 4000m a.s.l high Tibetan Plateau. The two currents reunite again off the east coast of China. The creation of the two branches is attributed to the disruptive effect of the topographic barrier on the airflow The branch over northern Pakistan and India corresponds to a strong latitudinal thermal gradient and it is probable that this factor, combined with the effect of the barrier to the north, is responsible for anchoring the southerly jet (Barry & Chorley 1970, Rao 1981) This southerly branch is stronger, with an average speed of 66 m s^{-1} at 200-300 mbar, compared with about 20 - 25 m s^{-1} in the northern branch (Hewitt 1968, Barry & Chorley 1970) Air subsiding beneath this upper westerly current brings dry northerly winds from the subtropical anticyclones to northwest Pakistan and India

Equally important is the steering of winter depressions over Pakistan-India by the upper jet. The lows, which are not usually frontal, appear to penetrate across the Middle East from the Mediterranean and are considered to be the main source of precipitation falling on the Upper Indus Basin (Hewitt 1968, Barry & Chorley 1970) These troughs in the westerlies are most marked in winter and give more precipitation in the Western Himalayas than in the eastern, the former being higher in latitude by four degrees (Rao 1981)

Rao (1981) and Boucher (1975) stated that on average 5 to 8 disturbances a month affect northern Pakistan and India from December to April However, the numbers vary greatly from year to year, and conditions depend upon scale, frequency and intensity of these westerly disturbances (Hewitt 1968) Rao (1981) found that winter precipitation around 76°E on Karakoram ranges reaches a maximum, as it is consistently higher than any other section, and well north of the main Himalayas According to him, the ranges in Pakistan north of 34° N receive approximately 1500 mm water equivalent of snow between October and May reaching maximum in January and February He also stated that 40% of the precipitation may occur as rainfall at places 2000-3000 m a.s.l in elevation Data collected by S.I.H.P indicate that on average 900-1500 mm water equivalent fall in the elevation band 4500-6500 m a.s.l., but much less below 3000 m Out of this total precipitation about 70% falls during winter and 30% in summer (Wake 1987)

In May or June the northern branch of the subtropical jet stream decreases markedly in strength from about 25 m s^{-1} to about 10 m s^{-1} and by mid-June is altogether diverted to the north of the Tibetan Plateau (Barry 1981) While this is occurring an area of high pressure is established over the Tibetan Plateau due partly to surface heating by the sun As warm air is less dense than cold air, pressure falls less rapidly with height over the Tibetan Plateau The formation of the Tibetan anticyclone is concurrent with the development of a high level easterly jet stream over southern Asia (at ~15°N) Wind speed in the core of this easterly jet stream often exceeds 51.5 m s^{-1} At the same time the summer monsoon begins to move across the Indian subcontinent.

2.4 ROLE OF MONSOONS IN THE WESTERN HIMALAYAS

The effect that the upper air flow has on the climate of the Karakoram is not clear, as the regional airstream continues to be influenced by the westerlies and westerly depressions may continue to affect the region throughout the year (Hewitt 1968) However, temporary destruction of the Tibetan anticyclone can result in the incursion of monsoonal air masses into the Karakoram, resulting in heavy precipitation The potential for fluctuations in the summer circulation creates an environment that is open to substantial variability

Several studies summarize the effect of monsoons in the Western Himalayas. notable amongst these are Finsterwalder (1960), Mayewski & Jeschke (1979), Mayewski et. al (1980), and Mayewski and Lyons (1983, 1984) Finsterwalder (1960) reported heavy rains in the Hunza Valley (Central Karakoram) and intensive snow storms at higher elevations due to the monsoon between July 2-5, 1959 A return period for such monsoonal intrusion is reported to be fifty years (Finsterwalder 1960; p 787) Mayewski et. al (1979, 1980), attributed the advances of Trans-Himalayan Glaciers during the period 1890-1910 to monsoonal currents and to secular variations in Indian rainfall In the most recent studies Mayewski and Lyons (1983, 1984), on the basis of snow chemistry, suggested two different sources of precipitation for lower and higher elevations in the Ladakh Himalayas They collected a series of fresh snow samples over a range of elevations in the Ladakh Himalayas in 1983 All fresh snow samples were collected within 24 hours, following the end of the precipitation event They noted a distinct change in chemical contents of fresh snow between 5250m and 5300m a.s.l On the basis of this they concluded that northerly and southerly flowing warm air masses dominate

precipitation input at low levels, and relatively cold air flowing easterly and westerly dominates at higher levels

The relative roles of westerly disturbances and monsoonal air masses during summer in the Karakoram Mountains are still not clear However we have observed significant amounts of precipitation under the influence of monsoonal incursions across the whole of the Karakoram Ranges during two of our field seasons in 1985, 1986, and heavy snowfall at higher altitude and rain at lower altitudes continued into August 1987 from westerly depressions (Hewitt Pers Comm 1988) Chemical results of snow samples collected in the summer of 1986 from Biafo Glacier's Snow Lake in the Central Karakoram, indicate alternating deposits of snow from Atlantic or Mediterranean Ocean (winter) and Arabian Sea (summer) (Wake 1987) Monsoonal influence is more prominent towards the south and east of the Upper Indus Basin as the Himalayan front ranges receive heavy precipitation during summer Further north and west this source of moisture is comparatively less important and westerly depressions give most of the precipitation However all of these regional climatic conditions are powerfully modified by topography and the great altitudinal ranges of the Karakoram Mountains

2 5 LOCAL CLIMATE The effect of altitude and topography

Altitude alone has a significant effect on local climate in addition to locational factors Wind speeds in the mid-troposphere are generally higher than those close to sea level, an important factor in relation to snow movement and packing, evaporation and the effect of topography-controlled wind systems in exposed mountain masses The density and composition of the atmosphere changes with

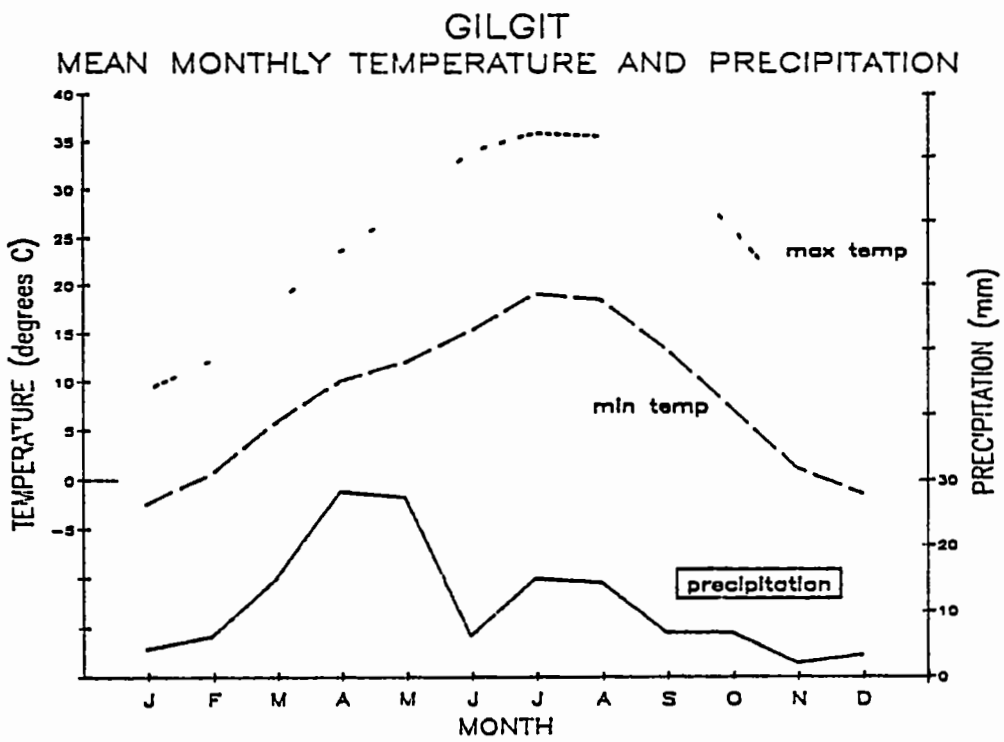
altitude, having a marked effect on solar radiation received at the surface Both the water vapour (chief absorber of solar radiation) and aerosol content decrease with elevation, so the direct solar radiation reaching the surface increases (Barry 1981) Air mass movements modify this picture locally, but diurnal and day-to-day fluctuations in solar radiation at the surface in high mountains are invariably very large (Hewitt 1968) Differences in receipt of radiation resulting from topographic factors are reflected in air temperature and snow cover duration South facing slopes receive direct radiation for much longer periods during the day, resulting in a greater degree of melting The transitory snow line on south facing slopes at any given time during the ablation period is usually higher than on north facing slopes This has important implications for the hydrological response of snow covered basins A basin with a large percentage of south facing slopes would produce meltwater sooner under clear sky conditions While radiation is probably the major source of energy in controlling melting on south facing slopes, air temperature is presumably more important on north facing slopes (Hewitt 1968) This is also evident from the number and extent of small cirque or hanging glaciers which occur less often at higher altitudes on south facing slopes in most of the valleys in Karakoram Mountains

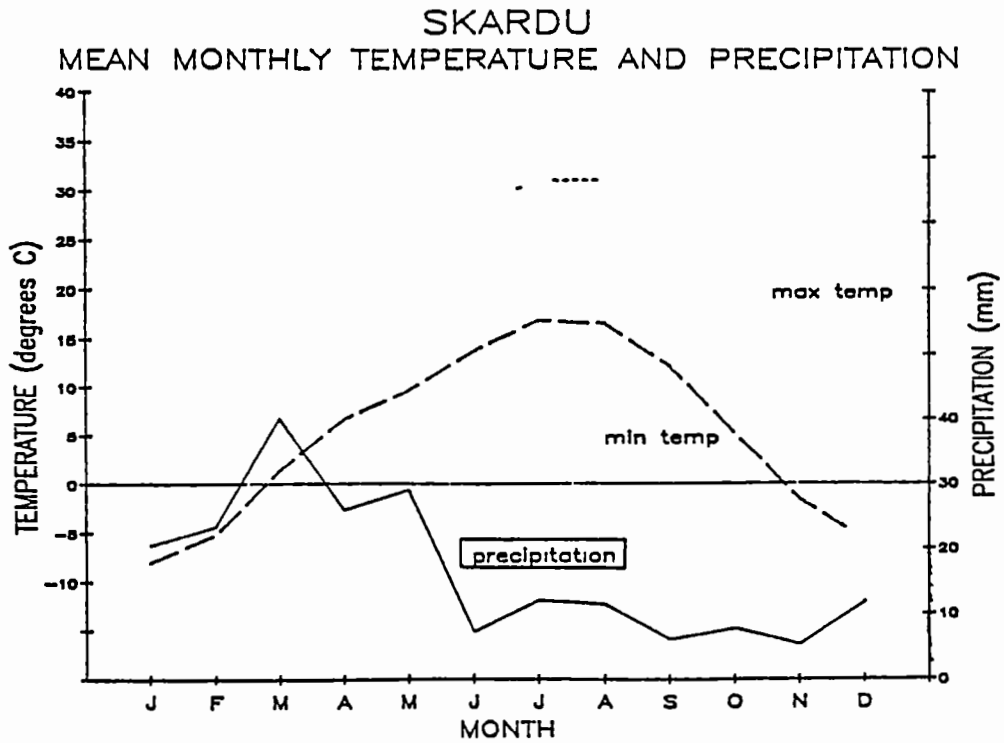
The interaction between topography and meteorological elements involves several basic characteristics of any relief feature The overall dimensions and the orientations of a mountain range with respect to prevailing winds are important for large scale processes, relative relief and terrain shape are particularly important on a regional scale, while slope angle and aspect cause striking local differentiation of climate

Snowmelt runoff provides a major portion of the volume of flow in the Upper Indus Basin rivers In order to understand how snowmelt converts into streamflow, it is necessary to know the altitudinal oscillation of the snow line during the year, as altitude, topography, orography and aspect have considerable effects on the local climate of the Karakoram Range and thence on snow hydrology Outstanding features of this variability are very significant increase in precipitation with elevation, rainshadow effects and the existence of desiccating valley wind systems (Hewitt 1985)

Most of the Indus valley in Pakistan has a moisture deficit, rainfall is exceeded by evapotranspiration Lockwood (1974) suggests that evapotranspiration is more than 1,000 mm per year, and considers the climate to be arid (Lockwood 1974, p 166) The main river valleys below about 3000m in the Karakoram are also extremely arid Mean annual precipitation measured in the valley bottom towns of Gilgit and Skardu is 132 and 202 mm respectively (Fig 2 4) Butz and Hewitt (1986), also calculated a negative moisture balance both for Gilgit and Skardu This negative moisture budget together with dry valley wind systems creates a severely desiccated landscape up to 3000 m a.s.l in the western and 5000 m a.s.l in the eastern Karakoram Any agriculture that does exist is supported by the water from the snow and ice melting at higher elevations (Butz 1987) The effects of altitude and topography on some important climatic parameters are discussed individually in the following paragraphs

FIGURE 2 4


GILGIT
MEAN MONTHLY TEMPERATURE AND PRECIPITATION
TEMPERATURE (degrees C)
PRECIPITATION (mm)
max temp
min temp
precipitation
J F M A M J J A S O N D
MONTH


SKARDU
MEAN MONTHLY TEMPERATURE AND PRECIPITATION
TEMPERATURE (degrees C)
PRECIPITATION (mm)
max temp
min temp
precipitation
J F M A M J J A S O N D
MONTH

2.5 1 Radiation

Radiation is considered the key factor in controlling snow melt and runoff in mountain areas where aerosol content tends to be low, and most of the water vapor is concentrated in the lowest 2,000-3,000 meters of the atmosphere (Barry 1981) The absence of aerosols with increasing altitude results in increased solar radiation being received on higher elevation slopes (Hewitt 1968, Barry & Claudia 1974)

The Upper Indus Basin in general has a high incident radiation level because of the rain shadow effect which reduces the cloud cover, especially in summer In the Barpu Basin, Central Karakoram, maximum shortwave radiation recorded on Miar Glacier near the junction of sumiayar Bar and Miar Glaciers at an elevation 3300 m a.s.l. was 1334 w/m^2 during summer 1987, which is 95% of the solar constant (1380 w/m^2) However net radiation in the mountainous areas tends to decrease with elevation as the increased duration of snow cover at higher elevations causes to reduce the absorbed radiation due to higher albedo of snow (Barry 1981)

2.5.2 Temperature

Increase in altitude in free air is, on average, associated with a temperature decrease However there are significant seasonal differences The environmental lapse rate along a mountain slope below 2000 m a.s.l may differ considerably from that in the free atmosphere, depending on the time of the day Above that, the lapse rate may be similar to that in the free air (Barry & Claudia 1974, Whiteman 1985) Temperature records available for northern areas Pakistan (Table 2 1), may not represent the true climatic conditions as most of the present climate data is recorded

Table 2 1

Mean Monthly Temperatures (°C)

Selected Upper Indus Basin Stations

Site	No of Years Record		J	F	M	A	M	J	J	A	S	O	N	D	Extreme Daily Value
Chilas	27	Min	0 5	3 2	8 5	13 6	18 1	24 3	27 5	26 9	22 8	14 6	6 7	1 8	-6 7
(1260 m)		Max	12 1	14 7	19 2	25 2	31 0	37 8	39 8	38 9	35 1	28 6	20 8	13 9	47 0
Gilgit	30	Min	-2 4	0 6	5 8	10 0	12 0	15 2	19 0	18 4	13 3	7 2	1 2	-1 4	-9 5
(1490 m)		Max	9 1	12 1	17 8	23 6	28 0	34 0	35 9	35 6	31 7	26 2	17 9	11 0	45 4
Chitral	20	Min	-0 7	0 4	4 2	8 6	12 6	18 3	20 3	19 3	13 3	7 7	3 1	-0 8	-12 3
(1500 m)		Max	8 7	9 8	14 9	21 8	27 1	34 8	36 2	35 0	31 1	25 0	18 4	11 4	44 8
Gupis	26	Min	-4 9	-2 8	2 2	7 6	11 3	16 1	19 2	17 5	13 5	7 2	1 7	-3 0	-11 2
(2144 m)		Max	4 0	6 6	12 2	18 3	22 9	29 1	32 0	31 1	26 3	20 0	13 5	6 0	40 3
Astore	25	Min	-7 2	-5 6	-1 2	4 0	7 3	11 6	15 0	15 1	10 6	4 5	-0 6	-4 7	-15 7
(2148 m)		Max	2 6	4 1	8 4	14 8	19 6	25 2	27 3	26 9	23 8	17 1	11 0	4 8	35 3
Skardu	29	Min	-8 0	-5 2	1 3	6 6	9 6	13 8	16 9	16 6	12 2	5 2	-1 6	-5 7	-18 5
(2197 m)		Max	2 6	5 1	11 4	17 9	21 6	28 3	31 2	31 1	26 6	20 3	11 7	5 5	40 0
Karimabad	7	Min	-4 0	-2 6	-2 4	7 2	10 6	13 9	16 4	17 2	11 5	7 7	2 6	-1 8	-6 7
(2405 m)		Max	2 1	4 3	9 0	16 1	20 2	25 8	28 5	29 4	23 8	18 1	10 7	4 3	37 8
Yasin	3	Min	-9 7	-7 4	-1 6	4 2	7 9	9 5	11 4	12 1	7 1	2 6	-1 9	-6 6	-15 0
(2450 m)		Max	-0 2	2 4	8 3	13 6	20 3	24 7	26 4	30 1	22 1	16 4	9 8	2 7	36 0
Naltar	2	Min	-9 7	-9 4	-4 7	0 7	4 1	8 7	9 6	12 1	9 6	2 6	-0 4	-6 1	-15 6
(2880 m)		Max	-2 8	-1 4	4 0	9 5	14 0	21 0	27 3	23 9	19 8	14 0	7 0	1 7	32 2
Babusar	2	Min	-14 7	-10 6	-3 0	3 6	(6 5)	10 7	14 5	13 4	9 4	0 4	-5 4	-12 6	---
(3003 m)		Max	-1 6	-1 8	6 0	12 6	(15 1)	20 1	25 4	23 9	19 1	10 1	3 4	-4 3	---
Misghar	17	Min	-13 2	-9 7	-5 4	-0 2	3 4	8 0	11 1	11 6	6 6	-0 1	-5 8	-10 6	-18 9
(3088 m)		Max	-1 1	1 7	7 2	12 3	16 3	21 2	24 6	25 2	20 8	14 0	6 9	0 5	32 8

Source Whiteman, 1985

from stations lying in the valley's bottoms below 3000m elevation (Hewitt 1986)

The annual variation in monthly averaged minimum and maximum precipitation and temperatures for Gilgit (1,494m a.s.l.) and Skardu (2,286m a.s.l) is illustrated in figure 2 4 Figures for these valley stations reveal large annual temperature range usually over 30°C, moisture regimes vary from subhumid to arid (Whiteman 1985)

Altitudinal gradients in air temperature are the basic control over snowmelt runoff in the Karakoram However, the most important factor is the seasonal and diurnal duration of melting temperatures which both decrease with altitude and are concentrated in fewer months of summer Data collected during the summer of 1987 by S I.H.P shows a decrease of the average diurnal temperature range in Barpu Glacier Basin from 9°C at 3510 m a.s.l to 6°C at 4572 m a.s l on southwest facing slopes.

Whiteman (1985) calculated environmental lapse rates of 6.5 - 7 8°C per 1,000 meters using the regression of mean monthly screen air temperature recorded at eleven stations over a range of elevation from 1,260 to 3,088 m a.s.l in the northern areas of Pakistan He also mentioned that minimum lapse rate is lower than the maximum except during the summer solstice period He noted however, that local aspect and topography cause sharp departures from these values The lapse rate calculated from the data collected by S.I.H.P during May to August, 1987 between elevation 3500m to 4572m a.s.l shows much higher lapse rates (i e more than 1°C per 100 meters) Comparatively higher results can be attributed to the fact that Whiteman calculated the lapse rate from the mean monthly temperature using

regression equations, and that the latter lapse rate is calculated from the summer data only This can also be attributed to the higher temperatures on the mountain slopes than in the free atmosphere

2 5 3 Precipitation

In general high-altitude terrain receives more precipitation than do the surrounding lowlands, because of forced, or orographic, lifting of air masses as they cross the highlands. Also precipitation falls as snow as a result of decreasing air temperature with increasing elevation (Alford 1985) The spatial distribution of precipitation in amount, form and to some extent timing is dominated by orographic conditions. Altitude induces upslope changes, topography induces variations due to exposure, obstruction and steering of moisture-bearing winds, and local heating and circulation effects Of spatial interest is the increase in precipitation with elevation and zone of maximum accumulation, since it controls the runoff available to different drainage units and the relative proportions of direct and melt-water runoff Record are poor for high altitude conditions, but it is clear that precipitation exceeds evaporation roughly from 3500m a.s.l in the study area of the Karakoram Above 3500m elevation precipitation is in the order of 1100 to 1600 mm per year (Hewitt 1968, p 49) A study on the Batura Glacier at an altitude above 5,000 m a.s.l shows net accumulation of snow between 1000 to 1300 mm water equivalent annually (Batura Glacier Group 1979) It is actually this snowfall at higher altitude that supports the vast snow fields and large valley glaciers in the Karakoram, and through subsequent melting contributes 70-80% of the surface runoff in the Upper Indus Basin Rivers (Hewitt 1986)

2.5 4 Wind

The most important characteristics of wind velocity over mountains are related to topography, rather than altitude effects In middle and high latitudes it is normal to expect that, on average, there will be an increase of wind speed with height, due to the characteristics of global westerly wind belts (Reiter 1963) Isolated peaks and exposed ridges of the Himalaya experience high average and extreme speeds as a result of limited frictional effect of the terrain on the motion of the free air

Recording stations across the Gilgit district measure average wind speeds of 0 9 m s^{-1} during summer months (Table 2 2)

Table 2.2
Wind at 2m height. (m s^{-1}) Selected stations in Upper Indus Basin

Site/elev	J	F	M	A	M	J	J	A	S	O	N	D
Chilas (1260m)	0 2	0.5	0 7	0 8	0 7	0 8	1.2	1 1	0 9	0.5	0.2	0 2
Gilgit 490m)	0 3	0.5	0 6	0 6	0 6	0 5	0 6	0 4	0 4	0 3	0.2	0 2
Gupis (2144m)	0 3	0.5	0 8	1 0	1 0	1 1	1 0	0 9	0 9	0 6	0 3	0 2
Astore (2148m)	0 4	0.5	0 7	0 7	0 7	0 7	0 9	0 9	0 9	0 8	0 7	0 2
Skardu (2197m)	0 3	0.5	0 8	1 0	1 0	1 1	1 0	0 9	0 9	0 6	0 3	0 2
Yasin (2450m)	0 6	1 1	1 3	1 3	1 2	1 1	0 9	0 8	0 7	0 7	0 8	0 9

Source Met Dept Lahore, except Yasin (FAO) (in Whiteman, 1985)

Wind speed measured during the summer of 1985 and 1986, on the mainstream Biafo Glacier ice at an elevation of 4080 m a.s.l ranges from 3-7 m s^{-1} (Wake 1987) However much stronger winds were observed at exposed areas with higher elevations such as 'Shark Col' (5660 m a.s.l) and Khurdopin Pass (5800 m a.s.l), Central Karakoram Maximum wind speed measured during summer 1987, in Barpu Glacier Basin Central Karakoram at an elevation of 3510m a.s.l. on a valley side is 7 3 m s^{-1} Due to marked local variability, these values are at best rough estimates of regional trends

2.6 HYDROLOGY

To appreciate the hydrology of the Upper Indus Basin, it is important to look at the course of the Upper Indus river and its longest tributary, Sutlej. These two rivers, though rising within 130 km of each other envelop the entire western Himalayas before they meet near Mithankot (Fig 1 1) The Karakoram Mountains lie almost entirely within the headwaters of the Indus river and contribute about 25% of its total flow from 15% of the catchment area, all derived from snow and ice melt (Butz 1987) The seasonal, short-term and year to year fluctuations of river flow in the Karakoram all appear distinctively alpine in character The Batura Glacier Group (1976) measured rainfall of 100 mm below the snout of the Batura Glacier Whereas, average annual runoff measured ranges between 300 - 1000 mm, which shows rapid increase in precipitation with elevation This is supported by the findings of the Batura Glacier Groups (1979), who reported that, at elevation 3400m a.s.l., precipitation doubled with an increase of 900m in elevation and also indicate net accumulation of 1030 - 1250 mm water equivalent Discharge begins to rise in

April at lower stations, and not until May in the upper catchment areas. Due to a greater glacierised area the Hunza river produces 30% more streamflow than the Gilgit River even though it has only 9% more catchment area (Table 2 3)

Table 2.3
Basin Characteristics of the Upper Indus Basin Rivers

No	Name	No of Years	Area ($km^2 x10^3$)	% Area Indus (Besham)	Mean Annual flow (m^3s^{-1})
1	Shyok (Yogu)	9	33 7	20 73	348 4
2	Indus (Kachura)	13	112 7	69 38	1011 0
3	Gilgit (Gilgit)	12	12 1	7 45	288 9
4	Hunza (Dianyor)	16	13 1	8 1	308 1
5	Gilgit (Alam)	15	26 1	16 11	666 2
6	Indus (Partab)	22	142 7	87 9	1730
7	Astore (Doyian)	6	4 0	2.5	121 0
8	Indus (Besham)	15	162 4	100 0	2351.5

No of years are taken from the data record given in Snow and Ice Hydrology Project Annul Report 1985

In terms of volume of water carried annually, the Indus ranks with the Columbia River in Canada and United States The annual flow of the Indus is about 209,691 6 million cubic meters, twice that of Nile, three times that of the Tigris and Euphrates combined (Gulhati 1968) Out of the total drainage area, 453,250 km^2 lie in the Himalayan mountains and foot-hills, which are the source of water supply, the rest lie in the arid plains of India and Pakistan which would mostly be desert

but for the waters of the Indus.

Runoff from snow melt depends upon the previous winter precipitation Snowmelt component is larger after an unusually snowy winter, but melting from the glaciers is delayed as snow insulates the ice surface for a longer period Stream flow also varies considerably from year to year For example, runoff from the entire Karakoram region (Upper Indus catchment plus Gilgit and Hunza rivers) was 390 mm in 1970 but increased to 540 mm in 1973 (Fig 2.5) This variation may sometimes depend less on winter snow fall than on weather conditions which largely control ablation rate in summer

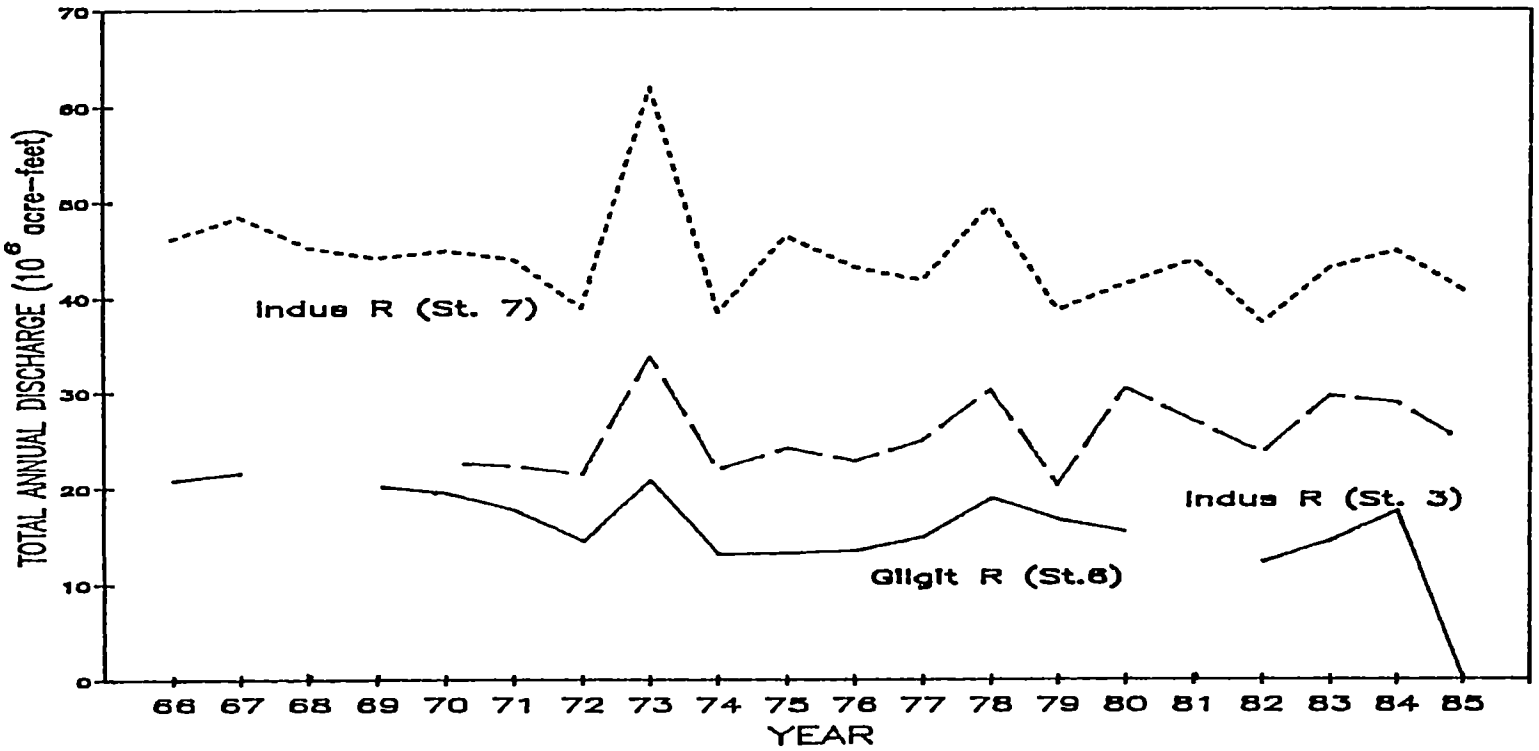
Snowfall during spring or prolonged cloud cover may interrupt the diurnal flow regime considerably by reducing glacier ablation Therefore, a sunny or comparatively hot summer yields high runoff at the expense of glacier storage Also seasonal variations in the Upper Indus Basin river's flow are affected by the creation of major natural dams and outburst floods as a consequence of either glacial, mudflow or landslide blocking and subsequent sudden outburst flooding (Hewitt 1982)

2.7 CONCLUSION

Climatic information of the Upper Indus Basin is scanty, particularly at high elevations. The mountains extend well into the troposphere and are influenced directly by the upper air system The effect of the upper air flow on the climate of the Karakoram is unclear The regional airstream continues to be influenced by the westerlies

FIG 2 5

UPPER INDUS BASIN
VARIATION IN ANNUAL DISCHARGE



VARIATION IN ANNUAL DISCHARGE OF THE UPPER INDUS BASIN

STATION•	MEAN ANNUAL DISCHARGE (1000 A-ft)	PERCENTAGE OF MEAN ANNUAL DISCHARGE 1966	67	68	69	70	71	72	73	74	75	76	77	78	79	80	81	82	83	84	85
3 Indus R at Katchura	25 210	—	—	—	—	88	86	82	131	85	94	91	99	120	81	121	107	94	118	115	98
6 Gilgit R at Alam Br	16 610	122	127	—	119	114	104	85	122	77	77	81	90	114	100	93	—	74	87	106	—
7 Indus R at Partab Br	43 160	104	110	102	100	101	100	88	140	87	105	100	97	115	90	96	102	87	100	104	94

• Station Number refers to Figure 2 2

and westerly depressions may continue to affect the region The number of these low pressure systems is extremely variable, resulting in seasonal periodic and aperiodic fluctuations in temperature and precipitation This variability is further increased when temporary destruction of the Tibetan anticyclone allows the incursion of monsoonal air masses into the Karakoram, resulting in heavy precipitation However, effect of monsoonal air masses is limited to the southeastern part of the Upper Indus Basin and further north, westerly depressions may continue to influence the climate during summer as well Some scientists have reported precipitation under both (westerly and monsoonal) air masses during summer in Ladakh, Himalayas (Mayewski et al 1983) The effect of these broad climatic regimes are further strongly modified by altitude and local topography

CHAPTER 3

Snow Accumulation And Ablation In Mountain Areas

3.1 INTRODUCTION

In the mountains, both the timing and volume of runoff from watersheds differ from that derived from lowland rain or snowmelt Typically, high-altitude terrain receives more precipitation than the surrounding lowlands, because of forced, or orographic, lifting of air masses as they cross the highlands (Barry 1981) The mountain hydrometeorological system can be examined at a variety of spatial and temporal scales that range over 10 orders of magnitude (Alford 1985, p 353) The smallest scale, the microscale, is distinguished from the other scales by the great importance of physical changes in air masses necessary to promote the formation of precipitation and by the point surface energy exchange processes. The micrometeorological problem is fairly well understood and is identical, at least in kind, with that of all other terrestrial environments (Anderson 1976) The factors interact in a complex way in areas of high local relief, such as the Karakoram Mountains, and water balance calculations cannot be based upon standard meteorological measurements in these areas (Alford 1985)

Snowmelt results from many different processes of heat transfer The quantity

of snowmelt is, moreover, dependent upon the condition of the snow pack itself As a consequence, the rigorous determination of snowmelt is quite complex and several assumptions are used in the practical computation of snowmelt The relative importance of the various heat transfer processes involved in the melting of snow packs vary with time and with locale As a result of this no single method or index for computing snowmelt has been found that is suitable to all areas and at all times of the year In order to select the best method of computing snowmelt for a given area a complete understanding of the snowmelt process is necessary In this chapter, the most important factors affecting snow accumulation, distribution and melting of snow packs in mountainous areas will be discussed

3.2 FACTORS AFFECTING SNOW COVER ACCUMULATION AND DISTRIBUTION

Snow cover comprises the net accumulation of snow It may include water from rainfall and various contaminants frozen in the cover as well as precipitation deposited as snowfall, ice pellets, and hoarfrost Its structure and accumulation patterns are complex and highly variable with space and time This variability results from a number of factors such as wind, temperature, and humidity at the time of precipitation and immediately following Radiative exchange alters snow structure, density, and optical properties such as reflectance etc Wind modifies snow density and crystalline structure, and causes scour and redeposit Topography and physiography also affect snow cover On steep slopes, mechanical creep and avalanches may greatly modify the snow cover All these factors contribute to accumulation, metamorphosis, ablation, and alter the snow cover's physical characteristics until it bears little resemblance to freshly fallen snow The effect of each factor is discussed below

3.2 1 Wind

The characteristics of the wind near the earth's surface are of major importance in determining the amount of snow movement and in determining the scour and depositional patterns Isolated peaks and exposed ridges of the Himalaya Mountains experience high average and extreme wind speeds as a result of limited frictional effect of the surrounding terrain on the motion of free air In some locations, terrain configuration may even increase wind speeds near the surface above those in the adjacent free air (Barry 1981, p 51) Snow accumulation patterns are a complex

function of deceleration and acceleration of the air stream, its velocity profile and the formation of separation vortices (Richter 1945, cited in Gray et al 1979) The presence of velocity gradients within the boundary layer implies the existence of shear stresses in the wind flow The shear stress is highest at the earth's surface and decreases with height, becoming zero in the geostrophic wind above the boundary layer (Kind 1981) It is the shear stress exerted by the wind on the surface which causes the movement of loose snow Virtually all natural surfaces act as rough surfaces with respect to wind. When the wind speed is high enough to induce drifting or movement of snow, the flow near the surface is dominated by the shear forces. Under such conditions Kind (1981), has mentioned the following relationship

$$\frac{U}{U^*} = 2\ 5 \ln \left(\frac{Z}{K}\right) + 5\ 5 - C(\lambda) \qquad (3\ 1)$$

where U = the mean wind speed at a height, Z,

U* = the "friction" or "shear" velocity, equal to $\sqrt{\tau o/\rho}$ in which τo is the shear stress at the surface and ρ is the density of air,

K = a roughness parameter, and

C (λ) = a constant whose magnitude depends on the non-dimensional spacing of the roughness elements, λ.

This velocity profile equation can apply to the surface over which roughness is fairly uniform and approximately valid for Z<fetch/20 or Z = 50 m, whichever is less (Gray et al 1979, p 10, Kind 1981, p 341) The equation is not applicable to

the profile near or below the tops of roughness elements, where the wind flow pattern is very complex and often three dimensional, for example where trees, buildings, ridges, and other obstacles affect the wind pattern

3.2 1 1 Transport of snow by wind

The force exerted by moving air per unit area is the shear stress, τ_o, existing between the moving air and the snow cover, which in turn, is a function of surface roughness, the mass density of air, and the wind velocity Before movement can occur, it is necessary that the shear stress attain some critical value to overcome the particle weight and inter-particle cohesive forces. In snow drifting, it is taken as the shear velocity, $U^* \equiv \sqrt{\tau o/\rho}$, instead of the shear stress The threshold shear velocity (U^{*th}) required to disturb the surface and transport particles is highly variable depending on the size, shape and weight of the snow crystals and the cohesive forces, the latter being dependent on the wetness of the snow The drifting process itself causes compaction and hardening of the snow and the threshold shear velocity increases progressively during drifting of snow by wind Oura (1967) reported an increase in the threshold velocity of freshly fallen snow from 0 22 m/s to 0 4 m/s after only several hours of aging

In the equation 3 1 shear stress, τ_o, and the shear velocity, U^* are related to the wind velocity, U Kind (1976) and Owen (1964) suggest that the following relationship can be used to describe the velocity profile over and within the saltation layer

$$\frac{U}{U^*} = 2\ 5\ \ln\left[\frac{Z}{U^{*2}/2g}\right] + 9\ 7 \qquad (3\ 2)$$

This equation is almost the same as equation 3 1 except they have replaced effective roughness height, K with $U^{*2}/2g$ Kind (1976) and Owen (1964) interpreted this as being proportional to the effective height of the rough surface

3.2 1 2 Modes of transport

The three most recognized modes of snow transport in the movement of snow are *saltation* - the bounding of particles along the surface travelling in curved trajectories under the influence of wind and gravity forces, *ground creep* - the sliding or rolling of particles along the surface, and *turbulent diffusion* - in which particles are held in suspension in the air stream without necessarily contacting the ground Generally, it is accepted that most of the snow is transported by saltation or turbulent diffusion or both (Kind 1976, Gray et al 1981) In effect, these two modes of snow transport have led to the development of two snowdrift theories, the dynamic and diffusion theories based respectively on the works of Bagnold (1941) and Schmidt (1925) (cited by Gray et al 1979) The dynamic theory views snow drifting as a near surface phenomenon due to small eddies in the lowest 10 cm producing mainly saltation, whereas the diffusion theory attaches the main importance to the larger eddies in the free air stream extending to tens or hundreds of meters above surface (Radok 1977)

Bagnold (1973) suggests that snow particles can only rise to great heights when the turbulent velocities become roughly equal to the terminal fall velocity of

particles and shear velocity, U^*, is about five times greater than the threshold velocity, U^*_{th} Generally, the major portion of the mass in the surface layer of the snow pack which can be blown by wind consists of particles having fairly similar size (nominal diameter) of about 0.5 mm, and the threshold shear velocities may range from 0 1 m/s to 0 2 m/s (Kind 1981, p 343) Owen (1964) has stated that for saltating, uniform, spherical particles, assuming no phase change and two dimensional flow, the mass transport rate or the ability of the wind to transport snow is approximately proportional to the 3rd power of shear velocity, U^{*3} The distribution of the horizontal mass flux of blowing snow with height is evident from figure 3 1 Approximately 90% of the total flux occurs within about 2 cm of the snow surface "saltation" (Kind 1981, p 348)

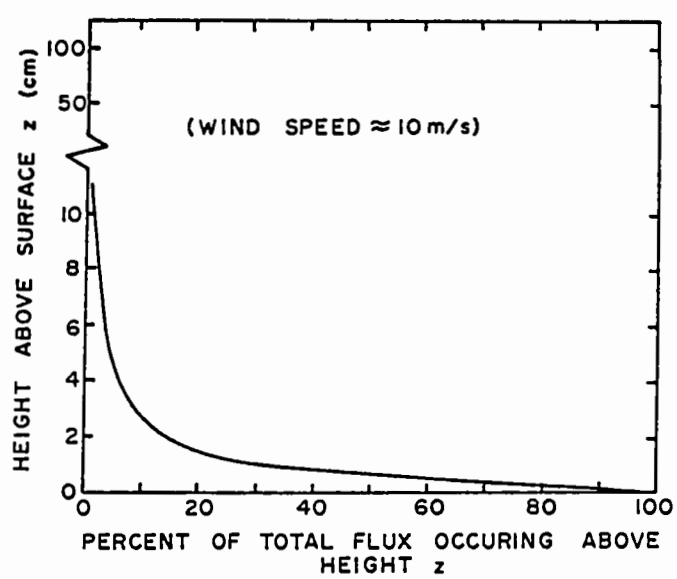


FIG 3 1 Approximate variation of horizontal mass flux of blowing snow with height (Kind 1981)

3.2.1.3 Effects of wind on snow density

In snow hydrology, the most critical factor is the effect of wind on snow density and subsequently on the snow water equivalent When snow crystals are moved by wind, their physical shape and properties are changed and they are redeposited with densities much greater than the parent materials. Gray et. al (1971) reported the density of newly fallen snow increased from 45 kg/m^3 to 230 kg/m^3 within a period of 24 hours under blizzard conditions For comparison, Goodison and Ferguson (1981) reported average increase in density of a freshly fallen snow at a sheltered site in Ontario from 104 kg/m^3 to 152 kg/m^3 within six and a half hours (1030 to 1700hrs) Table 3 1 shows the characteristic values of the densities of snow as influenced by wind

Table 3.1 - Densities of snow cover

Snow Type	Density (Kg/m^3)
Wild snow	10 - 30
Ordinary new snow immediately after falling in the still air	50 - 65
Settling snow	70 - 90
Very slightly toughened by wind immediately after falling	63 - 80
Average wind-toughened snow	280
Hard wind slab	350
New firn snow	400 - 550
Advanced firn snow	550 - 650
Thawing firn snow	600 - 700

Source Gray et al 1979, pp 14

3.2.2 Topographic Factors

The primary topographic factors affecting snow accumulation and distribution are elevation, slope and aspect. Of these three, in mountain terrain, elevation is normally considered the major factor At a specified location and within a given elevation interval, a linear association between snow accumulation and elevation is often found (U S Corps of Engineers, 1956) The transposability of these relationships from place to place is highly suspect, because the influence of elevation alone is indeterminate due to the interdependency of climate, slope, and elevation Each of these physiographic factors is discussed independently below

3.2.2.1 Elevation

The increase of precipitation with height on mountain slopes is a world-wide characteristic, although actual profiles of precipitation differ regionally and seasonally (Barry and Chorley 1985) Several studies demonstrate that the altitudinal increase is due to the combined effect of higher intensities and greater duration of precipitation (Atkinson and Smithson 1976, Hendrick et al 1979) Flohn (1974), stated that increase in precipitation along a mountain profile is dependent on the total vapor amount which condenses per unit of time during transport, and on the increase in wind velocity with height. Gray et al (1979) measured precipitation within selected elevation bands in Colorado for three consecutive years and found large variation between major physiographic areas and also spatial- temporal variations within a given area (Gray et al 1979, p 14)

Hendrick et al (1979) have made a comprehensive study of the spatial

distribution of precipitation with elevation at Mount Mansfield, Vermont between 327m and 1170m a.s.l During two winter-spring seasons (October to March 1976-1977 and 1977-1978), precipitation, snow depth, and snow water-equivalent observations over the mountain profile were analysed for elevation effects on weekly snow and water input, on event and hourly precipitation amounts and intensities The study shows that,

1 Total precipitation increases linearly with elevation, but individual events makes both seasonal averages and linear-regression inadequate for predicting higher elevation precipitation on an event basis

2 Hourly analysis of precipitation intensities showed that increased intensity at higher elevations during periods of concurrent precipitation accounted for 74% of the total increase of precipitation depth with elevation

3 Finally they suggested that wind and moisture observations are necessary to estimate snow accumulation with elevation on an event basis

Peck (1964) and Dingman et al. (1979) emphasized the influence of climatic factors or elements of parent weather systems in interpolating snow distribution and accumulation patterns. Dingman et al (1979) have analysed snow course data for 93 stations in New Hampshire and Vermount for the years 1964 - 1973 to determine distribution of snow depth, density and water equivalent with elevation Table (3 2) summarizes the results of the principal set of regression analysis. There is a significant (α = 05) relationship for water equivalent, depth with elevation for all three months (January, February and March) This is largely due to two elevation related climatic factors First, more precipitation occurs at higher elevations due to higher intensities of precipitation (Hendrick et al 1979) The second factor is

Table 3 2 – Results of Regressions of Snow Properties vs Elevation

	n*	r	r^2	a (cm)	b (cm/m)	Std error (cm)	95% C I (cm)
Water equivalent							
January	91 (11)	639	408	5 62	0107	1 93	80 - 2 29
February	92 (13)	696	484	8 23	0165	2 50	1 03 - 3 06
March	90 (65)	744	533	9 08	0224	3 03	1 27 - 3 62
Depth							
January	91 (11)	637	406	27 7	0409	7 40	3 08 - 8 83
February	92 (13)	675	456	34 1	0584	9 31	3 83 -11 33
March	90 (65)	721	519	31 1	0769	11 2	4 67 -13 32
Density					b (1/m)		
January	91 (11)	345	119	190	00006	0234	
February	92 (13)	115	013	251	00002	0196	
March	90 (65)	-029	0009	300	-000004	0223	

* Numbers in parentheses are numbers of stations for which only 9 years of data were available
(Source Dingman 1979)

a function of the vertical temperature gradient which minimizes the loss of water from melting

In Karakoram, orographic effects dominate the form and spatial distribution of precipitation and rapid increase is reported by many expeditions above 3000m a.s.l (Hewitt 1968, Batura Glacier Group 1979 and Wake 1987) The zone of maximum accumulation of snow has been observed between 4800 to 5200 m a.s.l and decrease above that up to 5800m a.s.l (Hewitt 1985, Wake 1987)

3.2.2.2 Slope

Mathematically, the orographic precipitation rate is predominately related to terrain slope and windflow rather than elevation If the air is saturated, the rate at which precipitation is produced is directly proportional to the rate of rise of the air The rate of rise of the air flowing over a upsloping terrain is directly proportional to the product of the wind speed and the magnitude of the slope (Gray et al. 1979) Gray et. al also suggests that winds of high velocities and long duration of snowfall are most important among the factors effecting distribution at higher elevations. That may well be a major factor in the Karakoram Mountains, where high winds and reduced cloud cover are observed at high elevations (Hewitt 1968)

Steppuhn (1978) shows the relative amounts of snow retained by level plains, gradual slopes and hill tops (all in summer fallow) to be of the order of 0 6, 0 7 and 0 2 respectively Woo and Marsh (1977) suggest, after a detailed snow survey at the end of the 1975-76 winter in four small High Arctic basins, that various types of terrain within the basins possessed distinctive characteristic snow depths and densities This suggests that division of terrain units with different topographic

features is necessary in a study of snow storage

3.2.2.3 Aspect

The importance of aspect on snow accumulation is evidenced by the wide differences found between the snow cover on windward and leeward slopes in mountainous regions In mountains, the major factors contributing to these differences are the interaction and influence of aspect in relation to the directional flow of snowfall producing air masses, the frequency of snowfall occurrences, and their effect on the energy exchange processes influencing snowmelt and ablation (Gray et al. 1979)

Meiman (1970) suggests that the effect of aspect appears to be predominantly a melt effect rather than an accumulation effect Goodell (1952) and Stanton (1966) have similar views that aspect does not affect the maximum snow cover in natural forest conditions where the melt opportunity is minimized, while the effects increased in areas where winter melt is common Similar arguments are summarized by Landals and Gill (1973) Stanton (1966) show the importance of aspect along the eastern slopes of the Rocky Mountains (Table 3 3)

Table 3.3 Mean accumulation of snow (cm) under different cover conditions in the Eastern Rockies as related to aspect (after Stanton, 1966)

	Aspect		
	N	S	E
Forest	41	41	39
Cut forest	45	53	65

(Source Gray et al (1979 p 16)

In the Himalaya, on regional scale, snowline on southern slopes is lower by 700-1500 meters than the cool northern slopes. While on a mesoscale the snow line is usually 1000-1500 meters higher on southern slopes during the melting season due to extended energy exchange as compared to relatively cool north facing slopes. This is also evident from many hanging glaciers on northern slopes in valleys of the Karakoram Mountains.

3.3 SNOWMELT IN THE MOUNTAIN ENVIRONMENT

3.3.1 Sources of Heat Energy

The principle fluxes of heat energy involved in the melting of the snow pack were discussed by Male and Gray (1981) Practically all of the heat utilized in the melting of snow can be ascribed ultimately to solar radiation (Fig 3 2)

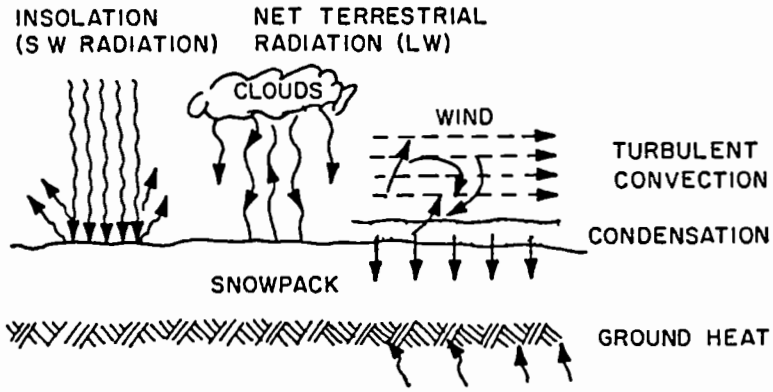


Fig 3 2) Sources of heat that generate snowmelt (Source Davar 1970)

Solar radiation may supply heat for snowmelt in several ways The principal ways are a) by direct incidence upon the snow, b) by reflected radiation resulting from incidence of solar radiation upon objects with or without conversion from short wave to long-wave heat, and c) indirectly as warm air, the temperature of which has been raised either by direct solar radiation or by contact with heated bare rocks, ground or trees or by the conversion of short wave radiation to long-wave heat Another source of heat available for snowmelt is the latent heat of condensation and also the heat produced by refreezing of melt water in the snowpack In the following paragraph each of the factors affecting snowmelt in mountainous areas will be discussed individually and also the effects of mountainous topography on each of the individual components will be described

3.3.2 Thermal Quality of Snow-pack

The amount of snowmelt resulting from a given quantity of heat energy is dependent upon the thermal quality of the snow pack The latent heat of ice is a well established quantity 80 cal/g (U S Army 1956) Only rarely is snow encountered which consists of pure ice at 0°C During the melt season the snow pack is generally isothermal at 0°C and also contains varying amounts of free water The actual condition of the snow pack with regard to the volume of water resulting from a given quantity of heat energy is designated as the "thermal quality" of the snow pack At sub-freezing temperatures snow will have a thermal quality greater than 100 percent as more heat energy is required to bring the snowpack at 0°C The thermal quality of snow is defined as the ratio of heat necessary to produce a given amount of water from snow to the amount of heat

required to produce the same quantity of melt from pure ice at 0°C The U S Army (1956) has determined by using calorimetric methods, that the thermal quality of snow ranged from 80 - 110 percent. Generally snow has low thermal quality values during times of high melt, when samples of snow contained melt water in transit or in excess of the liquid-water-holding capacity of the snow

3.3.3 The Radiation Balance at the Snow Surface

Radiation is most often the important factor in heat exchange at the snow surface Obled and Harder (1979), describe the following equation for a given unit area

$$N = I + D + C - T - R \tag{3.3}$$

where N is the net radiation at the snow surface, I is the direct solar radiation reaching the snow which has not been attenuated in the atmosphere by reflection, scattering or absorption, D is the diffuse radiation which has been scattered by the atmosphere or radiation reflected from the surface which has been backscattered by the atmosphere, C is the downward long wave atmospheric radiation, or counter radiation, emitted by the earth's atmosphere, T represents the terrestrial radiation emitted at the earth's surface, and R is the portion of incoming radiation which is reflected at the low surface

The direct and diffuse radiation which are often considered together and referred as global radiation, incident only during the daytime in wave lengths from 0 3 to 3 0 μm The effective radiation, which is the difference between the terrestrial

radiation and downward atmospheric radiation and defined as positive in outgoing direction because the terrestrial radiation is usually the larger of the two, falls in long wavelengths from 3 0 to 80 μm. The reflected component R of the net radiation balance depends upon the albedo of the surface For long wave radiation snow works as a blackbody, and it is generally accepted that less than 0.5% is reflected (Geiger 1965) Variation in long wave radiation due to topographical variation is also considered negligible with respect to the net radiation balance (Obled and Harder 1979)

3.3.3.1 Direct solar radiation

In the study of the effect of topography on the radiation balance it is the direct solar radiation component which has been widely studied (U S Corps of Engineers 1956, Dozier 1980, Barry 1974 & 1978 , Obled and Harder 1979, Ohmura 1970; Garnier and Ohmura 1970, and Williams et al 1972) The radiation received on slopes and their variations with altitude will be discussed in the following paragraphs.

3.3.3.2 Direct radiation received on slopes

In the northern hemisphere, the radiation incident on south facing slopes exceeds that on the north facing slopes At any given instant, the radiation on a sloping surface depends upon the geometry including the slope angle and it's aspect in conjunction with the solar altitude and azimuth In addition, there is the variability in the transmission of radiation with solar altitude due to differences in the optical air mass through which radiation must pass (U S Corps of Engineers 1956)

In figure 3 3, if we look from the direction normal to the plane formed by the direct rays of the sun and the normal to a horizontal surface of unit area (slope 2) we can find another surface (slope 1) having a unit area perpendicular to the rays of the sun

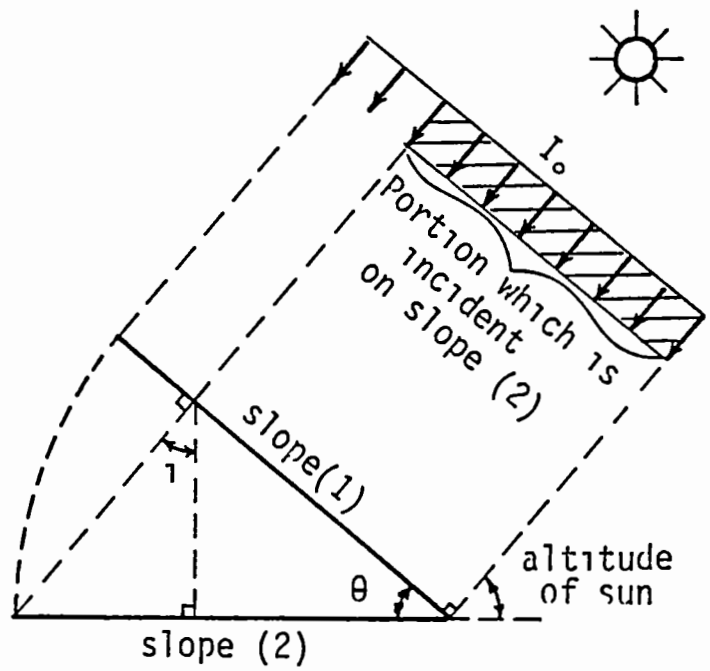


FIG 3.3 Direct radiation on slopes (Obled and Harder 1979, p 182)

In this case if the intensity of the sun's rays reaching the earth's surface is taken as I_o the slope 1 will receive the same radiation flux as I_o Here surface area of slope 2 is equal to slope 1, but the area seen by the sun's rays will be reduced by

a factor $\cos\theta$ where θ is the angle between the perpendicular surface and surface 2 (or any surface on which the intensity of radiation is required) Therefore, the radiation flux on the slope is also reduced by $\cos\theta$ For example if the sun is at 50°, then $\theta = 40°$ and the radiation received will be $I_o \cos 40°$ or $0.77\ I_o$ Similarly if a the sun is at 50° and a slope surface is tilted 30° from the horizontal towards the sun, the angle θ will be 10° and the radiation flux received will be $0.98\ I_o$ or 127% of the flux on the horizontal surface

From figure 3.3 it can be found that angle θ is also equal to angle i, the angle between the normal to the slope and the direction of the sun's rays. Therefore it can be deduced that the solar radiation flux on any slope will be equal to

$$Is = I_o \cos i \qquad (3.4)$$

This information and the value of direct radiation flux on a horizontal surface, equation 3.4 can be used to calculate the direct radiation on any slope if the value of direct radiation either measured or estimated is available (Kondratyev 1969)

3.3.3.3 Variations of direct radiation with altitude

Various field studies have shown that for clear sky conditions the direct solar radiation on a perpendicular surface increases between 0.05 and 0.20 cal/cm^2 min per 100 m of rise in altitude (Kuzmin 1961 cited by Obled and Harder 1979) Barry (1981) mentions that there is rapid increase up to 2000 m a.s.l, after which the rate of increase declines A study by Sauberer and Dirmhirn (1958) found 32% greater radiation at 3000m a.s.l than at 200 m in December, 25% greater radiation receipt in

March and September and 22% greater radiation receipt in June This increase is referred to as a decrease in optical air mass with the decreasing thickness of atmosphere, and occurs because of changes in atmospheric transmissivity due to the decreasing quantities of moisture and dust pollution with elevation (Barry 1981, Obled and Harder 1979, and Male and Gray 1981) The optical air mass (m) is corrected for elevation by Barry (1981) using the ratio of local pressure to normal sea level pressure, and is expressed in terms of optical air mass,

$$m = \frac{1}{\sin \theta} \quad \text{where } \theta = \text{solar altitude} \qquad (3\ 5)$$

where at sea level, the relationship between optical air mass and solar altitude will be for $m = 1$, $\theta = 90°$, $m = 2$, $\theta = 30°$ For comparative radiation calculation at different altitudes, the absolute optical mass (M),

$$M = m(p\ /\ po), \text{ where } p = \text{station pressure and } po = 1000 \text{ mb}$$

is used to allow for the effects of air density on transmission

William et al (1972) has estimated the optical air mass values for different elevations numerically by multiplying air mass with $(1-h/h_t)$ where h is the local elevation and h_t the height of the troposphere which they have taken as 10 Km Therefore, for a change in elevation of 1000 m, the optical air mass will be reduced by 10% With a transmissivity of p = 0 6 and an uncorrected m = 2, the correction will alter p^m and thus the radiation flux by $(0\ 6^{9m} - 0\ 6^{m})/P^{m}$ or about 10%

3.3.4 Diffuse Radiation

Obled and Harder (1979), stated that long term seasonal radiation totals measured at eight stations between latitudes of 40° and 60° show that diffuse radiation accounts for between 36% and 51% of global radiation in spring and between 24% and 41% in summer These figures show the importance of diffuse radiation and it's variation with topography Diffuse radiation affecting snowmelt comes from three sources (see Fig 3 4)

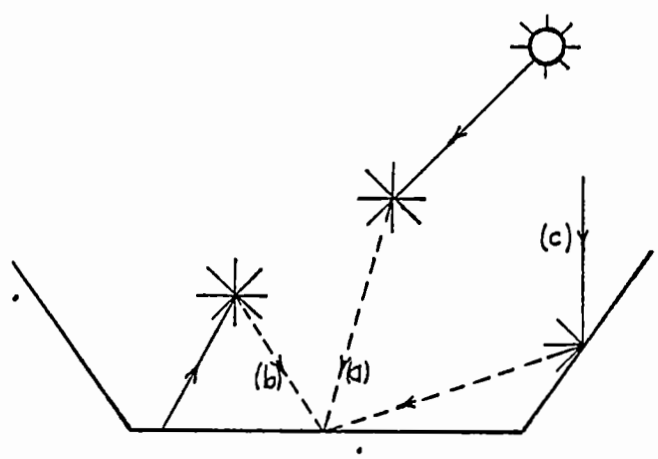


FIG 3 4 Sources of diffuse radiation

(a) Scattered incoming solar radiation
(b) backscattered radiation
(c) reflected global radiation

(a) incoming solar radiation which has been scattered in the Earth's atmosphere

(b) the portion of global radiation reflected at the Earth's surface which has been redirected downward by subsequent scattering and reflection in the atmosphere This is usually referred to as backscattered radiation.

(c) global radiation which has been reflected for other surfaces of different slopes or orientation

3.3.4 1 Diffuse radiation over snow surface

The most important source of diffuse radiation to the snow surface is the backscattered short wave radiation, i e., the portion of the incoming radiation reflected at the snow surface which is again scattered and reflected downward by the atmosphere A comprehensive backscattering model is presented by Dozier (1979) The critical parameter, in the model calculation he formulated, is the reflectance or albedo of snow surface

The difference in albedo is less marked for clear sky conditions because more reflected radiation escapes through the atmosphere without being backscattered Under cloudy conditions the flux of diffuse radiation depends on the sun's elevation angle and the amount and type of clouds Clouds reduce direct radiation but increase diffuse radiation by reducing reflected energy lost to space (Male and Gray 1981) It is reported that for cloudy skies and a snow covered surface the incoming global radiation can be twice as much as for a surface with an albedo of zero, because of successive reflections and backscattering between the snow surface and the clouds (de Brichambaut cited by Obled and Harder 1979)

The diffuse radiation which is received by reflection from surrounding surfaces of various slopes and aspects is unique to regions of topographical variations. The variations due to topographical features is attributed to the following five conditions.

(a) decreased sky dome because of surrounding topography

(b) the quantity of diffuse radiation received on a sloping surface

(c) reflected radiation from surrounding surfaces

(d) different amounts of radiation received on surfaces because of the anisotropic distribution of diffuse radiation

(e) the effect of altitude

3.3.4.2 Variation of diffuse radiation with altitude

Under clear sky conditions the quantity of diffuse radiation tends to decrease with altitude This is because of the decreasing optical air mass and atmospheric turbidity with altitude, which causes a decrease in the scattering of incoming direct radiation

3.3.5 Reflected Radiation or Albedo

For long wave radiation the snow surface acts as an almost ideal blackbody but for radiation of short wavelengths the reflectance of snow is much higher and

decreases rapidly with increasing wavelengths (O'Brien and Munis 1975) The integrated reflectance spectrum is called the albedo of the snow and can be expressed as the percentage of incident global radiation which is reflected by snow Factors affecting the albedo of snow include grain form, solid impurities, water content and surface irregularities (U S Corps of Engineers 1956, Male and Gray 1981) Figure 3.5 shows the degree to which albedo can vary for different types of snow

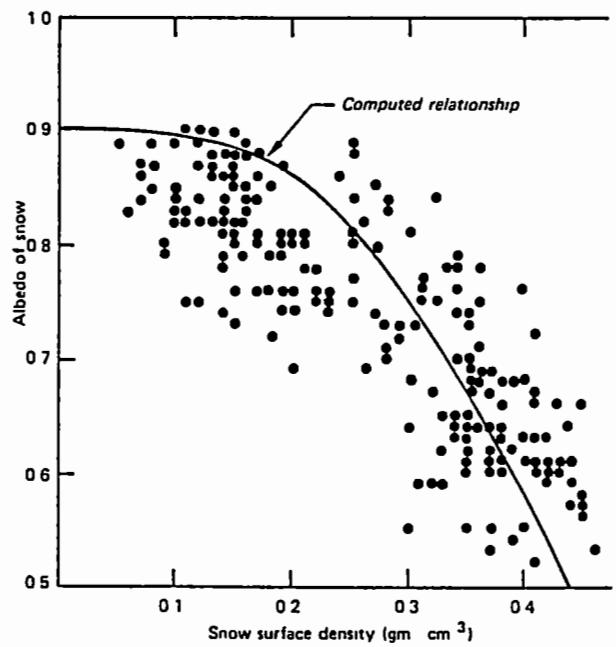


FIG 3.5 Albedo versus snow surface density (Anderson 1976)

These factors can effect the snow albedo on a diurnal and seasonal basis During melt periods a crust forms at night due to refreezing of melt water, by mid-day the

liquid water content at the surface often increases due to melted snow Changes in amount of reflected radiation are mainly the result of two phenomena.

(a) difference in the properties of the snow surface

(b) changes in the angle of incidence of direct solar radiation on slopes.

3.3.5 1 Changes in snow surface properties

Snowfall patterns, snowmelt and the thermal balance of the snow surface vary over regions of topographical variations. For this reason, the reflected term in the net radiation balance is one of the hardest to estimate from point measurements. The most important and most regular variation in albedo is the variation with elevation (Obled and Harder 1979) Melting snow in valleys has much less albedo than untransformed or fresh snow in the higher mountain slopes Kuzmin (1961) found that for neve basins in the Central Asian mountains the increase in albedo with elevation was 0 27% per 100 meters at the beginning of winter increasing to 1 24% at the end of the snowmelt season It is suggested that albedo can be calculated from the density measurements if sufficient measurements are available for this purpose (Obled and Harder 1979)

3.3.6 Effective Long-wave Radiation

The effective radiation totals describe a net loss in energy at the surface The intensity of the effective radiation is small during the day However the total quantity over a 24 hour period is highly significant (Fig 3 6) and at night the

radiation balance at the surface is determined by effective radiation alone (Obled & Harder 1979)

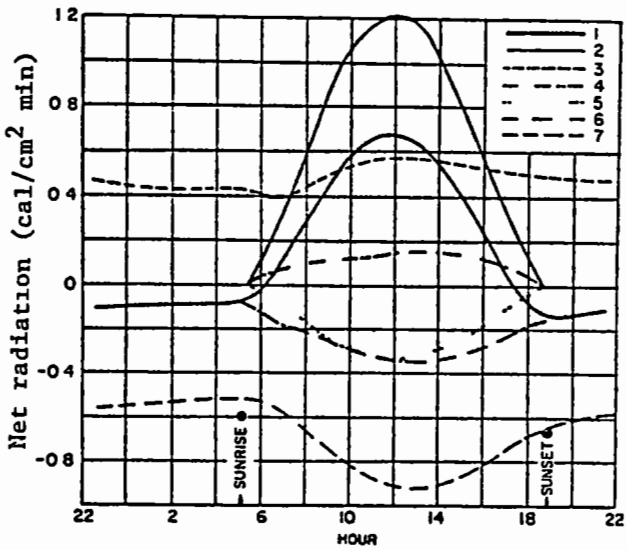


FIG 36 Diurnal variation of net radiation components (Kondratyev 1969)

(1) Net radiation *(2) direct solar radiation*
(3) downward atmospheric rad *(4) diffuse rad*
(5) reflected shortwave rad *(6) effective rad*
(7) terrestrial rad

Under clear skies downward atmospheric radiation is emitted by water vapor, carbon dioxide and ozone in proportion to the quantities present in the atmosphere and their temperature Geiger (1965), stated that atmospheric radiation from ozone represents only 2% of the total, and carbon dioxide which is almost constant in the

atmosphere, contributes about 17% of the total atmospheric radiation The remainder, more than 80% comes from water vapor This suggests that variation in atmospheric radiation depends largely upon the quantity of water vapor present in the atmosphere and the temperature of this water vapour Under cloudy conditions radiation is also emitted by the droplets forming the cloud base and the quantity depends on the type and height of the clouds as well as the temperature at the cloud base (Obled & Harder 1979)

Atmospheric radiation comes from all layers of the atmosphere, but the largest portion reaching the earth's surface is attributed to the lowest 100 meters (Geiger 1965) This phenomenon leads to the calculation of atmospheric radiation from temperature and vapor pressure measured at normal meteorological stations Marks (1979) developed an empirical formula for use in alpine areas under the assumptions that the relative humidity is constant with height and the temperature change with height is equal to the standard lapse rate, (Marks 1979, p 167) All these formulae are based on the assumption that the earth's surface radiates as a blackbody, and terrestrial radiation can be calculated from the Stefan-Boltzman law to be σT_s where σ is the Stefan- Boltzman constant and T_s is the temperature of the earth's surface Snow has been shown to reflect at most 0.5% of incident long wave radiation and therefore essentially behaves as an ideal black body (Geiger 1965)

The low thermal conductivity of snow helps in dropping the snow surface temperature rapidly which in turn drops thermal radiation This phenomena of low thermal conductivity of snow therefore has a regulating effect on terrestrial radiation (Obled & Harder 1979)

3.3.7 Turbulent Energy Transfer

Of secondary importance to radiation in the transmission of heat to the snowpack is the process of turbulent exchange in the overlaying air According to Kuzmin (1961) this heat exchange with the atmosphere accounts for approximately 40-50% of snowmelt at altitudes of around 3000 m a.s.l and at lower altitudes for up to 60-70%

Turbulent exchange between the snow and the atmosphere is driven by air temperature and vapour pressure gradients and by turbulence due to wind in the lower atmosphere With a downward temperature gradient there is a direct transfer of heat from the air to the snow, and with a downward vapour pressure gradient there is a direct transfer of moisture from the air onto the snow surface, releasing, in addition, its latent heat of vaporization The reverse processes occur as well In the absence of wind, the heat exchange between snow and the atmosphere is relatively low and increases considerably with increased wind speed Topography can affect the direction and the speed of wind at the surface either by altering the flow of large scale airmass movements or through the creation of local winds of thermal origin In the northern hemisphere the result of these modifications is that the horizontal component of the wind is accelerated and deflected to the left towards the region of lower atmospheric pressures on the windward slope and decelerated and deflected towards the right on the leeward slope (Fig 3 7) These modifications in wind flow have great importance in snowmelting For example, the chinook results in a strong, warm and dry wind often produces rapid snowmelt on the leeside of mountain ridges (Obled and Harder 1979) This is a consequence of the

airmass losing moisture while rising on the windward side and then heating by compression during the descent down the leeside

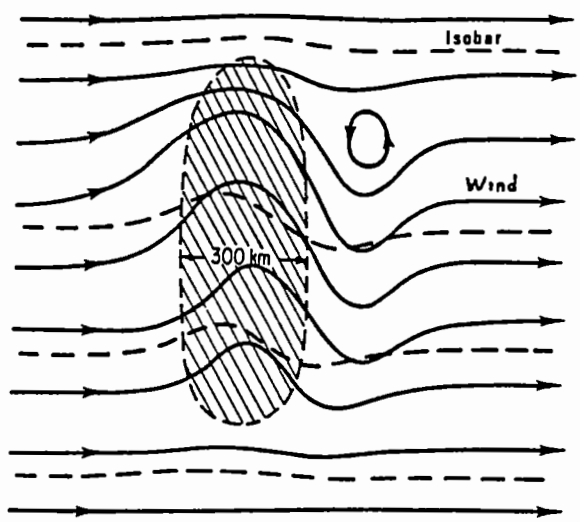


FIG 37 Deflection of large air masses over mountain barriers in the northern hemisphere

3.3.7 1 Air Temperature

The computation of turbulent exchange requires air temperature observations In the free atmosphere the decrease of air temperature with altitude assuming adiabatic conditions, is approximately 0 6°C per 100 m for unsaturated air For air saturated with water vapor the wet adiabatic lapse rate depends on temperature and the atmospheric pressure Above 20°C, the saturated adiabatic lapse rate (SALR) is reported less than o 5°C per 100 m whereas at sub-zero temperatures, the moisture content is reduced to the level where very limited amount of latent heat can be

released through condensation At -40°C, the SALR is almost identical to the unsaturated or dry adiabatic lapse rate (Barry 1981) Kuzmin (1961) has stated that with the change in air temperature from -20 to +20°C the value of the saturated lapse rate will change at 100-mb level from 0 87 to 0 43°C per 100 m, and at the 500-mb level from 0 78 to 0 33°C per 100 m

By comparison with the surrounding atmosphere, the slope air over mountains is affected by radiative and turbulent heat exchanges These processes modify the temperature structure over the massif and result in adiabatic lapse rates which differ from environmental lapse rate according to the time of the day (Barry 1981) Figure (3 8) shows the high variation in mean daily environmental lapse rate at different elevations within Zaravshan River Valley

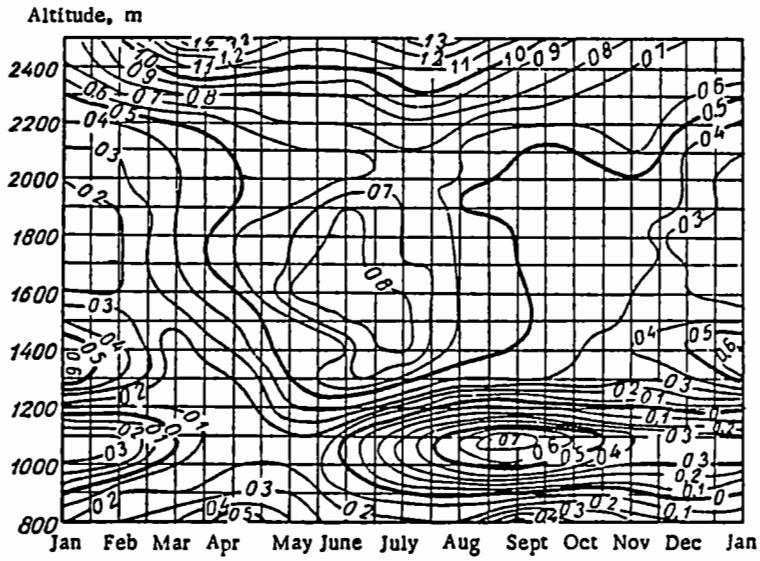


FIG 3 8 Isolines of the mean monthly lapse rates of daily temperatures in the Zaravshan River Valleys (Obled and Harder 1979)

3.4 AVALANCHES

The last important effect of the mountain environment on snow distribution and melt process is the frequent occurrence of avalanches which can cause the accumulation of large masses of snow in the valley bottom This process is important for two reasons. First, most of the avalanched snow concentrates in the gullies and has smaller surface area than the mountain snowpack but it has the same water content. Such accumulations are often exposed to less solar radiation than they would have been had the snow remained in the mountains. Second, avalanched snow increases the water yield of the basin by transfering the snow to the ablation areas, which otherwise may have remained frozen due to freezing temperatures on higher elevations without contributing to the basin yield

Due to steep terrain and great relief, snow avalanches are common phenomena in the Karakoram Mountains. These play an important role in snow hydrology by delaying the melting due to their higher density It is particularly true in the catchment of Kunhar River Northern Pakistan where avalanched snow supplies fresh water later in spring for agriculture and domestic use (de Scally 1986)

3.5 CONCLUSION

Mountains have strong influence upon the movement of air masses and the spatial distribution of precipitation In general, on a regional scale the variability is primarily dependent upon the altitude and geometry of the mountain ranges On a local scale it is due mainly to differences in elevation, slope, and aspect. However, this variability cannot be described solely in terms of macro and mesoscale process

since rugged topography creates countless topoclimates which differ widely with each other in their response to slope and aspect (Geiger 1965, p 455)

Slope angle and aspect are the key determinants of topo and micro climates. Differences in radiation receipt which result from topographic factors are reflected in air temperature, snow cover duration and consequently in the distribution of runoff Relief creates an unequal snow distribution which in turn causes an areal variation in the volume of spring melt This is particularly true in Karakoram where precipitation increases many times with increasing elevation and the melt pattern is controlled by the upward migration of melting temperatures with the progress of summer season

CHAPTER 4

Field Observation Program and Procedure

4 1 INTRODUCTION

The snow and ice Hydrology Project has undertaken research investigations in the Upper Indus Basin since summer 1985 During the course of the Project work, I visited Barpu Bualtar Basin in the summer of 1986 This basin was chosen for intensive scientific studies (both for research and monitoring purposes) during the forthcoming years and for other reasons explained by Young in S I H.P Annual Report 1986 The following are some main features which make Barpu/Bualtar Basin a suitable location for scientific studies.

a) Geographic location it lies on the southern flank of the Central Karakoram at the heart of the main source area of glacier meltwaters for the Upper Indus Basin

b) Available relief from the terminus of the Bualtar Glacier (2439m a.s l) to the maximum elevation of the Barpu Basin (7460m a.s.l) The 5021m of relief provides a range of elevations similar to other glacier basins in this part of the Karakoram including the zones of major snowfall and ablation

c) Glacier characteristics the basin contains a variety of glaciers having different characteristics For example Sumiayar Bar Glacier is a mainly avalanche fed glacier with an extensive debri covered area in the ablation

zone In contrast Miar Glacier is a white ice glacier being fed mainly by direct snowfall and Bualtar Glacier is an intermediate glacier fed both by direct and avalanched snow These different features represent the range of morphological and nourishment conditions of the Karakoram glaciers

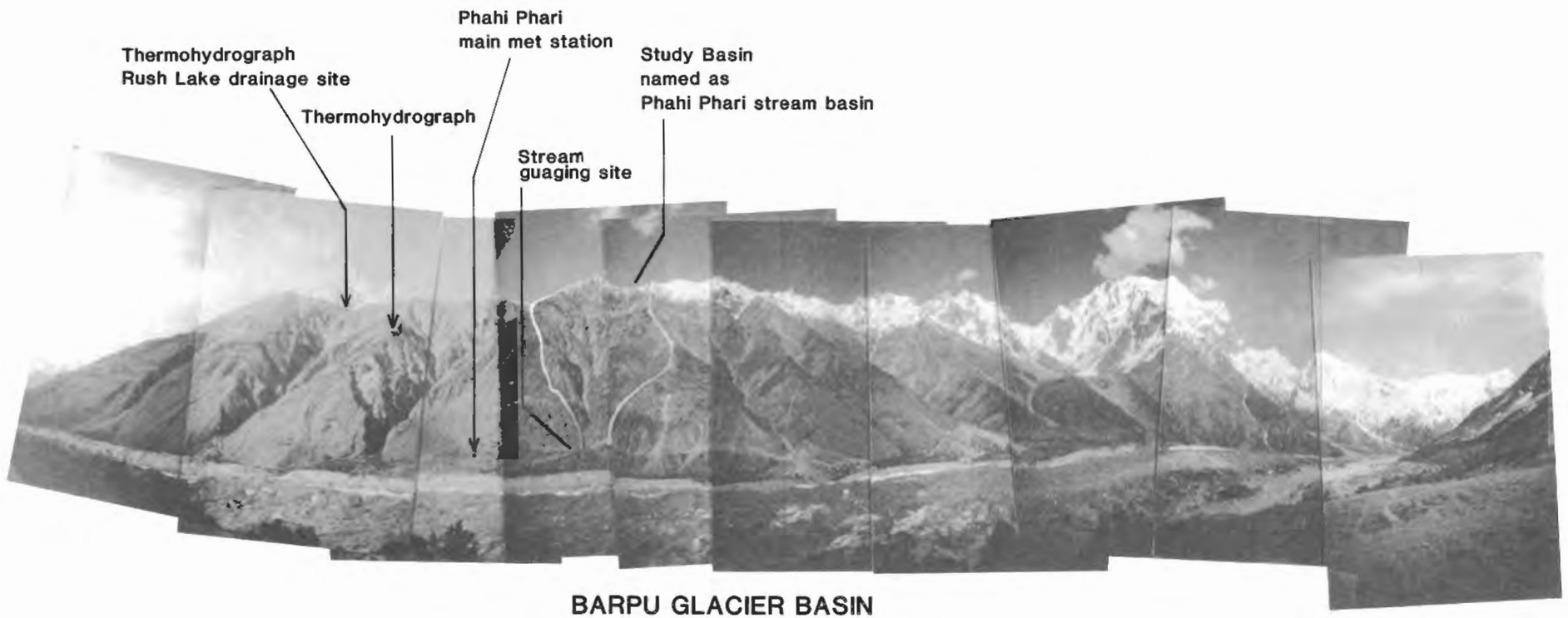
d) Most important for an effective research programme is the relatively easy accessibility of this basin

For the same reasons, the Barpu/Bualtar basin proved suitable as the site for this thesis research. During the 1987 summer field season, meteorological and streamflow data in the study basins were collected from mid May to mid September

4.2 STUDY AREA. The Barpu Glacier Basin Central Karakoram

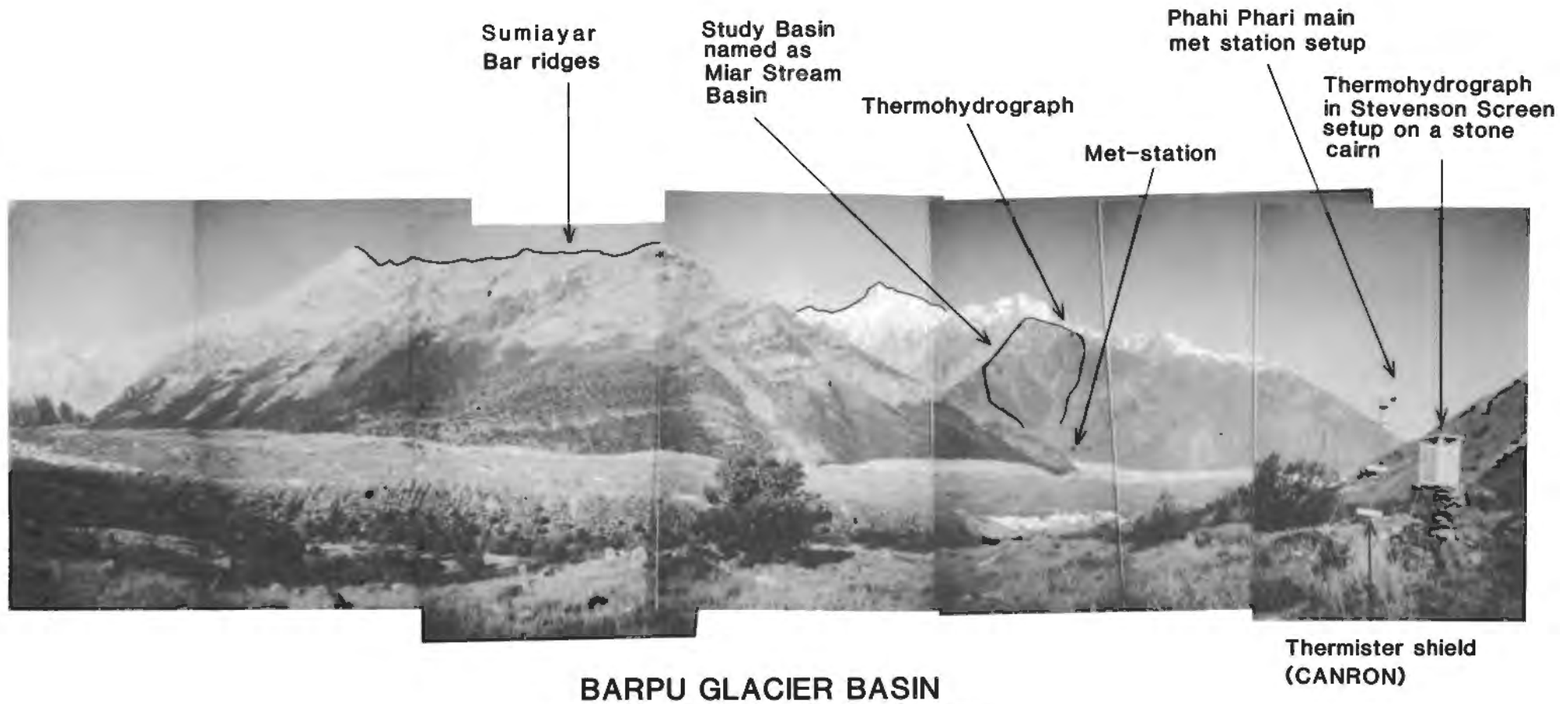
The basin is located by the coordinates longitude 74°47' to 74°49' and latitude 36°13' to 36°01' in the Central Karakoram The general features of the Barpu Glacier Basin are shown in figure 4 1 (a,b) Topographic features and location of measuring stations are shown in figure 4 2

The Barpu Glacier of the Barpu/Bualtar system descends northeast from the main crest of the Central Karakoram The main stream of this glacier is fed by two glaciers (Sumiayar Bar and Miar) which descend northwest and north, respectively, before combining to form the Barpu (Fig 4 2) Much of the Sumiayar Bar tributary is covered with debris, whereas the Miar glacier is a "white ice" glacier The glacier itself is a 29 km long valley glacier with an average slope of 1 6 72 from the main crest to its terminus. The elevational range is 4625 m The minimum elevation found at the glacier terminus is 2835 m a.s l The Basin covers an area of 415 Km^2,



BARPU GLACIER BASIN
Central Karakoram Mountains

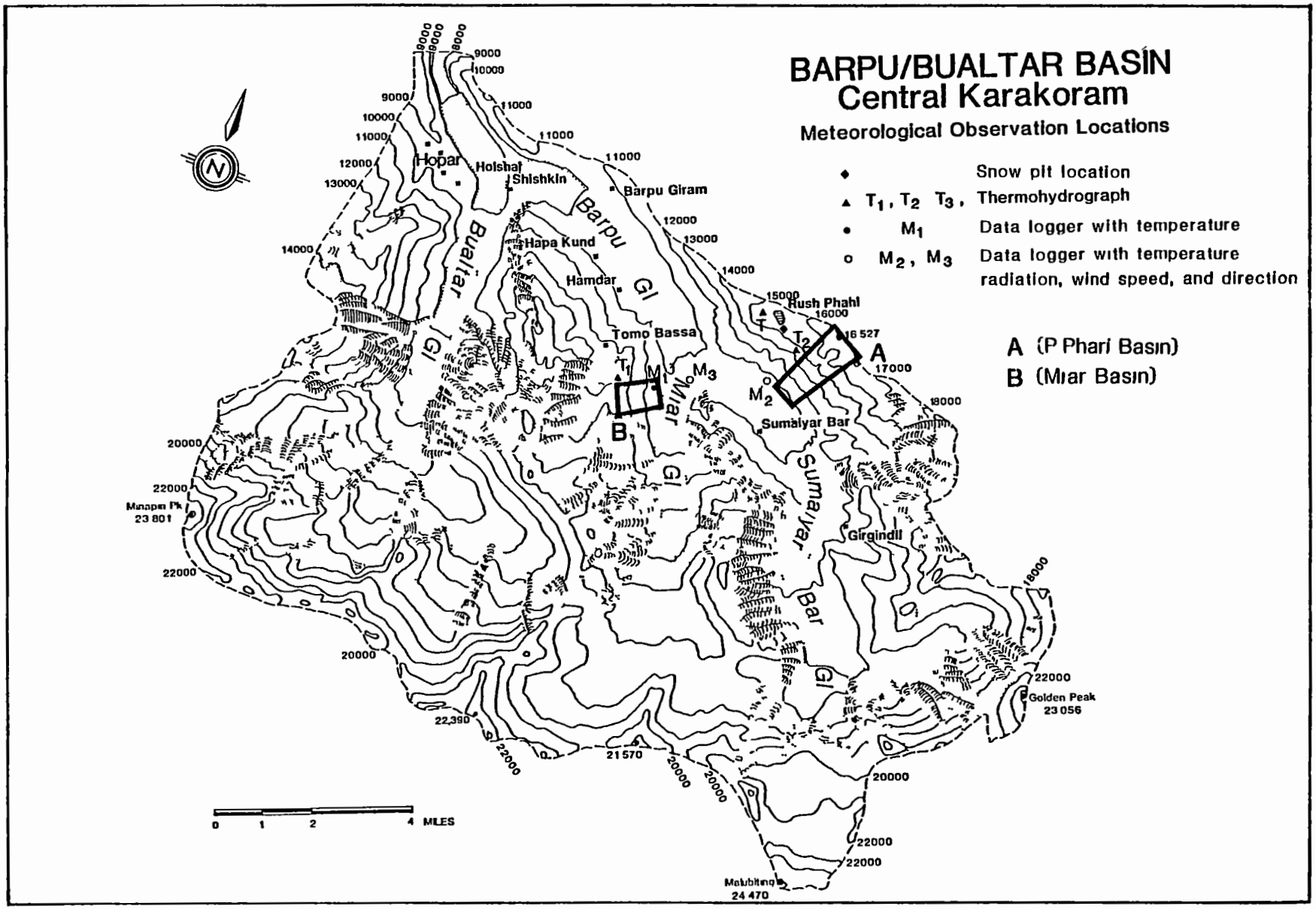
Phahi Phari – Yengutz Har Peak Ridge facing South–west,
July 6, 1987



BARPU GLACIER BASIN
Central Karakoram Mountains

Sumaiyar Bar and Miar Ridges facing North-east,
July 5, 1987

FIGURE 4 2


BARPU/BUALTAR BASIN
Central Karakoram
Meteorological Observation Locations
Snow pit location
T1, T2 T3, Thermohydrograph
M1 Data logger with temperature
M2, M3 Data logger with temperature radiation, wind speed, and direction
A (P Phari Basin)
B (Miar Basin)
Hopar
Holshal
Shishkin
Barpu Giram
Hapa Kund
Hamdar
Tomo Bassa
Rush Phahi
Sumaiyar Bar
Girgindil
Golden Peak
23 056
Minapin Pk
23 801
Malubiting
24 470
Barpu Gl
Bualtar Gl
Miar Gl
Sumaiyar Bar Gl
16 527
21 570
22,390
0 1 2 4 MILES

of which 125.56 Km^2 is glaciated The rest of the area remains covered with seasonal snowpack during winter (January through March) and gives most of the melt water during early spring

The Bualtar Glacier of the Barpu/Bualtar system descends northward from the main crest of the Central Karakoram to its terminus (Fig 4 2) It is 20.5 km long with an average width of 1 km. The main stream of this glacier passes very near to the snout of Barpu Glacier and the terminus lies approximately 3.5 km past the Barpu Glacier terminus (figure 4.2) The elevation difference between the snouts of these glaciers is 400 m According to the figures provided in the Pakistan Glacier Inventory for the Barpu/Bualtar Basin, the north facing slopes cover 34 9% of the total basin area, south facing slopes. 8 8%, east facing slopes. 32 8% and west facing slopes. 23.5% of the total area. A distinctive feature of the channels on SSW facing slopes is extensive debris flow fan deposits, whereas NNE facing slopes have comparitively less fan deposits which suggest that freeze thaw cycles are more prominent on southerly slopes During our 1987, field season summer, we witnessed massive debris and mud flows from NNE and SSW facing slopes in Barpu Glacier Basin

The melt water from the Barpu Glacier Basin drains into the Bualtar Glacier just above the terminus of the latter and the combined water of these glacier basins joins the Hispar river near Nagar village Near Aliabad, the Hispar river joins the Hunza River, a major tributary of the Indus which drains many large glaciers of the Hunza valley (Appendix A lists the glacier inventory data for Barpu/Bualtar Glacier Basin)

4 3 SELECTION OF STUDY BASINS

The characteristics of the mountain streams are directly related to the mountain climate (Young 1977) Differences in radiation receipts on different aspects and development of slope winds which result from topographic factors are reflected in air temperature, snowcover duration and consequently the distribution of runoff Among other factors, slope aspect plays a critical role in controlling both the local climate and through that hydrology The essential differences due to aspect are

(a) earlier disappearance of snow and longer snowfree period on southerly slopes than those of north facing slopes.

(b) greater moisture conservation (less evaporation) on northerly slopes and longer snow-covered period

(c) fewer freeze thaw cycles on northerly slopes, as these slopes receive comparatively less radiative energy (in Northern Hemisphere) than those of south facing slopes and also energy drive is controlled mainly by air temperature rather than solar insulation

(d) development of slope winds (katabatic and anabatic) due to difference in snow cover period (see Fig 5 1 4a)

Variations in snowmelt that result directly from slope angle are difficult to isolate Indirectly, however, slope does cause a number of variations The radiation incident on south-facing slopes (northern hemisphere) exceeds that on north-facing slopes For moderate slopes during the spring time, as a result of high solar altitude, the effect of slope is slight During the winter the effect is more pronounced At any given instant, the radiation on a sloping surface relative to the radiation received on a horizontal surface may be determined from the geometry of the

situation (the slope and its aspect in conjunction with the solar altitude and azimuth) Such a determination involves the integration of the solar path relative to the sloping surface It, of course, varies with the time of the year as a result of the changing solar path and is different for every slope and slope aspect.

Slope can affect the direction and the speed of wind by altering the flow of large scale airmass movement such as accelerating wind speed on windward slope and decelerating on the leeward slope and also through creation of local winds of thermal origin such as katabatic and anabatic In addition to these climatic modifications by slope, steeper slopes allow less time to overland flow for infiltration, and therefore greater runoff

To examine the effect of topo and micro climate on snowmelt runoff in Barpu Glacier Basin Central Karakoram, two small snow fed basins were selected on different slope aspects These basins are about 3 km apart on opposite sides of the Barpu Glacier, near the junction of its tributaries, the Sumiayar Bar and Miar Glaciers (Fig 4 2) The main characteristics of these small basins are given in table 4 1 and area-elevation relationship is shown in figure 4 3

Table 4 1 - Physical characteristics of the study basins

Exposure	Elevation range (m a.s.l)	Average slope (degrees)	Total area (km^2)
P Phari (southwest)	3600 - 5000	33 64	1 4
Miar (northeast)	3350 - 4250	41 30	0 995

(source Field observations S I H P 1987)

Area Elevation Curves for the
two Study Basins

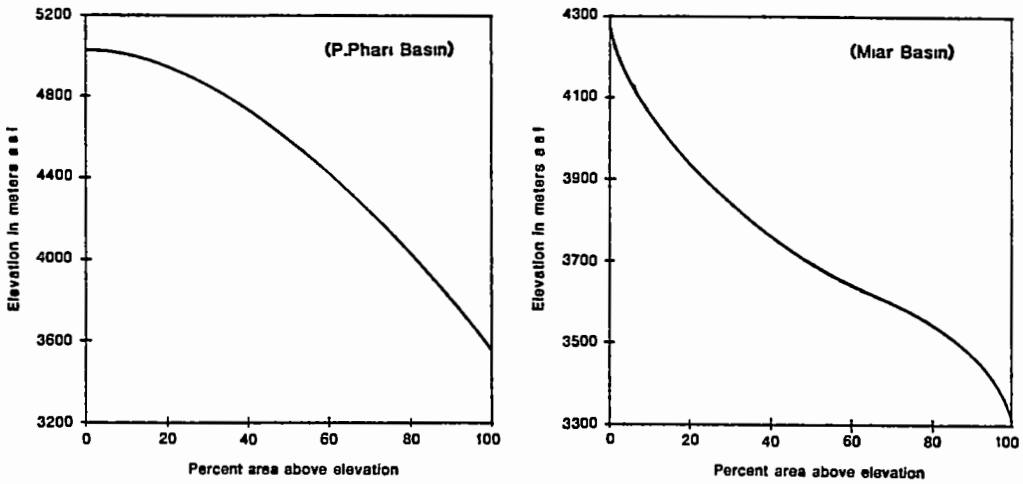


FIG 4.3 Area elevation curves for the two study basins

The basins were given names 'Phahi Phari' (southwest) and 'Miar' (northeast) according to the respective names of mountains in Simpton Map (Fig 4 1a,b) for future reference in this paper In each of these small snow fed mountainous stream basins, hydrological and climatological observations were taken at different elevations. The details of these measurements are discussed under individual headings.

4 4 STREAMFLOW MEASUREMENTS

The site location to measure water level in each of the streams was subject to a variety of constraints. For example, the Phahi Phari stream after leaving the hillside,

flows approximately one hundred and fifty meters in bouldery channel then diverges into two or three sub-channels, which flow in braided reaches on a relatively flat portion of land between the Sumiayar Bar Glacier lateral moraine and the hillside before joining the glacier itself It was decided that the stream had to be gauged as near as possible to the base of its rock-cut channel and above the debris fan to minimize the error due to loss of water to infiltration and evaporation A staff gauge marked in centimeters and a water level recorder attached to an electronic data logger "Ecobug" were installed one hundred meters downstream from the hillside on May 20, 1987 The "Ecobug", was programmed to record water level every 30 minutes (Wake 1986 S.I.H.P Annual Report 1986) Problems with equipment prevented data collection up to June 16 During this period, hourly staff gauge readings were manually recorded from 8 am to 5 pm Data were collected from May 25 to July 20 A large debris flow washed away staff gauge and electronic data logger on July 23

The stream coming from Miar (northest facing stream basin) leaves the hillslope in many small channels on a debris fan However, it converges into one channel after flowing about 170 meters parallel to the Miar Glacier between a lateral moraine and the hillslope The stream was gauged just below the convergence to one channel A staff gauge marked in centimeters was installed on May 26, 1987 Daily readings began on June 3 Hourly gauge data for this stream was obtained from 7 am to 8-9 pm daily from June 3 through July 11 On the latter date, the stream stopped flowing Flow measurements were made to develop stage-discharge rating curve using a Pigmy current-meter Nineteen and 14 discharge measurements were made at Phahi Phari and Miar stream respectively The stage-discharge relation was

determined by plotting discharge measurements against the recorded gauge height on rectangular coordinate graph paper and drawing curve through these points with the help of French curves (Appendix B shows rating curves for these streams) Rating tables were prepared from the rating curves in order to simplify the process of converting gauge heights to discharges Hourly discharges were computed directly from the rating tables. The summary of discharge measurements for each of these streams is given in table 4 2 and 4 3

4 5 METEOROLOGICAL STATIONS Location and description

A quantitative analysis of meteorological elements at key locations within a basin is central to improving understanding of how topographic factors effect local climate such as radiation, air temperature, wind speed and through that snow hydrology Meteorological stations were established with due consideration to studies relating to snow-cover runoff by Popov (1972), Anderson (1972), Logan (1972) and Jolly (1972) In total, five meteorological stations were operational at fixed locations for most of the summer season four "off ice" and one "on ice" These stations recorded meteorological variables such as radiation, air temperature, wind and relative humidity over elevation ranges from 3500 to 4572m a.s.l on the SSW facing slopes behind Phahi Phari Base Camp, and 3500 to 4200m a.s.l on the NNE facing slopes on Miar slope The on-ice S.I.H.P meteorological station was operated in the middle of Miar Glacier Meteorological instruments consisted of both electronic and mechanical equipment. The locations of the meteorological stations are shown in figure 4 2, where 'M' stands for meteorological stations attached with electronic data logger, and 'T' stands for thermohydrograph

Table 4.2

Summary of stream flow measurements (Phahı Pharı Stream)

No	Date	Gauge Height (m)	Velocity (m/s)	Discharge (m^3/s)
1	28/5/87	075	173	014
2	30/5/87	085	404	026
3	01/6/87	125	689	104
4	03/6/87	095	.500	063
5	05/6/87	105	607	099
6	07/6/87	060	204	012
7	07/6/87	128	744	273
8	15/6/87	040	080	007
9	25/6/87	050	221	015
10	28/6/87	090	281	024
11	29/6/87	115	652	126
12	02/7/87	127	.556	121
13	05/7/87	067	722	011
14	05/7/87	150	123	266
15	06/7/87	140	661	224
16	10/7/87	130	598	164
17	12/7/87	103	561	078
18	14/7/87	056	117	013
19	17/7/87	073	224	023

(Source Field Observations S I H P 1987)

Table 43

Summary of stream flow measurements (Miar Stream)

No	Date	Gauge Height (m)	Velocity (m/s)	Discharge (m^3/s)
1	04/6/87	103	279	012
2	05/6/87	070	064	001
3	08/6/87	105	284	013
4	12/6/87	110	312	015
5	14/6/87	097	177	007
6	18/6/87	085	095	004
7	24/6/87	090	109	006
8	26/6/87	100	084	007
9	27/6/87	115	340	017
10	30/6/87	087	112	005
11	04/7/87	095	155	006
12	08/7/87	075	076	002
13	10/7/87	073	068	001
14	11/7/87	080	080	002

(Source Field Observations S I H P 1987)

The setup of meteorological stations can be seen in figure 4 1 Anemometers were placed on top of 2 m high meteorological stands. All radiation instruments were mounted onto the aluminium meteorological stand rods at approximately one meter above the surface Pyranometers and net radiation instruments were fixed to the end of a one meter rod and placed on the south side of the meteorological stand to eliminate shading affects. Thermistors for air temperature were placed in custom made screens. These screen were made of white "CANRON" having a diameter of 7 6 cm and were mounted inside the other tube made of same material having a diameter of 10 2 cm The screen were mounted horizontally in a north-south orientation to protect the thermistor from direct sun rays and reduce the measurement errors.

All thermohydrographs were set in wooden Stevenson screens on approximately 1.5 m high stone cairns A min/max thermometer was placed in all thermohydrograph screens as a check against any "drifting" of sensors Elevation, period of operation, measurement interval and type of instrument used appear in table 4 4 The description of meteorological stations is given below

4 5 1 Southwest Slopes

The main meteorological station M_2 (Fig 4 2) was installed on May 18 just above the Phahi Phari Base Camp at an elevation 3510m a.s.l on southwest facing slopes The station was surrounded by scattered boulders of different size and between these boulders the ground was covered with wild grass and flowers. The station recorded hourly values of wind speed and direction at two metres

Table 4 4

Summary of Meteorological Stations in Barpu Glacier Basin

Site/Elev	Period of operation	Items	Measurement interval	Instruments
M1 (3550 m)	26/5-12/9	Air temperature (0 25, 1 0, 2 0 m)	Hourly	Thermister
M2 (3510 m)	18/5-14/9	Air temperature (0 25, 1 0, 2 0 m) Net radiation (1 0 m) S W Radiation (1 0 m) Wind Speed and direction (2 0 m) R Humidity (1 5 m)	Hourly Hourly Hourly Hourly Weekly chart	Thermister Net radiometer (Fritschen) Pyranometer (Kipp & Zonen) Wind anemometer (Taylor) Thermohydrograph (Casella)
M3 (3200 m)	23/5-12/9	Same as in M2	Hourly	Same as in M2
T1 (4200 m)	26/5-17/8	Air temperature (1 5 m) R Humidity (1 5 m)	Monthly chart	Thermohydrograph (Casella)
T2 (4267 m)	28/5-17/6	Air temperature (1 5 m) R Humidity (1 5 m)	Monthly chart	Thermohydrograph (Casella)
T3 (4572 m)	19/6-16/8	Air temperature (1 5 m) R Humidity (1 5 m)	Monthly chart	Thermohydrograph (Casella)

M1, M2 and M3 were attached with electronic data logger termed as "Ecobug" Sites refers to Fig 4 2

above ground surface using a Taylor wind anemometer, short wave radiation one meter above the ground surface with Kipp & Zonen pyranometer, net-radiation with Fritschen net radiometer and temperature with thermolinear thermistors at 0 25, 1 0 and 2 0 meters above the ground surface The station remained in operation from May 18 to September 14 In general the electronic data logger worked well except for the loss of data from July 18-21 due to failure of the external battery

The second station "T_2" on southwest facing slopes was located 757 meters higher than M_2 A thermohydrograph was installed here at 1.5 meters above the ground surface on a stone cairn at an elevation of 4267m a.s.l This was on a ridge coming down from an elevation 4600 meters On both sides of 'T_2' there are steep gorges due to cutting of channels During the measurement period the mountain slopes above an elevation 4400m a.s.l remained covered with snow This thermohydrograph remained in operation from May 27 to June 17 Later, the same thermohydrograph was moved to the highest station, T_3, on southwest facing slopes and installed at an elevation of 4572m a.s.l on June 18, 1987 (Fig 4 2) The Stevenson screen containing the thermohydrograph was placed on a 1 6 m high cairn on a gentle hill slope in the northwest edge of Rush Lake On the day of installation the ground surface was covered with 0.5-1 0 meter of snow After one month on July 17 most of the snow was melted away except for a few patches around this station. This thermohydrograph provided a continuous record of temperature and relative humidity until it was dismantled on August 19

4 5.2 Northeast Slopes

Station M_1 (Fig 4.2) consisted of three temperature thermistors at heights of 0 25, 1 0 and 2 0 metres above the ground surface These thermisters were attached to the data logger Three thermistors were installed to measure the magnitude of temperature inversion on slopes due mainly to downslope movement of cold air (katabatic) at night The station was installed on a hill slope about 290m above the Miar camp at an elevation of 3550 m a.s.l The ground around the site was sparsely covered with wild grass This station recorded air temperature every hour from May 26 to September 12 Data from this station is continuous with certain missing days (June 3, 4, 13-15, 27-29) during which data logger did not function properly The other station T_1 was set up on a ridge at 4200 m a.s.l A thermohydrograph was installed on a stone cairn 1 6 m above ground surface The depth of the snowpack measured around the site at the time of installation was more than 0 6 m By the end of June whole basin was snowfree except few scattered patches in gullies near the top This provided a record of temperature and relative humidity at charts with a minimum interval of 6 hours time The station was dismantled on September 12, but data could be obtained only up to August 21

4 5 3 Meteorological Station at Miar Glacier

This station was located in the middle of Miar Glacier about 0 6 km from the Miar base station on a relatively clean ice The anemometer was screwed on top of a 2 m high stand rod, pyranometer and net radiometer at 1 0 m, and temperature thermistors at 0 25 m, 1 0 m and 2 0 m above the ice surface All these instruments were attached to a data logger which recorded information at one hour intervals

The recording height of meteorological information is approximate as the actual height above the ice surface continually changes due to ablation which is approximately 6-7 cm per day during summer 1987 To minimize the instrumental height error, adjustments were usually made on every second day From June 26 the height of the instruments were changed as it was difficult to maintain the height and to keep the stand vertical due to vigorous ice melting The new height for the anemometer was 1 6 m, for pyranometer and radiometer 1 0 and 1 2 m and for temperature thermistors ice surface, 1 0 m and 1 5 m above the ice surface Data was collected from the May 23 to September 12 The major portion of missing data is from July 1-11 when the data logger did not function due to external battery failure Also manual measurements for air temperature, relative humidity (at 0 25 and 1 75 m above ground), wind speed (at the same two heights) and incoming radiation and net radiation were recorded from May 23 to July 10 at 08·00 and 17·00 hours so that a comparison with the electronic data logger measurements could be made

4 6 PRECIPITATION AND CLOUD COVER

A cylindrical metric rain gauge was installed near the base camp at Phahi Phari at an elevation 3450m a.s.l on May 18 and remained in operation till mid September Precipitation was measured daily at 0800 and 1700 hrs Total daily precipitation was calculated from 0800 to 0800 hrs Cloud cover was estimated visually in Barpu Glacier Basin and recorded at 0900, 1200, 1500, and 2100 hrs daily From May 18-25 data is available on an hourly basis from 0800 to 1800 hrs Data could be collected from May 18 to July 17 For purposes of analysis the daily mean is calculated from these measurements

4 7 SNOWMELT PATTERNS

A systematic program for monitoring snowmelt patterns using photography was carried out from May 18 to July 23 A photography station was set up behind the Phahi Phari base camp at an elevation 3510m a.s.l From here photographs were taken at weekly intervals except when storms blanketed the area with new snow and a day or two were necessary to allow the winter snowpack to reappear or if frequent cloud cover prevented useful photographs and a day or two would pass before clear shots could be taken Likewise a photo station was set up near a shepherd hut at Sumiayar Bar From here photographs of the Phahi Phari Basin were taken These weekly images were used to estimate the change in areal extent of snow cover and changing snowline elevation Detail of the analysis is given in Chapter 5 2

4 8 SNOWPIT STUDY

A snowpit was dug on June 18, 1987 at an elevation of 4570m a.s.l on the southern edge of Rush Lake to observe the meltwater percolation pattern (Fig 4 2) Snowpack stratigraphy was delineated on the basis of layer thickness, hardness and colour, and temperature Density was measured using 23 cm long stainless steel tube with a cross-sectional area of 41 7 cm^2 The physical characteristics observed are given in Table 4.5 Continuous ice layers around the snowpit walls and below freezing temperatures of snowpack from surface to bottom suggest that no conduit type of meltwater percolation occured before June 18 It was assumed that the contribution to streamflow from the snow at these elevations was minimal prior the date of observation

Table 4 5 Physical characteristics of snowpit dug on 18 June, 1987 in southern edge of Rush Lake at 4570 m a.s l in Barpu Glacier Basin, Central Karakoram

STRATIGRAPHIC CHARACTERISTICS					
Depth cm	Temp °C	Hardness/ Colour	Sample length	Density gr/cm³	Remarks
8.5	-2 0	vs/w	8.5	25	soft snow
9.5	-2.5	s/l b	9.5	38	do
1 0	-2.5	ice			
13.5	-2.5	m.h/w	14 5	.51	ice layer 1cm
1 0	-2.5	ice			
16.5	-2.5	s/w			
0 7	-2 5	ice	18 2	.50	ice layer 1 7cm
1 7	-2 6	s/w			
0 8	-2 6	ice			
20 0	-2 7	vs/w	22 5	49	ice layer 8cm
21 5	-3 0	s/w	21 5	49	
22 0	-3 0	s/w	22 0	45	
			Avg	44	Total 3.5cm

† s = soft, v.s = very soft, w = white, m.h = medium hard,

† l b = light brown, Weather conditions sunny, hot and calm

† Time of study 0900 to 1200 (local standard)

4 9 CONCLUSIONS

Meteorological and streamflow data in the experimental basins were effectively collected from mid May to mid September On the whole all meteorological stations ran for the better part of the summer over an altitudinal range of 3400m to 4600m a.s.l

CHAPTER 5

Analysis of Hydrometeorological Data and Results

This Chapter is divided into three parts The first part, 5 1, deals with temperature gradient on two slopes with different aspects within the Barpu Glacier Basin, the second part, 5.2, explains the factors causing variations in discharge regime in the two study basins, and section 5 3 defines and evaluates the relative importance of different climatic parameters in snowmelt runoff prediction models

5 1 ALTITUDINAL GRADIENT OF TEMPERATURE

5 1 1 Introduction

There have been very few hydrometeorological observations from upland areas in the northern regions of Pakistan, and those that exist vary widely in respect to both quality and period of record The scarcity of records is due to the inaccessibility of the region combined with distance and frequent adverse weather conditions The need for standard observations has not been considered important despite the fact that ever increasing demands are being made on the upland environment, by tourism, forestry and above all the water and power industry A good understanding of the upland climate, especially the climate prevailing in the

Himalayan sources of the River Indus, which is the major source of fresh water supply for Pakistan, is essential if the effects of these often competing activities are to be properly assessed.

Temperature lapse rate is undoubtedly the single most important aspect of mountain climates which controls the whole hydrologic system on mountain slopes. It is particularly important in stream basins with large elevation ranges. Decrease in air temperature with altitude favours precipitation input in the form of snow rather than rain, while seasonal freezing and thawing determines the release of meltwaters.

The variation of climate within this mountainous region, as in the free atmosphere, is primarily due to altitude and topography The interaction of the atmosphere with the uneven surface of the mountains introduces further variability By comparison with the surrounding atmosphere, the slope air over mountains is affected by radiative and turbulent heat exchanges These processes modify the temperature structure over the massif and result in adiabatic lapse rates which differ from environmental lapse rate according to the time of the day (Barry 1981) These variations are described below for the Barpu Glacier Basin, Central Karakoram and an attempt is made to account for them in physical terms

5 1 2 Observation Sites and Data

Temperature gradients are calculated for slopes of two different aspects (SSW and NNE) Site locations are given in figure 4 2 Elevational range of the study sites and period of record used are given in Table 5 1 1 Although the data collection

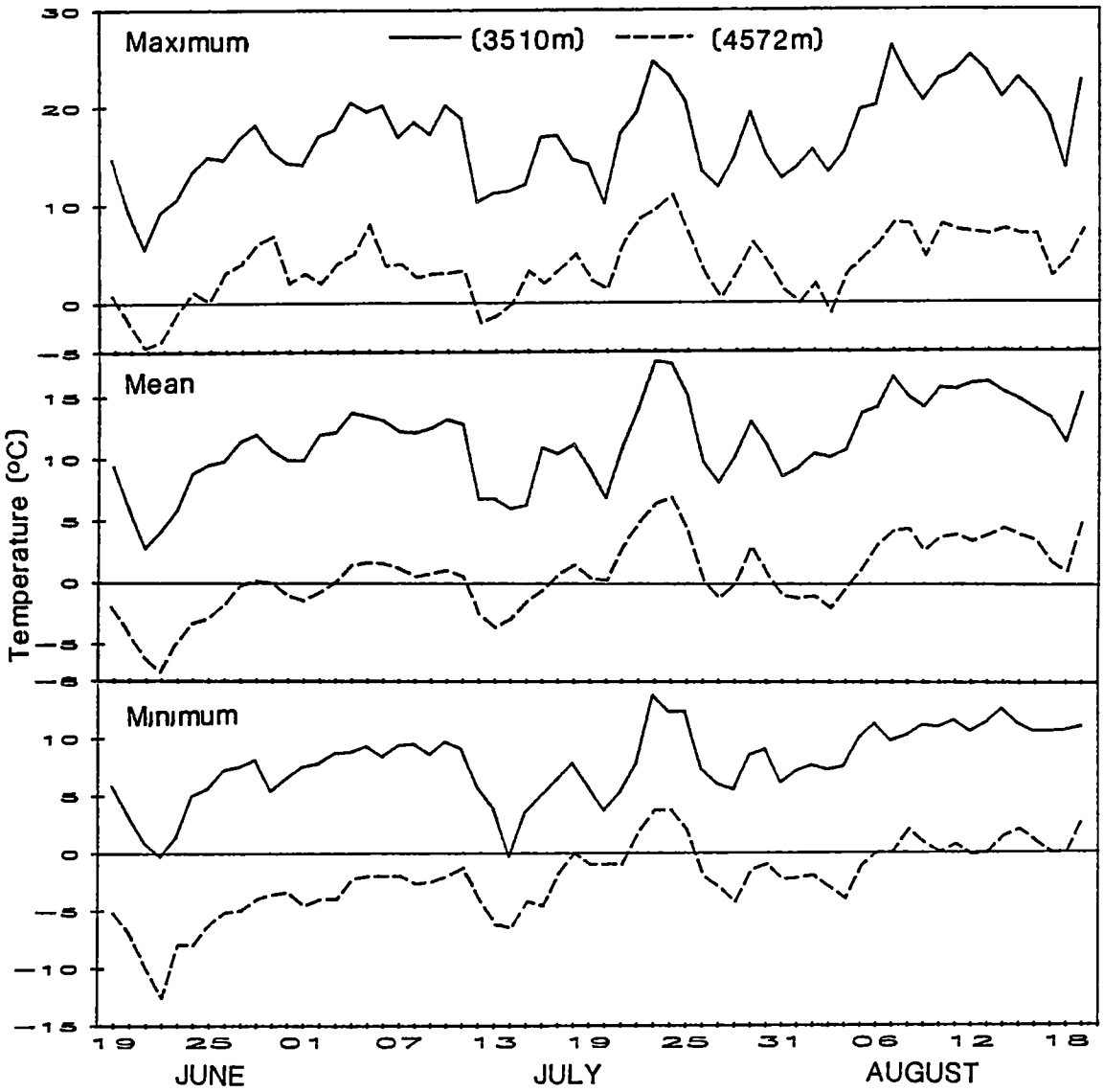
period was limited to one summer season the temperature gradients calculated are important for two reasons. Firstly, this is the first measured and sustained environmental temperature gradient study ever made from data collected above 3000m a.s.l in the Central Karakoram Mountains Before this, some scientists have calculated lapse rates from widely scattered temperatures observations (Schlagintweit 1972, Hewitt 1968, Whiteman 1985) Secondly, the environmental lapse rates calculated in the present study were measured directly from the hill slopes during summer and will aid significantly in explaining the snow melt phenomena on these slopes Moreover a major portion of summer snowmelt runoff is contributed by the snowpack above 3000m a.s.l. (Hewitt 1985) Therefore, this study provides a necessary basis for further studies in the Karakoram and may also be helpful in snowmelt runoff forecasting models.

5 1 3 Results and Discussion

Figure 5 1 1 (a,b) shows the trends of the temperatures plotted from lower and higher stations for each slope aspect. Results are given in Table 5 1 1 and are discussed separately on the basis of aspect

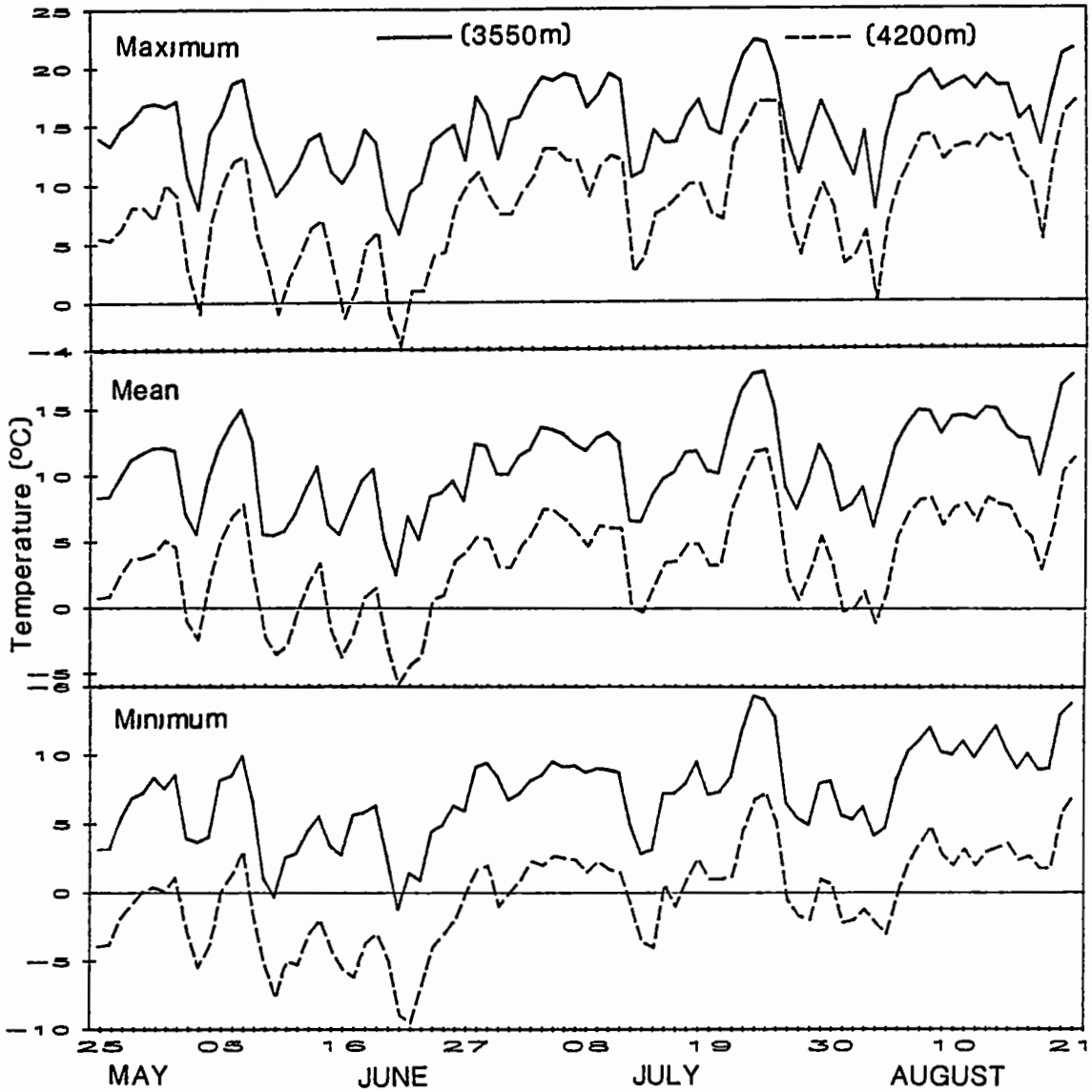
5 1 3.1 Southwest Slopes

Table (5 1 1) shows that the environmental gradient of maximum temperatures is considerably higher than the gradient of mean and minimum temperatures The standard deviation for maximum temperature gradient is also higher than those of mean and minimum temperatures. This greater standard deviation of maximum temperature gradients, 2 29 °C Km^{-1} (Table 5 1 1), corresponds well with the results



RELATION BETWEEN AIR TEMPERATURE PROFILES RECORDED AT 3510 and 4572m a s.l ON SSW FACING SLOPES IN BARPU GLACIER BASIN CENTRAL KARAKORAM

FIGURE 5 1 1b



RELATION BETWEEN AIR TEMPERATURE PROFILES RECORDED AT 3550 and 4200m a.s.l ON NNE FACING SLOPES IN BARPU GLACIER BASIN CENTRAL KARAKORAM

Table 5 1 1

Altitudinal Gradients and Standard Deviations for the Daily Max, Min and Mean Temperatures

	Period of record	No of days	Altitudinal gradient (°C/100 m)	Standard deviation (°C)
MAXIMA				
SSW	Jun 19 - Aug 19	61	1 24	2 29
NNE	May 25 - Aug 21	89	1 08	1 62
MINIMA				
SSW	same	same	0 93	1 76
NNE	same	same	1 15	0 94
MEAN				
SSW	same	same	1 03	1 30
NNE	same	same	1 12	0 95

Elevation range for southwest and northeast, is 3510 - 4572 m and 3550 - 4200 m respectively

observed by Harding (1978) during a study in the Northern Pennines Scotland He observed a standard deviation of 4 0°C Km^{-1} for the daily values of maximum temperatures and suggests that this variability decreases with increasing averaging period, and so the observed standard deviation decreased to 1 3°C Km^{-1} when monthly means are used to calculate environmental gradients of maximum temperatures (Harding 1978, p 193)

The gradient of minimum temperature is less than that of maximum temperature This is to be expected since large-scale drainage of cooled air down the hill slopes at night suppress air-temperature at lower elevation sites This shows that minimum temperatures are more dependent on the local topography and siting of station, topographic variations are greater in mountainous regions which result in larger spatial variation of minimum temperature (Harding 1979)

The gradient of maximum temperatures is especially dependent on sunshine duration Harding (1979), states that altitudinal gradient of maximum temperature increases linearly with the increase in sunshine duration at the lower elevation site He reports a 3°C Km^{-1} higher gradient of maximum temperatures during periods of appreciable sunshine in the Northern Pennines than in periods when there is no direct sunshine at the upland or lowland sites The same is quite evident from the results obtained in the present study on southwest facing slopes, where sites remain exposed to the sun for a much longer period of the day have greater gradient of maximum temperature (Table 5 1 1) as compared to the northeast facing slopes

The gradient of air temperature is also affected by the presence of snowpack as the snow surface never exceeds 0°C and has a cooling effect on air temperature To

elaborate on the cooling effect of snowcover on air temperature, the temperature gradients were calculated by splitting the total period of observation (Table 5 1 1) for each aspect into two parts In table (5 1 2) *A* represents the period when the ground surface around the higher elevation site was covered with snow and the lower site was snowfree, period *B* when both higher and lower sites were snowfree The results given in Table (5 1 2) shows lower gradients of maximum, minimum and mean temperatures during the period *B* than for the period *A* on both aspects A slightly higher gradient of maximum temperatures on SSW slopes for period *B* is due mainly to the location of the upper meteorological site This site was on snow covered slope just behind a ridge facing northeast and remains in shadow during second half of the day (for location description see Chapter 4, section 4.5) In addition the maximum temperature here are suppressed by downslope drainage of cooled air (katabatic) due to the presence of snow (see Fig 5 1 2b)

Table - 5 1 2 Comparison of temperature gradients calculated for two different periods for SSW and NNE slopes (°C/100 m) *A* when the ground was covered with snow around the higher site, *B* when both (higher and lower) sites were snowfree

	SSW Slopes		NNE Slopes	
	A	B	A	B
MAXIMA	1 21	1 26	1 25	0 97
MINIMA	1 02	0 87	1 22	1 09
MEAN	1 06	1 02	1 21	1 05

† Elevation range for both aspects are same as in table 5 1 1

† SSW A (June 19 - July 15) B (July 16 - Aug 19)

† NNE A (May 25 - June 30) B (July 01 - Aug 21)

These results suggest that snow has a cooling effect on air temperature and increases the gradient by decreasing air temperature at higher sites for the period when ground around the site remained covered with snow

5 1 3.2 Northeast Slopes

In contrast to the southwest facing slopes the environmental temperature lapse rate of maximum temperature is lower on northeast facing slopes than the gradient of mean and minimum temperatures (Table 5 1 1), although the standard deviation for maximum temperature gradient is higher than the mean and minimum temperatures similar to that of the southwest aspect The higher gradient of minimum temperatures is due to the fact that the higher site is on the summit, and a summit site has higher minimum temperature, which is generally bout 1°C higher than that which would be expected from a valley-side site at the same altitude (Harding 1978, p 43 & Harding 1979, p 194) These greater night-time temperatures are expected where air cooled radiatively drains down on the slopes away from the summit This indicates that the effect of local topographic variations around a site can be as great as the effects of altitude

The altitudinal temperature gradient near mountain slopes is dominated by the local drainage system of air, which cools on contact with the ground at night, this is recorded both at higher and lower sites Figure 5 1 2a shows the effect of the slope's wind on air temperature recorded at three different heights (0 25 m, 1 0 m and 2 0 m above the ground surface) on SSW facing slope at 3510 m a.s.l Here the temperature regime recorded at 0 25 m above the ground surface is higher than at 2 0 m during the day time due to the thermal effect of rocks and lower during

FIGURE 5 1 2

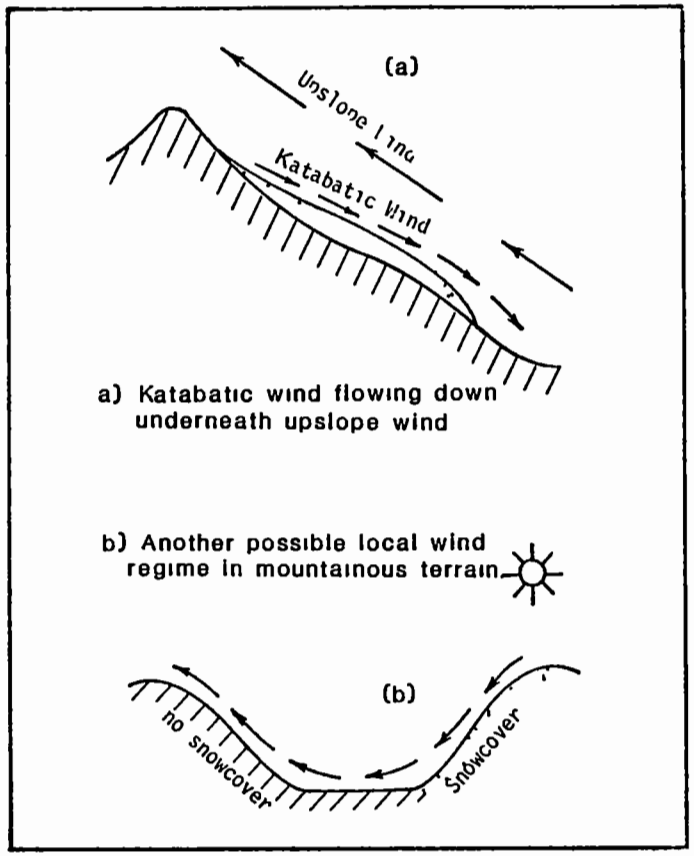


FIG 5 1 2b

Development of local wind regime in mountainous terrain in the presence of snow cover
(Source Obled and Harder 1979)

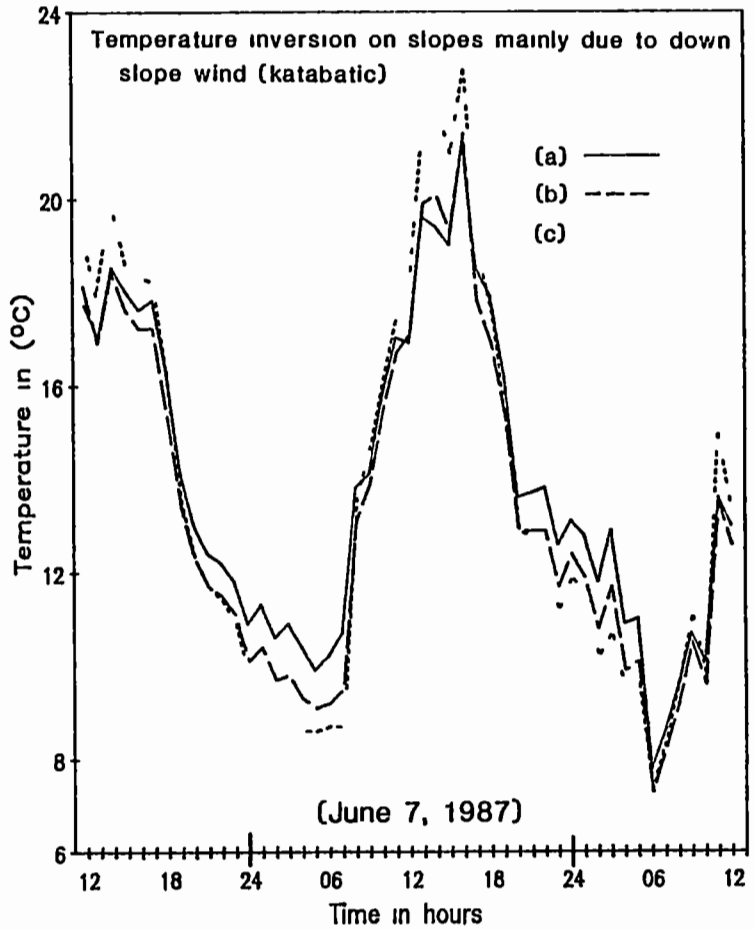


FIG 5 1 2a

Air temperature recorded at (a) 2m (b) 1m (c) 0 25m above ground surface on SSW facing slopes at 3510m a s l in Barpu Glacier Basin, Central Karakoram
(Source Field Observations SIHP 1987)

night times due to downslope drainage of cool air (katabatic winds) under the radiatively heated warm air rising upslope Also, the patterns of slope winds are given in figure 5 1 2b to illustrate the development of katabatic wind system in the presence of snow on sloping surfaces. This shows that the location of the sampling site is of prime importance for any scientific study in mountainous regions.

5 1 4 Maximum Isotherm

The transect between 3510 and 4572m a.s.l yielded a lapse rate, from daily maximum temperature, of 1 24°C/100 m and 1 08°C/100 m on southwest and northeast facing slopes respectively (Table 5 1 1) The upper limit of the 0°C daily maximum isotherm was shown to be at about 5700 m a.s.l These data correspond well with data gathered during the 1953 German Nanga Parbat Expedition which found the upper limit of the 0°C daily mean isotherm to be in the vicinity of 5300m a.s.l., and also with the maximum isotherm limit of 0°C calculated at 5800 m a.s.l in Rakhiot Valley (Gardner 1986) It may be noted that Nanga Parbat is subject to higher summer cloudiness and precipitation than the Karakoram (Hewitt 1985) In the last century, Schlagintweit (1867) demonstrated that main summer isotherms and snowlines rise from south to north within the main Himalayan Range across the Karakoram

5 1 5 Conclusions

The temperature gradient is reasonably constant on both aspects, provided certain effects of exposure are taken into account, particularly for the minimum temperatures. The estimation of higher elevation temperature from lower station

observations will be limited by the temporal variability of the gradients calculated for two different periods (Table 5 1.2) This variability is the result of complex interactions of the synoptic weather patterns. Therefore the use of simple synoptic indices and the mean seasonal variation removes only a proportion of the temporal variability, but it is to be expected that a better understanding of the mechanisms of the study area's climate will allow improved estimates to be made in future

5.2 VARIATION IN SNOW-MELT RUNOFF

5.2.1 Overview

This section describes the environmental factors that determine differences in the volume and timing of snow melt runoff Meiman (1970) has partitioned these factors in two groups. Firstly, there are those related to the atmosphere such as air mass characteristics influencing precipitation - e g precipitable water, temperature characteristics, circulation patterns, and precipitation processes - as well as those affecting the radiant, latent, and sensible energy exchanges in the snowpack and mass redistribution of snow after snowfall Secondly, there are land surface factors such as topography and vegetation cover

Price and Dunne (1976) mention topography as a major factor influencing the amount of snow falling over a watershed and it's spatial distribution They suggest that during the snow melt period, the main effect of topography is to increase the spatial diversity in snow melt rates resulting in a staggering in time in the release of meltwater over the watershed However, the effect of these factors cannot be differentiated as a rule, since the laws governing them are interdependent and their magnitudes are not, in general known That is especially true in the mountainous areas, where the effects of these factors are three dimensional (Alford, 1985) From a practical point of view the factors affecting the snowcover runoff can be grouped together into two main categories the first describing the existing state (variability) of precipitation input and the second determining its future melt or accumulation (Anderson 1972)

The areal distribution of the snow cover is considered the most important factor

affecting runoff regime (Rawls et al 1980) This is a consequence of meteorological conditions during snowfall, especially temperature and humidity, which affect the initial density and transportability of snow Wind velocity and direction during and after deposition lead to different areal patterns of snow accumulation Topographical and vegetational factors regulate the effects of meteorological factors and also directly determine many characteristics of snow accumulation Three different scales are recognized in the areal distribution of snow cover

1 Microscale variability This can be defined as the variation of snow – properties over a fairly homogeneous area. This area can be a section of horizontal, open field or a section of a slope with a constant angle in any terrain type Characteristic linear distances of microscale variability range from a few centimeters up to 100 m

2 Mesoscale variability This is caused by variation of physiographical – factors. terrain types, slopes, aspects, variations in vegetation cover etc The characteristic linear distance of mesoscale variability depends on the scale of variation of physiographical factors In the Karakoram Mountains it usually ranges from a few hundreds of meters to several kilometers

3 Macroscale variability This depends mainly on the variation of climatological factors over a region Typical scales in Pakistan are from a few kilometers to hundreds of kilometers.

These three types of variability have been used by several snow hydrologists (e g Gray et. al 1978, McKay & Gray 1981) though the definition of characteristic linear distances varies The extent of variability primarily depends upon the location, geometry and height of mountain ranges The Karakoram Mountains having significant elevation serves to magnify this variability Energy availability is highly diurnal due to location in the mid latitudes This energy input pattern also increases

variability during melt season The present study is concerned with micro and mesoscale variability, as none of the two basins is more than 1.5 km^2

5.2.2 Snow Disappearance and Runoff

In the computation of snowmelt runoff, attention is usually focused on determining snow melt rate However, accuracy can be improved by considering the changing snow coverage of the watershed This factor may be of little importance when considering flat areas with uniform snow cover or in forecasting only total snow melt volume But in mountain basins where the snow covered area gradually decreases from approximately 100 percent of the watershed towards zero, it plays a dominant role, especially where the snowpack depth and water equivalent increases substantially or the rate of ablation decreases with elevation

Practically all of the water yield from the two monitoring basins with the opposite aspect, is the result of snowmelt during spring A correlation between diminishing area of the snow cover and increase in streamflow is thus to be expected up to certain extent Furthermore, in the Karakoram Mountains, the timing of snow melt is largely a function of the intensity of incident solar radiation and altitude The former is, in turn, a function of aspect and slope

Normally snow covered surfaces become bare on south slopes first and subsequently north facing surfaces at lower elevations South slopes at high altitudes become bare earlier than north slopes at high elevations (Gartska 1958, Anderson 1972, Kuusisto 1984) This consistency in the progression of snow melt facilitates mapping and recording of the disappearance of the snowpack Such a record for the

two study basins was made in the spring of 1987 along with runoff measurements to indicate the relative contribution to streamflow with the percentage depletion of snow covered area.

5.2.2.1 Assessment of the areal extent of snowcover

In mountains with rounded peaks it may be possible to determine snow coverage by visual observation or by planimetering photographs. But in watersheds with rugged terrain and sharp mountain ridges such evaluation is impossible because the snowline is extremely variable and the snowpack consists of numerous dispersed high density snow patches, especially in gullies.

The following procedure was used to determine snow coverage in the monitoring basins at Barpu Glacier Photographs were taken at weekly intervals except when storms blanketed the area with new snow and brief delay was necessary to allow the pattern of the winter pack to reappear Because of the dispersed snow cover on the lower parts of the basins it was decided to estimate percentage of cover rather than defining the altitude of a snowline as is usual in past work in the Himalaya. The percentage of snow covered area was found by drawing matrix grid lines on photographs. From these grids proportions of white (snow covered ground) and dark (snow free ground) were calculated This information was then transferred to contour maps A series of maps was prepared defining the snow cover on specified dates for each basin (Appendix C and D for P.Phari and Miar basin respectively) Table 5 2 1 summarizes the plan area (percent) extent of snow determined from maps and adjusted for each elevation band for both monitoring basins.

Table 5 2 1

Comparision of snow cover percentage in each elevation band for the study basins

Elevation in (ms)	Dates						
	May 24	May 31	June 8	June 14	June 28	July 5	July 19
3350 - 3658	--/**	--/11	--/12	--/--	--/--	--/--	--/--
3658 - 3962	5/**	--/18	--/23	--/20	--/20	--/--	--/--
3962 - 4267	20/**	4/50	2/20	--/17	--/11	--/35	--/--
4267 - 4572	25/--	20/--	6/--	2/--	12/--	--/--	--/--
4572 - 4800	37/--	35/--	30/--	17/--	9/--	05/--	--/--
4800 - 5029	100/--	100/--	100/--	100/--	95/--	85/--	70/--

* Southwest facing basin/northeast facing basin

** data not available

-- snowfree

This is a rough estimate of snow covered area, because photographs could not show all the snow patches in the complex terrain The Phahi Phari stream basin was more rugged in terrain and more problematic in this respect than the Miar basin However, in the absence of aerial photographs this is considered to be an appropriate method to estimate snow covered area.

5.2.2.2 Relationship between snow cover depletion and runoff

The snow cover depletion during the melting period has an essential influence on meltwater production, especially in watersheds with considerable elevation range The shape of the areal depletion curve varies considerably from basin to basin Anderson (1977) has given examples of the characteristic shapes of depletion curves showing relationship between snowfree area and water equivalents (W/W_{max}) where W is the water equivalents at any time during the melt period and W_{max} is the maximum water equivalents at the time of one hundred percent snow coverage (Fig 5 2 1)

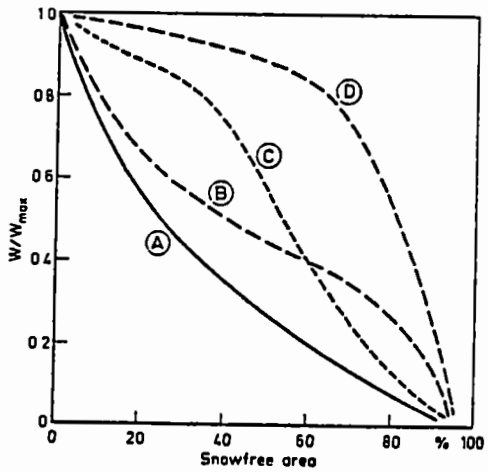


FIG 5 2 1 Some characteristic shapes of areal depletion curve of snow free area (%) vs water equivalents (Anderson 1977) Explanation in the text

Curve A indicates that bare ground appears at a continually increasing rate as snow cover abates This kind of curve is typical of basins where both snow accumulation and melting vary relatively evenly over the area. Curve B is similar to curve A in the early stages of melting, but later the rate of appearance of snow free areas becomes slower This indicates that a portion of the basin accumulates considerably more snow or has a much lower melt rate than the rest of the basin

Curve C is similar to curve A in the middle and at the lower end In the beginning, however, curve C indicates that the areal cover drops off very rapidly when the melting begins This suggests that a portion of the area accumulates much less snow or has much higher melt rate - or both - than the remainder Curve D is for an area which can be basically divided into two extremes, i e low accumulation and/or high melt rates and high accumulation and/or low melt rates Figure 5 2.2 shows the plotted values of snowfree area and subsequent runoff for the monitoring basins It is worthwhile to mention here that runoff figures for Phahi Phari are subject to the total streamflow measured up to July 20 when the basin still had some snow on higher elevations.

Data are not available for the early part of the melt season, but later parts (June-July) of these depletion curves are certainly like that of either B or D as explained above This shows that the rate of appearance of snowfree areas decreases and in this instance reflects the fact that a portion of the basin accumulates considerably more snow on higher elevations The plots for the two basins exhibit a uniform pattern (Fig 5 2 2) which suggest that the maximum snow covered area, before snow melt sets in, has a direct relationship to the subsequent snow melt runoff

Snowfree Area vs Runoff for the
Two Study Basins

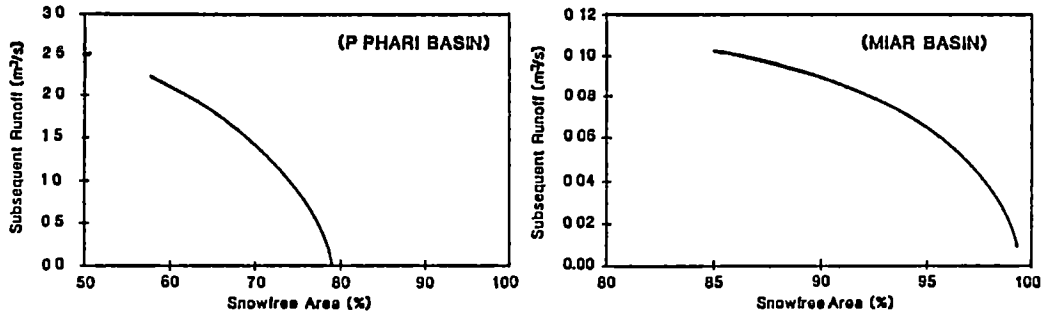


FIG 5.2.2 Snowfree area (%) and subsequent runoff for the two study basins

This relationship has reported elsewhere (Rango et al 1977, Gupta et. al 1982) However, it is observed in this study that, for each basin, there exists a separate rectilinear relationship between snow cover and snow melt runoff (Fig 5.2 3)

Relation Between Snow Covered Area and
Subsequent Runoff

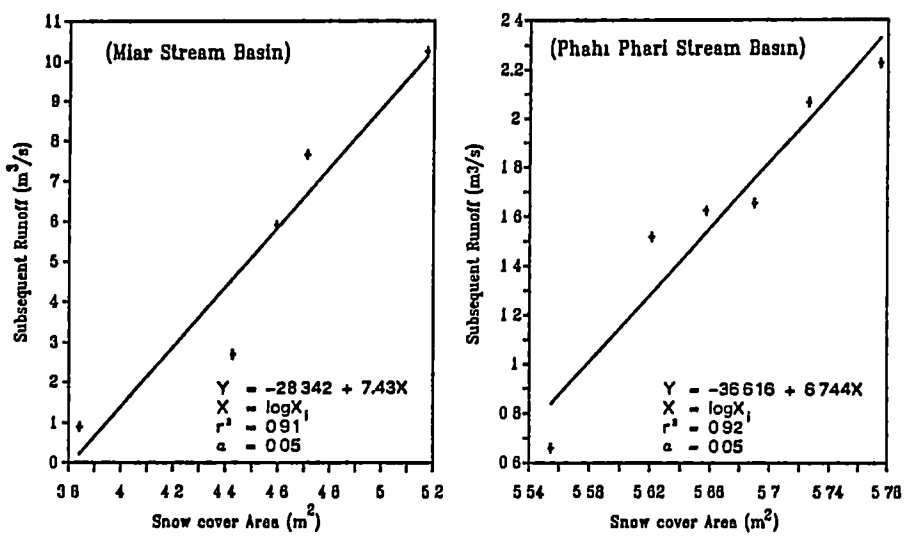


FIG 5 2.3 Relation between snow covered area and subsequent runoff

To illustrate the fact that more snow occurs at higher elevations, change in snow covered area (percent) and basin water yield are related in Table 5 2 2a and Table 5.2 2b for Phahi Phari and Miar Basin, respectively

Table 5 2 2a suggests that during May 24 - May 31, a 4.5 percent change in snow covered area produced 2 3 mm water yield over the whole basin per day By comparing this period with June 29 to July 6, it can be shown that slightly less change in snow covered area produced three times the water yield per day That is possible only when the snowpack at higher elevation have much greater depth Now table 5.2 1 shows that on June 28 there was hardly any snow left below elevation 4572m a.s.l suggesting that greater water yield measured in streamflow after June 28 is due to melting snow from above 4572 m a.s.l Peak discharge was also observed during this period on July 5

The northeast facing basin (Miar) has lower elevation range 3350 m to 4250m a.s.l and greater slope than Phahi Phari i e 41 3°, which shows a small difference in water yield as compared to that of the Phahi Phari stream basin (Table 5 2 2b) Figure 5 2 4 (a, b) are the composite charts consisting of hydrographs of Phahi Phari and Miar streams showing the snow coverage in relation to flow

5.2.2.3 Logarithmic relationship between snow-cover area and streamflow

The logarithmic relationship between the snow-cover area and snow melt runoff implies that early increments in snow cover area lead to smaller increases in snow melt runoff than later increments in snow-cover area of the same magnitude (Fig 5 2 3) This seems logical, since with the setting-in of snowfall season, the

Table 5.2.2a - Relation between change in snow covered area and basin water yield (Phahi Phari stream)

Period	Δ Snow covered (area %)	Mean Daily Q (m^3/s)	Water Yield (mm)
May 24 - May 31	4.51	0373	2.295
Jun 01 - Jun 08	3 03	0391	2 405
Jun 09 - Jun 14	2.53	0052	0 320
Jun 15 - Jun 28	2.59	0076	0 468
Jun 29 - Jul 06	4 21	1070	6.580
Jul 07 - Jul 19	4.54	0510	3 140

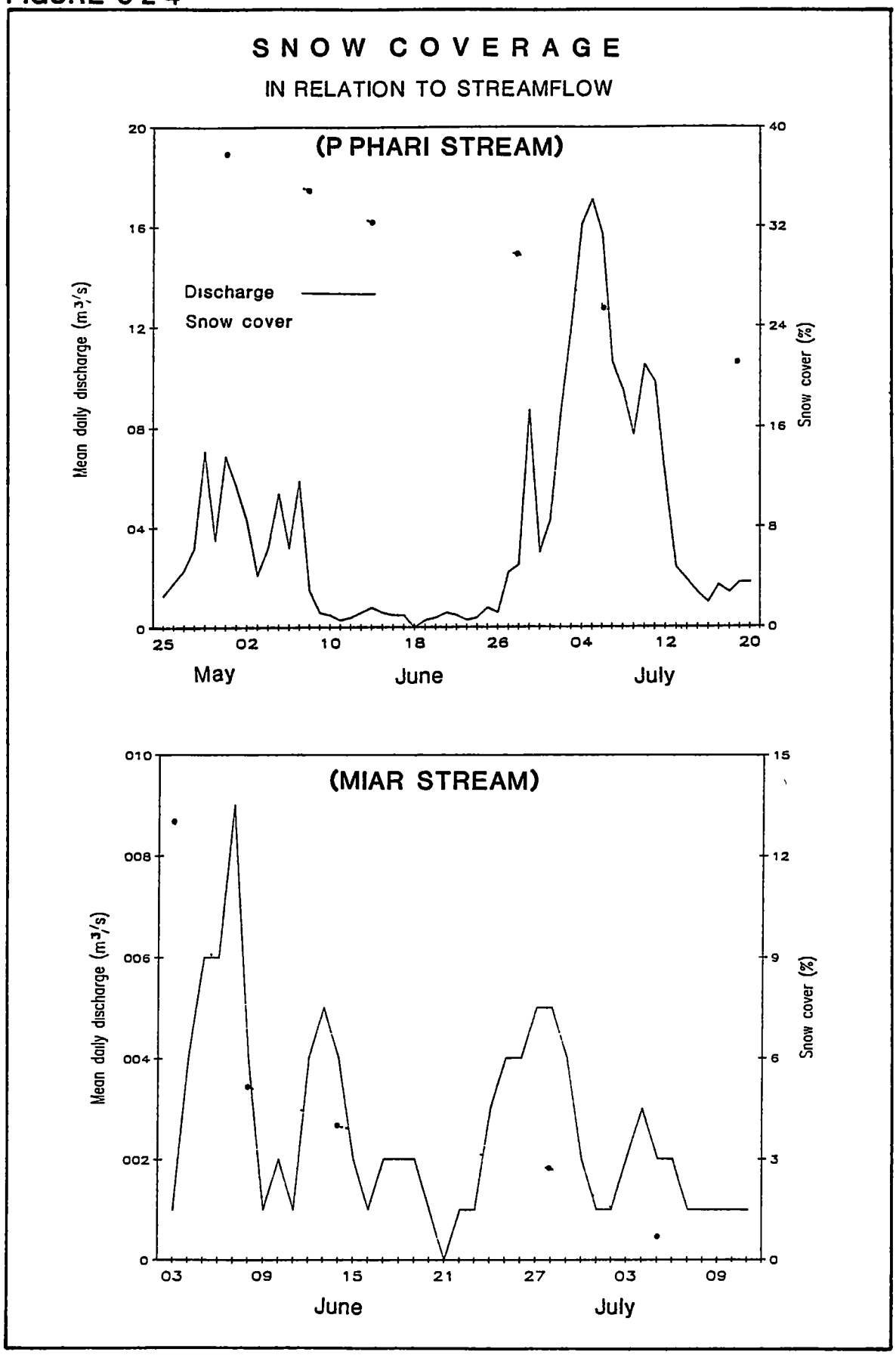
(Source Field Observations Ali 1987)

Table 5.2.2b - Relation between change in snow covered area and basin water yield (Miar stream)

Period	Δ Snow covered (area %)	Mean Daily Q (m^3/s)	Water Yield (mm)
May 31 - Jun 07	9 84	0052	0 45
Jun 08 - Jun 13	1 20	0030	0 26
Jun 14 - Jun 27	1 24	0023	0 20
Jun 28 - Jul 05	2 00	0025	0.22

(Source Field Observations Ali 1987)

FIGURE 5 2 4


SNOW COVERAGE
IN RELATION TO STREAMFLOW
(P PHARI STREAM)
Discharge
Snow cover
Mean daily discharge (m³/s)
Snow cover (%)
20
16
12
08
04
0
40
32
24
16
8
0
25
02
10
18
26
04
12
20
May
June
July
(MIAR STREAM)
Mean daily discharge (m³/s)
Snow cover (%)
010
008
006
004
002
0
15
12
9
6
3
0
03
09
15
21
27
03
09
June
July

seasonal snowline descends to lower elevations on the periphery of the permanent snowline But due to a strong precipitation gradient in the Karakoram Mountains lower elevations are thinly snow covered As a result, increase in snow-cover area takes place, but is also accompanied by an ever-increasing snowdepth at higher elevations Thus during the melt period, a marginal increase in snow free area at a later date, implies a large volume of seasonal flow Hence the relationship between snow-cover area and snowmelt runoff is logarithmic

5 2.3 Description of Snowmelt Runoff Pattern

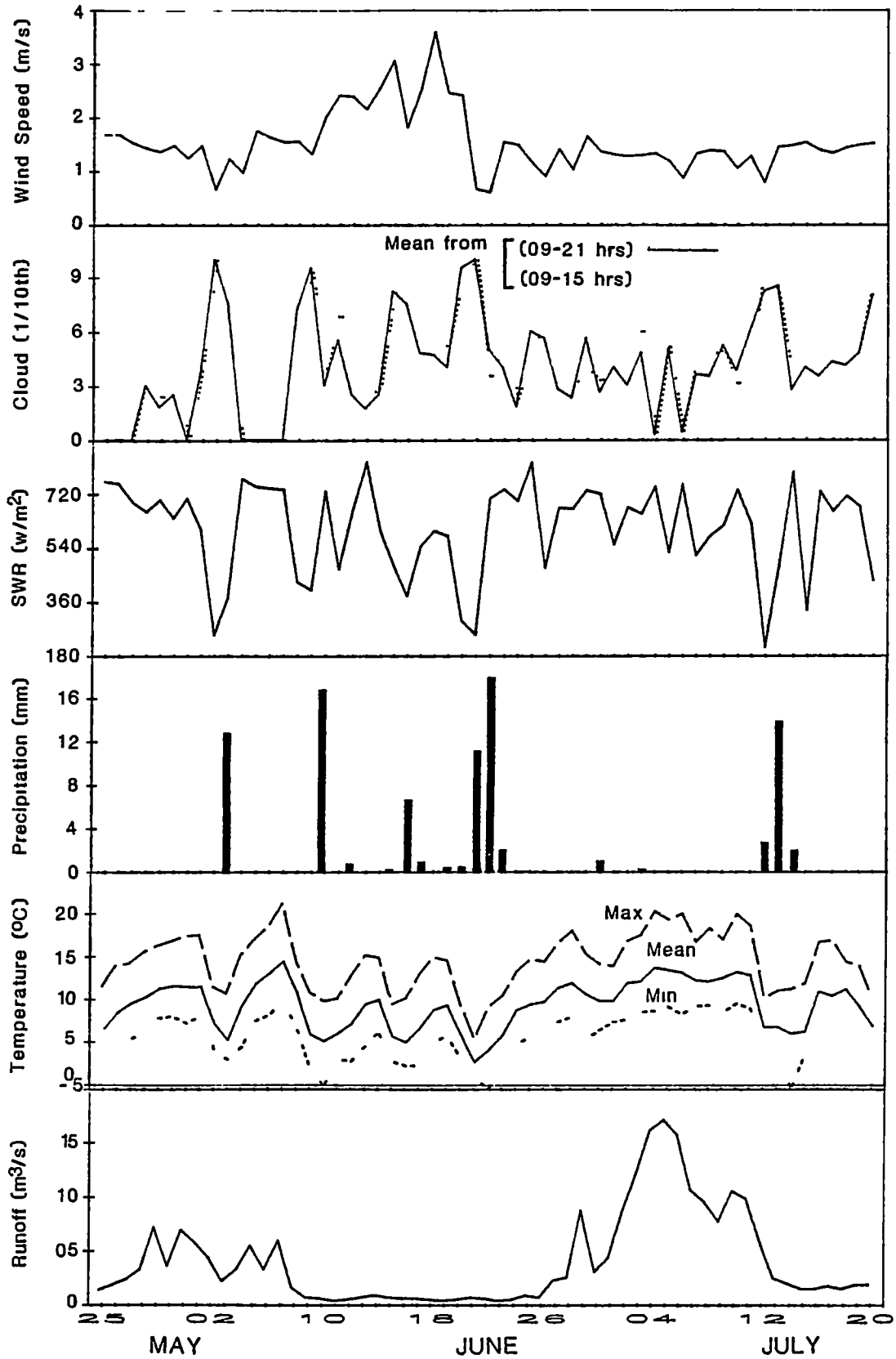
This section describes the stream flow patterns based on data collected during the spring of 1987, in Barpu Glacier Basin, Central Karakoram Runoff during the full snow melt period is examined first, followed by diurnal variation in discharge

5.2.3.1 Seasonal runoff pattern

Southwest facing stream (P.Phari)

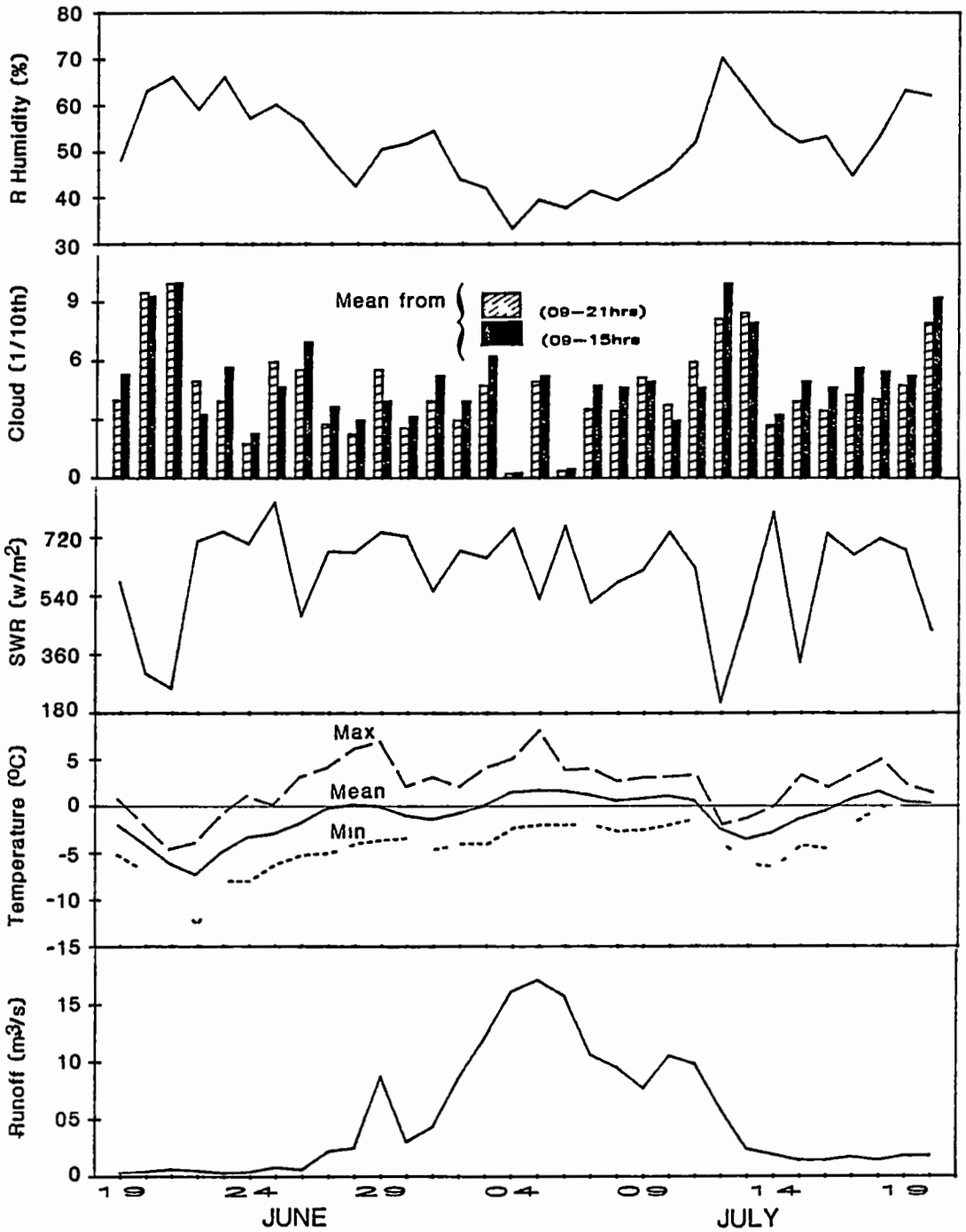
The shape of the hydrograph represents the interaction of the climatic and physical characteristics of the drainage basin Figures 5 2.5a & 5 2.5b show the seasonal and daily distribution of the volume of flow and related meteorological factors recorded at different elevations for the stream draining the southwest facing basin (Phahi Phari)

Monitoring of the stream commenced on May 25, 1987 At this time 42 4 percent of the basin area was covered with snow Almost all the snow was above 3658m a.s.l., except a few patches of avalanched snow in gullies at lower elevation (Table 5 2 1) The area above 4800m a.s.l is a depression which seemed to act as a



MEAN DAILY RUNOFF (PHAHI PHARI BASIN) AND RELATED CLIMATIC VARIABLES RECORDED AT 3510 m a s l.

FIGURE 5 2 5b



RUNOFF FROM P PHARI BASIN AND RELATED CLIMATIC VARIABLES RECORDED AT 4572 m a s l

snow trap The hydrograph for this stream can be roughly divided into three periods on the basis of flow volumes i e May 25 to June 8, June 9 to June 27, and June 28 to July 14 (Fig 5 2.5a)

May 25 - June 8

During this period the basin produced a small amount of its potential discharge, despite the maximum air temperature for the season being recorded on June 7, and on average mean daily short wave radiation for the period was 638 w/m^2 The low stream flow can be explained first, possibly an important portion of the runoff drained before May 25 when the monitoring began However, there was no indication of higher flows prior to our arrival. Alternatively, the snowpack may have been too thin at lower elevations to produce substantial runoff No detailed information is available for this period. However the results of section 5 2 2 and considering previous studies, (Batura Glacier Group 1979 and Wake 1987) which observed considerable increase in snow accumulation with elevation, the second factor would seems to be more relevant

June 9 - June 26

Table 5 2 1 shows that on June 8 there was little snow remaining below 4572 m, whereas mean air temperature recorded at lower site at 3510 m a.s.l show that there was enough energy to produce meltwater should there have been any snow remaining at these elevations (Fig 5 2.5a) In comparison to this there was enough snow above 4572m a.s.l but too little energy to produce snow melt runoff Air temperatures calculated from a thermohydrograph at an elevation 4572m a.s.l remained well below freezing level till June 26 (Fig 5 2.5b) This indicates that

melt water measured later than June 26 is mainly contributed from the snowpack above 4572 m. It is also supported by the fact that a snow pit dug at an elevation 4570m a.s.l in the south edge of Rush Lake on June 18 did not show any conduit type vertical channels and continuous ice lenses were observed around the pit walls (Table 4.5) This suggest that low flow period cannot be solely attributed to a downward trend in air-temperature, but also to the migration of snow line above the freezing levels

June 27 - July 20

This was the season's highflow period Energy input was at its maximum level On average mean daily shortwave radiation remained 618 w/m^2 due to minimum cloud coverage Wind velocity remained within the range of 1-2 meters per second (Fig 5 2 5a) All these factors increased the energy input to the snowpack at higher elevations hence the melt rate By comparing table 5 2 1 and figure 5 2.5a it can be seen that streamflow started increasing when the snowpack above 4500 m a.s l started contributing to flow The significant increase in flow came when the snow line crossed the 4800 m a.s.l elevation Peak discharge occurred on July 5 with a secondary surge on July 11 (Fig 5 2.5a) The mean air temperature plot recorded at 4572 m a.s.l dropped down below freezing on July 11 and started rising on July 16, however, no corresponding rise in streamflow was observed (Fig 5 2.5b) This indicates that there was insufficient snow left on the drainage basin to be available for melt. It also suggests that knowledge of whether there is sufficient snow remaining for snow melt is needed to develop any relationship between the snow melt causative variables and runoff The period of record used in regression analysis was thus reduced to July 14 (see Chapter 5 3)

Energy system

The results indicate that the snowpack on higher elevations started contributing to surface runoff when the mean air-temperature rose above 0°C This suggests a direct variation in melt rate with temperature and hence with elevation It is true that air temperature and vapor pressure tend to decrease with increasing elevation, thus reducing convection (heat transfer through conduction) and condensation (latent heat released by condensing water vapor) melt with increasing elevation (Barry 1981) However these melt components are part of the total melt. On the other hand solar radiation, the single most important source of heat in melting snow, tends to increase with increasing elevation due to reduced scattering and absorption by the air at higher altitude A comparison of snow melt rates at different elevations over the study period indicates lower melt rates at the higher elevations A partial explanation of these apparently contradictory statements may be obtained when the variation of albedo with elevation is taken into account As a result of the greater frequency of new snowfalls at higher elevations there is an increase in the mean albedo with elevation. It is primarily a consequence of this higher albedo at high elevations that the melt is reduced rather than as a direct result of the decreased air temperature However, the US Corps of Engineers (1956) have suggested that over extreme ranges, air temperature itself also has an effect and very little melt occurs with marked freezing temperatures. In short, for the melt season as a whole, the delay in ripening of thicker snowpack, the higher albedo of higher level snow (particularly in Phahi Phari basin), and the decrease in air-temperature and vapor pressure with elevation all result in decreasing melt rates with increasing elevation

Northeast facing stream (Miar)

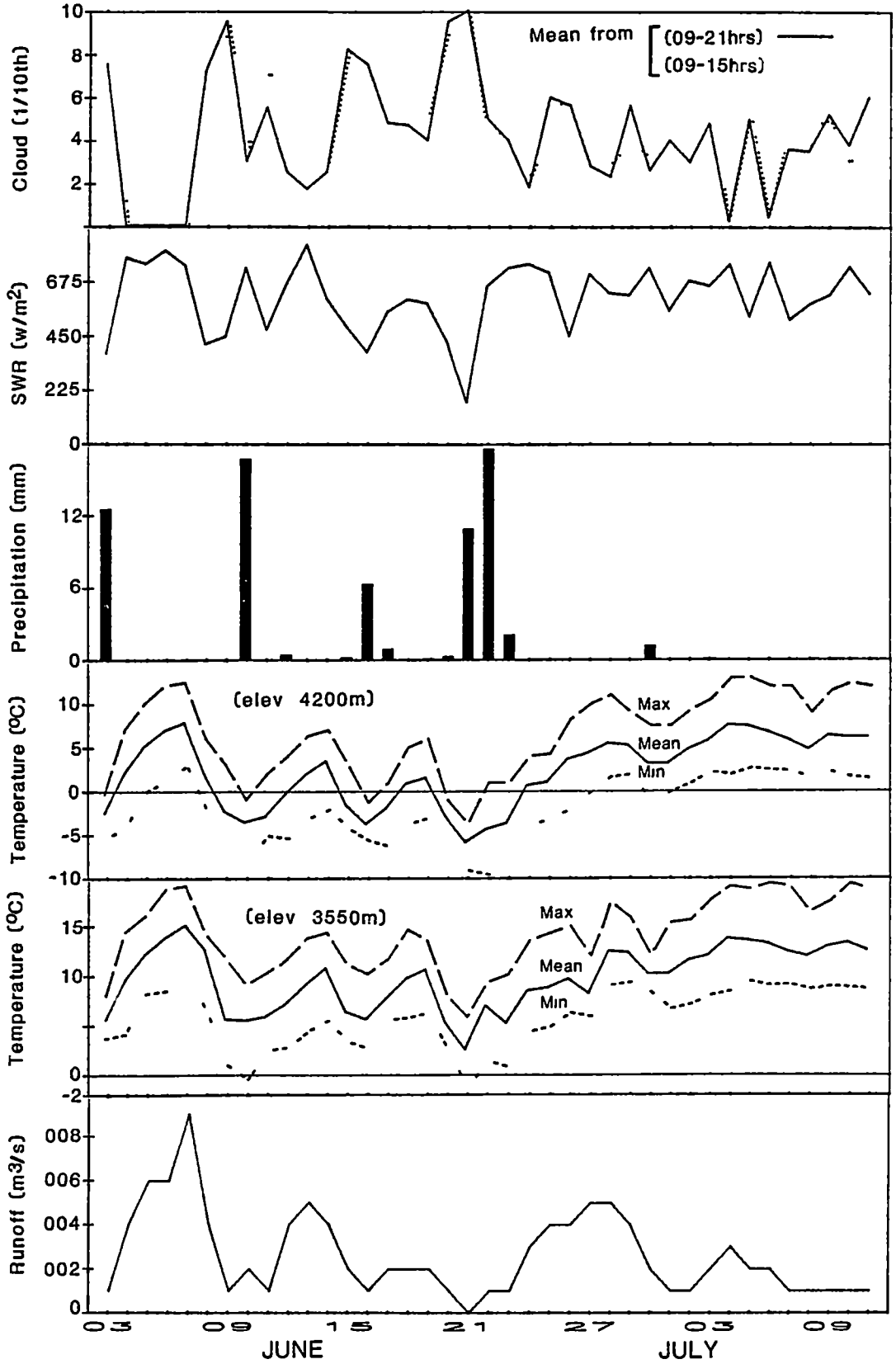
In contrast to the southwest facing basin, the northeast facing Miar stream had a peak discharge on June 7, a day when air-temperature recorded was maximum for the season and the average daily mean short wave radiation recorded between June 3 to June 7 was 674 w/m^2 due to minimum cloud cover during this period (Fig 5 2 6) This early peak flow can be attributed to the lower elevation range in comparison to the Phahi Phari basin The total area of Miar basin lies below 4267m whereas the Phahi Phari basin has two thirds of its area above this elevation Air temperatures recorded near the top of Miar basin at 4200m a.s.l show that the whole basin was under the influence of a similar temperature regime (Fig 5 2 6) thus producing a peak flow derived from snowmelt in all parts of the basin

Table 5 2 3 indicates the dates of maximum discharge, air-temperature, and radiation in the two study basins

Table 5.2.3 -Dates of maximum mean daily discharge, radiation and air temperature for the two study basins

Basin	Streamflow record	Temperature	Radiation	Discharge
Miar (northeast)	June 3 - July 11	June-7	June-13	June-7
P.Phari (southwest)	May 25 - July 20	June-7	June-25	July-5

(Source Field data S I H P 1987)



MEAN DAILY RUNOFF (MIAR BASIN)
AND RELATED CLIMATIC VARIABLES

Another distinctive feature of this seasonal flow regime is that even in the low flow period, from June 9-June 26, major snowfall events recorded on June 10, 21 and 22 did not disturb the melt pattern developed by energy input. Also, no significant rise in flow was observed after the precipitation events in the southwest facing basin, Phahi Phari However, the effect of these snowfall events was quite apparent from the hydrograph for the Miar stream and a rise in flow was observed three to four days later, after every snowfall event (Fig 5 2 6) This is mainly due to the difference in elevation of the two basins The Miar basin being lower in elevation lost most of its winter snowpack (95% of its area was free of snow on June 8) earlier than the southwest basin Each snowfall event therefore blanketed the Miar basin with thin snow cover which produced streamflow during subsequent thawing The Phahi Phari basin had all it's snowpack above 4572 a.s.l on June 8 Hence, the snowfall events added fresh snow to the existing snowpack and retarded the melting process by increasing the snow albedo Also air temperature recorded at 4572 m a.s.l remained below 0°C during lowflow period (Fig 5 2 6) Consequently no increase in flow was observed due to precipitation events until the thicker snowpack started producing meltwater due to greater energy input later than June 26 (Fig 5 2.5b)

5 2 3.2 Diurnal runoff patterns

Although daily flow patterns are not as significant an indicator of site variability in melt runoff as flow characteristics over the entire melt period, they nevertheless illustrate a number of environmental determinants The time of peak daily discharge is largely a function of the amount of solar radiation, which in turn

is dependent upon degree of slope, aspect, vegetation cover, and cloud cover conditions, the size of the individual drainage basin, it's elevation range and gradient

Northeast-southwest comparison

On the mesoscale, slope and aspect have a considerable influence on the components of the energy balance of the snowcover and subsequently on the runoff pattern In the case of some components, especially incoming short wave radiation, this influence is straight forward and relatively simple to measure On the other hand, slope and aspect also cause changes in wind velocity and other meteorological variables affecting turbulent energy transfer These changes can be very complex

It is a common observation that snow on slopes with a southern aspect melts much faster than that on a northern slopes (Anderson 1972, Meiman 1968, and Garstka 1958) This is mainly due to differences in incoming short wave radiation For example on a 10° slope at 50°N on April 1, a south facing slope receives approximately 40 percent more direct beam radiation than a north-facing slope (Male and Gray 1981)

During the spring of 1987, hourly incident short wave radiation was calculated for the middle elevation of each study basin with the help of a computer program described by Fuggle (1970) and is given in Appendix E The relation of slope and aspect to incident solar radiation is indicated in Table 5 2 4 Relative values of incident short wave radiation are presented for two aspects (NNE and SSW) and for two spring days The days are June 7 and July 5, when the sky was clear and maximum runoff was recorded for the northeast facing basin (Mair) and southwest

facing basin (Phahi Phari) respectively The values are expressed as percentages of the radiation incident on the south slope and represent percentages of maximum possible sunshine as computed from relative positions of earth and sun on the sample days.

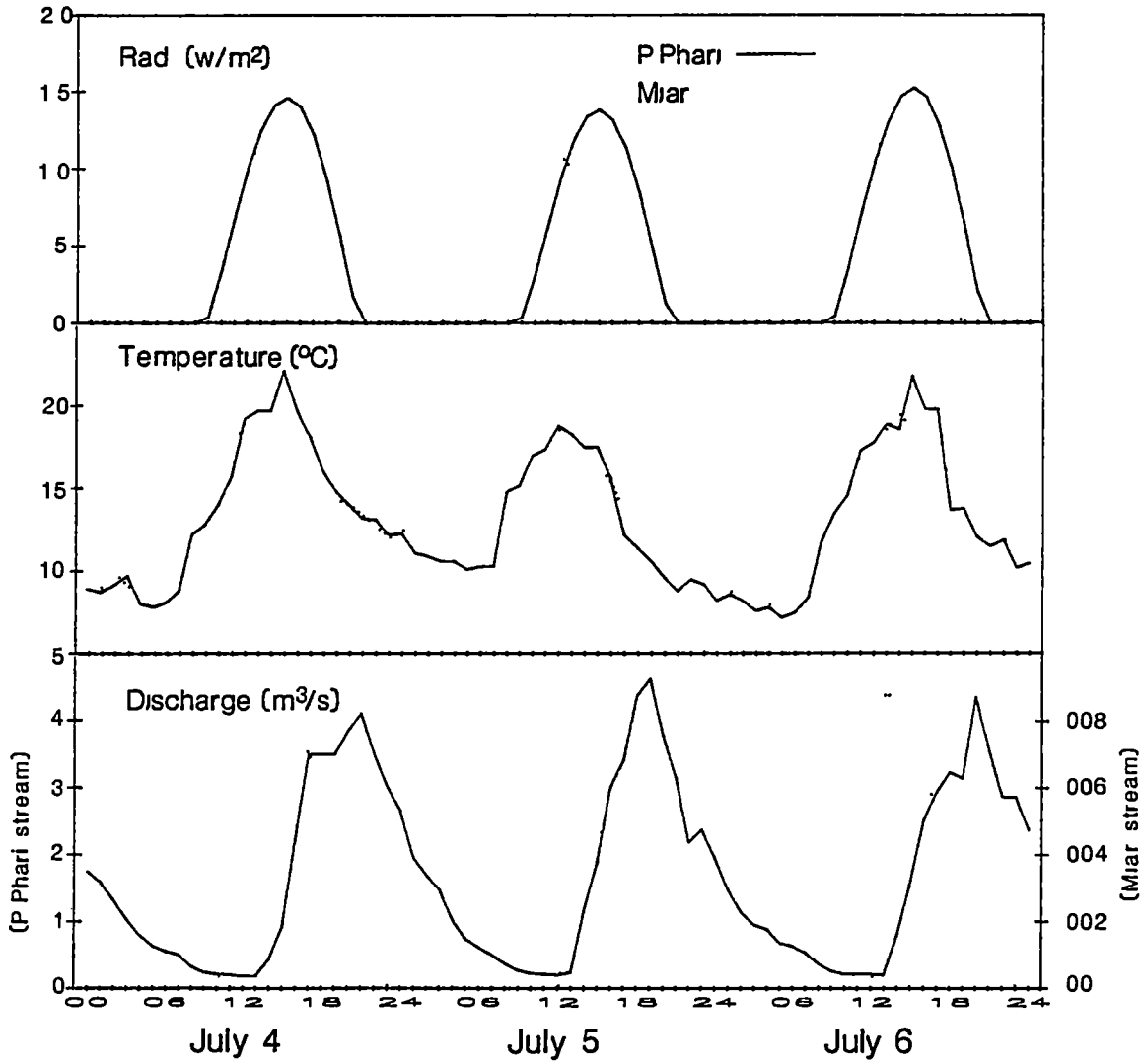
Table 5.2.4
Relative values of incident solar radiation on different aspects, 1987

	Exposure	
Date	southwest (33 6° slope) (percent)	northeast (41 3° slope) (percent)
June-7	100	78 3
July-5	100	87 9

(Source Field observations S I H P 1987)

Runoff showed prominent daily peaks in each basin during the days undisturbed by precipitation and with temperatures above the freezing point In Figure 5 2 7, comparison of diurnal variation in streamflow with radiation and air-temperature is given for the two monitoring basins for the days when no rainfall is recorded The days chosen are July 4 to July 6, when streamflow data are available for 24 hours, maximum discharge is measured on July 5 in Phahi Phari Basin The daily time range of maximum discharge, air temperature and radiation for each basin are compared below (Table 5 2.5)

FIGURE 5 2 7



HOURLY FLUCTUATIONS IN RUNOFF, RADIATION AND TEMPERATURE AT THE TWO MONITORING BASINS

Table 5.2.5 -Daily time range (hours) of maximum discharge, radiation and temperature

Month	Time in hours					
	Discharge		Temperature		Radiation	
	SW	NE	SW	NE	SW	NE
May	17-20	——	13-15	12-14	12-13	09-10
June	17-21	13-15	13-16	12-14	13-14	10-11
July	18-23	13-14	12-15	11-15	13-14	10-11

† SW (southwest facing basin), NE (northeast facing basin)

On an individual basis daily peak flow during May-June, in the stream draining the southwest facing basin was six hours later than maximum radiation and four to five hours later than maximum temperature This lag between energy input and runoff increased to seven and five hours respectively during July The increased lag time between energy input and daily peak discharge can be attributed to the greater thickness of snowpack at higher elevations and increased channel length to the gauging site later in the melt season In contrast, the northeast facing basin had its daily peak discharge three hours later than maximum radiation and one hour later than maximum recorded temperature in June and reduced to two and one hours respectively in early July

Similarly it can be seen from the Fig 5.2 7 that maximum temperature was reached one hour earlier, short-wave radiation three hours and daily peak flow about six hours earlier in northeast facing basin than southwest There is no simple way to explain this variation as it is a result of complex climatic and

environmental integration. However the most significant environmental factors that cause this local variation in melt runoff regime in the study basins area are discussed in the next paragraphs

Aspect

Whenever snow covers an area that has topographic relief, aspect is the main factor causing local variation in melt and runoff (Meiman 1970) The study basins exhibit runoff characteristics that are attributable to differences in aspect Figure 5 2 7 shows that peak flow occurred about six hours earlier in the northeast facing stream than in the southwest This is mainly due to the time difference in energy input (Table 5 2.5) Temperatures are greater on northeast facing slopes during the first half of the day as these slopes were exposed to the sun earlier than southwest slopes and lower in the second half as the former slopes remained in shadow A similar pattern is visible in the case of radiation (Fig 5 2 7) This pattern of energy input brings maximum air-temperature one hour and shortwave radiation three hours earlier on northeast facing slopes than southwest, which ultimately effect the flow regime and timing as discussed earlier

Slope

Variation in melt runoff that results directly from slope angle is difficult to isolate Indirectly, however, slope does cause a number of variations Steeper slopes have a lower storage capacity and therefore faster runoff The slope of a basin also affects the depth of soil accumulation and therefore quantity of infiltration that takes place and the rate of overland flow towards stream channels As a result,

basins with steeper slopes allow less time for infiltration and increase the surface flow The average slope from the upper divide to the basin outlet is 41 3° for Miar and 33 6° for Phahi Phari (Table 5 2 4) This slope difference helps to produce daily peak discharge six hours earlier at Miar

Elevation

The effect of altitude on a catchment is very marked in mountainous regions. The thicker snowpack generally found at higher elevations can affect the snow melt runoff in many ways. First, cooler temperatures at higher altitudes prolong the streamflow period as melting is delayed Secondly, melt water from the higher elevation snowpack has to travel longer distances to reach the gauging site Thirdly, greater snowpack depth increases the time required for melt water percolation through the snowpack thereby delaying contribution to streamflow This hysteresis is most pronounced during the early melt period This is also evidenced by the fact that the southwest facing basin, which is higher by 700 meters in elevation than the northeast facing basin and produced daily peak discharge six hours later during May-June and seven hours later in July

5 2 6 Conclusions

One preliminary conclusion of this study relates to the possibility of making short term streamflow forecasts based on the extent of snow-cover as the prime index Observations need to be made for several years for any firm relationship to be established However, the general agreement between the watershed area covered with snow and subsequent runoff indicates the possibility of estimating the

streamflow during the snow melt period on the basis of proportion of bare area The logarithmic relationship between snow-cover area and snow melt runoff, implies that snowdepth increases with increasing elevation

Patterns of snow melt runoff, examined in two contrasting environments within Barpu Glacier Basin, Central Karakoram, indicate that the volume and timing of discharge varied considerably Runoff patterns are found to be dependent upon the complex interaction of a number of factors

1 Topography influences the rate of spring snow melt in the following ways. aspect and degree of slope modifies the winter and spring snowpack by causing unequal rates of ablation, relief creates unequal distribution of snow which in turn causes areal variation in the volume of spring melt

2 Increase in the elevation range prolongs the period of seasonal flow and also shift the seasonal peak to later in the spring This is mainly due to two critical climatic factors, the increase in precipitation with elevation and decrease of air temperature with elevation

3 Weather conditions during the spring, particularly the quality of the cloud cover and daily temperatures, directly influence the rate of melt

4 Snow melt runoff is characterized by regular daily streamflow minima and maxima which reflect the daily fluctuations of solar radiation and temperature

5 Basin drainage density and snowpack thickness have a considerable affect on the timing of snow melt runoff

6 Finally, spring runoff patterns are governed by the interaction of all the foregoing physical and climate factors.

Although the above factors exert a control over melt runoff, aspect and elevation were found to be the most important Phahi Phari was free of snow up to an elevation of 4267 m a.s.l by June 10 whereas Miar retained snow up to this elevation later than July 5 This roughly indicates a time difference of three weeks in shifting snow line between the two basins.

Aspect has considerable effect on both seasonal and diurnal flow regime The daily peak from southwest aspect is delayed by six hours relative to that of the northeast On a seasonal basis, aspect prolongs the streamflow period by melting snow first, from south slopes at lower elevation and then from north slopes at lower elevation This pattern continues in different elevation bands till the complete depletion of snowcover

Elevation difference creates an unequal distribution of snow which in turn causes areal variation in the volume of spring melt It prolongs the period of seasonal flow and also shifts the seasonal peak later in the spring In the present study the southwest facing basin has a much higher elevation and produced its peak discharge on July 5 as compared to lower elevation basin (Miar), where peak flow occurred on June 7

Measurements of water yield for different periods suggest that there was a considerable increase when the zone above 4800 m a.s l started contributing to the streamflow

5 3 RELATIVE IMPORTANCE OF METEOROLOGICAL VARIABLES IN SNOW MELT RUNOFF

Since snowmelt runoff is the dominant source of streamflow in the Upper Indus Basin rivers during early spring, a thorough understanding of the relationships between meteorological variables and the snowmelt runoff is needed to improve both seasonal and short-term water yield forecasting Even small improvements in forecasting could result in better management of water resources

The meteorological factors affecting snowmelt runoff can be analysed through several approaches Statistical correlation analyses can be used to relate snowmelt runoff to its causes. Also, these relationships can be expressed in equations derived from non-statistical considerations of the observed physical phenomena of nature

5 3.1 The Statistical Approach

All hydrological phenomena are products of multiple causation Flood season discharge is associated with several antecedent variables with respect to time, which include accumulated precipitation (both snow and rain), ground water conditions, and energy components such as air-temperature, solar radiation, wind and humidity In high mountain areas, where energy availability for melting of snow is controlled by altitude, aspect and other complex microclimatic factors, the effect of each of these associated factors (independent variables) upon seasonal runoff (dependent variable) may be determined by multiple correlation By obtaining data on the independent variables, an estimate of the seasonal runoff may be made by use of the basic multiple linear regression equation

$$Y = a + b_1X_1 + b_2X_2 + b_3X_3 + \qquad + b_nX_n$$

in which

Y = estimated runoff

$X_1, X_2 \quad X_n$ = observed values of independent variables

a = constant (Y intercept of curve of the above equation)

$b_1, b_2 \quad b_n$ = regression coefficients showing relative contribution of each casual factor to the dependent variable

The forecasting problem to which the method of multiple correlation analysis is applied here concerns estimating in advance of flood season the probable runoff to be expected during the spring and early summer months from snowmelt

5 3 2 Correlation Computations

Both simple and multiple correlation analysis were computed relating runoff to the various factors causing snowmelt Correlations were performed using the method outlined by Ford (1953) The factors used in this series of analysis are described in Table 5 3 1

5 3.2 1 Effect of recognizing the recession flow

A study was made by using the Phahi Phari main meteorological station data to determine which hydrograph area of the four mentioned in Table 5 3 1 (Q_1- Q_4) would correlate best with the major causative variables The results tabulated in Table 5 3 2 show that mean daily runoff (Q_4) calculated from 1 day snowmelt

Table 5.3.1

List of variables used in statistical correlations

Identification	Description of variables	Units
Q_1	Total volume of runoff for 1 day as represented by the area under the hydrograph for one day and above a baseline of 0 flow bounded by vertical time lines at midnight	cubic cm/s
Q_2	Mean daily runoff calculated from 1 day's contribution as in Q_1	cubic cm/s
Q_3	Net volume of flow for 1 day as represented by the area under the hydrograph for one day above the recession of the preceding day	cubic cm/s
Q_4	Mean daily runoff calculated from Q_3	cubic cm/s
X_1	Mean diurnal air-temperature at 2 m height	°C
X_2	Max diurnal air-temperature at 2 m height.	°C
X_3	Min diurnal air-temperature at 2 m height.	°C
X_4	Mean diurnal air-temperature at 1.5 m height from thermohydrograph	°C
X_5	Max diurnal air-temperature at 1.5 m height from thermohydrograph	°C
X_6	Min diurnal air-temperature at 1.5 m height from thermohydrograph	°C
X_7	Relative humidity at 1.5 m height	percent
X_8	Mean day time incoming short-wave radiation 1 m above ground surface	w/m^2
X_9	Total day time incoming short-wave radiation 1 m above the ground surface	w/m^2
X_{10}	Mean daily wind travel at 2 meter height	m/s
X_{11}	Cloud cover in Barpu Basin	10ths

TABLE 5 3 2

Comparison of simple (Y = a + bX) correlation using Phahi Phari meteorological station data (facing southwest - elev 3510 m) with snowmelt-runoff (Q1-Q4) for the period May 25- July 14

Independent variables		Dependent Variables							
		(Q1)		(Q2)		(Q3)		(Q4)	
		Equation	r^2	Equation	r^2	Equation	r^2	Equation	r^2
X1	Mean temp	Y=4 183+ 140X1	461	Y=3 014+ 140X1	557	Y=4 155+ 143X1	480	Y=2 982+ 143X1	594
X2	MAX temp	Y=3 879+ 112X2	421	Y=2 715+ 111X2	506	Y=3 861+ 113X2	429	Y=2 683+ 113X2	534
X3	Min temp	Y=4 687+ 146X3	470	Y=3 523+ 145X3	558	Y=4 660+ 151X3	500	Y=3 499+ 148X3	598
X7	Humidity	Y=6 054- 012X4	081	Y=4 955- 014X4	128	Y=6 071- 013X4	087	Y=4 971- 015X4	140
X8	MeanSWR	Y=5 155+5 861E-04X5	022	Y=3 939+6 558E-04X5	034	Y=5 202+5 078E-04X5	017	Y=3 927+6 692E-04X5	036
X10	Wind	Y=6 409- 583X10	329	Y=5 066- 474X10	264	Y=6 397- 576X10	321	Y=5 078- 484X10	282
X11	Cloud	Y=5 694- 045X11	045	Y=4 562- 055X11*	083	Y=5 695- 045X11	046	Y=4 568- 058X11*	092

* All equations with temperature (max min mean) and wind are significant to 0 01 level and with humidity and cloud at 0 05

* Equations with Short wave radiation are not significant at all

* Y = log (Yi) For the description of other meteorological parameters see Table 5 3 1

* Number of Observations (N) = 50, Degree of freedom (DF) = 48

(area 1 of figure 5 3 1) gives better correlation coefficients with all the independent variables in simple regression analyses Similarly, better correlation coefficients resulted from multiple regression analyses when runoff (Q_4) was entered as a dependent variable, with seven independent variables in the equation (Table 5 3 3) Further, the result presented in Table 5 3 4 show an improved relationship through logarithmic analyses compared to straight-line regression analyses. These results indicate that the use of net volume of runoff for 1 day snowmelt as represented by the area under the hydrograph for 1 day and above the recession of the preceding day (Area 1 of figure 5 3 1) offers an improvement over the use of the flow as measured for the day from midnight to midnight and also the relationship can be better explained as a curvilinear correlation, both in simple and multiple regression analyses.

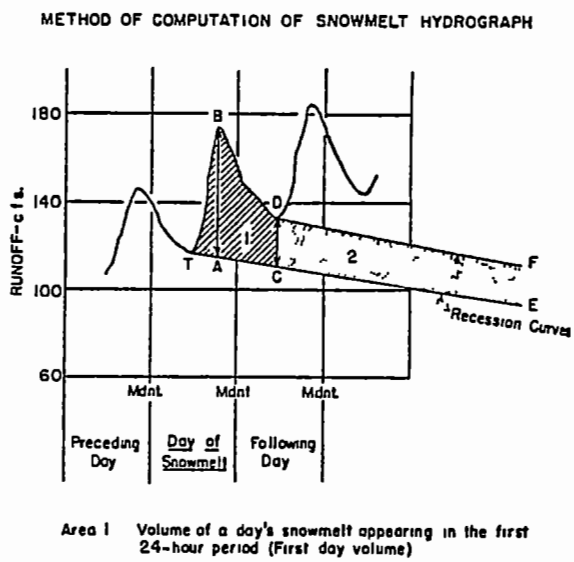


FIG 5.3 1 Snowmelt hydrograph showing area 1 = volume of day's snowmelt appearing in the first 24 hours period (Source Garstka et al 1958)

Table 5 3 3

Comparison of multiple correlation using Phahi Phari main meterological station (elev 3510 m) data with snow-melt runoff (Q1 - Q4) for the period May 25-July 14

		Independent Variables									
Eq No	Dependent variables	Mean T X1	Max T X2	Min T X3	Humidity X7	SWR X8	Wind X10	Cloud X11	a	r^2	F
P1	Q1	0 1854	0 0335	-0620	0 0075	-5 659E-04	-4334	-0075	4 2684	704	14 25
P2	Q2	0 2192	0 0024	-0609	0 0094	-7 708E-04	-3164	-0225	3 2085	736	16 77
P3	Q3	0 2074	0 0145	-0590	0 0075	-7 812E-04	-4194	-0030	4 4743	719	15 31
P4	Q4	0 2259	-0095	-0521	0 0097	-7 662E-04	-3215	-0281	3 2773	783	21 69

* Number of observations = 50, Degree of freedom = 42, F significance level 0 01

* Y = log (Y_i) For independent variables see Table 5 3 1

Table 5.3.4

Comparison between linear and log-linear regression results by using Phahi Phari main meterological station data for the period May 25- July 14

Independent variables	Dependent Variables			
	(Q_4)		log (Q_4)	
	Equation	r^2	Equation	r^2
X_1 Mean temp	$Y=-061+011X_1$	.514	$Y=2\ 982+143X_1$	.594
X_2 MAX temp	$Y=-090+009X_2$	.511	$Y=2\ 683+113X_2$	534
X_3 Min temp	$Y=-019+011X_3$	482	$Y=3\ 499+148X_3$	.598
X_7 Humidity	$Y=0\ 091-001X_7$	121	$Y=4\ 971-015X_7$	140
X_8 Mean SWR	$Y=0\ 008+5\ 623E-05X_8$	038	$Y=3\ 927+6\ 692E-04X_8$	036
X_{10} Wind	$Y=0\ 086-028X_{10}$	145	$Y=5\ 078-484X_{10}$	282
X_{11} Cloud	$Y=0\ 058-004X_{11}$	063	$Y=4.568-058X_{11}$	092

† All equations are significant to 0 01 level except equations with short wave radiation and cloud cover

† For the description of meteorological parameters see Table 5 3 1

† N = 50, DF = 48

From these results it is apparent that forecasts based on the linear regression equations could be improved by taking into consideration possible logarithmic relationships. Therefore, in view of the above results, it was not considered necessary to repeat similar correlation studies using the data from the higher site from this aspect

5 3.3 Results and Discussion

Snowmelt runoff from each of the two study basins (P.Phari & Miar) is correlated with meteorological variables collected at two different elevations. For example in Phahi Phari basin metrological observations were made on 3510 m and 4572 m a.s.l and regressed independently with stream runoff Similarly for Miar basin regression analyses are performed with data collected at 3550 m and 4200 m a.s.l Results are presented for each basin independently

5 3.3.1 Southwest

Runoff from the Phahi Phari stream draining a basin with an elevation range between 3550 m and 5030 m a.s.l., correlated with climatic variables recorded at 3510 m a.s.l and 4572 m a.s.l The lower elevation site is referred to as the Phahi Phari main metrological station while the higher station is given the name Rush Lake for the purpose of distinguishing the correlation results

Lower elevation site (P.Phari 3510 m a.s.l)

Table 5 3.5 presents the simple linear correlation coefficients between all the independent variables and snowmelt runoff used for this site The summary of

Table 5 3 5

Simple correlation matrix of meteorological factors and snowmelt runoff
Phahi pharı main met station (elev 3510 m)

Variables	Mean temp X1	Max temp X2	Min temp X3	Humidity X7	SWR X8	Wind X10	Cloud X11	Runoff Q4
Mean temp	1 000							
Max temp	0 975	1 000						
Min temp	0 945	0 890	1 000					
Humidity	- 725	- 734	- 570	1 000				
SW Rad	0 461	0 538	0 208	- 709	1 000			
Wind	- 180	- 147	- 222	- 092	0 033	1 000		
Cloud	- 563	- 619	- 376	0 833	- 820	- 019	1 000	
Runoff	0 717	0 715	0 695	- 348	0 195	- 382	- 251	1 000

* N = 50 (where N is the total number of measurements) For the description of variables see Table 5 3 1

multiple correlations using Phahi Phari main meteorological station data is given in Table 5 3 6 The variables are combined in equations having one, two, three, four, five, six and seven independent variables

The highest correlation coefficient r^2 = 0 783 resulted from the inclusion of seven independent variables in the equation To quantify the results indicated by factor analysis, stepwise linear regression was performed with all the variables, with runoff used as the dependent Equation (PP_7) resulted from stepwise analyses shows mean temperature and wind significant (α = 0 01) and humidity at the 95 percent significance level. Both of the above equations (with all variables & with three variables) explain 78% and 77% variance of the stream flow respectively

Among the equations, where only two independent variables are tested, mean daily temperature with wind produced equally good results to that where three and seven variables are in equations - Equation PP_{12} In simple correlation equations daily minimum air temperature alone gives the best correlation coefficient r^2 = 0 60 in equation PP_{14} This shows that in the absence of other factors temperature alone can be used in a predictive equation. It is also important as no other snowmelt causative variable alone gives a correlation coefficient sufficiently reliable to be used alone in the absence of air temperature However, mean temperature with the combination of other variables, especially wind, certainly improves the relationship (Table 5 3 6) The importance of wind lies in the way convective heat transfer is produced by the turbulent heat exchange between the air mass immediately above the snowpack This heat transfer is dependent on both wind and air temperature and particularly on the stability of the air mass above the snowpack (Quick 1987) Quick explains that a warm air mass above a cold snow surface tends to be stable,

Table 5 3 6

Summary of multiple correlation using data collected at an elev 3510 m with snowmelt runoff (Q4) from southwest facing stream (Phahi Phari stream basin) Period of record used from May 25 - July 14

	Independent Variables										
Eq No	Mean T X1	Max T X2	Min T X3	Humidity X7	SWR X8	Wind X10	Cloud X11	a	r^2	F	DF
PP1	0 2259	- 0095	- 0521	0 0097	-7 662E-04	- 3215	- 0281	3 2773	783	21 7	7,42
PP2	0 2026	0 0020	- 0443	0 0062	-5 274E-04	- 3310	———	3 1945	780	25 3	6 43
PP3	0 1802	- 0191	0 0034	0 0102	———	- 3156	- 0094	2 9662	774	24 6	6,43
PP4	0 1682	0 0138	- 0398	———	-8 121E-04	- 3577	- 0022	3 8158	773	24 4	6,43
PP5	0 0876	———	0461	———	———	- 3575	———	3 7953	759	48 1	3,46
PP6	0 1292	———	———	———	———	- 3691	———	3 6784	752	71 4	2,47
PP7	0 1611	———	———	0 0088	———	- 3219	———	2 9217	774	52 4	3,46
PP8	0 1429	———	———	———	-5 405E-04	- 3523	———	3 8509	771	51 5	3,46
PP9	———	———	1300	———	2 176E-04	- 3491	———	4 0092	738	43 2	3,46
PP10	———	———	1297	-9 703E-04	———	- 3496	———	4 1859	734	42 5	3,46
PP11	———	———	1280	———	———	- 3505	- 0117	4 2025	738	43 1	3,46
PP12	0 1292	———	———	———	———	- 3691	———	3 6783	752	71 4	2,47
PP13	———	———	1327	———	———	- 3444	———	4 1191	735	65 0	2,47
PP14	———	———	1488	———	———	———	———	3 4986	599	71 7	1,48
PP15	0 1427	———	———	———	———	———	———	2 9822	594	70 2	1,48

* Number of observations = 50, "DF" Degree of freedom, F significance level for all equations 0 01

* Y = log (Y_1) For description of independent variables see Table 5 3 1

resisting any downward transport of heat to the snowpack, unless there is enough wind to produce turbulent mixing In case where air temperature increases and wind remains moderate, the stability can increase to the extent where very little convective heat transfer can occur Convective heat transfer is therefore self limiting and becomes quite small at higher temperatures, unless there is a very strong wind Similarly in advective heat transfer, whether condensation occurs, releasing latent heat to the snow pack, or whether evaporation occurs, cooling the pack, depends on the relative vapor pressures of the air mass above the snow surface Wind is once again an important factor and therefore so is stability, as was discussed for convective transport (Quick 1987)

Higher elevation site (Rush Lake 4572m a.s.l.)

A simple correlation matrix of meteorological factors and snowmelt runoff is given in Table 5 3 7 and summary of simple and multiple regression analysis is given in Table 5 3 8 Short wave radiation used in this regression analysis was calculated with the help of computer programme written by Fuggle (1970) This program calculates hourly totals of incoming short wave radiation on slopes The basic input data for this program are latitude, sun declination, transmissivity, azimuth and average slope angle of the slope on which the radiation measurements are required The transmissivity factor was calculated using actual data recorded at the base of the slope (i.e Phahi Phari meteorological station elev 3510 m a.s.l) where incoming short wave radiation measurements were recorded The value of the solar constant is taken as 2 cal. cm^2 min^1 (Fuggle 1970)

Table 5 3 7

Correlation coefficient between independent variables and runoff using data from Rush Lake met station (elev 4572m)

Variables	Mean temp X1	Max temp X2	Min temp X3	Humidity X7	SWR X8	Cloud X11	Runoff Q4
Mean temp	1 000						
Max temp	0 887	1 000					
Min temp	0 951	0 766	1 000				
Humidity	- 841	- 789	- 684	1 000			
SW Rad	0 255	0 423	0 133	- 388	1 000		
Cloud	- 505	- 539	- 326	0 681	- 527	1 000	
Runoff	0 803	0 681	0 757	- 776	0 095	- 383	1 000

* N = 26 (Where N is total number of measurements) For the description of variables see Table 5 3 1

Table 5 3 8

Summary of multiple correlation using meteorological data collected at Rush Lake (elev 4572 m) with adjusted mean daily runoff (Q4) from southwest facing stream basin for the period June 19 - July 14

	Independent Variables									
Eq No	Mean T X4	Max T X5	Min T X6	Humidity X7	SWR X8	Cloud X11	a	r^2	F	DF
RH1	0 1046	- 0041	0 0664	- 0150	-3 550E-04	0 0230	5 8329	771	10 6	6,19
RH2	0 1916	———	———	- 0096	-3 943E-04	0 0296	5 3528	766	17 1	4,21
RH3	0 0852	———	0 0798	- 0150	———	0 0332	5 5906	762	16 9	4,21
RH4	0 0712	———	0 0898	- 0133	-4 562E-04	———	5 9688	766	17 2	4,21
RH5	0 2001	———	0 0152	———	-3 552E-04	0 0142	4 9936	760	16 6	4,21
RH6	0 0476	———	0 1122	- 0113	———	———	5 6491	753	22 4	3,22
RH7	0 1901	———	0 0242	———	———	0 0248	4 7505	753	22 4	3,22
RH8	O 1762	———	0 0323	———	-4 145E-04	———	5 1417	759	23 0	3 22
RH9	0 2177	———	———	———	-3 644E-04	0 0173	4 9385	760	23 2	3,22
RH10	0 1951	———	———	- 0084	———	0 0419	4 9948	757	22 8	3,22
RH11	0 1963	———	———	- 0045	-5 268E-04	———	5 3159	759	23 0	3,22
RH12	0 2108	———	———	———	-4 761E-04	———	5 0775	757	35 8	2,23
RH13	0 2181	———	———	———	———	0 0302	4 6513	752	34 9	2,23
RH14	0 2050	———	———	6 628E-04	———	———	4 7358	740	32 7	2,23
RH15	0 2027	———	———	———	———	———	4 7664	740	38 8	1,24
RH16	———	———	0 1826	———	———	———	5 3382	714	59 9	1,24

* Y = log (Y_1) For the description of climatic parameters see Table 5 3 1 Number of observations = 26
* All equations are significant to 0 01 (F distribution) "DF" degree of freedom

Equation RH_1, including six independent variables (Table 5 3 8) shows the highest correlation coefficient r^2= 0 771 It is worthwhile to mention here that the independent variable wind, which was one hundred percent significant (t- student test) in most of the multiple correlation equations computed by using the lower elevation data, is not available for the higher elevation site The physical phenomena and importance of wind were discussed above

Equation RH_2 including mean air-temperature, humidity, short wave radiation and cloud cover gave a correlation coefficient r^2= 766 Similarly equations RH_9 and RH_{12} with three and two variables (Table 5 3 8) give high correlation coefficients as with all the six variables in the equation In simple correlation analysis mean daily air-temperature explains 74 % variance of the dependent variable This is as good as that explained by all the variables in the equation r^2= 0 77 In stepwise regression analyses mean temperature is the only variable which stand significant (α= 0 01) level

Relation between temperature profile (Lower and Higher sites)

Tables 5 3 6 and 5 3 8 indicate that in simple correlations, minimum temperature produces the highest correlation coefficient at the lower elevation site whereas the mean air-temperature is best correlated from the higher elevation site To determine the relationship between these two temperature profiles, correlation analysis was computed using temperature data from June 19 to July 14 Figure 5 3 2 shows the relation of mean daily temperature from the higher elevation site (4572 m a.s.l) with minimum temperature from the lower elevation site (3510m a.s l)

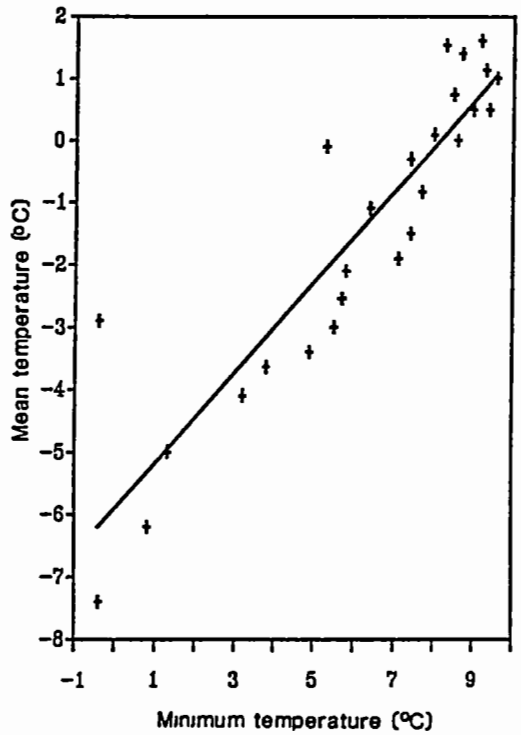


FIG 5 3 2 Relation between minimum and mean air-temperature recorded at elev 3510m and 4572m a.s.l respectively on SSW facing slopes

The correlation coefficient resulted from analysis is

r^2 = 0 821, for the equation

Y = -5 9047 + 0 7249X

where Y = Mean temperature from higher elevation site

X = Minimum air-temperature from lower elevation site

The results from this equation suggest that minimum temperatures recorded at valley bottom slope (southwest) can potentially be used if no temperature data are

available from the higher elevations Similarly, mean temperature is the single most important variable for higher elevation in the present case for mid basin elevation.

5 3.3.2 Northeast

Both simple and multiple statistical correlation analysis were computed relating snowmelt runoff to the various factors recorded at 3550 m and 4150 m a.s l separately The stream flow was strictly diurnal and ceases to flow during the night time This situation restricted the analysis to dependent variables Q_1 & Q_2 (Table 5 3 1) Period of record used in analysis is from June 3 to June 30, 1987 (28 days) as there was hardly any snow left in the basin later than June 30 (Table 5.2 1)

To determine which of two discharge (daily total or daily mean) gives better results using correlation analysis, a study was made by using the independent variables collected at a higher elevation (4200 m a.s.l.) named as Miar Hill Results are presented in Table 5 3 9 This shows that the use of mean daily runoff is a definite improvement over the use of daily total, as recorded at the southwest monitoring basin In view of this, all the simple and multiple correlation analysis were computed relating mean daily runoff to its causes.

<u>Lower elevation site (Miar 3550m a.s.l)</u>

The coefficient of correlation r between independent variables and stream flow is given in Table 5 3 10 and results of simple and multiple correlation analysis are given in Table 5 3 11

Table 5.3.9

Comparison of correlations using total one day runoff (Q_1) with correlations using mean daily runoff (Q_2) from northeast facing stream basin (Miar) with meteorological data collected at an elevation 4200 m a.s.l (Miar hill) for the period June 3-30, 1987

Independent variables	Dependent Variables			
	Q_1		Q_2	
	Equation	r^2	Equation	r^2
X_1 Mean temp	Y=4 697+ 0734X_1	717	Y=3 343+ 0743X_1	789
X_2 MAX temp	Y=0 025+ 0093X_2	699	Y=0 001+4 150E-04X_2	729
X_3 Min temp	Y=0 109+ 0114X_3	582	Y=0 005+5 198E-04X_3	638
X_7 Humidity	Y=0 184- 0018X_7	645	Y=0 008-8 068E-05X_7	667
X_8 Mean SWR	Y=0 111-7 306E-05X_8	021	Y=0 005-3 847E-06X_8	031
X_{11} Cloud	Y=0 120- 011X_{11}	489	Y=0 005-4 851X_{11}	468

† Y = LOG (Y1) N = 27, DF = 25 For description of variables see Table 5 3 1

Table 5 3 10

Pearson Correlation Coefficients for Miar metrological station data (3550m a.s l)

Variables	Mean temp X1	Max temp X2	Min temp X3	SW Rad X8	Cloud X11	Runoff Q2
Mean temp	1 000					
Max temp	0 937	1 000				
Min temp	0 918	0 849	1 000			
SW Rad	0 285	0 309	0 269	1 000		
Cloud	- 641	- 652	- 520	- 300	1 000	
runoff	0 835	0 851	0 737	0 069	- 716	1 000

* N = 28 (Where N is the total number of measurements) For the description of variables see Table 5 3 1 1

Table 5 3 11

Summary of multiple correlation using metrological data collected at an elevation 3550m a.s l with snowmelt runoff (Q2) from northeast facing stream basin (Miar) Period of record used from June 3 - 30, 1987

	Independent Variables								
Eq No	MeanT X1	Max T X2	Min T X3	SWR X8	Cloud X11	a	r^2	F	DF
MR1	1 533E-04	3 288E-04	-3 108E-06	-4 188E-06	-2 139E-04	6 904E-04	83	21 5	5,22
MR2	———	4 026E-04	6 813E-05	-4 252E-06	-2 268E-04	8 257E-04	83	27 4	4,23
MR3	———	4 583E-04	———	-4 229E-06	-2 226E-04	4 125E-04	82	37 5	3,24
MR4	4 525E-04	———	———	-3 883E-06	-2 429E-04	0 0022	80	31 8	20 3
MR5	2 919E-04	3 974E-04	-9 249E-05	-3 583E-06	———	- 0022	78	20 3	4,23
MR6	2 024E-04	4 050E-04	———	-3 610E-06	———	- 0020	78	27 8	3,24
MR7	1 627E-04	2 872E-04	———	———	-1 845E-04	- 0011	78	27 7	3,24
MR8	———	———	3 708E-04	-3 842E-06	-3 483E-06	0 0048E-04	74	22 8	2,25
MR9	———	4 270E-04	———	———	-1 933E-04	- 0015	77	41 6	2,25
MR10	———	5 857E-04	———	-3 632E-06	———	- 0025	77	40 7	2,25
MR11	4 283E-04	———	———	———	-2 116E-04	3 304E-04	75	38 0	2,25
MR12	5 952E-04	———	———	-3 121E-06	———	-4 442E-04	73	33 4	2,25
MR13	———	5 435E-04	———	———	———	- 0038	72	68 2	1,26

* N = 28, F = F distribution, DF = Degree of freedom All equations are significant to 0 01 level (F distribution)

* For the description of independent variables see Table 5 3 1

Although the best correlation coefficient is produced by equation MR_1 where all five independent variables are included, an equally good result is achieved by incorporating three variables in equation MR_3 (Table 5 3 11) With two variables in the equation i e equations MR_8 to MR_{12}, the maximum temperature with cloud cover gave the best correlation coefficient r^2= 769 Among the simple correlation analysis daily maximum air-temperature gave the highest coefficient r^2= 724 The equations MR_1 (with all variables) and MR_{13} (with three variables) which explain the variance of the dependent variable (mean daily snowmelt runoff) 83% and 72% respectively Also in stepwise analyses, maximum temperature with cloud gave the best correlation coefficient r^2= 769 in equation MR_9 (Table 5 3 11) These results indicate that the maximum temperature alone can be used in a predictive model if only one variable is available for prediction

Higher elevation site (Miar Hill 4200 m a.s l)

Six independent variables were used for this site with the addition of humidity which was not available for the lower elevation station (Table 5 3 12) The inclusion of humidity resulted in higher correlation coefficients The variables are combined in equations having six, four, three, two and one independent variable and the results are tabulated in Table 5 3 13

Again the best correlation coefficient is achieved with all six variables in the equation Equally good results are obtained with the three variables mean temperature, minimum temperature and short wave radiation in the equation (MH_2) With two variables in the equation, the best results are obtained by combining mean daily temperature and cloud cover equation (MH_9)

Table 5 3 12

Coefficient of correlation R between independent variables and runoff
Miar Hill metrological station data (4200m a.s l)

Variables	Mean temp X1	Max temp X2	Min temp X3	Humidity X7	SWR X8	Cloud X11	Runoff Q4
Mean temp	1 000						
Max temp	0 974	1 000					
Min temp	0 956	0 918	1 000				
Humidity	- 855	- 856	- 746	1 000			
SW Rad	0 226	0 290	0 220	- 255	1 000		
Cloud	- 664	- 686	- 506	0 846	- 230	1 000	
Runoff	0 891	0 866	0 818	- 831	0 069	- 716	1 000

* N = 28 (Where N is the total number of measurements) For the description of variables see Table 5 3 1

5 3 13

Summary of multiple correlation using meteorological data collected at an elevation 4200m a.s l with snowmelt runoff (Q2) from northeast facing stream basin (Miar) Period of record used from June 3 - 30, 1987

	Independent Variables									
Eq No	Mean T X4	Max T X5	Min T X6	Humidity X7	SWR X8	Cloud X11	a	r^2	F	DF
MH1	0 1160	- 0135	- 0475	- 0018	-4 301E-04	0 0084	3 5921	86	20 4	6,20
MH2	0 1285	———	- 0657	———	-4 068E-04	———	3 2989	84	43 2	3,23
MH3	0 0614	———	———	———	-4 673E-04	- 0262	3 7113	84	40 9	3,23
MH4	0 1098	———	- 0496	———	———	- 0112	3 2002	84	39 6	3,23
MH5	0 1126	———	- 0563	- 0017	———	———	3 2326	84	39 2	3,23
MH6	0 0509	———	———	- 0049	-5 144E-04	———	3 9313	83	38 7	3,23
MH7	0 0602	———	———	-7 808E-04	———	- 0212	3 4934	82	35 4	3 23
MH8	0 1282	———	- 0653	———	———	———	3 0866	83	60 0	2,24
MH9	0 0625	———	———	———	———	- 0241	3 4557	82	55 2	2,24
MH10	0 0551	———	———	- 0040	———	———	3 6043	81	51 3	2,24
MH11	0 0743	———	———	———	-3 987E-04	———	3 5532	80	49 2	2,24
MH12	0 0743	———	———	———	———	———	3 3433	79	93 4	1,25

* N = 27, "DF" degree of freedom For the description of independent variables see Table 5 3 1

* Y = LOG (Yi) All equations are significant to 0 01 level (F distribution)

Among the individual variables, mean daily air-temperature proves the best having $r^2 = 0.789$ for MH_{12} However, inclusion of any other variable results in a correlation coefficient which is more significant. In stepwise analyses mean and minimum temperature qualify to enter in the test and produced a correlation coefficient of 0.834 (eq MH_8 in Table 5.3.13)

Relation between temperature profile (Lower and Higher site)

Tables 5.3.11 and 5.3.13 suggest that the maximum temperature from the lower elevation site and mean air-temperature from the higher elevation site are the best among the individual parameters explaining 72 % and 79 % variation in stream flow respectively To determine the relationship between these temperature profiles, a study was conducted using temperature data from May 25 to August 19, 1987 The correlation coefficient for simple correlation analysis is

r^2 = 0.903, for the equation

$$Y = -12.246 + 1.04449X$$

where Y = Mean air-temperature from higher elevation site

X = Maximum air temperature from lower elevation site

The relationship between the mean and maximum air-temperatures is shown in figure 5.3.3

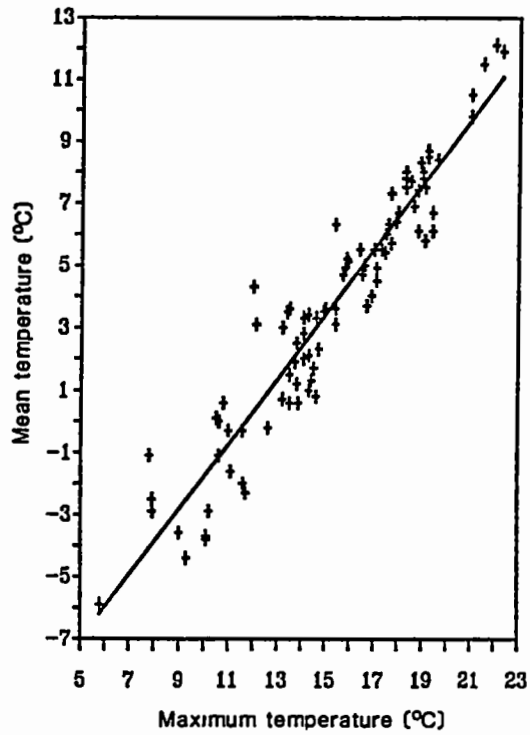


FIG 5 3 3 Relation between maximum and mean air-temperature recorded on NNE facing slope at elevation 3550m and 4200m a.s.l respectively

In general, the greater number of snowmelt causative variables in the equation the better the correlation coefficient is But apart from this, the results from the statistical analyses of these factors lead to the conclusion that temperature factor is at least as good as, and in many cases better than, a combination of other factors used in correlation analyses Therefore, in the development of practical applications of methods of forecasting runoff from snowmelt, particular attention should be paid to the temperature variable However, it should be kept in mind that the independent variables must be considered only as indices showing the effect of conditions where data were collected, since the variables are intercorrelated

5.3.4 Comparison of Results in Terms of Aspect

Results tabulated in Table 5.3.14 indicate that mean air temperature is the best index of streamflow, when it is recorded in the middle of the basin. The physical explanation of this phenomena is that mean temperature calculated from hourly measurements does take into account the effect of diurnal freezing of snowmelt water (US Corps of Engineers 1956). A similar view is presented by Kuusisto (1984), where he used mean daily temperatures (taking into account the negative temperatures) to calculate degree days as an index of snowmelt.

From the lower elevation stations minimum temperature gave a better correlation coefficient for the southwest aspect in curvilinear regression analysis. For the northeast aspect maximum temperature produced a higher correlation coefficient than any other variable. Relative humidity recorded at higher elevations on both aspects produced much better correlation coefficients when regressed against runoff as compared to the lower elevation stations. Incoming short wave radiation is better related to streamflow from the southwest than northeast facing slopes. This may be due to the fact that for all the other three stations, radiation was calculated with the help of a computer program which does not represent the true environmental effect. Similarly, cloud cover shows better correlation for the northeast facing basin than southwest, whereas radiation is better correlated with stream flow from the southwest facing basin.

Table 5.3.14

Comparison of correlation coefficients of simple regression with streamflow in terms of aspect.

Independent variables	Southwest		Northeast	
	Lower elev (3510m)	Higher elev (4572m)	Lower elev (3550m)	Higher elev (4200m)
Mean temp	0.717	0.803	0.835	0.891
Max temp	0.715	0.681	0.851	0.866
Min temp	0.695	0.757	0.737	0.818
Humidity	-.348	-.776	*****	-.831
Mean SWR	0.195	0.095	0.069	0.069
Wind	-.382	*****	*****	*****
Cloud	-.251	-.383	-.716	-.716

***** Variable not available for the analysis

† Southwest
Lower site (May 25 - July 14)
Higher site (June 19 - July 14)

† Northeast
Lower site (June 3- June 30)
Higher site (June 3- June 30)

5.3.5 Comparison of Observed and Predicted Hydrographs

Figure 5.3.4 shows the plot of best predicted equations for both monitoring streams. The solid-line hydrographs are the observed flow at the gauging site for each basin. The dashed line hydrographs represent the forecasted runoff using equation PP_1 (Table 5.3.6) with mean, maximum and minimum temperature, relative humidity, short wave radiation, wind, and cloud cover in the equation MH_1 (Table 5.3.13) with same variables except wind speed for Phahi, Phari and Miar stream respectively.

The dotted line hydrographs are the forecasted runoff derived by employing minimum and mean daily temperatures for Phahi, Phari and Miar streams respectively. Figure 5.3.4 also shows that the best fit forecast was obtained by using equations PP_1 and MH_1 as compared to the forecast based on air-temperatures only. These results suggest that although air-temperature is the best indicator in simple regression models, the forecast can be improved with the inclusion of other snowmelt causatives variables in analyses.

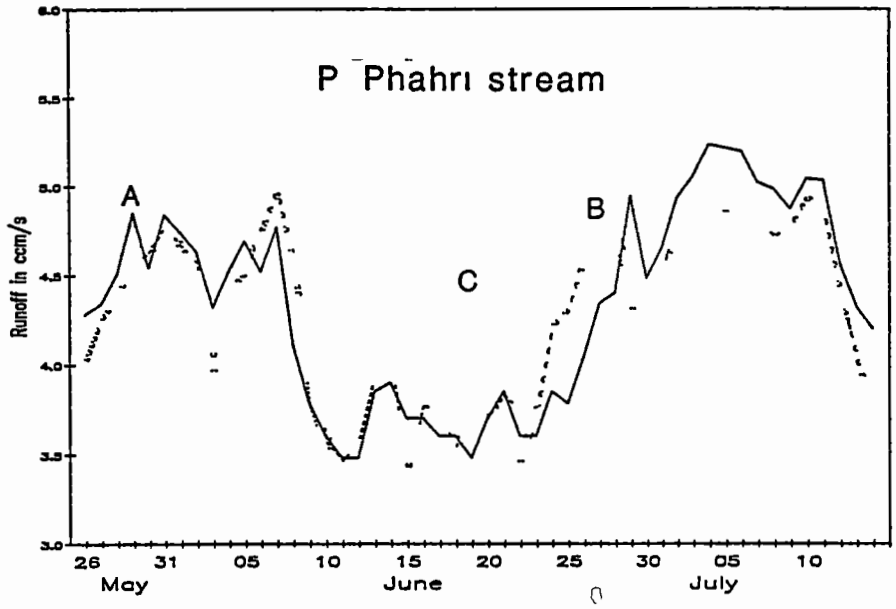
5.3.6 Discussion of Analytical Methods

There are two different prerequisites for judging the value of different methods for maximizing information about hydrologic problems. One is to get an approximate working tool, e.g., a prediction formula. The other is to acquire knowledge about the physical laws underlying the hydrologic phenomena.

In the first case, it can be concluded that the regression technique was successful. The formulas developed were able to explain most of the variation in

COMPARISON OF OBSERVED AND PREDICTED HYDROGRAPHS

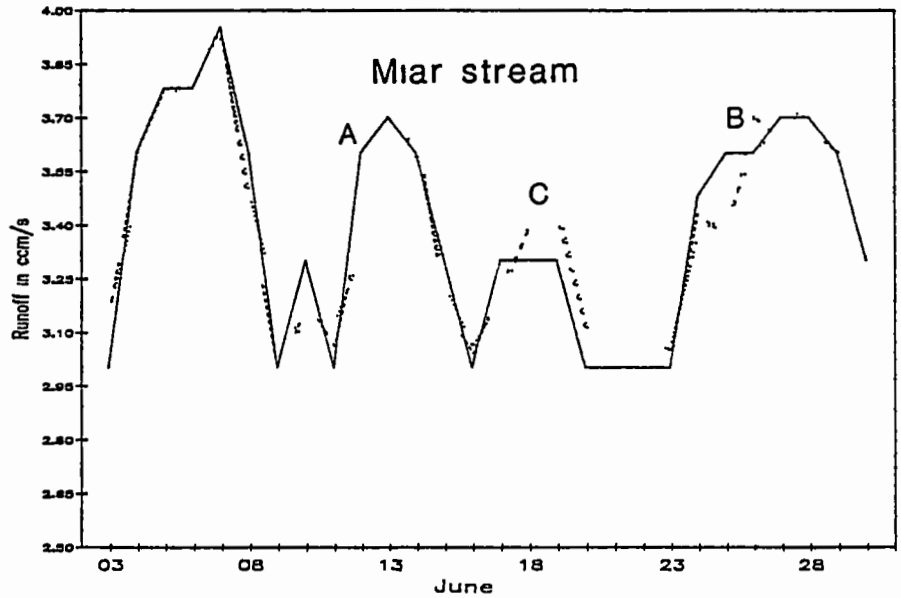
Snowmelt Season 1987



A ———— Observed hydrograph

B -------- By using equation PP1 (Table 5 3 6)

C By using equation PP11 (Table 5,3 6)



A ———— Observed hydrograph

B -------- By using equation MH1 (Table 5 1 13)

C By using equation MH13 (Table 5 1 13)

stream flow and can be used as prediction formulas under similar conditions that do not differ widely from those of the original data. However, it is very difficult to interpret the hydrological meaning and functional form of such indices Because these indices include the effects of various real but unmeasured climatic and hydrological factors, they are intercorrelated with each other These intercorrelations make the interpretation of the effect of the variable more difficult (Mustonen 1967) However, in hydrology problems like this one, to look for and explain the true individual effects of available climatic variables is difficult For instance, air temperature may have its own true strong effect, but it is so inextricably bound up with other measured as well as unmeasured factors, that it is too complex to compute and explain its real effect uninfluenced by any other factor (Mustonen 1967)

The true hydrologic law of stream flow is a very complicated combination of variables The variables used in this study are only indices of the effect of the true hydrologic factors Normal linear and curvilinear multiple regression analysis is a useful statistical method for increasing our understanding of many hydrologic problems

5 3.7 Conclusions

Normal linear multiple regression analysis is an appropriate method to study hydrological relationships Intercorrelations between independent variables do not invalidate prediction models when they are used under the same conditions as those under which the data were obtained Simple correlation coefficients from all the four sites indicate that if only one meteorological variable is available for snowmelt

runoff prediction, mean air temperature is the best predictor

The models developed in this study serve as a starting point for further attempts to explain important relationships in the hydrology of cold regions like that of the Karakoram Mountains.

CHAPTER 6

Summary and Conclusion

6 1 INTRODUCTION

Both in project planning studies and in the operation of water resource utilization projects, there is a need for the refinement of techniques to compute seasonal water-yield forecasts and estimate momentary seasonal peak discharge and daily streamflows The present study is part of an investigations, initiated by the "Snow and Ice Hydrology Project"[1] in the Karakoram Mountains since 1985 The purpose of this project is to develop and test methods of monitoring and forecasting volume and rates of runoff from snow and ice melt in the rivers basins of the Upper Indus Basin

This thesis has examined specific factors causing variation in snowmelt runoff, and the relative importance of hydrometeorological variables in prediction models As the study was based on field observations for only one summer season, it should not be considered adequate to establish firm long-term relationships but is a pilot study towards that goal It does indicate the requirements and problems of establishing reliable hydrological investigations.

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1 Snow and Ice Hydrology Project established in 1985 at Wilfrid Laurier University, Waterloo, Ontario, Canada

The results obtained provide useful information with regard to stream flow behaviour in drainage basins in the crucial elevation range between 3500 to 5030 m a.s.l in two monitoring basins with opposing slope aspect This study provides information about the zone of maximum accumulation and timing of meltwater release from this zone Formulae were also developed to evaluate the relative importance of hydrometeorological variables for predictive models. Section 6 2 summarizes the results achieved and some important interpretations Section 6 3 is devoted to concluding remarks and 6 4 to some recommendations which may be useful for the development of future monitoring programmes.

6 2 RESULTS SUMMARY

6 2 1 Variation in Runoff

Mountain streams exhibit strong and frequent fluctuations in flow regime as a result of variable conditions in their source areas Thus, on days with no cloud cover, runoff from a snow covered area depends on the distribution of snow and radiant heat supply to the basin On wet days, the entire basin receives precipitation input, but at higher elevation above freezing level precipitation may accumulate as snow storage instead of producing direct surface runoff (Woo 1972) If there is melting it depends largely upon sensible heat or the ambient air temperatures and advective heat The shape and dimensions of the hydrograph are controlled by the interaction of climatological and physical characteristics of the drainage basin The variation observed in hydrographs of the two monitoring basins are analysed with respect to aspect and elevation

6 2 1 1 Seasonal variation

Seasonal variation in runoff regime observed from the experimental basins suggests.

1 that elevation has considerable effect on the seasonal flow regime Increasing elevation range prolong the period of seasonal flow and also the seasonal peak to later in the spring This is mainly due to two critical climatic factors, the increase in precipitation with elevation and decrease of temperature with elevation. In the present case the southwest facing basin (P.Phari) which has a higher elevation, produced its peak discharge on July 5 as compared to the lower elevation basin (Miar) where peak discharge was observed on June 7, a difference of almost one month,

2 that in a year such as the pilot study year, deep and extensive late winter and early spring snow cover above an elevation 4600 m a.s.l is maintained by cloudy and cool weather well into summer, in this case to late June (Table 5 2 1),

3 that if the elevation of a basin spans the zone of maximum snow accumulation, that is 4800 m to 5200 m a.s l (see Hewitt 1986, Wake 1987), the seasonal peak will be shifted towards the time when the upward migration of temperature belts culminates, in this case mid to late July (Fig 5 2.5b),

4 that if the elevation of a basin increases above 4800 m a.s.l with moderate decrease in area, the increase in runoff volume will be many fold when the area above 4800 m starts contributing to streamflow Runoff measured for P.Phari stream shows that 38% of the total area, located above 4800 m a.s.l., contributed more than 70% of the streamflow after June 26 This is due to two reasons, first, precipitation increases with elevation and second, a higher portion of this basin is a relatively flat area surrounded by the 4877 m contour and seems to work as snow trap This is the elevation at which the zone of maximum snowpack was observed to begin by Wake (1987), in the accumulation area of the Biafo Glacier, Central Karakoram,

5 that a small change in snow cover area in Phahi Phari stream basin during June 29 to July 6 produced maximum daily water yield (Table 5 2 2a) This indicates much deeper snowpack above 4800 m a.s l However, there is always a point when the contributing area of a basin above a certain elevation will become too small for increased water

equivalent in the snowpack to compensate for it Then, runoff declines despite the progressively deeper snowpacks,

6 that precipitation events also cause variations in streamflow in the following ways

(a) they add fresh snow to the existing snow cover at higher elevations. That increases snowpack albedo and decreases the radiant energy input,

(b) they increase areal variability of snow accumulation by adding more snow to the existing snow cover at higher elevations The only quantitative measurements that exist in this regard is that of Wake (1987) which shows 30 - 50 percent of total snow accumulation can occur during the summer period This rate of snow accumulation at higher elevations implies that summer snowfall in glacierized basins substantially decreases the rate of melting for short periods of time (Young 1977, Collins 1982) In the present example, period of low flow was observed from June 9 to June 26 due to extensive cloud cover and precipitation events in P.Phari Basin,

7 finally, basins having lower elevations than those of the zone of maximum accumulation, that is lower than 4800 m a.s l, will tend to have their peak discharge proportionally earlier as in the case of Miar stream basin in early June (Fig 5 2 6) Obviously, they will also have a proportionally smaller net yield of water

6 2 1 2 Diurnal variation

Diurnal runoff cycles were pronounced throughout the study period in both streams They roughly follow the pattern of energy input However, timing of the peak discharges was different due essentially to differences in aspect, slope, elevation, and drainage density (Fig 5 2 7) Results described in section 5 2 4 2 suggest that while the northeast facing slope received less radiant energy as compared to the southwest facing slopes (Table 5 2 5), its maximum receipt was reached earlier both for solar radiation and air-temperature than in the case of the southwest site (Fig

5.2 7) The pattern of energy input, as well as steeper slope, reduced elevation range and shorter stream channel lengths, brought peak discharge, on average, six hours earlier than on the southwest facing basin It is difficult to differentiate the particular role of each of these factors, but elevation and aspect are the dominant controlling factors on snowmelt and snowline retreat The higher elevation of the southwest facing basin, its deeper snowpack was preserved much longer than that of the northeast basin As such the melt period of the southwest basin extended past that of the northeast basin to mid July The relatively deeper snowpack affects the streamflow pattern through storage Deep packs with prominent ice lenses have a large storage capacity and thus increase the retarding of flow (Woo and Slaymaker 1975) The increased lag between energy input and daily peak discharge during July in P.Phari can be attributed to the observed thicker snowpack above an elevation 4572 m a.s.l on June 18 (see Table 4.5)

6 2 2 Runoff Indexes

In hydrologic practice an index is either a meteorological or hydrologic variable whose variations are associated with those of the element it serves to estimate and which is more readily measured than the element itself (US CORPS of Engineers 1956)

The study suggests that mountain stream flow depends greatly on the thickness and distribution of snow cover Therefore it is important to know the variations of this value in time and space These are a function of the irregularity of snow deposition and melt regime due to the mountainous microclimate

6 2.2.1 Snow-cover area vs snowmelt runoff relation

The areal variability of snowpack has long been recognised as a primary hydrologic parameter related both to the average snowpack water-equivalent and to the snowmelt derived runoff (Rango and Salomonson 1977) Knowledge about the distribution of this primary factor over time and space is essential for any type of hydrologic study It may be of little importance in flat areas with uniform snow cover, but in mountainous areas it is most important as it plays a dominant role That seems especially true in the Karakoram Mountains where there are very significant increases in snow depth with elevation (Hewitt, 1985)

The rate at which the snow cover depletes is an index which is inversely related to the generated snowmelt runoff As the snow leaves the lower elevations of the watershed, the hydrograph begins to rise and continues to do so until the snowpack area reaches a critical value where meteorological snowmelt conditions cannot produce ever increasing amounts of snowmelt runoff The hydrograph then begins to recede until the remaining annual snowpack disappears The slower the snow line retreats up the watershed to the elevation where the hydrograph starts a downward trend, the greater the resulting runoff volume and usually peak flow

Although not a recent innovation, the use of aerial photography to estimate snow cover has become more important since the advent of satellite photography Many studies have been made attempting to forecast meltwater yield from snow-covered area measured on satellite images using regression equations (Rango et al. 1977, Tarar 1982, Gupta et al. 1982, Dey et. al 1983, Makhdoom and Solomon 1986) This purely statistical approach has proved surprisingly accurate in

forecasting, but only after a multi-year calibration period for every new catchment of interest

Rango and Solomonson (1975), used LANDSAT - I imagery to determine snow cover over the Upper Indus Basin of Pakistan, and the Wind River Mountains of Wyoming Using this data and seasonal streamflow, they developed regression equations that produced high correlation coefficients between snow cover area and seasonal runoff Similarly they achieved good results in 1977 by using the snow cover area for the Indus River above Besham and Kabul River above Nowshera (Rango and Solomonson 1977) On the basis of these results, they concluded that the snow covered area obtained from meteorological satellites over remote regions is significantly related to seasonal streamflow in regression analysis for the Indus above Besham, and Kabul River above Nowshera in Pakistan However, they suggest that successful application of these methods will depend highly upon "accurate ground truth data for calibration and verification"

The results obtained in section 5 2 3 also suggest that snow cover can be related to streamflow using regression analysis. However, accuracy of the results is largely dependent upon measurements of snow depth, water equivalent, and knowledge of meltwater release pattern with respect to elevation and slope aspect.

6 2 2 2 Correlation

The study has shown that multiple correlation is applicable to a variety of hydrologic problems. This example is intended to show the advantage of basing a forecast on the effects of several variables rather than on single index and to show

also the ultimate possibilities in approaches based on multiple relations.

It also shows that in general, the more causative variables employed, the better the correlation coefficient is. But apart from this general view, results obtained through equations PP_7 (mean temperature, humidity and wind), RH_9 (mean temperature, short wave radiation, cloud), MR_3 (maximum temperature, short wave radiation, cloud), and MH_3 (mean temperature, maximum temperature, short wave radiation) show that equally good results can be obtained by incorporating three variables among wind speed, radiation, humidity, and air temperature Wind speed is an important factor in turbulent energy transfer (Quick 1987, Obled and Harder 1979, U S Corps of Engineers 1956) Some of the simplified snowmelt equations involving wind and temperature were reported by U.S Corps of Engineers (1956) and Kuusisto (1984)

Another effective and widely used index of snowmelt runoff is air-temperature Although the sun is the main source of energy for melting of snow on high mountains, the exact manner in which solar radiation becomes available and active can be, at times, very intricate (Garstka 1958) Miller (1950) concluded that, during the melting hours, most of the heat applied to the snow came from solar radiation, but that as much as 300 calories per square centimeter per day from the insulation went to heat the air This is mainly due to the higher albedo of snow

6 3 CONCLUSIONS

Snow distribution with elevation

The exceptionally high and rugged mountain environment of the Karakoram affects snowfall occurrence Precipitation inputs vary greatly in quantity and regime with altitude, topography and aspect.

During this study it is observed that snowpacks at elevations lower than 4800 m a.s.l did not contribute substantial runoff to the hydrological system even though they cover a relatively large proportion of the basins studied. Above these elevations snowpacks of smaller areas contribute much greater runoff as a result of their greater thickness This is the conclusion from the Phahi Phari basin where the upper 38 percent of the basin area, which is over 4800 m a.s.l, contributes more than 70 percent of the seasonal runoff for that basin This indicates that for the development of a simulation model it is important to know the rate of increase in snowpack water equivalent with elevation, as the rate of increase in runoff is more than a simple linear function of area

Energy input

During day light hours air temperature is mainly a function of the total incident solar radiation and amounts of sensible and latent heat Nocturnal air temperatures depend primarily on the longwave radiation balance and wind velocity (Kuusisto 1984) It appears that air temperature is a secondary meteorological variable as opposed to radiation, humidity, and wind However, it is an important integrated measure of heat energy (Kuusisto 1984) In the present case

air-temperature proved to be the best explanatory variable in the simple regression models The mean air temperatures are the best indices for both sites, where they are recorded in the middle of the respective basins (Chapter 5 3) Correlations carried out between air temperatures recorded at base stations and at higher elevations indicate that the temperature at the base station can be used in snowmelt runoff forecasting models given the application of an appropriate lapse rate

Microclimatic effects

Patterns of snowmelt runoff, examined in the two contrasting environments within Barpu Glacier Basin, indicate that runoff is characterized by regular daily streamflow minima and maxima which reflect the daily fluctuations of the solar radiation and temperature

Weather conditions during the spring, particularly the cloud cover and daily air-temperature, directly influence the rate of snowmelt runoff

Precipitation events also cause variations in streamflow by adding fresh snow to the existing snow cover on higher elevations which increases snowpack albedo and decreases the radiant energy input, therefore suppressing the melting for short periods of time

Topoclimatic effects

Relief creates unequal distribution of snow which inturn causes areal variation in the volume of spring melt Aspect and degree of slope modifies the winter and spring snowpack by causing unequal rates of ablation

On a diurnal basis peak discharge occurs earlier in the basin with easterly exposure, but on a seasonal basis southwest facing slopes become bare before slopes facing northeast due to greater energy input Therefore, a basin with a large percentage of southwest facing slopes would seem to produce meltwater roughly 3 weeks earlier in clear weather than northeast facing slopes in the same elevation range

Increased elevation can effectively delay and increase the amplitude of runoff In the Central part of the Karakoram, if elevation of a basin exceeds 5,000 m a.s.l the peak discharge will occur when the snowpack above 4,800 m a.s.l. starts contributing to the streamflow, that is to say when the seasonal thermal conditions cause snowmelt on these elevations. On the evidence available it seems that peak discharge will not occur until the first week of July or, later However, clearly, this is subject to the weather conditions during the months of June and July

Snowcover/runoff relation

For hydrological analysis, the percentage snow cover in a basin is an primary variable It is an areal parameter and ground based techniques of snow cover measurements cannot detect significant variations Also in most of the mountain areas, accessibility limits the number of ground measurements, thus the representativeness of sampling sites can be a problem It is particularly true in the case of Karakoram mountain ranges where the region is very sparsely instrumented only during summmer However, the rapid development during the last 20 years of remote sensing techniques as applied to snow cover has provided new methods of observations, measurements and analysis But these techniques are yet to develop

enough to replace the ground based measurements and successful application of these new techniques still depend upon the "ground truth data for calibration and verification" The relationship developed between snow cover and streamflow indicate that snow cover area can be a good indicator of subsequent runoff subject to the accuracy of snow covered area measurements.

Snowmelt causatives and runoff relation

This investigation has attempted to identify the climatic and runoff variables which provide the best relationships between hydrometeorological elements and runoff from snowpacks, and has sought the best combinations of those variables to explain the variances of summer runoff Mutiple regression models developed between snowmelt causatives and summer runoff suggest that all wave radiation, humidity, air-temperature, and wind can produce satisfactory results in runoff forecasting However, if only one meteorological variable is available mean air temperature appear to be the best indicator

6 4 FURTHER STUDIES

The depletion of snowpack in a glacierized basin and the timing of melt water release on different aspect, and in different elevation range is not only important for the estimation of water yield, but also because it has a regulatory effect on glacier melt (Young 1981, Collins 1982, Hewitt 1985) After a snowy winter that generates above-average snowmelt will also retard the migration of the transient snowline up glacier and thus reduce amounts of ice melt (Krimmel and Tangborn 1974) Conversely, after a dry winter ice melt will begin earlier and yield extra

runoff to compensate for reduced snowmelt. Such information is vital in formulating runoff models for the Karakoram Ranges where there are great variations in runoff due to complex system of micro and topoclimatology This is essential if we are to confidently determine the relationship between snowmelt and runoff, as well as the role that snow plays in the hydrology of this highly glacierized region.

Several questions raised in this study cannot be answered with confidence due to the short period of the field record and other limitations, in particular difficulties encountered in the measuring of the physical properties of snow Many of these questions could be answered through a detailed knowledge of snow distribution in a more accessible basin before the onset of melt season if one can be found in this rugged area. The study involves measurements of snow depth and density in different elevation bands.

During the snowmelt period, measurements related to the factors affecting snowmelt runoff such as all-wave radiation, air temperature and wind speed should be carried out at higher elevations in addition On the basis of the data gathered, the following could be investigated

- micro and mesovariability of the snowcover, both before and during the snowmelt season

- areal variability of snow cover on contrasting aspects between the elevation range of 3,000 m to 6,000 m

- detailed formation of runoff from the onset of spring season

- effect of aspect on the runoff generation

- relationship between runoff and hydrometeorological variables

The results obtained could be used in the development of an areally distributed snowmelt model.

It is worthwhile to mention that within a particular tributary casual factors would possibly correlate more closely with discharge of the tributary rather than with total trunk discharge which reflects events throughout the total of the drainage area. This is particularly true in the Indus River system which drains heterogeneous climatic zones Therefore studies conducted simultaneously in at least two tributary basins are necessary to confirm the relationship applicable to whole drainage basin Such studies will also provide critical information with regards to the selection of hydrometeorological measurement sites in the rugged topography found in the Karakoram Mountains, and will undoubtedly be helpful during the operational phase of this project

The influence of temperature that have become apparent in the monitoring basins suggest that, in any investigation of forecasting procedures, temperature effects need further consideration The procedure described does provide an effective basis for studying temperature effects and accounting for such effects in forecasting runoff, and it seem desirable to repeat such a study in other basins

Continuous climatic measurements through the year, such as wind speed, air temperature and humidity, taken simultaneously in at least two glacierized basins of the Central Karakoram, would be representative of prevailing climatic conditions at

higher elevations of this region Such data will be important in formulating runoff forecasting models and in calibrating and verifying the remotely sensed data in future

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Appendices

Condensed From
PAKISTAN GLACIER INVENTORY+

MOUNTAIN AREA Karakoram GLACIER Barpu

SOURCES

Map Title and Number Hispar-Biafo Glacial Regions (The Geographical Journal, July-September 1950)

Compiled by E E Shipton Date 1950 (Surveyed 1939)

Scale 1 253,440 Contour Interval 250 feet

Map Title and Number India and Pakistan (Jammu and Kashmir) Topographic Series, Map NJ 43-14

Compiled by U S Army Map Service Date 1953 (Compiled from Survey of India, 1945)

Scale 1 250,000 Contour Interval 500 feet

TERMINUS CO-ORDINATES

Longitude 74°47' E

Latitude 36°13' N

MID-BASIN CO-ORDINATES

Longitude 74°48' E (Barpu/Bualtar Basin)

Latitude 36°07' N (")

ORIENTATION

Basin Long Axis 325° (NNW) Ablation Area 320° (NNW)

Accumulation Area 330° (NNW)

ELEVATIONS

Maximum Basin Elevation 7,460 m Glacier Terminus Elevation 2,835 m

Basin Elevation Range 4,625 m Ablation Line Elevation 4,116 m

Mean Accumulation Area Elevation (a) 5,788 m* (b) 5,428 m*

Mean Ablation Area Elevation (a) 3,475 m* (b) 3,500 m*

LENGTH AND AREA**

Maximum Glacier Length 29 km Mean Main Stream Width 1 5 km

Maximum Length Ablation Area 18 5 km Maximum Length Accumulation Area 10 5 km

Total Basin Area 414 98 km² Glaciated Area 125 56 km²

Connected Glacier Area 117 53 km²

Ablation Area 22 % 27 69 km² Accumulation Area 78 % 97 87 km²

SLOPE-ASPECT DATA TOTAL BASIN (Barpu/Bualtar Basin)

North facing 34 89 % 144 77 km² South facing 8 81 % 36.55 km²

East facing 32 77 % 36 55 km² West facing 23 53 % 97 65 km²

SLOPE-ASPECT DATA GLACIATED SLOPES (Barpu/Bualtar System)

Average Glacier Slope Barpu 1 6 72 Miar 1 4 59

North facing 58 07 % 122.01 km² South facing 3.67 % 7.71 km

East facing 20 17 % 42.38 km² West facing 18.09 % 38.01 km

+ Prepared by David Butz

* Ablation and accumulation area elevations are calculated in two ways (a) average of highest and lowest elevations, and (b) mean of area-altitude calculations above and below ablation line

** All area data refers to plan area

Condensed From
PAKISTAN GLACIER INVENTORY

AREA-ALTITUDE RELATIONSHIP

MOUNTAIN AREA Karakoram

GLACIER Barpu

Total Basin					Glaciated Area				
2,440-2,743m	6 88	km²	1 65	%	2,440-2,743m		km²		%
2,743-3,048m	15 52	km²	3 74	%	2,743-3,048m	3 08	km²	2 45	%
3,048-3,353m	26.41	km²	6 36	%	3,048-3,353m	8 63	km²	6 87	%
3,353-3,658m	31.24	km²	7 53	%	3,353-3,658m	8 22	km²	6 55	%
3,658-3,962m	35.97	km²	8 67	%	3,658-3,962m	8 22	km²	6 55	%
3,962-4,267m	40.29	km²	9 71	%	3,962-4,267m	6 89	km²	5 49	%
4,267-4,572m	42.44	km²	10 23	%	4,267-4,572m	7 91	km²	6 30	%
4,572-4,877m	47.89	km²	11 54	%	4,572-4,877m	10 79	km²	8 59	%
4,877-5,181m	48.20	km²	11 62	%	4,877-5,181m	16 11	km²	12 83	%
5,181-5,486m	42.35	km²	10 21	%	5,181-5,486m	17 48	km²	13 92	%
5,486-5,791m	26.61	km²	6 41	%	5,486-5,791m	12 03	km²	9 58	%
5,791-6,096m	21.58	km²	5.20	%	5,791-6,096m	12 34	km²	9 83	%
6,096-6,401m	14.39	km²	3.47	%	6,096-6,401m	7 39	km²	5 89	%
6,401-6,706m	10.81	km²	2.60	%	6,401-6,706m	7 90	km²	6 29	%
6,706-7,010m	3 38	km²	0 81	%	6,706-7,010m	1 34	km²	1 07	%
7,010-7,315m	0.71	km²	0 17	%	7,010-7,315m	1 03	km²	0 82	%
7,315-7,620m	0.31	km²	0 08	%	7,315-7,620m	0 10	km²	0 08	%
7,620-7,925m		km²		%	7,620-7,925m		km²		%
7,925-8,230m		km²		%	7,925-8,230m		km²		%
8,230-8,535m		km²		%	8,230-8,535m		km²		%

Condensed From
PAKISTAN GLACIER INVENTORY+

MOUNTAIN AREA Karakoram GLACIER Bualtar (Hopar)

SOURCES

Map Title and Number Hispar-Biafo Glacial Regions (The Geographical Journal, July-September 1950)

Compiled by E E Shipton Date 1950 (Surveyed 1939)

Scale 1 253,440 Contour Interval 250 feet

Map Title and Number India and Pakistan (Jammu and Kashmir) Topographic Series Map NJ 43-14

Compiled by U.S. Army Map Service Date 1953 (Compiled from Survey of India,1945)

Scale 1,250,000 Contour Interval 500 feet

TERMINUS CO-ORDINATES

Longitude 74°45' E

Latitude 36°15' N

MID-BASIN CO-ORDINATES

Longitude 74°48' E (Barpu/Bualtar Basin)

Latitude 36°07' N (")

ORIENTATION

Basin Long Axis 0° (N) Ablation Area 0° (N)

Accumulation Area 0° (N)

ELEVATIONS

Maximum Basin Elevation 7,275 m Glacier Terminus Elevation 2,439 m

Basin Elevation Range 4,836 m Ablation Line Elevation 4,268 m

Mean Accumulation Area Elevation (a) 5,772 m* (b) 5,352 m*

Mean Ablation Area Elevation (a) 3,437 m* (b) 3,354 m*

LENGTH AND AREA **

Maximum Glacier Length 20 5 km Mean Main Stream Width 1 km

Maximum Length Ablation Area 17 km Maximum Length Accumulation Area 3 5 km

Total Basin Area 414 98 km² Glaciated Area 84 53 km²

Connected Glacier Area 79 53 km²

Ablation Area 28 % 23 97 km² Accumulation Area 72 % 60 56 km²

SLOPE-ASPECT DATA TOTAL BASIN (Barpu/Bualtar Basin)

North facing 34 89 % 144 77 km² South facing 8 81 % 36 55 km²

East facing 32 77 % 136 01 km² West facing 23 53 % 97 65 km²

SLOPE-ASPECT DATA GLACIATED SLOPES (Barpu/Bualtar System)

Average Glacier Slope Bualtar 1 6 58

North facing 58 07 % 122 01 km² South facing 3 67 % 7.71 km²

East facing 20 17 % 42 38 km² West facing 18 09 % 38.01 km²

+ Prepared by David Butz

* Ablation and accumulation area elevations are calculated in two ways (a) average of highest and lowest elevations, and (b) mean of area-altitude calculations above and below ablation line

** All area data refers to plan area

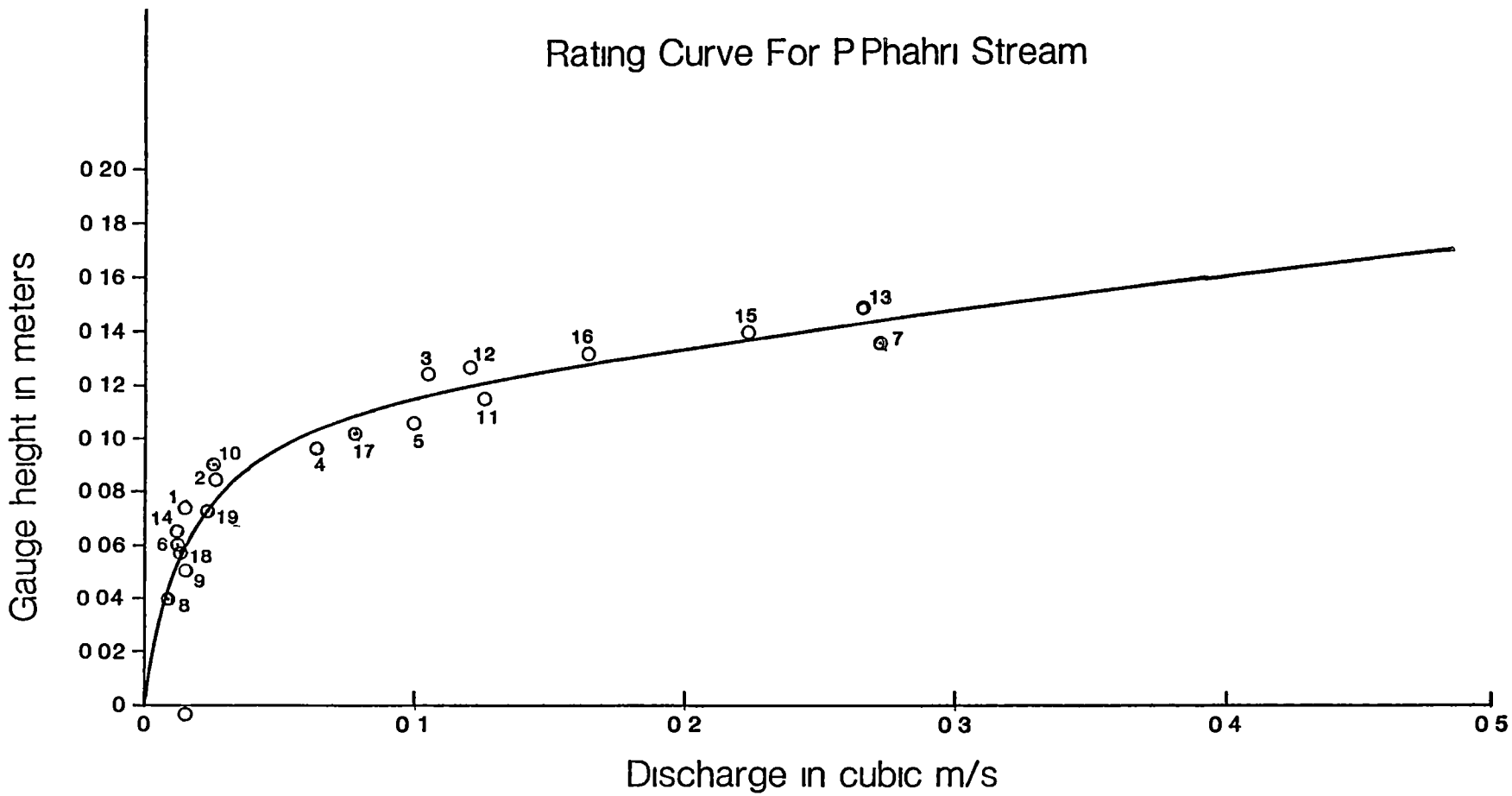
Condensed From
PAKISTAN GLACIER INVENTORY

AREA-ALTITUDE RELATIONSHIP

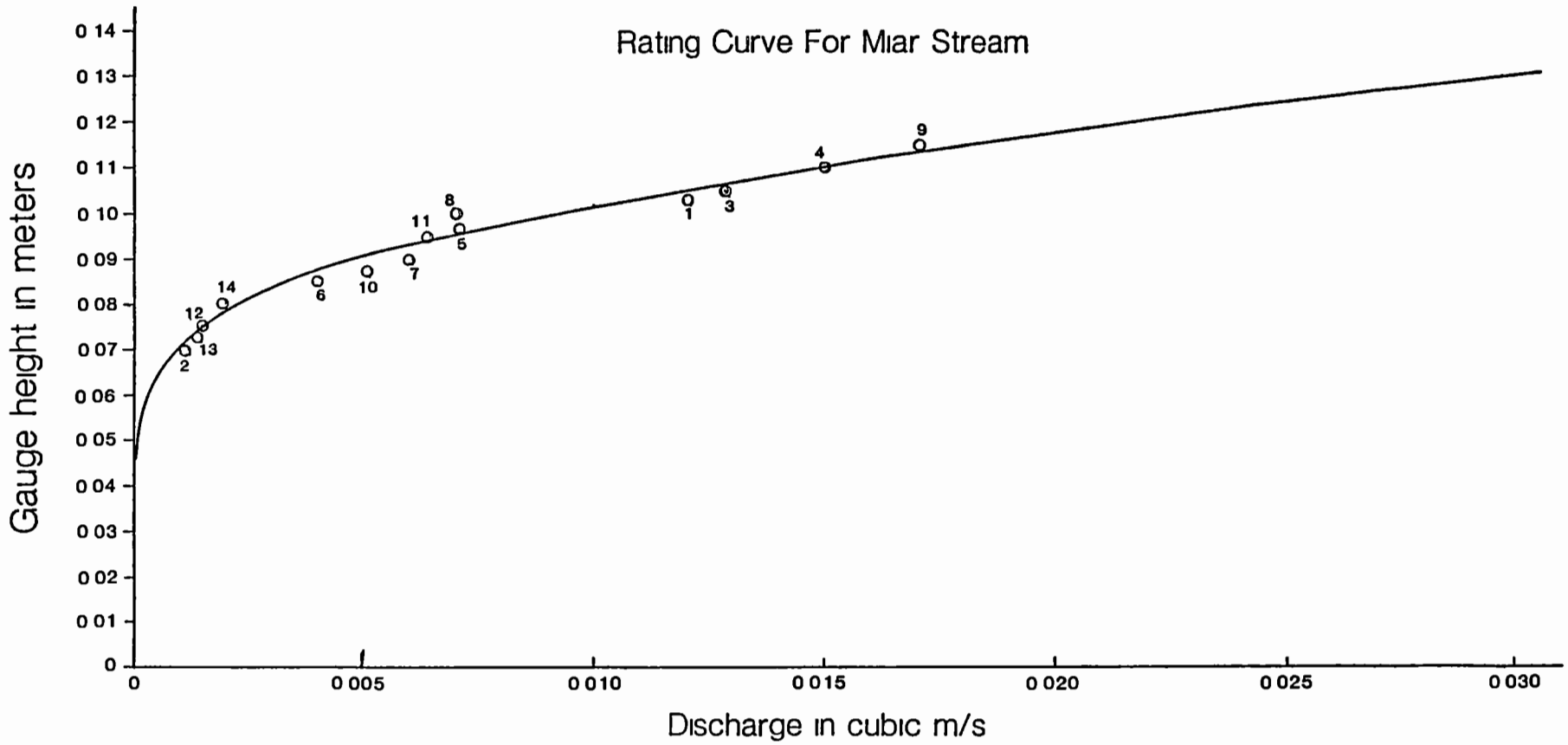
MOUNTAIN AREA Karakoram

GLACIER Bualtar (Hopar)

Total Basin				Glaciated Area			
2,440-2,743m	6 88	km²	1 65 %	2,440-2,743m	3 91	km²	4 62 %
2,743-3,048m	15.52	km²	3 74 %	2,743-3,048m	4 32	km²	5 11 %
3,048-3,353m	26.41	km²	6 36 %	3,048-3,353m	3 08	km²	3 64 %
3,353-3,658m	31.24	km²	7 53 %	3,353-3,658m	1 55	km²	1 83 %
3,658-3,962m	35.97	km²	8 67 %	3,658-3,962m	4 22	km²	4 99 %
3,962-4,267m	40.29	km²	9 71 %	3,962-4,267m	6 89	km²	8 15 %
4,267-4,572m	42.44	km²	10 23 %	4,267-4,572m	7 19	km²	8 51 %
4,572-4,877m	47.89	km²	11 54 %	4,572-4,877m	8 33	km²	9 85 %
4,877-5,181m	48.20	km²	11 62 %	4,877-5,181m	9 70	km²	11 48 %
5,181-5,486m	42.35	km²	10 21 %	5,181-5,486m	9 66	km²	11 43 %
5,486-5,791m	26.61	km²	6 41 %	5,486-5,791m	10 79	km²	12 76 %
5,791-6,096m	21.58	km²	5.20 %	5,791-6,096m	7 91	km²	9 36 %
6,096-6,401m	14.39	km²	3.47 %	6,096-6,401m	4 63	km²	5 48 %
6,401-6,706m	10.81	km²	2.60 %	6,401-6,706m	1.02	km²	1 21 %
6,706-7,010m	3.38	km²	0 81 %	6,706-7,010m	1 23	km²	1 46 %
7,010-7,315m	0.71	km²	0 17 %	7,010-7,315m	0 10	km²	0 12 %
7,315-7,620m	0.31	km²	0.08 %	7,315-7,620m		km²	%
7,620-7,925m		km²	%	7,620-7,925m		km²	%
7,925-8,230m		km²	%	7,925-8,230m		km²	%
8,230-8,535m		km²	%	8,230-8,535m		km²	%


Rating Curve For P Phahri Stream
Gauge height in meters
0 20
0 18
0 16
0 14
0 12
0 10
0 08
0 06
0 04
0 02
0
0
0 1
0 2
0 3
0 4
0 5
Discharge in cubic m/s
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19

APPENDIX B1

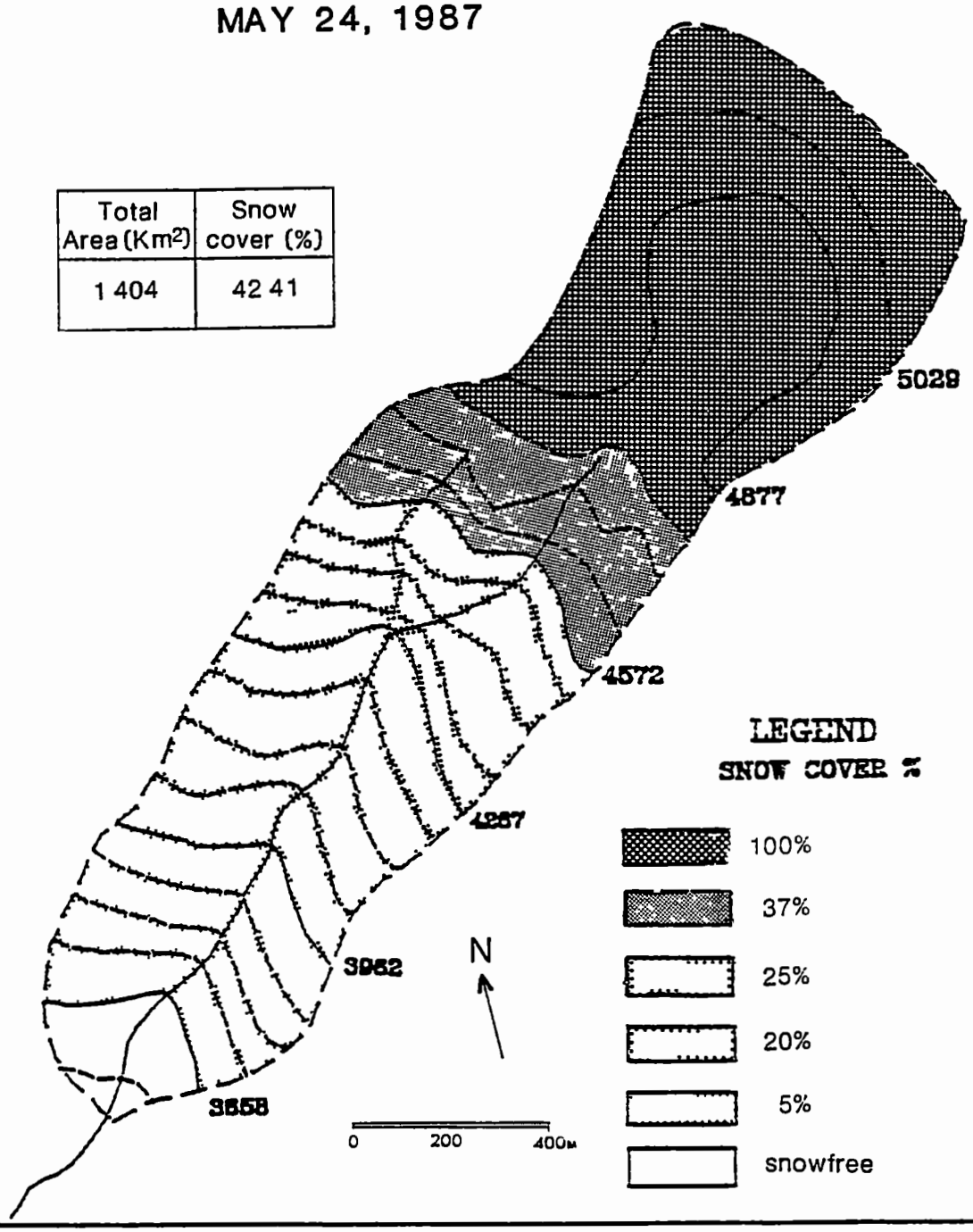

Rating Curve For Miar Stream
Gauge height in meters
Discharge in cubic m/s
0
0 01
0 02
0 03
0 04
0 05
0 06
0 07
0 08
0 09
0 10
0 11
0 12
0 13
0 14
0
0 005
0 010
0 015
0 020
0 025
0 030
1
2
3
4
5
6
7
8
9
10
11
12
13
14

APPENDIX B2

Phahi Phari Stream Basin
(southwest facing)

MAY 24, 1987

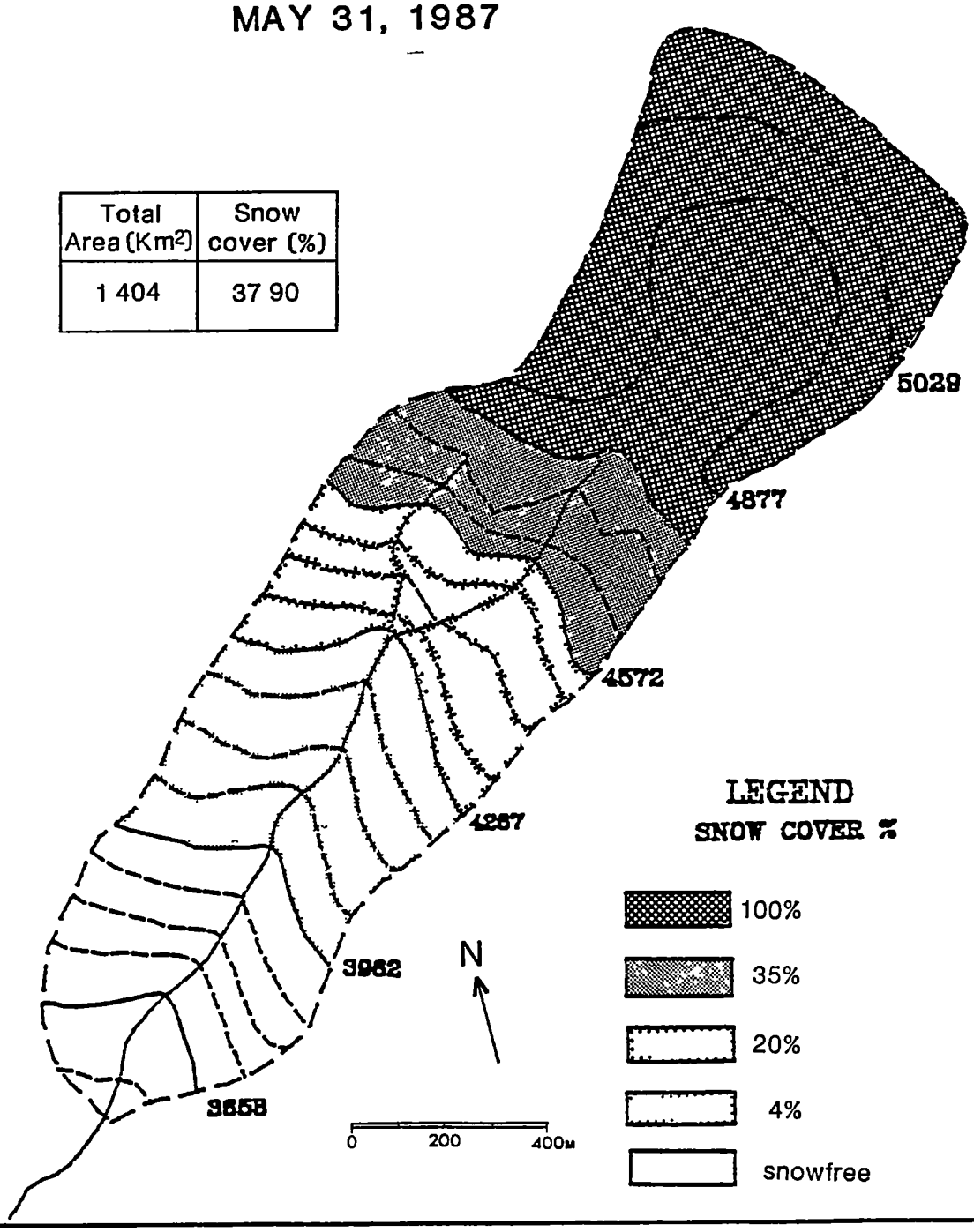
Total Area (Km^2)	Snow cover (%)
1 404	42 41



Phahi Phari Stream Basin
(southwest facing)

MAY 31, 1987

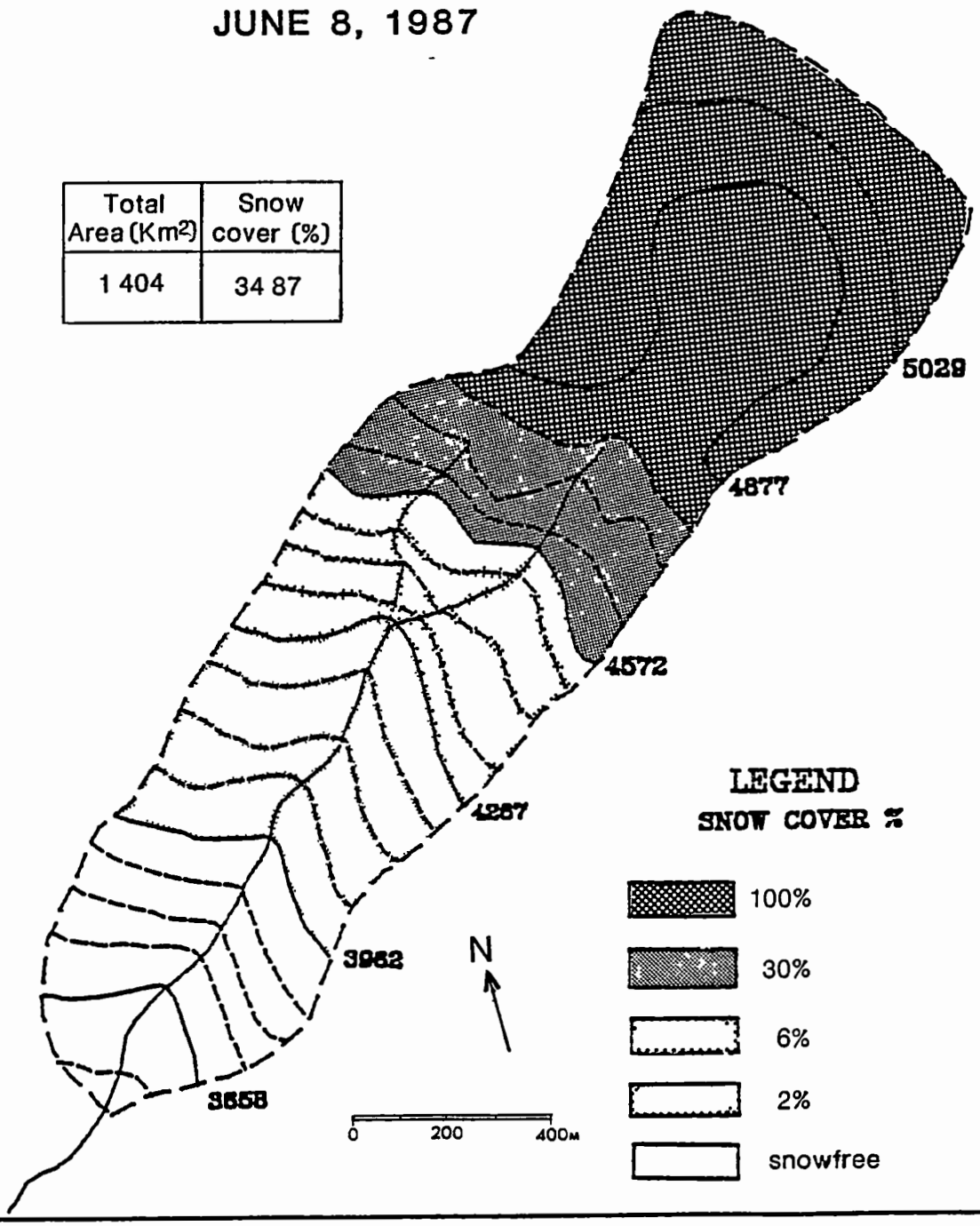
Total Area (Km2)	Snow cover (%)
1 404	37 90



Phahi Phari Stream Basin
(southwest facing)

JUNE 8, 1987

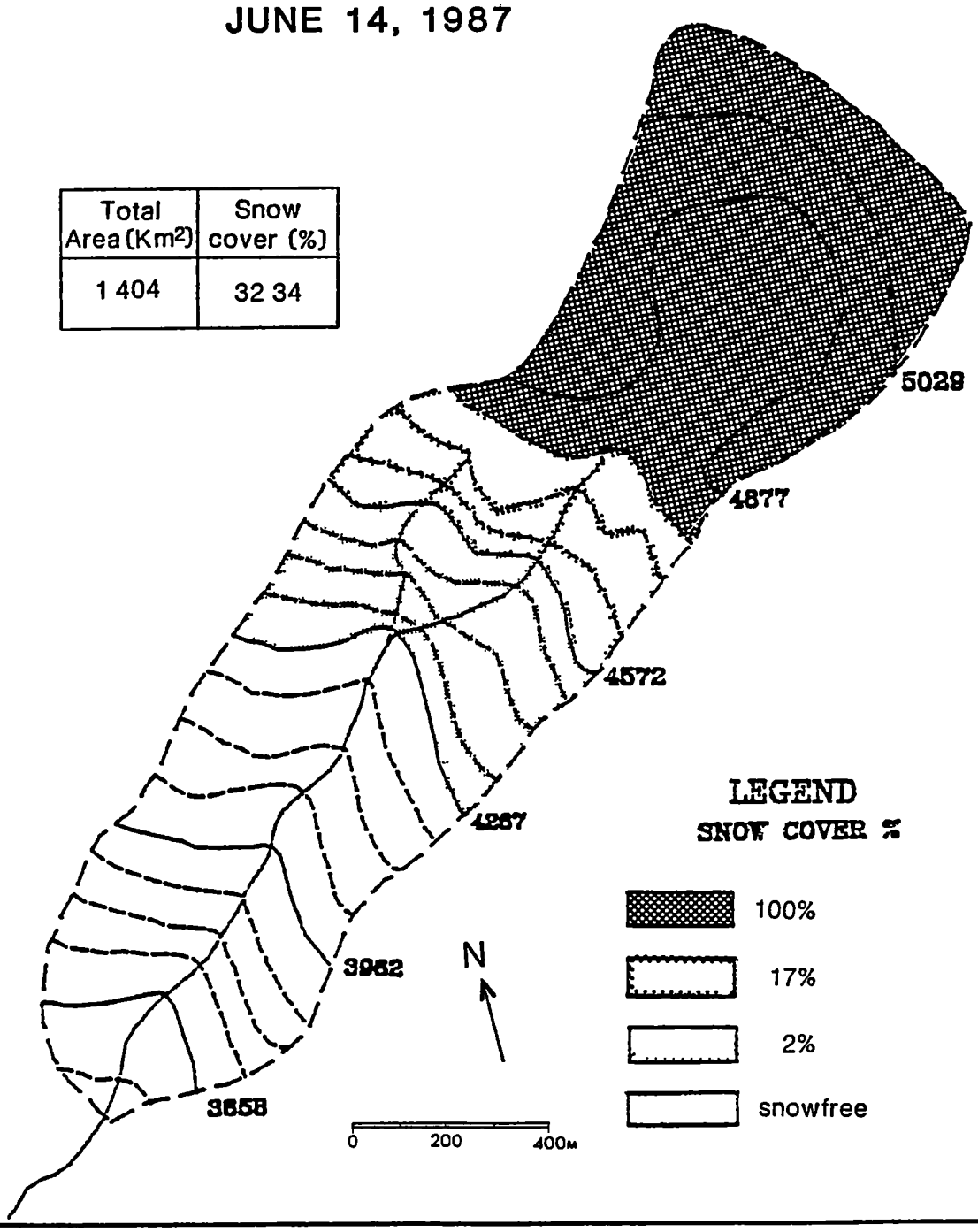
Total Area (Km2)	Snow cover (%)
1 404	34 87



Phahi Phari Stream Basin
(southwest facing)

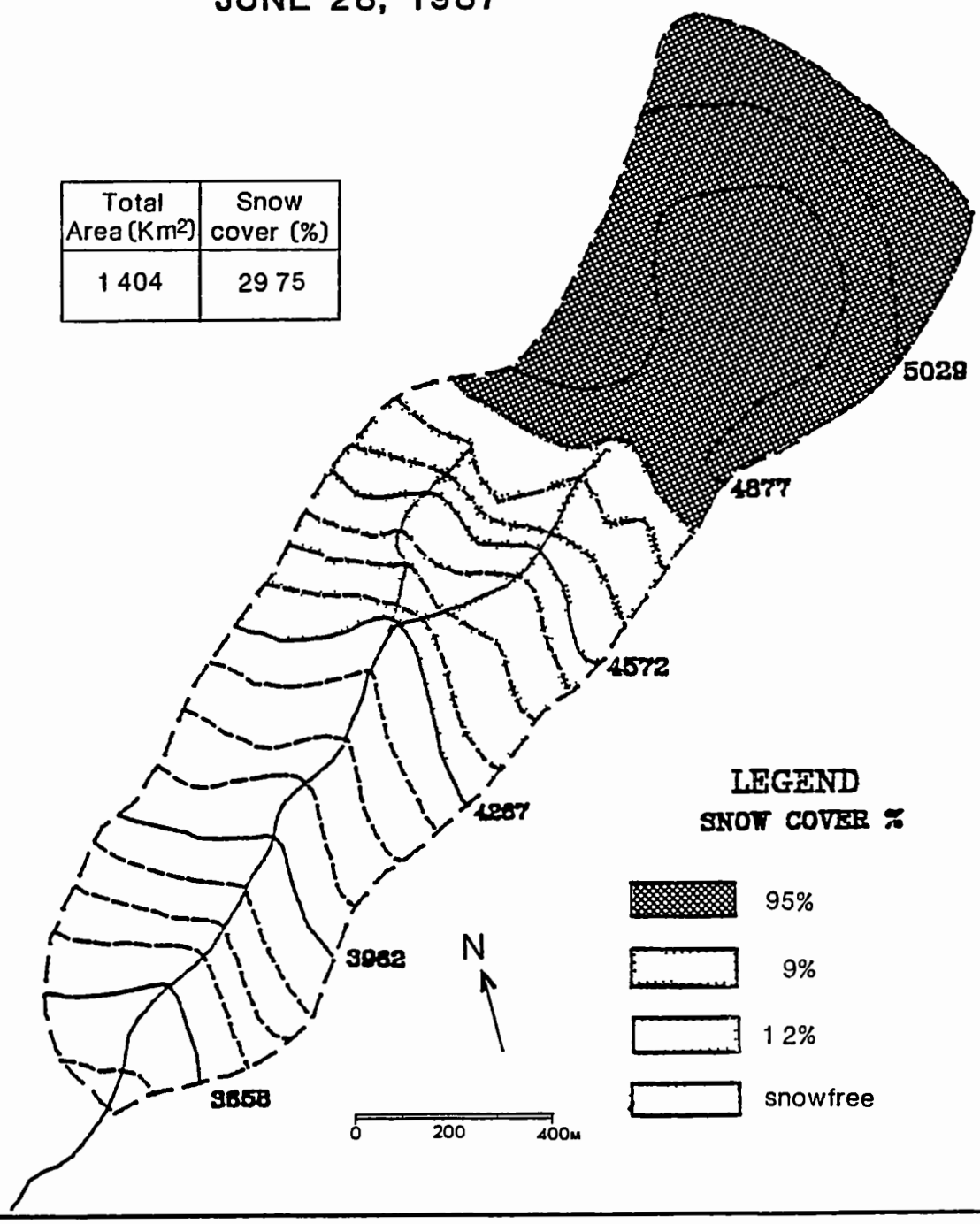
JUNE 14, 1987

Total Area (Km2)	Snow cover (%)
1 404	32 34



Phahi Phari Stream Basin
(southwest facing)
JUNE 28, 1987

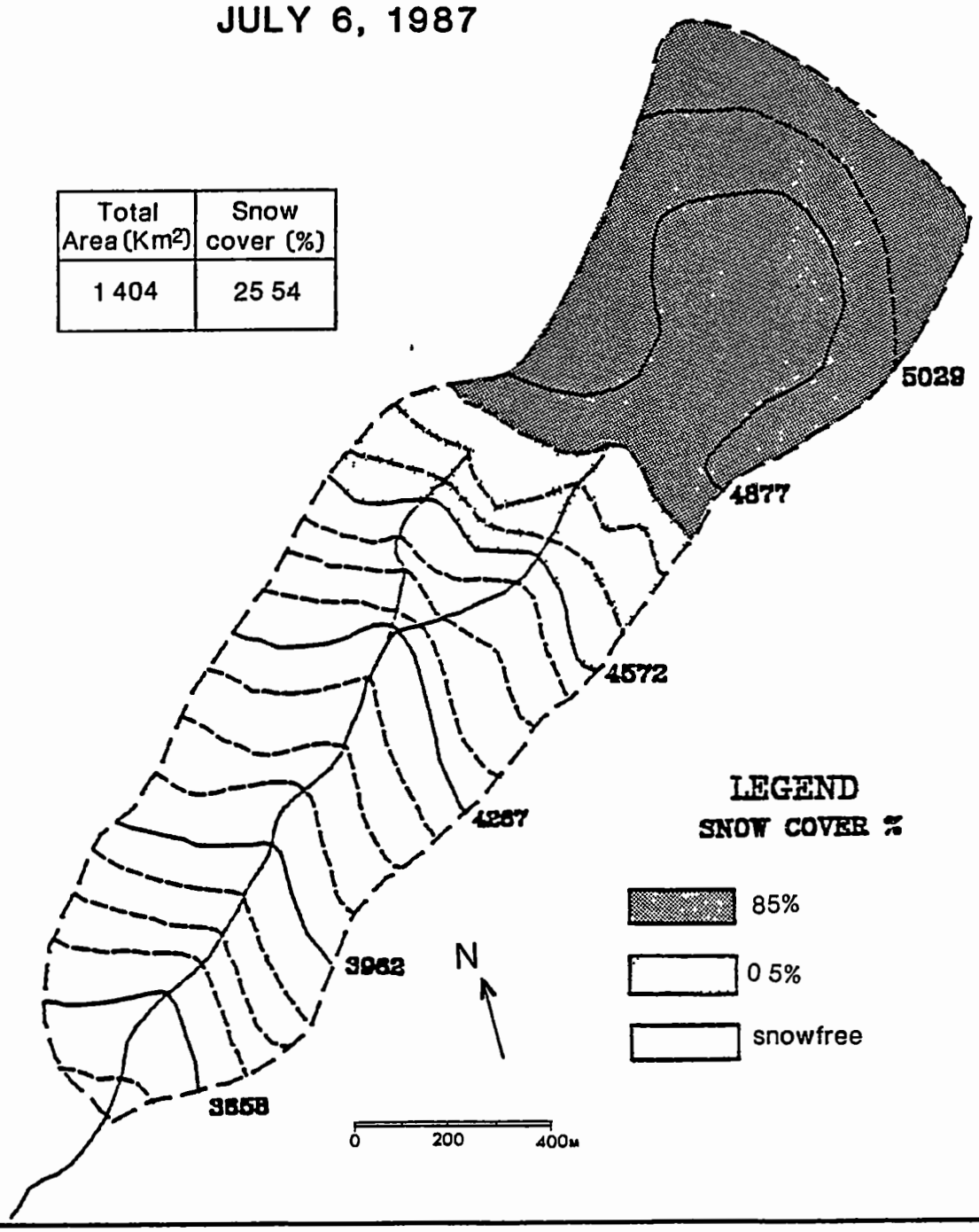
Total Area (Km2)	Snow cover (%)
1 404	29 75



Phahi Phari Stream Basin
(southwest facing)

JULY 6, 1987

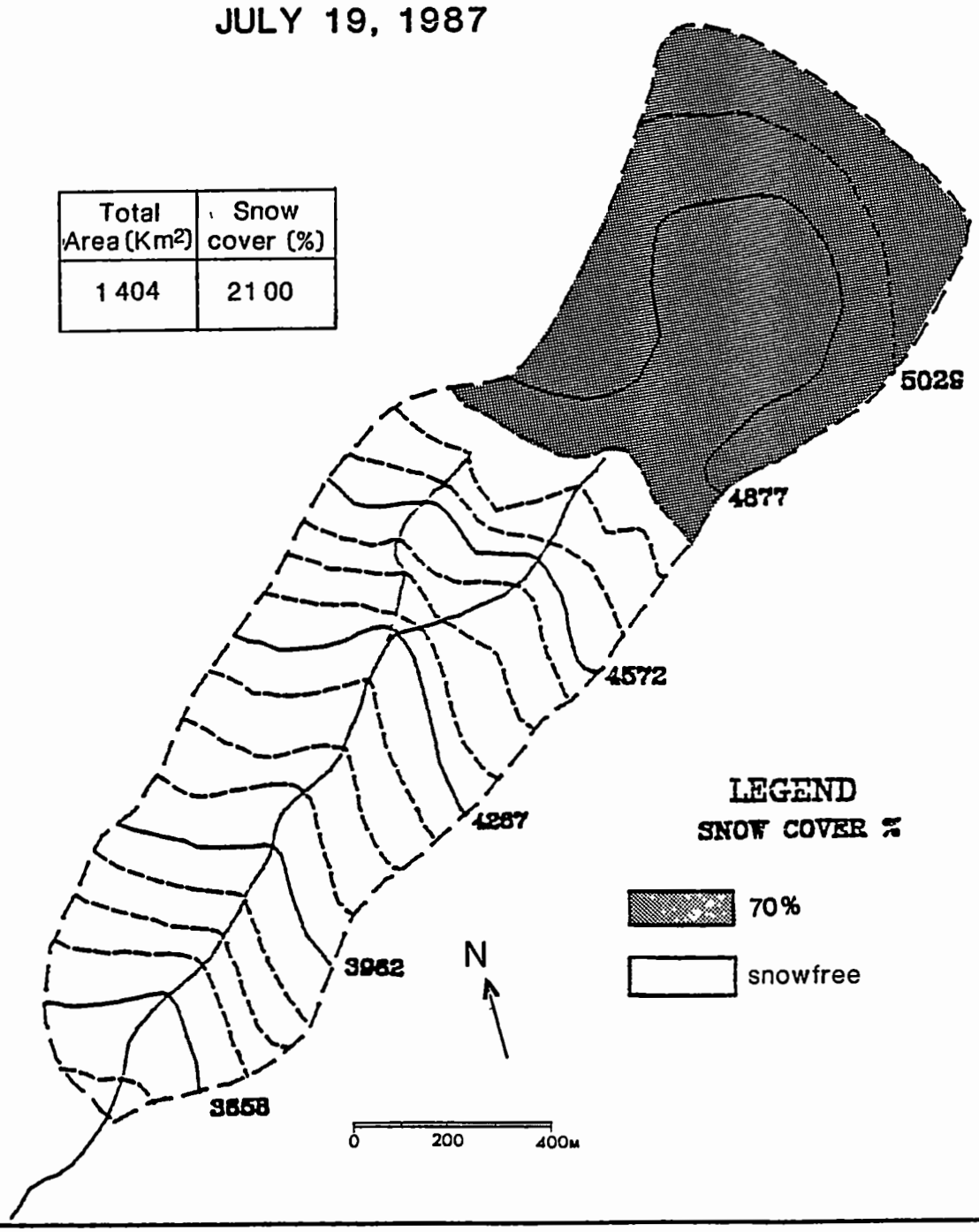
Total Area (Km2)	Snow cover (%)
1 404	25 54



Phahi Phari Stream Basin
(southwest facing)

JULY 19, 1987

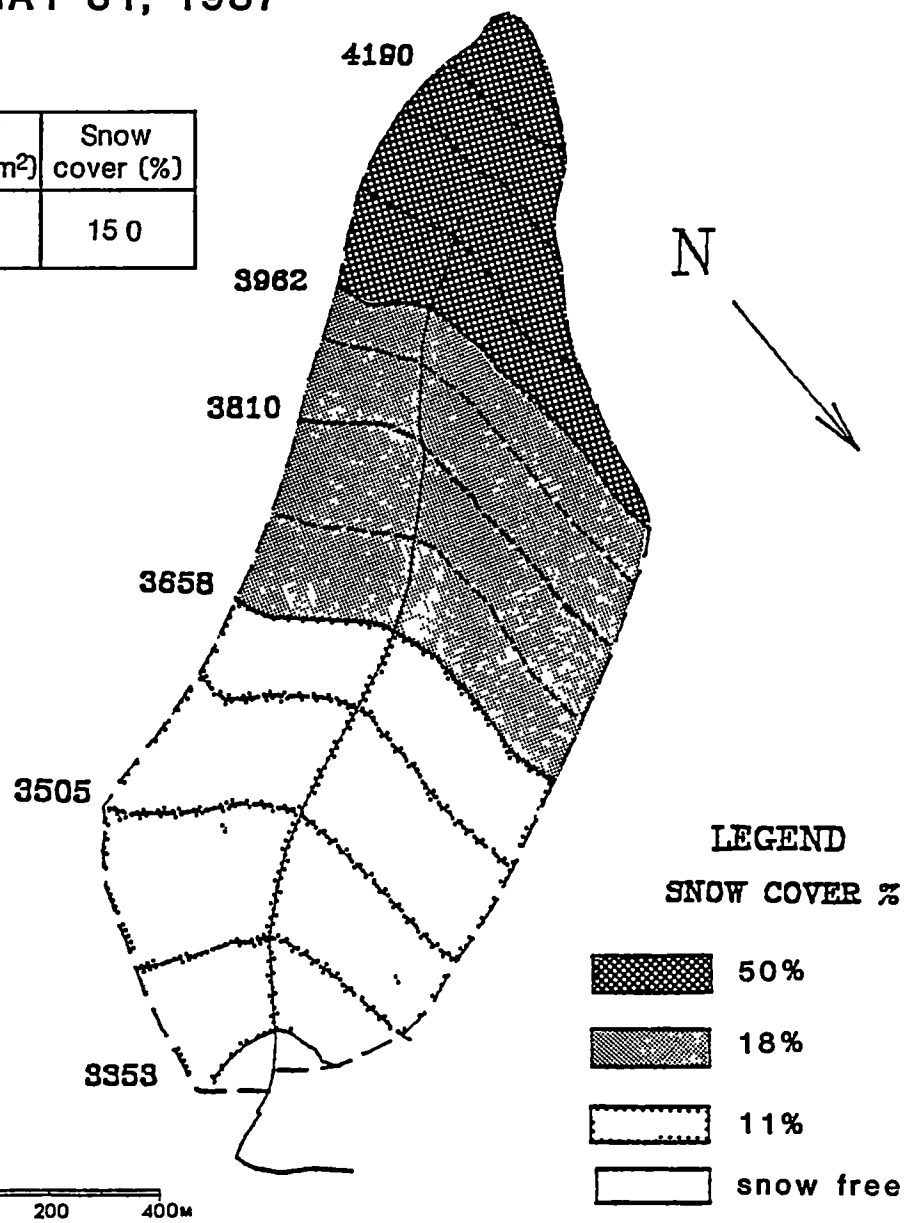
Total Area (Km2)	Snow cover (%)
1 404	21 00



Miar stream Basin
(northeast facing)

MAY 31, 1987

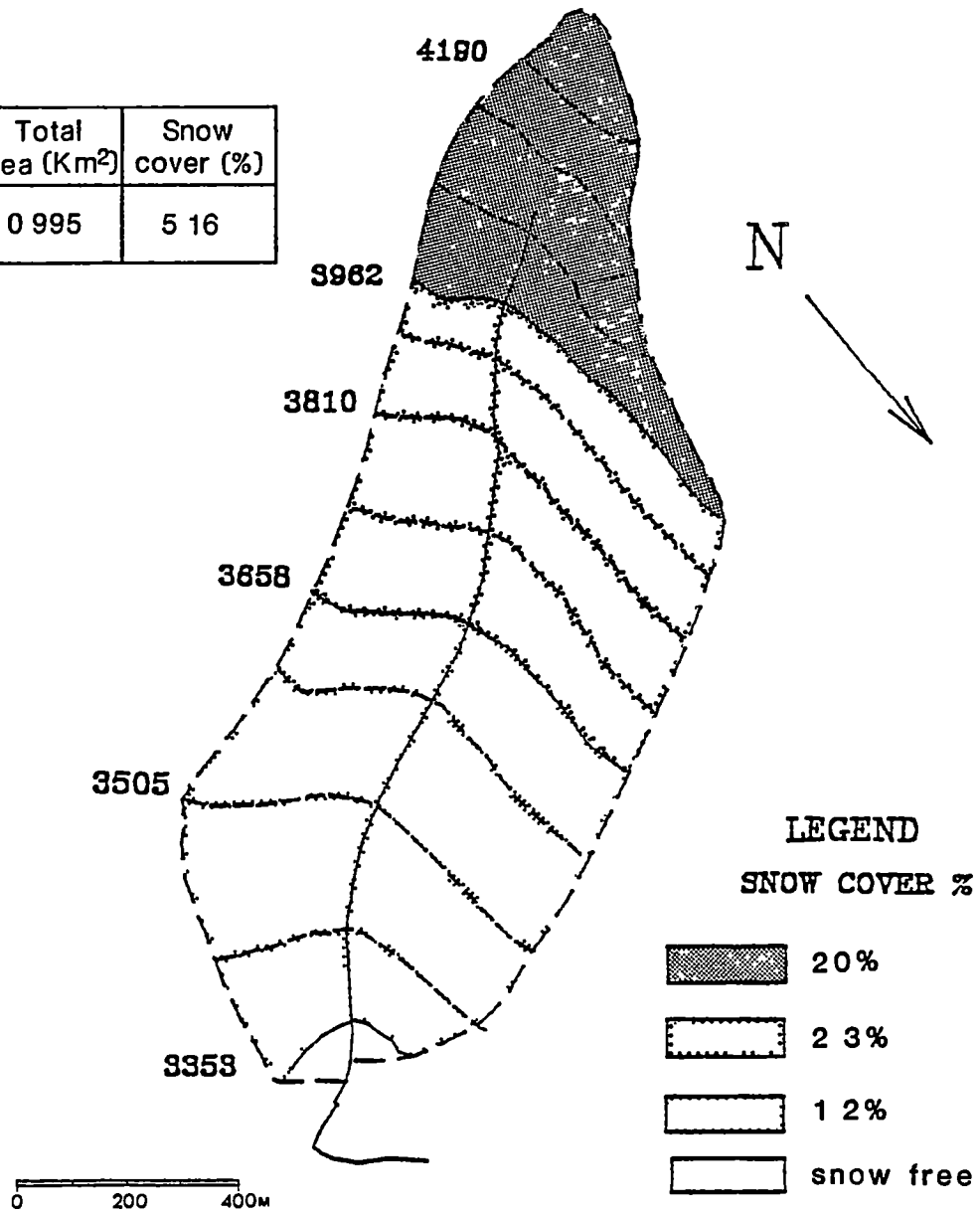
Total Area (Km^2)	Snow cover (%)
0 995	15 0



Miar stream Basin
(northeast facing)

JUNE 8, 1987

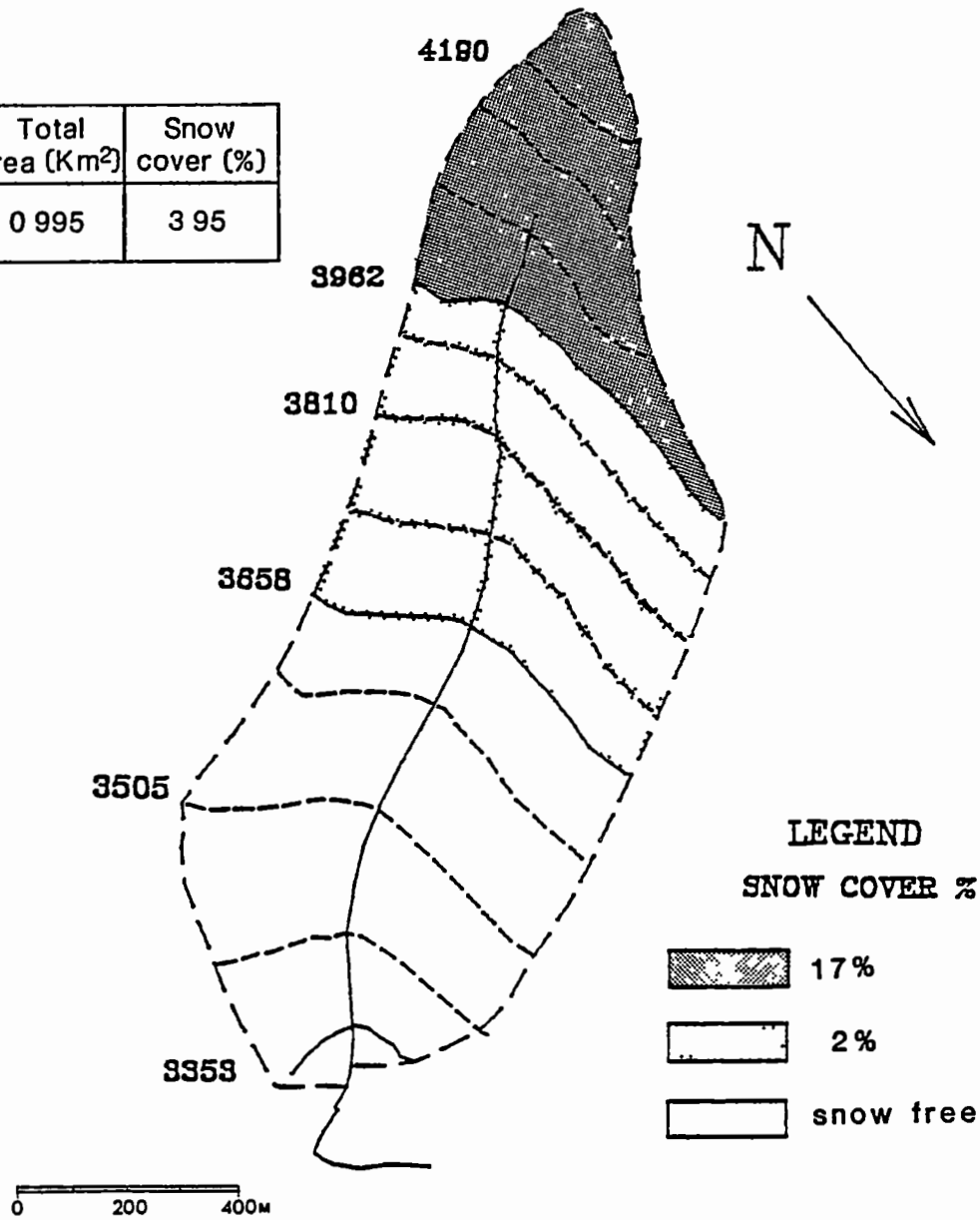
Total Area (Km^2)	Snow cover (%)
0 995	5 16



Miar stream Basin
(northeast facing)

JUNE 14, 1987

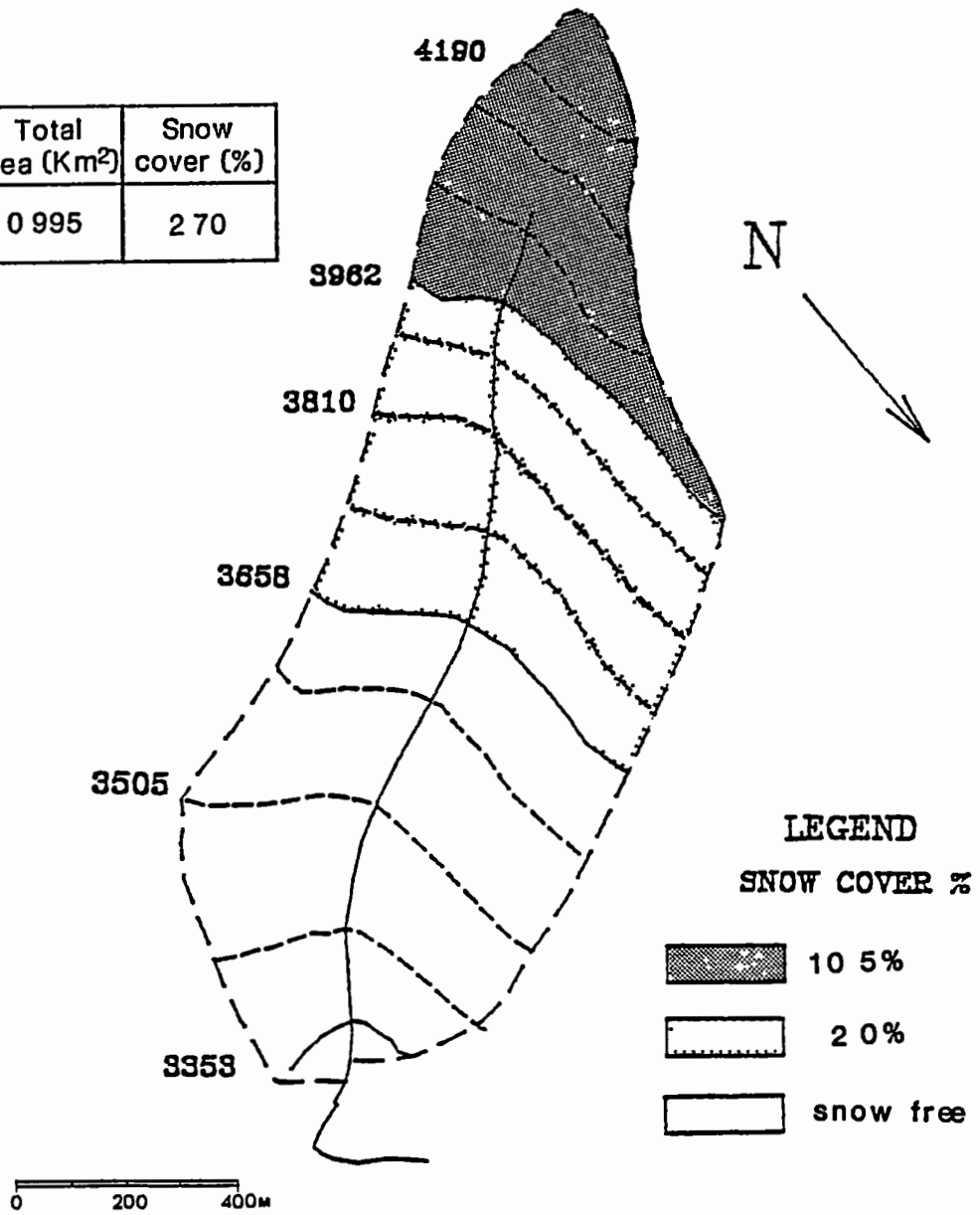
Total Area (Km^2)	Snow cover (%)
0 995	3 95



Miar stream Basin
(northeast facing)

JUNE 28, 1987

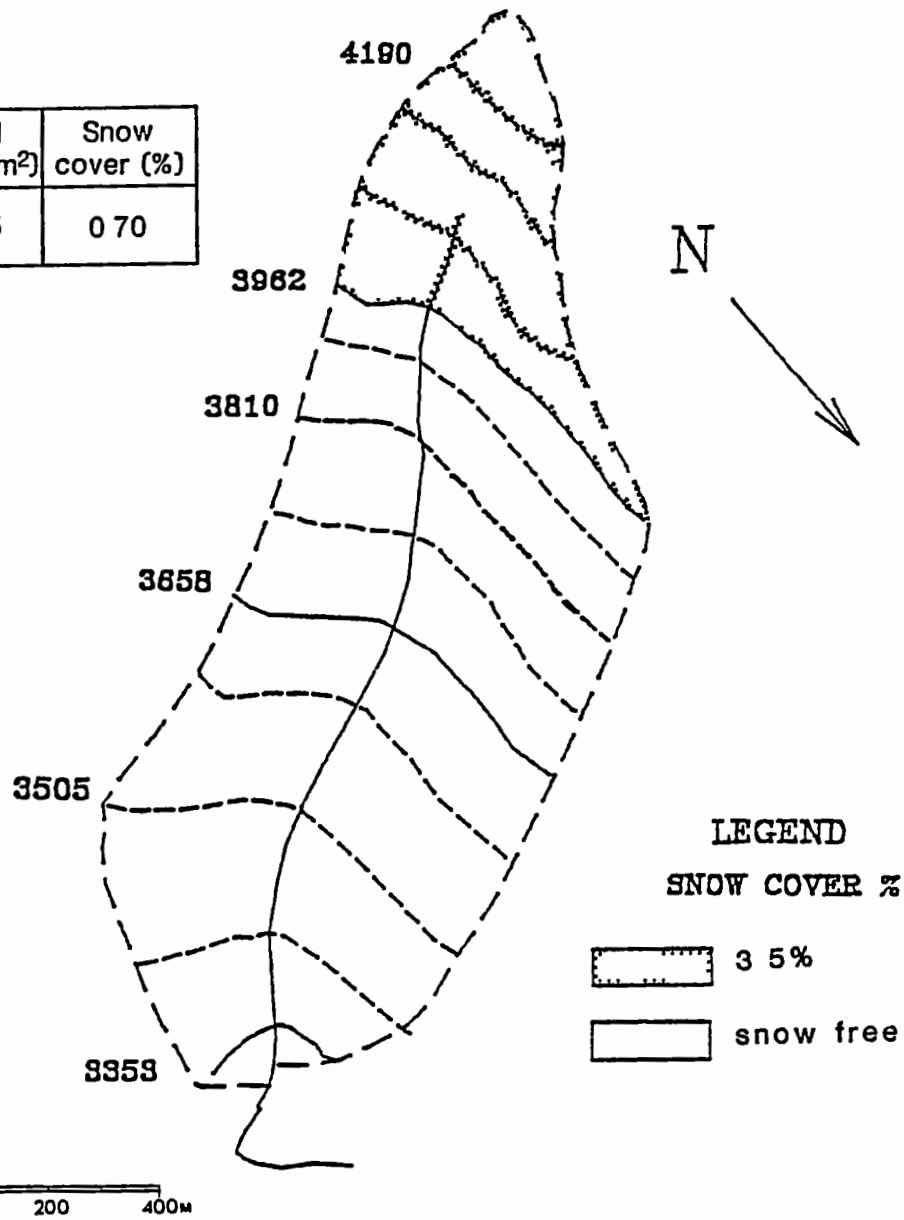
Total Area (Km^2)	Snow cover (%)
0 995	2 70



Miar stream Basin
(northeast facing)

JULY 5, 1987

Total Area (Km^2)	Snow cover (%)
0 995	0 70



C THIS PROGRAM WILL CALCULATE THE DAILY TOTAL OF DIRECT SOLAR RADIATION

```
C
C PARAMETERS
C              F     LATITUDE, IN DEGREES  SOUTH NEGATIVE
C              FO    ANGLE OF DECLINATION OF THE SUN IN DEGREES  SOUTH -VE
C              G     ATMOSPHERIC TRANSMISSIVITY, GIVEN AS A DECIMAL
C
C REMARKS      1)    THIS PROGRAM PRINTS A TABLE OF VALUES WITH SLOPE AND
C                     AZIMUTH IN TEN DEGREE INCREMENTS
C              2)    VALUES FOR THE SOLAR CONSTANT ARE SELECTED BY THE
C                     PROGRAM
C
C USAGE               THIS IS A MAIN PROGRAM AND REQUIRES NO SUBROUTINES
C
C METHOD              DESCRIBED IN GARNIER, B J  AND A OHMURA  "A METHOD OF
C                     CALCULATING THE DIRECT SHORTWAVE RADIATION INCOME ON
C                     SLOPES", JOURNAL OF APPLIED METEOROLOGY  VOL  7  1968
C
C
      DIMENSION Q(19),IQ(19)
C READ SPECIFICATION CARD (ANGLES IN DEGREES)  FORMATEIS (2F6 0,F4 0)
   99 READ (5,*,END=7)F,FO G
C HEAD PRINT OUT TABLE
      WRITE(6,101) F,FO G,(I,I=10 180,10)
C CONVERT DEGREES TO RADIANS FOR CALCULATION
      F=F*1745 3E-5
      FO=FO*1745 3E-5
C ESTABLISH SINES
      SF=SIN(F)
      SFO=SIN(FO)
C ESTABLISHE COSINES
      CF=COS(F)
      CFO=COS(FO)
C SELECTION OF SOLAR CONSTANT
      IF(FO GT 0 411 ) GO TO 6
      IF(FO LE 0 411 ) SOL=1 94
      IF(FO LT 0 341 ) SOL=1 97
      IF(FO LT 0 205 ) SOL=1 99
      IF(FO LT 0 068 ) SOL=2 02
      IF(FO LT -0 068) SOL=2 04
      IF(FO LT -0 205) SOL=2 06
      IF(FO LT -0 341) SOL=2 07
      IF(FO LT -0 411) GO TO 6
C INITIATE VALUES
      IB=0
      B=0 0
    1 A=0 0
      DO 4 I=1,19
      Q(I)=0 0
      X=-COS(A)*SIN(B)
      Y=SIN(B)*SIN(A)
      Z=COS(B)
      T1=(X*SF+Z*CF)*CFO
      T2=(-X*CF+Z*SF)*SFO
C SET HOUR ANGLE TO 0340 HOURS
      W=-2181 67E-3
C INCREMENT HOUR ANGLE BY 20 MINUTES
    2 W=W+8726 5E-5
C CHECK WHETHER HOUR ANGLE EXCEEDS 2000 HOURS
      IF(W GT 2 1) GO TO 4
```

```
C DETERMINE COSINE OF SUNS ZENITH ANGLE
      QD=CFO*CF*COS(W)+SFO*SF
C DETERMINE COSINE OF ANGLE BETWEEN SOLAR BEAM AND NORMAL TO THE SLOPE
      Q2=-Y*SIN(W)*CFO+T1*COS(W)
      QT=Q2+T2
C CHECK WHETHER SUN IS ABOVE HORIZON
      IF(QD LE 0 0) GO TO 2
C CHECK WHETHER SLOPE IS IN SHADOW
      IF(QT LE 0 0) GO TO 2
C DETERMINE OPTICAL AIR MASS, SECANT APPROXIMATION (0 TO 70 DEGREES)
      Q1=1 0/QD
      IF(Q1 LT 2 9) GO TO 3
C
C DETERMINE OPTICAL AIR MASS FOR 70 TO 90 DEGREES  (SMITHSONIAN TABLES)
      IF(Q1 GE 114 6) QQ=30 00
      IF(Q1 LT 114 6) QQ=26 96
      IF(Q1 LT 38 20) QQ=19 79
      IF(Q1 LT 22 93) QQ=15 36
      IF(Q1 LT 16 38) QQ=12 44
      IF(Q1 LT 12 74) QQ=10.39
      IF(Q1 LT 10 43) QQ= 8 90
      IF(Q1 LT  8 84) QQ= 7 77
      IF(Q1 LT  7 66) QQ= 6 88
      IF(Q1 LT  6 76) QQ= 6 18
      IF(Q1 LT  6 06) QQ= 5 60
      IF(Q1 LT  5 49) QQ= 5 12
      IF(Q1 LT  5 02) QQ= 4 72
      IF(Q1 LT  4 62) QQ= 4 37
      IF(Q1 LT  4 28) QQ= 4 07
      IF(Q1 LT  3 99) QQ= 3 82
      IF(Q1 LT  3 74) QQ= 3 59
      IF(Q1 LT  3 52) QQ= 3 39
      IF(Q1 LT  3 33) QQ= 3 21
      IF(Q1 LT  3 15) QQ= 3 05
      IF(Q1 LT  2 99) QQ= 2 90
      Q1=QQ
C
C COMPUTE TWENTY MINUTE RADIATION VALUES AND ADD TO TOTAL
    3 Q(I)=Q(I)+20 0*SOL*(G**Q1)*QT
C CHECH WHETHER HOUR ANGLE HAS REACHED 2000 HOURS
      IF(W LT 2 1) GO TO 2
C INCREMENT AZIMUTH BY 10 DEGREES, CONVERT TO RADIANS
    4 A=A+1745 3E-4
C ROUND OFF TOTAL RADIATION TO NEAREST WHOLE NUMBER  AND PRINT RESULT
      DO 5 M=1,19
    5 IQ(M)=Q(M)+0 5
      WRITE(6 102)IB IQ
C INCREMENT SLOPE ANGLE BY 10 DEGREES, CONVERT TO RADIANS
      IB=IB+10
      B=B+1745 3E-4
C CHECK SLOPE ANGLE DOES NOT EXCEED 90 DEGREES
      IF(B LT 1 6) GO TO 1
      GO TO 99
    6 WRITE(6 104)
      GO TO 99
    7 WRITE(6 103)
      STOP
C
  101 FORMAT(1H1/50X,'TABLE OF DIRECT RADIATION ON SLOPES //21X 1'
      LATITUDE' F9 2 20X,'DECLINATION',F9 2 20X,'TRANSMISIVITY ,F7 2///
      29H ANGLE OF50X, AZIMUTH OF SLOPE  /12H SLOPE      018I6/)
  102 FORMAT(1H ,I5,19I6)
  103 FORMAT(1H1)
  104 FORMAT( 0SELECTION SKIPPED BECAUSE OF UNALLOWABLE DECLINATION )
C
      END
```

APPENDIX F (List of data used for analyses)

Daily Figures of Hydrometeorological Data Recorded on SSW Facing Slpoes in Barpu Glacier Basin, Central Karakoram

Month/Day	Phahi Phari Main Meteorological Station (Elev 3510 m)								Rush Lake (Elev 4572 m)			
	Temperature (°C)			SWR (w/m²)	Wind (m/s)	Precip (mm)	R H (%)	Disch (m³/s)	Temperature (°C)			R H (%)
	Mean	Max	Min	Mean	Mean	Total	Mean	Mean	Mean	Max	Min	Mean
May- 25	6 6	11 4	2 1	763	1 66	0 0	18 5	013	—	—	—	—
26	8 5	13 8	3 6	755	1 67	0 0	23 0	018	—	—	—	—
27	9 6	14 1	5 2	694	1 52	0 0	23 5	023	—	—	—	—
28	10 3	15 4	6 1	662	1 42	0 0	28 0	032	—	—	—	—
29	11 3	16 2	7 8	701	1 35	0 0	32 0	071	—	—	—	—
30	11 6	16 7	8 0	641	1 47	0 0	35 5	035	—	—	—	—
31	11 5	17 3	7 2	706	1 23	0 0	39 0	069	—	—	—	—
June-01	11 5	17 4	8 1	603	1 47	0 0	43 0	057	—	—	—	—
02	7 2	11 4	4 2	251	0 65	0 0	70 0	043	—	—	—	—
03	5 2	10 6	3 0	374	1 22	12 5	71 0	021	—	—	—	—
04	9 2	15 1	4 4	773	0 96	0 0	38 0	032	—	—	—	—
05	11 8	17 0	7 5	747	1 74	0 0	25 0	054	—	—	—	—
06	13 1	18 5	8 2	741	1 62	0 0	23 0	032	—	—	—	—
07	14 5	21 2	9 9	736	1 53	0 0	22 0	059	—	—	—	—
08	10 8	14 3	6 7	429	1 55	0 0	43 0	015	—	—	—	—
09	5 9	10 8	1 9	401	1 31	0 0	69 0	006	—	—	—	—
10	5 1	9 8	-0 1	731	2 00	16 6	53 0	005	—	—	—	—
11	6 0	10 1	3 0	472	2 40	0 0	38 5	003	—	—	—	—
12	7 2	12 8	2 8	669	2 38	0 4	30 0	004	—	—	—	—
13	9 5	15 2	4 5	828	2 14	0 0	28 0	006	—	—	—	—
14	10 0	14 9	6 2	599	2 55	0 0	33 0	008	—	—	—	—
15	5 7	9 5	2 8	483	3 04	0 2	65 0	006	—	—	—	—
16	5 0	10 1	2 3	381	1 80	6 3	69 0	005	—	—	—	—
17	6 6	13 0	2 3	546	2 50	0 9	55 0	005	—	—	—	—
18	8 8	15 0	4 9	598	3 57	0 0	43 0	000	—	—	—	—
19	9 4	14 6	5 8	581	2 45	0 1	40 5	003	-2 1	0 7	-5 3	48 0
20	5 9	9 3	3 2	298	2 40	0 3	64 0	004	-4 1	-2 0	-10 0	63 0
21	2 7	5 4	0 8	252	0 66	10 9	74 0	006	-6 2	-4 7	-12 7	66 0
22	4 1	9 2	-0 4	706	0 60	17 5	49 0	005	-7 4	-4 0	-12 7	59 0

continue

Month/Day	Phahi Phari Main Meteorological Station (Elev 3510 m)								Rush Lake (Elev 4572 m)			
	Temperature (°C)			SWR (w/m²)	Wind (m/s)	Precip (mm)	RH (%)	Disch (m³/s)	Temperature (°C)			RH (%)
	Mean	Max	Min	Mean	Mean	Total	Mean	Mean	Mean	Max	Min	Mean
23	5 8	10 5	1 3	736	1 54	2 1	53 0	003	-5 0	-1 0	-8 0	66 0
24	8 8	13 3	4 9	698	1 48	0 0	39 0	004	-3 4	1 0	-8 0	57 0
25	9 5	14 8	5 5	826	1 16	0 0	49 0	008	-3 0	0 0	-6 0	60 0
26	9 8	14 5	7 1	476	0 89	0 0	49 0	006	-1 9	3 0	-5 2	56 0
27	11 4	16 7	7 4	675	1 40	0 0	45 0	022	-0 3	4 0	-5 0	48 6
28	12 0	18 1	8 0	671	1 02	0 0	38 0	025	0 1	6 0	-4 0	42 2
29	10 7	15 4	5 3	733	1 64	0 0	46 0	087	-0 1	6 8	-3 6	50 3
30	9 9	14 2	6 4	720	1 36	1 2	49 0	030	-1 1	2 0	-3 4	51 6
July-01	9 9	14 0	7 4	552	1 30	0 0	49 5	043	-1 5	3 0	-4 6	54 3
02	12 0	16 9	7 7	677	1 27	0 0	39 0	086	-0 8	2 0	-4 0	43 8
03	12 2	17 6	8 6	654	1 29	0 1	38 0	121	0 0	4 0	-4 0	41 8
04	13 8	20 0	8 7	746	1 32	0 0	34 0	161	1 4	5 0	-2 3	33 1
05	13 5	19 4	9 2	528	1 18	0 0	34 5	171	1 6	8 0	-2 0	39 3
06	13 2	20 1	8 3	753	0 86	0 0	37 0	157	1 5	3 8	-2 0	37 5
07	12 3	16 8	9 3	516	1 32	0 0	33 0	106	1 1	4 0	-2 0	41 2
08	12 2	18 4	9 4	578	1 38	0 0	34 0	095	0 5	2 6	-2 7	39 2
09	12 6	17 1	8 5	616	1 36	0 0	33 0	077	0 7	3 0	-2 5	42 6
10	13 3	20 1	9 6	734	1 04	0 0	37 0	105	1 0	3 1	-2 0	46 1
11	12 9	18 7	9 0	622	1 27	0 0	39 0	098	0 5	3 3	-1 3	51 9
12	6 8	10 2	5 7	209	0 78	2 8	68 0	059	-2 5	-2 0	-4 0	70 0
13	6 8	11 1	3 8	482	1 44	13 6	58 0	024	-3 6	-1 3	-6 2	62 9
14	6 0	11 3	-0 4	795	1 47	2 1	48 0	019	-2 9	0 0	-6 5	55 5
15	6 3	12 0	3 5	333	1 53	0 0	44 0	014	-1 5	3 3	-4 2	51 7
16	11 0	16 8	——	730	1 39	0 3	45 5	010	-0 6	2 0	-4 6	53 0
17	10 5	17 0	6 3	663	1 33	0 1	38 0	017	0 7	3 5	-1 8	44 6
18	11 3	14 5	7 7	714	1 43	——	41 0	014	1 5	5 0	0 0	52 9
19	9 3	14 0	5 7	678	1 48	——	57 0	018	0 4	2 3	-1 0	63 1
20	6 9	10 0	3 6	433	1 51	——	55 0	018	0 2	1 4	-1 0	62 0
21	11 0	17 2	5 2	——	——	——	——	——	3 0	6 2	-1 0	——
22	14 3	19 3	7 7	——	——	——	——	——	5 0	8 6	1 8	——
23	18 1	24 6	13 6	——	——	——	——	——	6 5	9 6	3 7	——
24	17 9	23 1	12 2	——	——	——	——	——	7 0	11 0	3 7	——

continue

Month/Day	Phahi Pharı Main Meteorological Station (Elev 3510 m)								Rush Lake (Elev 4572 m)			
	Temperature (°C)			SWR (w/m²)	Wind (m/s)	Precip (mm)	R H (%)	Disch (m³/s)	Temperature (°C)			R H (%)
o	Mean	Max	Min	Mean	Mean	Total	Mean	Mean	Mean	Max	Min	Mean
25	15 3	20 4	12 2	—	—	—	—	—	4 2	7 0	2 0	—
26	9 9	13 3	7 2	—	—	—	—	—	0 2	3 0	-2 0	—
27	8 2	11 7	5 9	—	—	—	—	—	-1 2	0 5	-3 0	—
28	10 3	14 7	5 4	—	—	—	—	—	0 0	3 2	-4 3	—
29	13 2	19 3	8 4	—	—	—	—	—	3 0	6 2	-1 5	—
30	11 3	15 0	8 9	—	—	—	—	—	0 8	4 0	-1 0	—
31	8 7	12 6	6 0	—	—	—	—	—	-1 0	1 3	-2 3	—
Augu-01	9 4	13 7	7 0	—	—	—	—	—	-1 2	0 0	-2 2	—
02	10 6	15 5	7 5	—	—	—	—	—	-1 0	2 0	-2 0	—
03	10 3	13 2	7 1	—	—	—	—	—	-2 0	-1 0	-3 0	—
04	10 9	15 3	7 4	—	—	—	—	—	-0 3	3 0	-4 0	—
05	13.9	19 6	9 9	—	—	—	—	—	1 3	4 6	-1 2	—
06	14 4	20 0	11 1	—	—	—	—	—	3 2	6 0	0 0	—
07	16 9	26 6	9 6	—	—	—	—	—	4 3	8 2	0 0	—
08	15 3	23 0	10 1	—	—	—	—	—	4 5	8 0	2 0	—
09	14 4	20 5	11 0	—	—	—	—	—	2 7	4 7	0 8	—
10	16 0	22 8	10 8	—	—	—	—	—	3 8	8 0	0 0	—
11	15 9	23 5	11 4	—	—	—	—	—	4 0	7 4	0 7	—
12	16 4	25 2	10 4	—	—	—	—	—	3 5	7 2	-0 2	—
13	16 5	23 6	11 2	—	—	—	—	—	4 0	7 0	0 0	—
14	15 7	20 9	12 4	—	—	—	—	—	4 6	7 5	1 4	—
15	15 1	22 9	11 1	—	—	—	—	—	4 0	7 0	2 0	—
16	14 3	21 2	10 4	—	—	—	—	—	3 6	7 0	1 0	—
17	13 6	18 8	10 4	—	—	—	—	—	1 8	2 7	0 0	—
18	11 6	13 6	10 5	—	—	—	—	—	1 0	4 4	0 0	—
19	15 6	22 6	10 8	—	—	—	—	—	5 2	7 4	3 0	—

SWR = Incoming short wave radiation R H = Relative Humidity Precip = Precipitation

(Source Field Observations SIHP 1987)

Daily Figures of Hydrometeorological Data Recorded on NNE Facing Slpoes in Barpu Glacier Basin, Central Karakoram

	Lower elevation site (elev 3550 m)				Higher elevation site (elevation (4200 m)					
	Temperature (°C)			Disch (m^3/s)	Temperature (°C)			RH (%)	SWR (w/m^2)	Cloud 10th
	Mean	Max	Min	Mean	Mean	Max	Min	Mean	Mean	Mean
May- 25	8 2	13 9	3 1	——	0 6	5 4	-4 0	——	——	0 0
26	8 3	13 2	3 1	——	0 7	5 2	-3 9	——	——	0 0
27	9 7	14 7	5 3	——	2 3	6 2	-2 0	——	——	0 0
28	11 1	15 4	6 8	——	3 6	8 0	-1 0	——	——	3 0
29	11 6	16 7	7 2	——	3 7	8 0	0 0	——	——	1 8
30	12 0	16 9	8 3	——	4 0	7 0	0 3	——	——	2 5
31	12 0	16 6	7 5	——	5 0	10 0	0 0	——	——	0 0
June-01	11 8	17 1	8 5	——	4 5	9 0	1 0	——	——	3 8
02	7 1	10 6	3 9	——	-1 1	2 5	-2 7	——	——	10 0
03	5 5	7 9	3 6	001	-2 5	-1 0	-5 5	85 6	500	7 5
04	9 5	14 3	4 0	004	2 1	7 0	-4 0	55 0	521	0 0
05	12 1	15 9	8 1	006	5 1	10 0	0 0	32 0	473	0 0
06	13 8	18 6	8 4	006	6 9	12 0	1 3	30 5	475	0 0
07	15 0	19 0	9 9	009	7 8	12 4	3 0	29 0	486	0 0
08	12 5	14 1	6 5	004	2 0	6 0	-2 0	72 0	507	7 2
09	5 5	11 7	1 0	001	-2 3	3 0	-5 4	90 0	744	9 5
10	5 4	9 0	-0 4	002	-3 6	-1 0	-7 7	67 0	531	3 0
11	5 8	10 2	2 5	001	-2 9	2 0	-5 0	62 0	577	5 5
12	7 1	11 6	2 8	004	-0 3	4 0	-5 3	47 0	534	2 5
13	9 0	13 7	4 5	005	1 9	6 3	-3 0	35 5	535	1 7
14	10 7	14 3	5 5	004	3 4	7 0	-2 0	45 0	496	2 5
15	6 3	11 1	3 4	002	-1 6	3 0	-4 2	93 0	396	8 2
16	5 5	10 0	2 7	001	-3 8	-1 3	-5 6	96 0	371	7 5
17	7 6	11 6	5 6	002	-2 0	1 0	-6 2	82 0	521	4 8
18	9 6	14 6	5 8	002	0 8	5 0	-3 7	60 0	589	4 7
19	10 5	13 5	6 3	002	1 5	6 0	-3 0	55 0	654	4 0
20	5 1	7 9	2 6	001	-2 9	-1 0	-5 0	88 0	393	9 5
21	2 4	5 8	-1 3	000	-5 9	-3 7	-9 0	91 0	95	10 0
22	6 9	9 3	1 4	001	-4 4	1 0	-9 5	71 0	508	5 0
23	5 1	10 1	0 9	001	-3 7	1 0	-6 8	86 0	560	4 0

continue

	Lower elevation site (elev 3550 m)				Higher elevation site (elevation (4200 m)					
	Temperature (°C)			Disch (m^3/s)	Temperature (°C)			RH (%)	SWR (w/m^2)	Cloud 10th
	Mean	Max	Min	Mean	Mean	Max	Min	Mean	Mean	Mean
24	8 4	13 5	4 4	003	0 6	4 0	-4 0	54 5	592	1 8
25	8 7	14 3	4 9	004	1 0	4 3	-3 0	69 0	591	6 0
26	9 6	15 0	6 0	004	3 6	8 2	-2 0	56 8	333	5 6
27	8 1	12 0	5 9	005	4 3	10 0	0 0	49 0	490	2 8
28	12 4	17 5	9 1	005	5 4	11 0	1 7	38 0	551	2 3
29	12 3	15 9	9 4	004	5 2	9 0	2 0	44 0	584	5 6
30	10 1	12 0	8 4	002	3 1	7 5	-1 0	53 0	703	2 6
July-01	10 1	15 4	6 7	001	3 1	7 5	0 0	——	446	4 0
02	11 5	15 7	7 2	001	4 7	9 3	1 0	——	602	3 0
03	12 0	17 7	8 1	002	5 7	10 5	2 3	——	532	4 8
04	13 7	19 1	8 5	003	7 5	13 0	2 0	——	500	0 3
05	13 5	18 8	9 5	002	7 4	13 0	2 7	——	460	5 0
06	13 2	19 4	9 1	002	6 7	12 0	2 5	——	528	0 4
07	12 4	19 1	9 2	001	5 8	12 0	2 4	——	526	3 6
08	11 9	16 5	8 7	001	4 7	9 0	1 5	——	616	3 5
09	12 9	17 6	9 0	001	6 3	11 5	2 3	——	522	5 2
10	13 3	19 4	8 9	001	6 1	12 4	1 7	——	521	3 8
11	12 5	18 8	8 7	001	6 1	12 0	1 5	——	499	6 0
12	6 6	10 5	5 0	——	0 1	2 5	-1 0	——	——	8 2
13	6 5	11 0	2 8	——	-0 3	3 8	-3 6	——	——	8 5
14	8 5	14 5	3 1	——	1 7	7 5	-4 0	——	——	2 8
15	9 8	13 4	7 2	——	3 5	8 0	0 6	——	——	4 0
16	10 3	13 5	7 2	——	3 6	9 0	-1 0	——	——	3 5
17	11 8	15 8	7 9	——	4 9	10 0	1 0	——	——	4 3
18	11 9	17 1	9 5	——	4 9	10 0	2 5	——	——	4 1
19	10 4	14 6	7 1	——	3 3	7 5	1 0	——	——	4 8
20	10 2	14 1	7 3	——	3 3	7 0	1 0	——	——	8 0
21	13 8	18 3	8 4	——	7 5	13 3	1 2	——	——	——
22	16 5	21 0	11 8	——	9 8	15 0	4 6	——	——	——
23	17 8	22 3	14 3	——	11 9	17 0	6 8	——	——	——
24	18 0	22 0	14 0	——	12 1	17 0	7 3	——	——	——
25	15 2	19 2	12 7	——	8 7	17 0	5 0	——	——	——
26	9 2	13 8	6 5	——	2 5	7 0	-0 5	——	——	——

continue

	Lower elevation site (elev 3550 m)				Higher elevation site (elevation (4200 m)					
	Temperature (°C)			Disch (m^3/s)	Temperature (°C)			RH (%)	SWR (w/m^2)	Cloud 10th
	Mean	Max	Min	Mean	Mean	Max	Min	Mean	Mean	Mean
27	7 5	10 8	5 4	—	0 6	4 0	-1 7	—	—	—
28	9 6	14 1	4 9	—	2 8	7 5	-2 0	—	—	—
29	12 4	17 0	7 9	—	5 5	10 0	1 0	—	—	—
30	10 8	14 9	8 1	—	3 5	8 0	0 6	—	—	—
31	7 4	12 6	5 6	—	-0 2	3 2	-2 2	—	—	—
Augu-01	7 9	10 6	5 3	—	0 0	4 0	-2 0	—	—	—
02	9 2	14 4	6 2	—	1 3	6 0	-1 2	—	—	—
03	6 2	7 8	4 1	—	-1 1	0 0	-2 2	—	—	—
04	9 3	13 8	4 7	—	1 2	6 7	-3 0	—	—	—
05	12 4	17 3	8 2	—	5 5	10 0	0 0	—	—	—
06	14 0	17 7	10 2	—	7 3	12 0	2 3	—	—	—
07	15 1	18 9	11 0	—	8 3	14 0	3 6	—	—	—
08	15 0	19 6	12 0	—	8 4	14 1	4 8	—	—	—
09	13 3	17 9	10 2	—	6 4	12 0	2 8	—	—	—
10	14 6	18 5	10 0	—	7 7	13 0	2 0	—	—	—
11	14 7	19 0	11 0	—	8 0	13 0	3 2	—	—	—
12	14 4	18 0	9 0	—	6 7	13 0	2 0	—	—	—
13	15 3	19 2	11 0	—	8 5	14 2	3 0	—	—	—
14	15 2	18 3	12 1	—	8 0	13 5	3 3	—	—	—
15	13 7	18 3	10 3	—	7 8	14 0	3 6	—	—	—
16	13 0	15 0	9 0	—	6 3	11 0	2 4	—	—	—
17	12 9	16 4	10 0	—	5 5	11 0	2 4	—	—	—
18	10 1	13 2	8 9	—	3 0	5 3	1 8	—	—	—
19	13 5	17 5	9 0	—	6 0	12 1	1 8	—	—	—
20	17 0	21 0	12 9	—	10 5	16 0	5 8	—	—	—
21	17 8	21 5	13 7	—	11 5	17 0	7 0	—	—	—

RH = Relative Humidity, SWR = Incoming short wave radiation, (Source Field Observations SIHP 1987)