PLIOCENE AND QUATERNARY ENVIRONMENTAL CHANGE

IN KASHMIR, NORTH-WEST HIMALAYA.

Thesis submitted for the degree of Doctor of Philosophy by

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ABSTRACT OF THESIS PRESENTED FOR THE DEGREE OF DOCTOR OF PHILOSOPHY BY JONATHAN A. HOLMES, HERTFORD COLLEGE, UNIVERSITY OF OXFORD. MICHAELMAS TERM, 1988.

Late Cainozoic environmental changes in Kashmir (33°30' to 34°30' N: 74°10' to 75°30' E) have been reconstructed using a range of techniques. The sedimentary record in Kashmir consists of a thick (>1000m) basin-fill sequence known as the Karewa group, together with glacial and related sediments in the surrounding mountain flanks. The Karewa sediments are fluviolacustrine in origin and comprise alternations of conglomerates, sands and clayey silts.

Work on the lower Karewa formation, which has previously been dated palaeomagnetically to between 4 and 0.4 MaBP, involved the semi-quantitative analysis of clay-mineral assemblages by X-ray diffraction. The clay minerals in the lower Karewa mudstones are interpreted as detrital clays which reflect weathering within Kashmir basin. The analyses showed a change in clay mineralogy between about 2.5 and 2.3MaBP, from abundant kaolinite to abundant smectite.

Work on the upper Karewa formation involved field description and mapping of facies, sedimentological analysis, dating using thermoluminescence (TL) and amino-acid racemization, and analysis of ostracod assemblages from lacustrine sediments. Areal restriction of the lake in Kashmir occurred about 0.4MaBP with the rapid uplift of the Pir Panjal Range. Sedimentological data show that aeolian dust formed a major input into the lake. Ostracod assemblages show that the lake itself was cool, shallow, alkaline and had abundant plant macrophytes, The lake drained between 120 and 80kaBP. Stratigraphical, sedimentological and faunal evidence suggests that this was a result of tectonically-induced drainage rather than climatically-induced desiccation.

The glacial history of the surrounding mountain flanks was reconstructed by field mapping of glacial sediments and dated using TL and radiocarbon methods. Present and past patterns of glaciation were assessed by the determination of equilibrium-line altitudes (ELAs), glaciation thresholds (GTs) and cirque altitudes. Glaciers extended to 2150 m a.s.l in the Great Himalayan flank and 2600 m a.s.l. in the Pir Panjal. There is evidence for only 2 pre-Holocene advances in Kashmir, the older of which predates 35kaBP. Present patterns of glacierization indicate a SW to NE rise in the height of ELAs and GTs suggesting topographic and precipitation control. An apparent reversal of trends during the past is explained by Quaternary uplift of the Pir Panjal Range.

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CHAPTER 1. THE SCOPE OF THE PRESENT STUDY.

1.1 Introduction.

Kashmir is an excellent area for the study of Late Cainozoic environmental change. Its location, at the fringe of current penetration of the South Asian summer monsoon, should make it a sensitive indicator of past changes in monsoon circulation. Furthermore, because it is located in the north-west Himalaya, which has been tectonically active during the Late Cainozoic, it should have recorded important palaeogeographical changes that have occurred due to tectonic uplift.

The valley of Kashmir is located between $33^{0}30'$ and $34^{0}30'$ north and $74^{0}10'$ and $75^{0}30'$ east in the northern Indian state of Jammu and Kashmir. The name Kashmir is used here to refer to the intermontane basin, although the political division covers a much wider area. The location of Kashmir basin is shown in figure 1. The basin is oriented north-west to south-east and developed within a series of thrust faults. It is about 180 km long and 120 km wide, with a mean basin-floor altitude of approximately 1600 m a.s.l.The Great Himalayan Range bounds the basin to the north-east. This range has a maximum altitude of about 5500 m a.s.l. and is largely a product of Mio-Pliocene orogeny. The Pir Panjal Range, which bounds the basin to the south-west, has a maximum altitude of about 4500 m a.s.l. This is largely a

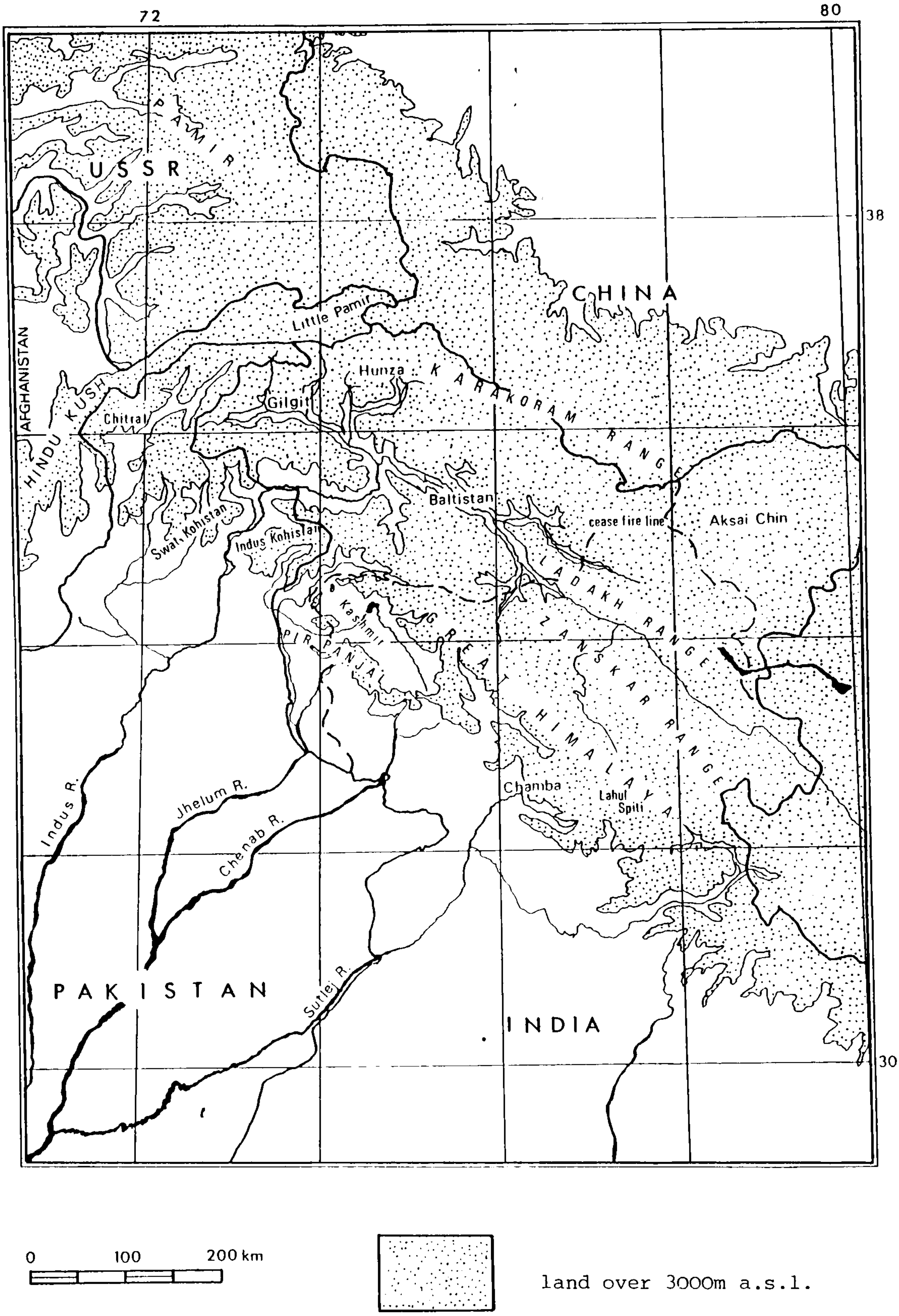


Figure 1

Location map of north-west Himalaya.

product of mid-Pleistocene thrusting along the Main Boundary Thrust Complex, which lies to the south-west of Kashmir.

Kashmir's marginal location with respect to the Asian summer monsoon is well illustrated by meteorological data. The absence of the monsoon in Kashmir is usually attributed to the rain-shadow effect of the Pir Panjal Range. At Jammu, immediately to the south of the Pir Panjal, about 750 mm of rainfall occurs during the summer monsoon months (July, August and September), which is about 67% of the annual total. At Srinagar, the main settlement in Kashmir, the total amount of rainfall occuring during the monsoon season is about 158 mm, or only 24% of the annual total.

There is abundant evidence to suggest that the South Asian monsoon circulation has undergone dramatic changes during the Quaternary. Work on oceanic sediments from the Arabian Sea and Bay of Bengal (eg. Cullen, 1981; Duplessy, 1982; Prell and Van Campo, 1986), lacustrine sediments in north-west India (eg. Singh et al., 1972), alluvial sequences from north-central India (Williams and Clarke, 1984) and numerical simulation models (eg. Manabe and Hahn, 1977; Kutzbach and Guetter, 1986) indicates a greatly weakened south-west (summer) monsoon at the last glacial maximum. There is also evidence that the north-east (winter) monsoon was enhanced at that time (Prell and Van Campo, 1986). However, there are few data from the north-west Himalaya itself to indicate palaeoenvironmental conditions at the last glacial maximum.

Opinions are divided as to when major uplift of the Himalaya took place. Some authors argue that substantial elevation occurred during the Palaeogene (eg. Patriat and Achache, 1984). However, there is increasing evidence to suggest that the bulk of uplift occurred during the latter part of the Pliocene, and the Quaternary (see, for example, Powell, 1986). If this is true, Kashmir has the potential to record the interplay between tectonic uplift and climatic change.

Over 1000 m thickness of sediments have accumulated in Kashmir basin, and these are exposed in the major river valleys on the Pir Panjal flank of the basin. These sediments are known as the Karewa group or 'Karewas'. They are essentially fluviolacustrine in origin and range from about 4.0 to 0.4 MaBP in age. Major lithological changes within the sequence, from fine-grained lacustrine sediments to coarse-grained fluvial conglomerate beds, reflect tectonic events along the basin margins (Burbank, 1982). Geological research on the Karewas began in the last century. However, substantial findings have only been made in the last ten years, with work by D.W. Burbank, an American geologist, and a team of scientists from the Physical Research Laboratory (PRL) in Ahmedabad, India.

The high surrounding mountain ranges of the Great Himalaya and Pir Panjal have preserved records of Quaternary glaciation. Despite numerous observations on supposed glacial deposits by many early geologists and explorers, the glacial record of Kashmir has not been reassessed in the light of current knowledge of mountain glaciation. Despite this lack of detailed knowledge, glacial tills are thought not to be present in the Karewas (eg. Singh, 1982; Burbank,1982) and direct evidence of glaciation is only found above the altitudinal limit of the basin-fill sediments. Furthermore, given the destructive nature of glacial advances, it is thought likely that glacial sediments found in the mountain flanks postdate much or all of Karewa sedimentation (Burbank, 1982). Archaeological remains have been found within sediments that overlie glacial tills in Kashmir (Joshi et al., 1974). Whilst the artefacts themselves have been studied by archaeologists, little is known about the age of the deposits in which they were found. Therefore, a reassessment of the Kashmir glacial record is of importance for archaeology.

1.2 Aims of This Thesis.

The extensive sedimentary record in Kashmir, and the complexities of interpretation, means that there are potentially many possible research problems that could be pursued. Following an initial field season of reconnaissance in Kashmir, four major topics were selected for detailed study. These topics are as follows.

1) To investigate environmental changes during the period of geological time during which the lower part of the Karewa group (the 'lower Karewa') was deposited, that is, the Late Pliocene to early Middle Quaternary.

2) To investigate environmental changes during the period in which the upper part of the Karewa group (the 'upper Karewa') was deposited.

3) To reevaluate the Late Quaternary glacial record of Kashmir.

4) To compare the results with other parts of Asia and with global data.

The methods used to investigate each part of the record were determined by the nature of the sedimentary evidence, the degree of resolution required and work in progress by other groups. The last of these is particularly important, since ongoing work by PRL scientists on the lower Karewa, particularly using pollen, diatoms, ostracods, and vertebrate fossils constrained the choice of techniques used in this study. The aim of studying the lower Karewa sequence was to investigate low frequency climatic changes occuring in the Late Pliocene to Middle Quaternary. For this part of the study, it was felt that the whole of the sequence should be sampled. Since the lower Karewa beds are over 1000 m thick, a relatively large sampling interval was adopted. Techniques of palaeoenvironmental reconstruction applicable to the whole of the sequence, which had not already been used, were required. Since the lower Karewa beds consist largely of mudstone - sandstone - conglomerate alternations, techniques based on the lithology and mineralogy of these facies were used. For the sandstones and conglomerates, scanning electron microscope (SEM) analysis of sand-grain microtextures was attempted. For the mudstones, semi-quantitative analysis of clay mineral assemblages was undertaken. As well as being hitherto unused on the lower Karewa sediments, these two techniques have the advantage of being applicable to the whole of the formation. This is in contrast to most of the other techniques previously used in studying this sequence. The chronology of the lower Karewa sediments has already been well established using magnetic polarity stratigraphy and fission track dating of volcanic ashes (Burbank, 1982; Agrawal, 1985).

The upper Karewa sediments are much thinner than the lower Karewa (<100 m), cover a shorter timespan and show fewer signs of intense tectonic disturbance. Therefore, it was felt that this part of the sequence might have preserved a more detailed and easily interpreted record of environmental change. The upper Karewa consists of three main facies: conglomerate, lacustrine silts and clays, and aeolian loess. For this part of the study, field logging and detailed sampling of a number of sections were carried out. Reconnaissance laboratory work showed that ostracods, which are microcrustaceans, were the only fossils found in abundance in the upper Karewa sediments. A detailed study of the ostracod fauna of one upper Karewa section forms the basis for palaeoenvironmental reconstruction. Extensive sedimentological work was also undertaken on the upper Karewa beds, particularly the lacustrine facies. The aeolian loess is the subject of a major research effort by PRL scientists, so was not studied in detail here. Because the chronology of the upper Karewa as a whole is poorly known, attempts were made to date the sequence.

Due to numerous studies on the glacial history of Kashmir during the early part of this century, the valley remains the type area for Himalayan glaciation. In this study, the glacial sequence was reassessed. Particular attention was placed on determining the nature of glacial and related sediments, and the magnitude and chronology of glacial advances.

For comparison of the Kashmir data with that from elsewhere, the most relevant studies include:

1) work on 'long' climatic sequences from Asia (eg. the Chinese loess) and elsewhere (eg. those deep-ocean cores that penetrate the Late Pliocene);

2) records of climatic change from other areas in Asia (eg. lakes in north-west India, alluvial sequences from north-central India and oceanic cores from the northern Indian Ocean);

3) other records of Himalayan glaciation;

4) climatic models.

1.3 Thesis Organization.

There are three discrete research aims in this thesis, each of which involves the use of separate techniques. Because of this, chapters 4 to 6 are relatively self-contained. There is no separate 'methods' chapter since most methods are used in only one part of the study. The details of individual methods are therefore discussed in the appropriate chapter in this part of the thesis. The thesis is divided into three parts and eight chapters. Part one comprises the present chapter and chapters 2 and 3. Chapter 2 reviews the environmental background to Kashmir: geology, climate and vegetation. Chapter 3 is a review of research into Late Cainozoic environments in Kashmir. Part 2 deals with the analysis of data and discussion of results. Chapter 4 is concerned with the lower Karewa, chapter 5 with the upper Karewa and chapter 6 with the Late Quaternary glacial history. Part 3 is concerned with synthesis and conclusions. Chapter

7 summarizes data from this thesis together with the results from other studies, in order to put forward an environmental history of Kashmir for the Late Cainozoic. It then considers this synthesis in a regional and, where appropriate, a global context. Chapter 8 examines areas for possible future research and lists the general conclusions from this study.

CHAPTER 2. PRESENT-DAY ENVIRONMENTS.

2.1 Geology.

a. Structural Evolution of North-west Himalaya.

The Himalayan Range is a result of the collision between the Indian and Asian continental plates (LeFort, 1975). The precise timing of this event remains uncertain, although intial collision may have occurred as early as 110 or as late as 50 MaBP. Between 55 and 40 MaBP, there was a major decrease in the rate of northward migration of the Indian subcontinent, from between 20 and 15 cma^{-1} to between 6 and 4 cma^{-1} (Windley, 1984). Shortening of the continental crust following collision is thought to have occurred in a number of ways. These include transcurrent faulting in the foreland of the indented Asian plate; crustal thickening in Tibet; underthrusting of the Indian plate under Tibet along the Indus-Zangbo suture; and extensive thrusting along the northern edge of the Indian plate (Windley, 1984).

Much of north-west Himalaya has been subjected to thrusting. The imbricated thrust sheets make up the Himalayan *schuppenstruktur*. Cainozoic sedimentation has occured in intermontane basins such as Kashmir, due to the ponding of fluvial drainage by thrusting, and in the Indo-Gangetic foredeep along the southern margin of the Himalayan Range. Various seismic and geomorphological investigations suggest that a series of thrust faults dips northwards beneath Kashmir (eg. Kaila *et al*., 1978; Seeber and Armbruster, 1979). The Kashmir Basin is located to the east of the north-west syntaxis. The syntaxis is defined by sharp bends in the trend of thrust faults, and by the convergence of the major structural features of the Himalayas, Karakoram and Hindu Kush (Wadia, 1932). There are three main thrust-fault complexes within North-west Himalaya: the Main Mantle Thrust (MMT); the Main Boundary Thrust (MBT); and the Main Frontal Thrust (MFT). The location of these faults and the structural setting of Kashmir are shown in figure 2.

Little recent work has been carried out on the mapping of faults in Kashmir. However, some of the earlier geologists mapped faults in the basin. For example, Middlemiss (1910) mapped a fault running along the north-eastern edge of the Pir Panjal Range. Although the exact age of the fault is unknown, it clearly postdates basin sedimentation since it dips below the Karewa deposits (Burbank, 1982). Wadia (1928, 1931, 1934) delineated two sets of north-dipping, imbricated thrust faults along the south side of the Pir Panjal Range and suggested that uplift of the range had occurred as a result of tectonic activity along the faults.

Faults have been inferred, although not physically identified, beneath the Himalayan flank of Kashmir. Chandra's (1978, 1979) seismic data suggest that active thrusting occurred beneath this flank. Seeber _et al_. (1981) inferred north-dipping faults beneath the Great Himalaya on the basis of seismic and geomorphological data. They suggested that activity along the faults was responsible for the abrupt topographic rise of the mountain front along the basin margin.

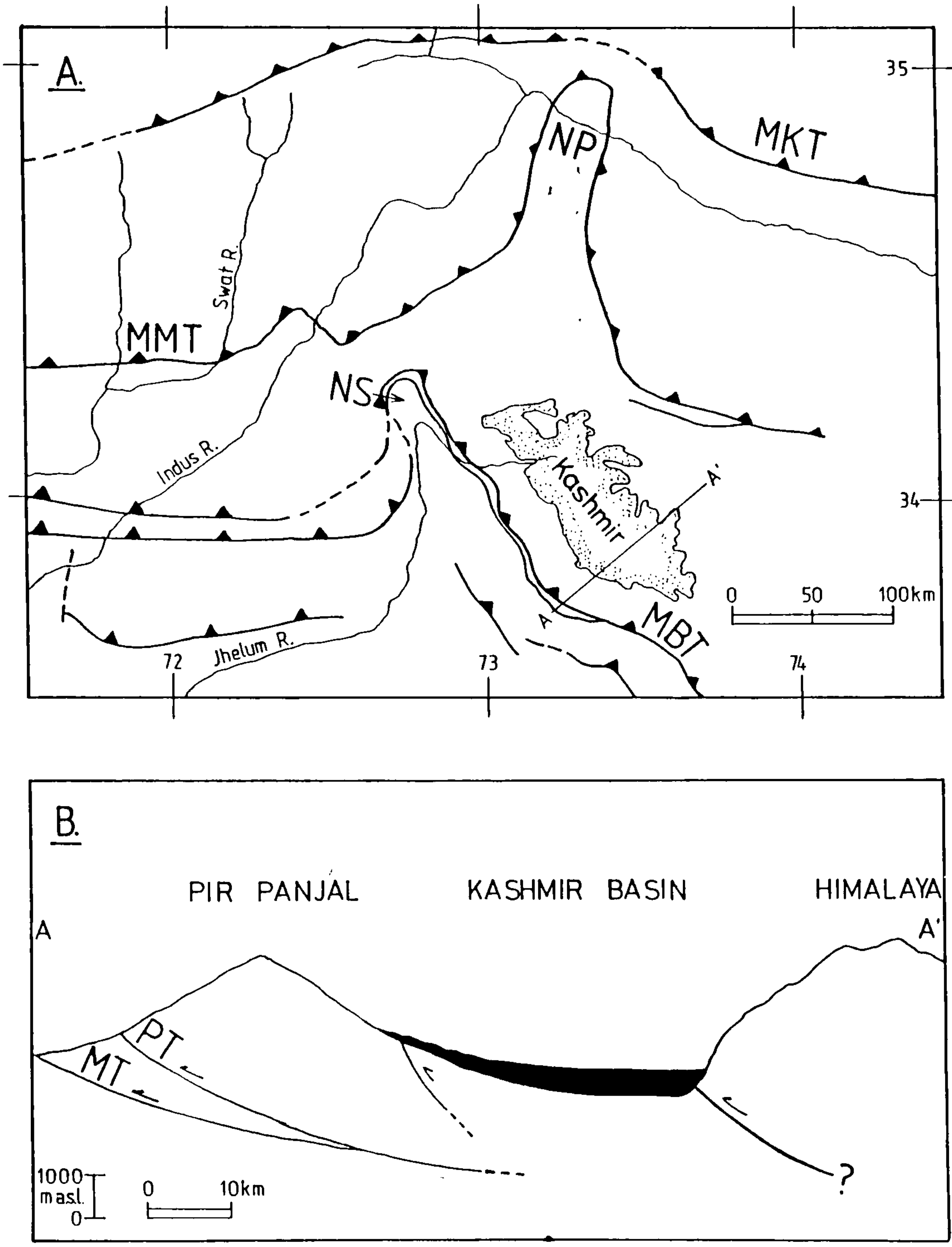


Figure 2

The tectonic setting of Kashmir. A. Map showing the major fault systems in north-west Himalaya. B. Cross-section of Kashmir, showing the major structural features. MBT = Main boundary Thrust, MMT = Main mantle Thrust, MKT = Main Karakoram thrust, NS = Nanga Parbat syntaxis, NS = North-west syntaxis, PT = Panjal Thrust, MT = Muree Thrust. From information in Middlemiss (1910), Wadia (1934) and Burbank (1982).

Burbank (1982) developed a chronological model for the tectonic evolution of Kashmir basin and the deposition of the Karewa beds. He argued that the initial ponding of drainage, and the subsequent formation of a large lake, resulted from the limited uplift of the Pir Panjal Range about 4 MaBP. Periods of fine-grained, lacustrine deposition, which represent quiescent phases, were punctuated by fluvial-conglomeratic deposition, inferred to represent active thrusting. From the onset of basin sedimentation, about 4 MaBP, until around 1.7 MaBP, the source of conglomerates was the north-eastern flank of Kashmir. This source was inferred by Burbank (1982) on the basis of clast imbrication in conglomerates and cross-bed orientation in interbedded sandstones. After 1.7 MaBP, there was a switch to a south-western source. Burbank (1982) interpreted the change in source area as a switch from tectonic activity along the north-east basin margin prior to 1.7 MaBP, to activity along the MBT, on the south-western margin, after 1.7 MaBP. Basin-wide sedimentation was terminated at around 0.4 MaBP by the rapid uplift of the Pir Panjal Range.

b. Rates of Uplift.

Rates of uplift of the Himalayas are of critical importance in this study for three reasons. Firstly, the substantial uplift of a mountain range of the magnitude of the Himalayas is likely to have had considerable effects on regional, and even global, climate. Therefore, in order to understand the links between tectonic uplift and climatic change, it is necessary to reconstruct the rate and timing of both. Secondly, the onset of glaciation in the Himalayas probably occurred when the range attained a critical altitude. Therefore, rates of uplift are an important line of evidence in determining when the range was first glaciated. Thirdly, it is important to identify areas and rates of Late Quaternary uplift since this will influence observed patterns of glacial advances.

Several methods have been used to calculate rates of Himalayan uplift. These include rock cooling-histories based on fission-track ages (Zeitler, 1985; Zeitler et al., 1982a, 1982b; Sharma, 1984); palaeontological data (Liu and Ding, 1984; Lakhanpal et al., 1983); palaeomagnetic evidence (Burbank, 1982); rates of erosion (Menard, 1961) and geodetic surveys (Narain, 1975). The timescales over which these rates apply vary from the present-day to periods of more than 40 Ma. For this study, it is fortunate that there are a number of published rates of Quaternary uplift of North-west Himalaya. In addition, there is increasing evidence for the rate of uplift of the range as a whole during the Cainozoic. However, the question of when the Himalayas attained substantial relief is still open to debate. The available evidence for rates of Himalayan uplift is tabulated in Table 1. Although the data are sparse, both temporally and spatially, they do provide some useful constraints on the timing and magnitiude of Himalayan orogeny that are of value for the present study.

Some tectonic models for the Himalayas and Tibetan Plateau imply that substantial relief was attained as early as the Eocene (eg. Patriat and Achache, 1984). Molnar (1984) argued that evidence for high rates of recent uplift only applies to the front of the

TABLE 1. RATES OF HIMALAYAN UPLIFT.

LOCALITY	METHOD	RATE (mm a^{-1})	PERIOD (MaBP)	SOURCE
Tibetan Plateau	Palaeontology & Geomorpholgy	1.5-2.0	2 - 0	1
Ladakh	Palaeontolgy	~0.1	38 - 14	2
		~1.0	3 - 0	2
Pir Panjal	Palaeomagnetism	3.4-10	0.4 - 0	3
Chitral	Fission Track	0.42	8.7 - 0	4
		0.19	18.2 - 8.7	4
Dir and northermost Swat	Fission Track	0.24	15.0 - 0	4
		0.07	41.8 - 15.6	4
Northern Swat	Fission Track	0.39	9.4 - 0	4
		0.18	19.7 - 9.4	4
		0.22	20 - 10	4
		0.08	55 - 20	4
West-Central Kohistan Arc	Fission Track	0.22	15.9 - 0	4
		0.07	44.5 - 15.9	4
S. Swat - Hazara	Fission Track	0.20	18.1 - 0	4
		0.39	22.8 - 18.1	4
N. Hazara	Fission Track	0.68	5.4 - 0	4
		0.13	19.2 - 5.4	4
Kaghan	Fission Track	0.75	4.9 - 0	4
		0.18	15.0 - 4.9	4
Nanga Parbat-Haramosh Massif	Fission Track	4.5	0.7 - 0	4
	Fission Track	1.6	2.0 - 1.7	4
	Fission Track	1.1	3.4 - 2.0	4
	Fission Track	0.3	6.8 - 3.4	4
East-central Kohistan Arc	Fission Track	0.5	7.3 - 0	4
	Fission Track	0.2	16.3 - 7.3	4
N.E. Kohistan Arc	Fission Track	0.87	4.2 - 0	4
	Fission Track	0.20	13.6 - 4.2	4
Pandoh-Baggi NW Himalaya	Fission Track	0.55	8 - 0	5

TABLE 1. CONTINUED.

LOCALITY	METHOD	RATE (mm a^{-1})	PERIOD (MaBP)	SOURCE
Kinnaur	Fission Track	0.08	20 - 31	5
NW Himalaya		0.15	11 - 20	5
		0.8	7 - 15	5
Chor	Fission Track	0.13	69 - 82	5
NW Himalaya		0.10	23 - 69	5
Karsog		0.13	12 - 60	5
NW Himalaya				
Dalhousie	Fission Track	0.10	6 - 74	5
Lesser Himalaya				
Ramban		0.3	8 - 22	5
Lesser Himalaya				
Rohtang	Fission Track	0.8	5 - 8	5
		0.8	11 - 13	5
		0.6	4 - 11	5
Lesser Himalaya	Geodetic survey	0.8	past 75a	5
Lesser Himalaya	Rb-Sr Age	0.7 - 0.8	15 - 0	6
Himalaya	Erosion Rate	0.2	40 - 0	7
E. Himalaya	Erosion Rate	0.62	present	7
W. Himalaya	Ersoion rate	1.00	present	7

<u>SOURCES</u>

1. Powell, 1986.
2. Lakhanpal <u>et</u> <u>al</u>., 1983.
3. Burbank, 1982.
4. Zeitler, 1985.
5. Sharma, 1984.
6. Mehta, 1980.
7. Menard, 1961.

range. However, recent evidence from Chinese scientists suggests that substantial uplift of the Tibetan Plateau may have occurred much more recently, in the Neogene rather than Palaeogene (see Powell, 1986, for a review). This evidence includes the occurrence of Hipparion fauna in elevated late Pliocene sediments and the existence of palaeokarst and well-developed erosion surfaces suggesting long periods of subdued relief during Eocene to Pliocene times. Models have been developed which account for this rapid uplift (eg. Klootwijk et al., 1985; Powell, 1986). They suggest not only that the idea of a 'high' plateau existing 40 MaBP is untenable, but that between 3 and 4 km of uplift of Tibet has occurred since 2 MaBP. This obviously has far-reaching implications for the Late Tertiary and Quaternary climatic evolution of Asia. However, it is not yet possible to place closer constraints on the uplift of the Himalaya and Tibetan Plateau as a whole.

Fission-track ages reported by Sharma (1984) and Zeitler (1985) for a number of areas in North-west Himalaya (see table 1) suggest that uplift rates have increased during the Late Cainozoic. This is supported by several other lines of evidence. On the basis of fossil plant finds in Tertiary rocks of Ladakh in North-west Himalaya, Lakhanpal et al.(1983) calculated a tentative uplift rate of about 0.1 mma^{-1} for the Oligocene and Middle Miocene, increasing to about 1 mma^{-1} for the period since the beginning of the Pliocene. Chronologies for intermontane-basin and molasse sediments indicate a migration of uplift and tectonic activity from the MMT during the Middle Cainozoic to the MBT during the Plio-Pleistocene (Burbank and Reynolds, 1984).

There is further evidence of uplift rates for North-west Himalaya for the Late Pliocene and Quaternary. Using fission-track dates on zircons, sphenes and apatites, Zeitler (1985) found high rates of Middle and Late Quaternary uplift (4.5 mma^{-1}) for an area centred on the Nanga Parbat syntaxis (see table 1). The zone of rapid uplift was found to extend beyond the MMT and onto the Asian landmass. To the south of the Nanga Parbat Massif, much lower rates of uplift were found. Lower rates of Late Pliocene and Quaternary uplift were also found for Swat Kohistan based on fission-track evidence. This is substantiated, for the Late Quaternary, by the glacial geology of the Swat Valley. Porter (1970) has shown that moraines of the penultimate glaciation were not overrun by those of the last advance, suggesting that uplift between the two advances was limited.

Burbank (1982) calculated rates of uplift for the Pir Panjal Range in Kashmir, based on altitudinal displacement of palaeomagnetically-dated Karewa beds. Uplift rates of between 3.5 and 10 mma^{-1} were suggested for the period between 0.4 and 0 MaBP. In Burbank's tectonic model of Kashmir, relative tectonic stability of the Great Himalayan margin is inferred for this period. Pulses of uplift of this flank were dated at around 3 to 3.5, 2.7, and 2.1 MaBP on the basis of conglomerate beds within the Karewa sediments. However, because dated, uplifted Karewa beds are not found in the Himalayan flank, no constraints could be placed on the magnitude and rate of uplift.

c. Pre-Cainozoic Geology of Kashmir.

Research into the bedrock geology of Kashmir has been in progress since the end of the last century. As such, the spatial and stratigraphic arrangement, and lithological character, of the main rock types is quite well known. A number of geological maps of the whole of Kashmir have been compiled (Gansser, 1964; Fuchs, 1975; Geological Survey of India, 1977; Shah, 1978; Thakur and Gupta, 1983) and there are a number of more detailed surveys of small areas within the basin (eg. Hashimi, 1971; Shah, 1972) and certain parts of the stratigraphic record (eg. Gupta, 1969; Fuchs and Gupta, 1971; Srikantia, 1973). Although many uncertainties remain (eg. Saxena, 1973; Fuchs, 1975) the geological background is adequately known for many of the purposes of this research.

The strata of Kashmir form a synclinorium and because of thrusting, the oldest, Precambrian, rocks overlie Permo-Carboniferous strata (Fuchs and Gupta, 1971). The youngest rocks found in the Kashmir basin are Jurassic. No equivalents of the Palaeogene to Early Neogene Muree and Siwalik beds of the Indo-Gangetic foredeep are found in Kashmir. A geological map and generalized stratigraphic columns for Kashmir are shown in figures 3 and 4 respectively. The lithological character of each unit is summarized from the literature below.

The Precambrian age of the oldest rocks in Kashmir, the Salkhala series, is based on the oldest overlying fossiliferous strata being of Lower Cambrian age (Fuchs and Gupta, 1971). The major lithologies found in the Salkhala series are phyllite and schist, although

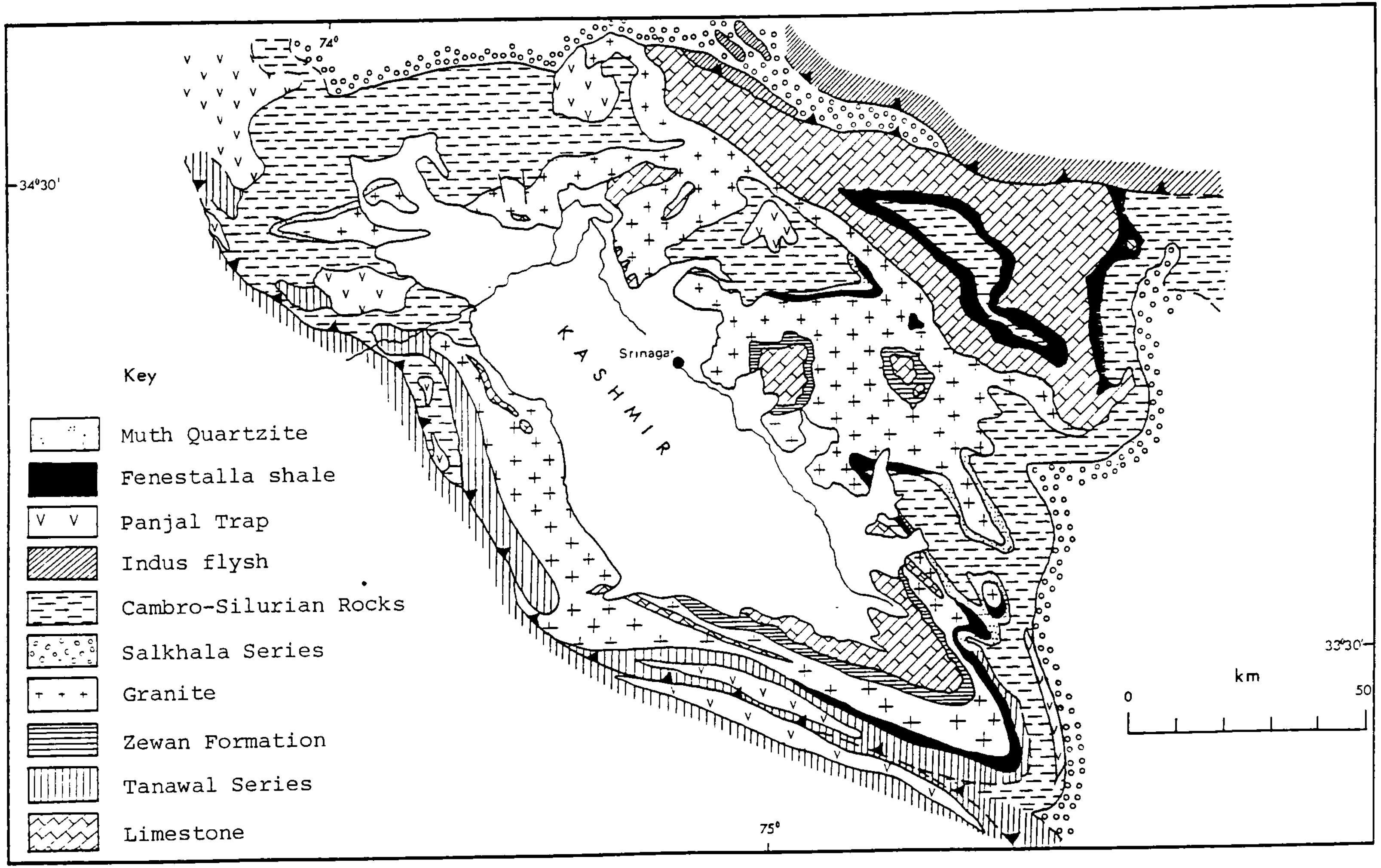


Figure 3
Simplified geological map of Kashmir.

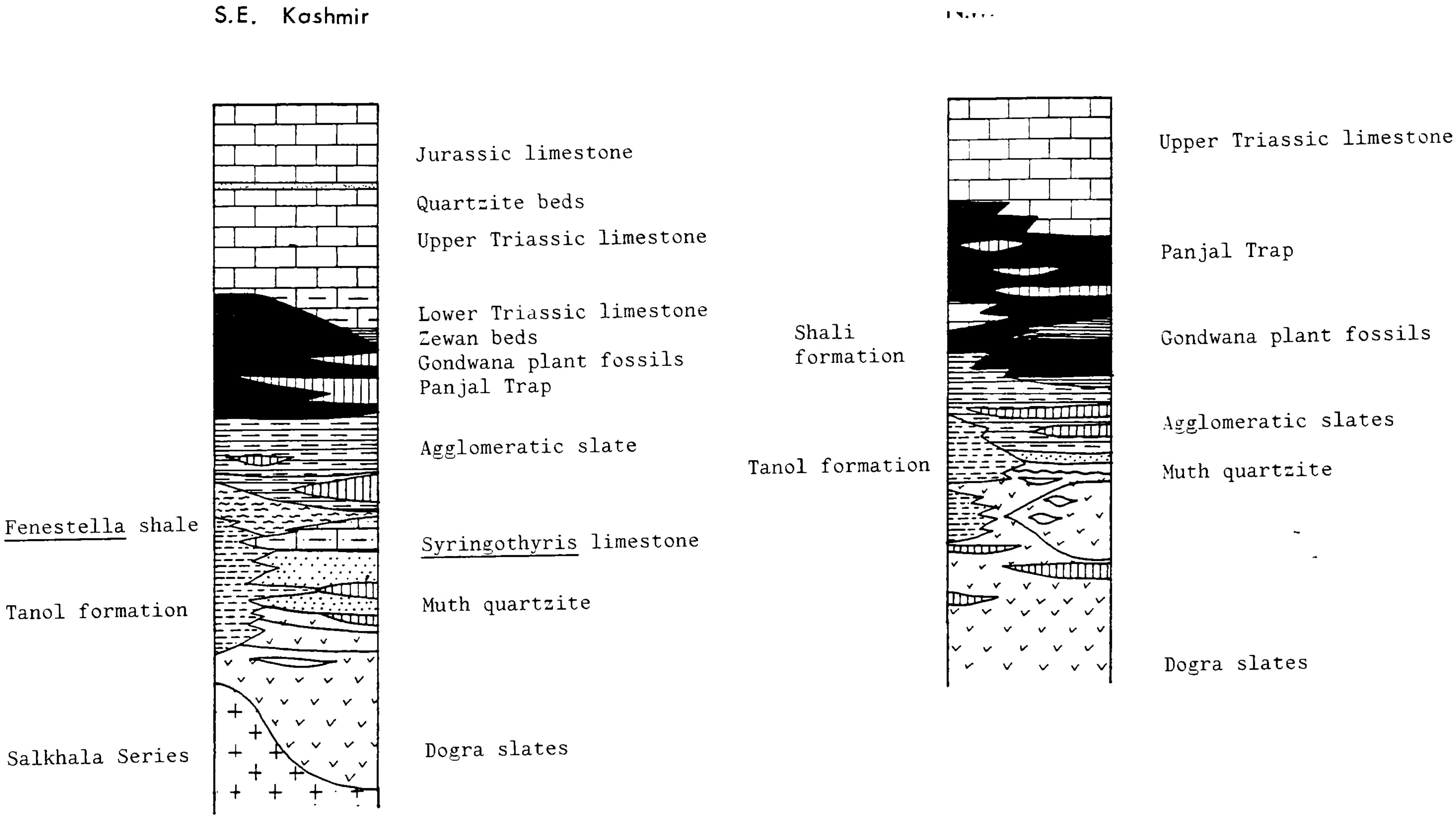


Figure 4.Generalized stratigraphic sections of the Pre-Cainozoic geology of Kashmir (after Fuchs, 1975)

limestone, dolomite and marble are also found. Granites associated with these rocks may be intrusive, or even older than the Salkhalas (Geological Survey of India, 1977).

In the western and south-western part of Kashmir, rocks of the Salkhala series are overlain by Dogra slates (Wadia, 1928). These beds are unfossiliferous, but are regarded as Precambrian to lowermost Cambrian in age. Lithologically, the Dogra slates comprise black to greenish slate and greenish phyllite, with occasional beds of quartzite and trap (Fuchs and Gupta, 1971). The last-mentioned of these is a product of contemporary volcanism.

Cambrian rocks are best exposed in the Baramulla area, where an unfossiliferous sequence, mainly of slates, is found. Elsewhere, unfossiliferous shales, sandstones, clays and limestones that occur beneath Ordovician-Triassic beds are considered to be of Cambrian age (Fuchs and Gupta, 1971).

Beds of Ordovician and Silurian age are found in the north-west and south-east parts of Kashmir valley, and also in the lower Sind Valley. These beds consist of ferruginous slates and quartzose greywacke. According to Fuchs and Gupta (1971), there is little lithological variation in the Precambrian to Ordovician strata. The rocks consist mainly of slates, shales, sandstones, siltstones and greywackes. Calcareous beds are rare.

Devonian age Muth Quartzite is found above the Silurian strata. It is assigned to the Devonian on the basis of fossil evidence. Where Silurian rocks are absent, the quartzite overlies Cambrian strata. Lithologically, it is a hard, white quartzite which is massive and, in parts, calcareous. The Tanawal series, found especially in north-western Kashmir, is also found to be of Devonian age (Geological Survey of India, 1977) although the rocks are unfossiliferous (Fuchs and Gupta, 1971). Lithologically, the Tanawals comprise quartzite, quartzose sandstone, phyllite, schist, gritstone and conglomerate.

Devonian strata are overlain by rocks of Carboniferous age. These include Syringothyris limestone, Fenestella shale and Panjal volcanics. Syringothyris limestone is a grey to dark-bluish, hard, flaggy limestone with bands of shale and quartzite. Intrusions of volcanic trap are sometimes found (Fuchs and Gupta, 1971). Fenestella shale consists of thickly-bedded, fossiliferous sand, and micaceous and calcareous unfossiliferous quartzite. Fenestella shales grade up into the agglomeratic slates of the Panjal volcanics and are overlain locally by Panjal Trap.

The agglomeratic slates and Panjal Trap constitute the Panjal volcanics. The presence of glass within the former lithology suggests a volcaniclastic origin, but with substantial subaerial reworking (Geological Survey of India, 1977). Lithologically, the agglomeratic slates comprise slates, sandstones, quartzites, conglomerates and tilloids. Rarely, limestone beds are found (Fuchs and Gupta, 1971). Panjal Trap is a greeenish, andesitic and basaltic lava of Upper

Carboniferous to Upper Triassic age. The greenish colour is imparted to the rock by chloritised and epidotised ferromagnesian minerals (Geological Survey of India, 1977). Panjal Trap is frequently overlain by Gangamopteris beds, which contain Gondwana fossils. Lithologically, these beds comprise cherts, siliceous shales, carbonaceous shales, thin-bedded limestones and flaggy quartzites. The uppermost Palaeozoic rocks of Kashmir are the marine fossiliferous limestones and shales of the Zewan series.

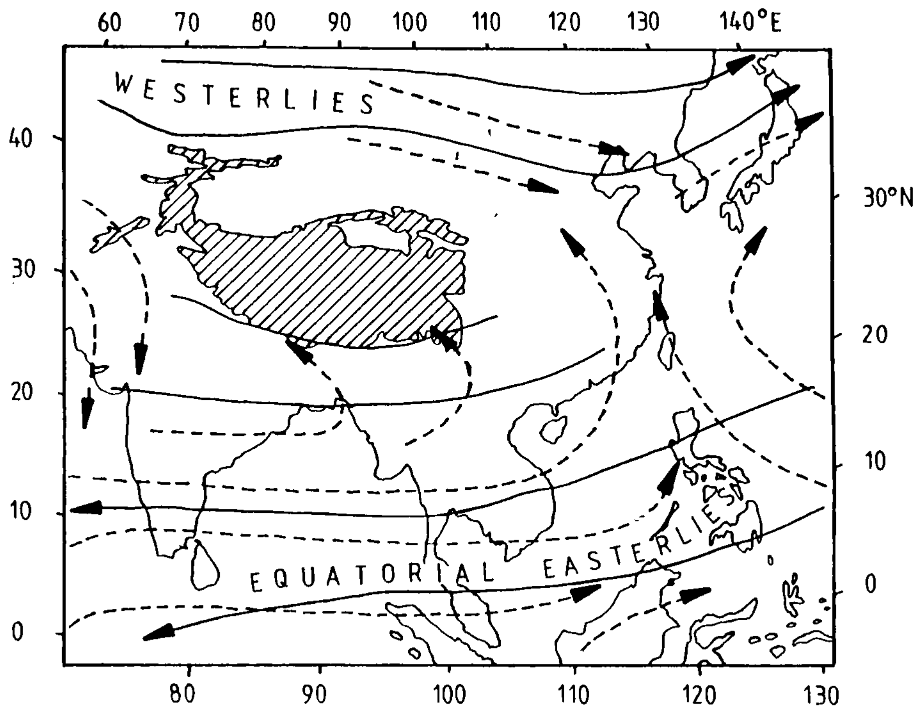
The Palaeozoic strata are overlain by Mesozoic rocks. These are mainly of Triassic age, although Jurassic rocks are also found. The Triassic rocks consist of limestones, argillaceous limestones and dolomites (Geological Survey of India, 1977). Jurassic strata are present in the Baltal syncline, which is in the upper Sind valley. These limestones and dark, pyritous shales are the youngest pre-Cainozoic rocks found in Kashmir.

2.2 Climate.

a. Present-day Climatic Systems.

Although Kashmir receives little of its precipitation during the summer-monsoon period, its climatic system forms part of the monsoon circulation. The South-Asian monsoon has been the subject of considerable study over the last 20 years. Although there are still many uncertainties, it is now known that the classical model of the monsoon as a giant, thermally driven land-sea breeze is inadequate. Current theories emphasize the importance of topography, and the relationships between the upper and lower troposphere, as well as thermal considerations. The South Asian year can be divided into summer and winter seasons on the basis of rainfall genesis. However, it is more usual to identify four seasons: the cool, winter season; the hot, pre-monsoon season, the season of monsoon rainfall; and the season of monsoon retreat.

During the winter season, an intense high-pressure cell develops over northern Asia, between about 40 and 60^{0}N. The high albedo, due to snow cover in this area, leads to very low temperatures and outflow of cold air to the south and south-east (Nieuwolt, 1977). However, the Himalayas and Tibetan Plateau form a protective barrier against the cold air and prevent it from reaching northern India. Although intense, the high pressure cell is shallow, and is overlain by a trough at the 700 mb level. Upper tropospheric airflow is dominated by the subtropical westerly jet at the 200 mb level. The upper westerly airflow bifurcates in the Tibetan Plateau region, with one jet-stream flowing to the north, and the other following the Himalayan front, across northern India. Although the effect of the Tibetan Plateau has often been cited as an explanation of this bifurcation, the northern branch of the jet is often found well to the north of the Plateau, and the jet often bifurcates well to the west of the topographic barrier (Barry and Chorley, 1982). The surface circulation is characterized by dry, outblowing winds: the so-called north-east or winter monsoon. This outflow originates from subsidence from the upper westerlies and subtropical anticyclone, situated over north-west India. The elements of winter circulation over Asia are summarized in figure 5.



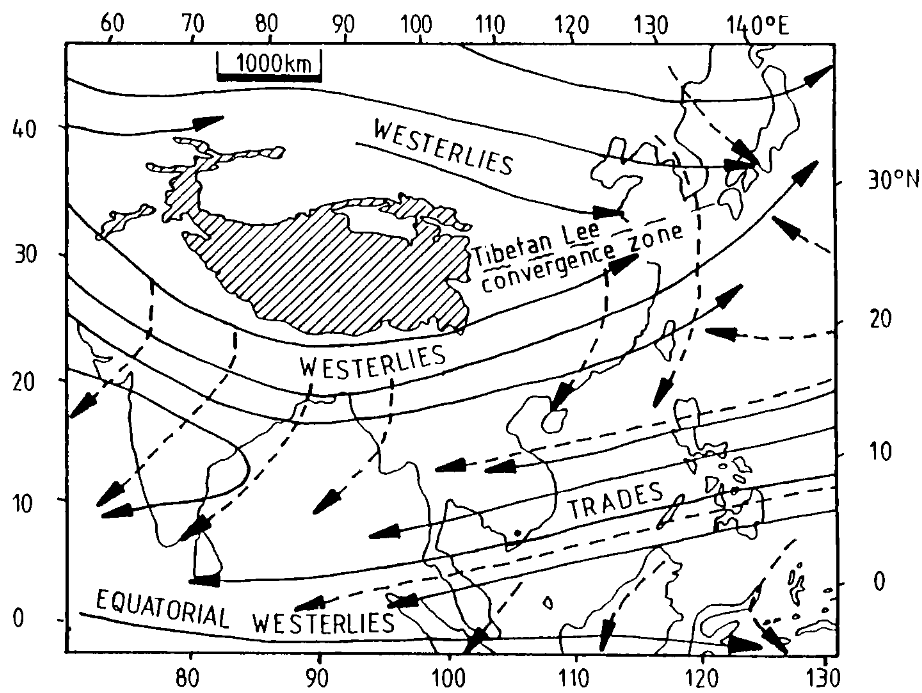


Figure 5

Monsoon circulation over Asia during (A) summer and (B) winter.

Depressions, steered by the southern branch of the upper westerly jet, are important sources of winter precipitation for Afghanistan, Pakistan and northern India. This precipitation is experienced in north-western and western Himalaya, but diminishes progressively towards the east. These 'Cyprus' depressions tend to be non-frontal and appear to originate over the Middle East. Pisharoty and Desai (1956) suggested that the life history of these depressions is similar to that of Atlantic and Pacific mid-latitude cyclones and that on their arrival over northern India, they are in an occluded state. Datta and Gupta (1967) proposed that frontogenesis results from the advection of air from a cold, surface high-pressure cell over the Ukranian region, to areas of warm, moist air over the Iranian Plateau. The depressions are then steered by the upper westerlies, with subsequent obliteration of surface fronts occuring, possibly as a result of their movement over the uplands of Afghanistan and Pakistan (Pisharoty and Desai, 1956). The depressions tend to move quite rapidly, up to 10 degrees of longitude per day (Datta and Gupta, 1967).

During the spring season of March to May, the pattern of upper airflow undergoes a transition. Due to increasing insolation, surface temperatures rise, and a thermal low-pressure cell situated over north-west India intensifies. As a result of this, the upper westerlies begin to migrate northwards. The southern branch of the upper westerly jet remains to the south of the Tibetan Plateau at this stage, but it is reduced in intensity. A number of mechanisms may cause limited precipitation over the subcontinent during this season.

These include westerly disturbances, tropical disturbances from the Bay of Bengal and the northward migration of the equatorial trough.

By the end of May, the southern branch of the westerly jet has migrated to the north of the Tibetan Plateau and the equatorial trough moves, reaching a maximum northerly position of 25°N by mid-July. A tropical easterly jet stream becomes established at the 150 mb level over Asia, at about 15°N. At surface level, humid, south-westerly air arrives and the monsoon 'breaks' with the onset of heavy rains. Although the changes in upper and lower tropospheric circulation are clearly linked, it remains unclear as to which is the primary causative mechanism (Walker, 1972). However, it seems that the role of the Tibetan Plateau in the change in circulation from winter to summer is crucial, although complex. Flöhn (1950) first suggested that the Tibetan Plateau, with an average altitude of 4000 m, might act as a high-level heat source. He argued that the excess of radiation received by the plateau would generate a thermally-driven upper-tropospheric anticyclone at 150 mb, with a shallow heat-low near the surface. The existence of this anticyclone has been demonstrated, and its development seems to be associated with the breakdown of the southern branch of the westerly jet and with the upper easterlies over northern India. The upper easterlies are, in turn, closely associated with onset of summer-monsoon rains. The circulation over Asia during the summer is summarized in figure 5.

The pattern of rainfall during the summer monsoon varies spatially, although much of the subcontinent receives more than 80% of its rainfall during this period (Rao and Ramamurthy, 1960). On the west coast of India, the major source of precipitation is the quasi-stationary, mid-tropospheric trough centred on Bombay. The development, intensification and dissipation of this system over periods of one to three weeks leads to periodic variations in precipitation over this area. Findlater (1969) has shown the existence of a low-level jet stream in the western Indian ocean, that crosses the equator in the region between 38 and 55°E. He suggested that variations in the speed of this flow may control summer rainfall in western India.

In central and north-eastern India, monsoon depressions are the major source of moisture. They form over the northern Bay of Bengal and track northward, north-eastward and west north-westward, bringing heavy rain to the adjacent landmass (Walker, 1972). These disturbances seem to form due to upper tropospheric divergence on the poleward side of high velocity cores in the tropical easterly jet over the Bay of Bengal.

During autumn, the summer monsoon circulation breaks down and there is a generally rapid reestablishment of the westerly jet to the south of the Tibetan Plateau. This change in circulation is accompanied by cool-season conditions over India. Easterly trades affect the Bay of Bengal at the 500 mb level and generate cyclonic disturbances at their confluence with the equatorial westerlies.

In contrast to the large summer component of rainfall in much of peninsular India, Kashmir receives only a small part of its precipitation during summer. The chief source of rainfall occurs during winter. Westerly depressions are the main bearers of rain and snow from November to February. This season is also characterized by high relative humidity and little bright sunshine (Raza et al., 1978). Snow is most frequent between December and February. The daily average snowfall at Srinagar is at a maximum (about 5 cm) during late January and early February (Bhan, 1954).

During the spring season in Kashmir, from March to mid-May, an increase in insolation leads to a rise in temperature. There is a high incidence of westerly depressions crossing Kashmir in early spring and March is the wettest month. However, towards the end of spring, northward migration of the westerly jet stream brings about a sharp decline in the number of westerly disturbances.

During the summer period, from mid-May to mid-September, the Pir Panjal Range prevents major incursion of monsoon rain into Kashmir valley. However, around 20% of the rainfall received at Srinagar falls during this period (Kaul, 1980). Ananthakrishnen and Bhatia (1960) mapped the tracks of monsoon depressions. They found that, occasionally, these depressions recurved from their north-westward track to a north and north-eastward path, across the Punjab and into Kashmir. This recurvature was linked to interactions between a deep trough situated over the Tibetan Plateau, and the monsoon depression moving towards the thermal low over north-west India. However, such occurrences are rare, and Ananthakrishnen and Bhatia (1960) identified only 3 over 30 years. On the other hand, Rao (1960) has suggested that the monsoon depressions influence rainfall over a wide area. Summer rainfall in Kashmir may, therefore, be linked to such depressions as they follow their normal course towards north-west India. Pisharoty and Desai (1956) suggested that the westerly movement of troughs over northern India during summer may also lead to the pulsatory extension of monsoon rainfall into Kashmir.

b. Climatological Data For Kashmir.

Meteorological observations have been made in Kashmir for much of this century. However, long records for both temperature and precipitation exist only for Srinagar. At Gulmarg, data are available only for the summer months. Recently, however, a new station has opened at the Gulmarg Research Institute, where observations are made all year round. Snowfall is only recorded at Srinagar.

Although there are many raingauges in Kashmir, the records are erratic. A raingauge station may close and reopen several years later for no apparent reason. For some years, many of the data for a particular station may be missing. For the purposes of this study, all of the available Indian Meteorological Department records were consulted. The locations of raingauges in Kashmir are shown in figure 6. Graphs of mean monthly precipitation are shown in figure 7 and temperature data for Srinagar listed in table 2.

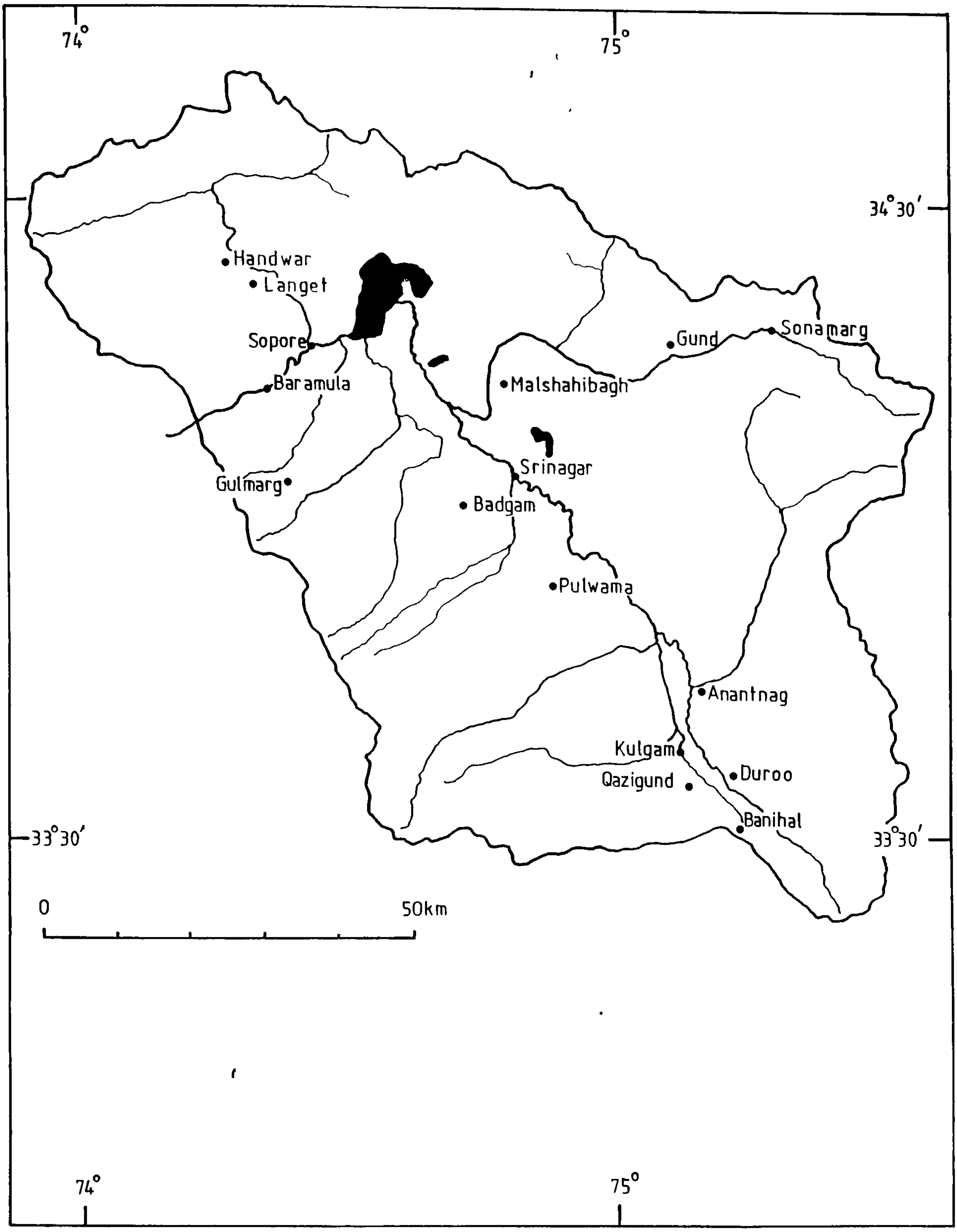


Figure 6

Location of meteorological stations in Kashmir.

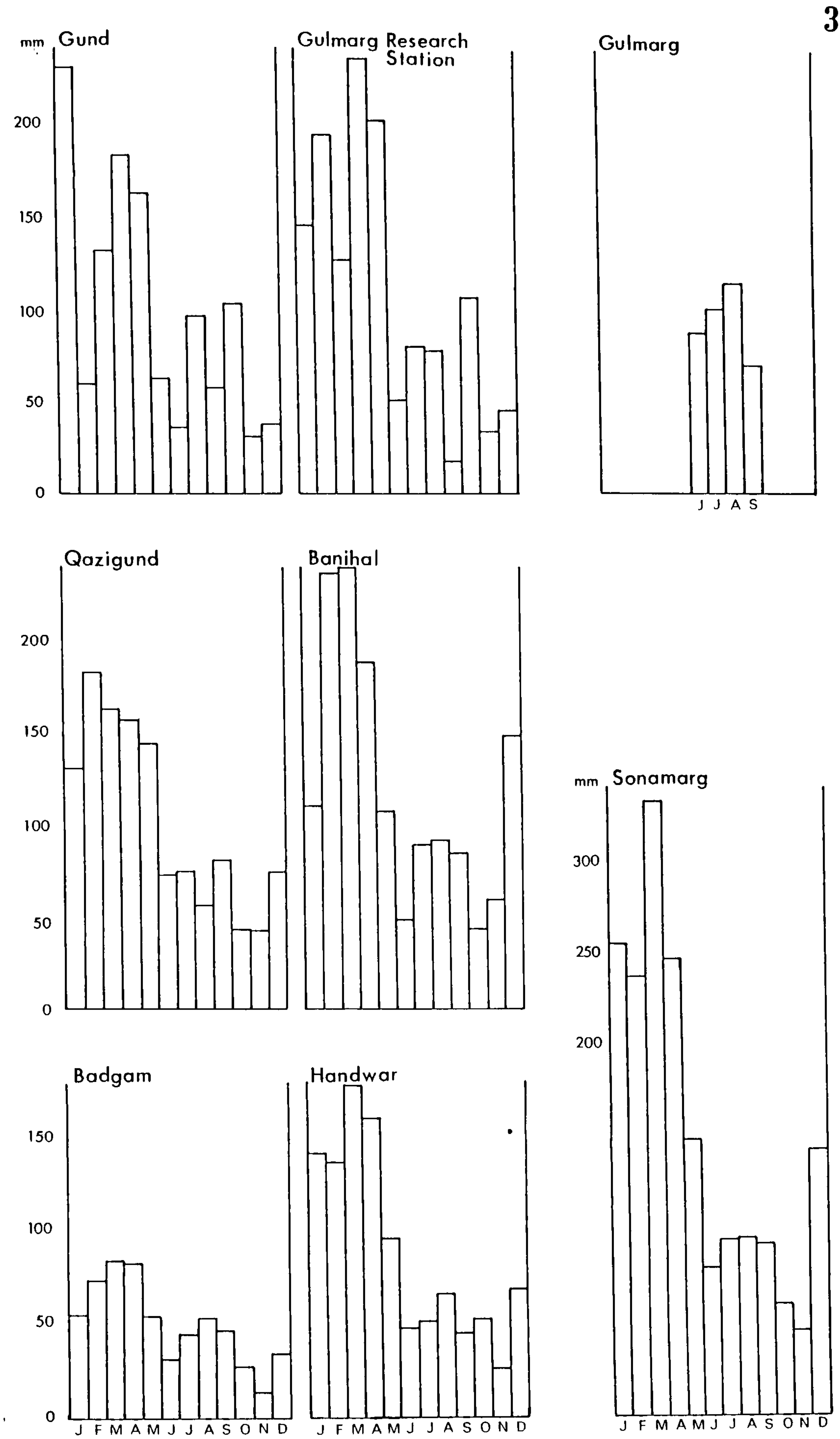


Figure 7

Climatic statistics for meteorological stations in Kashmir.

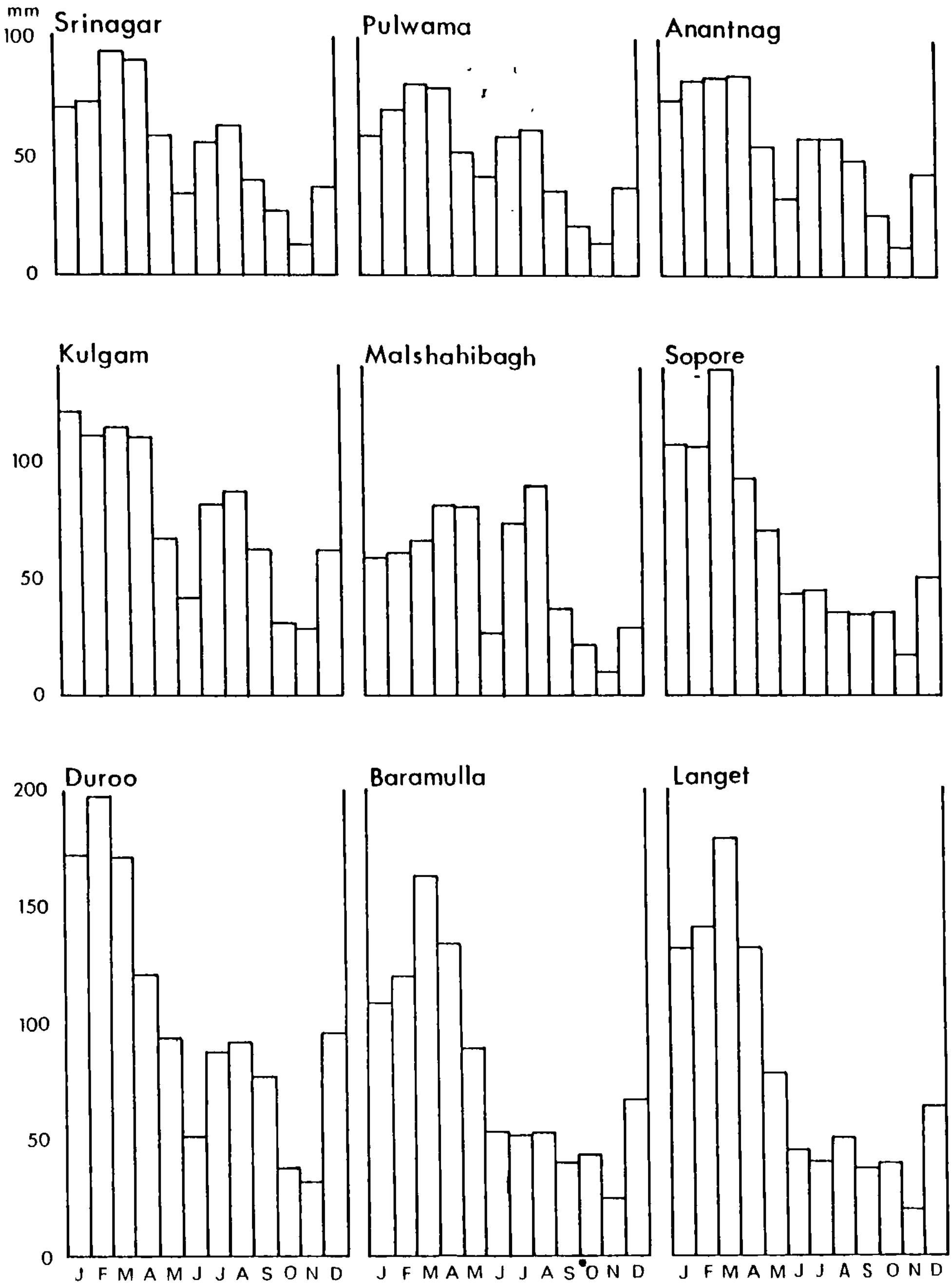


Figure 7 (continued)

Climatic statistics for meteorological stations in Kashmir.

TABLE 2. TEMPERATURE DATA FOR SRINAGAR (From Indian Meteorological Department (1967) Climatological observations of stations in India).

MONTH	MEAN AIR TEMPERATURE (^{0}C)			
	MONTHLY MAXIMUM	MONTHLY MINIMUM	DAILY MAXIMUM	DAILY MINIMUM
JANUARY	9.7	-6.7	4.4	-2.3
FEBRUARY	13.7	-4.8	7.9	-0.8
MARCH	20.1	-0.4	13.4	3.5
APRIL	26.3	-4.4	19.3	7.4
MAY	31.5	7.1	24.6	11.2
JUNE	34.4	10.5	29.0	14.4
JULY	35.5	14.5	30.8	18.4
AUGUST	34.0	13.7	29.9	17.9
SEPTEMBER	32.6	7.8	28.3	12.7
OCTOBER	28.5	1.2	22.6	5.7
NOVEMBER	20.4	-3.4	15.5	-0.1
DECEMBER	14.4	-5.4	8.8	-1.8

Length of record: 30 years.

2.3 Vegetation.

The flora of Kashmir has been the subject of interest since the end of the last century (eg. Duthie, 1893, 1894; Coventry, 1923-1930). The valley falls into the 'Western Himalayan' botanical province of most authors (eg. Clarke, 1898; Hooker, 1906; Chatterjee, 1939). Although the variations in regional climate, geology and topography render a strict altitudinal zonation of the flora difficult, there is a clear distinction between the temperate flora of Kashmir Valley, the intermediate subalpine zone and, above that, the alpine zone. The vegetation has also been affected by human activity, including cultivation, grazing and forest management.

Below about 2300 m a.s.l. is the temperate forest zone (Raza *et al*., 1978) which includes the forested area of Kashmir Valley itself. It extends into the mouths of the montane valleys on the Great Himalayan flank, and onto the uplifted Karewa surface on the Pir Panjal flank. The forest consists of mixed broad-leaved and coniferous forest. Common tree species include *Populus sp*., *Juglans spp*., *Ulmus spp*., *Pinus excelsa* and *Cedrus deodara*. Where neither cultivated nor forested, the temperate zone supports scrub and grassland vegetation or, in areas adjacent to lakes and on river floodplains, marsh vegetation (Singh and Kachroo, 1976).

Above about 2100 m a.s.l., there is a progressive decline in the broad-leaved component of the forest and between 2300 m and the alpine zone, coniferous forest predominates. Common arboreal species include *Pinus excelsa*, *Abies pindrow* and *A. webbiana*. Broad-leaved trees are present, however, and include Prunus spp., Acer spp., Salix spp. and Populus spp. (Dhar and Kachroo, 1983). Raza et al. (1978) note that Ulmus is also found in this zone up to altitudes of over 2700 m a.s.l.

The alpine zone occurs at altitudes above 2900-3200 m a.s.l. (Raza et al., 1978; Dhar and Kachroo, 1983). In its lower parts, up to about 3600m, Betula utilis is the most common tree species. Above about 3600 m, tree and shrub species tend to be stunted due to the extremes of temperature, frost and wind that are characteristic of the zone (Dhar and Kachroo, 1983). The commonest trees above 3600 m are Juniperus spp., although individuals tend to be stunted (Raza et al., 1978). Other plants in this zone include Rhododendron, Salix, Syringa and Lonicera. Meadows are also common in the alpine zone. These are partly a response to grazing, especially by sheep and goats, of the nomadic Gujar people. Temperate species such as Poa, Glyceria, and Festuca characterize these meadowlands (Raza et al., 1978).

CHAPTER 3. LATE CAINOZOIC ENVIRONMENTS.

3.1 The Early Research.

In Kalhana's Rajatarangini or Chronicle of the Kings of Kashmir a demon was said to have occupied a large lake that once filled the valley of Kashmir. When the gods descended and killed the demon, the lake was drained and the land of Kashmir created (Stein, 1900). These legends inspired the work of early scientists and explorers including Vigne (1842), Thompson (1852), Cunningham (1853) and Lawrence (1895). The term 'Karewa,' derived from the Kashmiri language, was adopted to describe the elevated mesas of lacustrine sediments.

Godwin-Austen (1859) first used the term Karewa in a geological sense: to describe the sediments as opposed to the landforms. In an 1880 paper, Godwin-Austen distinguished between the 'Hirpur series' which was exposed on the north-eastern flanks of the Pir Panjal Range and thought to be of Tertiary age, and a post-Tertiary sequence exposed in the south-east part of Kashmir basin. Drew (1875) recognized that the older beds were tilted and folded, whereas the younger sediments were virtually undisturbed. Lydekker (1876) first proposed the terms 'lower Karewa' and 'upper Karewa' for the older and younger beds respectively. Similarities between the lower Karewa beds and the high Siwalik strata on the south-west side of the Pir Panjal led Lydekker to suggest that they were near-contemporaneous.

Middlemiss (1910) observed remnants of Karewa beds at between 3350 and 3650 m a.s.l. in the Pir Panjal. Sahni (1936) discussed the origin of these beds in which he found leaf remains of temperate tree species. He suggested that their elevation was due to the rapid uplift of the Pir Panjal Range following deposition of the Karewa beds. Hora (1937) also found fish remains in the uplifted lower Karewa beds near Gulmarg, in the Pir Panjal. The existence of species characteristic of lacustrine and sluggish-river environments added weight to Sahni's idea of Pir Panjal uplift.

By the mid-1930s. observations in Kashmir had led to the development of several important ideas. Firstly, the idea of a former, large, lake had become well established. Secondly, a distinction had been made between tilted and folded lower Karewa beds and relatively undisturbed upper Karewa beds. Thirdly, elevated remnants of Karewa deposits suggested post-Karewa uplift of the Pir Panjal. Fourthly, contemporaneity between the lower Karewa and the upper Siwalik beds had been suggested.

Work on the supposed glacial deposits was also undertaken by a number of early writers. Lydekker (1879) found no evidence of ice expansion into the Kashmir basin itself, a view held by most subsequent authors. Lydekker identified the lowest evidence for glaciation at Kulan in the Sind Valley (about 2100 m a.s.l.) where supposed striations were found. Middlemiss (1911) described ancient moraines in the Pir Panjal Range at Gulmarg and

Toshmaidan. Oldham (1904) described the glacial geomorphology of the Sind Valley in some detail, arguing that the lowest unequivocal evidence for glaciation is at Gund in the Sind Valley, at 2080 m a.s.l.

A detailed study of the glaciation of the East Liddar Valley by Grinlington (1928) was subsequently included in a regional synthesis by de Terra and Paterson (1939). Since the publication of their monograph Studies on the Ice age in India and Associated Human Cultures Kashmir has been the focus of considerable research into Late Cainozoic environments. At this stage, it is convenient to review research on the Karewa, and that on glacial sequences, separately.

3.2 The Karewa Beds.

a. Introduction and Definitions.

Singh (1982 p.72) proposed the following definition of the Karewas:

> '...the more-or-less unconsolidated sedimentary succession overlying the Precambrian-Mesozoic basement and underlying the Holocene alluvium of modern rivers...the moraines of the mountain slopes are not included...'

Bhatt (1976 p.46) summarized the overall lithology of the Karewa beds as follows:

> '...a sequence of fluvial, lacustrine, glaciofluvial and aeolian rocks, deposited initially in the bed of a large lake, called the Karewa lake, that once covered the whole of the present-day Kashmir Valley floor; the Karewa

> succession is invariably capped by layers of continental loess that were generated later in stages when the draining out/desiccation of the lake had set in.'

The lower Karewa beds are best exposed in the river valleys draining the Pir Panjal Range. Here, over 1000 m of fine-grained lacustrine mudstones and fluviodeltaic sandstones are found. These are punctuated by 5 major fluvial conglomerate beds, which are generally several tens of metres thick. On the south-west side of Kashmir basin, a conglomerate unconformably overlies eroded lower Karewa beds. This conglomerate is less than 100 m thick and is regarded as the base of the upper Karewa. Towards the centre of the basin it pinches out, and the contact between lower and upper Karewa is presumably conformable, since the bulk of overlying strata are fine-grained. Conglomerate terraces are also present in the mouths of the Sind and Liddar Valleys, which drain the Great Himalayan Range. Upper Karewa lacustrine beds are exposed in the basin-centre and on the Himalayan flank, but are absent from the south-western part of Kashmir. They are less than about 50 m thick. The whole of the sequence is capped by loess. On the Pir Panjal flank, the loess overlies the upper Karewa conglomerate or isolated hummocks of lower Karewa lacustrine sediments: on the Himalayan flank and in the basin centre, it overlies upper Karewa lacustrine beds. It also caps the conglomerate in the mouths of the Himalayan valleys. The Karewa deposits have been extensively eroded following deposition, and remain as mesas standing above the floodplain of the Jhelum River and its tributaries. In the remaining parts of this section. the work on the formal lithostratigraphy, palaeoenvironmental interpretation and chronology of the Karewa beds is reviewed.

b. Stratigraphical controversies.

There have been several attempts to construct a formal stratigraphy of the Karewa group, but they have met with limited success. Lateral variations in facies render correlations based on lithology very difficult if, as is often the case, beds cannot be traced physically from one section to another. Such uncertainties in correlation have led to estimates of the thickness of the whole sequence ranging from 420 m (Godwin-Austen, 1864) to over 2200 m (Wadia, 1941).

The first attempt to construct a formal lithostratigraphy of the Karewa was made by Farooqi and Desai (1974). They gave 'group' status to the sequence and defined two formations. The lower Karewa was named the 'Pakharpura Formation' after the village of the same name in the Romushi Valley. However, the lowermost part of the sequence, the 'Hirpur Conglomerate,' was described and named in the Rembiara Valley. The upper Karewa was named the 'Shopian Formation' and based on four 'clay-capped' conglomerate terraces in the lower Rembiara Valley.

In a series of papers, D.K. Bhatt rejected Farooqi and Desai's lithostratigraphy and proposed an alternative (Bhatt, 1975, 1976, 1979, 1982a, 1982b; Bhatt and Chatterji, 1976, 1979).

Group status was retained, but the names of the constituent formations and members, together with the type sections, were rejected. Bhatt argued that Farooqi and Desai's 'Pakharpura Formation' was incorrectly defined, because the proposed type section did not display the whole of the stratigraphic sequence. The 'Shopian Formation' was rejected on the valid grounds that lithostratigraphic units should not be based on morphological features.

Bhatt (1976) proposed that the type locality for the lower Karewa should be in the Rembiara Valley and that the sequence should be termed formally the 'Hirpur Formation.' In his 1979 paper, Bhatt recognized two beds, or 'zones' of fine-grained, lacustrine sediments, separated by a thick conglomerate. In his PhD thesis, Bhatt (1982b) named each of them formally as the 'Dubjan Member', the 'Rembiara Member' (the conglomerate) and the 'Methawoin Member,' in ascending order.

Bhatt (1982a) argued that Farooqi and Desai's lithostratigraphy showed an insufficient understanding of upper Karewa palaeogeography and failed to identify the coexistence of lacustrine sedimentation in the north-east part of Kashmir basin, with fluvial conglomerate deposition in the south-west. Bhatt (1976) formally named the upper Karewa the 'Nagum Formation' after the type site at Qasba Nagum. Three members were informally designated: the 'gravel member', the 'laminated silt member' and the 'loam member', from the base upwards. They were interpreted as glaciofluvial, lacustrine and aeolian respectively. In his PhD thesis, Bhatt assigned formal stratigraphic names to the members: the 'Shupian Member', the 'Pampur Member' and the 'Dilpur Member' respectively.

The lower Karewa beds are exposed mainly along the south-western margin of Kashmir Basin, due to the incision of the uplifted strata by rivers draining the Pir Panjal. On the north-east flank of the basin, post lower Karewa uplift has apparently been limited, and the lower Karewa beds are buried beneath younger sediments. However, Singh (1982) maintained that blue-grey silty-clay deposits exposed at Wopazan, on the Himalayan flank, are lower Karewa beds that have been uplifted, and therefore exposed, by localized crustal warping.

The upper Karewa conglomerate bed truncates lower Karewa lacustrine strata on the Pir Panjal flank. This conglomerate thins progressively from the Pir Panjal mountain front towards tha basin centre. Well developed conglomerate terraces of uncertain stratigraphic position are also present in the mouths of the Sind and Liddar Valleys. Agrawal (1985) has suggested that conglomerates exposed at Malshahibagh, basinward of the Sind Valley mouth, may be distal remnants of the upper Karewa conglomerate on the Pir Panjal flank.

The angular unconformity between the upper Karewa conglomerate (the Shupian Memeber of Bhatt) and the lower Karewa lacustrine strata on the south-west basin margin forms the basis of the upper-lower Karewa division of many authors. Towards the centre of the basin and on the north-east flank, the marker conglomerate is absent and the boundary between upper and lower Karewa is presumably conformable. In these areas, the so-called upper Karewa lake deposits (the 'Pampur Member' of Bhatt) are found. They form distinct mesas of horizontally-bedded sediments which are almost invariably loess-capped. The spatial distribution of the upper Karewa conglomerate and overlying lake beds is held to indicate areal restriction of the lake during upper Karewa times as a result of tectonic uplift along the south-west basin margin (Bhatt, 1982b). The restricted lacustrine sedimentation was coincident with fluvial conglomerate deposition in the south-west part of the basin, which is regarded as either glaciofluvial (Bhatt 1982b) or tectonic (Burbank, 1982) in origin.

At the type sections proposed by Bhatt (1982a), the stratigraphical sequence of the Karewa group appears unambiguous. However, elsewhere in the basin, the situation is unclear. In the basin centre and in the north-east flank, the previously used distinction between upper and lower Karewa has been the presence or absence of blue-grey silty-clay facies. Where this facies is present, the sediments are regarded as lower Karewa. However, this is an inadequate basis for formal lithostratigraphic subdivision. A further problem relates to the relationship between the upper Karewa conglomerate and the lacustrine beds.

Although Bhatt (1982b) has assumed that the two units are coeval, the relationship between them has not been demonstrated convincingly. In addition, the stratigraphical position of the conglomerates in the outer Sind and Liddar Valleys is not fixed. Palaeomagnetic studies of the lower Karewa (Burbank, 1982; Kusumgar et al., 1985) have shown that the correlation of conglomerates in the Rembiara and Romushi Valleys by Singh (1982) is in error. In the light of this evidence, and the previous problems outlined above, the formal lithostratigraphical nomenclature for the Karewa should be dropped. The upper and lower divisions are retained in this study, however, but only for informal use.

c. Palaeoenvironmental Controversies.

Most authors agree that the Karewa beds are lacustrine deposits punctuated by fluvial sediments and capped by loess. However, considerable debate surrounds the more detailed interpretation of the record.

De Terra and Paterson (1939), whilst aware that tectonic uplift had occurred during the deposition of the Karewa, interpreted the sedimentary record largely in climatic terms. They argued that the conglomerates were of glaciofluvial origin and the fine-grained muds were interglacial deposits. In his earlier papers, Bhatt agreed with de Terra and Paterson (see, for example, Bhatt, 1975). In later works, however, his view changed (eg. Bhatt and Chatterji, 1976; Bhatt, 1979,1980,1982a) and the lower Karewa conglomerates were regarded as tectonically-induced fluvial sediments related to the uplift of the Pir Panjal Range. However, the upper Karewa conglomerates were still regarded as glaciofluvial. Burbank (1982) rejected a glacial origin for any of the Karewa conglomerates, arguing that they show no causative relationship with any moraines. Like Bhatt, Burbank interpreted the lower Karewa conglomerates as tectonic alluvial facies. The lower three beds were shown, on the basis of clast imbrication, to have been derived from the north-east margin of Kashmir, and so related to uplift of the Himalayan Range. Similar arguments were used to show that the upper three conglomerates, including the upper Karewa conglomerate, were derived from uplift of the Pir Panjal Range.

Recent work, then, has suggested that the major lithological changes in the Karewa sediments are a response to basin-margin tectonic activity. This is not to say, however, that there is no evidence for climatic change within the Karewa beds. However, any climatic interpretation must pay careful attention to the effects of tectonics.

Much of the palaeoenvironmental work undertaken in Kashmir has focused on the lower Karewa. A variety of different types of evidence have been used to infer environmental change including sedimentology, organic geochemistry and palaeontology.

Burbank and Grant (1985) examined particle-size variations in lower Karewa beds which span the time period from 2.6 to 1.6 MaBP. Using the palaeomagnetic chronology already developed for the lower Karewa by Burbank (1982) and assuming constant sedimentation rates between the palaeomagnetic boundaries, they were able to convert stratigraphic position to age. Particle size was assessed every 0.5 m by visual estimation. Using Fourier analysis, the authors identified three bands of spectral power in the particle-size data. Two of these bands , 20 to 24 ka and 40 to 50 ka were present throughout the sequence. In the part of the sequence older than 2.3 MaBP, a further band of power, centring on 100 ka, was found. Burbank and Grant (1985) suggested that these results indicate Milankovitch forcing of particle-size variations, with the 20 to 24 ka spectral band approximating to the 21 ka precessional cycle and the 40 to 50 ka spectral band approximating to the 40 ka tilt cycle.

Krishnamurthy et al. (1986) attempted to show fluctuations in the level of the lower Karewa lake using stable carbon-isotope and carbon:nitrogen analyses of organic matter in the lower Karewa mudstones. They sampled lower Karewa beds spanning 2.4 to 0.4 MaBP. 70 samples were taken from the 450 metre-thick section, and both carbon-isotope and C:N analyses were undertaken on each sample.

The ratio of the stable carbon isotopes ^{13}C and ^{12}C is generally regarded as a good indicator of the source of organic carbon. Aquatic vegetation will tend to have ^{13}C values that are distinct from those of terrestrial vegetation. This relates to the source of carbon dioxide used in photosynthesis. Submerged aquatic plants derive their CO_2 from the total dissolved inorganic carbon, which includes $CO_{2(aq.)}$, HCO_3^- and CO_3^{2-}. Terrestrial plants and emergent aquatics derive their CO_2 from the atmosphere. When CO_2 dissolves in water, there is preferential enrichment of the heavier isotope, ^{13}C. Thus, aquatic plants, if they are in isotopic equilibrium with the lake water, should consist of carbon that is isotopically heavier than that in terrestrial plants. For terrestrial plants, there are two categories of $^{13}C/^{12}C$, the values of which cluster around $-27^0/oo$ and $-14^0/oo$. These categories are found in plants with C3 and C4 photosynthetic pathways respectively. C3-type vegetation tends to be associated with humid conditions whereas C4 plants are found in arid and semi-arid environments.

Distinct C:N ratios are also associated with different types of vegetation. Lower plants, especially algae, tend to have high protein contents and thus low C:N ratios compared with terrestrial plants. However, dryland vegetation includes many leguminous species, which have lower C:N ratios than plants of more humid environments. From these two arguments, low ^{13}C values and high C:N ratios in a lacustrine environment tend to indicate a large input of terrestrially-derived organic matter to the lake whereas high ^{13}C values and low C:N ratios suggest that most of the organic matter was derived from aquatic plants and /or from terrestrial vegetation adapted to dry conditions.

Krishnamurthy et al. (1986) argued that high C:N ratios and low ^{13}C values indicate periods of low lake-levels, when a large amount of organic matter was derived from the adjacent terrestrial zone. Conversely, low C:N ratios and high ^{13}C values were interpreted as indicators of high lake-levels during which the organic matter was derived from the submerged aquatic vegetation. The low lake-levels were explained by a cold, arid climate and the high lake-levels by a warm, humid climate. Using this model, Krishnamurthy et al. (1986) identified three major lowstands of the lower Karewa lake.

However, there are a number of shortcomings with this study. Although the interpretation of high and low lake-levels from their analyses may be correct, the palaeoclimatic conclusions drawn from them are open to debate. Lake-level fluctuations can provide valuable estimates of water balance changes, but only under closely defined circumstances. In particular, it is assumed that the lake basin is closed,the only losses are evaporative, and the only inputs are due to precipitation over the basin and runoff into the lake. In practice, many lakes have inputs and outputs from subsurface flow and estimates of these must be made in order to draw palaeoclimatic conclusions from lake-level data. However, once surface outflow becomes important in the water balance of a lake, it is very difficult to interpret past fluctuations of lake-level in palaeoclimatic terms. Surface outflow can be estimated if the channel through which it occurred can be identified and assumed to have remained at constant altitude during the history of the lake. However, in the case of the Karewa lake, tectonic instability has led to variations in both source area and base level through time. Because of this, large variations in the level of the lake can be explained in terms of tectonic activity as well as climatic change. Burbank (1982) argued that a rise in the base level of the lake in the Baramulla Gorge area would pond fluvial drainage and lead to lake transgression. Conversely, uplift in the source area would lead to drainage in the gorge and lake regression. Clearly, lake-level fluctuations in Kashmir can be explained without recourse to climatic change. In reality, it is likely that both climatic and tectonic effects have influenced the lake-level record, but differentiating between the two is impossible.

Even if the lake-level changes could be attributed to climate, the argument that high lake-levels are a response to warm, humid climates and low lake-levels a response to cold, arid climates is simplistic. Any climatically-induced lake-level change is a function of both temperature and humidity. However, when there is no independent palaeotemperature estimate for any given lake-level record, it is impossible to determine the relative importance of precipitation and temperature. Clearly, the task of drawing meaningful palaeoclimatic information from the carbon isotope and C:N data for the lower Karewa is fraught with difficulty.

A considerable amount of palaeoenvironmental work on the lower Karewa has involved the study of pollen, although the results have been disappointing. Wodehouse (1935) concluded that the environment during lower Karewa deposition was little different from that of today. Vishnu-Mittre et al. (1962) analysed lower Karewa sediments from several sites. They interpreted the pollen record in terms of a change in climate from wet to dry, and then to moist with progressive cooling. However, conditions were thought to be temperate throughout.

Both of the above studies were undertaken prior to the development of a palaeomagnetic chronology for the lower Karewa. Recent work by PRL scientists had the advantage of being tied to a dated sequence. Samples were taken from a range of sediments of varying age from 4 to 0.7 MaBP. However, not all of the samples were found to contain pollen. The oldest pollen-bearing beds in the lower Karewa were found at Dubjan, in the Rembiara Valley (Sharma et al., 1985). By extrapolation from the magnetic polarity stratigraphy, it was concluded that the Dubjan beds are about 4 Ma old. The pattern of vegetation change was interpreted in purely climatic terms. Three vegetation zones were identified. The first of these was grassland, the second Chir Pine forest and the third grassland. The corresponding climates were interpreted as 'warm-temperate with moderate humidity', 'warm-temperate to subtropical with less humidity', and 'cool temperate with greater precipitation'. The next youngest section analysed was that from Hirpur, with a basal age of 3.9 MaBP. The sequence extended up to 3.58 MaBP. The analysis of pollen from this sequence was undertaken by Gupta et al. (1984).

The authors divided the pollen record of the Hirpur sequence into fifteen zones, with thirteen intervening barren zones. As with the Dubjan sequence, a palaeoclimatic interpretation was proposed. In the lowest eleven zones, the inferred climate fluctuated between 'subtropical warm' and 'warm temperate' with humidity ranging from 'dry' to 'wet' (Gupta et al., (1984). Stages twelve to fifteen showed a progressive cooling and drying.

Pollen analysis has also been undertaken by PRL scientists on younger lower Karewa sediments. The pollen sequence from Krachipatra, in the Rembiara Valley, was interpreted as indicating a'cool temperate and moist' climatic regime. The beds analysed date from about 1.5 MaBP (Sharma and Gupta, 1985). Two other sites, Wogahoma and Baltal, which preserve some of the youngest pollen-bearing lower Karewa strata, have both been dated to about 0.7 MaBP. In both cases, a 'cold and arid (glacial)' climate was inferred (Dodia et al., 1982; Dodia and Agrawal, 1985).

Taking the lower Karewa pollen sequences as a whole, Agrawal (1985) argued that the Late Cainozoic climate of Kashmir has followed a 'broadly global' pattern. The Late Pliocene was characterized by a warm, humid 'subtropical' climate. Progressive cooling followed in the Quaternary.

Following the publication of these pollen studies, the palaeoclimatic interpretations have been criticized. For example, Singh (1982) pointed out that the pollen spectra from Hirpur showed lithological control, in that sandy samples are dominated by Chir Pine pollen whereas lignite samples, stratigraphically close to the sands, contain aquatic taxa. Singh argued that the changes in pollen spectra were an artefact of the facies present, and therefore reflected changes in lake-levels or river channel-positions rather than climate. In the light of these criticisms, the climatic interpretation of the lower Karewa pollen sequence can only be accepted with reservation.

Despite several records of fossil diatoms in the lower Karewa, only one semicontinuous record has been published. This work was carried out on sediments from Baltal, in the Romushi Valley, by PRL scientists (Gandhi and Mohan, 1983; Gandhi et al., 1985). It was suggested that the diatom assemblage indicated the predominance of cool lacustrine conditions. The pollen spectrum from the same sequence at Baltal suggested a 'cold and arid' climate.

In contrast, studies of ostracods (Bhatia et al., 1985a), charophytes (Bhatia et al., 1985b), and plant macrofossils (eg. Lone *et al.*, 1985) have been too sporadic to be of value for the present purpose. Only vertebrate fossil remains have been studied in sufficient detail to provide useful palaeo-environmental information.

The study of vertebrate fossils from the lower Karewa has a long history. Godwin-Austen (1864) and Hora (1937) found fish remains: de Terra and Paterson (1939), Badam (1968, 1972) and Tewari and Kachroo (1977) all found various mammalian remains. However, only recently has the vertebrate record been studied in detail, following the doctoral research of B.S. Kotlia (Kotlia, 1984, 1985a, 1985b; Sahni and Kotlia, 1983, 1985; Kotlia *et al* 1982). Although the record is rather fragmentary, it provides some useful constraints on lower Karewa palaeoenvironments.

Kotlia (1984) found an abundance of megavertebrates, micromammalian and fish remains in lower Karewa sediments. Megavertebrate fossils belonging to Equidae, Elephantidae, ?Giraffidae, Cervidae and Canidae were found. Equid remains (*Equus sivalensis*) and Cervids (*Cervus punjabiensis* and *C. sivalensis*) were found in beds of lower Matuyama to Olduvai age (c. 2.48 to 1.67 MaBP). Elephants (*Elephas hysudricus*) and canids were restricted to the Brunhes chron while a ?giraffid was found just above the Matuyama-Gauss boundary. Micromammalian fossils were most common in sediments of Olduvai age (1.87 to 1.67 MaBP). They included murid rodents (Muridae), microtine rodents

(Microtidae) and shrews (Soricidae). Fish remains were found sporadically throughout the sequence.

Kotlia (1984) concluded that the lower Karewa records a change from warm humid and warm arid climatic conditions in the lower and middle parts of the sequence to an increasingly cold climate in the upper parts, particularly the exposures at Wopazan and Baltal. However, as with the pollen evidence, the problem of mixing of upland and lowland communities, and of facies control on the fossil record, exists. Kotlia (1984, p.280) pointed out that many of the fish and micromammalian fossils could have been washed into the lower Karewa lake, whereas the larger and heavier megevertebrate remains are probably more or less _in situ_.

The upper Karewa beds have received much less attention from geologists than the lower Karewa. On the basis of a sedimentological study, Singh (1982) argued that the upper Karewa conglomerate is a braided stream deposit and that the silt-clay beds were laid down in a shallow lake. The loessic origin of the overlying sediments was first suggested by Agrawal _et al_. (1979). Despite these studies, detailed sedimentological information for the upper Karewa is lacking.

Krishnamurthy _et al_. (1982) undertook stable oxygen-isotope measurements on carbonates from the lower Karewa lacustrine beds and also from pedogenic carbonates in the overlying loess. The lacustrine carbonates were found to be isotopically lighter than the pedogenic carbonates. On the basis of this result, Singh (1982) argued that the upper Karewa period was warmer than the period of pedogenesis that followed. However, Krishnamurthy et al (1982) suggest that meltwater from montane glaciers could have supplied isotopically light water to the upper Karewa lake. To some extent, this idea is supported by the studies of Pant et al. (1979). On the basis of scanning electron-microscope studies of quartz-sand grains from upper Karewa sand units, they suggested that the upper Karewa lake was fed by glacial meltwater.

Other useful palaeoenvironmental evidence, such as pollen and diatoms (Nair, 1960; Roy, 1974), is absent from the upper Karewa. Molluscs, vertebrate remains, charophytes and ostracods have been found, but only in the lacustrine beds: the conglomerate and loess are unfossiliferous. A number of studies have been undertaken on the the upper Karewa ostracod fauna (Bhatia, 1968; Bhatia and Singh, 1971; Singh, 1972, 1973, 1974). As a result of this, the species present in the upper Karewa are well known. However, the sampling in all of these studies had poor stratigraphical control and detailed species counts were not undertaken. Thus, the palaeoenvironmental value of the work is limited. However, Singh (1973) inferred that the upper Karewa lake was shallow, cool, and slightly alkaline from the ostracod species present.

Numerous studies have been carried out on the loess of Kashmir (eg. Agrawal et al., 1979; Kusumgar and Agrawal, 1985; Kusumgar et al., 1980, 1986). These studies have shown that the loess is up to 20 metres thick and in some sections, as many as 9 palaeosols have developed. These studies have argued that the loessic silt was produced by glacial grinding. Following studies of loess from elsewhere in the world (eg. Kukla, 1987), the palaeosols are regarded as indicators of warm, humid and interglacial conditions, and the non-pedogenically altered loess of glacial climates.

d. Chronological Controversies.

Early estimates for the basal age of the Karewa ranged from Miocene (Roy, 1975; Singh, 1977) and Pliocene (Lydekker, 1883, Oldham, 1893; Middlemiss, 1911; Wadia, 1941, 1951; Farooqi and Desai, 1974; Bhatt, 1975, 1976) to Pleistocene (de Terra and Paterson, 1939; Vishnu-Mittre et al., 1962). De Terra and Paterson argued that the top of the Karewa sequence corresponded to the '2nd interglacial' and that the loess was a postglacial deposit. This chronology was accepted by many subsequent workers, although Bhatt (1975, 1976) regarded the top of the Karewa as a time-transgressive surface of 2nd interglacial age on the Pir Panjal flank and Holocene age on the Himalayan flank.

The first attempt to develop a sound chronology for the Karewa was by Kusumgar (1980) using magnetic polarity stratigraphy (MPS) and palaeontological markers. The occurrence of fossils such as Equus sivalensis (Tewari and Kachroo, 1977) and Elephas hysudricus from a sequence of reversed polarity sediments led Kusumgar to believe that the beds were of Matuyama (reversed chron) age. Normally magnetized sediments below the aforementioned beds were interpreted as being of Gauss age. A further reversed chron, lower still in the sequence, was regarded as the Gilbert. It was suggested that Karewa sedimentation may have commenced 5 MaBP. The top part of the lower Karewa, together with the entire upper Karewa sequence that could be dated palaeomagnetically (the lacustrine beds and the loess) were found to be normally magnetized and interpreted as being of Brunhes age.

Burbank (1982) also carried out palaeomagnetic investigations on the lower Karewa beds. He had the advantage of finding volcanic ashes in several horizons. Fission track dates on zircons from these ashes provided much greater certainty to the MPS. From Burbank's chronology, a number of important details have emerged. Firstly, basin sedimentation was estimated to have begun about 4 MaBP. Secondly, accurate correlation between sections allowed a reliable thickness estimate for the lower Karewa of 1300 m to be made. Thirdly, although all of the upper Karewa was found to be normally magnetized, Burbank was able to place constraints on the onset of upper Karewa conglomerate deposition of 0.35 to 0.4 MaBP.

Following Burbank's work, a team of scientists from PRL published a palaeomagnetic analysis of the same sections (see Agrawal, 1985). Similar results were obtained, except that the lowest palaeomagnetic boundary was interpreted as Gilbert-Gauss by PRL, rather than Kaena-Gauss as suggested by Burbank (see appendix 1 for details of the palaeomagnetic timescale). Recently however, Burbank has expressed agreement with the PRL interpretation of the boundary (personal communication to author, 1988).

The chronology of the upper Karewa is poorly constrained.Because the entire sequence falls within the Brunhes chron, palaeomagnetic stratigraphy is of little value, with the exception of the chronological constraints on the onset of conglomerate deposition suggested by Burbank (1982) and mentioned above. Numerous radiocarbon dates have been carried out on loessic palaeosols (see appendix 2 of this thesis for a list of dates). The dates are interpreted to indicate a period of soil formation, and hence climatic amelioration, at about 18 kaBP (Kusumgar and Agrawal, 1985). However, many of these dates show considerable scatter around 18 kaBP and the technical and conceptual problems of dating palaeosols by radiocarbon dating are considerable. In particular, contamination with younger carbon tends to be a problem with dating palaeosols. Recent TL dates on the loess (see appendix 2: Bronger et al.,1987; Rendell, personal communication) suggest an alternative chronology for the

Kashmir loess, and confirm that the the palaeosol dated to 18 kaBP by radiocarbon is actually considerably older.

3.3 The Glacial History Of Kashmir.

a. The Onset of Glaciation.

In Quaternary research, glaciation was traditionally regarded as synonomous with the Pleistocene. However, in the absence of a reliable chronology, it was impossible to correlate the onset of glaciation and the onset of the Pleistocene convincingly. Recent work on oceanic sediments has suggested that substantial global cooling occurred in the Pliocene (eg. Shackleton et al., 1985). However, this does not necessarily imply terrestrial glaciation in specific geographical areas, although there is evidence for pre-Pleistocene glaciation (eg. Mercer et al., 1975).

The results of the pollen (Dodia, 1984; Sharma et al., 1985; Gupta et al., 1985; Sharma and Gupta, 1985) and diatom studies (Gandhi et al., 1985) reviewed in the previous section have been used to infer a 'glacial' climate for Kashmir beginning about 0.7 MaBP. However, the absence of any dated glacial deposits makes it impossible to state firmly that glaciers existed in any part of Kashmir Valley at that time.

The absence of any glacial deposits within, or in association with, the Karewa beds, led Burbank (1982) to suggest that significant glaciation of the Pir Panjal Range postdated the deposition of the lower Karewa beds. For the Pir Panjal Range, this is quite plausible, since during the deposition of the lower Karewa, the Pir Panjal Range was much lower than during the Late Quaternary. However, the Great Himalayan Range attained a higher altitude much earlier than the Pir Panjal. Furthermore, the absence of a topographic barrier to the south would have allowed the incursion of much more moisture into the Great Himalaya, which could have enhanced the growth of glaciers. However, glacial deposits of the earliest advances would be unlikely to have been preserved in Kashmir: early advances are only usually recorded in specific sedimentological situations, such as glaciomarine sequences, or where overlain by volcanic deposits. In the absence of glacial deposits in Kashmir, it is necessary to look for indirect evidence for the onset of glaciation.

b. Glacial Sequences of The Great Himalaya and Pir Panjal.

A number or early workers identified and mapped glacial deposits in Kashmir. Oldham (1904) and Norin (1925) worked in the Sind Valley; Grinlington (1928) in the East Liddar Valley; and Lydekker (1876) and Norin (1925) in the Pir Panjal Range. Following the Italian expedition to Central Asia in 1913-1914, Dainelli (1922) introduced the concept of four glaciations for Kashmir.

De Terra and Paterson (1939) built upon Dainelli's work in their monograph <u>Studies on the Ice Age in India and Associated</u>

Human Cultures.They recognised lower, middle and upper divisions of the Pleistocene in the Kashmir depositional record. The lower Pleistocene was thought to include, the 1st glacial and interglacial stages. Deposition in the lower Pleistocene consisted of sporadic, high-level moraines in the mountain flanks, thick gravel fans extending to the valley mouths and, in the basin itself, the lower Karewa beds. The Middle Pleistocene was thought to consist of the 2nd glacial and interglacial. De Terra and Paterson attributed widespread moraines in the mountain flanks and thick gravel deposits (the 'Karewa Gravel' which is equivalent to the upper Karewa conglomerate) to the 2nd glacial, and the upper Karewa lake beds to the 2nd interglacial. The Upper Pleistocene was thought to comprise the 3rd and 4th glacials and the 3rd interglacial. During the glacial stages, moraines were deposited in the mountain flanks: during the interglacial, terracing and erosion of older deposits took place.

Three main aspects of de Terra and Paterson's sequence should be emphasized. Firstly, the 1st and 2nd glacials were longer than the 3rd and 4th. Secondly, interglacials were held to be longer than glacials. Thirdly, the maximum altitudinal descent of ice was to about 1670 m a.s.l. In areal terms, this implies that glaciers extended to the mouths of the main Himalayan valleys and beyond the Pir Panjal slope and onto the uplifted lower Karewa surface. The moraines of de Terra and Paterson's glacial stages are shown in figure 8 and their sequence is summarized in table 3.

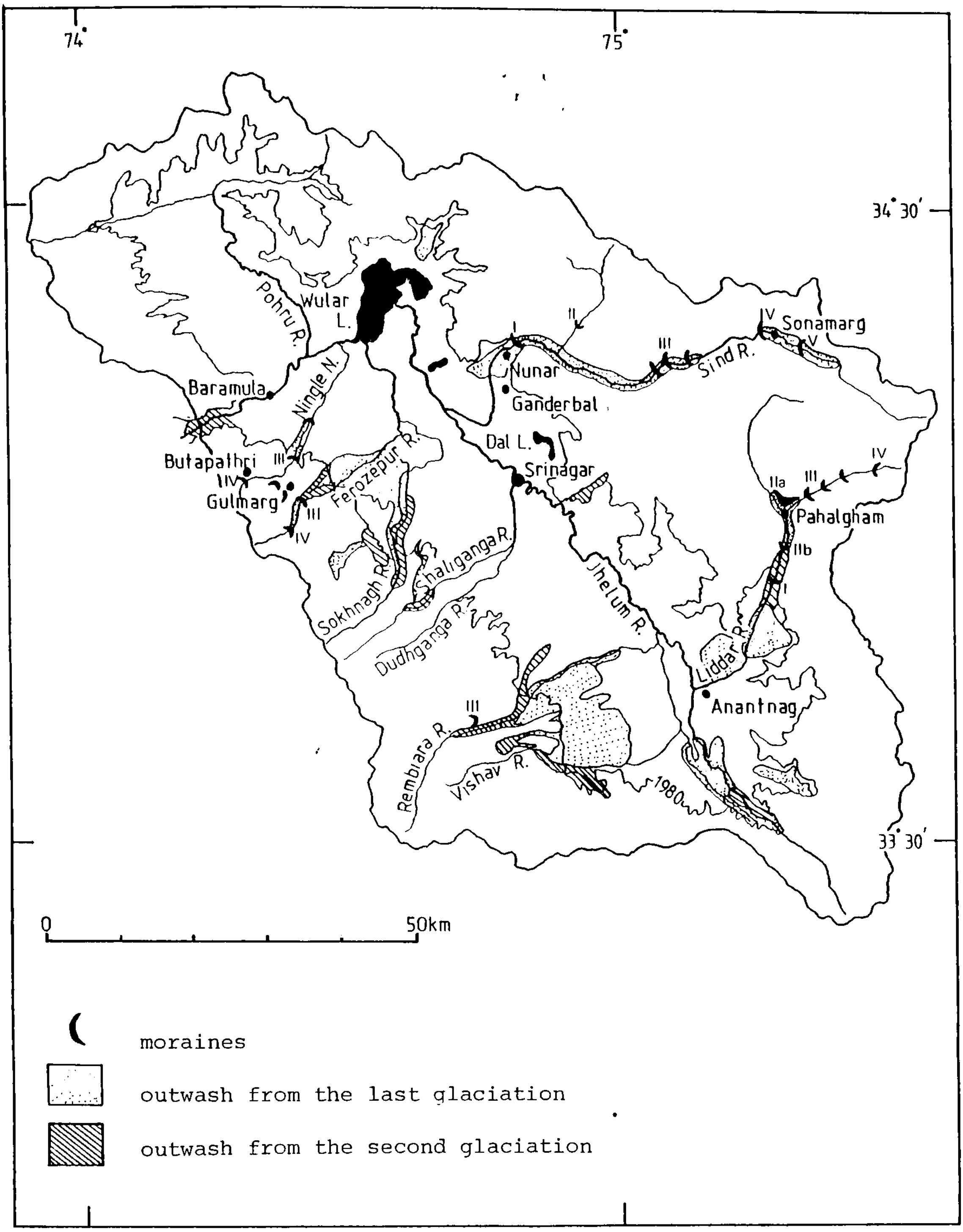


Figure 8

Glacial stratigraphy of Kashmir (according to deTerra and Paterson, 1939)

TABLE 3. A SUMMARY OF DE TERRA AND PATERSON'S GLACIAL HISTORY OF KASHMIR.

STAGE	HIMALAYAN FLANK		PIR PANJAL FLANK	
	SIND VALLEY	LIDDAR VALLEY	NINGLE VALLEY	REMBIARA & VISHAV VALLEYS
FOURTH GLACIAL	Sonamarg moraines & retreat stages in Baltal valley 2590 & 2700 m.	Moraines above Gorge & the outlet of Sishram Nag, 3353 m.	Moraines at edge of Gulmarg Basin & Butapathri 2740 m.	Terminal moraines at 2680 & 2670 m. loam deposition on terraces.
THIRD INTER-GLACIAL	Erosion & slight tilting-terrace formation	Erosion & slight tilting-terrace formation	Erosion & Tilting	Extensive terracing of gravels
THIRD GLACIAL	Moraines & outwash at Gund Pharao, Rezan & Gagangiyer 2070-2347 m	4 stages of terminal moraines between Nebatakun & Tanin. 2470-2774 m.	Moraines 1-3 1860-1980 m.	Terminal moraines at 2380-2225 m.
SECOND INTER-GLACIAL	Erosion & terrace formation. Upper Karewa lake beds in Kashmir basin and main valley mouths.		Upper Karewa Lake beds and aeolian deposition.	
SECOND GLACIAL	No terminal moraines but ice front at Mangom. Karewa gravel (outwash)	(i) Pahalgam moraine (ii) Lidru moraine Karewa gravel (outwash)	Moraines at the edge of the Karewa Lake. Karewa gravel (outwash) truncates Karewa surface	
FIRST INTER-GLACIAL	Lower Karewa Lake Beds - erosion and tectonic tilting			
FIRST GLACIAL	Terminal moraine at Mangom 1676 m. Malshahibagh conglomerate	Ganeshpur moraine 1860 m. Anantnag Conglomerate	Sporadic, high-altitude moraines	Sporadic, high-altitude moraines Thick gravel fans: basal bed of Karewas

Considerable doubt must be cast on de Terra and Paterson's ideas, however, as a result of general advances in Quaternary science and due in particular to work on the Karewa by Burbank (1982) and scientists from PRL (eg. Kusumgar et al., 1985).The shortcomings in de Terra and Paterson's sequence arose for a number of reasons. Their correlation between the glacial deposits and the Karewa beds, hampered by the lack of a sound chronology, is largely incorrect. Furthermore, many of the'moraines'are in fact of non-glacial origin. Their climatic interpretation of the Karewa beds was also largely incorrect. De Terra and Paterson's work was, of course, formulated within the framework of contemporary Quaternary science. In particular, their thinking was constrained by the ideas that the Pleistocene comprised four glacials and interglacials, that the Pliocene was warmer than the Pleistocene and experienced a tropical 'preglacial' climate; and that glacials were shorter than interglacials. All of these ideas are now known to be wholly or partly fallacious.

c. The Chronology of the Last Glaciation and the Holocene.

Globally, the last glaciation and Holocene have been the focus of considerable research effort. Sedimentary evidence is often well preserved, and the Holocene and latter part of the last glacial are within the range of radiocarbon dating. In Kashmir, a limited amount of work has been done on the chronology of this period. For example, de Terra and Hutchinson (1936) observed five lines of evidence for climatic change in the 'postglacial' period. These were recent moraines; terraces associated with postglacial glacier advances; terraces and beaches associated with lake-level fluctuations; ancient chronicles and prehistoric monuments; and rock patinations and engravings. Mayewski and Jeschke (1979) incorporated historical observations of the Kolahoi Glacier, which is at the head of the West Liddar Valley in Kashmir, into a general study of Himalayan and trans-Himalayan glaciers. They showed that the Kolahoi Glacier has been in retreat since the middle of the 19th century. However, there is no detailed information about the behaviour of Kashmir glaciers earlier in the Holocene. Unweathered and unvegetated moraines at the front of many glaciers indicate that Holocene advances have occurred. However, this problem is beyond the scope of this thesis.

Several sources of evidence have been put forward to support the idea that the end of the last glaciation in Kashmir had occurred by 18kaBP. The dates on loessic palaeosols in Kashmir basin were discussed earlier in this chapter and the dates themselves are listed in appendix 2. In addition, pollen profiles from two bogs in the Pir Panjal Range, at Toshmaidan and Butapathri, have been used to lend support to the idea of an early deglacierization. Singh and Agrawal (1976) claim to have found broad-leaved tree elements within the pollen profile from Toshmaidan Bog, the oldest date on which is about 15 kaBP. Dodia <u>et al</u>., (1984) also maintain that this is supported by a date of 17 kaBP for the occurrence of broad-leaved elements within the pollen profile from Butapathri. These dates were held to accord with the date of 18 kaBP for palaeosol development in the Kahsmir loess.

At first sight, this chronology presents an interesting palaeoclimatic problem. Global evidence suggests that during the period 21 to 18 kaBP, glaciers and ice-sheets were at their maximum extent (Denton and Hughes, 1981). However, there are a number of reasons why the chronology presented above for Kashmir, and its interpretation, may be incorrect. Firstly, the pollen profiles from Toshmaidan and Butapathri totally disagree. The dates themselves are also suspect, since there is an inversion in the basal part of the Toshmaidan chronology, and the Butapathri date of 17 kaBP was carried out on a 20 cm-thick section of the core (see appendix 2). Secondly, the dates on the loessic palaeosols have no direct bearing on deglacierization, since ice never extended into the Kashmir Valley itself. Furthermore, the dates on the loess must be regarded as suspsect, since radiocarbon dating of palaeosols is notoriously unreliable. The chronology of the loess has been substantially revised with the use of TL dating (Bronger _et al_., 1987; H. Rendell. personal communication). Thirdly, no plausible explanation has been put forward to support the idea of an early deglacierization. Clearly, much more evidence from Kashmir, a regional evaluation of data and a consideration of prevailing climatic systems are required in order to resolve this problem.

3.4 Evidence For Changes In The Monsoon During The Quaternary.

Evidence that the monsoon circulation has undergone dramatic changes during the Quaternary comes from six types of source: oceanic sediments from the Bay of Bengal, Arabian Sea and Indian Ocean; lacustrine sediments from north-west India; alluvial sequences from north-central India; supporting evidence from areas adjacent to south Asia; and climatic models. Broadly, this evidence suggests that at the last glacial maximum, the south-west monsoon circulation was weakened considerably and the north-east monsoon strengthened. An understanding of these changes is an essential prerequisite for the interpretation of Himalayan glacial records, since the advance and retreat of glaciers is controlled largely by temperature and precipitation.

Species assemblages and stable-isotope ratios of foraminifera from oceanic cores have been used to indicate changing salinity patterns in the northern Indian Ocean during Late Quaternary time (Cullen, 1981; Duplessy, 1982). The results indicate a much stronger north to south salinity gradient in the northern Indian Ocean at the last glacial maximum than at present. In the Bay of Bengal, this was explained by reduced discharge from the Ganges-Brahmaputra-Irrawaddy river systems. In the Arabian Sea, it was argued by Duplessy (1982) that evaporation over the north-eastern part of the sea and the Gulf of Oman was much stronger. Discharge of the Indus river was also reduced. Both Cullen (1981) and Duplessy (1982) maintained that this salinity pattern indicates much reduced precipitation during the last glacial maximum associated with the weakened south-west monsoon. Enhanced evaporation over the northern Arabian Seas was explained by reduced summer upwelling due to the weakened summer winds. Duplessy (1982) further suggested that the winter north-east monsoon was greatly enhanced at the last glacial maximum. He argued that this would increase winter upwelling in the Arabian sea, and so strengthen evaporation. This would add to the steep salinity gradient observed in the foraminiferal record.

Additional studies of oceanic sediments have provided more direct evidence of a strengthened 'glacial' north-east monsoon. Kolla and Biscaye (1977) found an increase in the quartz content of northern Indian Ocean sediments at about 18 ka BP. They inferred that this quartz had been blown from the adjacent South Asian continent, and represented enhanced aridity and windspeeds at that time.

Pollen records from Arabian Sea and Indian Ocean sediments show high numbers of saline-littoral and steppe taxa at the last glacial maximum (Van Campo _et al_., 1982; Van Campo, 1986; Prell and Van Campo, 1986). Pollen from a source in the Mediterranean steppe region, probably Iran and Afghanistan, was thought to have been wind-transported by an enhanced north-east monsoon. The low frequency of savanna and humid-tropical pollen at the last glacial maximum was held to indicate a much reduced south-west monsoon.

Fontugne and Duplessy (1986) suggested that the pollen evidence still provides only an indirect indication of an enhanced winter monsoon at the last glacial maximum. From work on carbon-isotope variations of organic matter within ocean sediments, Fontugne and Duplessy claimed to show more direct evidence of the glacial winter monsoon. In areas not affected by terrestrial input of organic matter, carbon-isotope ratios reflect oceanic palaeoproductivity. It was found that productivity was enhanced in the north-eastern part of the Arabian Sea at the last glacial maximum. Fontugne and Duplessy argued that this was a result of increased upwelling in this area due to the strengthened north-east monsoon.

Several studies in northern India have shown that the last glacial maximum was a period of extreme aridity. Analysis of small lakes in Rajasthan has indicated a hyperarid phase, prior to about 13 kaBP when the south-west monsoon was reestablished (Singh et al., 1972, 1974; Bryson and Swain, 1981; Swain et al., 1983; Wasson et al., 1984). In the Rajasthan desert, active dune formation occurred in areas where dunes are now stable between at least 20 kaBP and the mid-Holocene (Wasson et al., 1983). Williams and Clarke (1984) have shown that loess was accumulating in the Son and Belan Valleys of north-central India between 25 and 17 kaBP. Himalayan pollen within the loess suggested strengthened north-easterly winds during this period.

Pollen analysis has been undertaken on cores from several Himalayan bogs and lakes. Many of the studies, however, provide conflicting evidence and a number are undated. Visnhu-Mittre and Sharma (1984) analysed a bog in Kathmandu Valley, Nepal. Their results suggested a period of temperate climate prior to 40ka BP, characterized by the presence of oak pollen within the sediments. This was followed by an expansion of grassland vegetation between 40 and 17 kaBP, interrupted by an oak phase around 25 kaBP. The grassland vegetation was held to indicate a cold, dry environment. Vishnu-Mittre and Battacharyya (1980) also found evidence for vegetation change within the pollen from lake sediments in Ladakh, north-west Himalaya. The results showed two phases of <u>Juniperus</u> dominance, interpreted as an indicator of climatic amelioration, at around 30 kaBP and between 20 and 15 kaBP. The periods between 30 and 20, and 15 and 10 kaBP were characterized by alpine steppe vegetation.

A number of studies from south-west Asia have an important bearing on Quaternary climate of the Indian subcontinent. Roberts (1983) identified a major highstand of Lake Konya, in Turkey, spanning the last glacial maximum. This was attributed to the reduction in evaporation caused by lower temperatures rather than an increase in rainfall. Wasylikowa (1967) identified a high-stand of Lake Zeribar in Western Iran between 22.5 and 14 kaBP based on plant macrofossil evidence. Pollen evidence from the same lake (Van Zeist, 1967) indicated an <u>Artemisia</u> steppe vegetation during this period. From this, it was concluded that the lake high-stand was a result of reduced temperature, and therefore evaporation, rather than an increase in precipitation. However, this interpretation is open to debate. Water-balance considerations suggest that a massive drop in temperature would be required to sustain high lake-levels in conditions drier than present. Furthermore, Artemisia actually prefers slightly moister conditions than would have prevailed in the environment postulated above. It is possible, therefore, that the climate was slightly wetter than today along a belt stretching from the western Mediterranean through the near east and Central Asia to Tibet, which experienced higher lake-levels at the last glacial maximum than today (Street-Perrott and Harrison, 1985).

Simulations with general circulation models (GCMs) have also provided important information regarding changes in the monsoon (eg. CLIMAP, 1976; Gates, 1976; Manabe and Hahn, 1977; Kutzbach and Street-Perrott, 1985, Kutzback and Guetter, 1986). Manabe and Hahn argued that tropical glacial aridity was a result of stronger surface air outflow from, or weaker inflow to, the continents, due to the greater reduction of air temperature over the continents than over the oceans. In particular, they found that the weakened south-west monsoon at 18 kaBP was largely due to an increase in continental albedo. Kutzbach and Guetter's (1986) simulations lend some support to the idea that there was a belt of slightly wetter conditions during the last glacial stretching from the Mediterranean to Tibet as discussed above. Their 18 kaBP simulation indicated strengthened westerly storm tracks in both summer and winter passing over this zone. This effect, together with a slight decrease in temperature, could have led to slighlty wetter conditions during the last glacial, and provides an alternative explanation, to massive decrease in temperature and evaporation, of the lake-level data.

3.5 Archaeology.

Unlike the Siwalik molasse of the Himalayan foothills, the Karewa beds have not yielded hominid remains. In fact, the Karewa beds are devoid of evidence for human existence. The Neolithic of Kashmir has been studied quite intensively since the trial excavations at Burzahom by de Terra and Paterson (1939). However, the Neolithic in Kashmir occured during the Holocene and so postdates the period covered in this research.

De Terra and Paterson (1939) commented on the absence of stone tools from Kashmir and suggested that the climate may have been too inhospitable during the glacial period for human habitation. However, in 1969 excavations by H.D. Sankalia and coworkers revealed a number of Palaeolithic tools. Subsequent work in the early 1970s led to the discovery of over 20 tools from several sites in Kashmir (Sankalia, 1971; Joshi et al., 1974).

The original finds came from a boulder deposit near Pahalgam in the Liddar Valley (Sankalia, 1971) They included a hand-axe, scrapers and a flake. Sankalia proposed that the tools came from deposits of 1st interglacial to 2nd glacial age, using de Terra and Paterson's (1939) chronology. A borer, found in '3rd glacial' screes at Ganeshpur, in the lower Liddar Vallley, was regarded as a typological equivalent of the Middle Palaeolithic of Peninsular India. Additional finds were made in the Sind Valley. These included a 'hand-axe-cum-chopper' from terrace gravels at Nunar, a scraper from supposed glaciofluvial deposits at Prang, and a scraper from '2nd glacial' tills at Sonamarg.

R.V. Joshi and coworkers reinvestigated the site at Pahalgam in the Liddar Valley (Joshi _et al_., 1974). Both Sankalia and Joshi _et al_. identified two beds of gravel at Pahalgam, which were overlain by 'brownish silt.' The boulder deposits at Pahalgam, and hence the palaeoliths found therein, were regarded as Middle Pleistocene by Joshi _et al_. (1974) by means of correlation with bone-bearing beds at Sambur, in Kashmir Valley. However, as the authors readily admitted, it was not possible to date this deposit confidently, in the absence of associated radiometrically-datable material.

CHAPTER 4. EVIDEN ENVIRONMENTAL CHANGE FROM THE LOWER KAREWA FORMATION.

4.1 Introduction.

In chapter 3, the palaeoenvironmental work undertaken on the lower Karewa beds was reviewed. In order to reconstruct Late Cainozoic environmental changes in Kashmir, it is important to choose evidence that occurs more or less continuously throughout the sequence and which can be interpreted in palaeoenvironmental terms. Since no single indicator fulfils both of these requirements, it is necessary to pursue as many different lines of evidence as possible. The previous work on the pollen has shown that this is both discontinuous in the sequence and particularly difficult to interpret. Of the other fossils found in the Karewa beds, vertebrate remains are restricted stratigraphically (Kotlia, 1984). Ostracods have not been studied in sufficient detail to determine the extent to which they occur continuously in the lower Karewa. However, the analysis of lower Karewa samples during the course of this study showed that ostracods were absent from most of the 70 samples which cover the whole of the sequence. Charophytes were shown to be even rarer than ostracods.

Of the previous studies, only the work by Burbank and Grant (1985) on particle-size periodicities, and the carbon-isotope and C:N analyses of Krishnamurthy et al. (1986) span the whole sequence. In the present research, environmental changes in the lower Karewa were investigated by studying the clay mineralogy of mudstones and surface texture of quartz sands. Since sandstones and mudstones occur commonly throughout the lower Karewa sequence, these methods fulfil the requirement of continuity.

4.2 The Environmental Interpretation of Clay Minerals.

The rationale for using clay minerals as palaeoenvironmental indicators lies in the assumption that different climates lead to the formation of characteristic clay-mineral assemblages. Although this approach is not yet routinely employed, it has been used increasingly over the last 20 years (eg. Singer, 1980, 1984). Clay-mineral studies are frequently used to complement other lines of palaeoenvironmental evidence or, more rarely, on their own where other evidence is unavailable.

Since clays are ubiquitous and relatively easy to analyse, their use in palaeoenvironmental reconstruction holds considerable attraction. However, Singer (1980, 1984) has warned that they should be employed with care, and preferably alongside other types of evidence. Furthermore, the simple presence or absence of a clay mineral provides only low-order palaeoenvironmental data. For more detailed work, it is necessary to obtain an indication of the relative abundance of individual clay minerals and also indices of crystallinity and composition. Quantitative or semi-quantitative analysis of clay minerals is much more difficult than simple identification and should be approached with care. Indices of crystallinity and composition, whilst easier to determine, are not straightforward to interpret.

Clay minerals are formed in three ways. They may be inherited from other environments, neoformed by crystallization from solution or through the reactions of amorphous materials, or transformed by ion-exchange or layer transformation (Eberl, 1984). In using sedimentary clay minerals as paleoenvironmental indicators, it is assumed that the clays are transformed and neoformed in soils and weathering profiles, and that the type of minerals found depends on parent material and climate. The clays are then eroded, transported and deposited to form a sediment. It is assumed that little significant alteration of the clays occurs between erosion and analysis. Under this model, the sedimentary clays indicate palaeoenvironmental conditions in the surrounding basin where weathering and pedogenesis took place. Any study of clay minerals in sediments will inevitably involve a consideration of clays in soils and weathering profiles as well.

Singer (1980) suggested five assumptions that underlie the use of clay minerals in soils as palaeoclimatic indicators. Firstly, it is assumed that there is a direct and detectable relationship between climatic variables and clay-mineral assemblages. Singer suggested that precipitation may be the most important climatic control, since the rate of chemical weathering in a soil profile depends largely on vertical water movement or leaching. Higher rates of leaching will promote the production of more clay-sized materials and minerals that are related to more advanced weathering stages. Thus, Singer (1966) found a decrease in smectite in contemporaneous basaltic soils in the Galilee, Israel, with increasing rainfall. Smectite is found where soluble elements can accumulate, either in drier climates or in waterlogged soils. On a much larger scale, individual clay minerals are often regarded as being of high- or low-latitude origin. Chlorite and illite are usually associated with polar latitudes, and kaolinite with tropical latitudes (eg. Biscaye, 1965). According to this view, chlorite and illite are inherited products when they occur in high latitude soils or sediments, since they reflect minimal alteration of preexisting clay minerals in an environment of low weathering intensity. Kaolinite, on the other hand, indicates higher rates of leaching in an active chemical environment.

The second assumption made by Singer (1980) is that clay minerals remain stable once formed by weathering, in the absence of climatic change or tectonic rejuvenation. If clay minerals form by weathering of preexisting minerals, they should remain stable in the environment of formation. However, if they are inherited from elsewhere, they may undergo change, depending on the difference between the source and destination environments and the stability of the clay mineral in question.

Singer's (1980) third assumption is that clay minerals are distributed uniformly throughout the soil or weathering profile. However, he presents evidence to show that this assumption is incorrect and that the clay mineral assemblage of any soil depends partly on the depth of the sample in the profile since the effects of leaching vary with depth. Although this has obvious implications for the study of clays in soils and weathering profiles, it is of less importance in the study of sedimentary clays. In any case, it is impossible to determine the depth in the soil or weathering profile from which inherited sedimentary clays were derived.

Fourthly, Singer (1980) assumes that buried clays are stable. He argued that this assumption is valid for palaeosols, even when buried quite deeply, since the depth of burial would be insufficient to promote diagenesis and the interstitial waters, being meteoric rather than marine, would not be of suitable chemical composition to promote such changes.

Singer's (1980) final assumption is that clay minerals have a uniform sensitivity towards environmental change and that once formed, they preserve the palaeoenvironmental information throughout subsequent changes in the environment. This assumption is clearly not warranted in the case of pedogenic clays. As Singer points out, processes of clay-mineral degradation are much more rapid than those of aggradation. Thus, the clay-mineral assemblage in a soil will selectively record periods when degradation was active compared to those dominated by aggradation.

These five assumptions apply to sediments as well as to soils and weathering profiles. However, there are two additional considerations when dealing with sedimentary clays. Firstly, large sedimentary basins often consist of a variety of lithologies and geomorphological environments. Thus, the basin will receive an 'average' mixture of clay minerals that depends, not only of the types of lithologies present, but also on the climate of the source-area. However, unless there are significant changes in the source area, variations in the clay-mineral assemblage in the sediment can be interpreted in environmental terms. Secondly, deposition in the sedimentary column effectively seals off the clays from subaerial processes, thus reducing the possibility of postdepositional alteration by this means.

The problems of using clay minerals as palaeoenvironmental indicators are reflected in the fact that a number of nonclimatic factors are important in clay mineral formation. One of the most difficult problems is that of diagenesis subsequent to burial. Diagenesis is well documented in marine sediments. Some authors suggest that diagenesis is less pronounced in lacustrine environments, since the intersitial waters,which are of meteoric origin, are less reactive than those of marine origin. However, diagenetic alteration has also been noted in lake clays as well. Distinguishing between the detrital and authigenic components of an assemblage is often very difficult. In particular, the suite smectite - mixed-layer clays - illite is thought often to be a product of diagenesis (Singer, 1984).

There are a number of approaches to detecting diagenetic changes in clay-mineral assemblages. Highly-crystalline smectite is often regarded as authigenic, whereas poorly-crystalline smectite is assumed to be detrital (eg. Froget, 1981). However, this distinction is not always reliable and a detailed consideration of gradations and transitional phases in the assemblage is often required to make the distinction more reliable. In contrast to the clay minerals in soils, those in sediments are eroded from a source, transported and then deposited. The transfer path for each mineral does not always operate evenly, due to the variable erodibilty of different minerals. For example, finer-grained minerals such as illite and the mixed-layer clays have been found to be eroded selectively from a soil surface (Rhoton et al., 1979).

Finally, it is important to recognise the effects of tectonic uplift on clay-minerals. Uplift can affect the clay-mineral assemblage in a number of ways. Firstly, uplift itself may lead to a change in climate within a region. This effect may be real or apparent. Real effects arise from a significant change in the local or regional boundary conditions as a result of mountain building. Apparent effects arise from the elevation of a source area, which may change local climate without significant large-scale effects. This has already been documented for a latitudinal change in source area. A temporary increase in kaolinite from the Cape Verde Basin, around the Oligocene-Miocene boundary, was explained by plate tectonic movements which led to the migration of the basin into the equatorial zone at that time (Chamley, 1979). Presumably, altitudinal movement of the source area would have the same effect. However, the Himalayas play such an important role in global climate that significant uplift would have had large-scale effects as well.

The second effect of uplift on clay minerals is through its influence on rates of erosion. Uplift of a source area will tend to increase rates of erosion. This, in turn, will reduce the exposure time of rocks to weathering, possibly increasing the importance of physical over chemical means of rock breakdown. This effect is enhanced by the fact that uplift itself will also produce a colder climate.

Thirdly, due to higher erosion rates and progressive removal of strata, uplift may lead to a change in the rocks exposed at the surface. Furthermore, differential uplift within a basin may lead to a change in the locus of erosion. In geologically complex basins, these effects may be the same as changing the source area and can have significant impacts on sedimentary clay-mineral assemblages.

From the preceding discussion, it is clear that the palaeoenvironmental interpretation of clay minerals in sediments is often problematical. Careful consideration of non-climatic controls on mineralogy, particularly diagenesis, tectonics and variations in source area are most important. Furthermore, even if these factors can be accounted for, the type of palaeoenvironmental information derived from clay mineralogy must be regarded as very generalized.

4.3 Chronological Framework and sampling Design.

The chronology upon which this part of the study is based was devised largely by Burbank (1982), with additional information from Kusumgar et al. (1985), both of whom used palaeomagnetic stratigraphy and fission track dating of volcanic ashes. The chronology has already been reviewed in chapter 3 of this thesis.

In this study, samples were taken from the sections exposed near Hirpur in the Rembiara Valley (plate 1) and between Pakharpura and Baltal in the Romushi Valley (plate 2). Together, these two sections are over 800 m thick and cover the bulk of lower Karewa sedimentation. The age of each sample was determined using the sedimentation rates from Burbank's study, although the rate for the Gilbert and Gauss chrons was recalculated using the revised age for the lowest palaeomagnetic boundary. A rate of 25 mma^{-1} was used for samples taken from the Gilbert and Gauss chrons, 32 mma^{-1} for samples from the Matuyama chron up to the top of the Olduvai subchron and 16 mma^{-1} for samples from the post-Olduvai. The location of each sample was determined in the field relative to key marker beds and the age was calculated by linear interpolation between the appropriate geomagnetic boundaries. The decrease in sedimentation rate during the upper Matuyama chron is poorly constrained (see Burbank and Grant, 1985) so that the ages of samples from this part of the sequence may be subject to greater uncertainty than the older samples.

The aim of the study was to detect evidence for environmental change on at least a 100 ka wavelength. Therefore, sampling had to be carried out at intervals of less than 100 ka. An a priori estimate of the number and spacing of samples indicated that 30 samples taken at roughly 28 m intervals would be required to provide a frequency of one sample per 100 ka. However, to provide a better indication of environmental change, 70 samples, each of 250-300 g, were taken, with an average frequency of 1 sample per 40 ka. In some parts of the sequence, however, sampling was constrained by lithology, since only the mudstones are suitable for clay mineral analysis. Prior to sampling in the field, any obviously slumped and weathered material was removed. The location of each sample was fixed relative to an identifiable bed such as a major sand unit or conglomerate.

Samples of sand, each of between, 50 and 100 g, were taken from each major sand unit and conglomerate bed in the lower Karewa sequence for subsequent SEM surface-texture analysis. 75 samples of sand were taken, with an average sampling frequency of a little under 1 sample per 40 ka. Following the removal of any slumped or weathered material. the sand sample was scraped carefully from the section.

4.4 Analytical Procedures.

a. Sample Preparation and X-Ray Diffraction Methods.

The preparation of samples and the X-ray procedures employed differed little from accepted practice (see, for example, Wilson, 1987). About 10 g dry weight of each sample was dispersed with 50 ml of 'Calgon' solution and stirred with a high-speed stirrer for 15 seconds. It was not found necessary to treat any of the samples to remove calcium carbonate or organic matter in order to aid dispersion. The dispersed suspension was then made up to 1000 ml with deionised water, and the clay (< 2 micron) fraction was withdrawn by pipette after settling according to Stokes' law. 50 ml of the suspension was then washed three times with deionised water and centrifuged between each washing. A smooth, thick slurry was produced and smeared onto a glass slide using a 10 ml eye dropper in order to obtain an oriented mount. Triplicate slides were made from each sample.

X-ray analysis was undertaken using a Philips Diffractometer in the department of Earth Sciences, University of Oxford. XRD patterns were determined from 2 to 30^0 2θ (equivalent to d-spacings of 44.17 to 2.98 Å) using CuK_α radiation and a monochromator. Goniometer and chart speeds were 1^0 and 1 cm per minute respectively. Full-scale deflection was 1 k counts per second for most samples and the time constant was 1. Slit widths for divergence, receiving and anti-scatter slits were 1^0, 1^0 and 0.3 mm respectively.

For each sample, one slide was run after drying at room temperature and humidity, another was run after solvation with ethylene glycol for 48 hours and a third was run after heating to 550^0C. All of the samples were run on the aforementioned diffractometer settings. In addition, slow scans were carried out across key peaks for several samples. The peaks scanned were those at 7 and 3.5 Å, with the aim of differentiating between chlorite and kaolinite. This approach follows Biscaye, (1964) and involves separating out the chlorite and kaolinite components of the peaks under high resolution scanning conditions. However, no peak separation was observed for the lower Karewa samples and the slow scans were not carried out on the rest of the lower Karewa samples.

The 2θ angles of the peaks recorded on the diffractograms were corrected using the 20.85^0 2θ reflection of the quartz naturally present in the samples as an internal standard. The 2θ values were then converted to d-spacings using standard tables.

Six samples of bedrock were also subjected to X-ray analysis. Small samples of the rock were ground to a fine powder using a mechanical rock crusher. The resulting powders were then made into thick slurries using deionised water and transferred to slides using an eye dropper. Subsequent procedures for mineralogical analysis were similar to those described above for the lacustrine mudstones.

b. Identification of Individual Minerals.

The oriented mounts used in this study give basal spacings of the minerals on X-ray diffraction. Since the <2 micron fraction was used, the main constituents of the samples can be assumed to be clay minerals. However, minor amounts of other minerals, including quartz, feldspar, dolomite, calcite, iron oxides and zeolite are sometimes present in the clay-size fraction. Quartz is more or less ubiquitous in most sediments and the same is often true of feldspar. Preliminary tests with dilute hydrochloric acid confirmed the absence of calcium carbonate from most of the lower Karewa samples. Iron oxide was present in some samples, shown by reddish-brown colouration, particularly around fossil plant remains. The criteria for identifying each group of clay minerals in the lower Karewa samples are discussed below.

Smectite Group

Previously, the term montmorillonite was used as the collective name for this group of clays. However, this has recently been adopted for a specific mineral, and the term smectite used as the group name (Booy, 1981). This terminology will be used here.

Smectites are dioctahedral or trioctohedral clays with a 2:1 layer structure and exchangeable interlayer cations (Hall, 1987). The basal spacings of smectites are highly variable, depending on relative humidity and the nature of the interlayer cations. The expandabilty of these minerals is an important feature used in their identification. Smectites often have layer spacing of 10 to 15Å, which expands to 17Å on ethylene glycol solvation and collapses to 10Å on heating to 550^{0}C for about 1 hour. Smectites are frequently interstratified with other minerals to give mixed-layer clays (see below).

Illite Group

Illites have a dioctahedral or trioctahedral 2:1 layer structure that is about 10Å thick. Numerous problems surround the definition of illite. The term was originally used by Grim et al. (1937) for clay-sized minerals of the mica group. Since then however, many clay-sized micas have been discovered, including mixed-layer illite/smectite (I/S). The term illite has therefore been defined much more specifically as a

> '...non-expanding, dioctahedral, aluminous, potassium mica-like mineral which occurs in the clay-size (less than 4

micron) fraction.' (Środoń and Eberl, 1984, p.495).

According to this definition, the authors state that illite has d(001) of 10±0.05Å, and an intensity ratio, I_r (the ratio of the (001)/(003) peak intensity in air-dried samples to that of ethylene glycol solvated samples), of 1 showing that it is nonexpanding. Środoń and Eberl (1984) also introduce the term 'illitic material' to include all clay-sized material of approximately 10Å. This term is regarded as synonomous with the term 'illite' as originally defined by Grim et al. (1937). However, Reynolds (1980) has pointed out that illite with no expandable component is rarely found in nature. Since an expandable component (indicated by an I_r less than 1) was found for many of the illites in lower Karewa samples , the term used here refers not to the strict definition of Środoń and Eberl (1984) but to their less closely-defined term 'illitic material.'

Illite in the lower Karewa samples was identified by peaks at 10, 5 and 3.3Å. Ethylene glycol solvation and heating generally led to small change in peak intensities. A slight change in peak position was also sometimes seen. Heating of smectites to 550°C for 1 hour also collapsed smectites and the expandable components of mixed-layer clays to about 10Å, and this increased the height of the 10Å peak following heat treatment.

Chlorite Group.

The chlorite group minerals are usually trioctahedral with a 2:1:1 layer structure about 14Å thick. The group is chemically varied, with three common types: Fe-rich, Mg-rich and Al-rich. The group is characterized by basal spacings at about 14, 7, 4.75, 3.56 and 2.85Å, although only the first four of these would appear on an X-ray diffraction scan from 2 to 30^0 2θ. An additional type of chlorite is the swelling variety. Whereas most types show no response to ethylene glycol solvation, swelling chlorites have an (001) basal spacing of 28Å that expands to 32Å on treatment. On heating, the position of the (001) peak of nonswelling chlorites does not change from about 14Å, but undergoes a marked intensification. The (001) peak of Fe-rich chlorites disappears at about 600^0C, although Wilson (1987) notes that they may be thermally unstable at lower temperatures than this.

Kaolinite Group.

The kaolinite group consists of minerals with uncharged, dioctahedral units, with a 1:1 layer structure (Hall, 1987). Kaolinite minerals are characterized by basal reflections at about 7 and 3.5Å. Ethgylene glycol solvation has no effect on either of these reflections, but they generally disappear following heat treatment at 550°C for 1 hour. However, Brown and Brindley (1980) note that a weak, broad band occasionally occurs between 12 and 14Å following heating.

Chlorite and Kaolinite.

One of the most difficult problems in clay mineralogy is the distinction between chlorite and kaolinite in mixtures containing both of these minerals. The problem arises since the (002) peak of chlorite corresponds to the (001) peak of kaolinite and the (004) of chlorite to the (002) of kaolinite. Although odd-order reflections of chlorite (eg. 001 and 003) should be diagnostic since they do not overlap with kaolinite, in practice the peaks of other minerals, especially smectites, may obscure them. A number of approaches have been adopted in order to solve this problem.

1. In some clay mixtures, the peak at about 3.5Å shows slight separation, into a 3.58Å chlorite component and a 3.54Å kaolinite component. This separation shows up on untreated samples, particularly with slow scans across the peak (eg. Biscaye, 1964). However, this does not hold for all mixtures and will depend on the crystallinity of the kaolinite and chlorite, together with the occurrence of any other minerals in the mixture. Fast scanning, together with slow scans of several lower Karewa samples showed that this approach was not applicable to the distinction between chlorite and kaolinite in this study.

2. Acid dissolution methods are sometimes employed, on the basis of the the fact that many commonly encountered chlorites are composed by hydrochloric acid, whereas kaolinite is unaffected. since this approach is not applicable to all mixtures

and employs elaborate preparatory and analytical procedures, it was not used in this study.

3. Thermal treatments may be used to distinguish between kaolinite and chlorite. As a general rule, chlorites are thermally stable to above 550°C whereas kaolinites are unstable. However, as noted before, iron-rich chlorites may be thermally unstable at lower temperatures.

In this study, the presence of chlorite was diagnosed as follows. For air-dried samples, the (001) of any chlorite in the sample would be obscured by the broad smectite peak that was almost invariably present. Similarly, the (002) peak would be coincident with kaolinite. However, the (003) of chlorite, at 4.75Å, was seen in air dried samples. On ethylene glycol solvation,, smectite expands to about 17Å, revealing the unaltered peak of any chlorite present, at 14Å. The 14Å chlorite peak is also revealed in heated samples, since the smectite peak collapses to 10Å. Heating also tended to collapse the 7Å kaolinite/chlorite peak, leaving a residual, the height of which was roughly proportional to the 14Å peak-height. This residual 7Å peak was interepreted as chlorite.

<u>Mixed-layer clays</u>.

'Interstratified' or 'mixed-layer' minerals are composed of different minerals stacked in a sequence. The ordering of the individual minerals may be regular or random. Structures that show regular interstratification behave as individual minerals. They are characterized by a large basal d-spacing which is equal to the sum of the two components, and a rational sequence of higher order reflections. Randomly interstratified structures do not have large basal d-spacings. Their fundamental reflections fall between those for the component minerals and the higher-order reflections are non-rational.

Examination of the X-ray diffractograms of lower Karewa samples showed that any interstratified minerals present are of the randomly-ordered type, since large basal d-spacings are absent. The most likely mineral to be present is interstratified illite/smectite (I/S). This is a much-studied mixed-layer clay which is common in sediments. Systematic shifts in the $illite_{(001)}/smectite_{(002)}$ and the $illite_{(002)}/smectite_{(003)}$ peaks have been documented (eg Środoń and Eberl, 1984). The amount of migration depends on the mineral composition. Peak migration curves have been prepared by Wilson (1987) from data in Reynolds (1980) in order to make an approximate estimate of the composition of mixed-layer I/S.

c. Quantification Procedures.

Semi-quantitative analysis of the X-ray data for the lower Karewa mudstones was undertaken in order to determine the relative amounts of minerals in each sample. A truly quantitative assessment of mineral abundance is not yet possible for several reasons. Firstly, there are often variations in the operating conditions of the X-ray machines. Secondly, the thickness of the sample on the slide often varies. Thirdly, variations in the degree of sample orientation often exist.

Fourthly, the crystallinity and chemical composition of different samples of the same mineral may vary. The first three sources of variation can be held more or less constant by standardizing sample preparation and analytical conditions. However, small variations in these three factors that may occur, together with the variations in crystallinity and chemical composition that are beyond control, mean that the peak height on an X-ray diffractogram cannot be equated directly with mineral abundance. However, various methods based on peak intensity have been used to estimate mineral abundance. It is important, however, to emphasize the fact that the results are only semi-quantitative. Conclusions drawn from such data should bear this firmly in mind.

One of the most often-used means of quantification involves weighted peak heights or peak areas. The use of areas rather than heights provides additional information, but has often not been employed, because of the time involved in making area measurements. However, with the widespread availability of digital planimetric instruments, area measurements are almost as quick to undertake as linear ones.

Although several approaches heve been used in peak-area analysis, they are almost all variations on a single theme. The one used in this study comes from the work of Johns _et_ _al_. (1954) and Biscaye (1965). The area of the glycolated 17Å smectite peak is related to four times the 10Å illite peak area and twice the 7Å chlorite/kaolinite peak area. Similar approaches heve been used by Jacobs (1970), Jacobs and

Hays (1972), Bowles (1975), Zheng (1984), Robert and Chamley (1987) and Yemane et al. (1987). These methods have been used on oceanic, lacustrine and aeolian sediments ranging in age from Cretaceous to Quaternary.

In order to differentiate between the relative amounts of chlorite and kaolinite in the lower Karewa samples, an approach based on work by Schultz (1964) was used. In Schultz' (1964) study, the height of the 14Å chlorite peak following heating to 550^{o}C was found to be equal to approximately two-thirds of the overall chlorite contribution to the 7Å peak for air-dried samples. For the lower Karewa samples, a direct relationship was observed between the height of the 14 and 7Å peaks following heating to 550^{o}C. Therefore, two-thirds of the area of the 14Å chlorite peak following heat treatment was used to determine the amount of the untreated 7Å peak which was attributable to chlorite.

More detailed information regarding the crystallinity and chemical composition of minerals may be of considerable palaeoenvironmental value. Several method were used in this study. For illite, the 2θ values for the (002) and (003) peaks were plotted. Środoń (1984) has distinguished between an illite field and an I/S field on such a plot. The intensity ratio (I_r) was also calculated for the illite in each sample. This is given by:

$$I_r = \frac{I(001)/I(002)_{\text{air-dried samples}}}{I(001)/I(002)_{\text{glycolated samples}}}$$

This formula is a modification of the one used by Środoń (1984). The only difference to Środoń's formula is that the (002) peak of illite is used here in place of the (003). The reason for this is that the (003) peak in the lower Karewa samples overlaps with quartz. The crystallinity of illite can be expressed by a crystallinity index, I_c, such as that of Kubler (1964). This is the ratio of the (001) peak width at half height above background to the peak height. However, I_c depends on the quantity and nature of the expandable layers present in the mineral as well as the crystallinity. Thus, it is only a measure of crystallinity in true illites (Środoń, 1984). An estimate of the percentage of illite in mixed-layer I/S can be determined using the position of the (001)illite/(002)smectite and the (002)illite/(003)smectite on the diffractograms, and the data from Reynolds (1980). This information is presented graphically in figure 9.

Two measures were used to characterize smectites. Firstly, the 'valley to peak' ($S_{v/p}$) ratio was calculated. This is the ratio of the height of the (001) peak of of smectite above the valley on the low-angle side to the height above background (Biscaye, 1965). However, this ratio only gives an approximation of crystallinity and is also affected by the relative abundance of the mineral and the

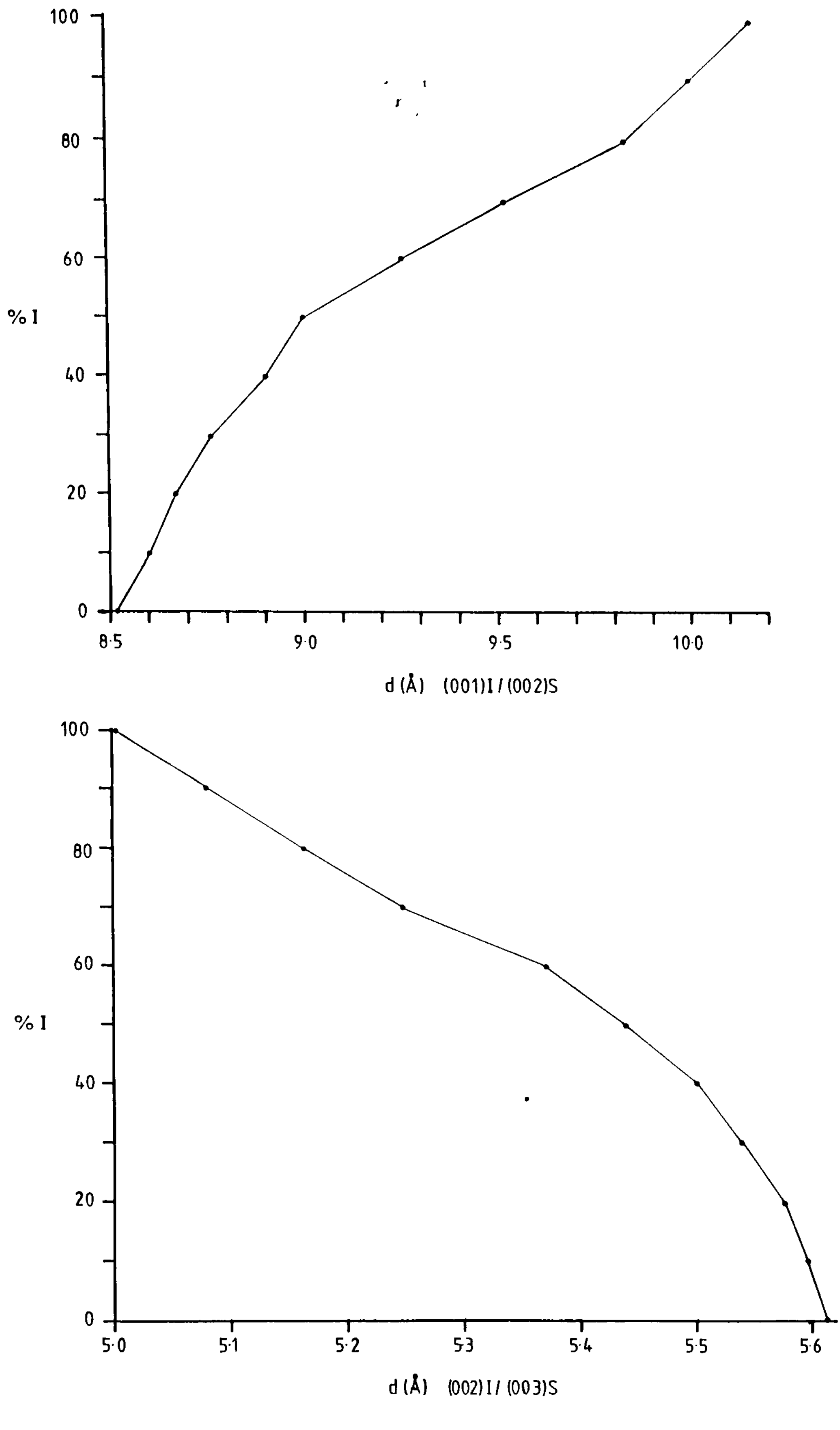


Figure 9

Peak migration curves for (001)I/(002)S and (002)I/(003)S for the estimation of the composition of mixed-layer clays (From data in Reynolds, 1980) presence of any interstratification. Eslinger and Savin (1976) described the ratio of peak height above the valley on the low angle side of the (001) of smectite to that on the high angle side (the $S_{l/h}$ ratio). This measure gives a rough indication of the proportion of expandable layers in the mineral. A larger ratio indicates a higher proportion of smectite layers. Eslinger and Savin (1976) used the computer-generated diffractograms of Reynolds and Hower (1970) to determine the relationship between the l/h ratio and the percentage of smectite layers. These data have been used to construct the curve in figure 10.

d. SEM analysis of sand grains.

The sand samples were sieved through test sieves of 250 and 500 micron sizes, in order to obtain the medium-sand fraction. About 1 g of this fraction was then boiled in dilute hydrochloric acid for 5 minutes, in order to remove calcium carbonate, which would otherwise bind the sand and obscure surface features on the quartz grains. Following rinsing in deionised water, the sand samples were then boiled in stannous chloride solution for 5 minutes, in order to remove any iron oxide coatings from the grain surfaces. The samples were then washed in deionised water and dried in an oven at 105^{0}C. 30 quartz grains were then picked randomly from the sample, and mounted on an SEM stub using double-sided adhesive tape. The mounted samples were then gold coated, prior to examination under a Cambridge Instruments Stereoscan SEM.

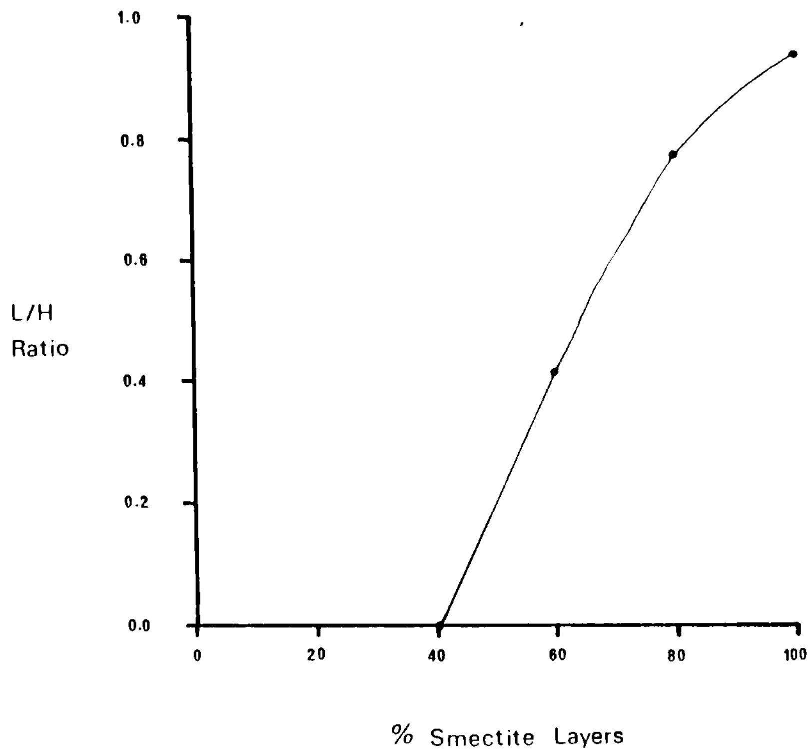


Figure 10

L/H ratio and percent Smectite in mixed-layer clays
(From data in Reynolds and Hower, 1970)

4.5 Results.

Most of the lower Karewa samples showed broadly similar diffraction patterns. Only the relative abundances, composition and crystallinity of minerals varied between samples. The trend in the relative abundance of each mineral in the lower Karewa is shown in figure 11 along with the changes in the indices of crystallinity and composition. The clay mineral data are summarized in appendix 4. The physical stratigraphy and magnetic polarity stratigraphy of the Karewa sequence shown in figure 11 are taken from Burbank (1982). Some examples of diffractograms for lower Karewa samples are shown in figures 12 to 14. In this section, the interpretation of the diffractograms and the trends in clay mineral data are discussed.

The first reflection on most diffractograms of untreated samples occurred at around 14Å. This peak was generally broad. Upon ethylene glycol solvation, it separated into two components: one at 16 to 17Å and another remaining at 14Å, often as a shoulder on the broader, higher angle peak. Upon heating to 550^{0}C, the 16 to 17Å peak collapsed to 10Å, so intensifying the 10Å peak. The 14Å peak remained unchanged in position, but generally intensified.

The 14Å peak on air-dried samples is interpreted as smectite and chlorite. The expansion of one mineral to 16-17Å indicates smectite and a generally high l/h ratio suggests a large number of expandable layers in the mineral. Since the 14Å peak remained unchanged on ethylene glycol solvation, the chlorite is of nonswelling variety.

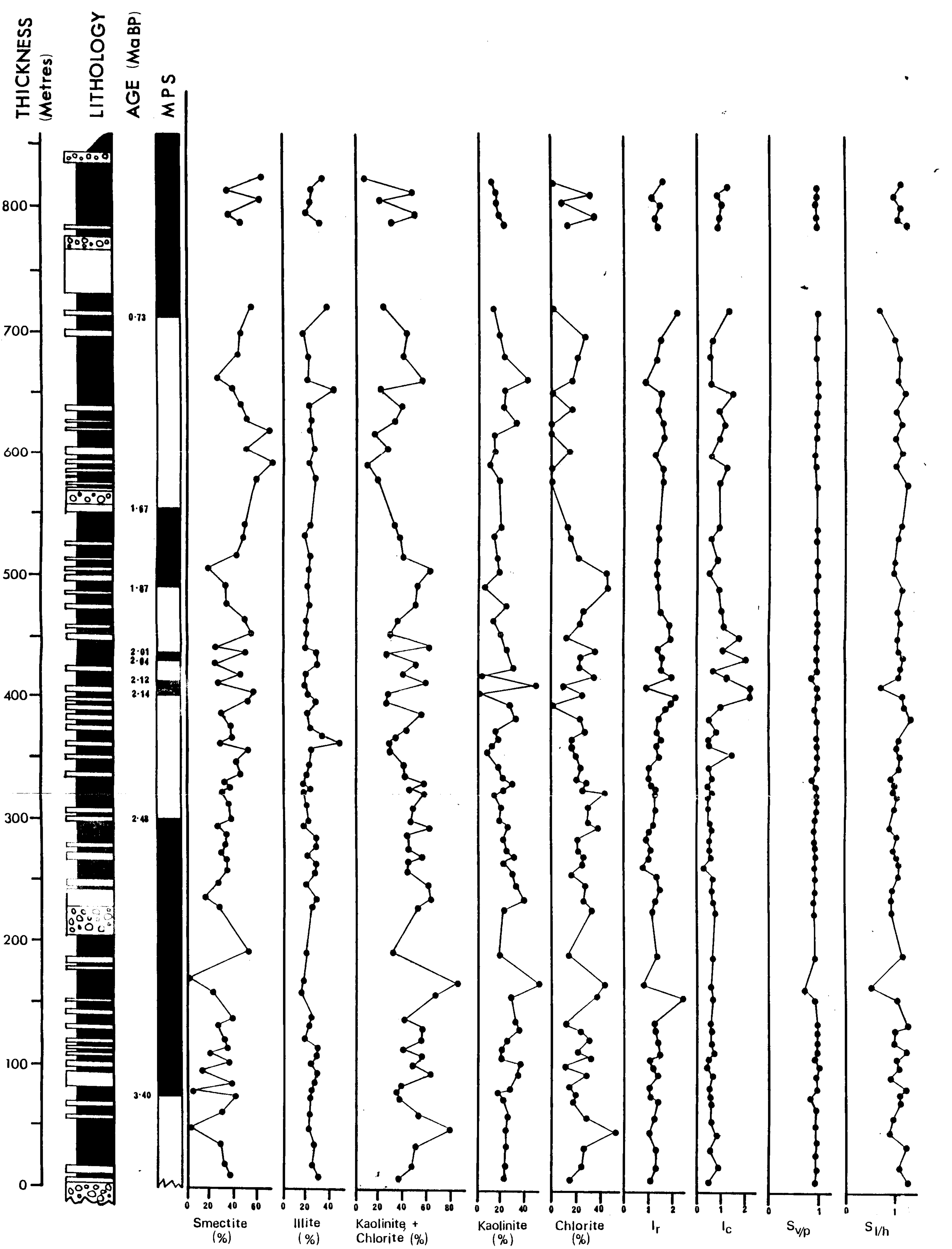


Figure 11. Clay mineralogy of the lower Karewa mudstones. (magnetic polarity stratigraphy and lithological information from Burbank, 1982).

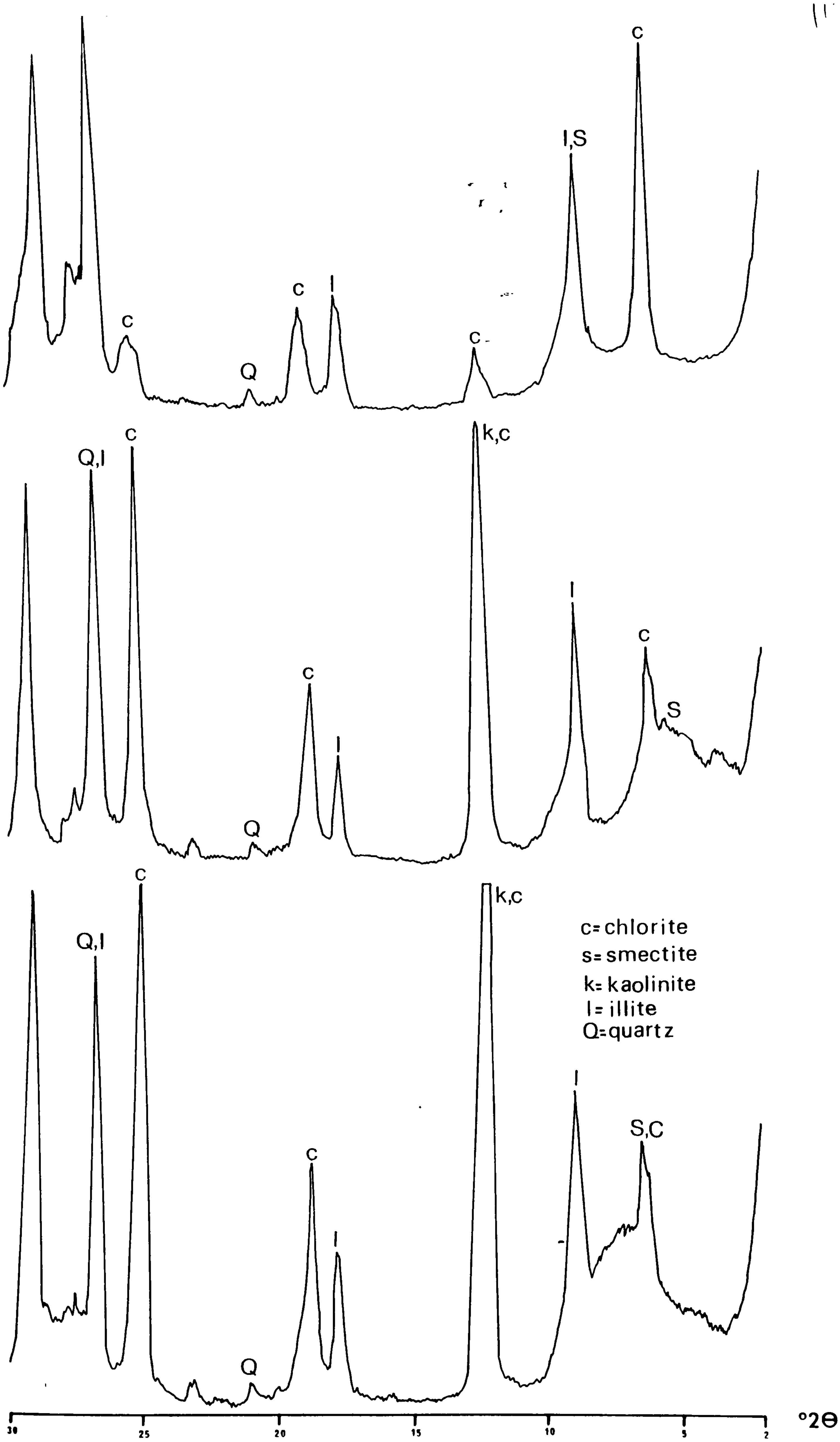


Figure 12

X-ray diffractograms for sample H32. Lower scan for air-dried sample, middle scan for glycol-solvated sample and upper scan for heated sample.

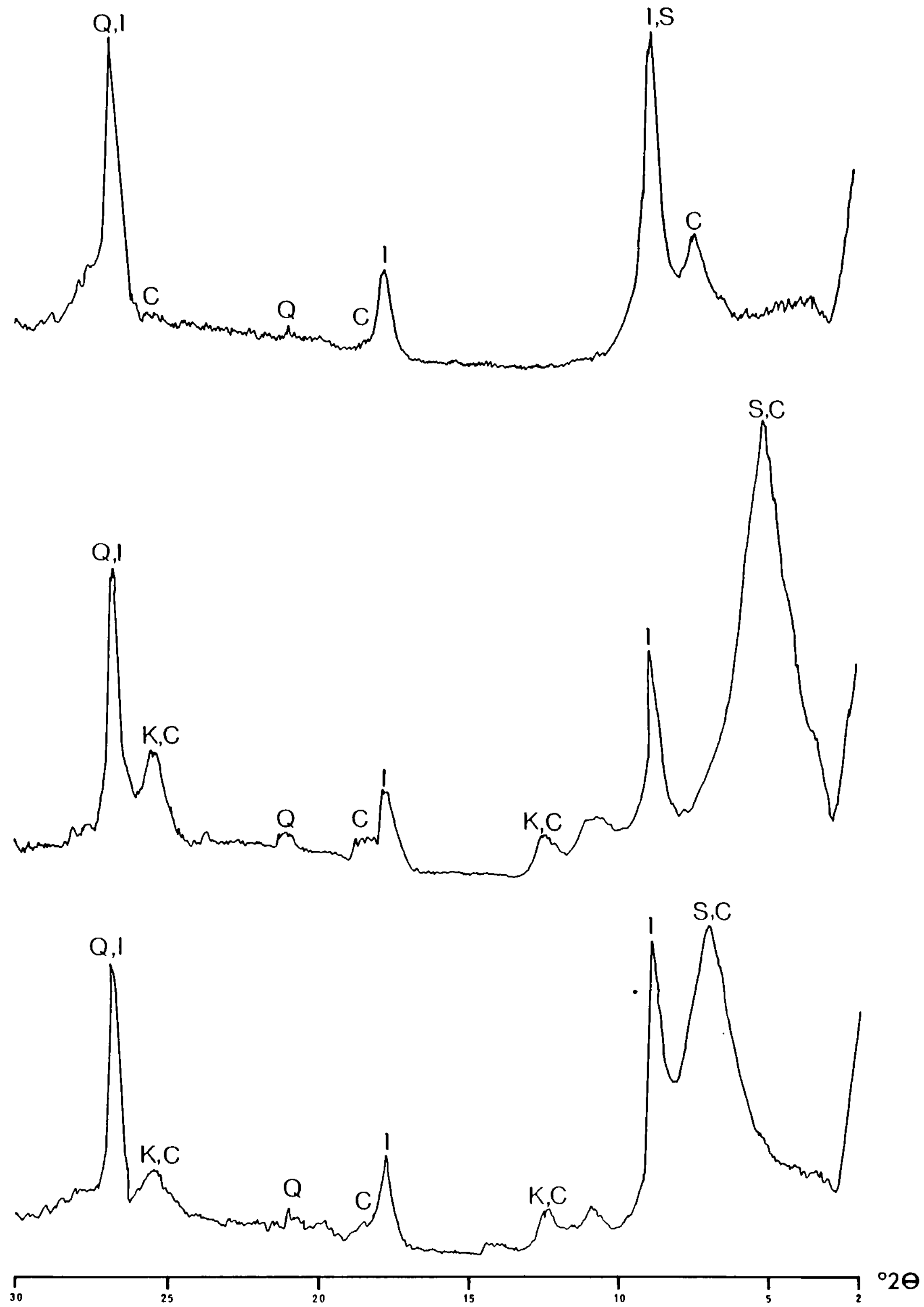


Figure 13

X-ray diffractograms for sample R27. Lower scan for air-dried sample, middle scan for glycol-solvated sample and upper scan for heated sample.

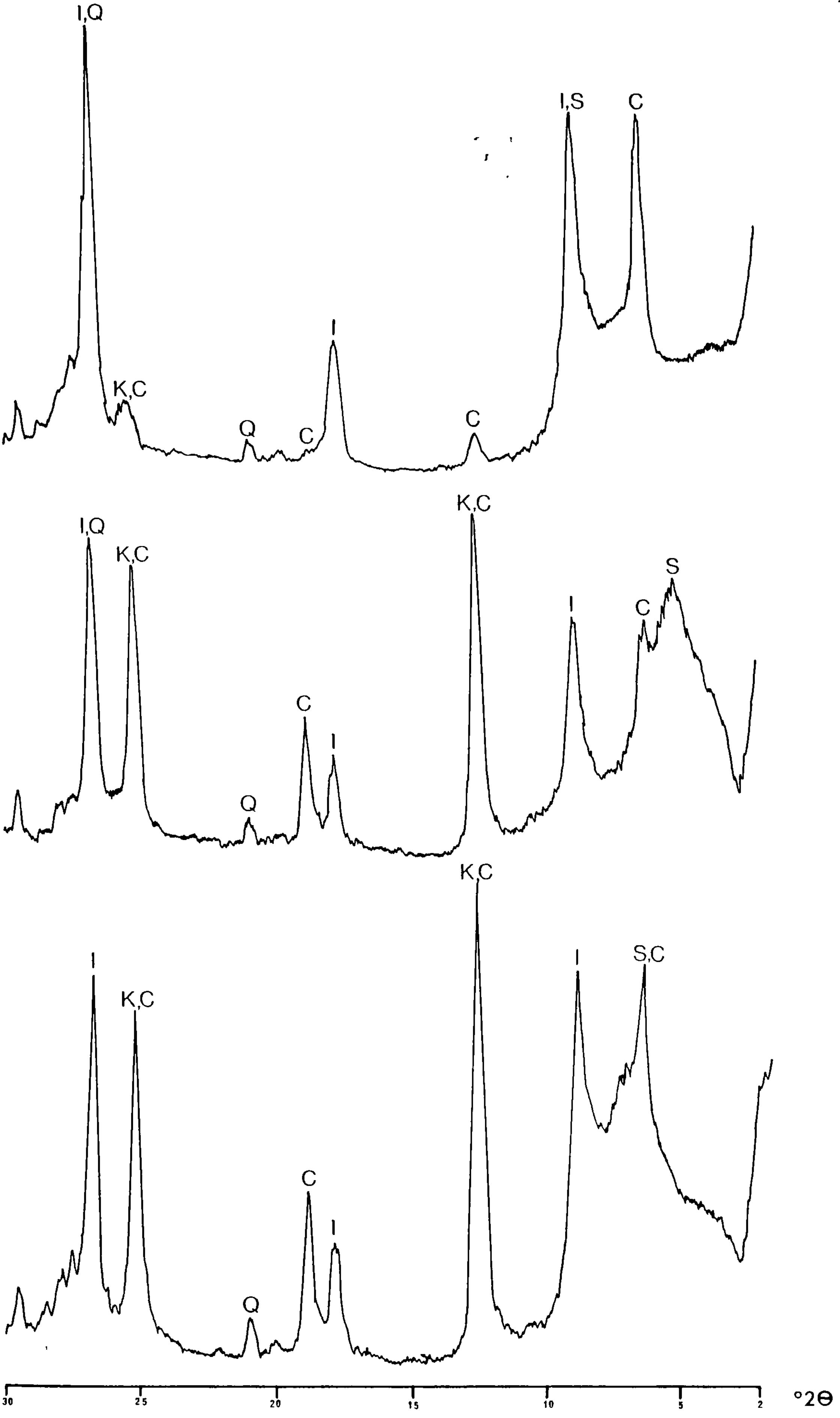


Figure 14

X-ray diffractograms for sample R19. Lower scan for air-dried sample, middle scan for glycol-solvate sample and upper scan for heated sample.

Upon heating, the smectite collapsed to 10Å but the 14Å chlorite peak remained unchanged. This suggests that the chlorite is not iron-rich.

The prominent peak at 10Å, which was unaffected by ethylene glycol solvation or heating, was interpreted as illite. However, the I_r value for the mineral is generally less than 1 suggesting the presence of expandable layers. As such, it is more properly termed 'illitic material' (Srodon and Eberl, 1984) although the term 'illite' will be used here for the sake of brevity.

A major peak occurs on all diffractograms at 7Å. This peak showed no change on ethylene glycol solvation, but diminished on heating to 550^{0}C leaving a residual peak of varying height. This peak is interpreted as chlorite and kaolinite. Diminution of the peak height on heating suggests the presence of kaolinite, since this mineral is thermally unstable. However, the residual peak suggests that chlorite is present as well and this is confirmed by the 14Å peak as discussed above.

A small peak at 4.9Å, showing no change on ethylene glycolsolvation or heat treatment, is interpreted as the (002) illite reflection. This peak was generally one third of the height of the (001) illite peak, at 10Å.

The presence of chlorite is additionally confirmed by the reflection at about 4.7Å. This suggests the (003) reflection of chlorite, which does not overlap with kaolinite. In general, this reflection showed no change upon ethylene glycol solvation or heating. However, in a few samples, the peak was reduced in height following heating even when higher order reflections remained unchanged. The interpretation of this slight thermal instability of the (003) of chlorite is open to question, although the chlorite in such samples may be iron-rich.

A small peak occurred at 4.25Å in most samples. Since this showed no change in intensity or position on either ethylene glycol solvation or heating, it is interpreted as quartz.

A peak at 3.5Å occurred in virtually all samples and showed very similar behaviour to the 7Å peak. No change was seen on ethylene glycol solvation, but the peak was reduced in height on heating, leaving a small residual. This peak is therefore interpreted as the (002) of kaolinite and the (004) of chlorite.

The final prominent peak on the diffractograms occurred at about 3.3Å. This generally large peak showed no change on either ethylene glycol-solvation or heating, and is interpreted as quartz and illite.

The samples of lower Karewa mudstone analysed in this study are, therefore, a mixture of minerals. They comprise smectite, kaolinite, chlorite, illite and quartz, in variable quantities. Some samples show an absence, or a very small relative proportion, of one particular mineral. Although the illite present generally contains an expandable component, the material from each sample falls within the illite field

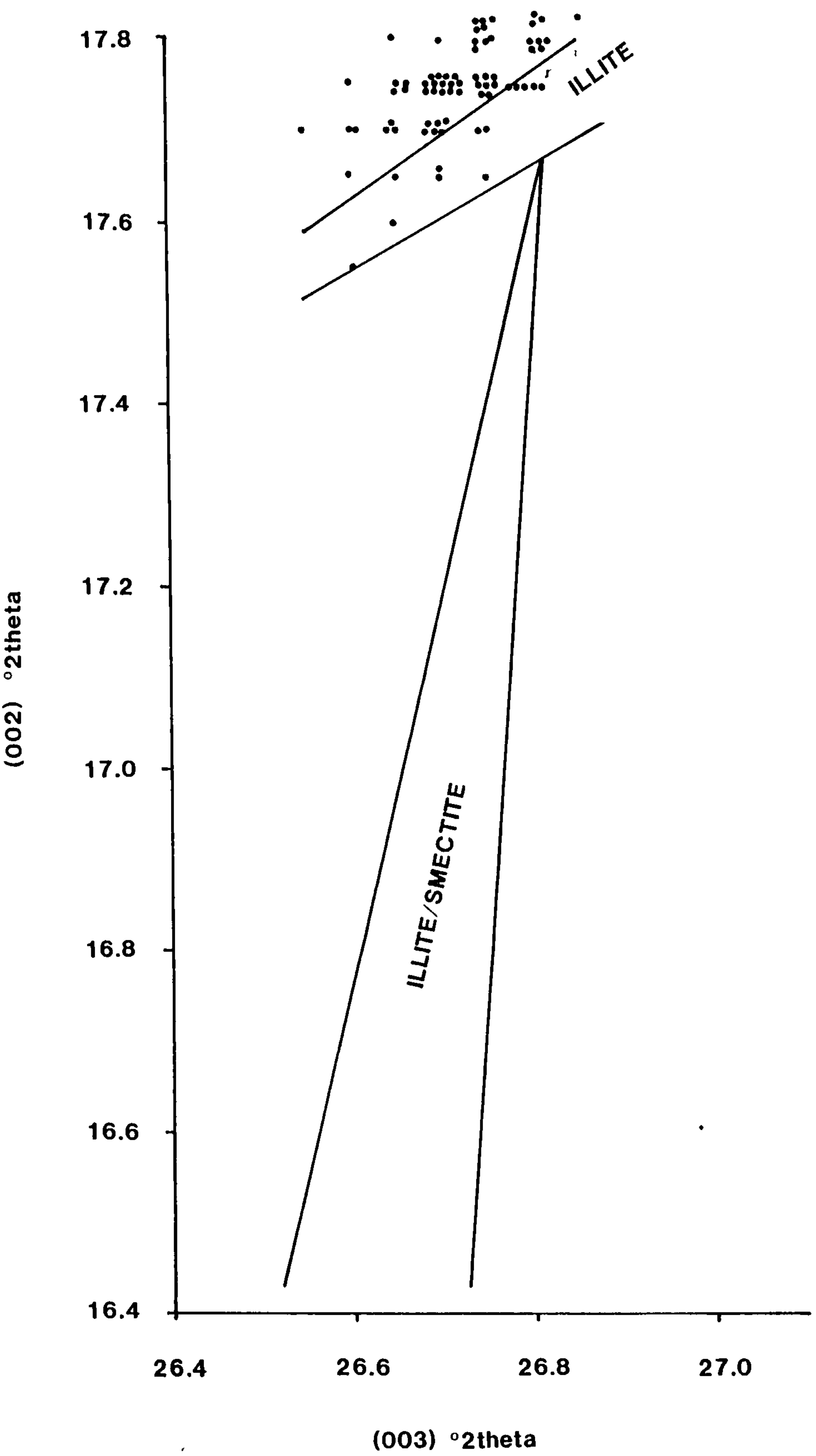


Figure 15

Plot of the (002) and (003) peak positions of illitic material (degrees 2-theta). Positions of the illite and mixed-layer fields follow Środoń (1984).

of Środoń (1984) (figure 15). This indicates that the mixtures are dominated by illite, and contain very small amounts of I/S (figure 9). A comparison of the lower Karewa smectites with the computed diffractograms of Reynolds and Hower (1970) suggests a very high proportion of expandable layers in the samples of this mineral (figure 10). The trends in the relative abundance of the minerals and mineral indices are dicussed below.

The abundance of smectite varied between 0 and 65%. On a broad scale, smectite increased during the Upper Cainozoic. During the upper Gilbert and Lower Gauss chrons, up to the 2nd conglomerate (around 2.92 MaBP), values of smectite are generally less than 40%. Sample H33, at about 190 m from the base of the sequence, seems to be an exception to this pattern. However, too much reliance should not be placed on a single sample. After the mid-Gauss chron, there was a steady rise in smectite, with lower-magnitude fluctuations than in the lower Gauss. By the lower Matuyama chron, at about 2.37 MaBP, there was a definite increase in the relative abundance of smectite, with values fluctuating around 40%. High but fluctuating values of smectite were sustained into the upper Matuyama and Brunhes chrons.

The main feature of the illite curve is its invariance, although the relative abundance of illite varied between 14 and 47%. There were, however, two small peaks of high illite abundance in the Matuyama chron. Illite also showed a slight increase during the Brunhes chron.

The joint contribution of chlorite and kaolinite varied from between 7 and 83%. The most striking feature of the curve is that it is an almost identical mirror-image of the smectite curve. Values of chlorite and kaolinite were generally greater than about 40% in the upper Gilbert and Gauss chrons. Values fell to below about 40% after the lower Matuyama, at about 2.37 MaBP. Chlorite and kaolinite values were particularly low after the mid-Olduvai subchron (about 1.8 MaBP). In the upper Matuyama and Brunhes chrons, the chlorite and kaolinite contribution fluctuated quite widely.

The individual relative abundances of chlorite and kaolinite varied from 0 to 54 and 0 to 50% respectively. To some extent, the mineral curves are mirror-images of each other, but there is not a perfect inverse relationship. Chlorite shows the most obvious inverse relationship with smectite. There was an overall, but slight, decrease in chlorite from the Gilbert-Gauss chrons up until the Matuyama chron. After the Olduvai subchron, chlorite abundance showed much larger fluctuations. Values of kaolinite were generally much greater than 40% prior to about 2.3 MaBP. After this time, there was a decrease in kaolinite with relative abundances less than about 40%.

The index I_r varied between 0.7 and 2.35. Since I_r is close to 1 for many of the lower Karewa illites, the index I_c, discussed below, is a true measure of illite crystallinity. As with the relative abundance of illite, I_r showed little variation in the sequence, with values fluctuating around unity. In the upper Gilbert and lower Gauss chrons, values of I_r were a little above 1, indicating slight expandability. An increase in I_r occurred at about 3.1 MaBP although this increase is seen in only one sample. From the upper Gauss to lower Matuyama chrons, I_r values remained very slightly above 1. After about 2.3 MaBP, I_r rose to slightly higher values, fluctuating around 1.5. Further increases in I_r occurred into the mid-Brunhes, possibly reflecting conditions more favourable for the formation of smectite.

Values of the illite crystallinity index, I_c, range from 0.32 to 2.17. From the upper Gilbert to the lower Matuyam chron, values of I_c were consistently low, indicating highly crystalline illite. After about 2.32 and up to 1.87 MaBP (at the base of the Olduvai subchron) values of I_c showed greater fluctuations. I_c decreased between 1.87 and 1.5 MaBP and showed less variation. However, values were higher in this upper part of the sequence than in the upper Gilbert to lower Matuyama.

The smectite indices showed less variation than those for illite. This is particularly true of $S_{v/p}$. The values of this index suggest that the smectite is highly crystalline throughout the sequence. However, Biscaye (1965) notes that the magnitude of $S_{v/p}$ depends not only on crystallinity, but also on the relative abundance of smectite. The $S_{l/h}$ index shows slightly more variability. Values fluctuated more in the Gauss-Gilbert chron than in the rest of the sequence. The values suggest that there was a slight increase in the smectite content of the 17Å complex after the Gauss-Gilbert, about 2.48 MaBP.

Because the solid geology of Kashmir is complex and spatially varied, it was impractical to sample all of the lithologies present. Instead, X-ray diffraction analyses of several of the most common rock-types in Kashmir, together with published descriptions, were used for the following discussion on the sources of primary minerals.

The most widespread mapped rock in Kashmir is Panjal Trap. X-ray diffraction analysis of trap samples from both the Himalayan and Pir Panjal flanks revealed that this lithology consists largely of plagioclase, quartz, mica (?muscovite) and chlorite, together with a range of unidentified accessory minerals in small quantities. In summary, the sedimentary rocks of Kashmir are mainly sandstone and limestone/dolomite. The igneous rocks include basalt, andesite, granite and granodiorite. The metamorphic rocks include gneiss, schist, phyllite, marble, slate and quartzite. The minerals that would be found in such rock-types are quartz, feldspar, calcite, dolomite, mica and chlorite. These minerals form the primary constituents from which the detrital clays within the lower Karewa mudstones would have been derived. More detailed information on primary mineral composition is currently unavailable. To some extent, however, it is not required for the present study. If the source-area can be assumed to have remained constant throughout the history of the basin, or any changes that may have occurred identified by means other than clay mineralogy alone, the fluctuations in clay-mineral assemblages can be explained by catchment weathering regimes.

Examination of the sand grains under the SEM suggested that they were unpromising as sources of palaeoenvironmental information. It was hoped, at the start of this study, that it would be possible to use differences in the shape and surface texture of sand grains to distinguish between grains that had been transported through a glacial transport path prior to subsequent fluvial transport and deposition, and those which had not.

However, there are two reasons why this was not possible for the lower Karewa samples. Firstly, most of the grains were found to be polycrystalline quartz, reflecting their metamorphic origin. Such grains are unsuitable for surface texture analysis (Bull, 1981). Secondly, the few monocrystalline grains that were found all exhibited similar features: they were typically angular to subangular with limited edge modification. Virtually all of the grains showed conchoidal breakage. There are two main ways in which such shapes and surface textures can arise. Firstly, through the physical weathering of quartz from bedrock, giving rise to first-cycle quartz. Secondly, they may arise from active transport within or beneath a glacier. In a tectonically-active basin, experiencing rapid rates of erosion, first-cycle quartz is likely to be common. In such a situation, the recognition of grains modified by glacial action would be impossible, since the surface textures imparted by the two modes of origin are the same (eg. Krinsley and Doornkamp, 1973). Because of these problems, it was concluded that surface-texture analysis of sand grains would not be diagnostic of glacial modification of lower Karewa sand grains. Therefore, a more detailed study was not undertaken.

4.6 Discusion

From the results of the clay-mineral analyses described in 4.5, the following trends emerge as the most striking. After about 2.5 to 2.4 MaBP, there was a change in the clay mineralogy of the lower Karewa mudstones. There was an increase in the abundance of smectite and a decrease in the abundance of chlorite + kaolinite and chlorite. Although illite abundance remained quite constant throughout the sequence, it became less well ordered and had a greater expandable component after 2.5 to 2.4 MaBP. Although it is tempting to seek a climatic explanation of the clay mineral record, non-climatic factors may also be important and cannot be discounted. In particular, the fact that Kashmir has been tectonically active over the last 4 Ma cannot be dismissed.

A major change in local source area occurred in Kashmir between 2.1 and 1.7 MaBP (Burbank, 1982). At this time, there was a switch in uplift from the Himalayan to the Pir Panjal flank. This switch, identified by Burbank (1982) on the basis of clast imbrication in conglomerates and the orientation of cross-bedded sandstone units, was confirmed by clast lithological changes in the conglomerates recorded in this study (table 4). In the conglomerates derived from the Himalayan flank, limestone clasts make up a small but significant percentage. In the conglomerates derived from the Pir Panjal flank, limestone is absent. This inferred change in source area is not, however, accompanied by a marked change in clay mineralogy. The major changes in the clay-mineral assemblages occurred rather earlier, between 2.5 and 2.4 MaBP. However, more subtle changes in the clay

TABLE 4. CLAST LITHOLOGICAL DATA: LOWER KAREWA CONGLOMERATES.

SITE	PERCENTAGE				N
	Panjal Trap	Quartzite	Limestone	Other	
Hirpur I	47.4	13.0	21.2	18.4	500
Hirpur II	52.0	8.4	27.4	12.2	500
Romushi I	59.4	13.0	9.0	18.6	500
Romushi II	72.2	12.6	0.4	14.8	500
Romushi III	82.4	9.0	0.0	8.6	500

clast size used: 5-10cm

mineralogy occurred around 1.8 MaBP. There was a quite dramatic increase in the abundance of chlorite at this time. This largely explains the decrease in the joint chlorite and kaolinite contribution, since the kaolinite contribution remained more or less constant. From this evidence, it seems that the changes in source area seem to have had a marked effect only on chlorite abundance.

The second nonclimatic factor that must be evaluated is the authigenic formation of minerals. One of the underlying assumptions of this discussion is that the clay minerals in the lower Karewa mudstones are detrital and that their type and nature reflects palaeoenvironmental conditions in the source area rather than in the lower Karewa lake. As outlined in section 4.2 of this thesis, lacustrine environments are regarded as less reactive than, say, marine ones. Therefore, neoformation or transformation of minerals was probably not of significant importance. Similarly, the lacustrine deposits are probably insufficiently thick, and also too young, for significant diagenetic alteration to have taken place. Both of these assertions are supported to some degree by the sequence of minerals present in the lower Karewa. For example, an important diagenetic reaction in mudstones is the formation of mixed layer I/S from smectite with increasing burial and depth (Eberl, 1984). However, figure 11 shows that there is no systematic decrease in the index $S_{l/h}$ with depth, which would signify a decrease in the number of smectite layers in the 17Å complex. Although it can never be certain that no authigenic formation or diagenetic alteration has occurred in the lower Karewa mudstones, it does seem likely that such reactions were subordinate, and that the bulk of the lower Karewa clay fraction is detrital.

Having satisfied the assumption that the lower Karewa clays are largely detrital and independent of changes in source area that are known to have occurred in Kashmir, it is necessary to evaluate the relative importance of climatic change and tectonic uplift. Although this is a difficult task, it is made rather easier by the existence of Burbank's (1982) chronology of uplift in Kashmir.

The increase in smectite after about 2.5 MaBP could indicate increasing aridity. The relative abundance of smectite does not appear to change significantly with the major tectonic events indicated by the conglomerate beds in the lower Karewa. The presence of illite in the mudstones suggests rapid weathering of rocks with little time available for chemical alteration of rocks. Thus, high values of illite could be associated with cold periods, during which rates of chemical weathering were reduced, or with tectonic uplift, when the exposure-time of rocks to weathering would have been limited. However, illite values within the lower Karewa are quite remarkably stable. These data suggest that limited chemical weathering and rapid erosion, and/or a cold climate, have prevailed throughout lower Karewa times. At first sight, this may seem to contradict the hypothesis that the lower Karewa clays are responding to climatic change. However, the high illite values may simply be a response to the relatively rapid rates of erosion, and therefore limited chemical weathering, that would be expected in a tectonically active area of high relief such as

Kashmir. The slight increase in expandable layers, and decrease in the crystallinity, of the illite after 2.3 to 2.1 MaBP is more difficult to explain. If the hypothesis of progressive aridity is supported, the illite would be expected to become more crystalline. However, increase in the relative abundance of smectite up the sequence suggests that conditions were becoming more favourable for the formation and preservation of this mineral.This may explain the existence of more expandable layers in the illite complex up the sequence. The change to less crystalline illite might be a response to a change in the source area, since the changes in I_c are most marked after about 1.8 MaBP, when tectonic uplift switched from the Great Himalaya to the Pir Panjal.

Kaolinite declined after about 2.4 MaBP. In addition, there is some correspondance between low kaolinite values and periods of rapid uplift as shown by the conglomerate beds in the lower Karewa. Since kaolinite tends to be formed under conditions of intense leaching, low kaolinite values would be expected in cold or arid climates and during periods of rapid uplift. Chlorite also decreased, but rather later than kaolinite, at around 1.8 MaBP. Since this was a period of change in source area in Kashmir, chlorite abundance may be responding to source rather than climate. Generally, chlorite is regarded as a cold-climate mineral which forms under conditions of minimal alteration of primary minerals. However, since chlorite is also present in the Panjal Trap rock, it would also be expected to occur detritally during times of rapid tectonic uplift. For the lower Karewa, the chlorite curve is best explained by uplift. Under this explanation, the chlorite is derived primarily from Panjal Trap, rather than secondarily, due to weathering.

From the above interpretation, a climatic-tectonic model of clay-mineral assemblages emerges. During humid and possibly warm periods, high values of kaolinite and low values of smectite are found. A similar situation also occurs during periods of quiescent tectonic activity. During arid and possibly cold periods, the abundance of smectite increases and that of kaolinite decreases. During periods of active tectonism, kaolinite abundance decreases. Climatically, the clay mineral evidence suggests progressive aridification during the Late Cainozoic, after 2.4 to 2.5 MaBP. Fluctuations in the clay mineral record may relate to climatic change on a shorter timescale. However, the clays provide insufficiently high-order palaeoclimatic data to allow this question to be answered. The general climatic trend is punctuated by tectonic uplift and this is shown in the clay mineral record.

The proposed climatic sequence from clay mineral assemblages shows some correspondence with other lines of evidence from the lower Karewa outlined in chapter 3. Work on the pollen, diatoms and vertebrate remains suggests cooling and increasing aridity, at about 0.7 MaBP, although oscillations in climate were inferred for the whole of the lower Karewa sequence on the basis of the pollen evidence. Burbank and Grant's (1985) study of grain-size variations suggests possible Milankovitch forcing on a 100 ka timescale but does not show a progressive climatic deterioration during the Late Cainozoic. The work by Krishnamurthy et al. (1986) on stable carbon isotope and C:N ratios shows no correspondence with the present data. One of the main reasons for this may be that the level of the Karewa lake was responding to changes in tectonics rather than climate. Furthermore, ^{13}C and C:N ratios are probably poor indicators of lake level and reflect, instead, changes in the source of carbon.

No attempt has been made to estimate the age of the onset of glaciation in Kashmir. Clearly, the Pir Panjal Range is young (Burbank, 1982) and on the basis of inferred rates of uplift, was of insufficient altitude to have been glaciated during lower Karewa times. However, the Great Himalaya was uplifted much earlier and so may have been glaciated during lower Karewa times. However, there is no positive evidence of glaciation in the lower Karewa sediments. Unfortunately, it is very difficult to separate first-cycle quartz grains from those of glacial origin using the surface texture methods attempted in this study. Furthermore, this is made even more difficult by the dominance of metamorphic polycrystalline quartz grains within the lower Karewa sand. However, the absence of direct evidence in the lower Karewa for glaciation does not confirm that the Himalayan Range was not glaciated during the Late Cainozoic. From the clay mineral record, it seems that the climate started to deteriorate around 2.5 to 2.4 MaBP. Furthermore, other mountain ranges at similar latitudes were glaciated during the Late Cainozoic. However, this question is discussed further in chapter 7, when data from outside Kashmir are considered.

CHAPTER 5. EVIDENCE FOR ENVIRONMENTAL CHANGE FROM THE UPPER KAREWA FORMATION.

5.1 Introduction.

The work summarized in chapter 3 illustrates the palaeoenvironmental potential of the upper Karewa formation. However, there are many problems remaining. Firstly, the overall stratigraphic and spatial distribution of the three major facies; the conglomerate, the lacustrine beds and the loess, is not well established. Secondly, the depositional environments of the upper Karewa have mainly been inferred on the basis of field evidence from isolated sections. No attempts have been made to back up these inferences with laboratory analyses. Thirdly, there exists considerable uncertainty over the palaeoenvironment of Kashmir during the upper Karewa times. Fourthly, the chronology of the upper Karewa is poorly known. These four problems are addressed in the present chapter.

The approach used in this study combines field and laboratory analysis. Initially, field reconnaissance and mapping were undertaken in order to reveal the limits to the upper Karewa lacustrine deposits and conglomerate and to determine the stratigraphical and spatial relationship between the three major facies. Once the reconnaissance and mapping had been completed, sections were chosen for more detailed analysis. The sections were located so as to cover the basin as widely as possible and include both marginal and central portions of the upper Karewa lake. Particular attention was paid to those parts of the basin where the lacustrine beds and the conglomerate are superposed. The sections themselves were located in quarries, gullies and river cuts. In all cases the slumped sediment was removed prior to description and sampling. Field description involved dividing the section into units based on lithology, particle-size and sedimentary structure. The moist colour of each unit was determined using revised Japanese Munsell soil-colour charts. Calcium-carbonate content of the sediments was determined in the field using dilute hydrochloric acid (Hodgson 1976). Samples of clasts were taken from upper Karewa conglomerates. At each sampled site, 10 samples, each of 50 clasts, were taken. Each clast was then assigned to a lithological group.Where mollusc horizons were found, bulk samples were collected for faunal analysis. Following field description, samples were taken every 25 cm up the section, with each sample consisting of 5 cm stratigraphic thickness. Subsequent laboratory analyses included calcium carbonate content (using a manometric calcimeter and dilute hydrochloric acid calibrated with pure calcium carbonate), organic matter content (by loss on ignition), particle-size analysis (by dry sieving and pipette or sedigraph analysis) and ostracod content. Selected undisturbed samples of silts and loess were taken for microfabric analysis. Samples from the upper portions of three sections of lacustrine sediment were taken for thermoluminescence (TL) dating. The upper parts of sections were sampled since previous TL dates on the overlying loess suggested that they would lie close to the maximum age-range for TL dating to be reliable, which is about 120 kaBP (H. Rendell, personal communication). Samples were also checked for the presence of pollen and diatoms, although these fossils were found to be absent.

The field reconnaissance, mapping and section descriptions form the basis for evaluating the palaeogeography of Kashmir during upper Karewa times. Depositional environments were inferred from the section descriptions and laboratory analyses. Palaeoenvironmental conclusions were drawn primarily from the analyses of the ostracod fauna. The most detailed analyses were carried out on the lacustrine beds, since the conglomerates are unfossiliferous and contain no material that could provide palaeoenvironmental information. The loess is also unfossiliferous; furthermore, it is the subject of a major research effort by numerous workers including Dr. R. Gardner (King's College, London: eg. Gardner, 1988), Dr. H. Rendell (University of Sussex) and a group from the University of Kiel, West Germany (eg. Bronger _et_ _al_., 1987) as well as PRL scientists.

5.2 Lithostratigraphy and Sedimentology of the Upper Karewa Formation.

a. General Description and Spatial Distribution.

The conglomerate, lacustrine beds and loess are the three lithological units of the upper Karewa that are generally mappable. However, the conglomerate is absent from the centre of the basin, leading to stratigraphical uncertainties in this area.

The upper Karewa conglomerates are exposed in all of the major river valleys draining the Pir Panjal (figure 16, plate 7). They thin

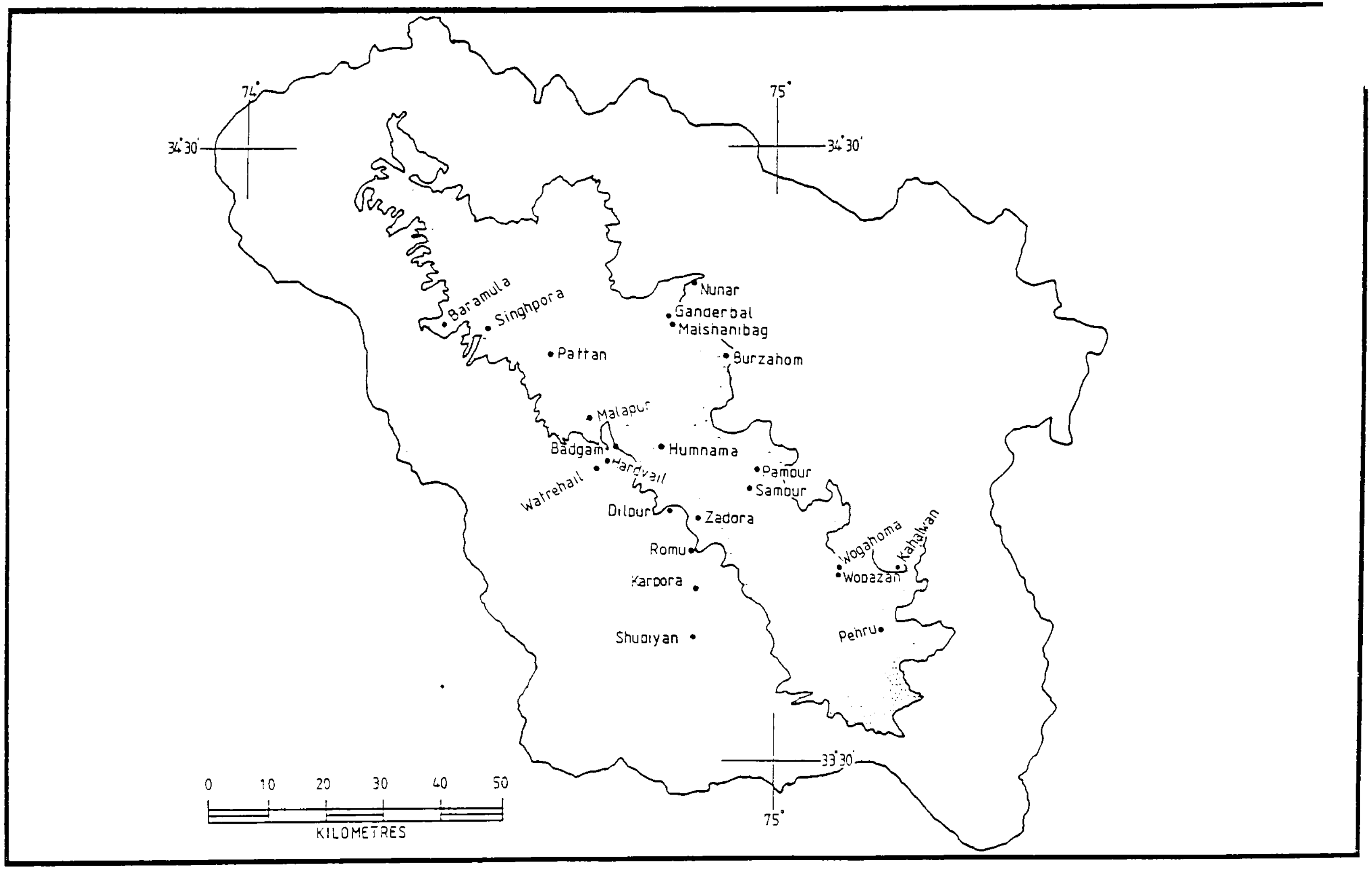


Figure 16
Location of main upper Karewa sites, and the inferred outline of the upper Karewa lake.

markedly towards the basin centre. The thickest exposures occur near Shupiyan, in the Rembiara Valley. Here, sections in the conglomerate exceed 100 m in thickness. In the basin centre, much of the sediment has been removed and the exact point at which the conglomerate pinches out cannot be seen. Sections in the mesa at Badgam, Zadora and Neu show more distal portions of the conglomerate. In the Badgam area, the upper Karewa conglomerate thins from about 4 m at the measured section to about 1 m at a location only 0.5 km away. At Dilpur, the conglomerate is about 8 m thick whereas at Qasba Nagum, 1 km to the east, it has thinned to only 1 m. In the Neu - Zadora area, the conglomerate is only 1 m thick.

The aforementioned sites are all located close to the north-easterly edge of the mesa that abuts the Pir Panjal flank. Between here and the Himalayan-flank mesa, a massive amount of sediment has been removed, although an outlier of upper Karewa sediments, upon which Srinagar airport is situated, occurs in the central portion of the basin. In the airport mesa, no beds of conglomerate were seen. Although detailed description of sections was impossible due to high security surrounding the airport environs, the absence of a conglomerate in this area is confirmed by Bhatt (1982a). These observations suggest that the upper Karewa conglomerate pinches out a short way to the east of the present position of the mesa that abuts the Pir Panjal flank.

In the upper reaches of the Pir Panjal Valleys, the upper Karewa conglomerate lies unconformably on fine-grained lower Karewa sediments. These beds are typically tilted and folded, and have been severely eroded either prior to, or during, the deposition of the upper Karewa conglomerate. Burbank (1982) has argued that the age of the uppermost preserved lower Karewa sediments is simply a function of the degree of uplift and subsequent erosion. Thus, in proximal locations, close to the Pir Panjal Range, uplift would have been more intense and the uppermost preserved beds older than in distal areas, where less post-depositional uplift would have occurred. The conglomerate beds themselves generally show no post-depositional alteration. However, in Badgam quarry 6, the conglomerate and underlying fine-grained sediments show numerous, small-scale, normal and reverse faults.

The occurrence of conglomerates on the Himalayan flank is much more limited. In the mouth of the Sind Valley, large conglomerate terraces occur on the southern side. These terraces extend upvalley to the confluence of the Sind with Wangat Nala, and exceed 30 m in thickness. Just beyond the mouth of the Sind Valley, between Malshahibagh and Ganderbal, a cemented conglomerate was found, interdigitating with upper Karewa lacustrine beds. Because there in no physical continuity between the cemented conglomerate and the terrace conglomerates in the Sind Valley, it is difficult to determine whether the two are stratigraphically equivalent. Agrawal (1985) has suggested that the cemented conglomerate at Malshahibagh may be the distal remnant of the upper Karewa conglomerates found on the Pir Panjal flank. However, this seems unlikely since the exposure at Malshahibagh is thicker (about 4 m) than the most distal portions of the Pir Panjal conglomerate (1 m). In this study, two hypotheses for the origin of the Sind Valley terrace conglomerates were proposed. Firstly, that the conglomerates exposed at Malshahibagh are distal remnants of the Sind Valley terraces. Secondly, that the two conglomerates are from different sources. Similar conglomerates are found in the mouth of the Liddar Valley. The terraces in both valleys are invariably capped by up to 10 m of aeolian loess.

Upper Karewa lacustrine beds form a major component of the mesas in the Kashmir basin, although they are absent from the Pir Panjal flank. Bhatt's (1982a) map of the distribution of the upper Karewa lacustrine beds suggests that the lake was areally restricted during upper Karewa times. In this study, mapping of the limits of upper Karewa lacustrine beds in the field revealed that they are not found above 1680 m a.s.l. The lowest part of the Kashmir Valley floor lies at about 1600 m a.s.l. The highest point in the Baramulla Gorge, through which the upper Karewa lake would have drained, is about 1500 m a.s.l. At sites such as Qasba Nagum, Zadora, Neu and Badgam where the conglomerate is thin, there is evidence of interdigitation between the conglomerate and the lacustrine beds. This indicates that the two facies are contemporaneous. Thus, the conglomerates and fine-grained lacustrine sediments represent braided streams feeding into a lake. Interdigitation would have resulted from changes in lake level and shifts in the axes of conglomerate deposition. In order to reconstruct the outline of the upper Karewa lake, the 1680 m contour was traced on U.S. Defence Mapping Agency 1:250 000 scale topographic maps (sheets NI43 6 to 7 and 10 to 11) (figure 16). The reconstructed outline conforms well with that drawn by Bhatt (1982a) on the basis of field evidence alone.

Most exposures show that the upper Karewa lacustrine sediments are horizontally bedded with no signs of faulting and folding. However, in quarry exposures at Humhama, towards the inferred centre of the lake, the beds show quite extensive folding and small-scale faulting. In the mesa around Wogahoma on the Himalayn flank, there is also evidence of post-depositional tectonic movement. At the base of the Wogahoma section, an exposure of blue-grey silt-clay is found. This type of lithology is not found anywhere else on the Himalayan flank and is regarded as lower Karewa lacustrine sediment (Singh, 1982). Although lithological similarity alone does not confirm stratigraphical equivalence, the occurrence of the blue-grey silt-clay facies, if it is of lower Karewa age, does suggest that older strata may heve been uplifted by localized tectonic movements in this area. This idea was confirmed in this study by simple levelling in the field with a clinometer. This showed that the top of the Wogahoma mesa was perceptibly higher than surrounding mesas, in which the blue-grey facies was not seen. In the light of this evidence, it would seem quite reasonable to accept Singh's (1982) idea of uplift.

There are problems of terminology in the basin centre, beyond the extent of the upper Karewa conglomerate. The basis for dividing the sequence formally into upper and lower formations has been the angular unconformity between the uppermost part of the lower Karewa sediments and the base of the upper Karewa conglomerate. However, where the conglomerate is absent, there is no basis for this subdivision since there is no distinction between the uppermost lower Karewa beds and lowermost upper Karewa beds. For this reason, the terms upper and lower Karewa should be retained in an informal sense only.

Loess is the uppermost unit of the basin sediments. In certain areas of the Pir Panjal flank, loess overlies fine-grained lower Karewa sediments. This occurs where lower Karewa sediments have been uplifted and eroded in such a way as to leave outliers, around which the deposition of the upper Karewa conglomerate took place. Elsewhere above 1680 m a.s.l., loess overlies the upper Karewa conglomerate. Below 1680 m a.s.l., loess overlies the upper Karewa lacustrine beds. In most exposures examined in this study, the boundary with the upper Karewa lacustrine sediments was found to be conformable, suggesting a gradual change from lacustrine to aeolian deposition. In some sections, however, extensive gullying of the lacustrine sediments clearly occurred prior to the onset of loess deposition. The quarry section near Sambur village shows an excellent example of this. In this section, a loess-filled gully about 20 m deep cuts into the underlying lacustrine strata. The sequence of palaeosols in the gully fill is similar to that in the adjacent loess that caps undissected upper Karewa beds. This observation tends to favour the gullying having occurred prior to the onset of loess deposition rather than after it.

The loess is of variable thickness. In the Burzahom area, for example, it is only about 4 m thick. Three palaeosols are seen in the profile. At Karpora, the loess is over 20 m thick, with nine palaeosols (Gardner, 1988).

b. Field Descriptions of Individual Sections.

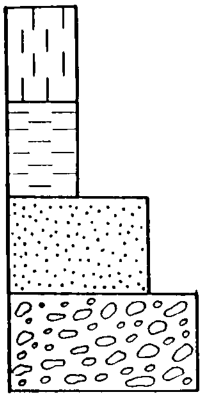
The sections described in this section are located on figure 16. A total of 25 sections was measured. 9 of these were measured and sampled in great detail. These were the sections at Badgam, Burzahom, Karpora, Sambur, Sambur Village, Pampur, Pattan, Pehru and Wogahoma. They were chosen since they were potentially the longest exposures in each area. The remaining 16 sections were studied in less detail, in order to support the conclusions drawn from the detailed sections and to clarify the stratigraphy in certain areas. For each of the detailed sections, graphic log diagrams were drawn, together with curves of calcium-carbonate content, where this was determined. A key to the logs is shown in figure 17. The precise location of each section is listed in appendix 3. The field descriptions are combined with the sedimentological data to provide an interpretation of each measured section which appears later in this chapter, in section 5.2d.

<u>Pattan</u>

One section was measured in the Pattan area (figures 18 and 19, plate 3).The section was located close to the head of a gully about 3 km to the south-west of Pattan village. At this site, about 8 to 10 m of loess overlies conformably the upper Karewa lacustrine beds (R. Gardner, personal communication). A total of 29 m of loess and upper Karewa lacustrine beds was exposed by digging into the side of the gully. The gully extends a further 10 to 15 m below the base of the

KEY TO LITHOLOGICAL SECTIONS

LITHOLOGIES



Aeolian loess and pedogenic features

Lacustrine and fluvial clay and silt

Sand

Gravel

SEDIMENTARY STRUCTURES AND BED CONTACTS

Massive

Parallel bedding

Parallel lamination

Ripple bedding

Ripple lamination

Parallel & ripple bedding

Parallel & ripple lamination

Calcareous concretions

Lenticular bedding

Flaser bedding

Cross-bedding: trough

Cross-bedding: planar

Sharp to clear

Gradual to diffuse

Straight

Wavy/irregular

BIOGENIC FEATURES

Location of measured section

Shells

Root casts

Vertebrate fossils

Figure 17

Key to graphic logs.

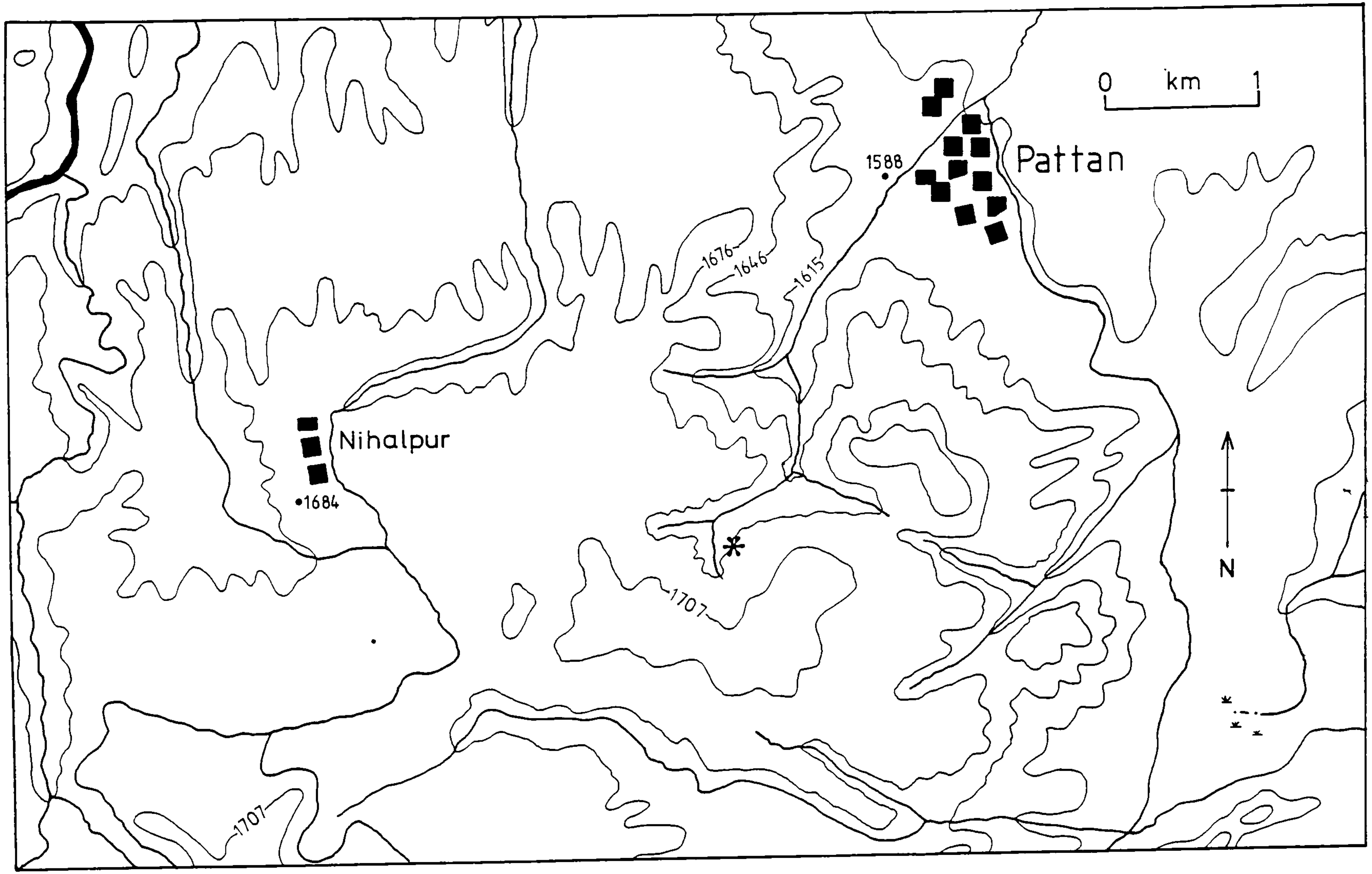


Figure 18
Location map of the Pattan section.

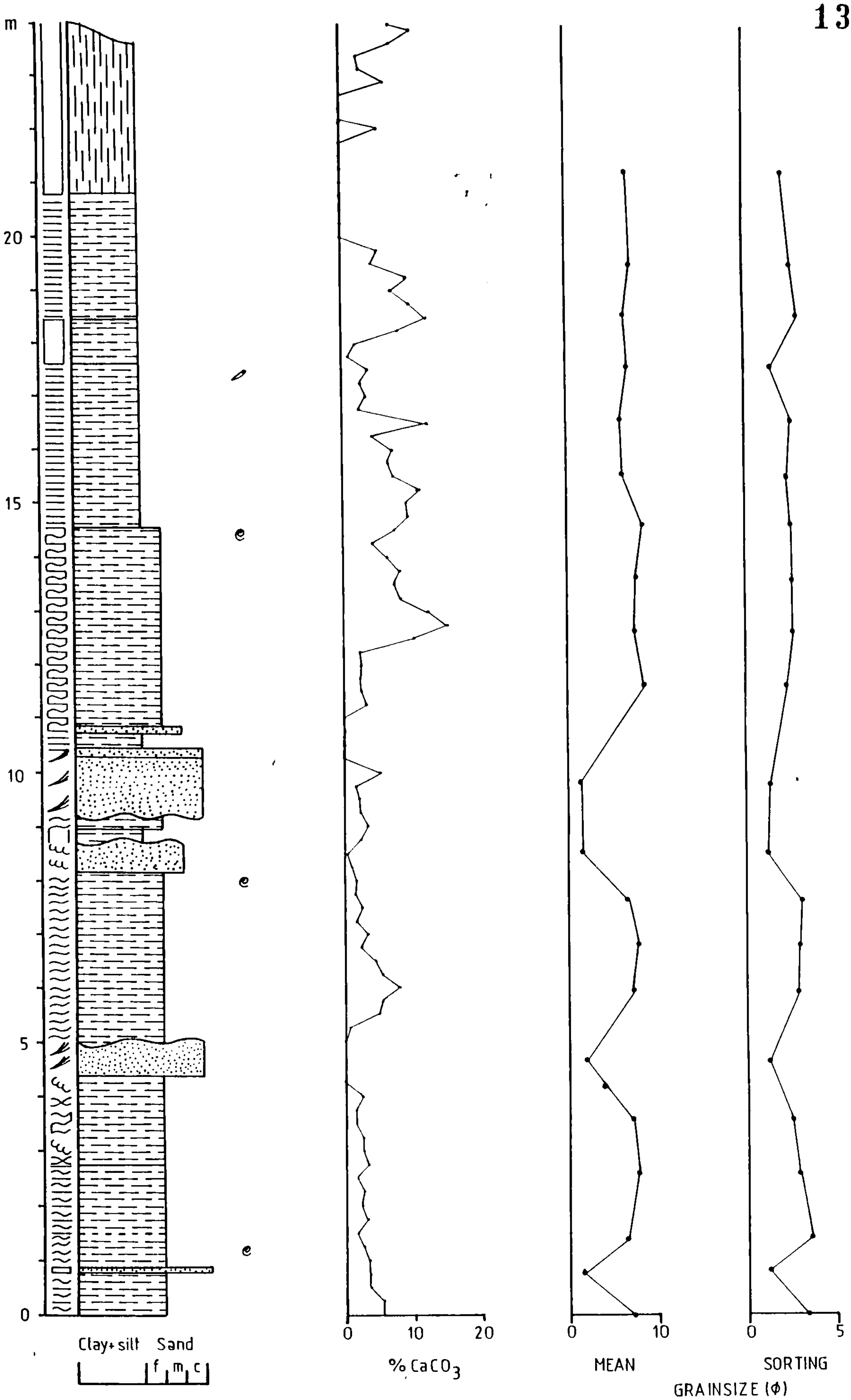


Figure 19
Graphic log diagram and calcium-carbonate curve of the Pattan section.

measured section, but the accumulation of slumped and washed material in the base of the gully made further excavation impossible. No conglomerate bed was found, despite an extensive search in the gully. The base of the excavated section lay 29 m below the top of the mesa. The section included 21 m of lacustrine beds and 9.5 m of loess. The Pattan section is one of the longest of the sections measured in detail. On the basis of field description the section was divided into units which were numbered from the bottom upwards.

The Pattan section consists of fine-grained sediments punctuated by minor units of sand. The lowermost unit in the section consists of alternate laminae of silty clay and very fine sand. No macrofossils were present in this unit, but it was subsequently found to contain abundant microscopic plant remains and ostracod valves, in common with most of the samples from the Pattan section. In its lower half, the laminae are parallel and straight, whereas in the upper half, ripple laminae are found. The silt and fine sand is typically light yellowish brown (2.5Y 6/4) but concentrations of iron oxide along the laminae margins give a characteristic reddish yellow (7.5YR 6/8) colour.

The second unit in this section is a thin bed of massive, unfossiliferous medium to coarse sand. It also shows iron oxidation along the upper and lower boundaries, which are both sharp and smooth. The next three units in the section are silty clay and very fine sand. Unit 3 is very similar to the upper part of unit 1, but contains abundant snails, identified as <u>Valvata</u> <u>piscinalis</u> (Müller) (D.S. Brown, personal communication), which is a widespread, palearctic littoral species. Iron oxide staining occurs commonly along the boundaries of the laminae, which become thinner towards the top of the unit. Unit 4 is unfossiliferous; it consists of alternating ripple and parallel laminations of silt and fine sand, but is otherwise similar to unit 3. Unit 5 is more complex. Sedimentologically, it is similar to units 1, 3 and 4, consisting of silty clay and very fine sand. However, it consists of a mixture of wavy, lenticular and flaser bedding and lamination. Iron oxidation is concentrated along the boundaries of the larger-scale bedding features, but is not found along the laminae. Some of the larger fine-sand bodies in this unit have load structures present on their upper margins.

Unit 6 is a sand bed. It has a sharp and straight boundary with the unit below and a sharp and wavy boundary with the unit above. Despite the sharp boundary between units 6 and 7 visible in the field, the upper 5 cm of the sand bed contain an increasing proportion of silt and clay. The sand shows distinct planar cross-bedding, with iron oxide staining along the individual bed surfaces.

Unit 7 marks a return to fine-grained sediments. The unit has irregular ripple lamination throughout, although in the top metre, the laminae become thin beds. As with the other fine-grained units, iron oxidation is concentrated along the boundaries of laminae and beds.

Units 8 to 14 comprise an alternation of sand and fine-grained sediments.Unit 8 consists of isolated silt flasers within troughs of medium-sand ripples. It has a wavy and sharp upper boundary with unit

9, which is a thin, massive silty clay. Unit 10 is similar to unit 9, but shows ripple laminations with alternating silty clay and fine to very fine sand. Unit 10 has a sharp and wavy, erosive boundary with unit 12, which is a coarse, planar cross-bedded sand. Individual cross-beds within the sand vary considerably in their dominant particle-size from fine to coarse sand, although within each bed the sediment is well sorted. In the top 5 cm, the sand is weakly indurated (unit 12). It shows wavy bedding and intense, yellowish brown (5YR 5/8) colouration due to iron oxide staining. This sand unit is separated from the upper sand in the section (unit 14) by a thin, parallel-laminated silty clay, which has iron oxide staining along the laminae boundaries. All of the units 8 to 14 contain no macrofossils, but are rich in ostracods.

The remaining 10 metres of upper Karewa sediments in the Pattan section are fine grained. Field observation suggested fining upwards in the sequence, but this was not supported by subsequent particle-size analysis. Unit 15 consists of parallel and ripple-bedding, with alternations of silty clay and fine to very fine sand. In common with units lower in the section, distinct iron oxide staining follows the individual bed surfaces. Sparse shell fragments were found in hand specimens, but they were too fragmentary to identify. Unit 16 consists of very faint parallel laminations distinguishable in the field on slight colour variations. Iron Oxide staining was much less apparent than in unit 15. Unit 17 is similar to unit 16 in texture, but is massive. Medium to fine root or shoot casts are quite abundant in this unit and the concentration of iron oxide around these remains produces a strong mottling effect. Unit 18 is marked by a return to very faint parallel lamination. Mottling is much less intense than in the underlying unit. The boundary between unit 18 and the overlying loess is clear and straight. In the Pattan section, 9.5 m of loess overlies the lacustrine sequence. A triplet of fossil Ah horizons occurs in the lowermost 3 metres of the loess. This is overlain by 3.5 metres of weathered loess, followed by another pair of palaeosols between 6 and 7 m. About 2 m of loess separates these soils from the modern soil at the top of the section (R. Gardner, personal communication).

Singhpora Road

This is a short, 4 metre-long exposure on the extreme north-west fringe of the inferred upper Karewa lake at about 1680 m a.s.l. (figure 16). The section was located on the road from the national highway linking Srinagar and Baramulla, and the village of Singhpora. The section consists of laminated silty clay overlying planar cross-bedded medium to coarse sand. It is otherwise unremarkable, except for the occurrence of several unidentifiable vertebrate bones within the sand unit.

Burzahom

One section was measured in the Burzahom area (plate 5). Burzahom village is located in an embayment in the former upper Karewa lake, close to the Himalayan front (figure 16 and 19). The section was dug in a disused quarry, beneath a Neolithic dwelling and stone circle. The archaeological sequence at Burzahom has been studied in detail (eg. de Terra and Paterson, 1939), but the underlying upper Karewa

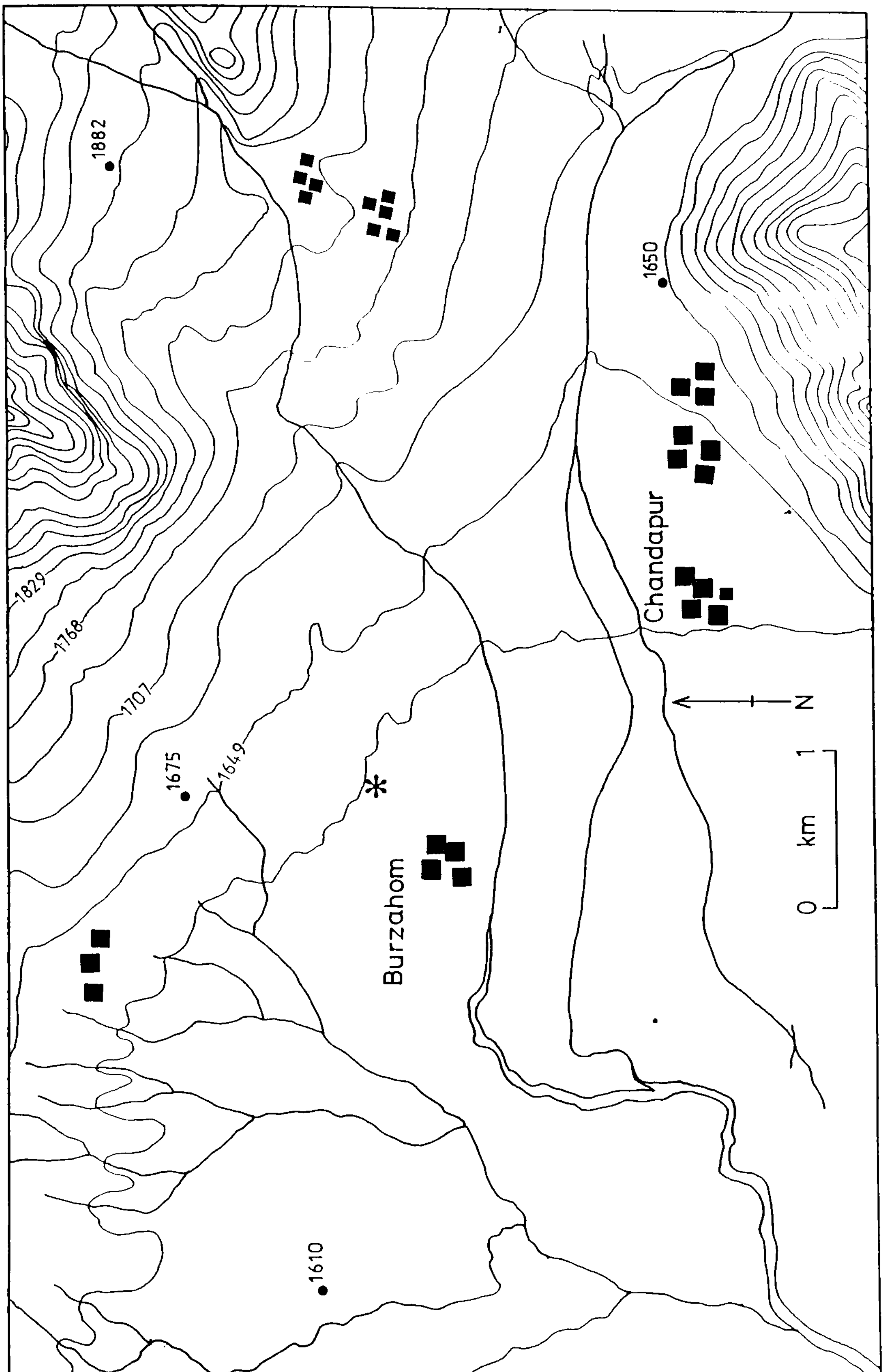


Figure 20
Location map of the Burzahom section.

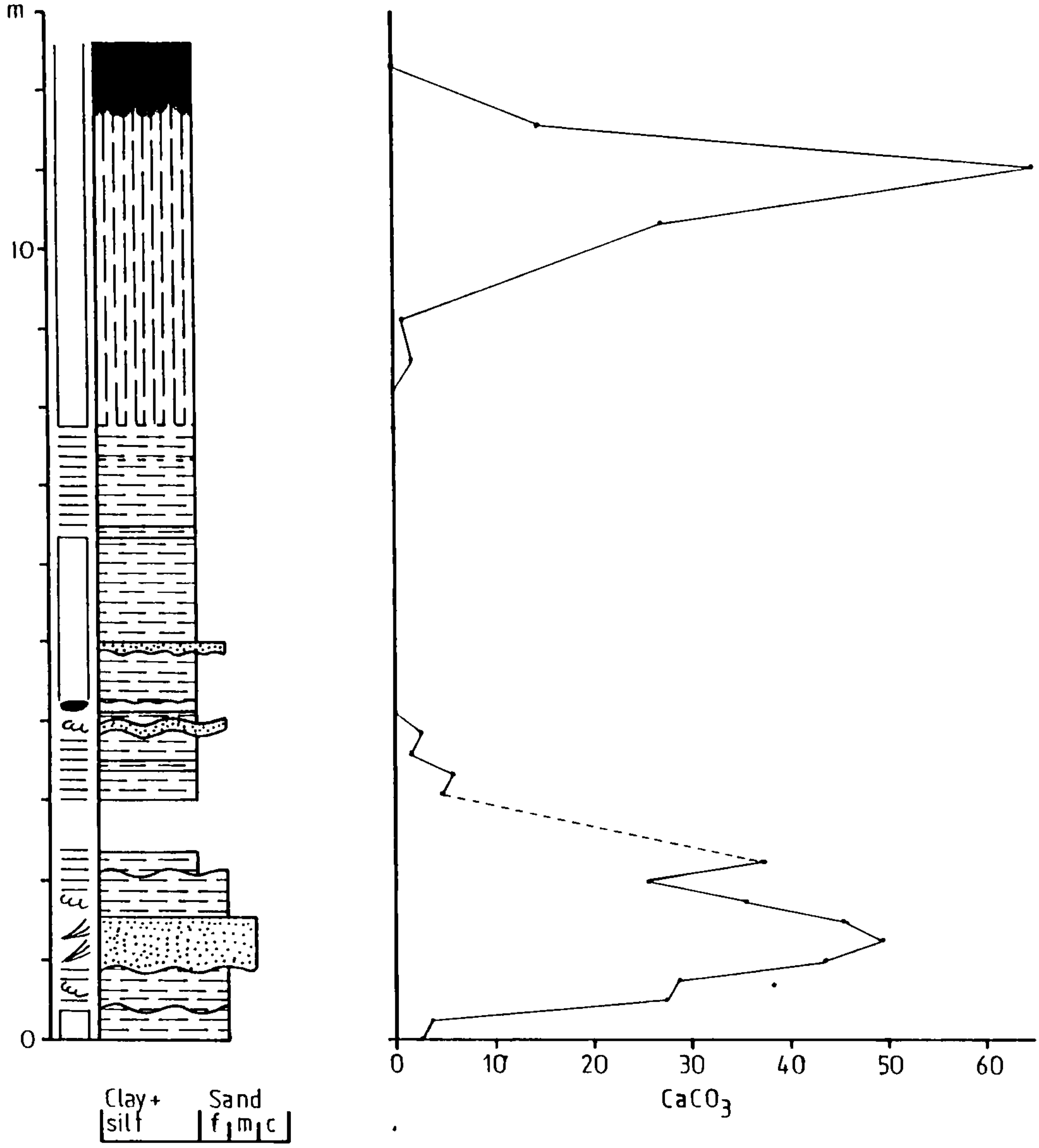


Figure 21

Graphic log diagram and calcium-carbonate curve of the Burzahom section.

sediments have received scant attention. The section at Burzahom consists of two parts: one above and one below a small terrace left by quarrying, which is now cultivated. The two parts of the measured section are separated by a hiatus of about 0.6 m. (figure 20).

The lowermost unit at Burzahom (1a) consists of massive sandy silt. Traces of plant rootlets or shootlets were found, and quite intense iron oxide staining is associated with vertical fissures in the sediment caused by the rootlet or shootlet remains. The upper boundary of this unit is sharp and slightly wavy. It is overlain by a silty sand (unit 2a) which has fine parallel laminations and disturbed flaser bedding. Diffuse brownish yellow (10YR 6/8) staining was found along the laminae. Unit 3a overlies unit 2a with a sharp and wavy boundary. Unit 3a is a well sorted, uncemented, planar cross bedded sand, with occasional mud clasts. The cross beds are picked out by strong brownish yellow (10YR 6/8) iron- and possibly manganese-oxide staining. An abrupt and slightly wavy boundary separates this unit from unit 4a, which consists of silt to silty sand. It is characterized by flaser bedding in the lower half and fine parallel lamination in the upper half. The silts of this unit are light grey (5Y 7/2) but show mottling, due to strong brown (7.5YR 4/6) to yellowish brown (10YR 5/8) iron-oxide staining. The boundary between units 4a and 5a is sharp and wavy to irregular. Unit 5a is finer-grained than the underlying units. It consists of clay-silt with faint parallel lamination and vertical root casts. The laminae and root casts are emphasized by iron oxidation. Occasionally, calcareous concretions were found.

Unit 1b is the lowermost unit above the terrace. It is very similar to unit 5a suggesting that there is no major facies change between the two units. It consists of faint, thin parallel-laminated clay silt with abundant root or twig fossils. Calcareous concretions are quite common in the lower part. Sparse snail shells were found in this unit. Most individuals were Valvata piscinalis, although a species of Lymnaea, probably L. auricularia, (D.S. Brown, personal communication) was also found. Units 2b and 3b are similar to unit 1b. Unit 2b contains moderately abundant mollusc shells which were exclusively V. piscinalis. Tabular calcareous concretions and fossil root or twig remains were also found in this unit. Unit 3b was found to be unfossiliferous in hand specimen. Unit 4b consists of of a thin bed of fine sand, with sharp and wavy upper and lower boundaries. The lower part of the unit is coarser-grained with disturbed parallel lamination and flaser bedding. The upper part is more clay-rich and has thin to thick parallel lamination. No fossils were found in hand specimen.

Units 5b to 8b are all clay silts, with occasional laminae of very fine sand. Most of the sediments in these units are massive, although some very fine colour banding was apparent to the top of unit 8b. Sparse fragments of ?V. piscinalis were found in unit 6b and near the base of 8b. Unit 7b is a horizon of very strongly cemented, calcareous concretions. The cemented layer is laterally discontinuous across the section and has sharp, irregular and broken boundaries with the over- and under-lying units. Unit 9b is a thin, massive bed of silt and very fine sand. Units 10b to 13b are all clayey silts. Unit 10b is massive and is overlain by two pale bands, units 11b and 12b. Unit 13b is the uppermost part of the lacustrine beds in the Burzahom section. It consists of faint parallel laminations stained by iron oxide along the laminae surfaces. Mottling is pronounced in the lower part of the unit but becomes less abundant towards the top. In the upper part of the unit, there is evidence of pedogenesis. This includes leached calcium carbonate along fissures, weak ped development and clay coats on the ped surfaces. These features become more pronounced in the upper 0.5 m of the unit. This unit is thus regarded as the Cox horizon of the first soil developed in the overlying loess.

The loess was described in more detail in this study at Burzahom than elsewhere. Over 4.5 m of sediments overlie unit 13b of the upper Karewa lacustrine beds. This thickness includes aeolian loess and neolithic spoil. Four palaeosols can be seen quite clearly in the aeolian loess. The lowest soil consists of a fossil Ah horizon, a textural B horizon and an oxidised horizon showing faint signs of pedogenesis. The last mentioned of these is unit 13b of the upper Karewa lacustrine beds, described above. The Ah horizon is darker (dark brown-10YR 4/3) than the underlying Bt horizon (yellowish brown-10YR 5/6), due to the organic coats on the peds within the former. Argillans are also common, especially in the Bt horizon, where they are quite well developed. The second and third soils from the base of the loess are similar in character to the first. The uppermost soil, however, lacks an Ah horizon and the Bt horizon is overlain directly by Neolithic spoil. The archaeological horizon contains charcoal, pottery, reworked loess and, presumably, the Ah horizon of the uppermost soil.

Badgam

Several exposures were examined in the Badgam area. This area includes an unnamed river valley a little to the west of Badgam village and a series of quarry exposures found in the mesa to the east of Badgam extending as far as the village of Humhama (figure 16). The section in the unnamed river valley was measured in detail is named here the 'Badgam section.' This section is of particular importance since it records the interplay between the upper Karewa conglomerate, the lacustrine beds and the loess (figure 22 and 23). The quarry sections are also interesting since they expose sediments from towards the centre of the basin. In addition, they also show folding and faulting.

The Badgam section is exposed in a river cliff on the eastern side of the unnamed river. Further digging was necessary in the base of the cliff in order to expose the lower 10 metres. The lowermost unit in this section consists of dark greenish grey (10G 4/1 to 5G 3/1) clayey silt. The unit has constant lithology throughout. Very thin parallel beds were seen, picked out by olive brown (2.5Y 4/6) iron staining. Unit 1 is overlain by a horizon of strongly cemented tabular calcareous concretions (unit 2) which run continuously across the section. Lithologically, this unit has every appearance of unit 1, except for the cementation which follows the parallel bedding of the unit. Unit 3 is slightly coarser in texture than the underlying units.

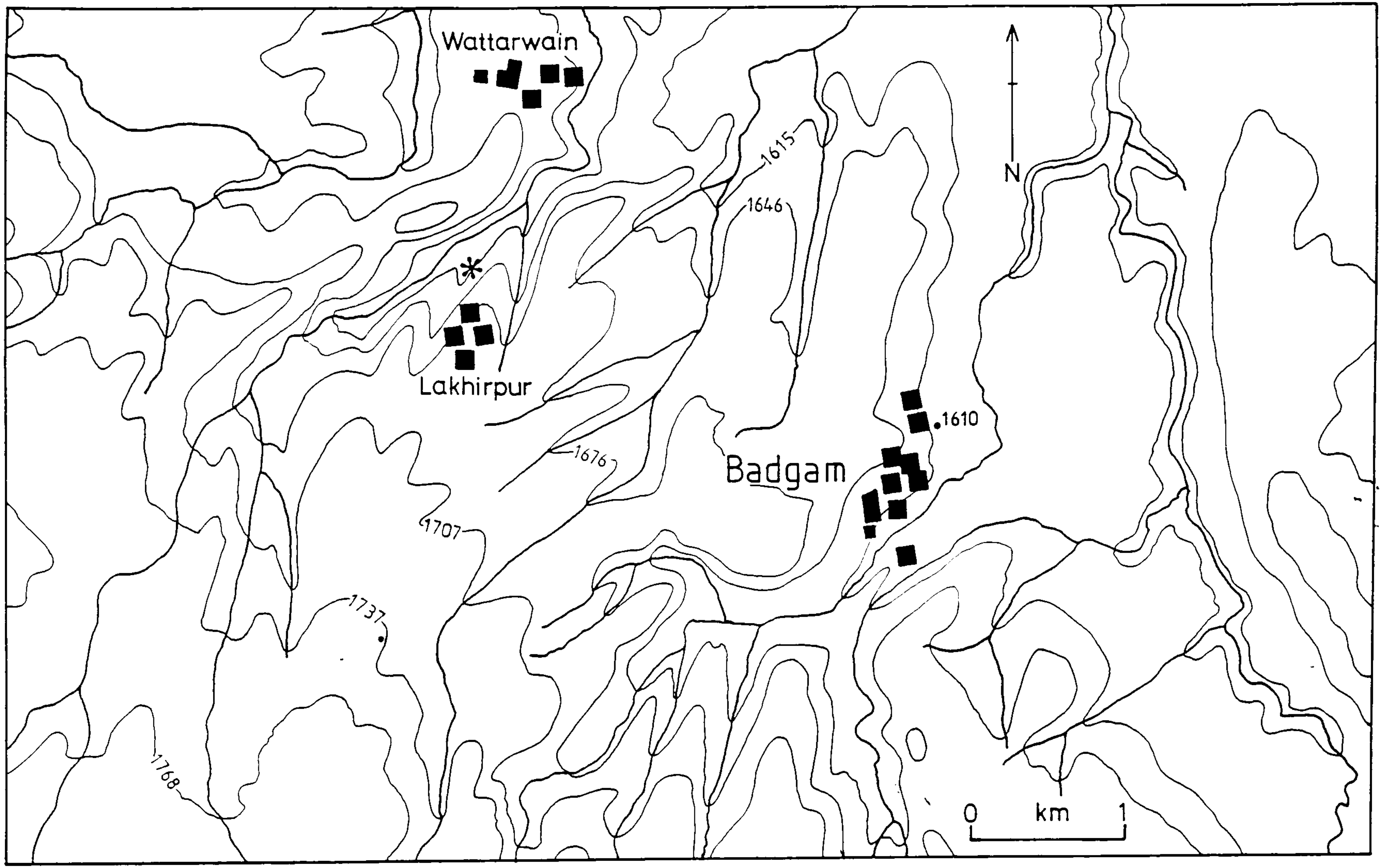


Figure 22
Location map of the Badgam section.

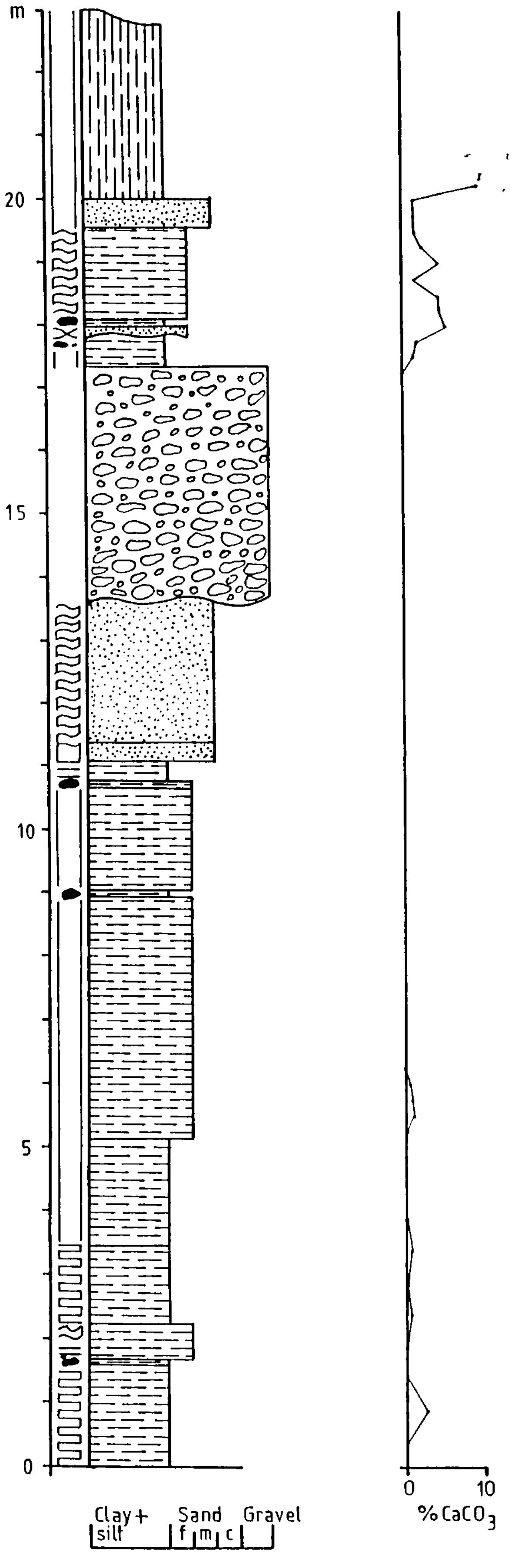


Figure 23

Graphic log diagram and calcium-carbonate curve of the Badgam section.

It consists of very thinly bedded, greenish grey (5G 5/1) silty clay and very fine sand. The unit shows ripple and parallel bedding and lamination. Iron oxide staining follows the laminae boundaries in the finer-grained layers. The boundaries between units 1, 2, 3 and 4 are all sharp and straight. Unit 4 is a greenish grey (5G 5/1) clayey silt with individual very thin to thick beds picked out by iron oxide staining. There is a gradual change from very thin bedding in the lowermost part of the unit to thick bedding at the top. This eventually grades into the massive unit 5 above, with a diffuse boundary.

Between units 5 and 6 there is a diffuse boundary. This is shown in the field by a colour change from greenish grey (5G 5/1) to olive (5Y 5/4) and a change in texture from clayey silt to silty clay and fine sand. Although mainly massive, unit 6 does have occasional thin beds of sand. Unit 7 is a horizon of very strongly cemented, finely laminated calcareous clayey silt. Unit 8 above is a massive, olive (5Y 5/4) silt with some yellowish brown (10YR 5/8) mottles of iron oxide. An irregular, but mainly continuous horizon of strongly cemented, nodular calcium carbonate concretions occurs in the middle of this unit. Unit 9 consists of a horizon of tabular, calcareous concretions which are very strongly cemented. This is overlain by unit 10, a noncalcareous, parallel-laminated, silty clay.

Units 11 and 12 are both well sorted, olive (5Y5/4) noncalcareous sands. Unit 11 is predominantly massive, but has some thin parallel beds which are picked out by reddish yellow (7.5 YR 6/8) iron oxide staining. Unit 11 has sharp and straight upper and lower boundaries. Unit 12 is wavy bedded medium to fine sand and silty clay. The silty clay is strongly iron stained. On the basis of field observation, no fossils were found in units 1 to 12. The top of this unit has an erosive contact with the conglomerate, unit 13, above. The conglomerate is 4 metres thick and unconsolidated. It is clast supported and has a sandy matrix. The clasts are generally 10 to 15 cm long, although locally, clasts double this length were seen. The conglomerate is crudely bedded. Some small (3 to 4 m in length) lenses of sand and fine gravel were seen within the unit. Locally, lateral and vertical fining was seen. The clasts were seen to have crude imbrication, with long axes oriented in a south-west to north-east direction. The conglomerate unit was traced both up and down the river valley from the measured section. It extended upvalley for at least 5 km as far as Watrehail (figure 16). A steady increase in the thickness of the conglomerate was seen from this point, where it was about 12 m thick, to the edge of the mesa, where it was less than 1 m thick. The olive-coloured silty clay unit found beneath the conglomerate at the measured section, was found to be absent at Watrehail, where the conglomerte overlies greenish grey clayey silt.

At the measured section, the conglomerate is overlain by a massive, olive brown (2.5Y 4/1) clayey silt. Fossil root or shoot casts were quite abundant in this unit. Bright brown (7.5YR 5/8) iron oxide is concentrated around these remains. Traces of black (10YR 1.7/1) material, probably manganese concentrations, were also found in this unit.

An abrupt and wavy boundary separates units 14 and 15. Unit 15 is a thin bed of sand bodies in clayey silt - lenticular bedding. The margins of the individual sand bodies are picked out by strong iron oxide staining. Rare, poorly preserved and unidentifiable snail shells, were found. Unit 16 is a very strongly cemented horizon of nodular calcareous concretions containing poorly preserved and unidentifable mollusc fragments. It is overlain by unit 17 which is ripple bedded clayey silt and fine to medium sand. Unit 18 is a massive, well sorted, noncalcareous sand overlain conformably by loess.

Units 14 to 18, which lie stratigraphically above the conglomerate, vary in collective thickness along the river valley. Upvalley from the measured section, they increase in thickness whereas downvalley they decrease. Over a distance of less than 0.5 km upvalley from the measured section, these units pinch out and the conglomerate is overlain directly by loess.

The six quarries examined in the Badgam area are found along the road linking Srinagar and Badgam. The road divides at Humhama, where one branch leads to the airport. Quarries 1 to 5 are in the central mesa upon which the airport is situated. Quarry 6 is in the most distal portion of the mesa which abuts the Pir Panjal front and in which the Badgam section, described above, is also located.

Quarry 1 is located at Humhama itself. The base of the exposure consists of parallel and ripple laminated silts typical of many other upper Karewa units. The units show many other 'typical' upper Karewa lacustrine bed features including iron oxide staining, and root and twig fossils. Overlying these beds is a unit of fractured, but otherwise massive, silt clay. It is variable in colour from dull yellow orange (10YR 6/3) in the lower part to dull yellowish brown (10YR 4/3) in the upper part. he individual fractured blocks are usually less than 10 cm^3 and almost invariably coated with a very thin layer(<1mm)of dark brown (10YR 3/3) to brownish black (10YR 2/3) material. The fractured mud was 2 to 3 m thick.

A similar vertical sequence of beds was found in quarries 2 to 5. However, in quarry 5, a laterally persistent sand and very fine gravel unit occurs at the base of the fractured silty clay. The laminated unit and the fractured silty clay unit are also quite severely contorted in this quarry. The surface of the fractured unit has been gullied and subsequently infilled with loess. Quarry 6 has a similar vertical sequence to quarry 5, except that the sand unit is replaced by a strongly iron-stained conglomerate less than 1 m thick. This appears to be the distal remnant of the conglomerate seen in the Badgam section. The sediments in quarry 6 show a complex range of small scale faults (plate 13).

<u>Wogahoma</u>

The village of Wogahoma is located about 4 km to the north of Bijbiara in the south-east of Kashmir basin (figure 24). The measured section was dug in the western limb of a large gully that cuts into the mesa to the north-east of Wogahoma (figure 25). 28 metres below

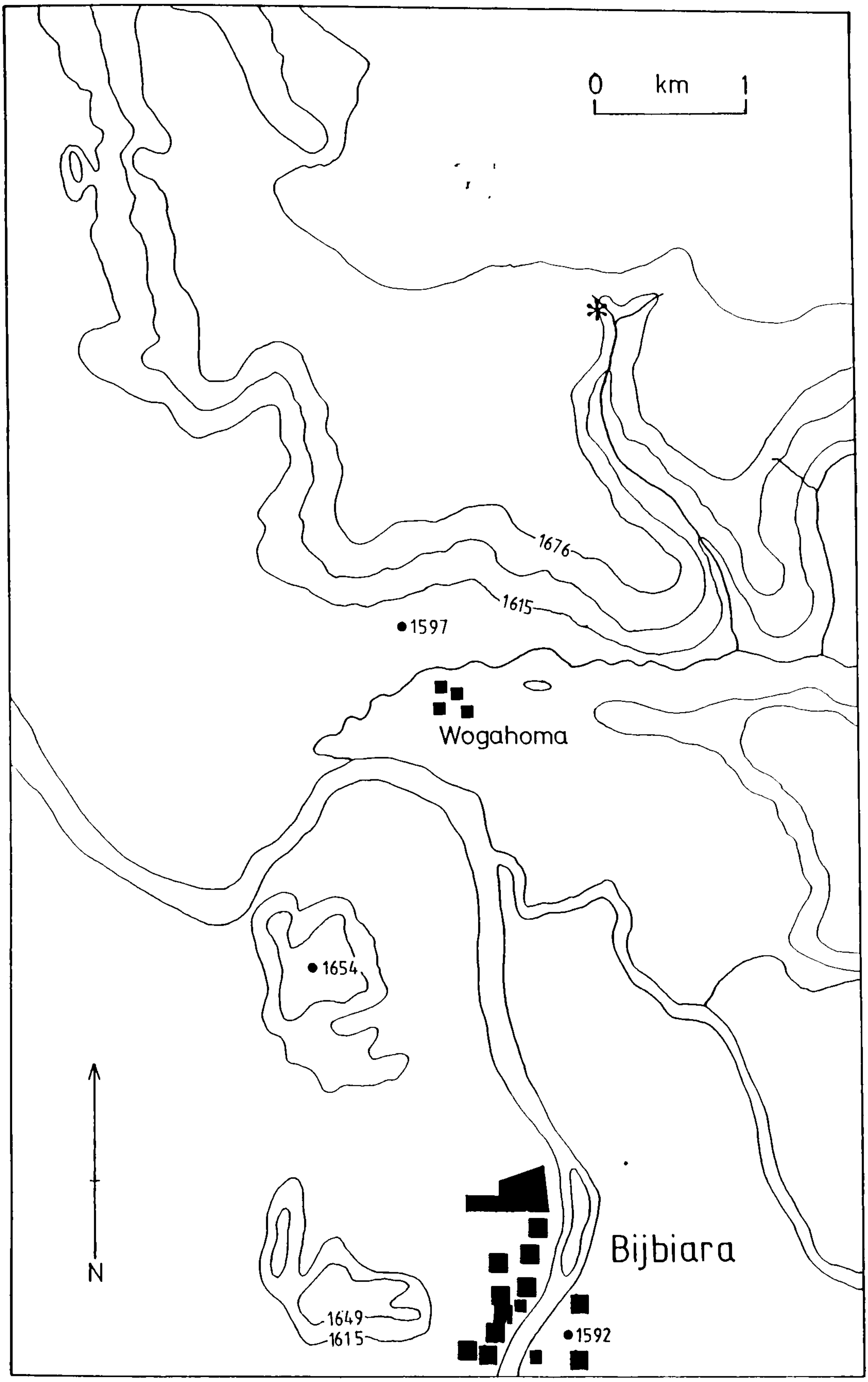


Figure 24

Location map of the Wogahoma section.

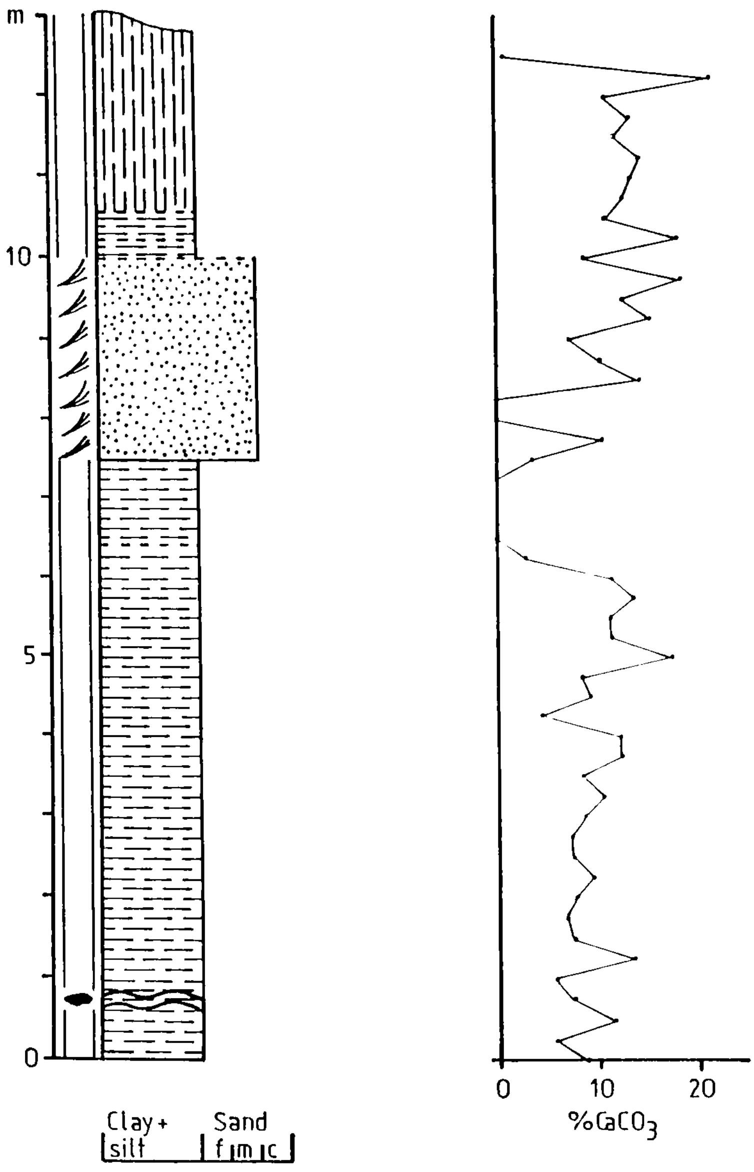


Figure 25

Graphic log diagram and calcium-carbonate curve of the Wogahoma section the base of the measured section, dark greenish grey (10G 4/1) silty-clay sediments are exposed. These are overlain by olive yellow (7.5Y 6/3) silts. The exposed sediments are about 8 m thick and are covered by slumped material, that obscures 20 m of overlying sediments.

The basal 6.5 m of sediment measured at Wogahoma are unremarkable. Unit 1 consists of a massive silty clay seen in hand specimen to be rich in mica (probably muscovite). Unit 2 is a laterally dicontinuous horizon of strongly cemented calcareous nodules overlying unit 1 with a sharp and irregular boundary. Unit 3 is similar to unit 1. Typically, units 1 to 3 are olive brown (2.5Y 4/4) to dark olive (5Y 4/3) with strong brown iron oxide staining (7.5YR 4/6), particularly in the basal part of unit 3. Unit 4 shows a progressive upward darkening in colour from dull yellow (2.5Y 6/4) at the base, through yellowish brown (2.5Y 5/4) in the centre, to brownish black (7.5YR 3/2) towards the top. There is also an increase in mottling, iron staining and abundance of organic matter from the base upwards. The upper part is also fissured.

Unit 5 overlies unit 4 with a sharp and straight boundary. It consists of fine to coarse calcareous sand with faint, low-angle cross bedding and size-sorting within the structures. Parts of the unit are very strongly cemented. The cemented horizons, which are less-easily erodible, stand out as prominent, sub-horizontal layers.

The upper 20 cm of unit 5 show an increasing admixture of fine silt together with a gradual decrease in the mean and maximum particle size of the sand. Unit 6, which overlies unit 5 with a diffuse boundary, appears to be a mixing zone. It is a massive, sandy silt, with a gradual upper boundary grading into the B horizon of the overlying loess. Apart from the fossil plant remains found in unit 3, no fossils were found in these sediments.

Pehru

Pehru is a small village 4km to the east of Anantnag, just beyond the mouth of the Liddar Valley. A 12 metre-long section was dug in the plateau edge close to the village (figure 26). In general, the Pehru section is characterized by a high proportion of sand units and a large number of highly calcareous beds (figure 27).

The basal unit in Pehru consists of massive, bright yellowish brown (2.5Y 6/6) silty clay with moderately abundant fossil roots. Bright brown (7.5YR 5/8) iron oxide is concentrated around the fossil organic remains. The top 5 cm of unit 1 are indurated with calcium carbonate, forming a prominent, strongly cemented horizon. A clear and straight boundary separates units 1 and 2. Unit 2 consists of parallel and ripple laminated silt clay. The top 7 cm of the unit are strongly cemented with calcium carbonate, the upper margin of which forms a straight and sharp boundary with the unit above.

Unit 3 is the first of 4 major sands exposed in the Pehru section. It consists of flaser bedded, calcareous sand. Drapes of silty clay lie on medium to coarse sand ripples. The unit is very strongly cemented in the top 10 cm. The boundary between units 3 and 4 is sharp

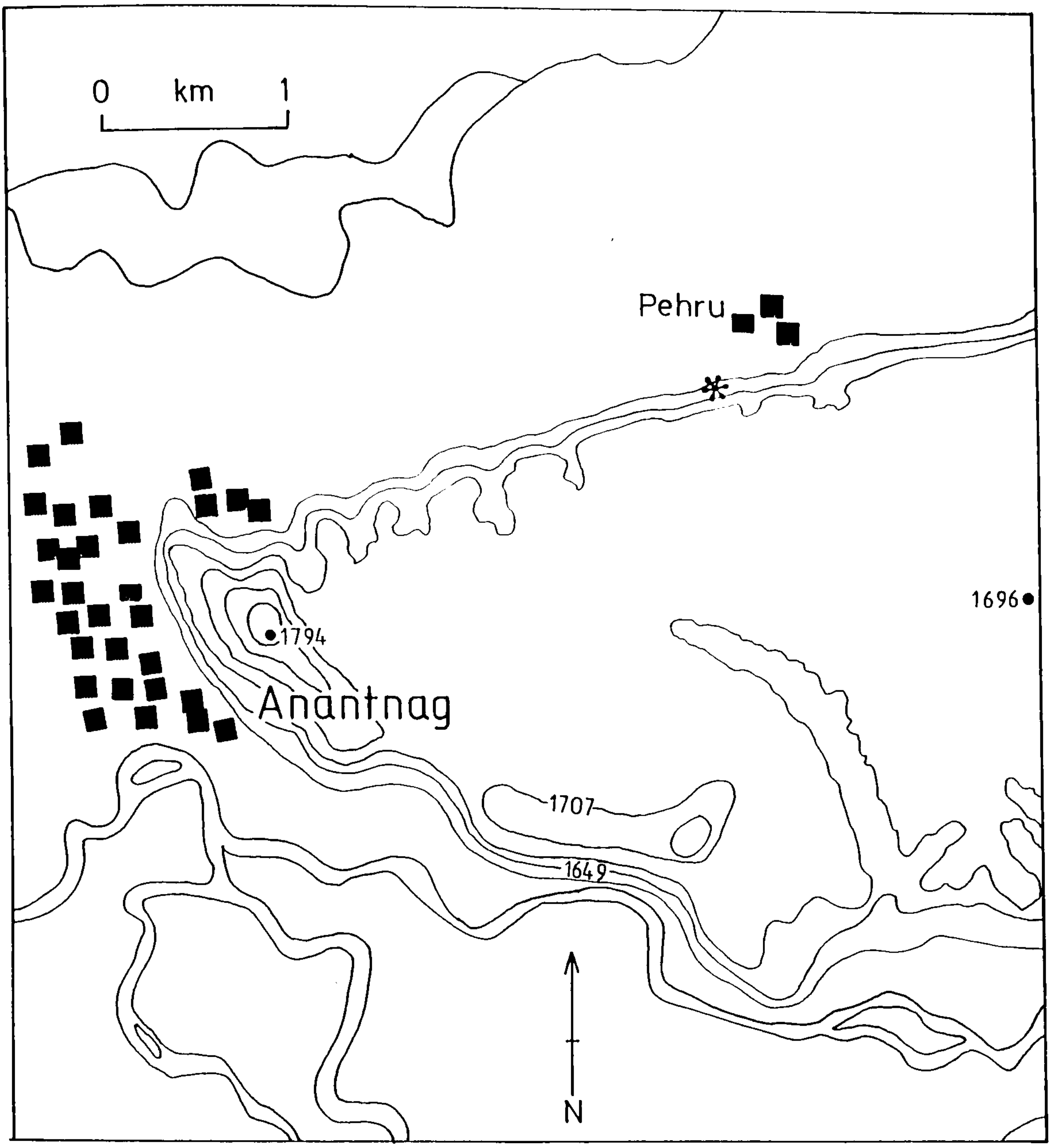


Figure 26

Location map of the Pehru section.

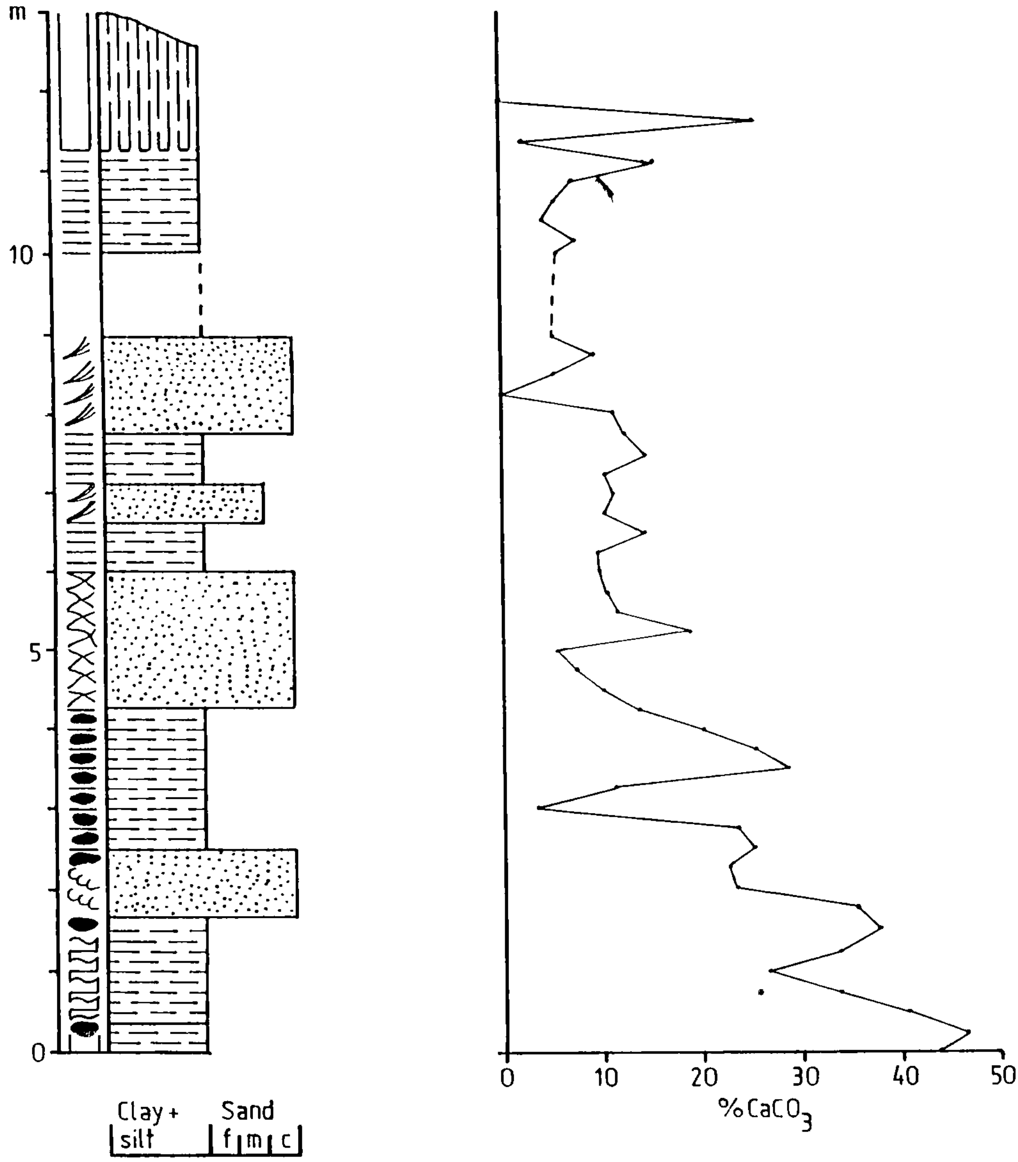


Figure 27
Graphic log diagram and calcium-carbonate curve of the Pehru section.

and straight. Unit 4 consists of silty clay with faint parallel lamination and 8 individual horizons of tabular calcareous concretions. The tabular horizons are very strongly cemented and vary in thickness from 2 to 20 cm. Unit 4 shows an increase in sand in the top 20 cm, although the boundary between it and the overlying sand bed, unit 4, is sharp and straight.

Unit 5 consists of medium to very coarse sand. It shows a type of lenticular bedding structure: small bodies of coarse sand are isolated within medium sand. The boundary between units 5 and 6 is sharp and straight. Units 6 and 8 are of similar character, separated by another sand unit. Units 6 and 8 consist of finely laminated silt clay, with iron oxide staining around moderately abundant fossil root casts. The sand, unit 7, is medium grained, with low-angle cross bedding. Unit 9, the uppermost sand in the section, is similar to unit 7 in lithology and structure.

A hiatus of about 1 m separates the uppermost sand from unit 10, which is a parallel-laminated silty clay, very similar to units 7 and 9. Unit 10 has a diffuse upper boundary and grades into structureless, aeolian loess.

Sambur

3 sections were measured in the Sambur area: a 24 metre-long section about 3 km from Sambur village (plate 4), a short section

behind the village itself and a section 2 km from the village near to Pampur (figures 28 to 31). The first of these is referred to here as the 'Sambur section,' the second as the 'Sambur Village section' and the third, the 'Pampur section.'

The lowest strata exposed in the Sambur section lie 4.3 m above the Jhelum River floodplain. The bottom 2m of sediment, which comprise units 1 to 7, consist of alternations of silty clay, very fine sand and medium sand. The boundaries between each unit are sharp and wavy. The units are typically olive (5Y 5/4) with faint ripple lamination. They are all unfossiliferous. Unit 8 is a ripple-laminated sand that is rather thicker than the underlying units. It marks the end of the coarse-fine sequence represented by units 1 to 7. Unit 8 shows distinct fining and darkening upwards. It is olive (5Y 5/4) in the lower parts darkening gradually to yellowish brown (10YR 5/6) towards the top. Unit 9 shows a similar fining and darkening tendency, but is finer-grained than unit 8. Units 10 and 11 also show the upward change in particle size and colour.

Unit 12 is a major bed of fine to coarse sand. It has a clear and straight boundary with the underlying unit. Unit 12 has a complex sequence of sedimentary structures. On a large scale, it is faintly cross bedded. However, towards the centre of the unit, there is a pronounced 10 cm-thick horizon of trough cross-bedded medium sand which is strongly iron-stained. In the upper 10 cm of unit 12, there is an increase in the silt content within the sand. The upper boundary of the unit is diffuse and shows lenticular bedding. Unit 13 is

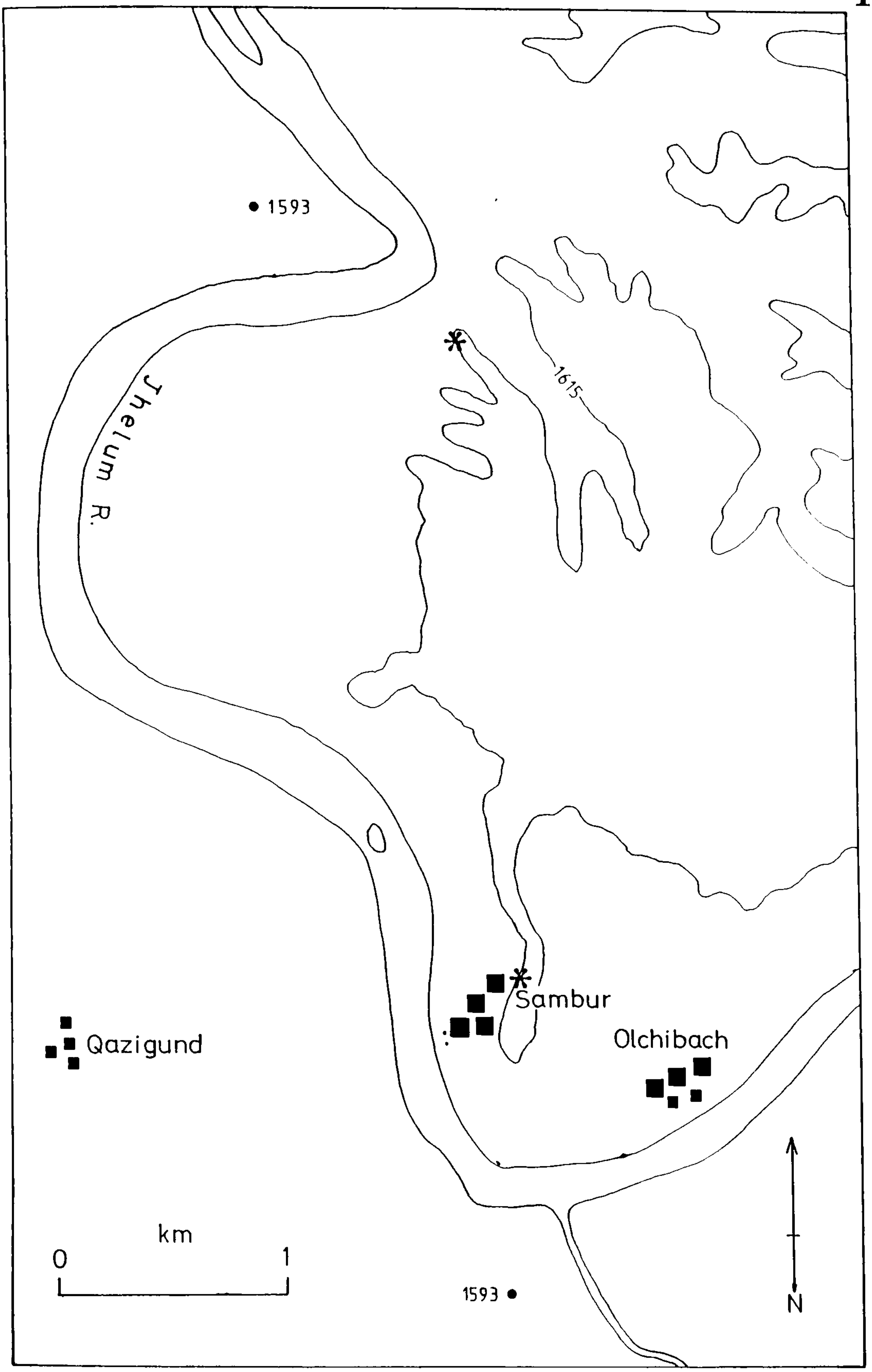


Figure 28
Location map of the Sambur section.

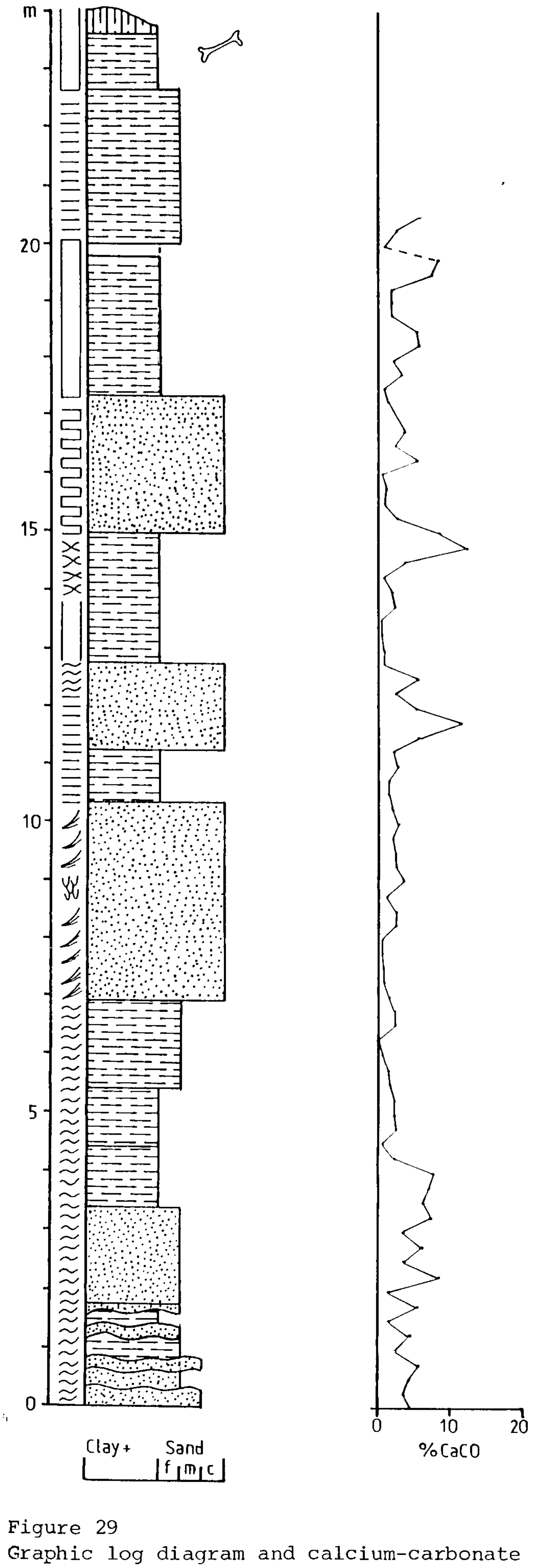


Figure 29
Graphic log diagram and calcium-carbonate curve of the Sambur section.

similar to units 9 to 11. It is fine-grained and shows darkening upwards. It has parallel lamination in the lower part, but the laminae become faint in the upper part. Unit 14 overlies unit 13 with a sharp and straight boundary. Unit 14 consists of fine to coarse sand with parallel bedding and lamination.

Unit 15 marks a return to fine-grained deposition. It consists of silty clay and fine sand. The unit is massive in the lower half but shows lenticular bedding in the upper half. Faint, bright brown (7.5YR 5/8) iron oxide staining is concentrated along the margins of the sand bodies. Unit 16 is the final sand bed in the Sambur section. It overlies unit 15 with a sharp and straight boundary. It consists of fine, medium and coarse cross-bedded sand which is well sorted. It is olive (5Y 4/6) becoming finer and lighter (eg. bright brown-7.5YR 5/8) towards the top. In the upper 20 cm, the sand is interbedded with fine silt to form wavy bedding. Near the base of unit 16, a large vertebrate bone was found. This has been identified as part of a vertebra of <u>Elephas hysudricus</u> (B.S. Kotlia, personal communication).

The rest of the Sambur section consists of fine-grained sediments. Unit 17 is a massive, unfossiliferous silty clay. Unit 18, which is parallel-laminated, is also unfossiliferous. It has clear and straight upper and lower boundaries. Bright brown (7.5YR 5/8) iron oxide staining follows the laminae boundaries. Unit 19 is a massive, fractured clayey silt to silty clay, which overlies the unit below with a clear and straight boundary. Unit 19 is dull yellowish brown (10YR 4/3) with bright yellowish brown (10YR 6/8) iron oxide coatings

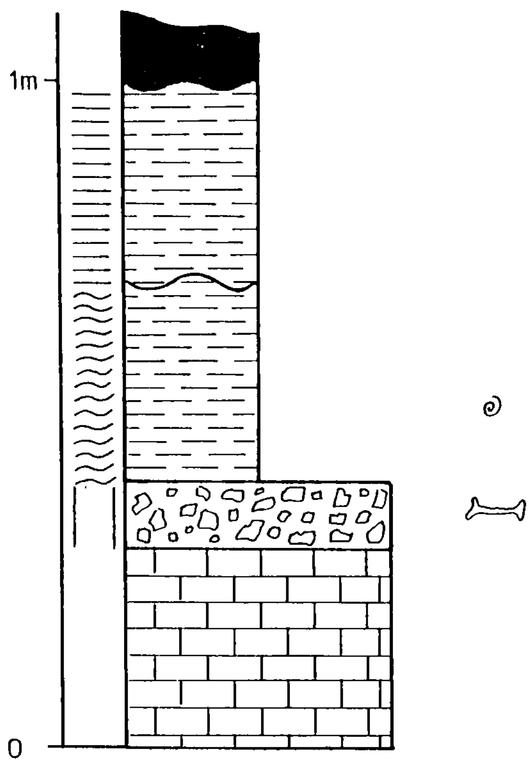


Figure 30

Graphic log diagram of the Sambur Village section.

and grey (5Y 6/1) veneers interpreted as reprecipitated calcium carbonate, on the surfaces of the fractured blocks. Unit 19 has a clear and straight upper boundary with overlying aoelian loess.

The Sambur village setion was located in the mesa edge behind the village of Sambur (figures 29 and 30). The sediments of this section rest on Triassic limestone. The uppermost surface of the bedrock has been intensively eroded, to such an extent that it breaks quite easily in the hand. Overlying the limestone is unit 1 of the section, a thin horizon of fossiliferous breccia and gravel. This is the 'Bone bed' originally described by de Terra and Paterson (1939). It contains a mixture of well-rounded pebbles and very angular clasts, both types of the local parent material. The matrix of this material is poorly sorted sand. Unit 2 of this section is ripple-laminated silty clay. Iron oxide staining follows the laminae boundaries. This unit was found to be very rich in fossil mollusca. The majority of specimens found were Valvata piscinalis, although specimens of Gryaulus convexiusculus and Lymnaea ?auricularia (Identifications by D.S. Brown, personal communication) were also found. The boundary between units 2 and 3 is sharp and wavy. Unit 3 is similar to unit 2, except that it is parallel-laminated and was found to be unfossiliferous. Unit 3 is overlain by Neolithic archaeological spoil.

The Pampur section is located about 2 km south of Pampur Village on the eastern side of National Highway 1A. It was dug in a disused quarry (figures 31 and 32). The base of the measured section lies 3 m above the base of the Jhelum floodplain.

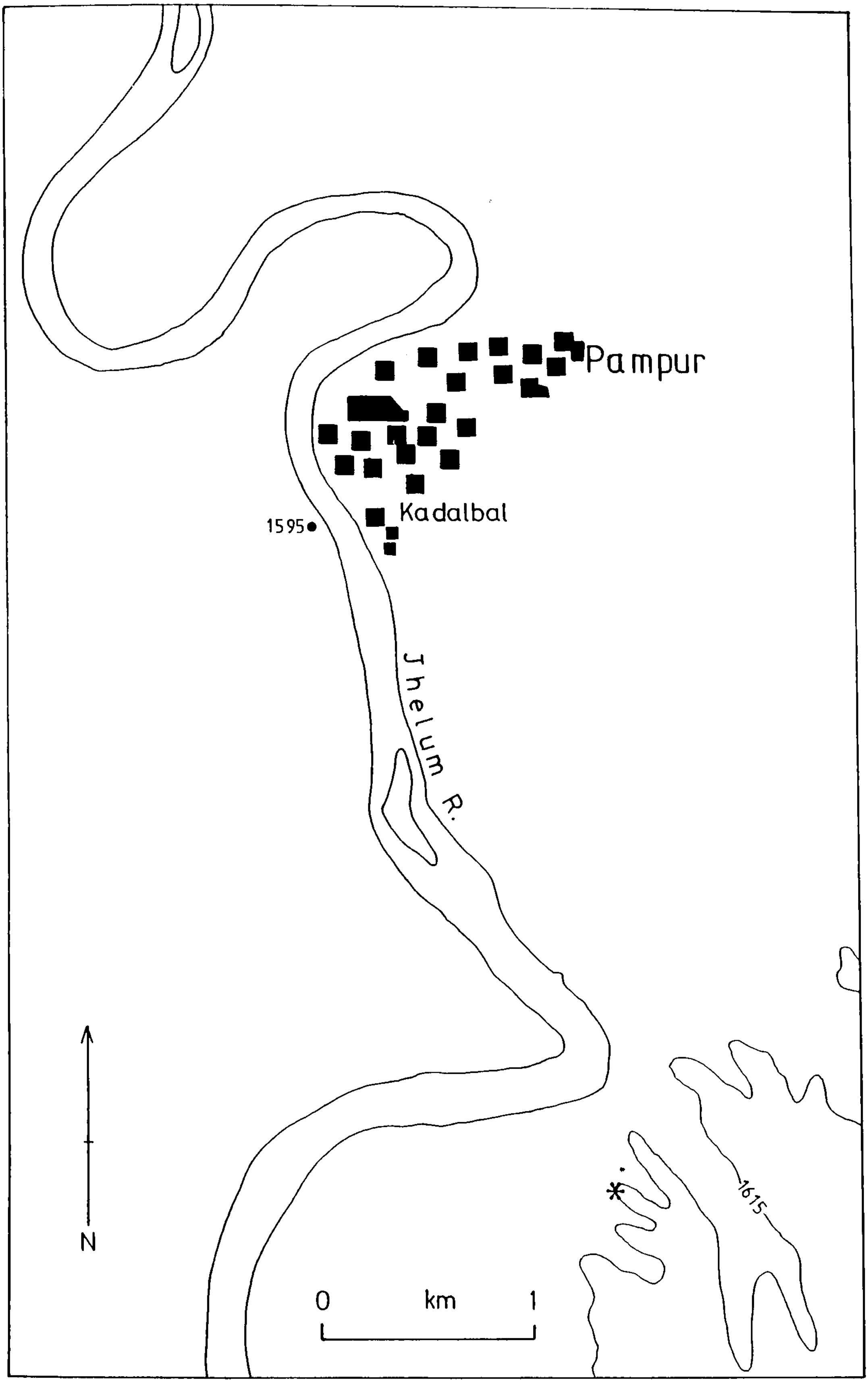


Figure 31

Location map of the Pampur section

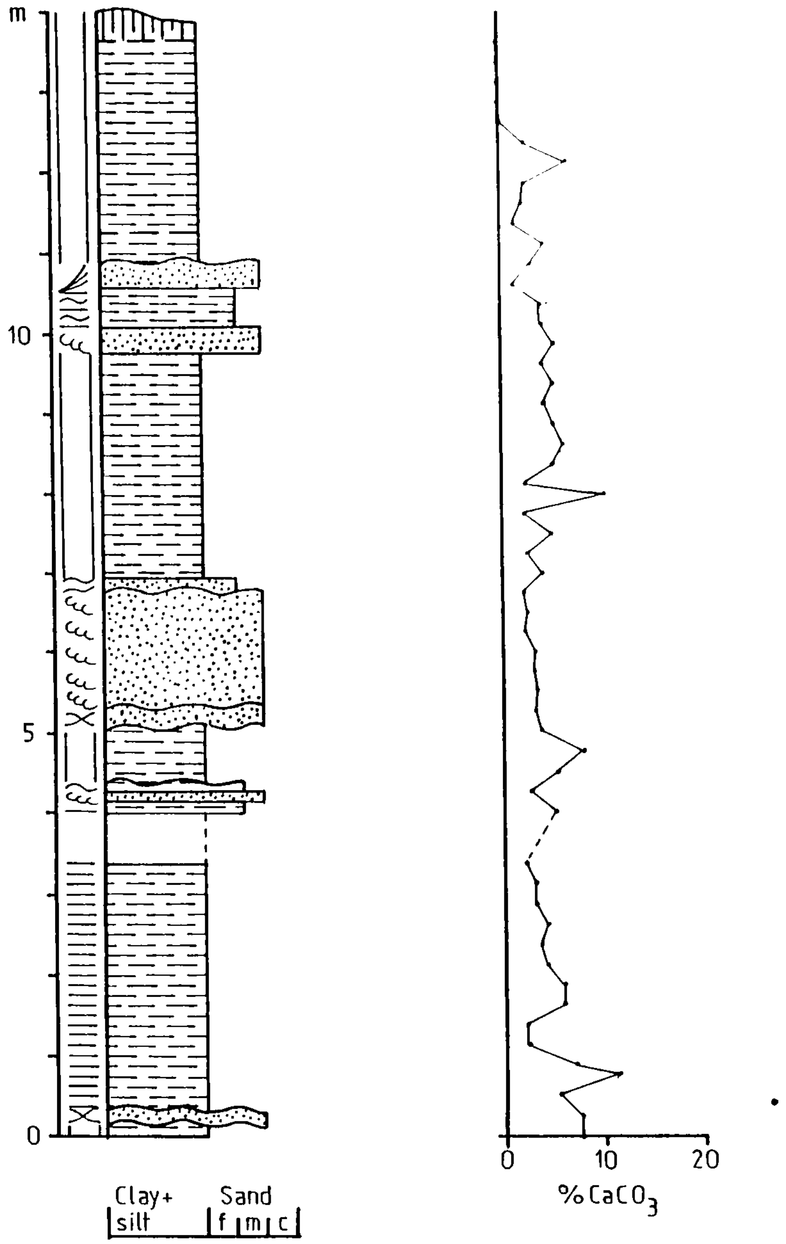


Figure 32

Graphic log diagram and calcium-carbonate curve of the Pampur section.

Unit 1 of the Pampur section is massive clayey silt with unidentifiable mollusc fragments and medium to fine iron-stained mottles. Unit 2 overlies unit 1 with a sharp and wavy boundary. Unit 2 is lenticular bedded . It consists of small bodies of fine to medium sand in silty clay. Unit 3 is a clayey silt to silty clay. It is greyish yellow brown (10YR 5/2), shows parallel lamination with occasional layers of fine sand and rarely, yellowish brown (10YR 5/6) mottles. The boundary between units 3 and 4 was not seen in the field as there is a short hiatus in the section at this point. Unit 4 is a thin bed of faintly laminated silty clay and very fine sand. It is unfossiliferous in hand specimen, calcareous and yellowish brown (2.5Y 5/6). Unit 5 shows flaser bedding. It has silty clay drapes on ripple-bedded medium sand. The individual sand layers vary in thickness from 1 to 4 cm. The mud drapes are 1 to 3 cm thick.

A smooth and sharp boundary separates units 5 and 6. Unit 6 consists of ripple-bedded silt to very fine sand which is unfossiliferous in hand specimen. It has a sharp and wavy upper boundary separating it from unit 7 which is a massive, fossiliferous silty clay. Unit 7 has quite intense bright brown (7.5YR 5/8) to reddish black (10R 2/1) iron oxide staining along possible former vertical root casts. Small, individual reddish black (10R 1.7/1) traces, possibly manganese stains, are also common. Fragments of unidentifiable molluscs are moderately abundant.

Units 8 and 9 consist of fine to medium sand, and silty clay. Unit 8 shows lenticular bedding, with sand bodies 1 to 4 cm thick. Bright brown (7.5YR 5/8) iron oxide staining is concentrated along the margins of the individual sand bodies. Unit 9 shows flaser bedding. The sandy layers become thinner towards the top of the unit and the mud drapes thicken, giving way to wavy bedding in the upper part of the unit. A sharp and wavy boundary separates units 9 and 10. Unit 10 is silty clay and fine sand which is ripple laminated. The unit shows brown (10YR 4/6) staining in the upper 10 cm.

Unit 11 is a massive clayey silt, which has a sharp and straight boundary with the unit below. It is greyish olive (5Y 3/3) and contains sparse fragments of molluscs that were unidentifiable. Iron oxide staining is concentrated around sparse fossil root remains. A sharp and straight boundary separates units 11 and 12. Unit 12 is a medium sand that is flaser bedded. The silty clay drapes on the sand ripples themselves show fine parallel lamination. Iron oxide staining is commonly concentrated along the margins of the laminae. Unit 13, which overlies unit 12 with a sharp and straight boundary, consists of parallel and ripple laminae of silty clay and very fine sand. The unit is unfossiliferous in hand specimen. Iron oxide staining is commonly concentrated along the laminae boundaries. Unit 13 is overlain by unit 14, a fine to medium sand. The sand shows faint, planar cross-bedding. Iron oxide staining is concentrated in the top 5 cm of the unit. Unit 14 has a sharp and wavy upper boundary. Unit 15 is a massive, unfossiliferous silty clay with faint iron oxide staining. Its upper boundary is diffuse.

Unit 16 is a dull yellowish brown (10YR 4/3) clayey silt. It is well fractured and shows bright yellowish brown (10YR 6/8) and grey (10Y 6/1) coats on the surfaces of the fractured blocks. These coats are interpreted as iron oxide and reprecipitated calcium carbonate respectively. This unit has a diffuse upper boundary; it grades into aeolian loess.

Dilpur-Qasba Nagum

A series of sections between Dilpur and Qasba Nagum was examined, but not in detail (figure 16). The section at Dilpur is exposed in a gully. Almost 20 m of loess overlies a conglomerate unit which is about 8 m thick. Beneath this conglomerate is a sequence of contorted fine-grained sediments. At Qasba Nagum, a similar sequence was seen, except that about 1 metre thickness of fine-grained, massive and parallel-laminated sediments occur between the conglomerate and the loess. At Qasba Nagum, the conglomerate is only 1 m thick. This sequence was also observed by Bhatt (1982).

Neu-Zadora

On the road between the villages of Neu and Zadora (figure 16), a similar sequence to that at Dilpur-Qasba Nagum was seen. Over a horizontal distance of about 300 m from east to west, laminated silty clay beds found between conglomerate and loess, pinch out. In the western part of this exposure, the loess therefore lies directly on conglomerate.

Nunar-Malshahibagh

Conglomerates forming the Sind Valley terraces were examined at Nunar (plate 6). The conglomerates are well exposed in this area in cuttings made for irrigation channels on the left hand side of the valley. At Nunar, the terraces are 27 m above the Sind River floodplain, although they thicken considerably upvalley. The conglomerates are invariably capped by loess which, at the measured section, is 9 m thick.

The conglomerate is clast supported and noncalcareous, although a poorly sorted coarse sand matrix fills the interstices between the clasts. The clasts are subrounded to rounded and typically between 15 and 30 cm long. Exceptionally, however, boulders more than 100 cm were seen. Generally, the conglomerate is massive, with a crude down-valley imbrication of clasts. Locally, however, lenses of moderately well sorted sand, up to 2 m across and 0.3 m thick, were seen.

A conglomerate is also exposed at Malshahibagh, beyond the mouth of the Sind Valley. However, physical continuity could not be established between this exposure and the terrace conglomerate at Nunar. The exposure at Malshahibagh is on the righthand side of the main highway from Srinagar to the Sind Valley. The exposure is well vegetated and partly hidden behind a retaining wall.

The conglomerate consists of rounded to well rounded clasts, which are very strongly cemented in a sandy matrix. No bedding structures were seen in the field, although this may be due to the limited extent of the exposure. Digging only slightly below the base of the exposure, revealed massive, fine-grained sediments. Careful excavation revealed that these were underlying the conglomerate and not simply banked up against it. The conglomerate was traced about 1 km northwards along the edge of the highway as far as Ganderbal. In a quarry at Ganderbal, the conglomerate was seen underlying fine-grained sediment.

Kahalwan

Terrace conglomerates, similar to those found in the outer Sind Valley at Nunar, were exposed in the outer part of the Liddar valley. The terraces are much lower and more degraded here than in the Sind. At Kahalwan, on the righthand side of the valley, a unit of noncalcareous, clast supported conglomerate was seen underlying fine-grained sediments. The conglomerate bed is overlain directly by a unit of planar cross-bedded, well sorted, medium sand. This, in turn, is overlain by massive clayey silt. The fine-grained unit contains numerous root fossils with concentrations of iron oxide. The sequence is capped by aeolian loess.

Karpora

The Karpora section is located in a river cliff exposed in Kanchi Kol Nala, close to the village of Karpora (figure 16). In contrast to the other sections described here, Karpora lies above 1680 m a.s.l. and hence beyond the inferred limits of the upper Karewa lake. The section was measured in order to test the stratigraphic framework for the upper Karewa established as a result of the measurement of other sections (figure 33). The beds at Karpora are tilted, and dip in a

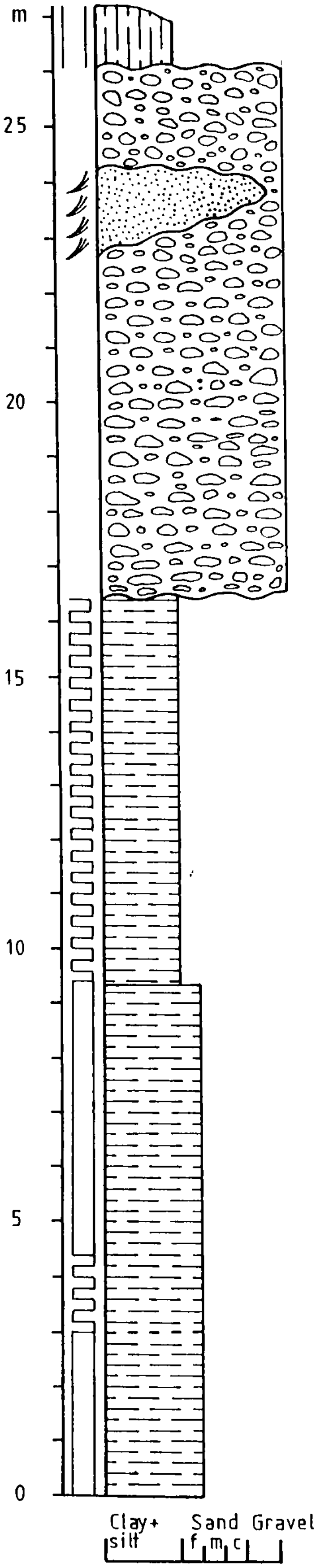


Figure 33

Graphic log diagram of the Karpora section.

north-easterly direction. Because of this, it was possible to trace the strata simply by moving up the river valley.

The individual units in this section were much thicker than the other units of upper Karewa lacustrine strata, with the exception of the lowest units in the Badgam section. Unit 1 consists of a massive silty clay to very fine sand which is grey (10Y 4/1) in colour. This is overlain by a greyish olive (5Y 4/2) unit and the boundary between the two is diffuse. The basal metre of unit 2 consists of parallel thin bedding picked out by yellowish brown (2.5Y 5/6) colouration. A clear and straight boundary separates this from unit 3, which is a parallel-laminated to parallel-bedded clayey silt. The colour of this unit varies from dark bluish grey (10BG 3/1) to olive yellow (7.5Y 6/3).

A sharp but wavy, and clearly erosive contact separates unit 3 from the overlying conglomerate, unit 4. The conglomerate is loose and has a poorly sorted, sandy matrix. In parts, the conglomerate is massive, although elswhere it has crude, planar and trough cross-bedding with vertical fining-upwards. In some parts, the conglomerate is a single unit. However, in much of the exposure, it is divided by a unit, about 1 metre thick, of coarse sand. The conglomerate unit is capped by loess throughout the section although rarely, a very thin sand unit occurs on top of the conglomerate. Throughout the measured section, the sediments were found to be noncalcareous and unfossiliferous.

c. Sedimentological Analysis.

Particle size analysis was carried out on samples of upper Karewa fine-grained sediments, loess, and conglomerate matrix. Samples were selected from many of the sections described in this chapter (appendix 5). In addition, the Pattan section was sampled and analysed systematically (figure 19). Analysis of clast lithologies from upper Karewa conglomerates was undertaken at 12 sites. Selected samples of upper Karewa fine-grained sediments and aeolian loess were taken for microfabric analysis.

<u>Particle-size analysis: upper Karewa fine-grained sediments and loess</u>

Observations on the fine-grained upper Karewa units in the field suggested that they were similar in texture to the overlying loess. Particle-size analysis was carried out as an initial step in testing the hypothesis that the upper Karewa lacustrine deposits may be partly loessic in origin. Summary particle-size diagrams are shown in figures 34 to 40.

The particle size envelope (figure 34) clearly shows that there is wide textural variation between samples. The sand content varies between 0 and 14%. The silt content varies from about 50% to over 90%. The clay content is between 5 and over 50%. (figure 35) Mean particle size ranges from less than 6 phi to about 10 phi. However, there are three clusters recognisable on the basis of mean size, together with one anomalously fine sample (number 107). The first of the cluster centres on 6 phi, the second on 7 phi and the third on 8.5 phi. On the

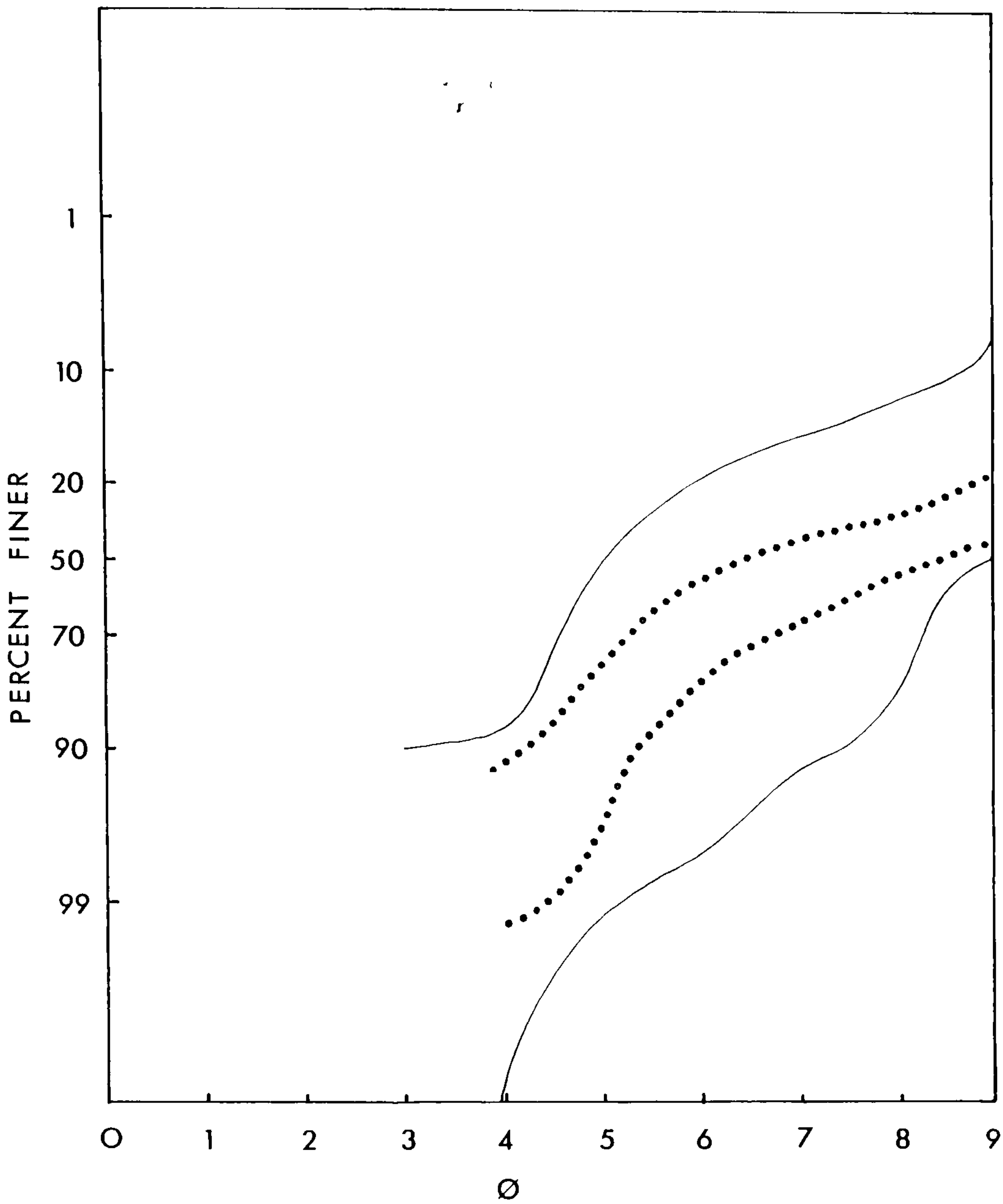


Figure 34

Particle-size envelope: upper Karewa silts (solid line) and aeolian loess (dotted line).

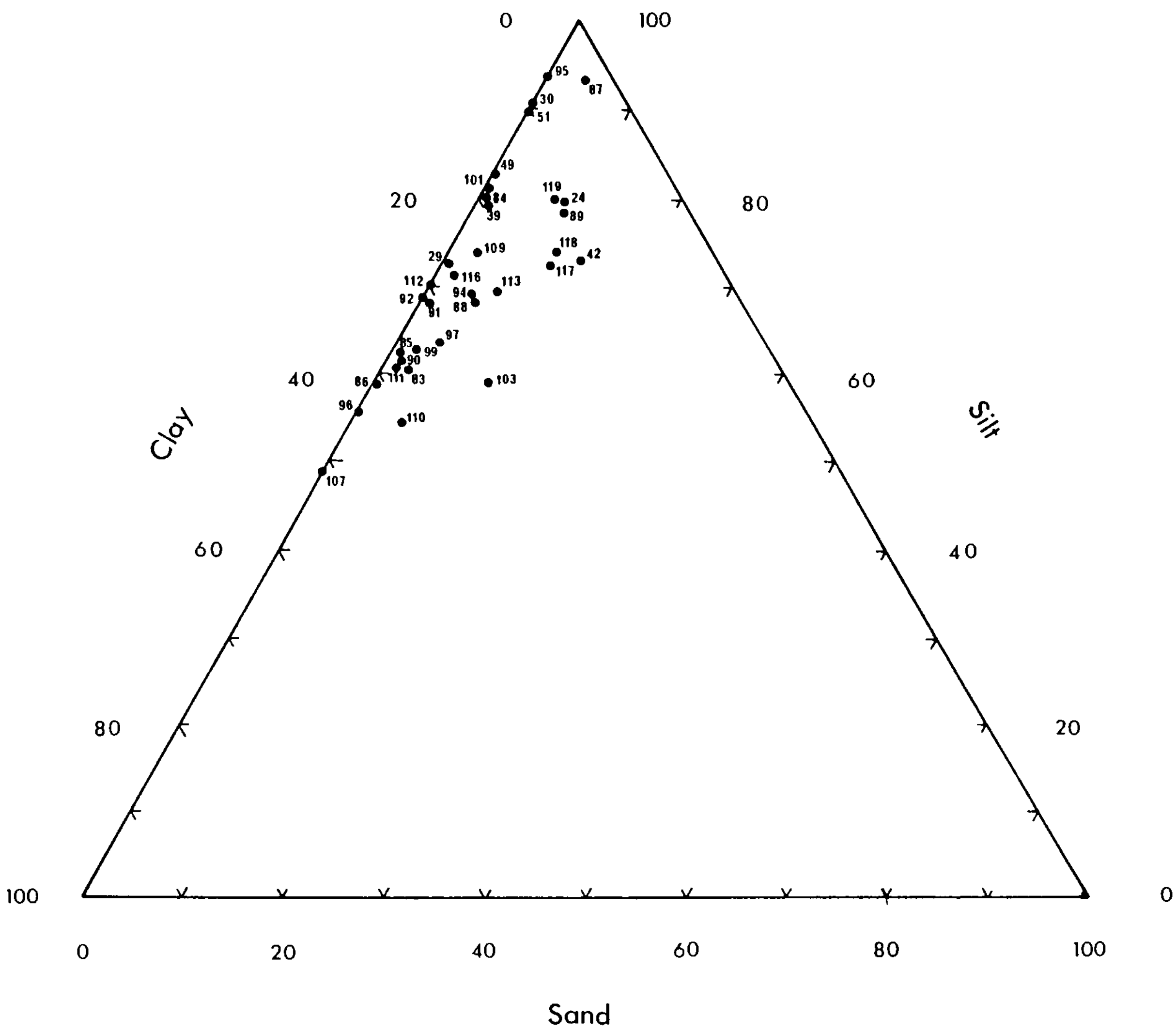


Figure 35

Ternary diagram of percent sand, silt and clay: upper Karewa silts.

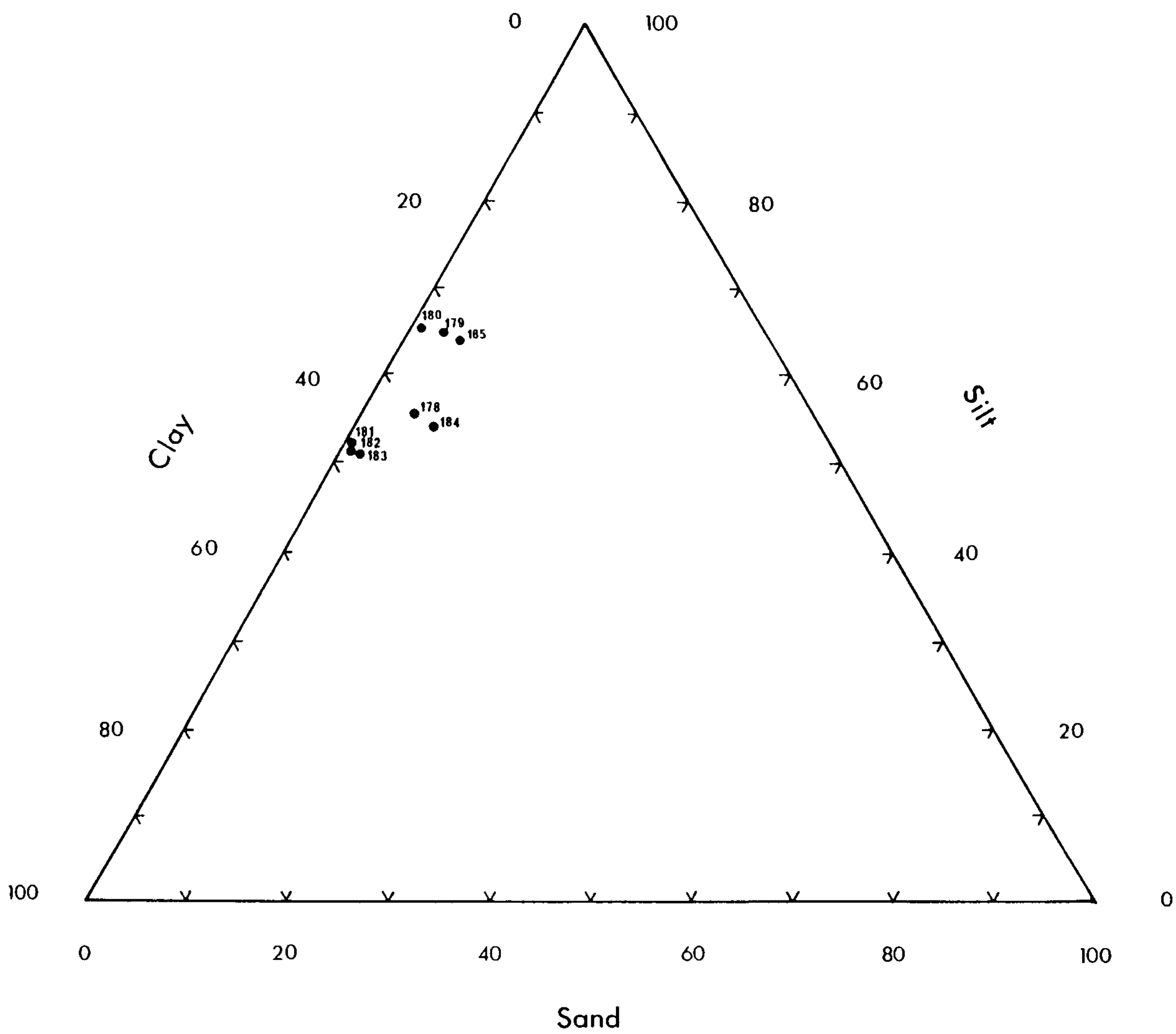


Figure 36

Ternary diagram of percent sand, silt and clay: aeolian loess.

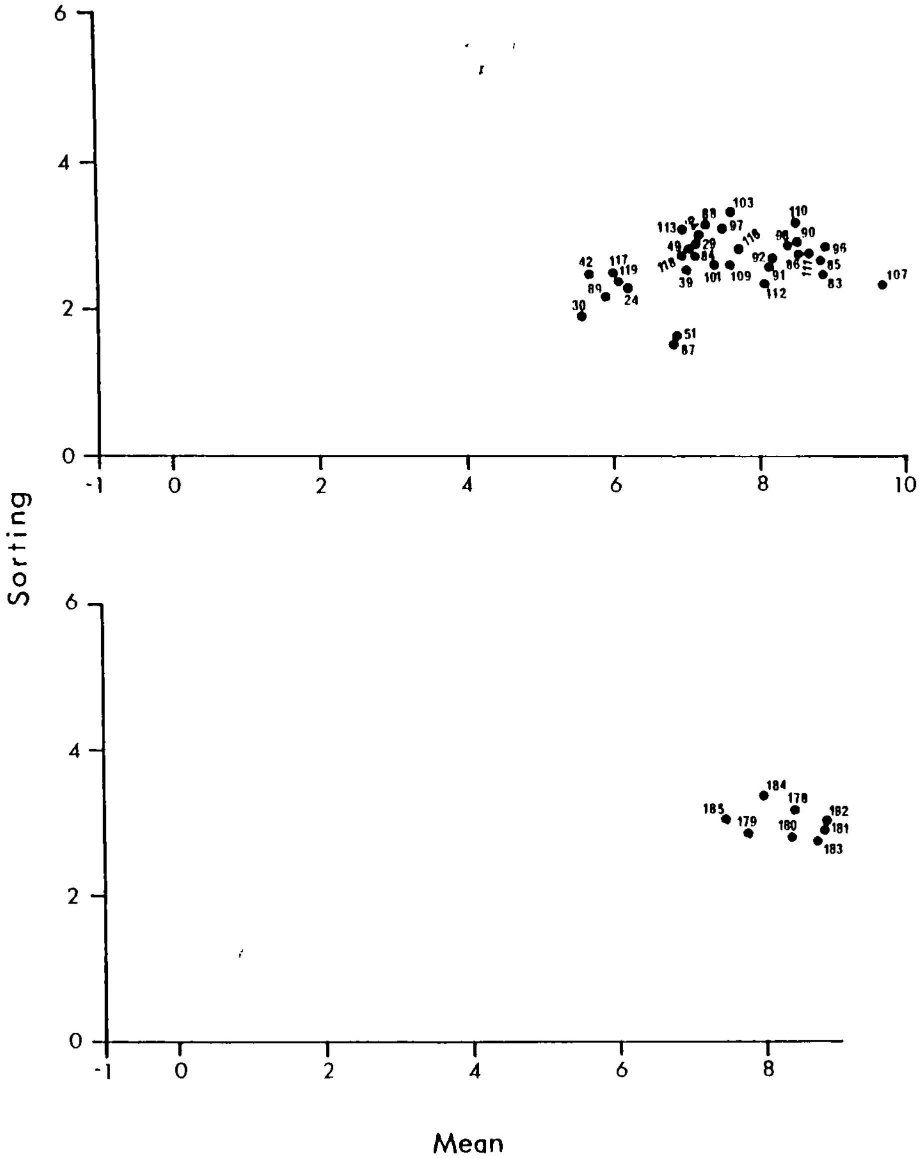


Figure 37

Bivariate scattergram of mean and sorting: upper Karewa silts (top) and aeolian loess (bottom).

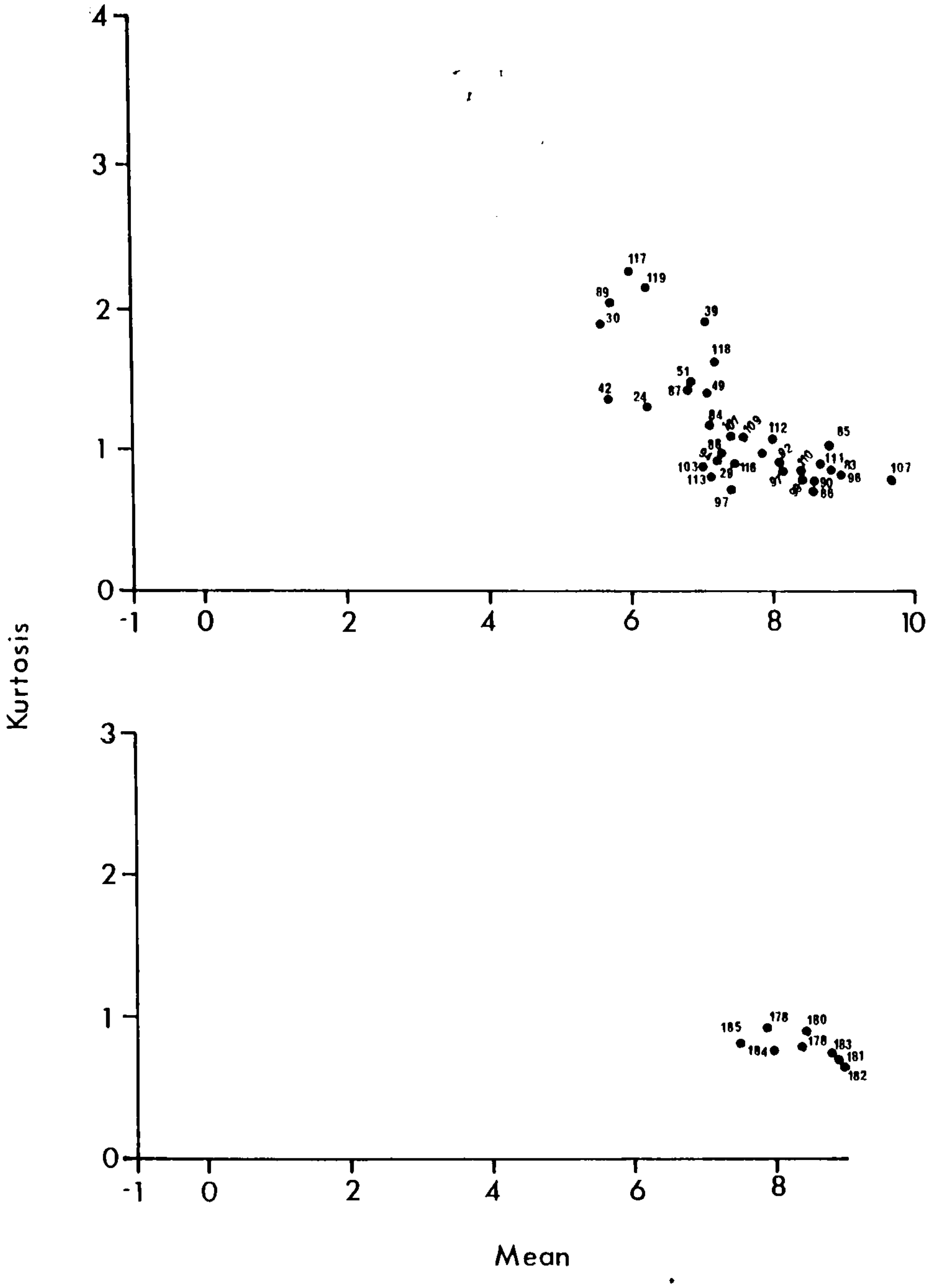


Figure 38

Bivariate scattergram of mean and kurtosis: upper Karewa silts (top) and aeolian loess (bottom).

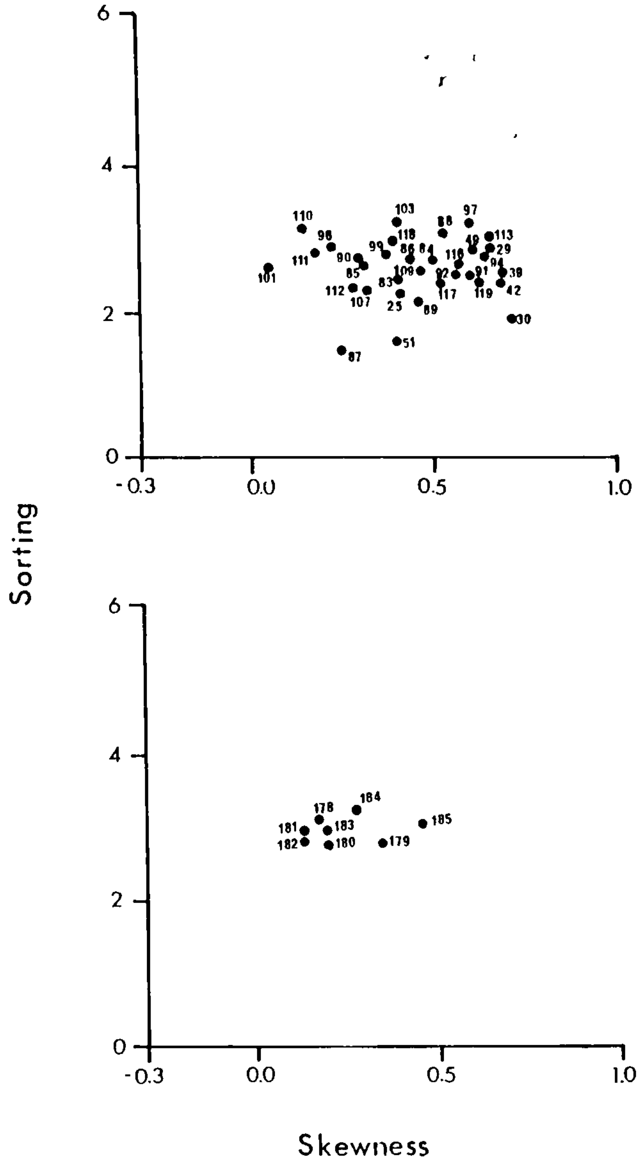


Figure 39

Bivariate scattergram of Skewness and sorting: upper Karewa silts (top) and aeolian loess (bottom).

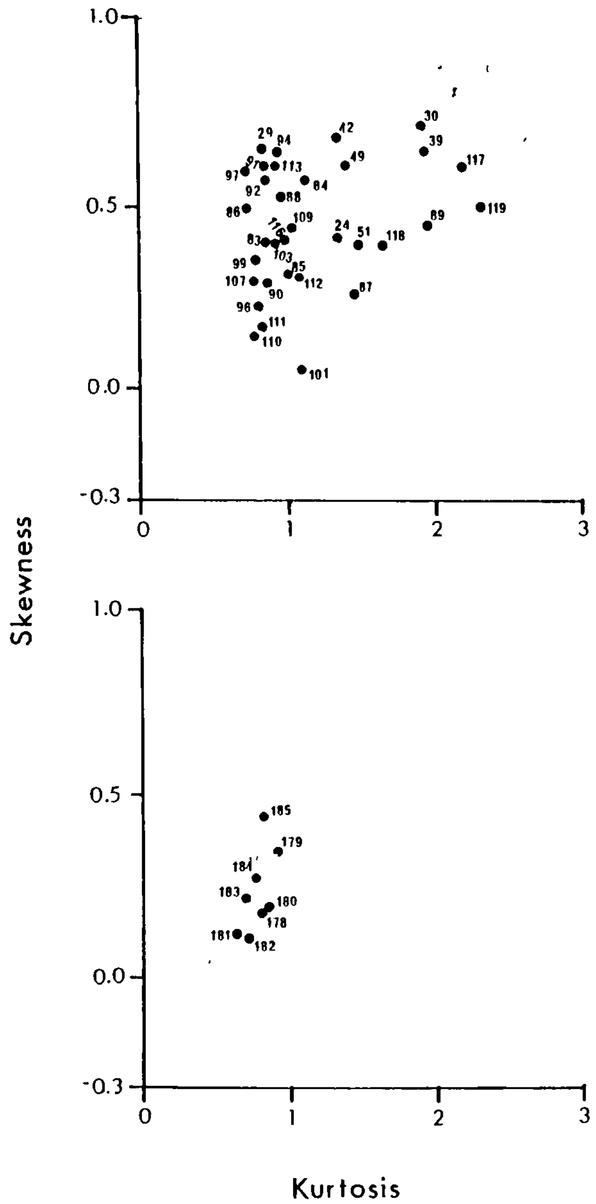


Figure 40

Bivariate scattergram of kurtosis and skewness: upper Karewa silts (top) and aeolian loess (bottom).

basis of mean particle size, the sediments are classified as medium to fine silts.

All of the samples fall into the poorly sorted or very poorly sorted categories of Folk and Ward (1957), with sorting values of around 2.25 phi (see, for example, figure 37). However, there is a considerable scatter of values. The most well sorted sample has a sorting value of 1.48 (sample 87) and the least well sorted sample a value of 3.25 (sample 103) Most of the samples analysed are fine-skewed or near-symmetrical. Values of kurtosis range from abpout 0.5 to 2.5 (very platykurtic to very leptokurtic).

Sample number 95, taken from the Wogahoma section, is an exception to most of the above generalizations. It is plotted on the ternary diagram but has been omitted from the bivariate scattergrams since its inclusion caused scaling problems. The sample is almost a pure silt. It contains no sand and only 6% clay. Over 80% of the particles fall within the 4 phi to 2 phi range (fine silt). Sample 95 has a sorting value of 0.68 phi (moderately well sorted), a skewness value of -0.59 (strongly coarse-skewed) and a kurtosis value of 3.83 (extremely leptokurtic).

Particle-size analysis was carried out on 8 loess samples from the Burzahom section.The particle size envelope for the loesses (figure 34) overlaps with that for the upper Karewa samples described above. A ternary diagram showing the loess samples is shown in figure 36 and various bivariate scattergrams of particle size statistics in figures

37 to 40. The mean particle size for the samples varies from 6.55 phi (sample 185) to 8.66 phi (sample 181). (medium to fine silts). Sorting values range from 2.76 to 3.3 phi, (very poorly sorted), skewness values range from 0.12 to 0.66 (fine to strongly fine skewed) with kurtosis values between 0.69 and 1.83 (platykurtic to very leptokurtic). The samples have low percentages of sand (0.6 to 5%) which falls exclusively in the fine sand grade. Silt content varies between about 50 and 65%; clay content between 29 and 47%.

Particle-size analysis: modern fluvial sands and upper Karewa sands

Summary particle-size diagrams are shown in figures 41 to 47.Sediment samples were taken from the beds of several modern rivers in Kashmir including the Sind, the Liddar and the Hakraj (a tributary of the Romushi River). Samples were invariably taken from thin surface layers to avoid possible vertical mixing of different sediment types. The samples came from a number of different fluvial environments. These included a proglacial area (West Liddar Valley), a braided reach of a river (the Hakraj River) and a meandering reach (The Middle Sind River). The particle size envelope shows wide variation in sample characteristics. Most of the samples contain more than 60% sand, a varying amount of silt and, usually, no clay. Sample 52, which is a fine-grained sediment from a small bedform in the meandering reach of the Sind River, is an exception to this pattern. The sample is composed mostly of silt, containing no sand and little clay (see figure 42).

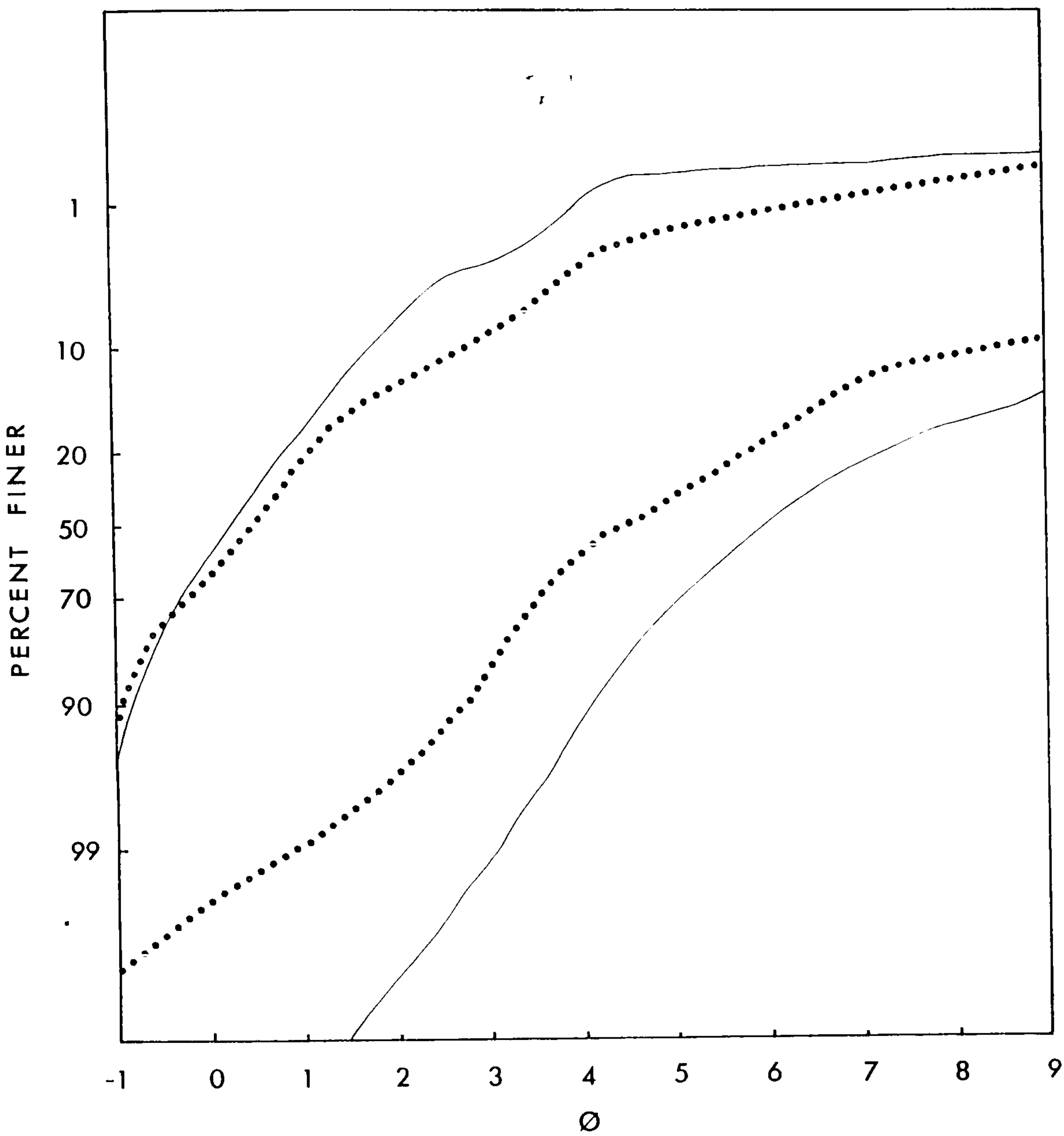


Figure 41′

Particle-size envelope: modern fluvial sands (dotted line) and upper Karewa sands (solid line).

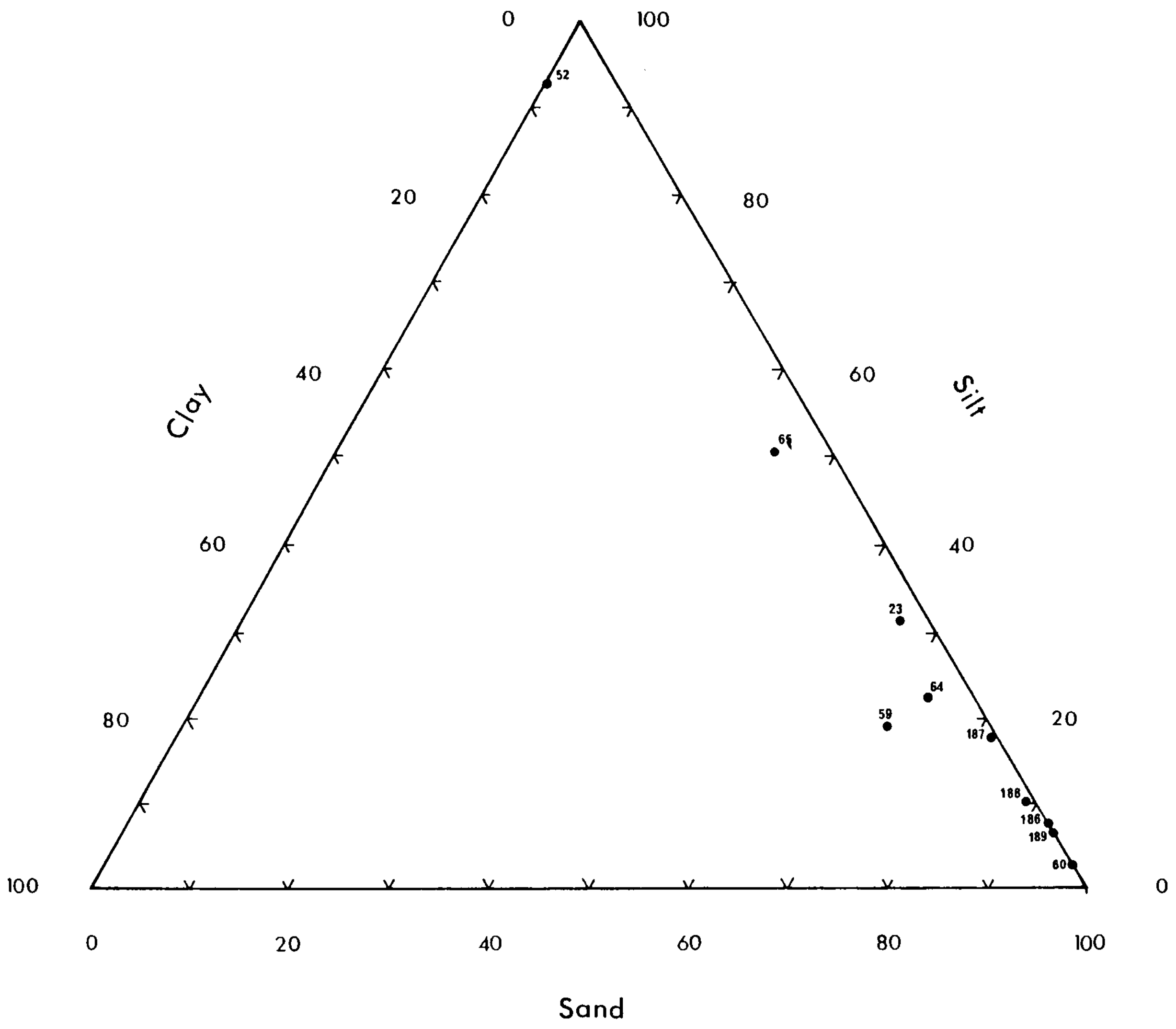


Figure 42

Ternary diagram of percent sand, silt and clay: modern fluvial sands.

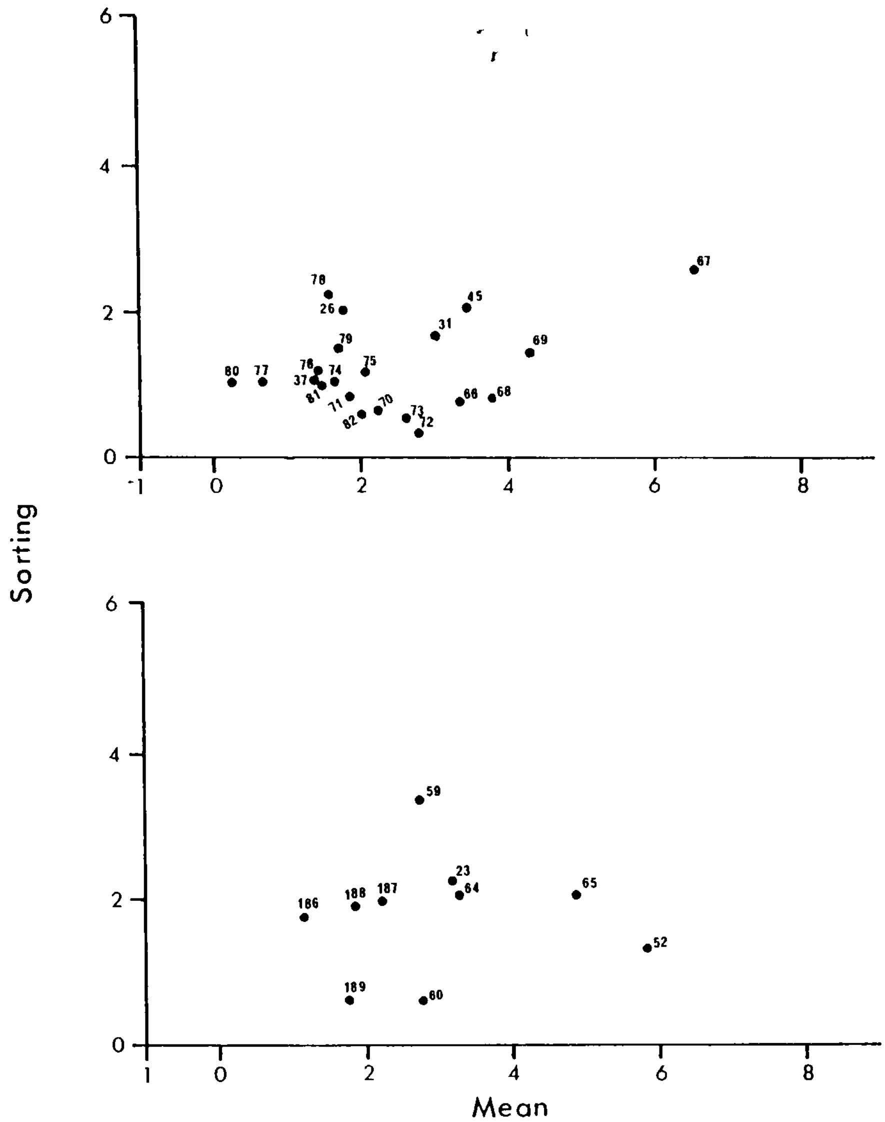


Figure 43

Bivariate scattergram of mean and sorting: modern fluvial sands (bottom) and upper Karewa sands (top).

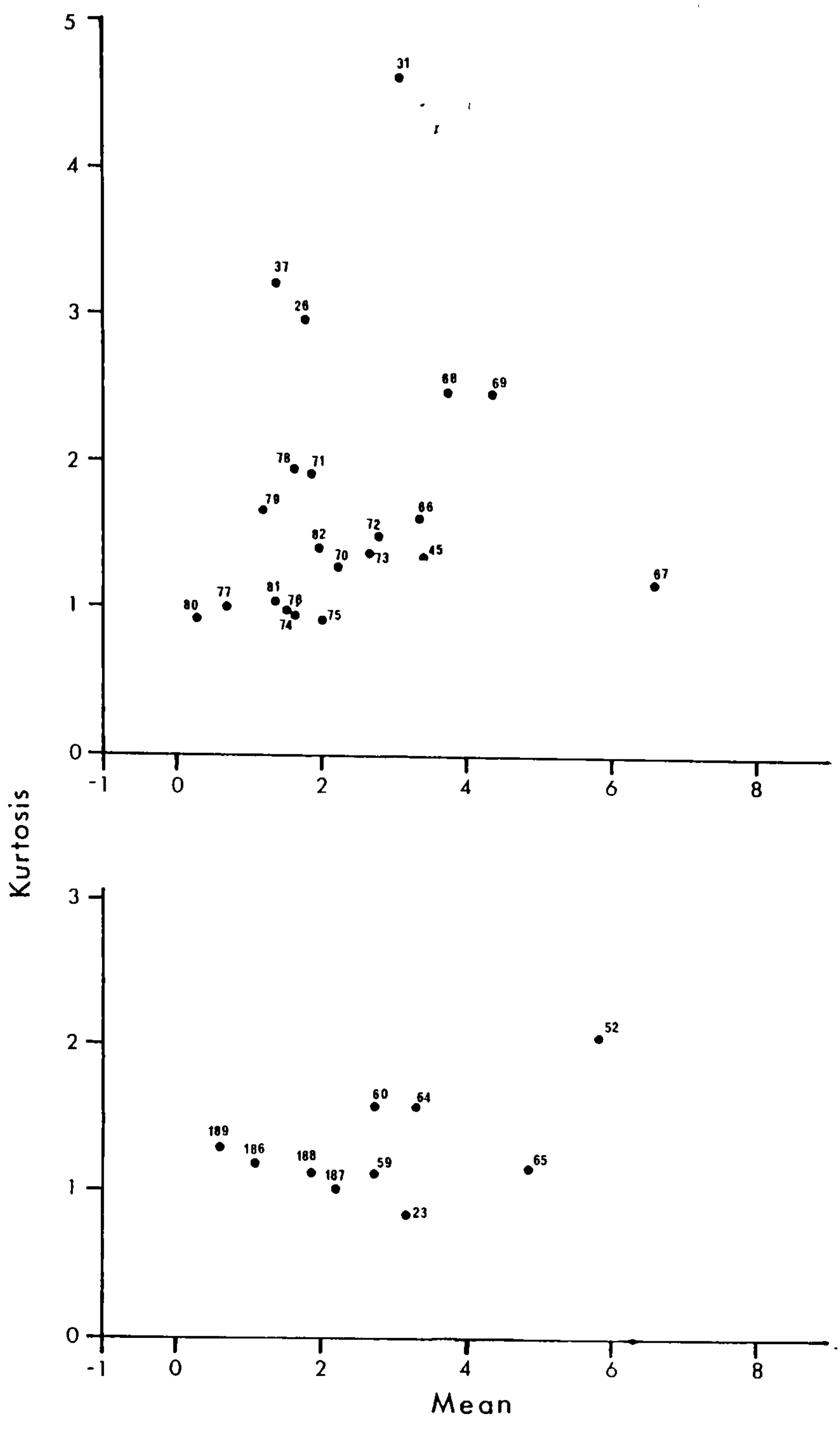


Figure 44

Bivariate scattergram of mean and kurtosis: modern fluvial sands (bottom) and upper Karewa sands (top).

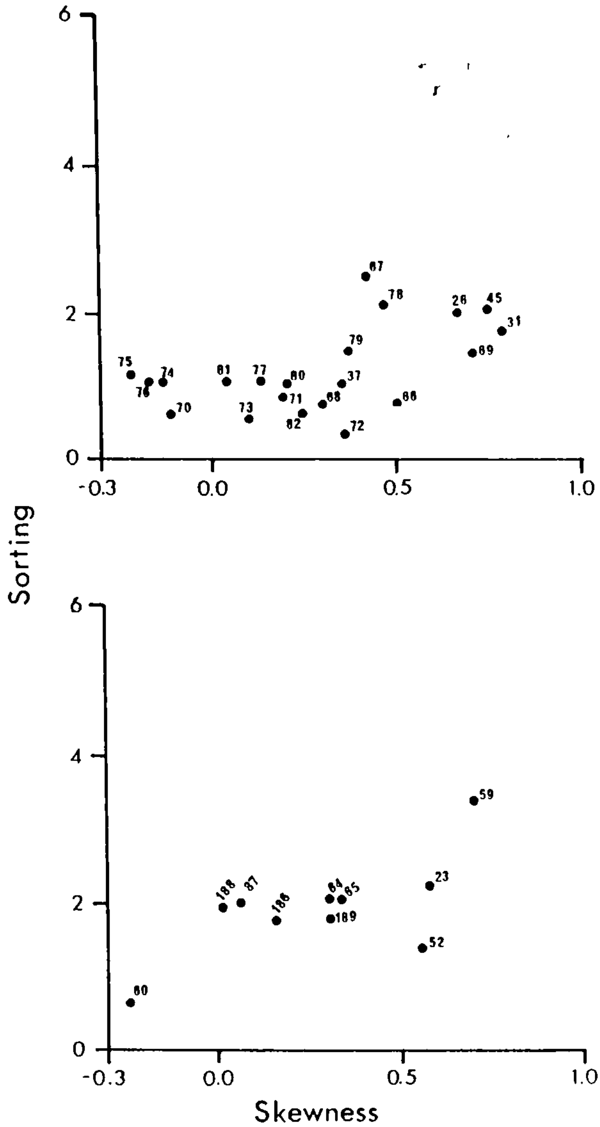


Figure 45

Bivariate scattergram of Skewness and sorting: modern fluvial sands (bottom) and upper Karewa sands (top).

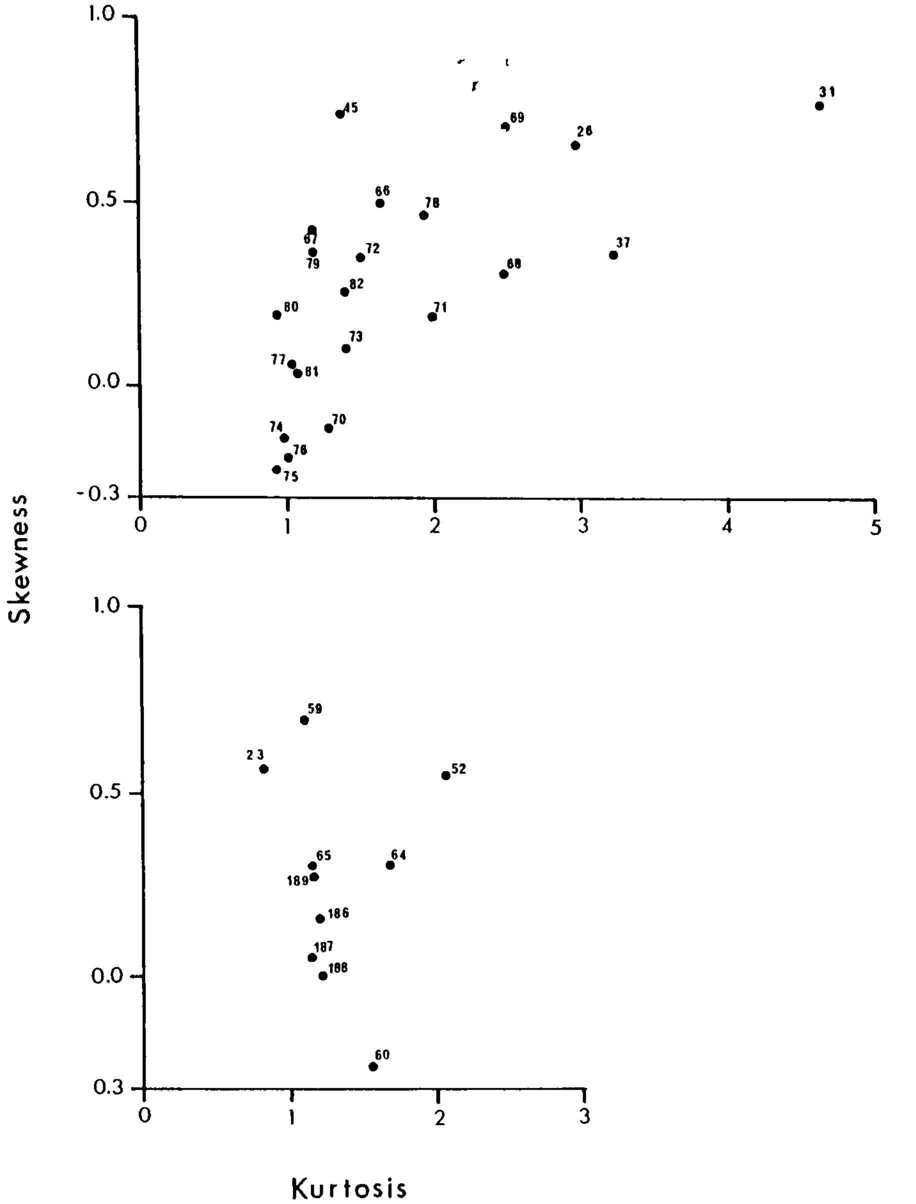


Figure 46

Bivariate scattergram of kurtosis and skewness: modern fluvial sands (bottom) and upper Karewa sands (top).

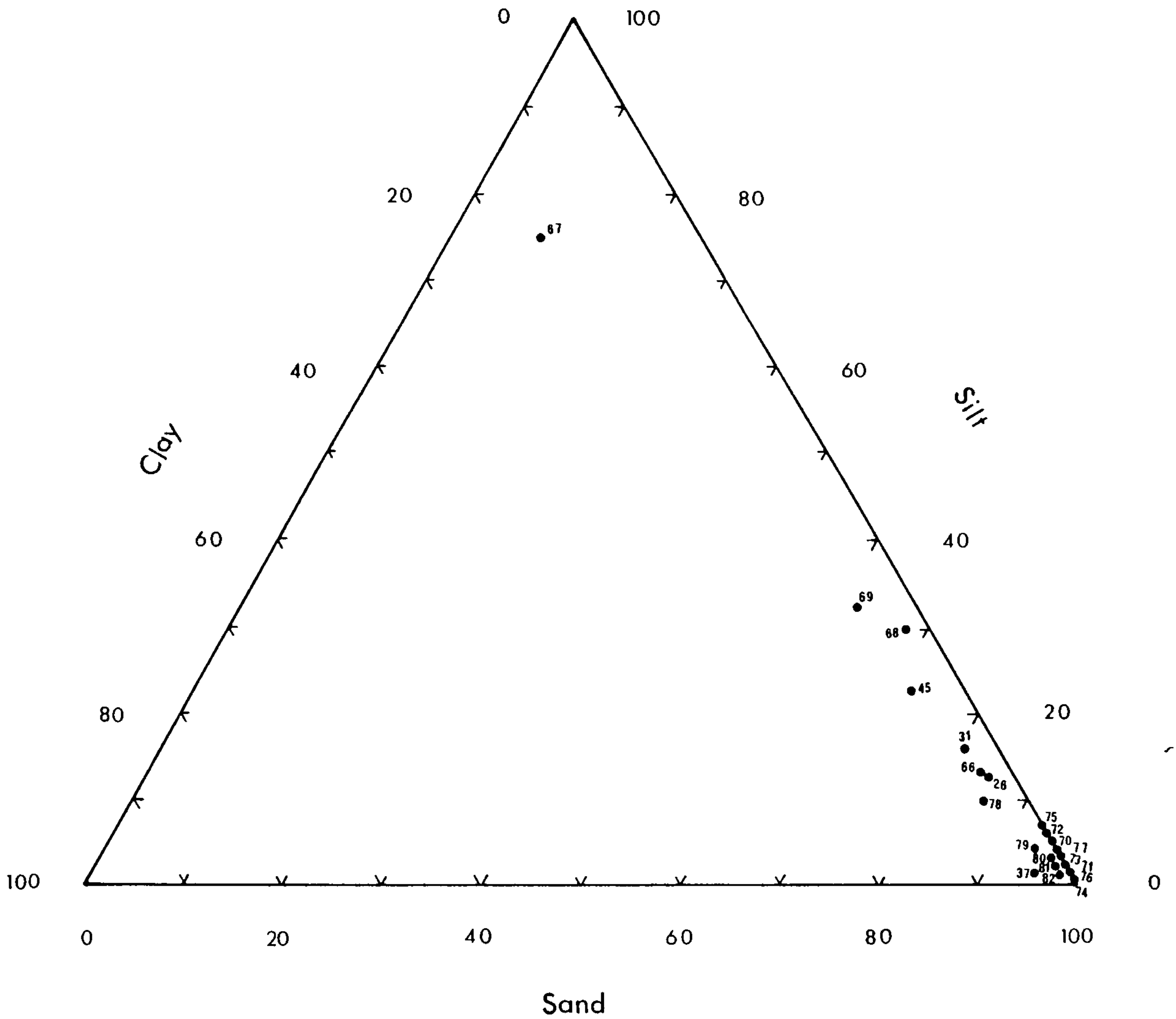


Figure 47

Ternary diagram of percent sand, silt and clay: upper Karewa sands.

The mean particle size of the modern fluvial sands shows considerable variation, ranging from about 1 phi to 6 phi(medium sands to medium silts). Sorting values range from about 0.5 to 3.5 phi (moderately well sorted to very poorly sorted). Skewness values lie between -0.25 and 0.7. Most of the samples are near-symmetrical. A few, however, are coarse-skewed (eg. 87, 188, 60) and some show fine-skew (23, 52, 5). Values of kurtosis range from under 1 to over 2 (mesokurtic to very leptokurtic).

The particle size envelope of upper Karewa sands (figure 41) shows a similar, though wider, spread in characteristsics to the modern fluvial sand samples also shown in this figure. The ternary diagram (figure 47) shows that there are three groupings of samples, together with one sample that does not fall into any group. Many of the samples fall in the bottom righthand corner of the ternary diagram, indicating that they contain a high percentage of sand. The second and third groups contain progressively less sand and more silt. The clay content for samples in all groups, however, is less than 5%. Sample 67 falls into none of these groups. Comparison with figure 35, the ternary diagram for upper Karewa fine-grained sediments, suggests that it should have been included in this category and was wrongly classified in hand specimen. However, no other samples were misclassified, and the sand and fine-grained sediments facies of the upper Karewa have two distinct and easily separable particle-size distributions on the basis of sand, silt and clay content.

The mean particle size of the upper Karewa sands (excluding sample 67) ranges from 0 to over 4 phi (coarse to fine sand). Many of the samples cluster around the 2 phi (medium sand) grade. Values of sorting range from 0.38 to 2.53 phi (well sorted to very poorly sorted) (see figure 13 and 14). Skewness values show a particularly wide spread. A small cluster of samples falls into the coarse-skewed category. A second, less-well defined cluster spans the near-symmetrical to strongly fine-skewed categories and a third cluster falls exclusively in the strongly fine-skewed category. Values of kurtosis range form under 1 to over 1.5 (mesokurtic to extremely leptokurtic).The wide range in the particle-size statistics is reflected by the lack of clustering on the bivariate scattergrams.

The particle-size variations in the Pattan profile are shown in figure 19. The mean particle size for all of the fine-grained samples ranges from 6.46 to 8.86 phi (medium to fine silt). Sorting values range from 1.48 to 3.42 phi (poorly to very poorly sorted, but with most samples falling in the former category). The mean particle size for the sand units ranges from 0.68 to 1.99 phi (medium sand) and the sorting from 1.06 to 1.13 phi (poorly sorted).

Interpretation of the particle-size data

Interpretation of these results requires some discussion of three points. Firstly, concerning the geological meaning of the particle size statistics. Secondly, concerning the influences on the particle size distribution in fluviolacustrine sediments. Thirdly, concerning the possible sources of sediment for the upper Karewa. However, particle size cannot, alone, provide a comprehensive diagnosis of sedimentary environment. This will be undertaken in section 5.2d, where all of the field and laboratory sedimentological data are synthesized.

Sahu (1964) considered the geological significance of the main particle size statistics and the following discussion draws on his work. The mean particle size is a measure of the 'average' particle size of the sediment. Ideally, then, it should indicate the average kinetic energy or velocity of the depositing agent. In reality, however, the mean particle size will also depend on the nature of the source material. The standard deviation provides a measure of sorting of the sediment. Sorting is regarded as an indicator of the fluctuations in kinetic energy, or velocity, of the depositing agent. However, the size distribution of the material will also affect the sorting of the sediment. Skewness provides a measure of the asymmetry of the particle-size distribution. It can also be regarded as the relationship between the mean and the median. The median itself, which is the central point of the frequency distribution, is not a useful summary statistic since it fails to account for skewness. Because of the nature of the phi-transformation used in particle size calculations, positively-skewed samples are skewed towards the fine tail of the distribution and negatively skewed ones to the coarse tail. Sahu (1964) suggested that coarse skewness in a sediment indicates that depositional velocity was greater for a longer period of time than average velocity, and/or velocity fluctuations occurred towards higher values more often than towards average values. Kurtosis is often regarded as a measure of the 'peakedness' of a frequency distribution (Pettijohn et al., 1972). However, Sahu (1964) argued that it should be regarded, more correctly, as a ratio of sorting in the central 90% of the distribution to sorting in the central 50%.

The shape of particle size curves itself has sometimes been used as a diagnostic tool (eg. Van Andel and Postma, 1954). However, classification of curve shapes has been largely subjective and problems arise when transitional forms are encountered. In this study, the particle size curves have been grouped to produce envelopes for each type of sample. Thus, the information provided by the individual curve is lost and the curves are presented simply as an illustration of the form and variation of the frequency distribution of samples within each group. Ternary diagrams provide a simple, but useful, illustration of sediment texture. They allow groupings within samples to be identified and they also highlight any anomalous samples within a group.

Sahu's (1964) account of the geological meaning of particle size statistics implies that there are other controls on the particle size distribution besides the kinetic energy of the depositing agent. Such controls were listed by Solohub and Klovan (1970) as: the geographical distribution of the source material; the mineralogy and texture of the source material; the rate of sediment supply; postdepositional sediment removal; sorting, mixing and remixing of separate populations; and diagenesis. Before examining the relative importance of these controls, it is important to discuss the possible sources of sediment for the upper Karewa lake. In view of the high energy environment in the mountainous areas surrounding Kashmir, it seems likely that large amounts of clastic material would have been deposited in the upper Karewa lake. There are five possible sources of clastic material.

1. Large, perennial rivers draining the Great Himalaya and Pir Panjal, including glacial outwash.
2. Gullies eroding uplifted and faulted lower Karewa strata in the Pir Panjal foothills along the western margin of the basin.
3. Mass-movement occuring along the steep mountain fronts adjacent to the lake, particularly along the Great Himalayan flank.
4. Wave erosion of sediment and bedrock along the lake shoreline.
5. Aeolian input from the weathering and erosion of uplifted lower Karewa beds, mass movement deposits and sources outside Kashmir Basin. This may have included salt weathering of rocks and sediments

Authigenic sediments may also make up a proportion of the upper Karewa sediments, particularly the fine fraction. In the upper Karewa lake environment, calcium-carbonate minerals are the most likely authigenic minerals to be present. Precipitation of calcium carbonate from solution may result either from physically induced chemical changes within the lake or from biogenic activity, although the latter is more common (Eugster and Kelts, 1983). Structural remains of organisms may contribute calcium carbonate to the sediments. In the case of the upper Karewa lacustrine deposits, ostracods and molluscs contribute calcite and aragonite respectively. In those parts of the basin adjacent to limestone bedrock, calcium carbonate may also be deposited as clastic material, due to the mechanical erosion of the limestone. Clearly, this is quite likely to have been an important source of calcium carbonate for the upper Karewa, since extensive outcrops of limestone are found in the Great Himalaya. Unfortunately, it is rarely possible to quantify the relative contributions of any of the above-mentioned sources.

The particle-size envelopes for the upper Karewa clayey silts and the loesses (figure 34) overlap.This lends some support to the field evidence suggesting that there was an input of aeolian dust into the upper Karewa lake. There are several possible sources of dust that could have supplied the lake. The first, and most obvious, is the large area of uplifted and eroded lower Karewa strata along the inferred south-west margin of the upper Karewa lake. Field observations showed that fine-grained lower Karewa sediments were intensively gullied prior to, and during, the deposition of the upper Karewa conglomerate. The widespread mobilisation and redeposition of this sediment would have provided large volumes of silt-sized material that could subsequently have been deflated. A second source of silt could have been the mass-movement deposits that extend from the Great Himalayan front. Today, fossilized screes can be seen in many localities along the mountain front. Interstitial fines in the screes could have been a source of silt. A third source is glacial outwash.These sources are all internal to the Kashmir basin. However, Kashmir is surrounded by areas which are currently arid and may also have been so during the period of upper Karewa deposition in Kashmir.

To the south-west, is the loess area of Potwar Plateau. Ladakh and Zanskar lie to the east and the Karakoram range to the north. However, it will probably never be possible to determine the relative contributions of internal and external sources to the loessic and lacustrine sediments of Kashmir since there are no known lithological, geochemical or mineralogical markers in the various source areas. However, given the extensive potential sources, there is no reason to expect that either the internal or external ones contributed all of the silt.

The bivariate scattergrams of mean versus sorting for the upper Karewa samples and the aeolian loess (figure 37) indicate that, although there is some overlap between the two groups of sediments, almost half of the upper Karewa samples are somewhat coarser than the loess. The larger mean particle size of the upper Karewa sediments suggests a greater velocity of the depositing agent compared to the loess. If a purely aeolian origin is proposed for the upper Karewa silts and clays, this implies higher windspeeds during the period of upper Karewa deposition and/or a change in the particle size distribution of the source material. However, it is highly unlikely that the upper Karewa lake was fed only by aeolian dust; the coarser samples may have been transported fluvially or simply reworked in the lake by wave action. Sorting values for the loess and upper Karewa samples partially overlap, although the upper Karewa samples tend to be better sorted. This may be a reflection of water sorting in the upper Karewa sediments. It could also be due to the fact that coarser sediments tend to be better sorted than finer ones. However, this is not supported by figure 37, which shows little relationship between sorting and mean particle size. An initially surprising result is that the upper Karewa samples tend to be more fine-skewed than the loesses, despite being coarser. An explanation of this may be found in the bivariate scattergrams that include kurtosis (eg. figures 38 and 40). These show that the loesses generally have kurtosis values less than 1 whereas the upper Karewa samples generally have values greater than 1 although there is some overlap. The loesses, therefore, have distributions that range from being mesokurtic to platykurtic whereas the upper Karewa sediments range from being mesokurtic to leptokurtic. A leptokurtic distribution suggests that the sediment is relatively better sorted in the central part of the distribution than in the tails. For a platykurtic distribution, the reverse is true (Folk and Ward, 1957). In a mesokurtic distibution, there is no difference between sorting in the centre of the distribution and that in the tails. On the basis of this argument, the loesses tend to show better sorting in the tails of the distribution than in the centre whereas for the upper Karewa samples, the reverse is true.

From the particle-size data presented in this study, it appears that an aeolian input into the upper Karewa lake was important. Variations between particle size statistics for true aeolian loess and the upper Karewa samples may relate to fluvial input, reworking in the lake and changes in the size distribution of the source material through time. Part of the silt fraction within the upper Karewa samples may be composed of authigenic calcite crystals. However, because authigenic calcite falls within the size-range of the clastic material, it cannot be identified by particle-size analysis. The analyses were undertaken on bulk samples, without the removal of the carbonate fraction. The reason for this was that a significant proportion of the clastic fraction of the sediment is likely to be calcium carbonate, derived from mechanical erosion of limestone. Furthermore, in none of the samples analysed did the presence of calcium carbonate prevent dispersal of the sediment.

The occurrence of sand beds within the upper Karewa formation suggests a change from aeolian-lacustrine deposition to fluvio-deltaic. However, sands may also represent beach environments. Some authors (eg. Friedman, 1961) suggest that dune, beach and fluvial environments can be distinguished on particle-size evidence alone. However, there is considerable doubt that this is universally true (eg. Solohub and Klovan, 1970). Unfortunately, it was not possible to test the discriminatory power of particle-size analysis in modern environments since many of those thought to be present in the upper Karewa formation do not have modern equivalents in Kashmir. However, the analysis of modern fluvial sands was undertaken as a partial solution to this problem.

The envelopes of particle size curves (figure 40) show that the variation in the distributions of the upper Karewa samples is greater than that for the modern fluvial sands. However, the ternary diagrams (figures 35 and 36) show that the scatter of samples is similar, particularly if samples 52, 65 (modern sands) and 67 (upper Karewa sand) are excluded from the plots. Samples 52 and 67 fall within the upper Karewa silt/clay envelope whereas sample 65 is transitional. If these three samples are excluded from the discussion, the bivariate scattergrams show that the upper Karewa sands and modern fluvial sands coincide on the basis of mean, sorting and skewness. However, a number of the upper Karewa sands are more leptokurtic than the modern fluvial sands. As outlined before, this shows a greater degree of sorting in the central part of the distribution compared with the tails.

From these data, it is difficult to state conclusively whether more than one depositional environment is represented by the upper Karewa sands. It is possible that all of the upper Karewa samples (except 67) come from environments that were similar to those of the modern perennial rivers in Kashmir. However, it may be that particle size analysis is an inadequate discriminator in this part of the study, and that other environments are represented as well. This problem is pursued later in this section when the laboratory data are considered alongside the field evidence for sedimentary environments.

<u>Clast lithologies from upper Karewa conglomerates</u>

Samples of clasts were taken for lithological counts at 7 sites. 4 of these were in sites in the upper Karewa conglomerate in the south-west part of the basin and 3 were in sites in the conglomerates of unknown stratigraphic origin in the Sind and Liddar Valleys. 10 samples, each of 50 clasts, were taken form each site. Each clast was assigned to one of four lithological groups, namely Panjal Trap, limestone, quartzite and 'other'. Although this lithological classification is grossly simplified, it avoids the error that would almost certainly have resulted form a more complex scheme. The most important lithology for this part of the study was limestone, since this rock is abundant on the Great Himalayán flank, but almost absent on the Pir Panjal flank. Thus, the presence or absence of limestone in conglomerates is potentially important in provenance studies. The value of limestone in this respect has already been demonstrated for the lower Karewa conglomerates (see section 4.6) suggesting that it may be useful for the upper Karewa as well.

The results of the clast counts are shown in table 5. Samples from Badgam, Karpora and Shupiyan all have similar lithological compositions. This is not surprising, since these three conglomerate units are regarded as equivalents of the upper Karewa conglomerate, the source of which lay in the Pir Panjal flank. The conglomerate at Malshahibagh had a small, but significant percentage of limestone clasts. Firstly, this effectively discounts Agrawal's (1985) suggestion that the Malshahibagh conglomerate is a distal remnant of the upper Karewa conglomerate in the south-west part of the basin. During the fieldwork for this study, it was initially thought that the Malshahibagh conglomerate was a distal remnant of the Sind Valley terraces. However, the clast samples form Nunar were completely devoid of limestone. A similar situation exists in the outer part of the Liddar Valley. A cemented conglomerete, containing limestone clasts, was observed at Anantnag (figure 16) by de Terra and Paterson (1939) although this was not sampled in this study. However, the conglomerates sampled at Kahalwan, in the mouth of the Liddar Valley, were found to be limestone-free (table 5). These results tend to

TABLE 5. CLAST LITHOLOGICAL DATA: UPPER KAREWA CONGLOMERATES.

SITE	PERCENTAGE				N
	Panjal Trap	Quartzite	Limestone	Other	
Ganderbal	45.2	3.4	19.0	32.4	500
Nunar	78.0	3.8	0.0	16.4	500
Romushi IV	86.8	5.2	0.0	8.0	500
Shupiyan	70.8	9.2	0.0	20.0	500
Karpora	82.4	4.8	0.0	12.8	500
Kahalwan	83.0	1.8	0.0	15.2	500
Badgam	80.6	2.6	0.0	16.8	500

Clast size used: 5-10cm

discount the hypothesis that the conglomerates at Malshahibagh and Anantnag are distal remnants of the terraces in the Sind and Liddar Valleys respectively.

In the Sind Valley, the terraces extend from Nunar to the mouth of Wangat Nala (plate 6). The terraces do not extend beyond this point, but are well developed in Wangat Nala itself. Excavations for a hydro-electric power project in the floor of the Sind opposite the mouth of Wangat Nala reveal at least 30 m of conglomerate extending below the current floodplain. This conglomerate is of similar appearance to the terrace conglomerate in the lower Sind and Wangat Nala, although it was not possible to take samples for lithological analysis. From the extent of the terraces in the outer Sind, it appears that the major source for this conglomerate was Wangat Nala. The catchment of Wangat Nala is completely limestone-free, in comparison to the upper reaches of the Sind Valley (see figure 3). The source of limestone for the conglomerete at Malshahibagh could either have been the upper reaches of the Sind Valley or a more localized source along the Himalayan front where limestone outcrops are common. In the Liddar Valley, the terraces are less well defined and it was impossssible to trace their extent.

Fieldwork for this research showed that the conglomerate at Malshahibagh was interbedded with upper Karewa lacustrine strata. Therefore, the two facies are coeval. The Sind Valley terraces and Malshahibagh conglomerates are therefore of different sources. It is suggested here that they are also of different ages, the terraces being older than the Malshahibagh conglomerate. The considerable age of the Sind Valley terraces is suggested by the thick loess cover on the surface. If this hypothesis is accepted, a change from a non-limestone to a limestone source area must have occurred during upper Karewa times. This may have occurred in several ways. Firstly, progressive erosion of calcareous bedrock may have occurred, eventually revealing limestone strata. However, physical tracing of the terraces leads to a source in Wangat Nala, where there is no limestone today. More convincing is the idea that the sources have changed geographically. If, by analogy with the Pir Panjal conglomerate, the terraces in the outer Sind Valley are interpreted as tectonic alluvial facies, their accumulation may have resulted from uplift in the Wangat Nala area. Because the conglomerate at Malshahibagh has a much more limited exposure, it was not possible to trace its source physically. Limestone occurs in the upper parts of the Sind and Liddar Valleys and clasts could have been derived from these areas either as a result of tectonic activity or glaciofluvial activity. Aternatively, the limestone may have been derived from outcrops within the basin itself. Unfortunately, the lithological data provide no further information to help answer this question.

Microfabric analysis

Microfabric samples were taken from upper Karewa clayey silt sediments exposed at Burzahom, Pampur and Wogahoma. Samples of loess were taken from the sequence overlying the Sind Valley terraces at Nunar. Field sampling involved the excavation of small (10x5x3 cm) oriented and levelled blocks from the exposure, following the removal of disturbed material from the section face. The blocks were allowed to dry at laboratory temperature of about 14^{0}C. Since the material will almost certainly have undergone numerous wetting and drying cycles when in situ, it is unlikely that the laboratory drying will have imparted artificial fabrics to the material (Derbyshire et al., 1987). Following drying, the blocks were subsampled and an observation face up to 1.5 cm^2 was obtained by natural breakage. This surface was then cleaned using a jet of compressed air and peeling with adhesive cellophane tape. The cleaning process was employed to remove particles that became dislodged during breakage. The sample was then mounted on an SEM stub, gold-coated and examined under SEM. Two machines were used; a Hitachi Stereoscan and Cambridge Instruments. Observation was carried out at 9 or 15 kV and magnifications ranging from 100 to 5000x. Electron Dispersive X-ray Analysis (EDAX), available on the Hitachi machine, was used for mineralogical determinations.

5 categories of features were examined under the SEM. These were the grains (size, shape, surface-features, mineralogy); cements, bridges and coats (structure, disposition, composition); aggregates (structure, composition); fabric (anisotropy, packing, voids); and biogenic features. Loesses from a number of localities have been subject to microfabric analysis (eg. Derbyshire, 1983a, 1983b; Derbyshire et al., 1987; Derbyshire and Mellors, in press; Billard et al., 1987). From these analyses, a number of generalizations concerning loess microfabrics can be made. Loess consists mainly of subrounded to subangular quartz silt. The mean size of the silt will depend partly on age and partly on the nature and location of the

source. Other minerals make up a smaller component of the loess. These include feldspars, micas, heavy minerals and clay minerals. The silt-sized particles make up the skeleton of the sediment. Fine-grained components occur in a number of ways. In true aeolian loess as opposed to reworked loess, silt-sized aggregates are often found. These may be composed of clay minerals, calcium carbonate or silica. They may have arisen by translocation within the loess profile or, particularly in the case of clays, the deflation of silt-sized clay minerals from pans and river floodplains. In true aeolian loess, the skeletal grains tend to show a preferred orientation and exhibit a random, 'face to face' fabric. Despite the fine-grained content of the sediment, the silt particles tend to be very clean, with little adhering material on the surfaces. Clays may also be found as buttresses and bridges between silt grains. Calcium carbonate and silica are also found in this disposition. This arrangement leads to an 'open fabric' with a high voids ratio (Derbyshire et al., 1987). It has been suggested (eg. Derbyshire, 1983a) that the compaction of loess in a dry state will lead to shearing of the silt-grain contacts and local disruption of the clay buttresses and bridges, but will not cause dispersion of the fines. In palaeosols, the silt fabric is largely masked by translocated clay and carbonate cement (Billard et al., 1987).

Loess which has been reworked by water following primary aeolian deposition will tend to have microfabrics that are distinct from those in true aeolian loess. This is because hydroconsolidation causes dipersion of the fines (Derbyshire, 1983a). A reworked loess will tend to have a lower voids ratio than a primary loess. Movement by water will tend to align the particles leading to preferred grain orientation. This is often detectable by visual observation of the sample under the SEM. Water action will also tend to break down the clay bridges and buttresses and the resultant clay is dipersed and will commonly adhere to the larger grains.

Microfabric analysis of the Sind Valley loess samples (Plate 14) shows that they are completely different to loessic samples found elsewhere in the world. The difference is so pronounced as to question a true loessic origin for the sample. The sediment consists of abundant clay-sized material and carbonate cement, which together obscure any skeletal grains. Even with such mantling, it is usually possible to see the outlines of skeletal grains. However, in this sample they are rare. Carbonate cementation has preserved abundant, fine organic remains, which are probably former rootlets. The lack of silt-sized material in this sample, together with the dispersed clay fabric, suggests that it is not a true loess. Alternatively, it may be a loess that has been subjected to intense pedogenesis. The latter hypothesis is supported by the presence of organic remains seen under the SEM. In either case, this sample does not provide a very useful basis with which to compare the upper Karewa samples. Therefore, samples of upper Karewa clayey silt were compared to aeolian and reworked loesses from elsewhere in the world, particularly China (Derbyshire et al., 1987).

The upper Karewa samples show a number of similar microfabric features. The samples have a pronounced silt skeleton.The size of the individual silt grains varies between samples, from fine to medium silt. The grains are typically very angular and show little edge modification. They are also clean, with few adhering clay particles. However, the actual amount of adhering material seems to depend on the clay content of the sample. For example, sample BZ-TL2 (plate 16) shows greater clay adherence to its silt grains than sample BZ-D2 (plate 15) and also has twice the clay content. This suggests that some reworking of clay aggregates may have taken place, either postdepositionally or following aeolian transport but prior to deposition. However, clay bridges, buttresses and coats are quite common in the samples, particularly sample BZ-TL2 where bioturbation is also in evidence. Much of the silt skeleton consists of quartz, although feldspars, micas and heavy minerals were also found. Mica varies in abundance from being a minor component (eg. BZ-TL2) to a major component (eg. sample BZ-D2: plate 15). Many of the features commonly destroyed by hydroconsolidation are found in the upper Karewa samples, particularly bridges and aggregates of clay. The fabric is open and the voids ratio high. Locally, individual grains or groups of grains were cemented with calcium carbonate. In the casts of fossil rootlets, the skeleton fabric has been altered and the grains wrap around the rootlet hole (plate 17). In all of the samples, there is evidence of anisotropy. In some samples, this consists simply of edge to face arrangement of the silt grains. However, in sample BZ-D2, there are well-defined microlaminae of alternate open and closed fabric.

The microfabrics of the upper Karewa samples show a number of features that liken them to true aeolian loess. These include angular, clean silt grains; clay-sized aggregates, clay bridges and buttresses; and calcium-carbonate cement. However, in contrast to primary loess, the samples show quite strong anisotropy. Thus, the grain fabric could reflect particle alignment due to deposition in water. However, reworking following deposition must have been quite limited, since the clay features, commonly destroyed by hydroconsolidation, are quite common in all samples. There is evidence of quite strong bioturbation of the sediments which has locally disturbed the fabric.

d. Sedimentary Environments of the Upper Karewa Formation.

In this section, the lithological and sedimentological data already discussed are used to reconstruct the sedimentary environments of the upper Karewa beds. In the field, individual units were assigned to facies. This classification was based initially on field evidence alone. In this section, the origin of the facies is discussed. Each type is not discuussed in isolation, however, since it is rarely possible to assign facies unambiguously to one single depositional environment. Therefore, facies assemblages are also considered. Also in this section, the occurrence and significance of calcium carbonate and iron oxide are discussed. The presence and interpretation of various structural features is examined. Finally in this section, this information is used to interpret the depositional environments of the main sections measured in this study.

The facies found within the upper Karewa beds are described in Table 6. The facies codes and interpretations are based partly on Miall (1978) and Rust (1978). However, since these authors were concerned primarily with fluvial deposition, some modifications have been included to allow for the classification of lacustrine and aeolian sediments. The classification distinguishes firstly between gravel, sand, 'fines' (silt and clay)and carbonates. Finer division within each group is made on the basis of sedimentary structures and the presence of organic remains that are observable in the field.

The upper Karewa conglomerates conform almost exclusively to the facies Gm (plates 6 and 7). They are either massive, or have crude bedding structures. The average clast size varies considerably, presumably depending on whether the exposure is in a proximal or distal position within the depositional system. Some exposures had maximum clast sizes around 15 cm.Elsewhere, maximum size was much smaller, typically only 5 cm. Visual comparison of the shape of over 2000 individual clast showed that they were predominantly subrounded to rounded.

According to the classification, facies Gm is a clast supported conglomerate. Although clast support was present in all of the upper Karewa exposures examined, the amount of matrix varies quite considerably. Conglomerates exposed at Nunar, in the Sind Valley, had a relatively high matrix content whereas those in distal positions of the Pir Panjal conglomerates, for example at Badgam, had a much lower content of matrix.

TABLE 6. SUMMARY OF FACIES PRESENT IN THE UPPER KAREWA FORMATION. Facies codes from Miall (1978) and Rust (1978) with relevant additions and amendments.

FACIES CODE	LITHOFACIES	SEDIMENTARY STRUCTURES	INTERPRETATION
Gm	massive or crudely bedded gravel	horizontal bedding, imbrication	longitudinal bars, lag deposits, sieve deposits
Sm	massive sand, medium to coarse	none	very rapid deposition from suspension
Sh	sand, fine to v. coarse	horizontal lamination, parting or streaming	planar bed flow (l.& u. flow regime)
Sp	sand, fine to v. coarse	planar crossbeds	linguoid transverse bars, sand waves (lower flow regime)
Sl	sand, fine	low angle crossbeds	scour fills, crevasse splays, antidunes
St	Sand, medium to coarse	trough crossbeds	dunes
Sr	sand, fine to coarse	all types of ripple marks	ripples (lower flow regime)
Fl	fine sand, silt, clay	fine lamination, ripples	overbank, waning flood or shallow lacustrine deposits
Fw	fine sand, silt, clay	wavy, lenticular and flaser bedding	levée and splay deposits
Fcf	silt, clay	massive or broadly banded	lacustrine deposits
Fsc	silt, clay	massive or broadly banded, molluscs, oxidised plant remains	lacustrine deposits
Fm	silt, clay	massive, fissured, carbonate on fissure surfaces	overbank deposits or deposits of drying lake
C	carbonate	strongly cemented nodular or tabular concretions	diagenetic or pedogenetic features
Fa	silt, clay	granular, pedogenic features, carbonate	aeolian loess

Facies Gm has been interpreted as a fluvial gravel or conglomerate (Miall, 1977). It forms a major component of the 'Scott-type' braided stream sequence. The facies is regarded as a deposit of longitudinal bar deposition (Miall, 1977, 1978). This environment contrasts with alluvial fans, which would tend to contain a high proportion of matrix-supported, debris-flow deposits, particularly in the proximal zone (Bull, 1972). However, alluvial fans tend to grade distally into braided streams, with a zone of transition between the two environments.

A variety of different sand facies is found within the upper Karewa. Medium to coarse grained massive (facies Sm) sand is quite common, usually as quite thin (<20 cm thick) beds. Massive deposits arise either due to structureless deposition, or because any structures that were formed on deposition have since been obliterated. If deposition itself involved the formation of massive beds, it is likely that this was a rapid process with no time for structures to form. A situation such as this may occur when a heavily sediment-laden current undergoes rapid deceleration. Grains arrive at the bed so quickly that they are buried and immobilised before further movement can take place (Collinson and Thompson, 1982). Destruction of preexisting bedding can occur in a number of ways. These include intense bioturbation and the mobilization of liquified beds. However, it is normally possible to recognise such changes in sediments. Since no obvious signs of reworking were seen in the massive upper Karewa sands, they are interpreted as primary massive beds resulting from rapid deceleration of sediment-laden waters. This could be explained by a sand-laden stream entering the upper Karewa lake. Singh (1982) has suggested that rivers entering the upper Karewa lake formed small Gilbert-type deltas. He argued that, because of the shallow depth of the upper Karewa lake, sand would have been deposited rapidly. Such conditions could have led to the formation of massive sand units.

Facies Sh , horizontally-bedded sand, is formed in specific flow conditions for particular particle sizes. For very fine sands, this implies flows greater than 60 cms^{-1} and for coarse sands, flows greater than 30 cm^{-1}. These are the upper and lower plane bed flow regimes respectively. In the former case, bedforms such as ripples and dunes are destroyed and turbulence is supressed. In the latter case, flow is greater than the critical erosion velocity, but less than that required for the formation of bedforms. Experimental work suggests that mica in sand can also explain the absence of ripples, since their formation and preservation depends partly on avalanching down the lee face of the ripple (Collinson and Thompson, 1982). Therefore, in mica-rich, horizontally laminated sands, assumptions about flow conditions cannot be made.

Facies Sp, planar cross-bedded, fine to very coarse sand, is also related to specific flow conditions. Planar cross-bedded sand results from the formation of sandwaves. These are formed by flow velocities that are slightly higher than those required to form ripples for any given particle size. Sandwaves are common bedforms in sand-bed rivers.

Facies Sl consists of fine sand with low-angle cross bedding. It is distinguished from facies Sp by the low angle of the foresets, which commonly dip at 5 to 10 degrees. Epsilon cross bedding also has low angle foresets. However, this structure , formed by the migration of point bars in streams and rivers, can usually be identified by the presence of smaller-scale internal structures which run parallel to the strike of the inclined beds (Collinson and Thompson, 1982). Reineck and Singh (1983) suggested that low angle cross bedding in fine sand can result from the migration of bars in a braided, sand-bed stream. Picard and High (1972) have suggested that this type of structure might be formed in shallow, ephemeral streams. Rust (1978), on the other hand, interpreted low-angle cross bedding as a result of high velocity, shallow flows into low relief scours and crevasse splays, and due to antidune migration. Facies St consists of trough cross-stratified medium to coarse sands, which result from the formation and migration of dunes.

Facies Sr consists of all types of ripple bedding in fine to coarse sand (eg. plate 11). They form under lower flow regime conditions when velocity exceeds the critcal for erosion, but is insufficient for dune and antidune formation. Subaqueous ripples may arise from unidirectional currents, oscillatory waves, or a combination of the two. Although ripples that are asymmetrical in cross-section are often regarded as products of unidirectional currents, and symmetrical ripples the product of waves, this is a simplification and cannot provide a basis for the diagnosis of origin (Collinson and Thompson, 1982; Reineck and Singh, 1983). The internal structure of ripples consists of various types of cross bedding and lamination. The nature of the internal structures reflects the type of ripple which, in ancient sediments, may not be preserved. Hence, the study of internal structures is of critical importance in diagnosing the flow conditions under which the bed was formed.

The terminology surrounding ripples and related structures is complex and confusing. Since there is no unambiguous link between process and form, any classification of structures must be either genetic or morphological. For the purposes of this discussion, a simple morphological classification is adopted. Following Jopling and Walkey (1968), cross lamination is defined as cross stratification in which the individual foreset laminae are less than 1 cm thick. Ripple cross lamination is the genetic term for cross lamination in which a definite ripple topography is seen. In some instances, a ripple may 'climb' up the stoss side of the ripple immediately downstream. This feature is termed ripple-drift cross lamination. In symmetrical ripples, laminae may be continuous or discontinuous across the ripple profile. Where lamination is continuous, the term sinusoidal ripple lamination is used (Jopling and Walker, 1968). Symmetrical wave ripples show a distinctive internal structure. The laminae are discontinuous across the ripple profile and may be stacked, chevron-fashion, in the ripple crests and troughs, or may show the development of unidirectional foresets, similar to asymmetrical ripples (Reineck and Singh, 1983). Wave-formed ripples may also climb if sufficient sediment is available to bury the ripple following formation.

From the above discussion, ripples can be divided into symmetrical and asymmetrical types. Within the symmetrical group, internal lamination may show chevron-structures, have unidirectional foreset laminae or show continuous lamination across the ripple profile. Asymmetrical ripples generally have a steep lee side and gentle stoss side. The main body of such a ripple, whether wave- or current-formed, generally consists of foreset laminae. Asymmetrical ripples may also climb. If the angle of climb is greater than the angle of the stoss-side slope, there is erosion between sets amd only the lee-side laminae are preserved. If the angle of climb is greater than the stoss-side angle, laminae willl be continuous from stoss to lee side. These types of ripple drift cross lamination are termed Type A and Type B respectively (Jopling and Walker, 1968).

Wave ripples arise from the oscillatory movement of water particles in waves. When the water velocity exceeds a critical value, particles begin to move and ripples eventually form. Wave action is analogous to repeated current action in opposite directions. If the velocity of motion is greater in one direction than the other, the ripples may be asymmetrical. Clearly, this renders the distinction between current- and wave-formed ripples on the basis of symmetry invalid. However, Reineck and Singh (1983) have argued that asymmetric ripples less than 4.5 mm in size are only known from wave environments.

Ripple-drift is a response to migration of ripples with a net supply of sediment. Jopling and Walker (1968) have shown that, for a non-cohesive bed, Type A ripple-drift cross lamination is a response to a low net sediment supply. In such a case, the ratio of suspended to bed-load is low, and burial insufficiently rapid for stoss-side preservation. The authors further suggested that an increase in the supply of suspended sediment would produce a transitional type and then, with a further increase in sediment, type B would form. In type B, the supply of suspended sediment is sufficient to bury, and hence preserve, the stoss side laminae. Jopling and Walker maintained that sinusoidal ripple lamination could be interpreted as a product of wave action. Hopwever, for their field example (a kame delta from the Pleistocene in North America), sinusoidal ripple lamination was more likely to have been the product of unidirectional currents since it is intimately interbedded with undoubted current, rather than wave structures and there is absence of independent evidence for wave activity in the former glacial lake. They suggested that the preservation of sinusoidal rather than asymmetrical form was due to a high suspended- to bed-load ratio, which would have caused rapid burial of ripples, together with a cohesive bed of fine material in which there was limited bed load transport where lee side faces do not develop and symmetrical rather than asymmetrical ripples are preserved.

Although ripples tend to be associated with sands, they also occur in finer sediments (plates 9, 10 and 12). This is true of both current- and wave-formed ripples. Facies Fl consists of fine to very fine sand, silt and clay in varying proportions, which may show ripple cross-lamination or fine parallel lamination. Parallel lamination may be highlighted by particle size variation within or between the individual laminae. This suggests that cyclic variations occurred in the input of sediment. Flat, parallel lamination suggests that erosion velocities were less than critical values for particle motion. In the case of lacustrine deposits, this may be because the lake was sufficiently deep for the bed to be below the influence of waves or simply because wind velocities were too low to generate waves of sufficiently high energy. Laminated fine sediments may also be found on flood plains, particularly as a response to waning flood conditions.

Cross lamination that arises from mixed sand and mud lithologies is varied in nature, but classified under the facies type Fw. The terminology of such features is discussed by Reineck and Wunderlich (1968). However, there are three basic tyes that can be identified: flaser bedding, lenticular bedding and wavy bedding. Flaser bedding consists of cross bedded or cross laminated sand with drapes of fine material frequently occuring in the ripple troughs, but sometimes also on the peaks. The mud drapes are preserved by the overlying sand bed. Lenticular bedding consists of sand ripples that are vertically and horizontally discontinuous. In this type of structure, the fine-grained sediment predominates. Wavy bedding consists of alternations of sand and mud, in which the mud overlies the ripple crest and more or less fills the troughs as well. Thus, both the sand and mud layers are laterally continuous. The sand layers in these three types of bedding are formed by wave or current action and show the internal structures which will reflect the mode of formation. The mud layers are deposited during periods of low velocity flow. Thus, the development of this type of bedding requires alternations of high and low energy conditions and a source of sediment with both sand and mud fractions.

Reineck and Wunderlich (1968) stated that structures of mixed sand and mud lithologies are characteristic of sub-and inter-tidal zones where the tidal rhythm generates the high and low velocity flows. However, this clearly does not apply to the upper Karewa sediments. Reineck and Singh (1980) quoted examples of flaser and lenticular bedding in lake sediments in front of advancing deltas. Burbank (1982) in his study of the lower Karewa sediments, suggested that flaser bedding could develop in levée and splay deposits, associated with river floodplains.

Facies Fcf consists of massive or broadly banded silt and clay (plate 8). A similar facies containing molluscs and oxidized plant remains is termed facies Fsc. Both facies are interpreted as lacustrine. Although these facies could also be overbank deposits of a meandering river, a lacustrine origin is thought more likely for two reasons. Firstly, the facies often contain freshwater, lacustrine fossils. Secondly, evidence for subaerial weathering and incipient pedogenesisis are rare. The oxidized plant remains do not appear to have caused extensive postdepositional disturbance to any bedding structures that are present and in agreement with Singh (1982 p.97), they are regarded as remains of subaqueous rather than terrestrial plants. Because facies Fcf and Fsw are horizontally banded and fine-grained, it is suggested that they are distal deposits of relatively deep water. Where banding is present, it is horizontally persistent.

Facies Fm consists of silt and clay which is intensively fissured but otherwise massive. Films of reprecipitated calcium carbonate frequently coat the fissure surfaces. Iron oxide staining is also quite common. This facies could be interepreted as an overbank deposit. However, because of its association with lacustrine sediments,it is interpreted here as a distal lake deposit that has undergone emergence and desiccation.

Facies C commonly occurs within facies Fcf or Fsc. It consists of strongly cemented, tabular or nodular, calcium carbonate concretions. The origin of calcium carbonate within the upper Karewa is discussed later in this sction.

Facies Fa is interpreted as aeolian loess following Pant _et_ _al_. (1978). It is granular, but otherwise structureless. It frequently shows signs of weathering and pedogenesis.

The two types of chemical sediment seen in the field in upper Karewa sections were calcium carbonate and iron oxide. Calcium carbonate is detectable where it occurs as nodular or tabular concretionary horizons. Frequently, these horizons stand out in the section since they are more resistent to erosion than sub- and superjacent beds.

There are four possible sources of calcium carbonate. These are detrital, chemical, biogenic and diagenetic. It is reasonable to expect that some of the calcium carbonate in the upper Karewa sediments is detrital, since limestone bedrock is quite common in Kashmir. Furthermore, the sections showing highest content of calcium carbonate are close to limestone outcrops. The chemical precipitation of calcium carbonate may occur as a responce to changes in the temperature and pH of lake waters. Primary calcite precipitation in lake waters is classically assumed to be of rhombohedral form. However, this is a misconception, and the form of precipitates may reflect other factors, including the rate of crystallization and the chemistry of the lake waters (Kelts and Hsü, 1978). The presence of primary carbonate precipitates is therefore not always obvious, particularly if the crystals are very small. Thus, although rhombohedral crystals were not seen in any upper Karewa samples, many of the small clumps of calcite, shown for example on the SEM micrographs, may be authigenic. Authigenic precipitation of calcite in the upper Karewa lake may have occurred as a result of localized photosynthetic activity or solution of limestone bedrock.

The biogenic component of the calcium carbonate in the upper Karewa lacustrine deposits consists of mollusc shells, ostracods and occasional charophyte oogonia. Numerically, the ostracods are most abundant. However, the molluscs are important sources of carbonate in the discrete horizons where they are found. Ostracods and charophytes are composed of calcite, wheraes molluscs are composed of aragonite. Early diagenetic carbonate is most clearly seen in the concretionary layers, which are quite common in the upper Karewa beds. Within any upper Karewa sample, it is difficult to assess the relative amounts of each calcium-carbonate source present. Therefore, the calcium carbonate curves for each section must be regarded as composite curves of varying, but unknown, amounts from each source.

The concretionary layers require further discussion. Nodular concretions are strongly cemented, discrete nodules that occur in laterally persistent horizons up to about 10 cm thick. Tabular concretions generally follow any preexisting bedding, without having caused any disruption. Their presence is sometimes only shown by preferential erosion of under- and over-lying, non-cemented beds.

Clearly, the two types of concretion have a different origin. However, they are interpreted here as of diagenetic, rather than pedogenetic origin. Singh (1982) has suggested that calcareous horizons in upper Karewa beds around Anantnag are pedogenic calcretes that formed under a warm, semi-arid climate. However, there is no evidence of pedogenesis in most of the sections studied here. In most instances, the concretions had not caused substantial disruption of the host beds. These observations suggest that they are better explained by diagenesis. The occurrence of tabular concretions is probably related to variations in the porosity of the beds, since they are generally found above relatively clay-rich units. These units would have impeded the downward percolation of water and provided sites for precipitation of carbonates. Nodular concretions may have formed due to the precipitation of carbonates around a nucleus. In some cases, this was confirmed by the occurrence of a mollusc shell within the nodule. In neither case, however, is the pedogenic hypothesis of concretion formation supported.

Concretions and stains of iron (III) oxide are quite common in the upper Karewa lacustrine sediments and indicated by distinct red colouration. According to Pye (1983), red soils and sediments are distinguished by having Munsell hues redder than 5YR, values in the range 4 to 7 and chromas between 4 and 8. However, the colour of a sediment containing iron (III) oxide depends not only on the presence of the ferric mineral, but also on the precise mineralogy of the iron compound, and the presence of matrix and organic matter.

According to Pye (1983) red sediments or 'red beds' may be _in situ_ or detrital. The latter arise from erosion, transport and deposition of preexisting red sediments. Since there is no source of preexisting red beds in Kashmir, the red colouration in the upper Karewa sediments is unlikely to be detrital. _In situ_ red beds may be pure chemical precipitates, or may form due to diagenesis or pedogenesis. Pure chemical precipitates of iron are rare in Quaternary sediments, and are not thought to be present in the upper Karewa beds. The most likely explanation of red colouration in the upper Karewa is either pedogenesis or diagenesis.

Subaerial pedogenesis is thought to be unlikely in most instances in the upper Karewa sediments, since unambiguous evidence for pedogenesis is generally lacking. There is rare evidence that pedogenesis may have occurred in upper Karewa sections and in such cases, any associated red colouration may be pedogenetic. However, in the majority of cases seen in the field, the red colouration is better explained by diagenesis.

According to Pye (1983), a simple model of iron oxide formation in sediments invloves four stages. These can be outlined as follows.

1. Iron-bearing minerals in sediments are weathered.
2. weathering leads to the formation of iron in solution.
3. The dissolved iron is reprecipitated to form colloidal ferric oxide.
4. The colloidal ferric hydroxide is <u>either</u> recrystallized to geothite, which is then dehydrated to haematite <u>or</u> dehydrated internally to haematite.

Whether iron is preserved in the oxidised or reduced state depends on the redox potential(Eh) and the hydrogen-ion concentration (pH) of the interstitial waters. Oxidation at depth in these sediments is favoured by free drainage. Under such conditions, iron (III) oxide will form except at very low pH. Under less strongly oxidising conditions. Iron (III) oxide may be stable at higher pH values (Pye, 1983). The abundance of calcium carbonate in the upper Karewa sediments, together with the excellent preservation of ostracod shells, suggests that the upper Karewa lake waters were alkaline. Thus, iron (III) oxide would have been preserved, even under less than maximum oxidising conditions.

In the upper Karewa sediments, red colouration is associated with the margins of laminae and beds, with fossil plant remains and with coarse-grained sand beds. Bed and laminae boundaries seem to have been optimum sites for the precipitation of iron within a unit. This may have been due to small changes in particle size, and hence permeabiltiy, within a unit. In the case of organic remains, the fossil casts seem to have acted as nucleii, around which precipitation occurred. In the sand units, oxidizing conditions would have been enhanced by the relatively free movement of interstitial waters.

Individual facies cannot usually be attributed unequivocally to a single sedimentary environment. Facies are predominantly the response of a particular type of sediment to a set of energy conditions and, as such, may occur in more than one environment. In order to interpret a sedimentary sequence, it is important to consider the association of facies present. Walther's law of the superposition of facies states that environments found adjacent in modern situations will be next to each other in the stratigraphic record, in the absence of unconformities. This can be used to help interpret the origin of a unit on the basis of the underlying and overlying sediments, as well as the nature of the unit itself. As a prelude to the interpretation of each measured upper Karewa section, facies models of each environment thought to be present in the upper Karewa are discussed. These models are based heavily on work from similar sedimentary environments elsewhere in the world. They include gravel-dominated braided river, sand-dominated braided river, delta and lake. The facies associated with each of these environments are listed in table

7. The facies listed are those from the upper Karewa sequences. Similar environments from other sedimentary sequences may have slightly different facies assemblages for one reason or another. As Miall (1978) warns, such facies models are not a 'universal panacea' for interpreting depositional sequences and facies models should be 'designed in a flexible way to accommodate...improvements.'

Gravel-dominated Braided River

Several reviews of the braided river depositional environment have proposed facies classifications and vertical profile models (eg. Miall, 1977, 1978). Of the six vertical profile models proposed by Miall (1978), four refer to predominantly gravel-bed rivers. Of these, the most relevant to the upper Karewa conglomerates is the 'Scott-type' model. This consists mainly of facies Gm, massive or crudely-bedded gravel, with minor trough and planar cross-bedded gravel. Sand and fines, showing a range of structures, are also present. This contrasts with the 'Trollheim type' model, which includes debris-flow deposits and the 'Donjek type' model which contains a substantial proportion of sand in the vertical sequence and thus less gravel. The Trollheim profile is the product of proximal rivers, mainly alluvial fans with debris flow and braided stream activity. The Scott profile is also a proximal river sequence. However, this consists mainly of braided river gravels and distal alluvial fan deposits. Debris flow sediments are thus rare. The Donjek profile is a distal cyclic sequence of sands and gravels.

TABLE 7. FACIES ASSOCIATED WITH DEPOSITIONAL ENVIRONMENTS IN THE UPPER KAREWA FORMATION.

ENVIRONMENT	MAJOR FACIES	MINOR FACIES
Braided River (Gravel dominated)	Gm	Gp, Sp, St, Sl
Braided River (Sand dominated)	Sh, Sl	Sp, Sr
Delta	Sh, Sp, Fw	St, Fl
Lake	Fcf, Fsc, Fm, Fl	C

There is, however, considerable variability in the upper Karewa conglomerates. This is best illustrated in sections in the Pir Panjal conglomerates. The steady transition, from proximal to distal sequences is apparent by comparing sections close to the mountain front, for example at Shupiyan, to those nearer the centre of the basin, such as at Karpora. In the proximal sequence, the conglomerate is very crudely bedded, relatively matrix rich and contains large, subangular clasts. In distal sequences, the conglomerate shows better defined bedding features. It has a lower proportion of matrix and the clastsare smaller and more rounded. This transition was also noted by Bhatt (1982b) and Singh (1982).

Sand-dominated Braided River

Miall (1978) proposed three vertical profile models for sandy braided alluvium. The 'South Saskatchewan' model is a cyclic, sandy deposit, the 'Platte model' a virtually non-cyclic sandy deposit and the 'Bijou Creek' model a deposit characterized by ephemeral rivers. or perennial rivers subject to flash floods.

The South Saskatchewan sequence is dominated by trough cross-stratified sand, formed by the migration of dunes under the lower flow regime. Other sand facies and fine-grained sediments are minor components of this model. The Platte model is dominated by planar and trough cross-stratified sands. Miall (1978) suggested that this sequence may have arisen from deposition in large bars and sand waves. The Bijou Creek model is characterized by sands which show low angle cross bedding and planar lamination. Both of these facies indicate high energy conditions. Ripple-bedded and planar cross-bedded sands also occur in this sequence.

Delta

The existence of extensive deltaic deposits has been noted both in the upper and lower Karewa beds (Burbank, 1982; Singh,1982; Tandon et al., 1982) although the type of delta present in the lower Karewa differs from that found in the upper Karewa. Singh (1982) argued that small 'Gilbert-type' deltas occurred in the upper Karewa, whereas those in the lower Karewa were birdsfoot deltas with well-developed interchannel swamps. Much of the literature on deltas is concerned with large features in coastal environments. However, small Gilbert deltas described in glaciolacustrine systems provide a useful model for the interpretation of the upper Karewa sediments.

Studies of the sedimentology of glaciolacustrine deltas and associated sediments have been undertaken by a number of authors including Jopling and Walker (1978), Ashley (1975), and Gustavson et al. (1975). The basic model of the glaciolacustrine delta proposed in these studies conforms to the classsic model of delta sedimentology in having bottomsets, foresets and topsets. In a delta from the Pleistocene of Massachusetts, Jopling and Walker (1968) found that bottomsets and foresets consisted of ripple-drift cross-laminated sand, although the actual distinction between bottomsets and foresets was unclear since the sequence of deposits was more or less continuous. Topsets were represented by cross-bedded, pebbly sands. In the distal prodelta zone, horizontally laminated clays and silts were found. Jopling and Walker (1968) argued that the actual type of ripple-drift cross lamination present in the sequence would depend on the ratio of suspended to bed load in the channel, a fact that has already been mentioned in this thesis in the discussion of individual facies.

Without considering the details of the individual sections at this stage, it is clear that this model is not wholly applicable to the upper Karewa sediments, since ripple-drift cross-lamination is rare in the sequences. However, Singh (1982) proposed an alternative model for the evolution of the upper Karewa deltas. This stresses the shallow depth of the lake and the resulting wave influence on the sediments. According to Singh's model, the topset beds deposited in the delta channnels consist of current-bedded, medium-grained sand with evidence of modification by wave action. The delta foresets consist of wave ripple-bedded and ripple-laminated fine sands and silts, in contrast to the glaciolacustrine model. The bottomsets consists of fine silts and clays which are frequently rhythmitic and often show ripple lamination together with wavy, lenticular and flaser bedding except in deeper water sequences. Although ripple-drift was observed in this study, it was not common. According to this model, the upper Karewa deltaic environment is characterized by planar and cross-bedded sand facies; wavy, lenticular and flaser bedding with sand and mud; and rare ripple-drift cross-lamination.

Lake

Deposits interpreted as lacustrine are intimately associated with deltaic sediments and there is a blurred transition between delta bottomsets and proximal lacustrine sediments. On the basis of wave-ripple structures in the sediments, Singh (1982) argued that the upper Karewa lake was only several metres deep. Only in restricted areas and for relatively short time intervals was the lake deep enough for the bed to be below the zone affected by waves.

Structural features in the upper Karewa formation

The structural features found in the upper Karewa sediments fall into three categories, which are faults, folds and gullies. These features are largely confined to the upper Karewa lacustrine strata. Structural disruption of the upper Karewa formation is much less common and less dramatic than in the lower Karewa formation and the features listed above were only found in a few sections. For the most part, the upper Karewa strata are horizontally bedded and undisturbed.

Faults are found in only a few areas. The best examples occur in Badgam Road Quarry 6 (plate 13). Here, small down-faulted blocks are bounded usually by normal faults with throws of less than 1 m. Since this faulting affects all of the beds in this section, it must have been postdepositonal. It may have resulted from localized tectonic activity on the area, or from the slumping and settling of unstable sediment nearby.

Folding and contorting of beds has also occurred in the beds exposed in Badgam road Quarries. The folding clearly occurred prior to the deposition of the loess and the overlying fractured mud, since only the underlying strata have been affected. As with the faults, the folding may have been cause by instability in the sediments themselves or localized tectonic activity.

There is also evidence for gully formation in the top of the upper Karewa lacustrine beds. In the central part of the Kashmir Basin, particularly in the Badgam Quarry and Humhama exposures, but also in the Sambur area, gullies up to 10 m deep and 20 m wide have been cut into the upper Karewa strata. They have been infilled with loess. The soil sequence within the gully fill is the same as that in the adjacent loess, indicating that the gullying occurred after upper Karewa deposition, but before the onset of loessic sedimentation.

The final part of this section consists of an interpretation of the sedimentary environments present in each measured section. This is concluded with some general points concerning the upper Karewa depositional environments.

<u>Pattan</u>

The lowermost 3 metres, comprising units 1 to 4, are interpreted as deposits of a shallow lake. Ripple-laminated fine sand and silt suggest wave or current action. The small unit of poorly sorted, gravelly sand (unit 2) could be the result of deposition on a high-energy beach or more likely, since it is structureless, due to rapid deposition from a small, possibly ephemeral stream. Units 5 and 6 are interpreted as deltaic. Unit 5 represents the delta foresets and unit 6 the topsets. Unit 5 has ripple lamination, which could be the result of wave or current action, interbbeded with flaser and lenticular bedding. Units 5 and 6 conform to the model of delta formation suggested for the upper Karewa by Singh (1982).

Unit 7 is interpreted as delta bottomsets and proximal lacustrine facies. Although there are no obvious foresets in this part of the sequence, they must be present if Walther's law is to be satisfied. However, distinction between the deltaic environments is often difficult, as noted by Jopling and Walker (1968). Units 8 to 14 are also interpreted as deltaic deposits, similar in structure to units 5 and 6, but more complex. In common with the lower delta, this sequence shows ripple bedding that could represent the action of waves in a shallow lake. Units 15 to 18 mark a return to lacustrine conditions. The change in sedimentary structures, from ripple bedded to parallel laminated to massive and then back to parallel laminated suggests that the lake may have undergone a transgressive - regressive cycle. The transition between the faintly parallel-laminated silty clay and the aeolian loess occurs over 5 cm of stratigraphic thickness. The boundary is conformable, showing no evidence of any break in sedimentation. This suggests that the end of the upper Karewa lake occurred quite rapidly.

The Pattan section has generally low values of calcium carbonate. The low values of calcium carbonate in the sand units probably reflects their source, which would have been in the Pir Panjal Range. The calcium carbonate content of the fine sediment may reflect input of fine clastic material from suspension or authigenic precipitation. A comparison of the calcium carbonate curve with the ostracod concentration curve shown in figure 18 later in this chapter suggests that variations in ostracod content may be a major control on calcium carbonate content.

The particle size characteristics of the Pattan sediments are remarkably consistent, the only significant variations being between the fine units and the sand units. The mean particle size of the fine-grained sediments is 6.91 phi and the values range from 6.46 to 8.86 phi. Values of sorting range from 1.48 to 3.42 phi, with a mean sorting value of 2.68 phi. The sand units are typically better sorted than the fine units. The relative consistency of the particle size characteristics of the fine units suggests that the mode of sedimentation was quite constant. This probably reflects the fact that aeolian input was occuring throughout the time-period represented by the section. Small variations in the particle-size characteristics can be explained by inputs from fluvio-deltaic systems migrating periodically into the lake and reworking of sediment on the lake bed, possibly by wave action.

A further feature of the Pattan section is the persistence of lacustrine strata. Compared to the other sections measured, the Pattan section shows a greater proportion of lacustrine sediments and fewer sand units. This is best explained by the position of the Pattan section within the former upper Karewa lake (figure 16). It is located quite centrally, particularly when compared to the other sections. Furthermore, it is not located close to the mouths of any major valleys that would have entered the lake.

Burzahom

The sequence of depositional environments at Burzahom is similar to that at Pattan. The lowest 1 units are interpreted as deposits of a small, Gilbert-type delta into the upper Karewa lake. This is followed by a change to deeper-water conditions shown by unit 5a. This environment is continued above the hiatus. A further phase of delta activity is represented by units 3b to 9b. This is followed by deeper water lacustrine conditions up to unit 13b. As in the Pattan section, the change from lacustrine conditions to aeolian deposition occurs over a short thickness of sediment, suggesting that the end of the upper Karewa lake occurred quite rapidly in this section.

The lowest 3 m in the Burzahom section have a high calcium carbonate content. In contrast to the Pattan section, the sand units also have very high contents of calcium carbonate. This is due to the proximity of limestone to Burzahom indicating that the carbonate is detrital. The upper part of the section, above the hiatus, has low calcium carbonate contents, which fall to zero above unit 6b. Unit 6b itself is a horizon of laterally continuous nodular concretions associated with an impervious clay-rich layer below. Since the concretions do not disrupt the host bedding, they are interpreted as diagenetic. It is likely that calcium-carbonate saturated water from overlying units percolated down through the section as far as the impervious clay horizon, above which precipitation took place. The variable calcium carbonate content of the loess is a reflection of carbonate mobilisation and reprecipitation during pedogenesis.

Badgam

The lowermost 10 m of this section are interpreted as lacustrine sediments. The darkish blue, broadly-banded sediments in unit 1 are probably deepwater deposits formed in a reducing environment. By contrast, the paler overlying beds such as those in units and 5 and 6 are interpreted as products of shallower water. Above unit 10, there is a change to sand and conglomerate deposition. There are two possible interpretations of this sequence. Under the first interpretation, units 11 and 12 are regarded as delta topsets, representing the migration of a delta into the lake. This was followed by a hiatus during which erosion of underlying units took place and then conglomerate deposition on the underlying sediments. Under the second interpretation, units 11, 12 and 13 are regarded as integral parts of a single braided river system, which was deposited following an erosional hiatus after the lacustrine phase of units 1 to 10. The sedimentary structures present in the sequence suggest that the second interpretation is more likely to be correct. This also accords well with observations elsewhere in the Kashmir Basin, further towards the Pir Panjal margin. In the Kanchi Kol Nala, for example, fine-grained strata beneath the upper Karewa conglomerate had clearly undergone intense erosion and folding prior to the conglomerate deposition. This suggests that a considerable time interval separated the end of lacustrine sedimentation and the onset of conglomerate deposition. At Badgam the sediments beneath the upper Karewa conglomerate have not been folded. This is explained by the fact that the Badgam site is some distance form the Pir Panjal flank, which was uplifted at the beginning of upper Karewa deposition.

Units 14 to 17 mark the end of conglomerate deposition and the beginning of fine-grained sedimentation. This sequence is interpreted as lacustrine. Unit 18, a well sorted sand, may have been deposited by a small, high-energy stream entering the lake which was, at this stage, very shallow. The loess overlies this sand unit. The onset of loess deposition would have occurred once the fluvial sand became stable.

The part of the section below the conglomerate bed has a very low calcium carbonate content. The sand unit and the conglomerate itself are carbonate-free, reflecting the fact that the source area of these sediments was in the Pir Panjal Range.

According to the 'accepted' model of Karewa sedimentation discussed in chapter 3 of this thesis, units 1 to 10 of the Badgam section would be regarded as lower Karewa and units 11 to 18 as upper Karewa. The Badgam section provides an excellent illustration of the change in the nature of lacustrine deposition below and above the upper Karewa conglomerate, as suggested by Bhatt (1982a) and Singh (1982). The reduction in size of the Karewa lake during upper Karewa times, suggested by Bhatt and Singh's model, is well illustrated by the change from deepwater lacustrine facies to braided stream and then shallow-water lacustrine facies shown in the Badgam section. It may also be reflected by the calcium carbonate curve for this section. If the calcium carbonate in the sediments is authigenic then the increase in the carbonate content in the upper part of the section compared to the lower part, may be due to the reduction of lake-volume and, hence, saturation of the lake water.

Badgam Quarries

The stratigraphic sequence observed in these quarries has important bearings on the overall palaeogeography of Kashmir during the upper Karewa period. The lowermost sediments exposed in each quarry are typically shallow-water lacustrine beds. In quarries 1 to 4, these are overlain by facies Fm, fractured silty clay, with abundant calcium carbonate and iron oxide staining. Facies Fm is overlain by aeolian loess. The presence of facies Fm suggests deposition under increasingly low energy conditions together with desiccation or draining out of the lake. This is in contrast to many of the other upper Karewa sections, where the end of the upper Karewa lake appears to have been rapid.

In quarries 5 and 6, the distal thinning of the upper Karewa conglomerate was seen. In quarry 5, the conglomerate is replaced by a coarse, gravelly sand. The conglomerate and sand are inetrepreted here as braided-straem deposits belonging to the same depositional system. It is suggested here that the conglomerate and sand are distal remnants of the conglomerate exposed in the Badgam section. The sand and conglomerate units are overlain by facies Fm in quarries 5 and 6.

The sequences described in the Badgam sections and the Badgam quarries suggests the following sequence of events. Initially, the lower Karewa lake was relatively shallow. During this period, shallow-water lacustrine facies were deposited in the central portion of the basin, as represented by sections in quarries 1 to 4. Towards the western margin of the basin, braided stream activity led to the formation of conglomerate proximally, as shown in the Badgam section and quarry 6, and coarse sand distally, as indicated by quarry 5. This was followed by a period of lake transgression, so that shallow-water lacustrine facies were deposited over the conglomerate in the Badgam section. This was followed by a third phase, during which the lake level fell. In the Badgam quarries 1 to 4, a remnant of the lake seems to have remained and undergone slow drying, prior to the onset of loess deposition.

Before the deposition of loess began, the upper Karewa beds underwent structural disruption. The faulting, found in quarry 6, cuts through the shallow lacustrine deposits, the conglomerate and the overlying blocky mud. However, where the folding is seen in the other quarries, it has not affected the blocky mud. This suggests that, if the blocky muds in each section are stratigraphically equivalent, the faulting postdates the folding by an undetermined period. The surface of the blocky mud has been quite intensively eroded in certain areas. These gullies have then been filled with loess. The phase of gullying occurred after the deposition of the mud and before the onset of loess deposition. Overall, this period seems to have been one of enhanced geomorphological activity in parts of Kashmir Valley.

Wogahoma

The Wogahoma section shows a relatively simple sequence. The lower three units consist of relatively deep-water lacustrine deposits. Unit 4 shows some evidence of limited pedogenic activity, with progressive darkening in the unit and an upwards increase in organic content and bioturbation. The unit is massive, but in the uppermost part shows blocky fracturing similar to that found in the upper part of the Badgam quarry sections. Unit 5 at Wogahoma is a 2.5 metre-thick bed of sand and very fine gravel, showing low angle cross bedding. This is interpreted as a fluvial sand from a high energy, possibly ephemeral, river. The upper part of the sand unit fines upwards very rapidly, showing an increasing admixture of loessic silt in unit 6. Unit 7 is essentially a Bt horizon, showing leached calcium carbonate and argillans coating granular peds.

The calcium-carbonate content of the Wogahoma section is quite high. Unit 2 consists of a horizon of nodular concretions, interpreted as diagenetic. The fluctuating calcium-carbonate curve for much of the section presumably reflects detrital input, since Wogahoma is located close to limestone outcrops. This is supported by the high values of calcium carbonate found in the sand. The absence of carbonate from unit 4 is probably due to leaching during pedogenesis.

Pehru

This section is located on the inferred margins of the upper Karewa lake. This location explains the high proportion of sand units in the section. The four sand units present are interpreted as deposits of small Gilbert deltas that migrated into the lake. Units 1 to 3 represent the development of the first delta, with unit 4 indicating the return to deepwater conditions. A similar situation is seen in connection with sand bed units 5, 7 and 9. However, the sequences associated with the upper two sands are less characteristically deltaic than those lower in the section. Unit 10, above the hiatus, shows the return to deepwater conditions. This is followed by a relatively abrupt transition to the loess.

Sambur

The Sambur section is about the same thickness as that at Pattan. However,it shows a contrasting sedimentary sequence. The Sambur section has a large proportion of sand-dominated units. Pronounced cycles of delta-building have already been noted in the Sambur Plateau area by Singh (1982). Units 1 to 11 in the measured section consist of current- or wave-rippled sediment ranging from fine sand to silty clay. This part of the section is interpreted as proximal lacustrine deposits, prodelta deposits, and, possibly, delta foresets. Units 12 to 14 represent two phases of delta migration. Both of the sand units have a range of particle sizes and unit 14 contains some fine gravel. The coarse texture of these units suggests a high-energy environment, presumably proximal delta topsets or a distal sandy braided stream environment.

Unit 15 marks a return to deeper-water conditions, particularly in the lower parts of the unit. The upper parts of unit 15 shows lenticular bedding and is overlain by the horizontally-bedded sand of unit 16. Horizontal bedding in coarse sand and very fine gravel indicates an upper plane bed flow-regime, which involves a high velocity fluvial environment. Units 17 to 19 show a return to deeper water conditions. Unit 19 is a thin unit of blocky mud, which grades into aeolian loess.

The Sambur section has quite low values of calcium content. Reconnaissance sample-preparation for the study of ostracods revealed few valves in this section. This probably reflects the high energy fluvial environments common in this section. Such environments may have been unfavourable for ostracods and for the preservation of valves.

Sambur Village

This short section consists of a thin conglomerate resting on Triassic limestone. The conglomerate hes been interpreted as a mixed beach and scree deposit (de Terra and Paterson, 1939; Singh, 1982). This interpretation is supported by the mix of angular and rounded clasts found in this study. The occurrence of a rich fossil assemblage within the conglomerate was originally noted by de Terra and Paterson (1939). Kotlia (1984) reported the presence of broken limb bones of Elephas hysudricus, together with jaws, teeth and spines from fish (subfamilies Schizothoracinae and Cyprininae) in this bed. The conglomerate is overlain by fine-grained lacustrine deposits.

Pampur

This section shows the lake-delta sequence found in many other sections. The lowest three units in the section are mainly shallow lacustrine deposits, although the first two units may mark the end of a delta-building phase. Two other pronounced delta sequences are seen. One of these is between 1 and 7 m up the section, the other between 10 and 11 m. They are separated by a massive mud interpreted as a deeper-water lacustrine sediment. The uppermost lacustrine unit at Pampur is a blocky mud similar to that found in the Badgam Quarry and Sambur sections. The mud grades fairly abruptly into aeolian loess. The calcium-carbonate curve in the Pampur section shows fairly low values.

Nunar-Malshahibagh

The terrace conglomerates at Nunar consist largely of facies Gm, with occasional sand-dominated facies such as Sp and St. The presence of these facies suggests that the conglomerate was deposited by a large braided-river complex that existed in the Sind Valley. The subrounded clasts, the large clast size, the relative abundance of matrix and poor organization of the conglomerate indicate a high sediment to water ratio and suggest a proximal location.

The cemented conglomerate at Malsahibagh contains small, rounded clasts and rather less matrix. Poor exposures prevented detailed examination of the conglomerate over a wide area. However, the available information suggests a more distal location than the conglomerate at Nunar.

Karpora

the Karpora section is similar to that at Badgam, in that the upper Karewa conglomerate overlies lacustrine strata. However, in the Karpora section, truncation of the lacustrine strata is more pronounced since they have been subjected to more intense folding and faulting. This is because Karpora is closer to the Pir Panjal flank than Badgam.

The lower part of the Karpoa section is relatively straightforward. It consists of lacustrine units representing lakes of varying depth. The conglomerate is interpreted as a braided stream conglomerate. Unlike at Bagam, the loess overlies the conglomerate directly, with no intervening lacustrine strata. This is because Karpora lay beyond the inferred margin of the upper Karewa lake.

Although no laboratory calcimteric analyses were made, simple field tests showed no calcium carbonate in any of the units.

Summary

The sedimentary environments of the upper Karewa formation can be summarized as follows.

1. The sedimentary framework proposed here consists of braided stream, delta, lake and aeolian environments. This agreees broadly with that proposed by Singh (1982).
2. Deltas and sandy braided rivers dominated marginal areas, whereas in the centre of the basin, the lacustrine facies are more persistent.

3. The upper part of the sequence suggests that the upper Karewa lake underwent a quite rapid demise. However, marsh conditions, represented by the blocky mud facies, persisted in the central part of the basin. The marshes presumably occurred in topographic lows in the basin.

4. The upper Karewa lake was generally a shallow water body. Singh (1982) suggested that the facies generally indicate a lake shallower than 2 m, which was, very exceptionally, more than 10 m deep.

5. Much of the silt within the uper Karewa lacustrine strata had an aeolian source. This is suggested by particle size analysis and confirmed by SEM microfabric studies.

6. The calcium carbonate may be detrital, authigenic, biogenic or diagenetic. There is little evidence in many units to indicate the relative importance of each source. Hence, the calcium carbonate curves must be regarded as a complex function of variation in each of the sources.

7. Prior to the onset of loess deposition in the centre part of the basin, the upper Karewa sediments underwent structural disruption and localized gullying: this may have been caused by small-scale, localized tectonic activity.

5.3 Ostracod Fauna of the Upper Karewa formation.

a. Methods.

82 samples, each of about 300 to 500 g, were taken from the Pattan section for the analysis of ostracod faunas. 5cm-thick samples were taken at 25 cm intervals. The Pattan section was chosen for sampling since it is one of the longest sections measured in this study. Furthermore, it has a large proportion of lacustrine strata in the vertical sequence. Reconnaissance studies indicated that most of the sequence contains abundant ostracod valves. This is in contrast to other long sections, such as the one at Sambur, which is dominated by sand units, few of which contain ostracods.

In the laboratory, 200 g of each sample was dispersed in deionised water, typically overnight. The samples from the Pattan section were generally unconsolidated and needed neither chemical nor mechanical dispersion. The dispersed samples were washed through a 63 micron sieve, and the residue collected and dried at 30^{0}C. Preliminary scanning under a low-power binocular microscope was than carried out, in order to determine the abundance of ostracod material within the sample. Because of the wide variation in abundance between samples, 200 g of sediment sometimes contained a massive number of valves. In such cases, the sample was split using a mechanical splitter. The splitting factor was then used to correct the final species counts to values that would have been obtained had the full 200 g been used.

Every ostracod carapace, valve and fragment was then picked from each sample. In order to facilitate picking, the dry residue of the sample was first sieved through 355, 250 and 180 micron meshes and each subsample was picked in turn. This made picking much easier by restricting the size-range of material being picked at any one time. The material from each size range was than amalgamated prior to taxonomic sorting. The sub-180 micron fraction was not picked, since it only contains juveniles, which cannot usually be identified to specific level.

Wherever possible, individuals were identified to species level. In some samples, numerous small fragments were present that were too small to identify. In such cases, the approximate number of fragments was noted and no identification attempted. The ostracod material was affixed to a Franke slide using a weak solution of gum tragacanth.

Chemical analyses were carried out on water samples from Nagin and Manasbal Lakes. Attempts were made to collect living ostracods by dredging amongst aquatic vegetation and in the lake substrate, but no individuals were found. However, some information on the modern ostracod fauna of these lakes is available from Singh (1974) and this is summarized in appendix 8 of this thesis.

b. Systematic Description of Taxa.

In the systematic descriptions that follow, the species are described in the taxonomic order adopted by Hartmann and Puri (1974). Additional species encountered in the Pattan section which were not described by these authors are included at the appropriate point in the taxonomic scheme. Data on the material are taken from this study. Information on ecology, geographical distribution and stratigraphical range comes form a number of studies, which are referenced, where appropriate, in the text. In the descriptions that follow, and also on the plates, the abbreviations C = carapace, V = valve, R = right, L = left, F = fragment are used.

Kingdom ANIMALIA

Superphylum ARTHROPODA

Phylum CRUSTACEA

Subclass OSTRACODA Latreille, 1806

Order PODOCOPIDA Müller, 1894

Suborder PODOCOPINA Sars, 1866

Superfamily CYPRIDACEA Baird, 1845

Family CYPRIDIDAE Baird, 1845

Subfamily CYPRIDINAE Baird, 1845

Genus EUCYPRIS Våvra, 1891

Eucypris sp.

Plate: 18 (1)

Material: 20 V.

Remarks: 4 species of Eucypris have already been reported from the upper Karewa (Singh, 1974). These are E. awantipurensis, E. sp. cf E. clavata. E. sp. cf E. moguntiensis, and E. zenkeri. However, the present species conforms to none of the above.

Ecology: since the present species is unnamed and because it is rare in the Pattan section, little comment can be made about its ecology. The Eucyrprids already found in the upper Karewa are found in sluggishly flowing waters, ponds and lakes, with rich vegetation (Singh, 1974).

Genus PARASTENOCYPRIS Hartmann, 1964

Parastenocypris delormei Singh, 1974

Plate: 18 (2)

1974 Parastenocypris delormei Singh, pp.107-108

Material: 555 V, 1 C

Ecology: Found, by Singh (1974)in abundance, in a shallow, cemented channel of cold, permanently flowing spring-water. Associated with Candona neglecta, Cyclocypris laevis, Cypria opthalmica, and Ilyocypris bradyi. The depth of the water was 0.60 m, the temperature at time of colllection (May) 12^{o}C, pH was 7.50 and total dissolved solids content (TDS) 220 ppm (Singh, 1974).

Geographical distribution: Quaternary, Kashmir.

Stratigraphical range: Quaternary.

Subfamily CYPRIDOPSINAE Kaufman, 1900

Genus CYPRIDOPSIS Brady, 1868

Cypridopsis cf C. aculeata (Liljeborg, 1853)

Plate: 18 (7)

1853 Cypris aculeata Liljeborg, p.117

Material: 28 V

Ecology: according to Singh (1977a), this species occurs in mildly brackish, even oligohaline to mesohaline, waters.It is commonly found in shallow pools and marshes. Although a good swimmer, it favours bottom sediments.

Geographical Distribution: Kashmir and the central Plains of North America (Singh, 1977a; Delorme, 1970).

Stratigraphical range: Quaternary.

Cypridopsis vidua (Müller, 1776)

Plate 18 (4)

1776 Cypris vidua Müller, p.198

Material: 21 V

Ecology: known from a variety of habitats. Often found in permanent, still waters with rich vegetation. In lakes, it is usually found in the shallow marginal zone. It is an active swimmer, but prefers muddy substrates. Also found in running waters, especially with rich vegetation (Singh, 1977a). Generally prefers warm waters and although characteristic of freshwater, it can tolerate sub-brackish to brackish conditions (Singh, 1977a; DeDeckker, 1979).

Geographical distribution: Kashmir (Singh, 1974) and Europe (DeDeckker, 1979). The recent distribution is cosmopolitan.

Stratigraphical range: Late Pliocene to recent (Swain, 1963; DeDeckker, 1979).

Genus POTAMOCYPRIS Brady, 1870

Potamocypris sp.

Plate: 20 (3)

Material: 58 V

Remarks: various species and subspecies of Potamocypris have been reported from the upper Karewa (see Singh, 1974, and appendix 8 of this thesis). However, the present species did not conform to any of these, possibly becasue the individuals are juveniles. Those species reported from Kashmir tend to inhabit cool, shallow water.

Subfamily CANDONINAE Daday, 1900

Genus CANDONA Baird, 1845

Candona candida (Müller, 1776)

Plate: 20 (4 & 5)

1776 Cypris candida Müller, p.198

Material: 64 V

Ecology: a typical cold-water form, resistant to slight changes in temperature. Occurs in most types of water body and springs. May even be found in saline water. Frequently found in marshy vegetation (DeDeckker, 1979).

Geographical distribution: North America, all holarctic regions Scotland, Scandinavia.

Stratigraphical Range: Quaternary.

Candona neglecta Sars, 1887

Plate: 20 (1 & 2)

1887 Candona neglecta Sars, p.107

Material: 264 V, 8 C

Ecology: currently found in all types of water body, often in marshy vegetation. It has been recorded in waters with temperatures between 5 and 8^{0}C (DeDeckker, 1979). Like most other members of the genus, it is a burrowing form and is often found amongst lake-bottom detritus

Preece *et* *al*., 1986).

Remarks: juveniles of *C*. *neglecta* have sometimes been regarded as a separate species under the name *C*. *lactea* (Preece *et* *al*., 1986).

Geographical distribution: Europe, Central Asia, North Africa.

Stratigraphical Range: Quaternary.

Candona *spp*. juveniles

Plate: 18 (9)

Material: 8849 V, 36 C

A large number of juveniles and fragments were found in the Pattan section. However, they could not be identified to specific level. Their association with *Candona* *candida* and *Candona* *neglecta* suggests that they may be juveniles of this species. Other species of *Candona* have been reported from the upper Karewa (Singh, 1974; appendix 8, this thesis) and some of the specimens encountered may belong to these groups. However, species of *Candona* are particularly difficult to identify, and it was preferred to leave much of this group at the generic level.

Family ILYOCYPRIDIDAE Kaufmann, 1900

Subfamily ILYOCYPRIDINAE Kaufmann, 1900

Genus ILYOCYPRIS Brady and Norman, 1889

Ilyocypris *bradyi* Sars, 1890

Plate: 19 (3 & 4)

1890 *Ilyocypris* *bradyi* sars, p.59

Material: 6938 V, 39 C

Ecology: usually regarded as a species typical of vegetation-rich, unning waters (Staplin, 1963; Delorme, 1970; Singh, 1977; DeDeckker, 979). It is also found in other shallow water-bodies including pools, onds, ditches, springs, marshes and rice-fields. It is less common in akes, except where the lake is sufficiently shallow for the bed to be iffected by wave action. It can tolerate quite high salinities, but only slight changes in temperature (DeDeckker, 1979). It is a sluggish swimmer (Robinson, 1980).

Geographical distribution: holarctic regions, North America, North-west Europe.

Stratigraphical range: Quaternary.

Ilyocypris gibba (Ramdohr, 1808)

Plate: 19 (1 & 2)

1808 Cypris gibba Ramdohr, p.91

Material: 728 V, 2 C

Ecology: found in small water-bodies that are not subject to desiccation. Also found in quiet flowing waters with temperatures between 4 and 19°C (DeDeckker, 1979).

Geographical Dsitribution: Europe and North America.

Stratigraphical range: Middle Oligocene to recent.

Ilyocypris kashmirensis Bhatia, 1968

Plate: 19 (5)

1968 Ilyocypris kashmirensis Bhatia, p.476

Material: 43 V

Ecology: uncertain, but found with I. bradyi and I. gibba.

Geographical distribution: Quaternary.

Stratigraphical Range: has only been found in the Kashmir Quaternary.

Superfamily DARWINULACEA Brady and Norman, 1889

Family DARWINULIDAE Brady and Norman, 1889

Genus DARWINULA Brady and Roberston, 1885

Darwinula stevonsoni (Brady and Robertson, 1870)

Plate: 19 (8 & 9)

1870 Polycheles stevonsoni Brady and Robertson, p.25

Material: 998 V, 15 C

Ecology: found in a variety of habitats, although occurs chiefly in permanent lakes (Singh, 1977). It is a non-swimmer, and burrows in the upper surface of lake floors, particularly amongst vegetation detritus (Robison, 1980). Can tolerate TDS levels up to 175 ppm (Delorme, 1964).

Geographical distribution: cosmopolitan.

Stratigraphical Range: Quaternary.

Superfamily CYTHERACEA Baird, 1850

Family CYTHERIDEIDAE Sars, 1925

Subfamily NEOCYTHERIDEIDINAE Puri, 1957

Genus CYTHERISSA Sars, 1928

Cytherissa lacustris (Sars, 1863)

Plate: 18 (6)

1863 Cythere lacustris Sars, p.222

Material: 30 V

Ecology: common in deep (>3 m deep) freshwater lakes (Delorme, 1970).

It con tolerate moderately saline environments and has, more rarely, been found in the littoral zone (Staplin, 1963). Sylvester-Bradley (in Shotton and Osborne, 1965) has suggested that the species prefers relatively cold water.

Geographical distribution: North America, Northern Europe.

Stratigraphical range: Quaternary.

Family LIMNOCYTHERIDAE Klie, 1938

Subfamily LIMNOCYTHERINAE Klie, 1938

GENUS LIMNOCYTHERE Brady, 1868

Limnocythere _franki_ (Bhatia, 1968)

Plate: 19 (6 & 7)

1968 _Limnocythere_ _staplini_ Bhatia, pp.478-479

Material: 1215V, 4 C

Ecology: uncertain, but occurs in the Quaternary of Kashmir in association with _I_. _bradyi_, _C_. _neglecta_, _C_. _candida_ and _D_. _stevonsoni_.

Geographical distribution: Kashmir.

Stratigraphical distribution: Quaternary.

Family CYCLOCYPRIDIDAE Kaufmann, 1900

Genus CYCLOCYPRIS Brady and Norman, 1889

Cyclocypris _spp_.

Plate: 18 (8)

Material: 1209 V

Remarks: This group is thought to comprise _C_. _laevis_ (Müller, 1776) and _C_. _ovum_ (Jurine, 1820), although individuals of each species were not counted.

Ecology: both C. laevis and C. ovum occupy a broad range of habitats, although they are most common in small water bodies, particularly those with marshy vegetation (DeDeckker, 1979). Both species can withstand changes in water temperature and salinity.

Geographical distribution: holarctic regions.

Stratigraphical range: Miocene (C. laevis) and at least Pleistocene (C. ovum) to recent.

Genus CYPRIA Zenker, 1854

Cypria opthalmica (Jurine, 1820)

Plate: 18 (3)

1820 Monoculus opthalmicus Jurine, p.178

Material: 33 V

Ecology: occurs in a wide variety of freshwater environments including ditches, pools, ponds and shallow lakes (Singh, 1977). It is a good swimmer, but seems to prefer water bodies with well-vegetated beds. It can also tolerate mildly saline waters (Singh, 1977).

Geographical distribution: North America, Northern Europe, North India.

Stratigraphical range: Quaternary.

c. Results.

The modern lakes in Kashmir valley, such as Dal Lake, Wular Lake, Nagin Lake and Manasbal Lake are presumably remnants of the upper

Karewa lake. Water samples analyzed from two of these lakes, Manasbal and Nagin (table 8) show that they are alkaline, with low salinities. The modern ostracod fauna of the lakes is characterized by species that are poor swimmers which prefer to crawl on vegetation and burrow in muddy substrates. These include Candona spp., Ilyocypris spp., Darwinula stevonsoni, and Potamocypris spp. However, some swimming species, such as Cyclocypris spp. and Cypria opthalmica, were also found (singh, 1974: see appendix 8). The occurrence of these taxa reflects the fact that the modern lakes in Kashmir are shallow, with muddy substrates and abundant aquatic macrophytes, such as Elodea.

The abundances of the ostracod taxa identified in the Pattan profile, calculated as the number of valves per 200 g of dry sediment, were plotted as a function of depth (figure 48). The actual numbers of valves counted, together with the splitting factors used in each sample, are listed in appendix 8.

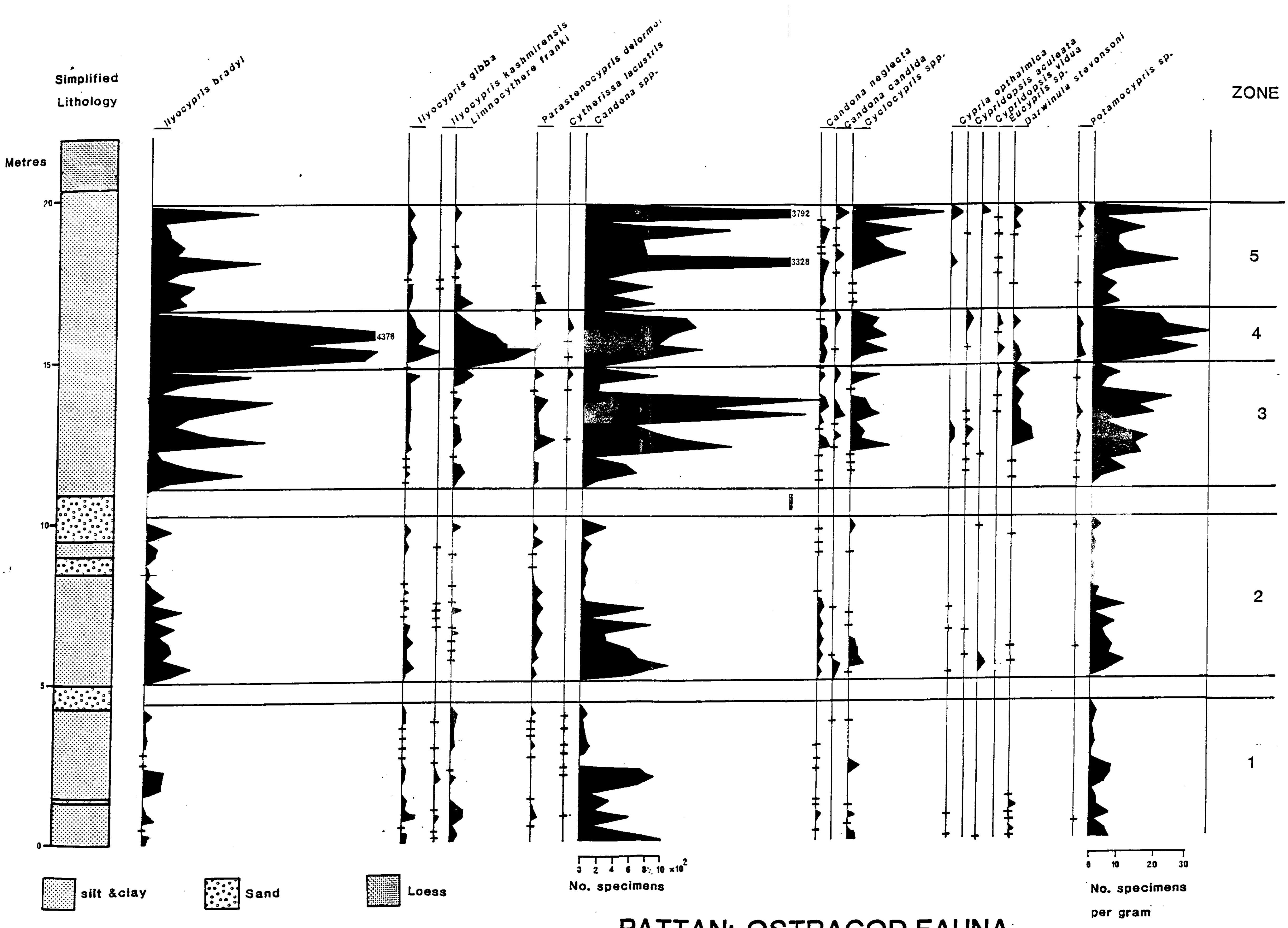
In the case of some groups, identification was only to generic level. In the case of Candona spp., many of the individuals found were juveniles. Juveniles of Candona are notoriously difficult to identify to species level. Although some palaeoecological information is lost using this procedure, this is preferable to the errors of interpretation that could arise through mis-identification.

The species-abundance curves on figure 48 show an upward increase in species diversity and the total number of valves. The number of valves per gram is particularly high in the upper 9 metres of the

TABLE 8. PARTIAL CHEMICAL ANALYSIS OF MODERN LAKE WATERS

SAMPLE	Ca^{2+} (ppm)	Mg^{2+} (ppm)	Sr^{2+} (ppm)	K^{+} (ppm)	Na^{+} (ppm)	pH	Alkalinity	Conductivity	Total Hardness
1	4.93	7.0	0.05	1.03	6.44	7.2	70.5	152.9	98.5
2	3.22	8.7	0.06	0.93	6.56	7.6	76.0	155.8	89.5
3	2.47	10.3	0.06	0.55	7.28	7.8	68.5	135.5	82.0
4	4.04	12.2	0.07	1.05	6.48	7.5	95.5	169.3	99.0
5	4.55	12.6	0.07	1.00	6.24	7.3	89.0	171.3	100.5
6	4.96	12.8	0.07	0.99	6.24	7.2	89.0	171.7	102.5
7	3.50	11.7	0.04	0.88	6.84	8.1	64.5	140.6	77.0

Samples 1 - 3 from Nagin Lake
Samples 4 - 7 from Manasbal Lake



+ = less than 5 specimens

Figure 48. Pattan: ostracod fauna.

section, although large fluctuations are also apparent. The main components of most of the samples are I. bradyi and Candona spp. The abundances of valves in these groups seem to covary, although the maximum peaks of one group are generally associated with slightly reduced peaks of the other. Curves for I. gibba, Limnocythere franki, and Cyclocypris spp. also follow the flucatuations in I. bradyi although these species make up a much lower percentage of the total fauna in any one sample. The candonids identified to specific level occur in small numbers, although they are more abundant in the upper part of the section. Most of the other species occur only sporadically in the section, but are more common in the upper 9 metres than in the lower part.

The ostracod diagram has been divided into 5 zones, based on the species present and their relative and absolute abundance. The 5 zones are delimited by barren horizons which occur at 4 to 5 m, 10 to 11 m, and 16.5 m up the profile.

Zone 1 is characterized by low abundance of ostracods, particularly in the upper half of the zone. The concentration of valves does not exceed 7 valves g^{-1} and in the upper part of the zone, falls below 0.5 valves g^{-1}. In the lower part of the zone, the assemblage is dominated by Candona spp., with small numbers of Limnocythere franki, Parastenocypris delormei, Cyclocypris spp., and Darwinula stevonsoni. In the upper part of the zone, I. bradyi, L. franki and Candona spp. occur in approximately equal numbers. Species diversity is high at the base and top of this zone, but declines dramatically in the centre.

Zone 2 has a similar ostracod assemblage to zone 1, although the concentration of valves is slightly higher. The zone is dominated by Candona spp., although large numbers of I. bradyi also occur. Species diversity is highly variable in this zone.

Zone 3 shows a sharp increase in the number of ostracods, with concentrations exceeding 25 valves g^{-1} in the upper parts. Dominance alternates between I. bradyi and Candona spp. Although the peaks in the taxa tend to coincide within the zone, large peaks of one taxon tend to be associated with reduced peaks of the other. Zone 3 also shows an increase in species diversity, including the appearance of a number of swimming taxa. In particular, Cyclocypris spp. is quite abundant towards the top of the zone. Darwinula stevonsoni is also moderately abundant, but makes up only a small proportion of the total fauna. The barren zone separating zones 3 and 4 is not accompanied by any lithological change in the sediment.

Zone 4 shows a similar assemblage of species to zone 3. The zone is dominated mainly by I. bradyi, although large numbers of Candona spp. and L. franki also occur.

Zone 5 marks the return to high, but fluctuating numbers of ostracods. Candona spp. is dominant over I. bradyi throughout this zone, and Parastenocypris delormei is absent. Swimming species are moderately abundant. Darwinula stevonsoni ocurs only sporadically.

d. Interpretation.

In the palaeoenvironmental interpretation of any fossil assemblage, it is vital to distinguish between autocthonous and allochthonous components, since interpretations based on the latter may be inaccurate. Kashmir, which is surrounded by high mountains, will have a wide range of very different habitats that may have been sources of ostracod valves for the lake basin. If palaeoenvironmental interpretations are based on allochthonous assemblages, they may simply reflect variations in source area rather than environmental changes in the lake itself. However, for ostracods, the distinction between autochthonous and allochthonous components is relatively straightforward. A method based on the age structure of ostracod assemblages has been described by Whatley (1983). He reported studies of the age structure of recent and Quaternary ostracod assemblages. In recent, low-energy environments, faunas were characterized by a population age-structure denoted type A. This consisted of adults of both sexes and juveniles of various moult stages. It is unlikely that any physical process could amalgamate all of the moult stages, since they have such different hydrodynamic properties. The presence of such an assemblage in the fossil record suggests that the sample is autochthonous. In recent, high-energy environments, faunas were characterized by a Type B structure, consisting of adults and larger juveniles, the earlier moult-stages having been removed by current action. Type B assemblages are rather more difficult to interpret, since they could represent autochthonous or allochthounous communities. Type C assemblages consisted only of small juveniles, a clear indication of allochthonous components in the fossil record.

In the Pattan samples, most samples were of age-structure type A, suggesting that they are autochthonous. This interpretation is consistent with the predominantly fine-grained character of the Pattan sediments, which suggests a low-energy environment. However, in samples above and below sand units, smaller juveniles are absent, suggesting that the fossil assemblages may contain allochthonous components. This must be considered in the palaeoenvironmental interpretation of the Pattan record.

Important palaeoenvironmental information can sometimes be inferred from the ratio of valves to carapaces in the ostracod assemblage. Disarticulation of carapaces is usually a response to high-energy conditions. In quiet depositional conditions, articulation tends to be preserved. When moulting, however, ostracods almost always disarticulate their carapaces. The presence of a large number of juvenile carapaces is often taken to indicate high infant mortality (Whateley, 1983). Such a situation suggests rapid onset of unfavourable environmental conditions. Articulation of adult ostracods is a better indicator of energy conditions. However, articulation may be preserved by rapid sedimentation, since the carapaces are buried before disarticulation can take place. In the Pattan samples, very few of the ostracods were articulated. This suggests low sedimentation and/or a high-energy environment. There are no constraints on sedimentation rate for the Pattan section but the proposal of a high-energy environment is at variance with the age-structure of the ostracods. However, it is possible that wind action in a shallow lake would have been sufficient to disarticulate valves lying on the lake bed, but insufficient to transport valves any great distance.

Several general interpretations can be drawn from the Pattan ostracod assemblages. The presence of I. bradyi suggests running or moving water. For the Pattan section, this is interpreted as indicating a shallow lake, the bed of which was affected by wave action. Many of the other species present indicate a shallow lake environment. These include Limnocythere franki, Parastenocypris delormi, I. gibba, Darwinula stevonsoni and Potamocypris spp. The presence of Cytherissa lacustris, however, suggests rather deeper-water conditions.

Most of the species referred to above are either non-swimmers or poor swimmers. Many of the members of the genus Candona are burrowers, found in the upper layers of vegetal detritus on the lake bed. I. bradyi and I. gibba are sluggish swimmers and, like Candona spp. , tend to be found in vegetation-rich environments. Limnocythere franki has only been found in Quaternary deposits in Kashmir. Hence, little is known about its ecology. However, by analogy with other members of the genus, it probably inhabited the upper layers of organic muds on lake beds. Darwinula stevonsoni is also a non-swimmer and tends to burrow in vegetation-rich detritus on lake beds. The presence of all of the above-mentioned species suggests that the upper Karewa lake was rich in vegetation with an organic-rich bed. In addition, there was some aspect of the lake that favoured non-swimming taxa. Although a few swimming taxa were found (eg. Cypridopsis spp., Cypria opthalmica), these were rare. There are several possible explanations for this. Firstly, it is possible that swimming species were selectively eaten by fish. Benthic forms would tend to be camouflaged in a vegetation-rich lake (R. Whatley, personal communication). However, there is no evidence that this occurred in the upper Karewa lake. Indeed, in the Pattan section, no fish bones were found, suggesting that the hypothesis is unlikely. A second explanation for the lack of swimmers is that swimmers tend to prefer a low-energy environment (R. Whatley, personal communication). In the case of the upper Karewa lake, the lack of swimmers suggests that the water was shallow, and under the influence of waves. The second of these hypotheses is the most consistent both with the ostracod evidence and that from sedimentology.

The species composition suggests that cool-water conditions predominated in the upper Karewa lake. Both Candona candida and C. neglecta are cool-water species. For example, C. neglecta has been found in waters of between 5 and 8^{0}C (DeDeckker, 1979). I. bradyi and I. gibba are also cool-water species: I. bradyi cannot tolerate wide variations in temperature, but I. gibba has been found in waters with temperatures between 4 and 19^{0}C. Parastenocypris depormmei was found in modern environments in Kashmir, in water of 12^{0}C (Singh, 1978) suggesting that it is also a cool-water form. However, its occurrence is limited to Kashmir. Cytherissa lacustris has been reported fom cool waters in northern Canada, in areas that experience a mean annual air-temperature regime of -10.1 to +9.5^{0}C.

The ostracod assemblages also place some constraints on the palaeochemistry of the upper Karewa lake. Although all of the species found are characteristic of freshwater environments, some, notably Candona candida, I. bradyi, Cytherissa lacustris, Cyclocypris spp., Cypria opthalmica, and Cypridopsis vidua can tolerate varying degrees of salinity. However, the presence of Darwinula stevonsoni suggests low salinity, since it has only been found in waters with up to 175 ppm TDS (Delorme, 1964).

In general, the Pattan ostracod sequence represents a cool, shallow lake, the bed of which was under the influence of wave or current action and supported abundant vegetation. This conclusion is supported by the more general work on the upper Karewa ostracod fauna by Singh (1978) and also by the sedimentological data presented in this thesis. The ecological interpretations of each of the ostracod zones are considered below.

Zone 1 is dominated by benthic forms. Swimming species are nearly absent. This suggests that the lake was sufficiently shallow for movement of the water at the bed by the wind. This conclusion is supported by the sedimentological data, which suggest a shallow-water lacustrine environment for the lower part of zone 1. Conditions do not seem to have favoured the occurrence of ostracods in the upper part of zone 1. The sedimentological data indicate that this zone was characterized by proximal lacustrine - prodelta deposits grading upwards into delta bottomsets, foresets and topsets. This indicates a progressive change to a high-energy environment, which might not have the lake must have been of low salinity, which is consistent with it being deeper, and therefore more dilute.

The barren zone at the top of zone 3 suggests that there was a rapid change to unfavourable environmental conditions in the lake. However, this change was relatively short-lived. This is followed by zone 4, which is dominated by I. bradyi, Candona spp. and L. franki. The postulated environment is similar to that for zone 3.

The barren zone separating zones 3 and 4 represents a dramatic, but short-lived environmental change to conditions unfavourable for ostracods. The dominanace of Candona spp. and I. bradyi suggests a vegetation-rich lake, However, swimming species are moderately abundant, suggesting that wave or current action was limited. Darwinula stevonsoni occurs only sporadically. This may mark an increase in salinity.

At the top of zone 5, there is a sharp change from high ostracod abundance to a barren zone. This is almost coincident with the change in lithology from lacustrine to loessic sediments, which presumably accompanied the drainage or desiccation of the upper Kaewa lake. The ostracod record suggests that this was an abrupt event, since there is no progressive change in the faunal assemblages below.

5.4 The Chronology of the Upper Karewa Formation.

The problems of dating the upper Karewa formation were outlined in chapter 3. In this study, one of the major aims was to develop a sound chronological framework for the upper Karewa sediments. Magnetic polarity stratigraphy is of no use for this purpose, since the whole of the upper Karewa formation falls within the Brunhes normal chron (Kusumgar, 1980). Radiocarbon dating is of no value either. Previous dates on loessic palaeosols, reviewed in chapter 3, are unreliable due to contamination of the soils with young carbon. The upper Karewa lacustrine sediments are virtually inorganic and beyond the range of radiocarbon dating. An attempt was made to date mollusc shells from Burzahom using uranium-series dating. However, this was unsuccesful, since uranium appeared to have been leached from the shells following the death of the molluscs (A. Rae, personal communication).

The dating in this study was fulfilled using two techniques: thermoluminescence (TL) and amino-acid dating. The dating in this study was carried out on the upper Karewa lacustrine beds, since the upper Karewa conglomerate contains no datable material and the dating of loess has already been undertaken as part of a wider study on this deposit in Kashmir. However, the dates on the loess are of critical importance for this study, and will also be discussed.

a. TL Dating.

TL dating, originally used to date archaeological pottery, has now been applied to a range of different types of sediment (Wintle and Huntley, 1982). TL is the light emitted from a mineral crystal when heated following exposure to ionizing radiation. Heating, or exposure to sunlight, will reduce the stored TL within a mineral to a residual component. With the passsing of time, providing there is no reexposure to heat or light, TL will build up again. The amount of TL present is, therefore, a function of the age of the material. The TL age of a sediment is is essentially the date of its last exposure to light. Assuming that there has been no reworking or reexposure of the material, the TL date will be the date of deposition (Wintle and Huntley, 1982).

Aeolian loess is particularly suited to dating by TL, since it can be assumed that the mineral grains are 'bleached' of stored TL whilst airborne. Sediments exposed on unvegetated river floodplains and in shallow rivers can also be bleached. Under the assumption that the upper Karewa lacustrine sediments were deposited either from aeolian suspension or in shallow-water rivers, they should be datable by TL.

TL dates have been undertaken on upper Karewa lacustrine sediments and loess from Burzahom, Humhama, Wogahoma, Pehru, Dipur, Karpora and Pattan. All TL measurements were undertaken by Dr. H. Rendell (University of Sussex). I collected samples from Burzahom, Pattan and Pampur. However, only the Pattan sample was subsequently dated. The other dates on the upper Karewa lacustrine sediments and loess were on samples collected by Dr. Rendell. TL dating of the loess has also been carried out at Karpora, Dilpur and Burzahom by Singhvi et al. (1987).

Sample collection involved the removal of slumped debris from the face of the section, together with at least 0.3 m of undisturbed sediment. A block of material was then excavated from the sediment, under shade. This block was then wrapped in aluminium foil and sealed in plastic, prior to shipment to the UK for analysis.

b. Amino-acid Dating.

Amino-acid dating is based on the fact that amino-acid molecules can exist in optically different forms called stereoisomers. The stereoisomers are known as D (dextro) and L (levo), depending on the way that they rotate plane polarized light. Amino acids in living organisms are virtually all L-form. On death, however, there is a progressive change to D-form, by a process known as racemization. The extent of racemization, shown by the L:D ratio, increases with time after death. Other environmental factors also affect the rate of racemization, notably temperature and pH (Miller and Hare, 1980). Racemization rate will also depend on the matrix in which the amino acid is found. In the case of mollusc shell, for example, the rate is genus-dependent (Miller and Hare, 1975). The extent to which amino acids are bound to each other is a further control on racemization. The position of any binding is also important. Rates of racemization are lower for free amino acids (ie. unbound) than for bound ones. Bound and free fractions are generally separated prior to analysis, which is by gas or liquid chromatography.

Although attempts have been made to use amino-acid racemization as a geochronometric dating technique, this approach is hindered by the lack of knowledge of the thermal history of the sample. However, D:L ratios can be used as indicators of relative age, if it can be assumed that all of the samples in the study have similar thermal histories.

This usually means that they have similar depths of burial below the ground surface and have been subjected to similar external temperature regimes.

In this study, six samples of fossil mollusc (V. piscinalis) were used for amino-acid analysis. The samples were collected from horizons where molluscs were abundant. These included Burzahom (units 1b, 6b, and 8b), Sambur Village (unit 2) and Pattan (unit 3). Analyses were undertaken by Dr Bill McCoy (Amino Acid Geochronology Unit, University of Massachusetts). For each sample, the ratios of alloisoleucine to isoleucine were determined for the free fraction and total acid hydrolysate.

c. Results.

All of the samples in this study were dated using the regeneration method of TL dating (Wintle and Prószyńska, 1983). Only the Pattan sample could be dated additionally by the additive dose method (H. Rendell, personal communication). For sediments older than 80 to 100 kaBP, the regeneration method gives an indication of the relative age of the sample, with increasing divergence between the TL age and the true age for older samples. Analysis was restricted to samples from close to the top of the upper Karewa lacustrine sediments, since a priori considerations of the age of the upper Karewa lacustrine sediments, based on the thickness of the loess overburden, suggested that the upper parts of the sequence would be close to the age-limit for TL dating. This assumption turned out to be correct. The TL dates, therefore, provide chronological constraints on the terminal phase of the upper Karewa lake. The Pattan sample provides some indication of the difference beteween minimum and true age estimates, since both additive dose and regeneration methods were employed. Dates on the loess are also relevant to this study. The results of TL dating are shown in figure 49. Individual dates are listed in appendix 2.

The TL dates from Singhvi et al. (1987), also shown on figure 49, differ in some respects to those of Rendell. From Rendell's dates, the end of upper Karewa lake seems to have occurred between about 115 and 79 kaBP. The only date from Singhvi et al. (1987) that relates directly to the upper Karewa lacustrine beds comes from the basal loess at Burzahom, which is dated at 101 kaBP. Although there is some discrepancy between the two dates, there is broad agreement that the upper Karewa lake began either to drain or desiccate at least 115 kaBP. The two sets of dates disagree more seriously about the age of the onset of loess deposition. Rendell's dates indicate that the oldest loess in Kashmir, at the base of the Karpora section, is at least 144 kaBP. However, Singhvi et al. (1987) argue that the onset of loess deposition occurred much earlier. They maintain that the basal soil at Burzahom, together with the fifth soil from the base of the profile at Karpora and Dilpur, date from the last interglacial. The presence of four soils beneath this at Dilpur and Karpora led Singhvi et al. (1987) to argue that the lower part of the loess sequence spanned at least four interglacials. Thus, it must be at least 300 Ka old. Since Singhvi et al. did not date samples from the lower part of the loess profiles at Dilpur or Karpora, it is difficult to sustantiate this claim. However, since Kashmir is in the subtropics,

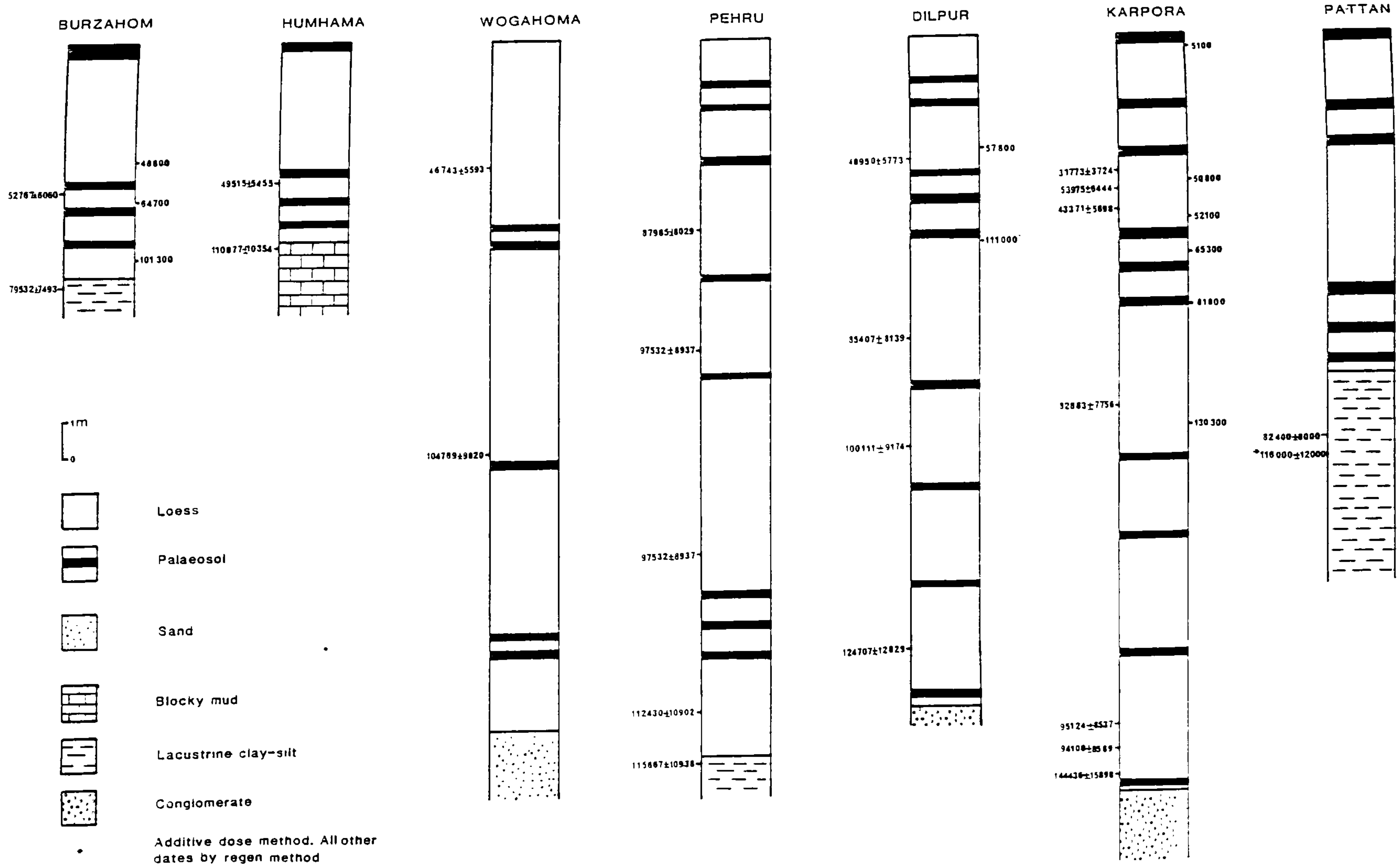


Figure 49 Summary of TL dates for the upper Karewa.
Dates on left of columns from H. Rendell (pers. comm.)
Dates on right of columns from Bronger *et. al.* (1987)

it would be expected that warm periods would occur approximately every 21 ka, in response to precessional insolation maxima (Berger, 1978). If the soils in the Kashmir loess formed in response to these maxima, then the base of the loess at Karpora and Dilpur should date roughly to about 180 kaBP, since there are nine soils in the profile. Whereas, the dates from Sighvi et al. correspond to a pattern of one soil every 21 ka, their estimate of the minimum basal age for the loess of 300 kaBP seems too old. Although TL dating has provided an initial chronology for the upper Karewa formation, more research is required to resolve the apparent contradictions between the two sets of dates.

The results of amino-acid analyses are shown in table 9. For each sample, the ratios of alloisoleucine to isoleucine are given in the free fraction ('free') and total acid hydrolysate ('Hyd'). For the samples from Burzahom, there is a slight decrease in the ratio for the 'hyd' fraction, which suggests a small difference in age that is consistent with the stratigraphy. However, this is not supported by ratios from the 'free' fraction, where the youngest sample stratigraphically has the highest ratio. 'Hyd' ratios are generally more reliable indicators of age, but are usually supported by 'free' ratios (W.D. McCoy, personal communication).

Based on these analyses, and the ratios for the Pattan sample, it appears that unit 1b at Burzahom correlates with unit 3 at Pattan. However, this assumes that the thermal histories of the samples are the same. The depth of burial and depth below the ground surface of the samples were sufficiently large in each case to render differences

TABLE 9. AMINO-ACID RACEMIZATION ANALYSIS RESULTS.

Site	Bed No.	Sample No.	AIlo/Ilo ratios Free Fraction	Total acid Hydrolysate
Pattan	3	AGL 530	~0.44	0.31±0.02
Sambur Village	2	AGL 531	0.77±0.15	0.59±0.09
Burzahom	1b	AGL 532	0.40±0.04	0.33±0.03
Burzahom	6b	AGL 533	0.38±0.02	0.30±0.03
Burzahom	8b	AGL 534 AGL 535	0.45±0.05*	0.27±0.04*

*Mean of samples AGL 534 and AGL 535.

in thermal history due to depth of burial insignificant. Furthermore, despite the absence of meteorological data for both sites, it is unlikely that the temperatures at each site were markedly different since the sites are quite close to one another, fall under the same climatic regime and are at almost identical elevations above sea level. On this basis, the correlation suggested by the L:D ratios is considered valid and not a function of differences in thermal history.

The sample from Sambur Village, however, shows a wide variation in both the 'free' and 'hyd' ratios. This is typical of samples that were shallowly-buried (W.D. McCoy, personal communication). Because of their proximity to the ground surface, the mollusc shells would have experienced a relatively high effective temperature, even though the thermal history of the Sambur Village section is unlikely to differ markedly from that at Pattan or Burzahom. The high effective temperature would appear to have led to high alloisoleucine:isoleucine ratios, precluding meaningful correlation with the other two sites.

The chronological information derived fom TL and amino-acid dating is used to provide a temporal framework for environmental changes that occurred during the period of upper Karewa deposition. This information is synthesized in the final section of this chapter, which follows.

5.5 Conclusions.

The onset of upper Karewa sedimentation began with the rapid uplift of the Pir Panjal Range (Bhatt, 1982a). This event led to the areal restriction of the lake in Kashmir basin. Braided-stream conglomerates were deposited in the south-western part of Kashmir basin concurrently with fine-grained lacustrine deposits in the central part of the basin and on the north-eastern flank. The onset of upper Karewa sedimentation has a minumum age of 350 to 400 kaBP (Burbank, 1982). Upper Karewa lacustrine sediments are interbedded with the conglomerate close to 1680 m a.s.l. Limited exposures of cemented conglomerate occur on the Himlayan flank. The fact that these conglomerates are also interbedded with upper Karewa lacustrine deposits suggests that they are of similar age to the conglomerates in the south-westen part of Kashmir basin. However, the contrasting lithological composition suggests different source-areas.

Towards the centre of the basin, beyond the extent of the upper Karewa conglomerate, the boundary between upper and lower Karewa formations occurs in fine-grained sediments and must presumably be conformable. This illustrates the problems of using lithological character as a means of correlation in the Karewa beds. Because of this, the terms upper and lower Karewa should remain informal. It is also difficult to estimate accurately the thickness of the upper Karewa. However, on the basis of measurements from this study, the lacustrine unit seems to be about 50 m thick, at maximum. Singh (1982) proposed a thickness of 100 m for the thickest exposures of the upper Karewa conglomerate and work on the loess (R. Gardner, personal communication) has shown that the thickest exposures, in the Kanchi Kol Nala, are about 20 m thick.

Several periods of delta formation occurred in the upper Karewa lake. The deltas were of Gilbert type, and relatively small. This observation supports Singh (1982). The upper Karewa lake was generally shallow, although individual sections show stratigraphic evidence of fluctuations in depth. Whereas some of these fluctuations may have resulted from changes in lake water-volume,others were probably the result of migration of deltas due to changes in sediment input.

The lithological character of the uppper Karewa silts and clays strongly suggests a significant input of aeolian silt into the upper Karewa lake. This suggests that there was a lack of vegetation cover over parts of the surrounding basin. This may have been localized, for example, on outwash plains of large gravel fans, or a more widespread phenomenon. Silt particles could have been derived from uplifted lower Karewa strata, poorly-sorted braided-stream deposits, screes and debris fans, and possibly also from glacial deposits. An unknown proportion of the silt may also have been derived from outside Kashmir.

The most detailed study of upper Karewa environments was undertaken on the Pattan setion. The top of the section was dated to a maximum of less than 116 kaBP using TL. The base of the profile, some 19 m below the dated unit, appears to be beyond the range of TL dating. However, an estimate of age was determined by calculating a mean sedimentation rate for the upper Karewa at Pattan. If the total thickness of upper Karewa strata at Pattan is 50 m, and the onset of upper Karewa sedimentation as 400 kaBP (Burbank, 1982), then the mean sedimentation rate in the upper Karewa lake would have been 17 $cmka^{-1}$. This is very close to Burbank's (1982) estimate of 16$cmka^{-1}$ for the upper part of the lower Karewa. However, it must be emphasized that the calculated rate for the upper Karewa is poorly constrained by the available dates, and a high degree of accuracy cannot be claimed. Using the rate of 17 $cmka^{-1}$, the base of the Pattan section would be dated to about 240 kaBP. This compares with a mean sedimentation rate for the loess of between 10.5 $cmka^{-1}$ (using Rendell's dates) and 7 $cmka^{-1}$ (using the dates of Singhvi et al., 1987).

The ostracod evidence suggests that the upper Karewa lake was shallow with abundant aquatic vegetation, cool and non-saline. The similarity between ostracod assemblages in the upper Karewa lake and those in modern Kashmir lakes suggests similarity between the two environments, although the upper Karewa lake was much larger. The ostracod record shows no dramatic change between 240 and 120 kaBP, which is the temporal span of the record. Some of the barren zones are explained by changes in the local depositional environment in the Pattan section, related to the migration of deltas. This is supported by sedimentological changes. However, two of the barren zones occur where there is no lithological evidence for a change in sedimentary environment. In these cases, the barren zones may be related to short-lived, but relatively dramatic, changes in the lacustrine environment. The occurrence of loess within the lacustrine sediments suggests that the climate of Kashmir was relatively dry during upper Karewa times.

The upper Karewa lake seems to have disappeared rather abruptly, some time between 120 and 80 kaBP based on TL dates. However, many of the TL dates for the loess and the upper Karewa lacustrine deposits provide only a minimum age for this event. In most areas, the loess overlies the lacustrine beds conformably. However, in some areas, particularly in the central part of the basin around Humhama, the lacustrine strata were gullied prior to the deposition of the loess. The ostracod record suggests no major change in the lacustrine environment towards the top of the sequence, prior to the actual cessation of the lake. Presumably, the lake either underwent very rapid desiccation as a result of climatic change, or rapid drainage due to tectonic activity which opened the Baramulla Gorge. The post-lacustral phase was dominated by terrestrial loess deposition on the upper Karewa conglomerates as well as the exposed lacustrine beds. However, since loessic deposition was already occurring while the lake was in existence, this fact alone does not suggest increased aridity. In the absence of palaeontological evidence for a climatic explanantion of lake cessation, it is suggested that the lake was drained by tectonic uplift. This hypothesis is supported by limited faulting around Humhama and by gullying in the upper part of the lacustrine beds. The gullying would have been caused by a rise in base level in the Kashmir basin. The absence of basin-wide gullying and faulting is perhaps puzzling. However, only slight changes in basin topography would be required to drain the lake.

CHAPTER 6. THE LATE QUATERNARY GLACIAL HISTORY OF KASHMIR.

6.1 Introduction.

In the report of the International Geological Correlation Programme project 24, Quaternary Glaciation in the Northern Hemisphere (Šibrava *et al.* (eds.) 1986), there is no chapter dealing with the glaciation of the Himalayas. Apart from work carried out in a few restricted areas, the mountain range remains poorly studied. Very little recent work has been undertaken and most of the knowledge of glaciation in this region comes from late ninteenth and early twentieth century accounts. In this chapter the Late Quaternary glacial record in Kashmir is reassesed.

6.2 Aims and Methods.

The aims of this chapter are to reconstruct the timing and magnitude of Late Quaternary glacial advances in Kashmir and to assess their climatic and tectonic implications. This reconstruction involved 5 stages.

1. Diagnosis of the major glacial sedimentary units, principally till and outwash.
2. Mapping the extent of these units.
3. Age-differentiation of the units using RD (relative dating) methods.
4. Geochronometric dating.
5. Assessment of the climatic and tectonic implications of the results

by determining the timing and magnitude of former glacial advances and associated changes in equilibrium lines compared with present patterns.

In this research, the fieldwork was concentrated in three valleys: the Sind and the Liddar on the Himalayan flank and the Ningle on the Pir Panjal flank. Government restrictions precluded access to any aerial photgraphs and satellite imagery that may exist. 1:50 000 scale maps of Kashmir have been produced by the Government of India, but they are also restricted. However, these maps are enlargements of 1:63 360 scale maps produced by the Survey of India prior to independance. These maps are easily available in the UK. They depict drainage and topography with acceptable accuracy and have a contour interval of 100 feet. Enlargements of these maps were generally used for the mapping of sedimentary units. Certain key altitudes were also determined in the field using a pocket altimeter. The one inch maps were also used to determine present-day patterns of glaciation for Kashmir and for the reconstruction of past ice-cover.1:250 000 scale maps, published by the Defence Mapping Agency, Washington DC, were also available. However, they have a contour interval of only 100 m and depict relief and drainage more generally. These maps were, therefore, only used for reconnaissance.

In each of the three valleys, the occurrence of glacial and non-glacial sediments was mapped. Particular attention was paid to localities studied in previous work, but many other new sites were also investigated. The methods used to determine the origin of each sediment type are discussed in 6.3. Relative age of sedimentary units was determined using a range of boulder-weathering and soil-development measures. In several localities, material was collected for radiometric dating. The procedures used are discussd in 6.4. Using the information from field mapping and from the one inch topographic sheets, present and past patterns of glaciation were reconstructed. The climatic significance of present-day patterns was used to evaluate the reconstructed patterns for Late Quaternary climate and tectonic uplift.

6.3 Glacial and Related Sediments in Kashmir.

a. Introduction and Definitions.

This section deals with the recognition of till and related sediments in Kashmir and discusses their distribution and sedimentology in each of the valleys studied. This information is then used in the reconstruction of the Late Quaternary glacial history of Kashmir.

The distinction between tills and other diamictons is often difficult because non-glacial processes which form till-like sediments tend to be common in mountainous regions, particularly those subject to recent tectonic uplift. Furthermore, till itself may be reworked by non-glacial processes. Because of the variability of till, it is not surprising that no single feature is considered diagnostic. Before examining the criteria that have been used in till diagnosis, it is necessary to consider some definitions.

The term 'diamicton' is non-genetic, and has been defined by Frakes (1978, p.262) as follows.

> 'Terriginous sedimentary rocks ranging in particle size from clay to boulder dimensions...the strictly descriptive definition relates to particle size and not to abundance of any or all size classes."

The term 'till' is generally agreed to be genetic and is defined by Boulton (1972, pp.378-379) as

> 'An aggregate whose particles have been brought together and deposited by the direct agent of glacier ice, which, though it may suffer post-depositional modification by flow, does not undergo subsequent disaggregation and redeposition.'

Although there is a burgeoning literature relating to the terminology of tills, daimictons and related sediments, no attempt is made to review this here. For the purpose of this study, the terms 'till' and 'diamicton' will be used as defined above.

b. The Occurrence of Till.

Till is deposited in a glacial environment, which is defined as

> '...one in which the principal agents of sediment transport are glacial ice or glacially-derived meltwater whose nature reflects its source.' (Boulton and Deynoux, 1981, p.398).

Three major groups of processes operate within the glacioterrestrial environment: glacigenic processes, which are glacial in the strict sense; glaciofluvial processes; and glaciolacustrine processes.

Boulton and Deynoux (1981) summarize the main types of till as follows.

> 'Lodgement till (Chamberlain, 1894)... deposited beneath an actively moving glacier, as a result of retardation of debris particles or debris-rich ice masses by friction against the glacier bed...' (p.400).
>
> 'Melt-out till (Boulton, 1970) and Sublimation till (Shaw, 1977)...melt-out till is deposited by slow melting of masses of buried stagnant ice and retains some of the structures of the englacial debris from which it is derived. It may be deposited on the surface of a melting ice mass (supraglacial melt-out till) or beneath it (subglacial met-out till). Sublimation till forms on the surface of glaciers in very cold and dry environments where ice is lost by sublimation...' (p.400).

The three types of till mentioned above are known as primary tills. Three types of secondary till are also identified;these are defined by Boulton and Deynoux as follows.

> 'Flow till (Hartshorn, 1958)... when debris melts out on a glacier surface as melt-out till, the high water content of the nascent till mass makes its potentially unstable and it flows readily...' (p.400).
>
> 'Deformation till (Elson, 1961)...substantial deformation of subglacial sediments produced by the drag of the moving glacier sole is known to occur...Banham (1977) has suggested...that bedrock, and presumably unlithified sediment, penetratively deformed beneath a glacier should be termed 'till'...Elson's (1961) term 'deformation till' should be applied to such material.' (p.403).
>
> 'Supra-tills.' In many valley glaciers, large quantities of debris are introduced onto the glacier surface from the flanking mountains. This material is then transported passively at a high level within the glacier system to produce a melt-out till or flow till on the glacier surface in the terminal area...' (p.403).

A further category of secondary till is the 'slide facies' (see Derbyshire, 1981), which is deposited in a similar depositional setting to flow till, but by sliding under gravity directly from the glacier slope, rather than by flow.

c. The Occurrence of Non-glacial Diamictons.

Non-glacial diamictons may result from subaerial mass-movement, subaqueous turbidity currents, fluvial deposition, faulting and differential chemical weathering. The products of all of these processes may, in certain circumstances, be confused with till.

Subaerial mass-movemnt comprises a number of processes that may form diamictons including debris flows and slides, mudflows and mud avalanches. In many respects, these terms describe variants of a single process (Johnson, 1984), and the blanket-term 'debris flow' will be used here. Lahars and solifluction deposits are also important sources of diamictons.

Debris flows are mixtures of granular solids, entrained water and air, with only minor amounts of clay. They move readily on low-angle slopes (Johnson, 1984). The resultant debris-flow deposit is often a coarse diamicton, the precise characteristics of which will depend on the local bedrock-lithology, the distance of travel and the local relief. When seen in vertical section, debris-flow deposits may look similar to till. Recently deglacierized valleys are highly susceptible to debris flows, since they are often characterized by unstable slopes with an abundance of transportable material, a lack of vegetation cover and availabitiy of water (Selby, 1982). Tills themselves are liable to be reworked by debris flows.

Fluvial gravels are classically regarded as well sorted, clast-supported deposits with well-rounded clasts in a sandy matrix showing characteristic bedding features such as scour-and-fill and clast imbrication. Tills, by comparison, are generally regarded as poorly sorted, matrix-supported sediments containing subangular to angular clasts in a silty-clay matrix, with little or no bedding other than preferred orientation of clasts. In many situations, this distinction holds, and fluvial gravels can easily be distinguished from tills. However, sediments classified as being fluvial can be almost as variable as till and instances where they could be mistaken for till do exist. In high-energy environments, for example, where large quantites of sediment are transported in a high sediment to water ratio flow, the resulting deposit may be poorly organised and lack many of the supposedly typical fluvial features. This may be particularly so in the proximal zone of deposition. Such conditions often exist in a proximal proglacial environment and in basins adjacent to tectonically active mountains. In each of these cases, till-like deposits may be formed. Both of these situations may be prevalent in Kashmir, which has undergone glaciation during a phase of tectonic activity. Because deglacierization and tectonic uplft both create high-energy fluvial environments and because sediment character is linked more directly to energy conditions than to climate, these two environments may produce similar sorts of sediments. As a further complication, tectonic uplift and glaciation may have occurred concurrently. As a result of this, the tectonic and climatic record in sediments will be interrelated, and must be interpreted with extreme caution.

There are two additional instances in which diamictons may be produced. A fault breccia results from the breakdown of rock adjacent to faults. The nature of such a breccia will clearly depend on the characteristics of the bedrock and the dynamics of faulting. However, in certain circumstances, a fault breccia may be confused with till. This is particularly the case for an older till, which would be expected to persist only in isolated patches and lack surface morphology. Differential chemical weathering under warm, humid climates may also produce diamictons. Derbyshire (1983c) reports such deposits in the Lushan Massif of south-east China, which for several decades were regarded as till.

One of the potentially intractable problems of till diagnosis is that of polygenesis. In studies such as the present one, the major aim is to identify the last major process of transport and deposition. The concern is not, therefore, with the entire history of the sediment from primary erosion to final deposition. Nor is the concern with minor postdepositional reworking. However, diagnosis of the origin of a sediment is often hindered by the fact that it may have inherited characteristics from previous processes that are unrelated to the final mode of deposition. It may also be affected by postdepositional reworking. Therefore, it is essential to use a range of criteria in the diagnosis of the origin of a sediment.

d. Criteria for the Distinction Between Tills and Other Diamictons.

Many early studies of glacial stratigraphy paid little heed to the question of till identification. This appears to have been the case in

Kashmir where de Terra and Paterson, for example, mapped many non-glacial diamictons as 'moraines'. Recent studies have emphasized not only the importance of careful distinction between tills and other diamictons, but also the genetic diversity of till itself. Despite attempts to distinguish between different types of till and the impressive increase in the amount of such work over the last 20 years or so, the field of glacial sedimentology is regarded as poorly developed (Derbyshire, 1981).

From the point of view of glacial stratigraphy, it may seem sufficient to diagnose a sediment as till and unnecessary to indulge in further classification. However, an awareness of the different types of till recognised in the literature is important because it emphasizes the wide variability of many till characteristics. This variability must be taken into account in an attempt to distinguish between tills and other diamictons. Furthermore, Derbyshire (1984c) has suggested that there may be links between till type, glacier type and climatic regime.

Many of the methods used to distinguish between different types of till are the same as those used to distinguish between tills and other diamictons. The range of criteria used in the past is outlined below.

Recent work on tills has tended to favour a genetic classification which relates the till to the debris source, its position within, and transport path through, the glacier, and the type of landsystem that arises upon deposition and the recession of ice. (eg. Boulton and

Paul, 1976; Eyles, 1983). This diagram illustrates the generally-held view that various types of till have different characteristics and that different scales of approach can be adopted for their study. These scales, which provide a useful framework for the study of non-glacial diamictons as well as till, are as follows.

1. The sedimentological scale, which includes the size and shape of sedimentary particles and the textural, mineralogical and petrological characteristics of clasts and grains.
2. The stratigraphical scale, which includes features such as clast and grain fabrics, patterns of jointing and faulting, internal bedding, and colour.
3. The morphological scale, in which the sediment is examined as a landform. In this case, attributes such as size, shape and orientation are examined. Recently, the assemblage of individual landforms into a 'landsystem' has been stressed as important in the palaeoenvironmental reconstruction of glaciated areas (Boulton and Paul, 1976; Eyles, 1983).

Many studies of diamictons have examined sedimentological properties, especially particle-size distributions. The use of particle-size analysis in this way assumes that diamictons of different origin have distinctive sources and/or transport modes and/or depositional mechanisms. It also assumes that these differences will influence the particle-size distribution. This approach has been used quite widely both to distinguish between different types of till and between tills and non-glacial diamictons. It is, however, important to recognise that most sediments will be affected to some degree by post-depositional modification. This may affect the particle-size distribution and so lead to false conclusions being drawn. In particular, the fine fraction of a sediment may be removed by percolating water, soil formation and aeolian action (Boulton <u>et al</u>., 1974). Conversely, a coarse, fines-deficient gravel may be susceptible to the inwashing of fines if low-energy, sediment-laden waters flow across its surface soon after deposition.

Boulton (1978) distinguished between 'active' and 'passive' transport paths in glaciers. He argued that, in the basal zone of traction, material is actively comminuted. In contrast, material in the supraglacial zone or the high-level englacial zone is transported passively. Material within a glacier can be supplied either from the glacier bed (basally) or from nunatak and the valley-side slopes (supraglacially). Supraglacial debris is likely to be transported in either the supraglacial or englacial transport zone and is likely to be transferred to the subglacial path only if it is entrained well above the equilibrium line of the glacier. Subglacially-derived material will be transported tractionally and in suspension in the basal zone of the glacier, but may be transferred to englacial or supraglacial positions (Boulton and Paul, 1976).

Boulton (1978) was able to distinguish between tills that had been actively transported and those that had only been passively transported, on the basis of the textural characteristics of the till. Actively-transported tills were found to be enriched in silt and clay and depleted in the coarser, sand fraction. Conversely, passively-transported tills were found to be fines-deficient. Boulton argued that the absence of comminution in the passive transport path was the main reason for this.

Derbyshire (1984) reported work on the Hooker Glacier, New Zealand. The dominance of melt-out processes on this glacier produces tills that are fines-deficient. This was thought to be due to mechanical eluviation of fines by percolating water and aeolian action. Li et al. (1984) found similar depletion of fines in the melt-out tills of the Hunza Valley, Karakoram Range, and were able to distinguish between tractionally-transported and melt-out tills on the basis of particle-size analysis. Eluviated tills were found to be transitional between subglacial and melt-out types. The clear distinction between the active and passive transport paths in glacial sediments is also shown in the particle-size plots of German et al. (1979), Dreimanis (1976), Rabassa and Aliotta (1979), Schubert (1979) and Vorren (1977).

Particle-size analysis has also been used to distinguish between tills and non-glacial diamictons. For example, Passega (1964) used bivariate plots of the coarsest single percentile in a particle-size distribution against the median particle diameter, in millimetres. These plots were succesfully used to distinguish between sediments deposited by turbidity currents and those carried tractionally in a glacier. Landim and Frakes (1968) used bivariate plots of measures such as mean particle-size, sorting, skewness and kurtosis and were able to distinguish between tills, outwash and alluvial fan sediments. Debris-flow sediments and alluvial fans sediments showed depletion of fines compared to till. Outwash and paraglacial fan sediments were more poorly sorted, and contained a wider range of particle sizes than true fluvial gravel, but were better sorted than till. However, proximal outwash may pose diagnostic problems, since it is often poorly sorted.

Boulder and grain shapes have been used to identify tills and various subenvironments. Boulton (1978) argued that boulder shape is a better indicator of till-type than particle-size distribution, because it is less prone to postdepositional modification. However, shape can be inherited from a previous cycle of erosion and deposition. Both clast shape and, more rarely, grain shape, have been used to distinguish between active and passive transport paths in glaciers. Studies using shape as a means of distinction between glacial and non-glacial diamictons are rarer. However, King and Buckley (1968) were able to distinguish between end moraines, kames and eskers on the basis of clast roundness alone. Boulton and Eyles (1979) used incipient edge-rounding on clasts as a means of distinguishing between two very similar facies, proximal outwash and supra-till.

Fabric analysis has been used in the study of tills and other diamictons. Following Derbyshire et al. (1976), the 'total' fabric of a sediment consists of macrofabric (disposition of folds, thrusts and fissures), a mesofabric (clast disposition) and a microfabric (oganisation of the matrix). Although most studies have concentrated on mesofabric, there is an increasing awareness of the importance of all three aspects in the study of diamictons.

The lithology of clasts and matrix, whilst of little value in distinguishing between different types of till, may help to distinguish between tills and non-glacial diamictons. The value of this discriminator rests on the assumption that the source of rock within sediments can be identified and characterized. If this assumption can be fulfilled, debris-flow and solifluction deposits will tend to contain higher proportions of locally-derived material than tills. Ryder (1971) also suggested that paraglacial fan material will tend to show less variation in lithology than true fluvial gravel, due to differences in the size of the source area.

The surface morphology of a deposit, its orientation and relationship with other landforms can often have considerable diagnostic value. In glaciated mountain areas, mass-movement sediments may be identifiable by their fan shape and genetic relationship with valley sides. However, older and more degraded features, often buried beneath other facies, may be more difficult to identify. Alluvial fans tend to have a distinctive planform. However, as with mass-movement deposits, these sediments may have the appearance of moraine at the low, oblique angles of observation available in the field and in the absence of aerial photography.

e. The Distribution and Sedimentology of Tills and Related Sediments in Kashmir.

Because Kashmir is well vegetated, good exposures of glacial deposits are not common. In this study, sediments were examined in river- and road-cuttings wherever possible, or in hand-dug pits. The exposures examined included each of de Terra and Paterson's 'moraines'. At each site, particle size, clast shape, calcium-carbonate content and moist colour were determined in the field and in hand specimen. Estimates of plasticity and stickiness of the sediment matrix were carried out to determine texture (Birkeland, 1984). Colour was determined using revised Munsell colour charts. Calcium-carbonate content was estimated using dilute hydrochloric acid (Hodgson, 1976). Evaluation of clast orientation, bedding, internal structure, external boundaries, bed superposition and overall morphology was also carried out in the field. Bulk disturbed samples were then taken for more detailed sedimentological analysis. Undisturbed samples were not collected, since most of the diamictons are poorly consolidated.

The shape and lithology of clasts were determined at base camp in Srinagar. This involved the visual estimation of roundess (Powers, 1953, see appendix 9 of this thesis) and lithology of 50 to 100 clasts. Typically, more than one sample was take from each sediment unit. All other sedimentological analyses were carried out in the UK. Particle-size analysis involved dry sieving of the sand fraction (> 63 micron) and wet pipette analysis for the silt and clay grades. Roundness of quartz grains was determined using the above-mentioned visual estimation methods under low-power stereomicroscopy. The medium-sand grade, retrieved after dry sieving, was used for this purpose. The percentage of calcium carbonate in the matrix (< 2 mm) was determined using a manometric calcimeter calibrated with pure calcium carbonate.

The sediments were classified provisionally in the field into three categories: till, outwash and non-glacial diamicton. Over 90 samples were subsequently analysed in the laboratory, with the aim of testing the classification and providing greater insight into the sedimentology and genesis of individual deposits. The results of field observations and laboratory analyses are discussed in this section, initially by site. The aim of this is to evaluate the extent of glaciers in Kashmir during the Late Quaternary. Following this, statistical analysis of the sedimentological data is used to test the field diagnoses. Finally in this section, the sedimentological characteristics of the glacial sediments are discussed. The data upon which this section is based are listed in appendices 6, 10 and 11.

The Sind Valley

In the Sind Valley, the lowermost occurrence of diamicton is near Mangom (figure 72, plate 22). The large conglomerate terraces in the mouth of the Sind Valley have already been discussed in chapter 5, and a glacial origin discounted.

At Mangom, a large gravel fan enters the main Sind Valley from a large tributary that lies to the north. The fan surface is well vegetated and despite dissection by several small streams, is largely inactive. The observed section was exposed in a road cutting in the distal portion of the fan. In the lower part of the section, the diamicton is interbedded with massive silt and sand beds. However, the junction between this unit and the upper part of the section was obscured. The upper part of the section consists of matrix-supported diamicton, showing crude alignment of clast long-axes transverse to the main Sind Valley. The diamicton consists of angular to subangular clasts mostly less than 15 cm long, although exceptionally, clasts of more than 100 cm length were seen. The predominant lithology of the clasts was Panjal Trap. The matrix was found to be non-calcareous, dark yellowish brown (10YR 3/4) to dark brown (7.5YR 4/4), with a substantial clay and silt component. The medium sand grains were found to be predominantly very angular to angular.

Field evidence suggests that the feature at Mangom is not a moraine. Its morphology suggests that it is a small alluvial fan with a source area in the mountain flanks of the Sind Valley, immediately above Mangom. However, the poorly-sorted character of the deposit and the angular clasts and grains do not support a fluvial origin. The sediments are better explained by debris flow. Fluvial sediments may be found in association with the small channels on the fan surface and other exposures may reveal fluvial activity in the past as well. However, the high proportion of debris-flow sediments in the observed section is consistent with a proximal fan environment.

A similar fan is seen on the northern flank of the Sind Valley at Kangan (plate 21), close to the tributary of the Sind and Wangat Nala. No structures were seen in the roadcut exposures, although the clasts were aligned crudely transverse to the main sind Valley. The diamicton is matrix-supported, with angular clasts of Panjal Trap and unidentified lithologies in a yellowish brown (10YR 5/6) noncalcareous matrix. Medium sand grains were found to be very angular to angular. Morphologicallly, the feature is similar to the one at Mangom and the same mode of origin is proposed. The lack of incipient rounding of the clasts or grains compared to Mangom is consitànt with a smaller tributary area and, hence, shorter distance of transport.

Exposures of diamicton occur semi-continuously between Pharao and Rezan and were examined at Gund and Rezan. The morphology of the features suggests that they were formed by the coalescence of the distal portions of discrete debris fans.

At Gund, a matrix-supported diamicton is exposed in a road cutting (plate 23). This consists of angular to subangular clasts of Panjal Trap, Quartzite and other, unidentified lithologies in an olive (5Y 5/3) to dark yellowish brown (10YR 3/4) matrix containing moderate amounts of silt and clay, and very angular to angular medium-sand grains. One sample contained a small proportion of limestone clasts, was locally cemented and had a slightly calcareous matrix. However, extensive field assessment together with laboratory testing of the other samples from this exposure showed the Gund diamicton to be predominantly noncalcareous. The exposures at Rezan are in sediments of similar morphology and internal structure to those at Gund, although the clasts and grains are rather more angular. This suggests a shorter distance of transport than at Gund. The diamicton is matrix-supported with very angular to angular clasts of Panjal Trap, quartzite and unidentified lithologies. The olive (5Y 5/3) to dark yellowish brown (10YR 4/4) matrix was found to be noncalcareous with very angular to angular medium-sand grains.

At Gagangiyer, fans enter the main Sind Vallley from both the north and south flanks. The fan entering the valley from the north, at a point just beond the western end of Gagangiyer Gorge, was sampled. The Gorge itself has been dominated by erosion, and no deposits remain. The sediment at Gagangiyer is a matrix-supported diamicton, with an overall fan morphology and no internal structure. It is composed of mainly Panjal Trap clasts in a noncalcareous, dark greyish brown (2.5Y 4/2) matrix containing prodominantly angular medium-sand grains.

A variety of depositional features is seen in the Sonamarg Basin. At the western end, close to the village of Shitkari, a large hummocky feature stretches across the valley at the confluence of the Sind, the Tajiwas and Lashimarg Rivers (plate 32). These rivers have dissected the feature, and terraces have been formed at a lower level, but are apparantly banked up against it.

On morphological grounds, the hummocky feature is interpreted as a moraine. It may have been deposited by a former glacier extending from the Tajiwas and Lashimarg Valleys as well as the main Sind. The sediments are not well exposed and were mainly examined in shallow, hand-dug pits. These pits revealed a matrix-supported diamicton containing very angular to subangular clasts mainly of Panjal Trap. Other lithologies, including a small percentage of limestone, were also found. The limestone content reflects the limestone bedrock found in the Tajiwas Valley and the upper Sind above Sonamarg. However, the dark yellowish brown (10YR 3/3 to 10YR 3/4) matrix was generally noncalcareous and only one sample contained calcium carbonate. The surface soils on the moraine were also found to be noncalcareous. This is rather surprising, in view of the abundance of limestone bedrock nearby and the presence of limestone clasts in the till. However, the noncalcareous samples came from shallow pits dug in the moraine surface, and may have been decalcified since deposition. This idea is supported by the fact that the calcareous sample came from a temporary exposure near the base of a hummock, a site that would not be so susceptible to rapid decalcification. The medium sand grains within the sediment were found to be very angular to angular.

Terraces extend from Shitkari, upvalley beyond Sonamarg to Baltal, at varying heights above the Sind River, but lower than the height of the moraine itself. These terraces are banked up against the moraine where it has been truncated by fluvial erosion. Exposures in these terraces revealed crudely bedded, clasts-supported gravel (plate 24). Most of the clasts were less than 100 cm long although some were larger than this, up to to several metres in length. Limestone was found to be the most abundant lithology, with a minor component of Panjal Trap. The clasts are angular to subrounded. The matrix was found to be generally sandy, highly calcareous and dark greyish brown (2.5Y 3/2) to pale olive (5Y 6/3) in colour. Medium-sand grains vary from angular to subrounded. The terrace gravels are interpreted as outwash.

Immediately upvalley from Sonamarg, a series of large debris fans is found. Exposures in the distal parts of these fans showed that the sediments are matrix-supported diamictons, with some crude parallel bedding. The clast were found to be predominantly limestone, reflecting the bedrock of the source area. The dark greyish brown (2.5Y 3/2) to dark brown (10YR 3/3) matrix is highly calcareous and contains medium sand grains which are typically very angular to angular. The morphology of these features, in contrast to the others described in the Sonamrg Basin, is fan-shaped, and the sediments are interpreted as products of mass movement. Similar features are also found in the Tajiwas Valley.

In the Sind Valley, sediments of undoubted glacial origin were not found below the Sonamarg Basin. The diagnoses were based mainly on morphological evidence, although lithological composition and the alignment of clasts transverse to the main Sind Valley support a non-glacial origin for the diamictons in the lower part of the Sind Valley. Since a glacier extending down the Sind Valley would have passed over areas of limestone bedrock, glacigenic sediments would be expected to contain limestone clasts and have a calcareous matrix. In general, the inferred glacial sediments of the Sonamarg Basin are indeed calcareous. The sediments of the lower Sind Valley tend to be noncalcareous, reflecting their more localized source areas of rocks other than limestone. However, several of the inferred tills from

Sonamarg are noncalcareous and one of the inferred mass-movement deposits, at Gund, contains calcium carbonate. It is possible that the absence of calcium carbonate from most of the lower Sind Valley mass-movement deposits is due entirely to decalcification following deposition. However, this is unlikely, since samples were taken from points stratigraphically low in the deposits in relatively recently-dug exposures. The presence of calcium carbonate in the Gund section may reflect a very localized source of limestone or, perhaps, mixing with material from further upvalley. In either case, the low abundance of calcium carbonate in a single sample is not thought to be very significant.

The Liddar Valley

In the Liddar Valley, the first major occurrence of diamicton is at Ganeshpur, where a large, fan-shaped landform extends from a series of tributary valleys to the east, into the Liddar Valley itself (figure 73, plate 25). Exposures reveal interbedding of matrix-supported diamicton with fine-grained sediment. Individual beds of each facies are of the order of 1 metre thick. The clasts within the diamicton are very angular to subangular, with the dominant lithology being Panjal Trap. The matrix was found to be generally noncalcareous, although one sample did have a very slight carbonate content. The colour of the matrix is dark brown (7.5YR 4/4) to dark yellowish brown (10YR 3/4). Medium sand grains are very angular to subangular. The deposit at Ganeshpur was mapped by de Terra and Paterson (1939) as waterlain till. In this study, it was interpreted as a mass-movement deposit with interbedded loess. The angularity of clasts and grains in the diamicton suggests a modest transport distance. The presence of small clasts within the fine material suggests a colluvial rather than primary aeolian origin, indicating that the loess is probably reworked.

Futher upvalley from Ganeshpur, large terraces occur on both sides of the Liddar Valley. These are particularly well-developed between Batakut and Pahalgam. The morphology of the terraces suggests that they were formed by the coalescence, and subsequent truncation, of large debris fans. Exposures show that they consist of a poorly sorted, matrix supported diamicton with no internal structure other than crude alignment of clasts long-axes transverse to the direction of the main Liddar Valley. The sediments were found to be noncalcareous and to consist of angular clasts, which are predominantly Panjal Trap. The dark brown (10YR 4/3) matrix contains angular medium sand grains.

The Liddar River bifurcates at Pahalgam into east and west branches. At Pahalgam, a large, irregular, lobate feature extends downvalley about 1 km from the confluence of the two valleys. On the eastern flank of this deposit, a poorly sorted, matrix-supported diamicton has been exposed by the road leading from Pahalgam village to the golf club (plate 27). This material is mostly unbedded, although it does contain several small (less than 0.5 m long and 0.2 m thick) lenses of well sorted silt and fine sand. The diamicton consists of very angular to subangular clasts of Panjal Trap and limestone in a light yellowish brown (2.5Y 6/4) to pale olive (5Y 6/4) calcareous matrix. The medium sand grains were found to be very angular to subrounded. This deposit is interpreted as a till. Its compact nature, together with the presence of the small lenses suggests that it might be a lodgement till. On the western flank of this deposit, and at the front of the lobe, the sediments are quite different (plate 28). They are noncalcareous, and consist of angular to subrounded clasts in a dark yellowish brown (10YR 3/4) to dark brown (7.5YR 4/4), clay-rich matrix. The clasts are mainly Panjal Trap and quartzite, but with no limestone. Medium sand grains are angular to subrounded.

Exposures in the western and eastern part of the lobate feature at Pahalgam are clearly in two contrasting facies. Whereas the eastern flank of the lobe is composed of till the western flank is made up of outwash. The contrasting lithology of the till and outwash suggests different source areas. The high calcium-carbonate content of the till suggests a source area in the east Liddar Valley, which is cut into limestone. The west Liddar, by contrast, is cut mainly into noncalcareous bedrock. This is likely to have been the source for the outwash. The high clay content of the outwash is rather atypical for sediments of this origin. However, the incipient rounding of clasts and grains, and the tendency towards clast support argue against it being a till. The high clay-content could be explained by the outwash being proximal to a wasting glacier. Alternatively, the clay fraction may have been deposited by a later event into the interstices in the framework of a preexisting, well-sorted, outwash gravel.

Inferred outwash, of similar sedimentological characteristics, also extends a little over 1 kilometre downvalley from Pahalgam (plate 26). Subsequent downcutting by the Liddar River has formed outwash terraces, upon which the village of Pahalgam itself has been built. This outwash is commonly massive, although it sometimes shows crude horizontal bedding. It is noncalcareous and contains very angular to rounded clasts which are mainly of Panjal trap. The olive brown (2.5Y 4/4) matrix has quite a high clay content, although this does not seem to be as high as for the sediment in the Pahalgam lobe, described above.

Upvalley from Pahalgam, work was concentrated in the west branch of the valley. Between Pahalgam and Aru, gravel terraces occur discontinuously, perched on valley-side benches up to 100 m above the level of the present river (plate 29). These gravels are well exposed at Mondlan, where they were described and sampled. The deposit is crudely bedded, with down-valley clast imbrication. The clasts are typically less than 1 metre in diameter, although very large boulders, several metres across, are not uncommon. The clasts are angular to subrounded, and mainly of Panjal Trap. The gravel is clast-supported, although the infilling light olive brown (2.5Y 5/6) to olive yellow (2.5Y 6/6) matrix was found to be quite clay-rich. No limestone clasts were found and the matrix is noncalcareous. This gravel has much in common with the outwash around Pahalgam, and a similar origin is proposed.

At Aru, the Tson and Girwar Valleys join the West Liddar and a complex of depositional features is seen. The situation is analogous to that at Sonamarg in the Sind Valley. A large hummocky cross-valley feature occurs at the confluence of the three valleys. This has been downcut by the action of the rivers, and terraces of diamicton are banked up against it where this has occurred. This diamicton is exposed at various places around the village of Aru. The diamicton is clast supported and shows crude horizontal bedding. It consists of angular to subrounded clasts of Panjal Trap and several unidentified lithologies. However, no limestone was found and the matrix is noncalcareous. The clasts are typically less than 30 cm, although occasionally, larger clasts were found. The dark yellowish brown (10YR

4/4) to olive brown (2.5Y 4/4) matrix is quite sandy, although clearly contains a small amount of silt and clay. Medium sand grains within the matrix are very angular to subangular. The morphology and internal structure of this sediment suggests that it is outwash.

The large hummocky feature that occurs above the level of the outwash terraces appears to have been derived at least in part from the tributary valley, Girwar Nar. The sediment itself is not well exposed, and was examined in hand-dug pits. It consists of matrix-supported diamicton which contains angular to subangular clasts in a dark brown (7.5YR 3/2) to brown (10YR 4/3) matrix which is noncalcareous. A layer of structureless silty clay, of variable thickness, overlies the diamicton. The hummocky feature is interpreted as till; the overlying deposit as loess.

Another hummocky feature is found at Liddarwat, some 10 km up the West Liddar valley from Aru. This occurs at the confluence between the West Liddar and Tarsar Valleys. The sediment is poorly exposed and was examined mainly in hand-dug pits. It consists of a matrix-supported diamicton containing angular to subangular clasts which are predominantly Panjal Trap, although a small proportion of quartzite was also found. The olive brown (2.5Y 4/4) to dark brown (10YR 3/3) matrix is clay-rich in hand specimen and noncalcareous. This material is interpreted as till. Towards the eastern flank of the hummocky feature, exposures show the diamicton has a sandier matrix and more-rounded clasts, suggesting that terraces of outwash gravel flank the moraine at Liddarwat.

In the foreland zone of the present Kolahoi glacier, some 12 km from Liddarwat, a complex of small, transverse and linear ridges is found. These are unweathered and lack vegetation cover. They are presumably till ridges of recent, Holocene, advances of the Kolahoi glacier. Sediments exposed in these ridges consist of matrix-supported diamicton containing very angular to subangular clasts of Panjal Trap, quartzite and other lithologies (particularly slate) in a black (2.5Y 2/0) to olive (5Y 5/4) noncalcareous matrix. Medium sand grains were found to be very angular to angular. The sedimentological characteristics of these features are consistent with a glacigenic origin, and most of the ridges are clearly moraines. However, in some cases, the sediment is rather coarser and clasts and grains show incipient rounding. This suggests the action of running water.

The Ningle Valley

In the Ningle Valley on the Pir Panjal flank (figure 74), the sediments were not well exposed. In the upper reaches of the valley, military presence prevented digging. In the lower part of the Ningle Valley, de Terra and Paterson mapped moraines at Durhom and Dandamuh. Sections in these supposed moraines at both sites.revealed a crudely bedded, clast-supported gravel containing subrounded to rounded clasts overlying contorted lower Karewa silt-clay facies. The sedimentological character of this gravel, together with its stratigraphical position, above lower Karewa strata, suggests that it is equivalent to the upper Karewa gravel exposed all along the Pir Panjal flank (plate 30). In the lower part of the Ningle valley, the gravel has been eroded to give a hummocky surface expression that could be taken for moraine. However, the internal structure of the deposits discounts a glacial origin.

At Gulmarg, a small basin is infilled with a hummocky feature, which is made up of matrix-supported diamicton (plate 31). The occasional exposures in this material show that it consists of angular to subangular clasts, mainly of Panjal Trap, in an olive brown (2.5Y 4/4) to dark yellowish brown (10YR 3/4) noncalcareous matrix. The matrix contains moderate amounts of clay and very angular to angular medium sand grains. Morphological and sedimentological evidence suggests that this material is till.

Close to the outlet of Gulmarg Basin, gravel terraces are apparently banked up against the moraine. The terraces are composed of matrix-supported diamicton, with angular to subrounded clasts in a dark yellowish brown (10YR 3/4 to 10YR 3/6) sandy, noncalcareous matrix. Medium sand grains are very angular to subangular. Some crude, horizontal bedding was seen in the exposed section. Because of the intimate association of this deposit with till, it is interpreted as proximal outwash. Similar hummocky moraines were seen higher in the Ningle valley at Butapathri. Because of military activity in the area, it was impossible to examine these features in detail. However, they have similar morphology to the moraines at Gulmarg and a similar origin is proposed.

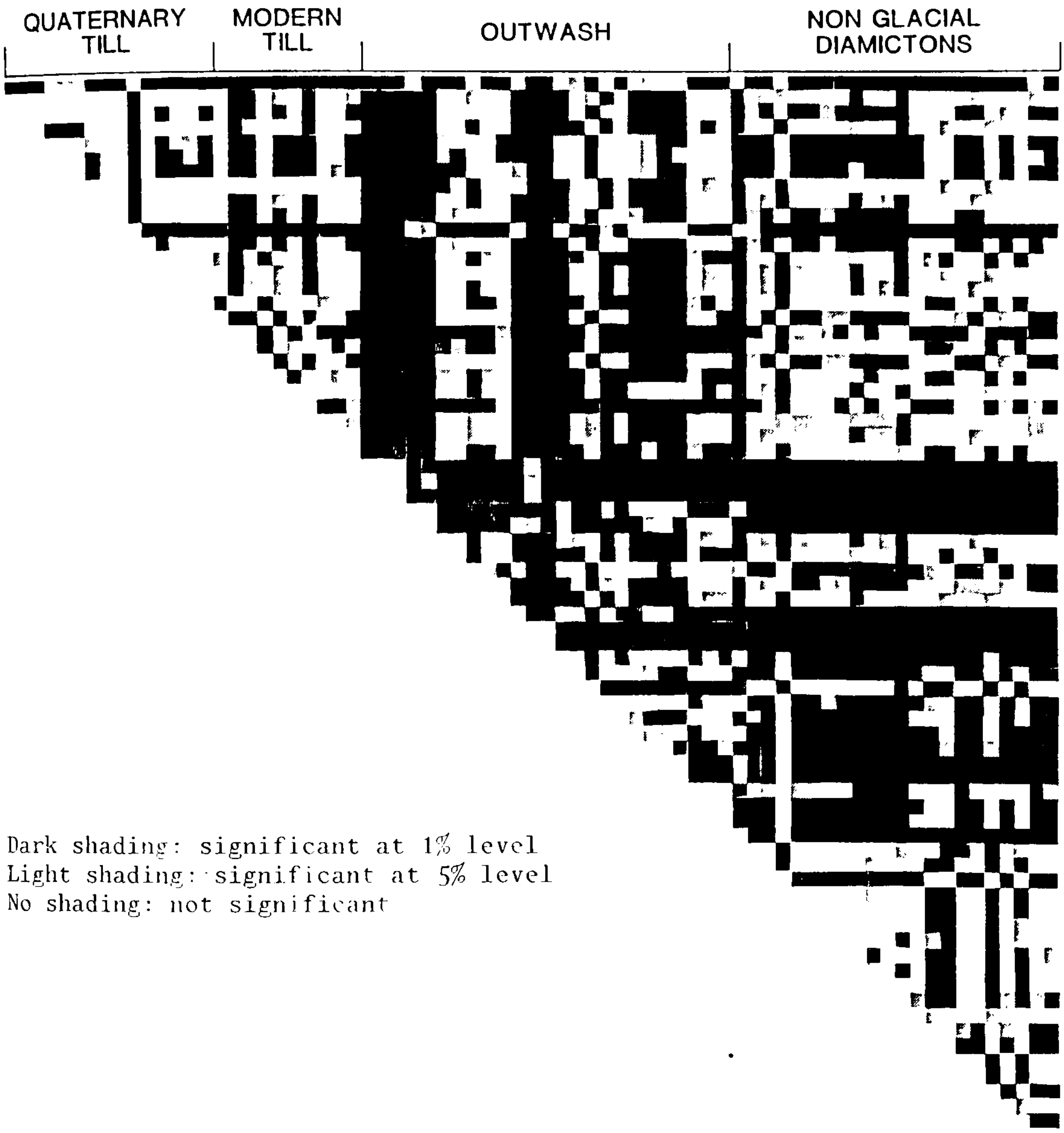
In this study, the maximum downvalley occurrence of undoubted moraines is at Sonamarg in the Sind Valley, Pahalgam in the Liddar

Valley and Gulmarg in the Ningle Valley. Outwash is generally found in close association with the moraines. All of the sediments downvalley from these points mapped previously as moraines are better explained by non-glacial processes.

The classification of sediments described above was tested using particle-size and particle-shape data. For both clast and grain roundness, pairwise Kolmogorov-Smirnov tests of association were carried out between each pair of samples. For the particle-size data, particle-size distribution envelopes, ternary diagrams of sand, silt and clay content, and bivariate scattergrams of particle-size statistics were plotted. Pricipal components analysis (PCA) was carried out using the SPIDA statistical package (appendix 12). Plots of the 1st and 2nd principal components were then drawn to look for groupings of samples of the same inferred origin.

The results of the pairwise Kolmogorov-Smirnov tests for the clast roundness are shown in figure 50 and those grain shape data in figure 51. These matrices show, quite clearly, that neither shape index entirely supports the conclusions drawn on the basis of field observations.

For the particle-size data, envelopes are shown for each type of sediment in figures 65 to 67. The bivariate scattergrams and ternary diagrams of percentage sand, silt and clay are shown in figures 52 to 64. All of these diagrams show that there is no tight cluster of coordinates for samples of similar origin. Particle-size distribution



Dark shading: significant at 1% level
Light shading: significant at 5% level
No shading: not significant

Figure 50. Matrix showing the results of pair-wise Kolmogorov-Smirnov tests on clasts.

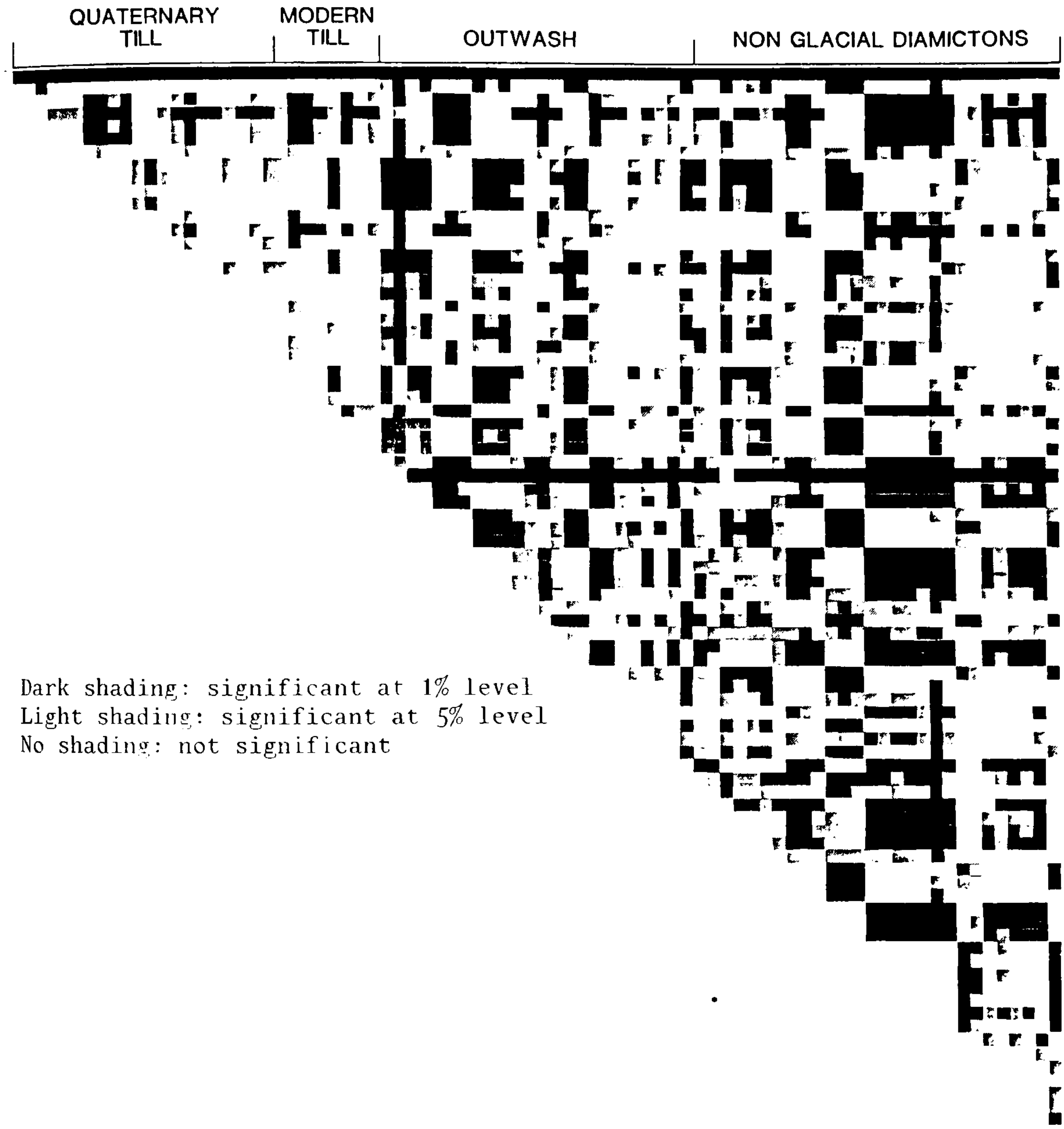


Figure 51. Matrix showing results of pair-wise Kolmogorov-Smirnov tests on sand grains.

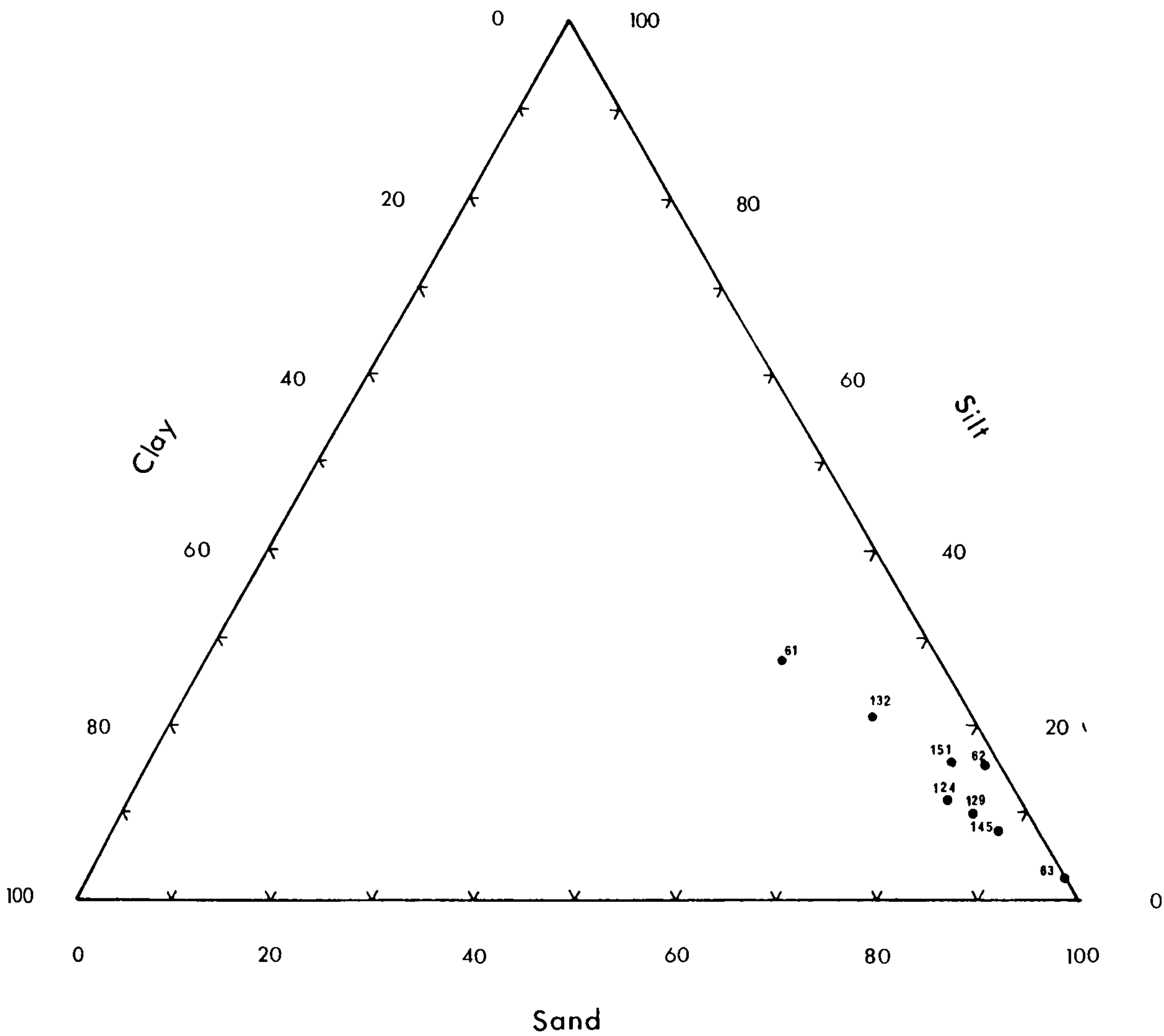


Figure 52

Ternary diagram of percent sand, silt and clay: modern tills.

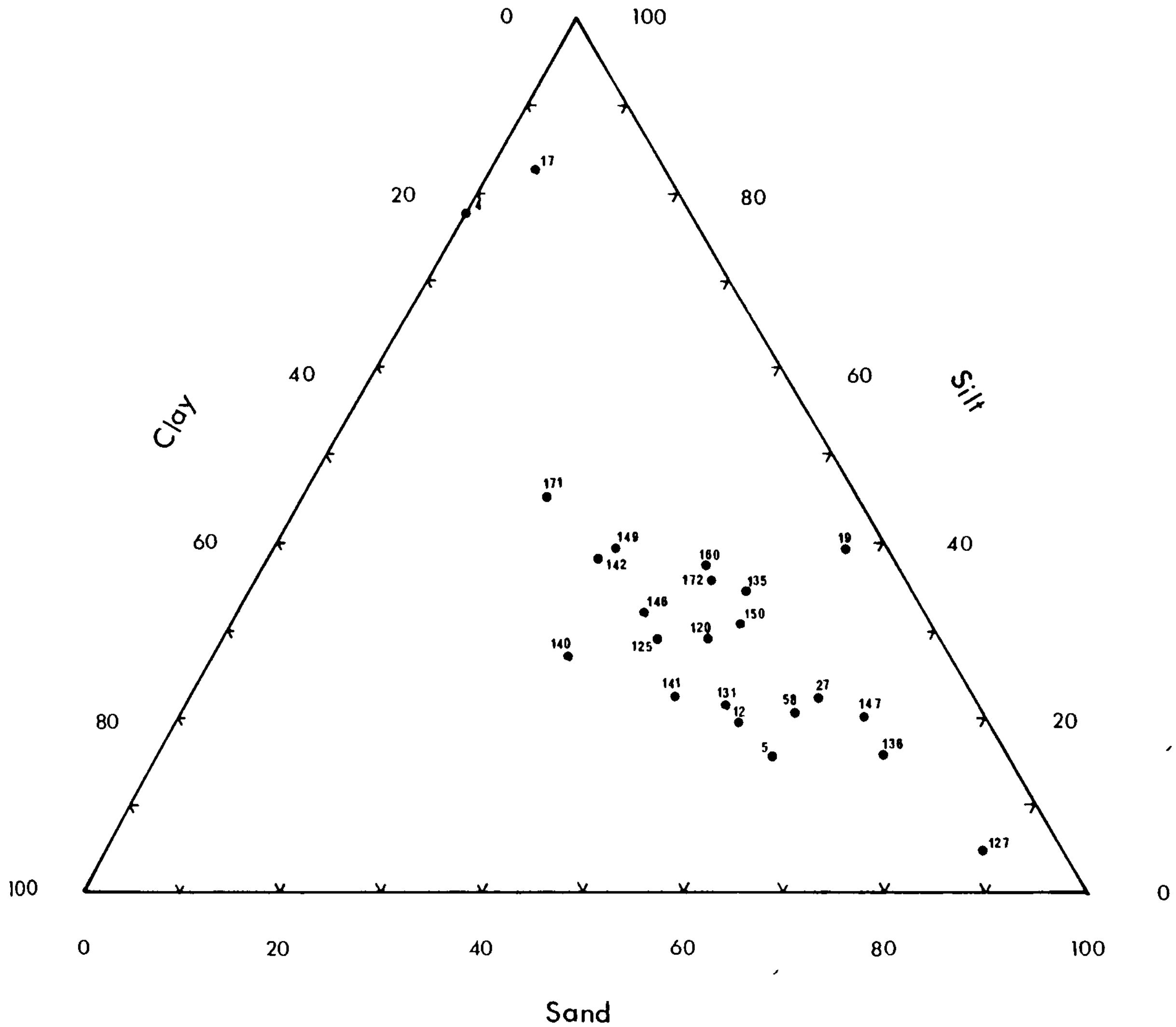


Figure 53

Ternary diagram of percent sand, silt and clay: Quaternary tills.

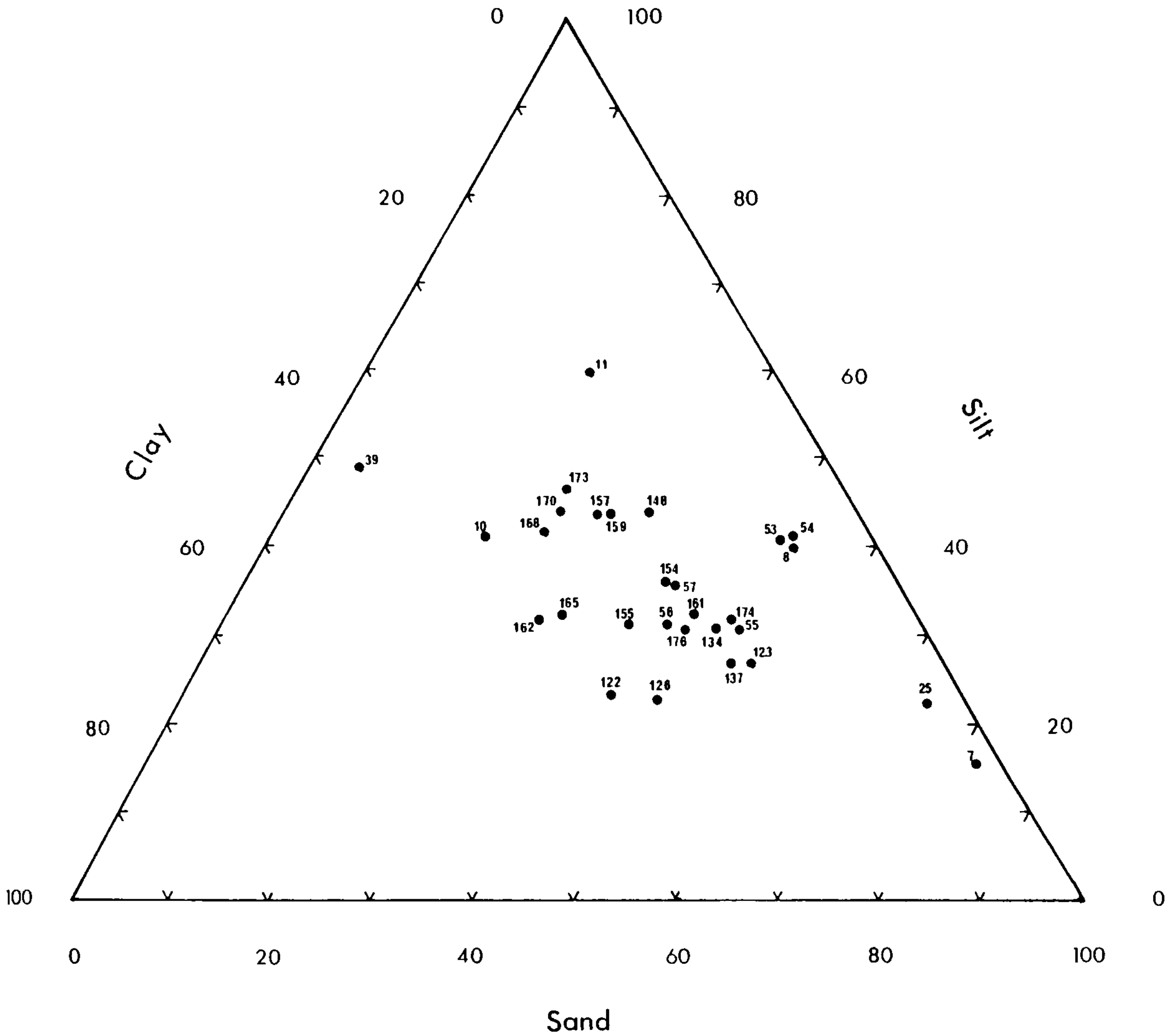


Figure 54

Ternary diagram of percent sand, silt and clay: non-glacial diamictons.

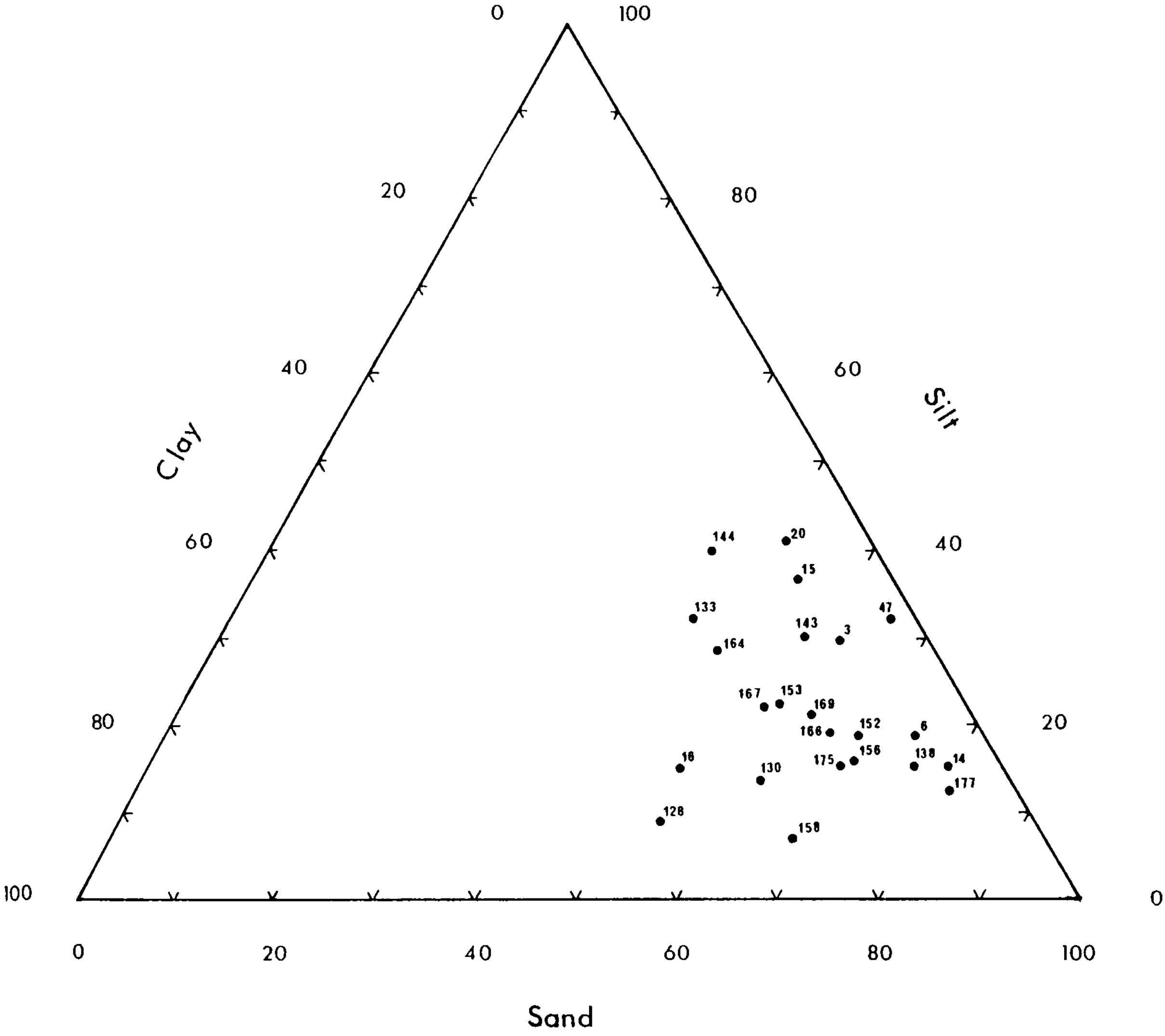


Figure 55

Ternary diagram of percent sand, silt and clay: Outwash.

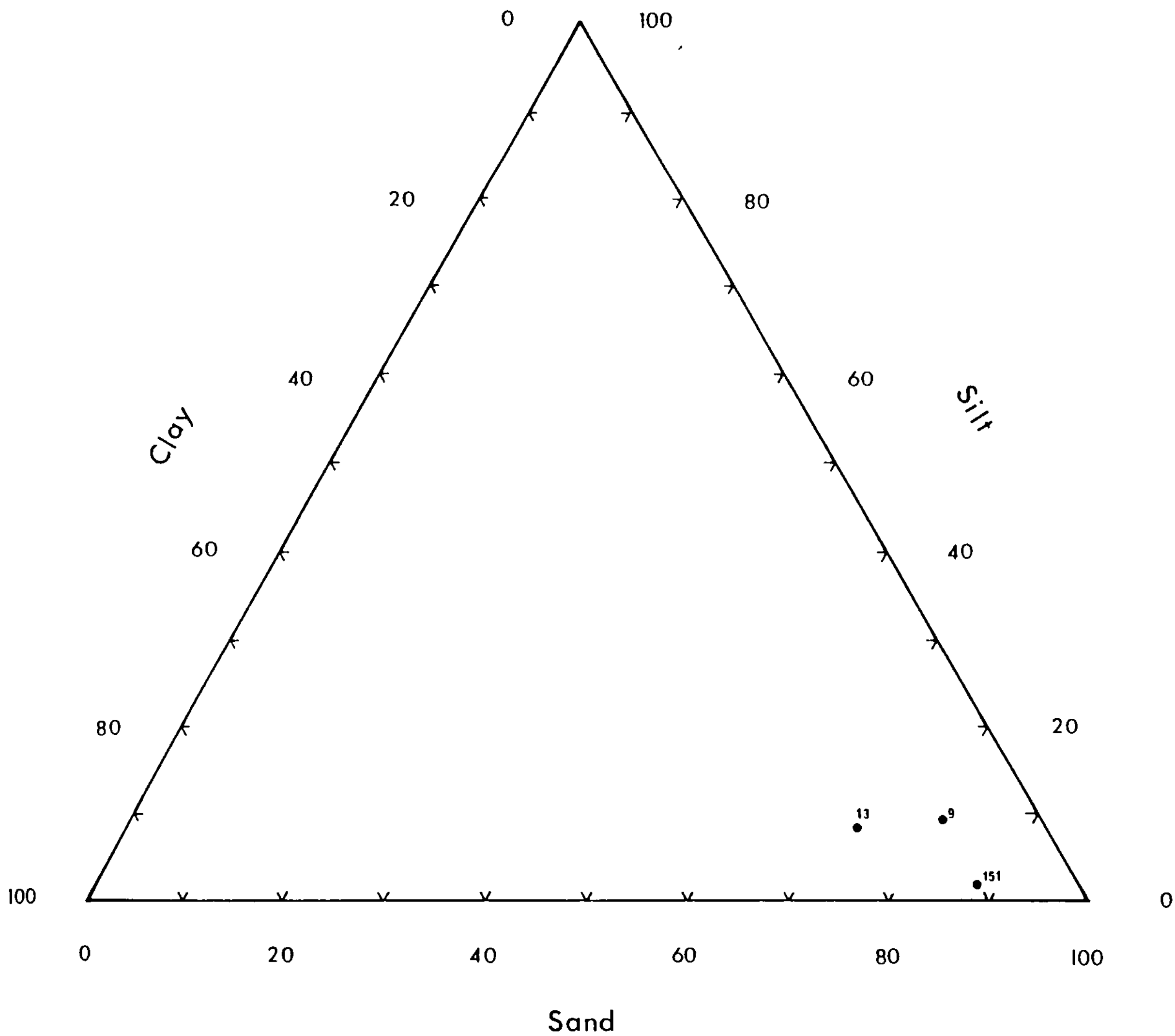


Figure 56

Ternary diagram of percent sand, silt and clay: upper Karewa gravel.

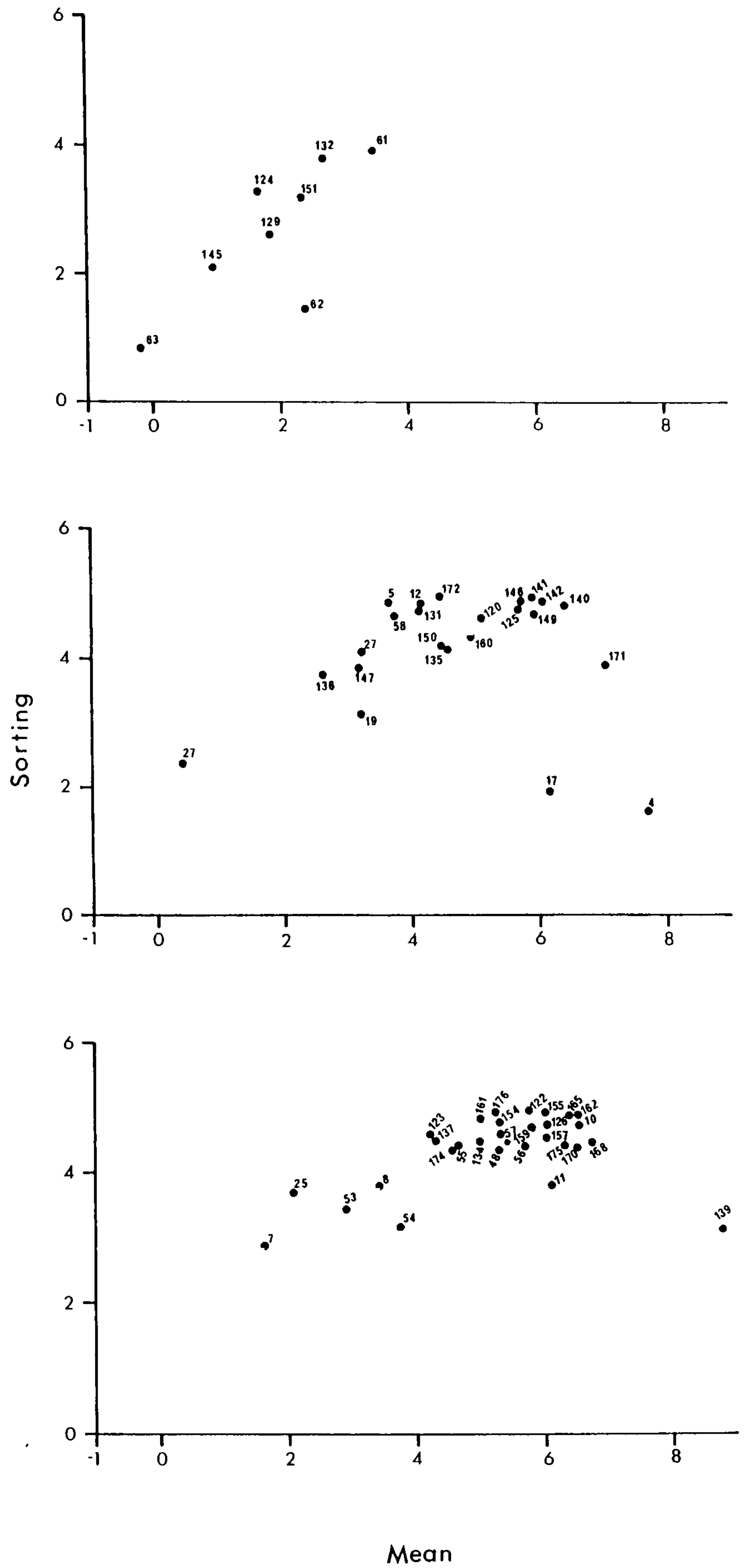


Figure 57

Bivariate scattergram of mean and sorting: modern tills (top) Quaternary tills (middle) and non-glacial diamictons (bottom).

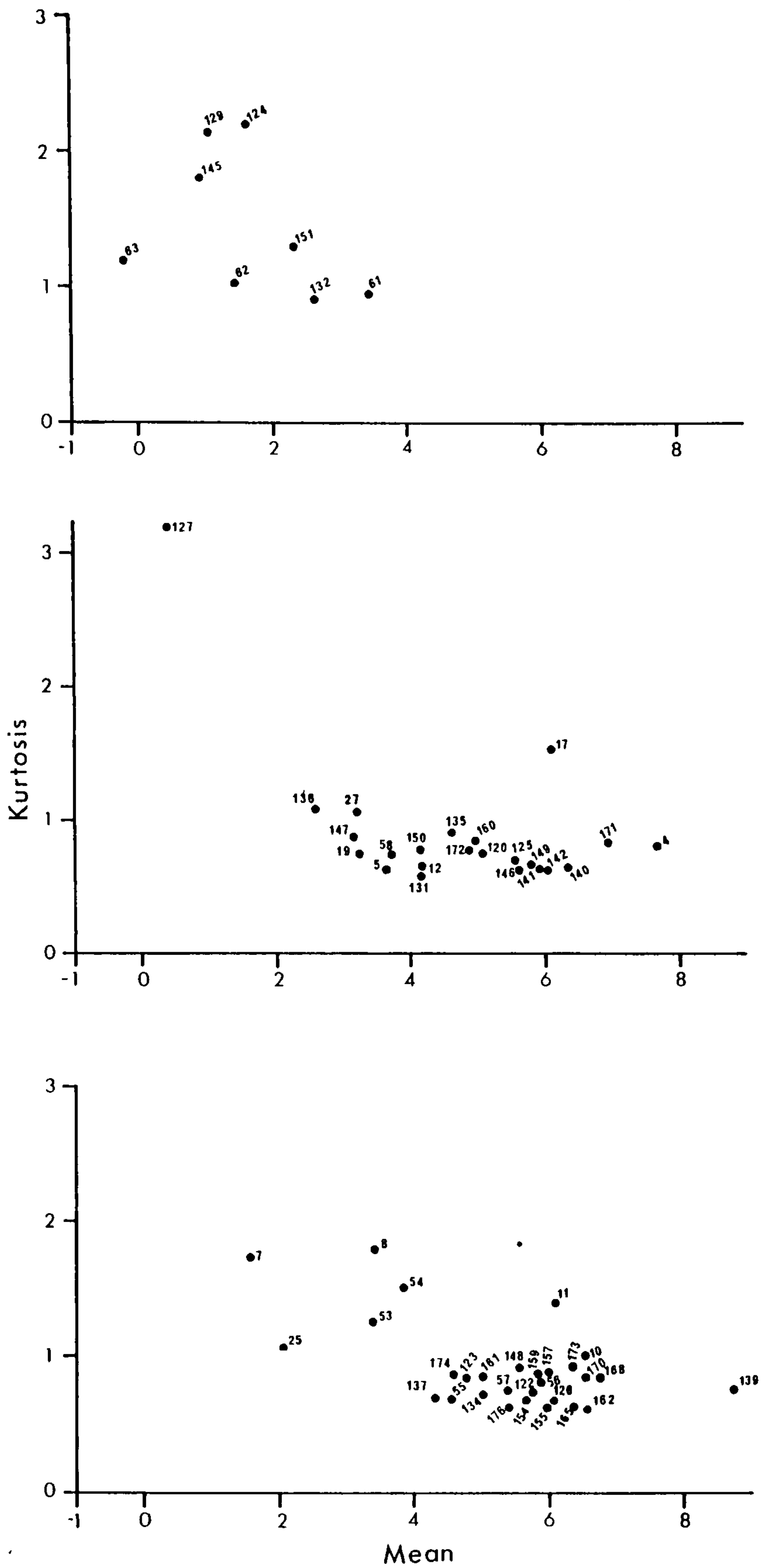


Figure 58

Bivariate scattergram of mean and kurtosis: modern tills, (top) Quaternary tills (middle) and non-glacial diamictons (bottom).

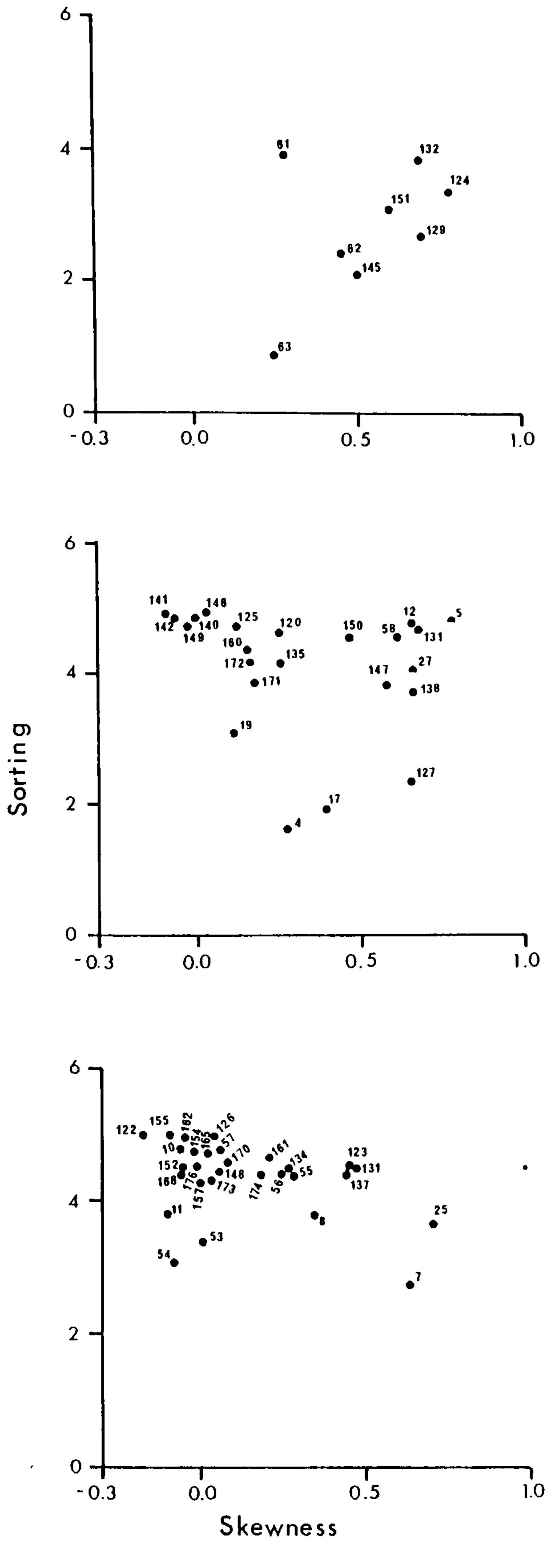


Figure 59

Bivariate scattergram of Skewness and sorting: modern tills (top), Quaternary tills (middle) and non-glacial diamictons (bottom).

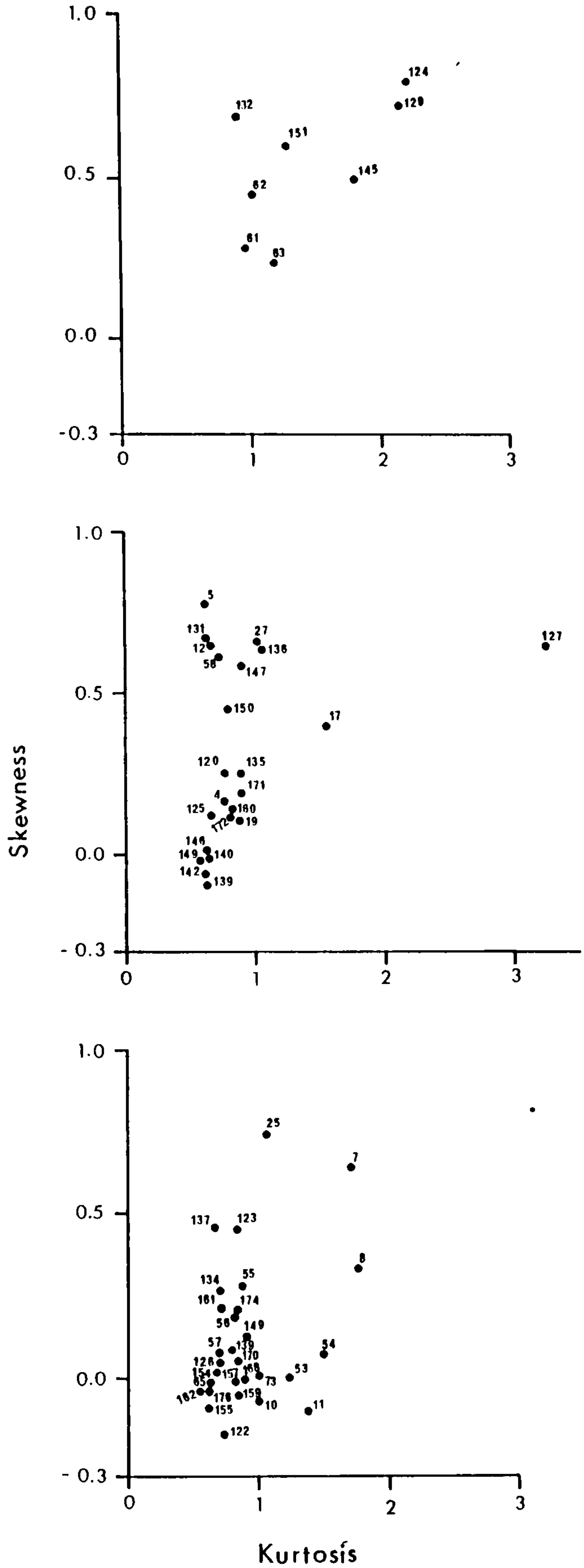


Figure 60
Bivariate scattergram of Kurtosis and Skewness: modern tills (top), Quaternary tills (middle) and non-glacial diamictons (bottom).

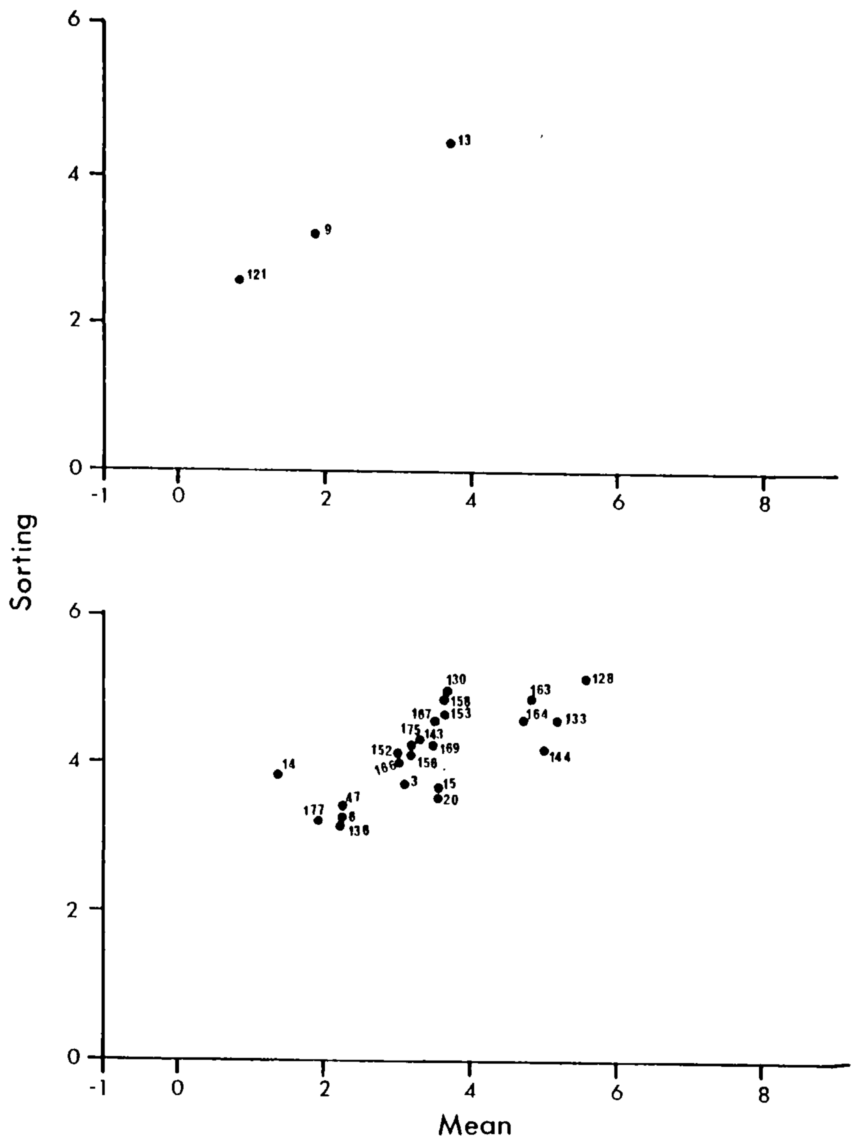


Figure 61

Bivariate scattergram of mean and sorting: Outwash (bottom), and upper Karewa gravel (top).

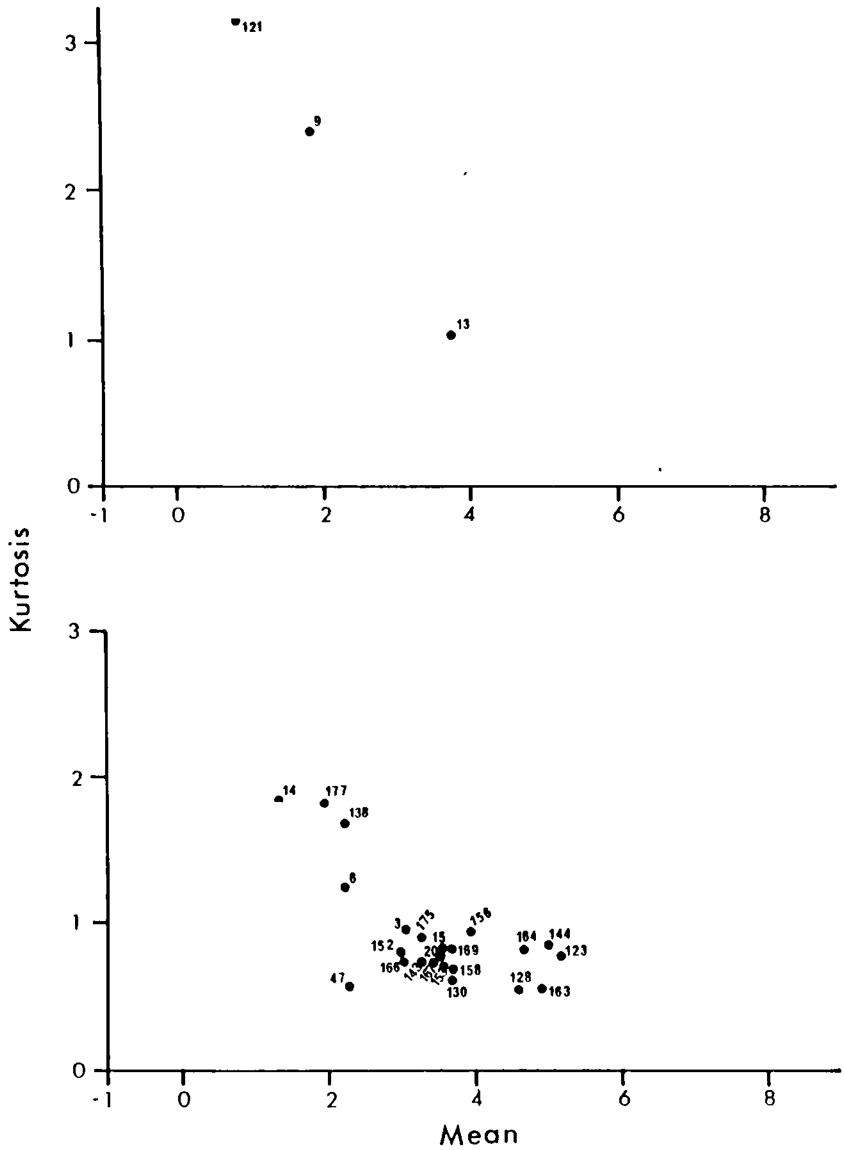


Figure 62

Bivariate scattergram of mean and Kurtosis: Outwash (bottom) and upper Karewa gravel (top).

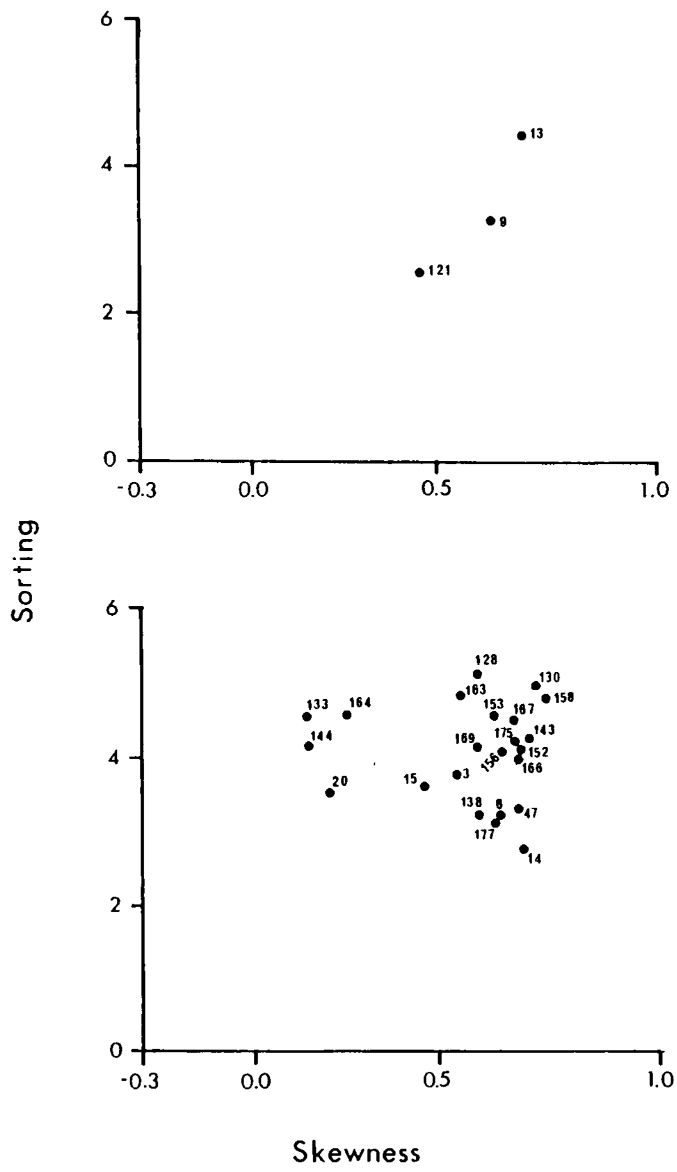


Figure. 63

Bivariate scattergram of Skewness and sorting: Outwash (bottom) and upper Karewa gravel (top).

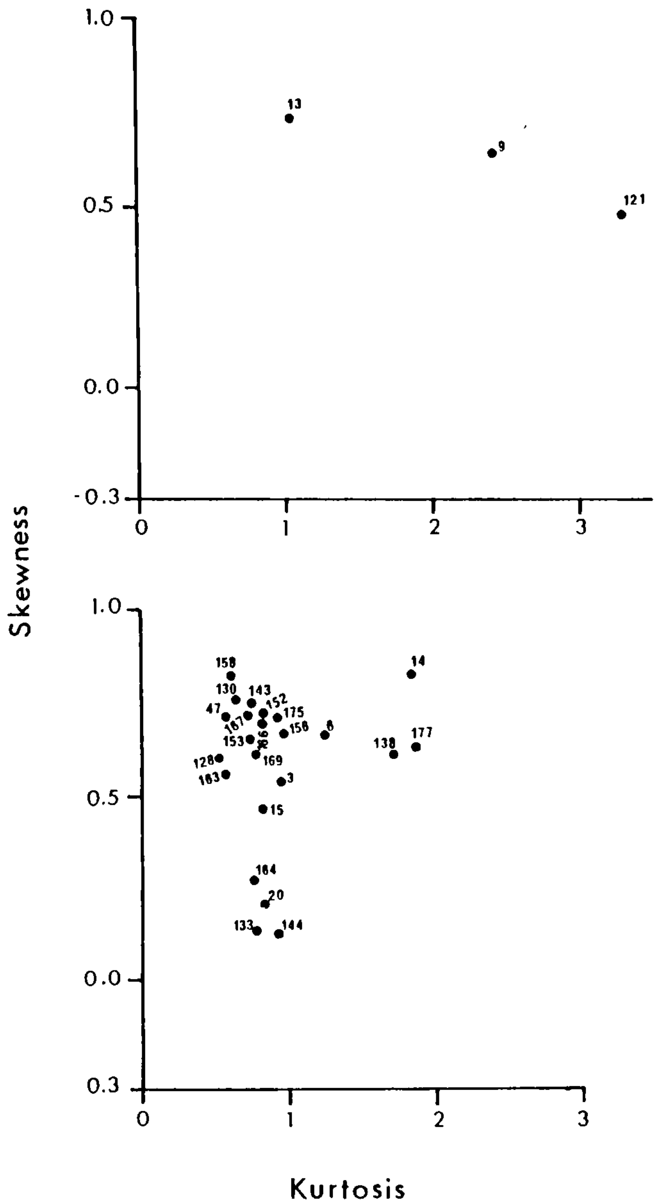


Figure 64

Bivariate scattergram of Kurtosis and Skewness: Outwash (bottom) and upper Karewa gravel (top).

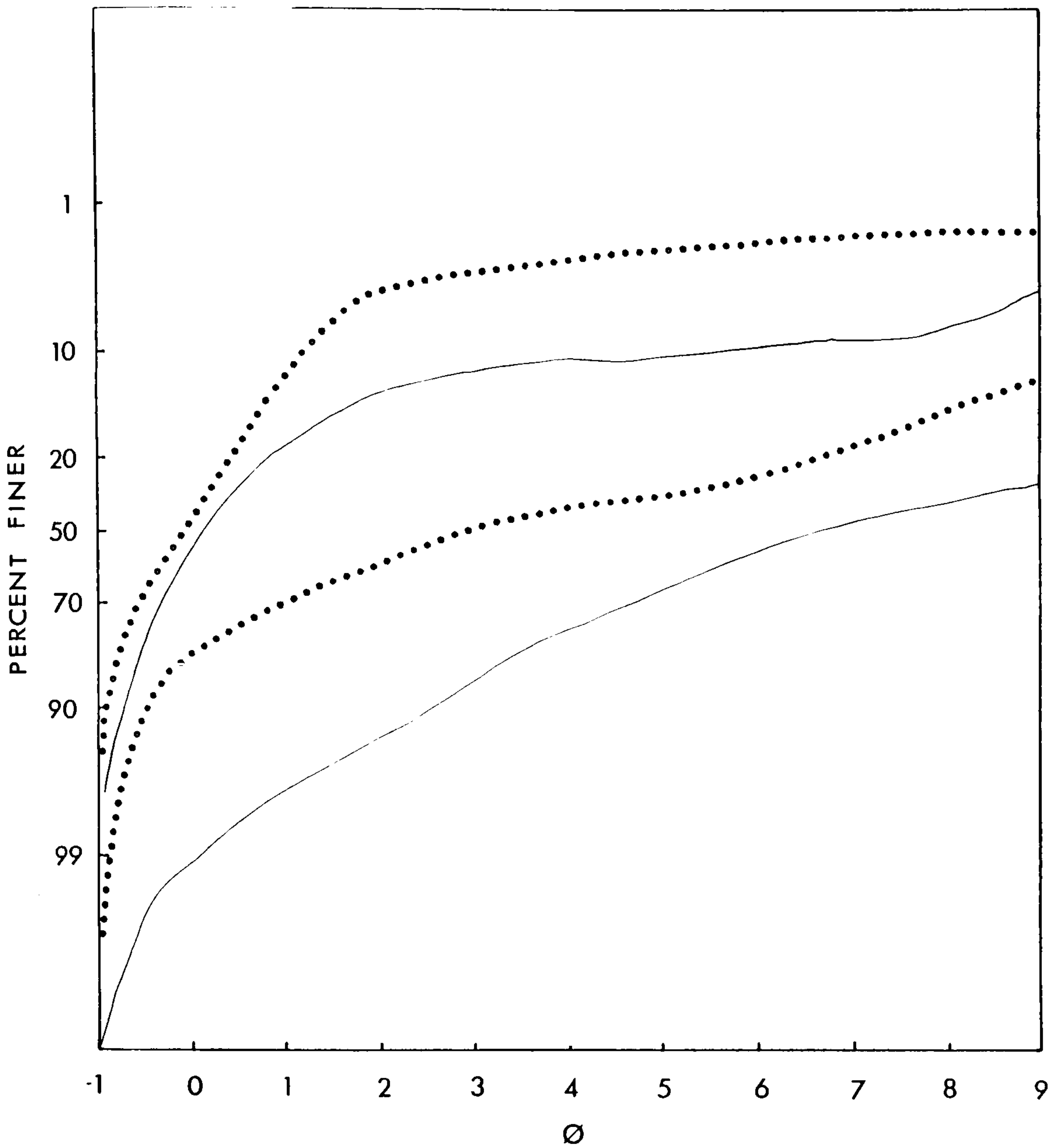


Figure 65

Particle-size envelope: modern tills (dotted line) and Quaternary tills (solid line).

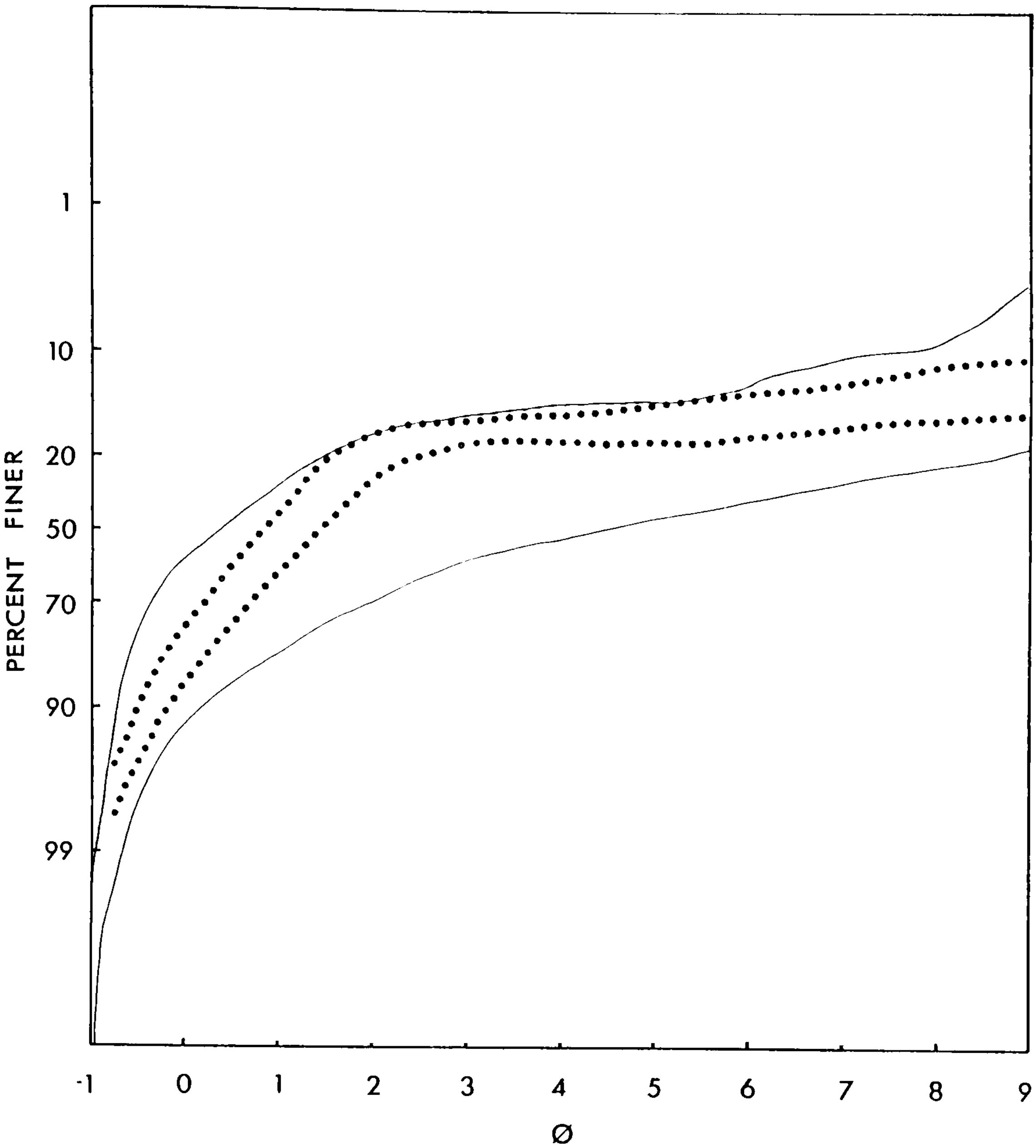


Figure 66

Particle-size envelope: Outwash (solid line) and upper Karewa gravel (dotted line).

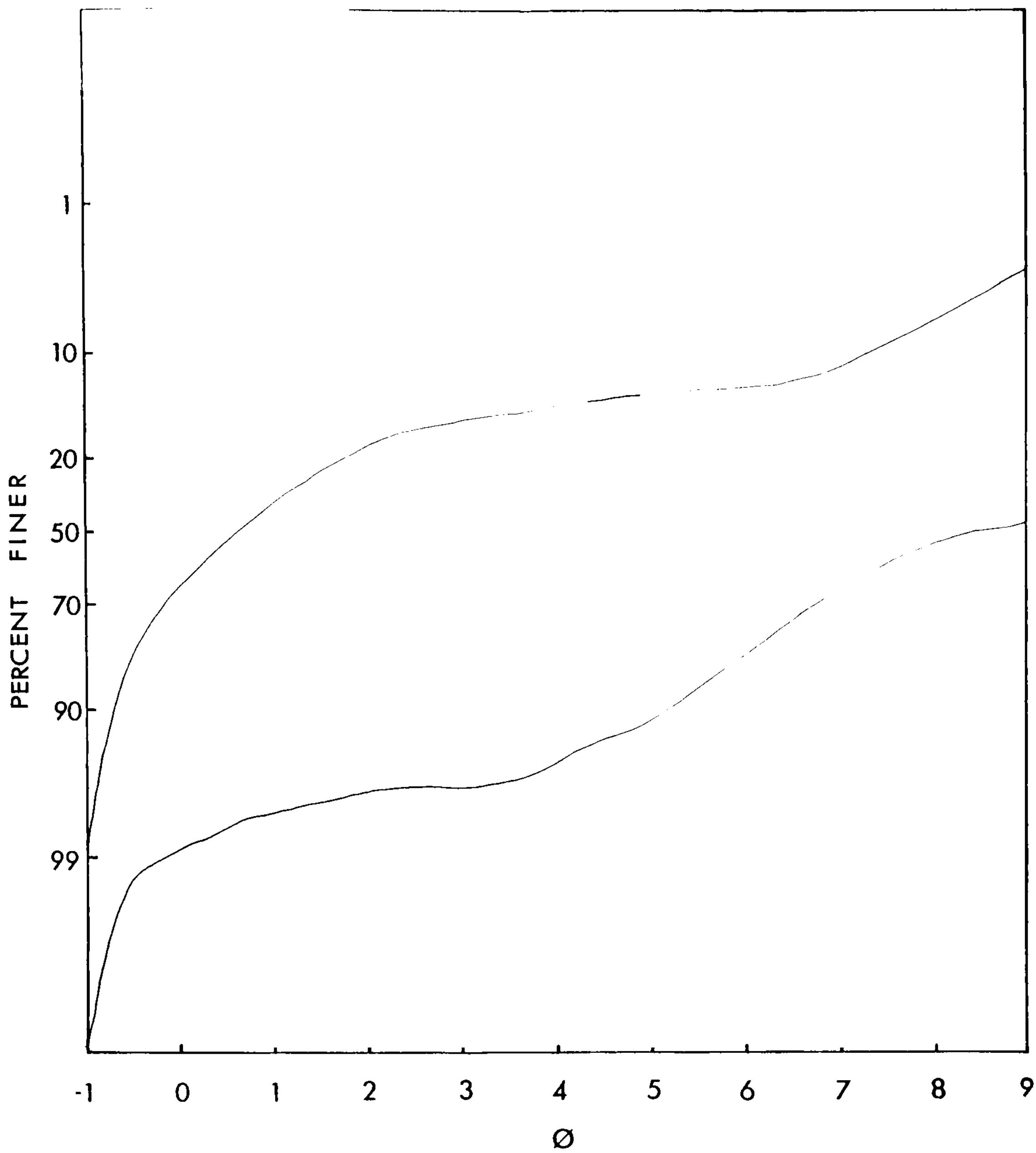


Figure 67

Particle-size envelope: non-glacial diamictons.

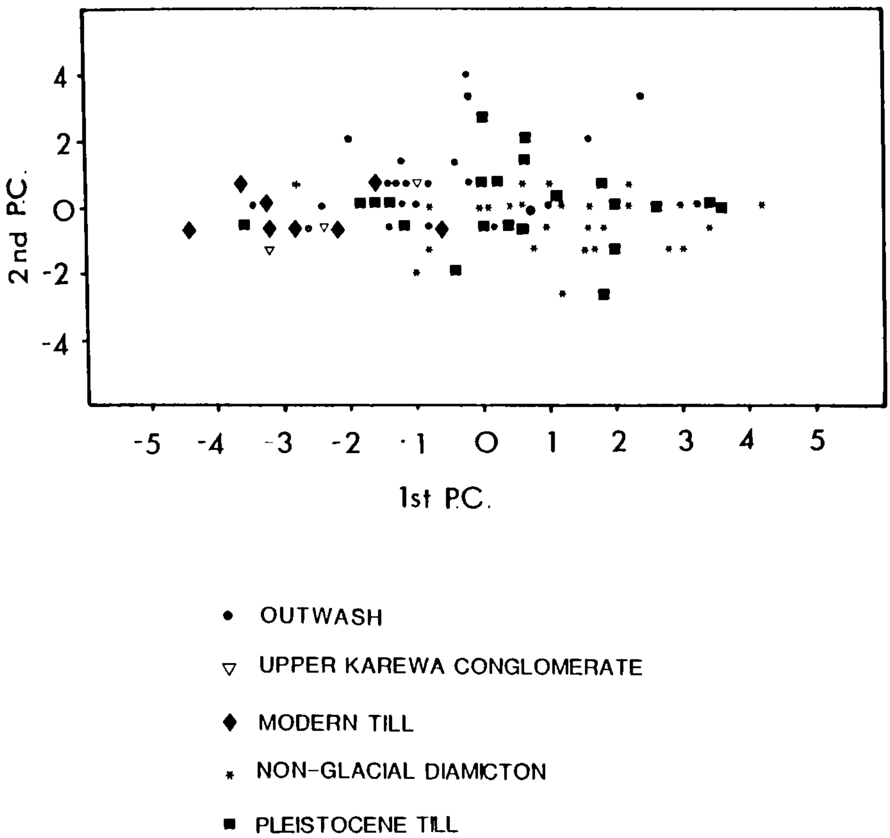


Figure 68

Plot of the 1st and 2nd principal components for particle-size data from diamictons.

envelopes also show considerable overlap. PCA revealed that the mean particle size and sorting accounted for the most variance in the data. Plots of standardised scores for the 1st and 2nd principal components (figure 68) do not distinguish between the inferred groupings either.

On the basis of any one of the above techniques, it is clearly not possible to lend full support to the field classification of diamictons. There are four reasons why this may be so and each requires discussion before firm conclusions can be drawn regarding the origin of the sediments.

1. Error in the field classification. However, the morphological distinction between tills, outwash and mass-movement sediments was always very pronounced. This was partly supported by clast and grain-shape data. Particles in mass-movement deposits tended to be very angular, those in tills subangular to subrounded and those in outwash subangular to rounded. It is not thought that errors in field classification explain the apparent disparity between field and laboratory data.

2. The greater influence on sedimentology by source-area geology than differences in process. Rock characteristics will determine the rate and degree of shape alteration during transport. They will also affect the size-distribution of particles produced during erosion. However, the effect of source on sedimentology is very difficult to test, since the bedrock-mapping of Kashmir is quite generalized. Panjal Trap forms a major component of most of the clast samples, but tests showed that

there was a statistically significant effect of lithology on clast roundness for the rock-types encountered in Kashmir (see appendix 13). Some of the differences between samples may, therefore, be lithologically controlled. The effect of source on grain shape and particle-size distribution is difficult to test and must remain undetermined.

3. Differences in the nature and degree of each individual process. Differences in the sedimentological properties of tills depending on the transport path through the glacier and the mode of deposition were reviewed earlier in this chapter. It is significant, for example, that the modern tills are coarser-grained than the Pleistocene tills, even when samples from only one valley are considered. This can be explained by the larger and thus more active Pleistocene glaciers which would have transported debris further from source than modern and Late Holocene glaciers causing greater comminution of fines. For both mass-movement sediments and outwash, the sedimentological characteristics will depend on the distance and rate of transport. Thus, proximal outwash, for example, will tend to contain more angular particles and have a higher silt and clay content than distal outwash. In mass-movement deposits, clast and grain rounding will increase with distance of transport. Large variations within each sediment type could, therefore, be a reflection of variations in the magnitude of the processes.

4. Particle size and particle shape are poor indicators of process. There is no firm agreement amongst workers as to the reliability of

sedimentological information in this respect. Undoubtedly, sedimentological chracteristics are responses to energy conditions rather than specific geomorphological processes and similar energy conditions can occur in rather different sedimentary environments. When this is added to variations in source-area and variability in the magnitude of processes operating in each broad sedimentary environment, the results obtained in this study are perhaps not surprising.

Observations in the accessible, lower reaches of the Rembiara and Vishav Valleys on the Pir Panjal flank revealed no undoubted glacial deposits. Although the surface expression of the landforms in the lower Rembiara Valley, near Hirpur, is hummocky, exposures revealed that the sediments are clast-supported diamictons, with predominantly subrounded to rounded clasts in a sandy matrix. Exposures in this gravel showed that it was continuous with the upper Karewa conglomerate in the Rembiara valley. It is therefore not a till and the hummocky morphology is better explained by postdepositional reworking of fluvial deposits rather than glacial deposition. In the lower Vishav Valley, much of the depositional evidence has been eroded away. Below the gorge, extensive gravel units interbedded with fine-grained sediments occur. These sediments have every appearance of the upper part of the lower Karewa overlain by upper Karewa conglomerate, as seen elsewhere along the Pir Panjal front. Thus, there is no clear depositional evidence for glaciation in the lower Vishav Valley.

The absence of glacial deposits in the lower reaches of the Rembiara and Vishav Valleys accords well with the lowermost occurrence of moraines in the other valleys studied. However, further information regarding the former glaciation of the south-western part of Kashmir will remain unavailable until the upper reaches of these valleys become open for research.

6.4 Mapping and Dating of Glacial deposits.

a. Introduction.

The next stage in revising the glacial history of Kashmir involves the subdivision and dating of deposits diagnosed as glacial in origin. This is necessary to determine the timing and extent of glacial advances and for the correlation of reconstructed advances in Kashmir with those in other areas. Suitable material for radiometric dating is not often found in association with glacial sediments. As such, a set of techniques known collectively as relative dating, or RD, has often been used to subdivide glacial sediments in terms of their age. Radiometric dates are, however, extremely useful if suitable material can be found. Such dates place temporal constraints on glacial sequences whereas RD only indicates relative age. Radiometric dates are vital for correlation with glacial sequences elsewhere and with other forms of evidence for climatic change.

b. Relative-Dating Techniques.

RD is based on the premise that rock weathering, soil development and biological growth are time-dependent. It also assumes that effects of non-temporal factors that govern these processes can be minimised, so that any difference in RD measures is truly a reflection of difference in age. The destructive nature of glacier advances means that glacial deposits will often only be preserved if subsequent advances are of lesser extent. Thus, the occurrence of deposits from more than one ice advance overlying each other in vertical sequence is rare, and the standard stratigraphical procedure of superposition cannot often be used to determine relative age.

RD techniques fall into four categories: measures of surface or subsurface boulder weathering; measures of soil formation and the accumulatiom of other, non-glacial, surficial sediments; measures of landform morphology and postdepositional alteration; and measures of vegetation growth on the surface of the deposit. Moraines and, less commonly, outwash, have been studied using RD techniques. Most RD work has been undertaken in North America, particularly in the Western Cordillera, Alaska and Arctic Canada (eg. Birkeland, 1973; Carroll, 1974; Burke and Birkeland, 1979; Péwé, 1975). However, a number of studies have been conducted elsewhere including New Zealand (eg. Chinn, 1981; Gellatly, 1984), East Africa (eg. Mahaney, 1984), South American Andes (eg. Porter, 1981; Wayne, 1984), the European Alps (eg. Porter and Orombelli, 1982), Northern Pakistan (Porter, 1970; Derbyshire <u>et al</u>., 1984) and Nepal (Iwata, 1976).

Individual RD techniques vary considerably in ease of data collection, subjectivity, precision and temporal range. The use of RD in glacial stratigraphy and the choice of method used in this research are explored in more detail in the rest of this section.

c. RD and Glacial Stratigraphy.

Andrews and Miller (1980, p.264) applied the Jenny soil equation to RD measures as follows:

State of Weathering = f(T,C,P,V,R)

Where T = time, C = climate (principally precipitation and summer temperature), P = parent material, V = vegetation, R = relief.

If weathering and soil development are to be used as RD measures, it is essential to minimise variations in non-temporal factors. Several authors have considered the influence of non-temporal factors on rock-weathering rates.

Colman and Pierce (1981) showed that the environmental factors affecting soil development were the same as those controlling rates of rock weathering. By careful site selection, they were able to demonstrate that age accounts for the major amount of rock weathering as indicated by the thickness of weathering rinds. Similarly, Chinn (1981), in a study of rock weathering rinds on dated landslide deposits, found that age explained 98% of the variation in rind thickness. Porter (1975c) also found that, whereas there was some overlap in the range of rind thicknesses on deposits of different ages, the standard deviations of values were mutually exclusive. Measurements of weathering-rind thicknesses tend to be precise and objective. However, weathering rinds are only found on certain types of rocks, principally granites and basalts. Many other measures of rock weathering have been used in RD studies, but the arguments put forward above apply as much to these as to weathering rinds.

Many studies have shown that the rate of weathering of a rock surface decreases with time, even without changes in environmental controls. Ollier (1969) pointed out that the rate of weathering of a rock will eventually reach zero, due to the protecting effect of the weathered layer. This observation suggests that the discriminating power of RD mthods based on rock weathering will decrease with time. However, the actual temporal span of any RD method will depend on the factors in the Jenny equation, particularly rock type and climate.

In an RD study, it is necessary to hold non-temporal factors constant between sites. This is usually achieved by careful site selection. In the case of climate, it is generally assumed that this will be constant within the area of study, providing sites with similar orientation and exposure are chosen. The rock-type factor in the Jenny equation can be held constant simply by making measurements on a single lithology, although this is not straightforward in valleys that have complex bedrock geology. Vegetation and topography can be held constant by choosing sites that are in similar topographic positions and have similar vegetation cover.

Sampling in an RD study, like in many other aspects of Quaternary geology, is often governed by the availability of exposures. Whereas surface boulders and soil cover on moraines can be examined easily, the examination of unweathered till is often confined to sections exposed by human activity or natural erosion, particularly where the till has been oxidized to a considerable depth. As such, true random sampling is rarely possible in RD studies. This has obvious implications for the statistical analysis of the data.

Meierding (1984) argued that the average of the highest 5 to 10 values of any RD measure has more meaning in terms of the age of a deposit and its correlation than the average of all the available data of randomly selected values. This is because, whereas many processes may make a deposit appear younger than it actually is, few will make it apparently older. The removal of weathered material, a process that would make a deposit appear younger than it is, is common on glacial landforms. The incorporation of older, previously weathered material, which could conceivably make a deposit appear old, is less likely. This process of data maximisation is, as Meierding (1984) points out, detrimental to any subsequent statistical analysis, since the data will be biased and the search-time for high values of RD measures will depend largely on preconceived ideas about the age of a deposit.

Over the last 15 years, there has been a tendency to use more than one RD measure in a study: the so-called 'multi-parameter' approach. There are several reasons to favour this approach in glacial stratigraphy. Firstly, the subjective nature of many RD measures combined with the effects of non-temporal factors yields data with a large variance. By using more than one measure, it is possible to increase tha amount of available information and so strengthen the conclusions drawn. This is well summarized by Blackwelder (1931, p.880):

> 'of all the criteria, only a few can usually be applied in any one place. One of them alone affords only a tentative opinion, but when several of them all point to the same conclusion, confidence is much strengthened.'

Secondly, criteria which work well in one valley may have poor discriminating power in another. For correlation between valleys, it is clearly necessary to use more than one measure. Thirdly, since individual RD measures have different timespans, the use of more than one measure will increase the temporal span of RD as a whole.

Meierding (1984) is possibly the only recent writer to question the validity of using more than one measure in an RD study. Whereas Burke and Birkeland (1979) argued that fewer samples are required in a study involving more than one RD measure, Meierding (1984) maintained that this is only the case if all of the measures confirm the age of the deposit. He argued that, if some of the measures provide conflicting results, the reason may simply be that too few samples have been taken to overcome the large variance in the data. Meierding proposed that individual research workers should concentrate on generating large data sets using single RD measures. This suggestion is applicable to the American Rocky Mountains, with respect to which it was made. In this range, many RD studies have been undertaken and the framework of the glacial stratigraphy is well understood. The results from previous studies provide a useful basis for evaluating the most applicable techniques to the area. However, in areas where the glacial stratigraphy is poorly known, it is essential to use more than one RD measure so that a stratigraphical framework can be determined and the most useful RD measures identified for future work. The value of this approach in such instances is exemplified by the studies of Porter (1970) in Swat Kohistan, Pakistan and Derbyshire _et al._, (1984) in the Hunza Valley, Karakoram range.

Once RD data have been collected, it is necessary to subdivide the deposits and decide upon the degree of variation that should allow for the definition of major glaciations and advances of lesser order. Birkeland _et al._ (1979) contended that only those glacial deposits that have been dated numerically or characterized by adequate RD data should be given stratigraphic names and assigned type localities. They also stated that second-order glacial advances (stadials) within first-order advances (glacials) should not normally be named unless there is adequate RD evidence to do so. They recommended that letter-designations should be adopted for deposits of different ages, since this avoids uncertain correlation with better-studied type areas and prevents the accumulation of a large number of informal names.

The use of formal stratigraphic nomenclature, as set out by the American Commission on Stratigraphic Nomenclature (1961) is, according to Birkeland _et al._ (1979) unworkable for mountain glacial deposits. Lithostratigraphic units are hard to define, since most deposits within a valley will be lithologically identical. Chronostratigraphic units are hard to apply since glacial deposits are always time-transgressive, superposition at an exposure is rare and numerical ages uncommon. Since Birkeland _et al._ wrote, there have been a number of attempts to rectify this problem by incorporating the degree of postdepositional alteration, which is the basis of RD, into formal stratigraphical procedure (eg. North American Commission on Stratigraphic Nomenclature, Draft Stratigraphic Code, 1981, pp.33-35).

The problem of defining first- and second-order glacial advances using RD data has not been adequately addressed. From independent evidence, it has been shown that first-order glaciations occur on a 100 ka cycle. However, the degree of weathering, soil development and biological growth that occurs during this period will depend on the factors in the soil equation. Since the magnitude of these factors will clearly have fluctuated since the deposition of a sedimentary unit, and because long-term rates of weathering, soil development and biological growth are poorly known, there are no predetermined values of RD measures that can be quoted to represent first- or lesser-order magnitude glacial advances. Burke and Birkeland (1979) divided glacial stratigraphers into 'splitters' and 'lumpers'. Splitters divide the glacial sequence as much as possible, based on subtle variations in RD measures. Lumpers, however, require gross changes in RD measures which are demonstrable in several valleys, before they subdivide the sequence. Burke and Birkeland suggested that a doubling in the numerical value of an RD measure is required before deposits are assigned to separate first-order glaciations. However, they conceded that this generalization is probably not applicable to all measures. Burke and Birkeland (1979) proposed that this approach will separate first-order glacials, but that advances of lower order cannot be recognised with any confidence. Hence, only first-order glaciations are generally named and formalized on the basis of RD.

d. Methods Used in This Study.

The Survey of India 1:63 630 topographic maps were used as the topographic base for this part of the study, but were substantially enlarged for use in the field. Because Kashmir is a politically sensitive area subject to considerable military activity, it was thought unwise to attempt to use surveying equipment other than a compass, clinometer and altimeter. However, detailed surveying was not important for this part of the study, since the main information required was the approximate distribution and down-valley limits of sediment units. Mapping therefore involved fixing the locations of sampling sites and the areal extent of sediment units on the enlarged topographic maps with reference to key local landmarks using the simple instruments listed above. These procedures provided adequate information for this study.

Following reconnaissance in Kashmir, the following RD measures were chosen for further use: the frequency of surface boulders on moraines, the percentage of split surface boulders, the percentage of pitted surface boulders, the depth of pits on pitted boulders, the maximum height of upstanding quartzitic or mafic intrusions above the surface of boulders, the maximum corner angularity of subsurface clasts, thickness of loess on moraines, and the nature and properties of soil development on moraines. The working definitions of these measures and the procedures used in the field are outlined in appendix 14.

The choice of measures was based on several considerations. Initial reconnaissance indicated that the only lithology common to all sites is Panjal Trap. This rock was therefore chosen for all measures involving surface or subsurface boulders. Panjal Trap is a widespread volcanic rock, generally a basic pyroxene-andesite or basalt of quite uniform mineralogy (Geological Survey of India, 1977). It is a very hard, resistant rock. The use of this lithology in RD studies precludes the use of any RD measures designed for granitic rocks, since the two lithologies are dissimilar. The resistance of Panjal Trap appears to have prevented the development of weathering rinds, since none were found even in the oldest deposits. Some of the surface boulders showed signs of weathering interpreted here as pitting and splitting. Although such features may not be the same as those defined for granitic rocks by, for example, Burke and Birkeland (1979), this is not in itself important. It is crucial, however, that a consistent definition is maintained for all of the sites in a study, and that the definition is described adequately. This was achieved in this research by collection of data by one operator. Resistant quartzitic intrusions were seen in boulders at some, but not all, sites. This suggests that there is some lithological variabiltiy in the Panjal Trap.

Soil pits were typically dug on moraine crests. One of the shortcomings of the soil data in this study arises from the variability in the nature of the parent material, the underlying till.

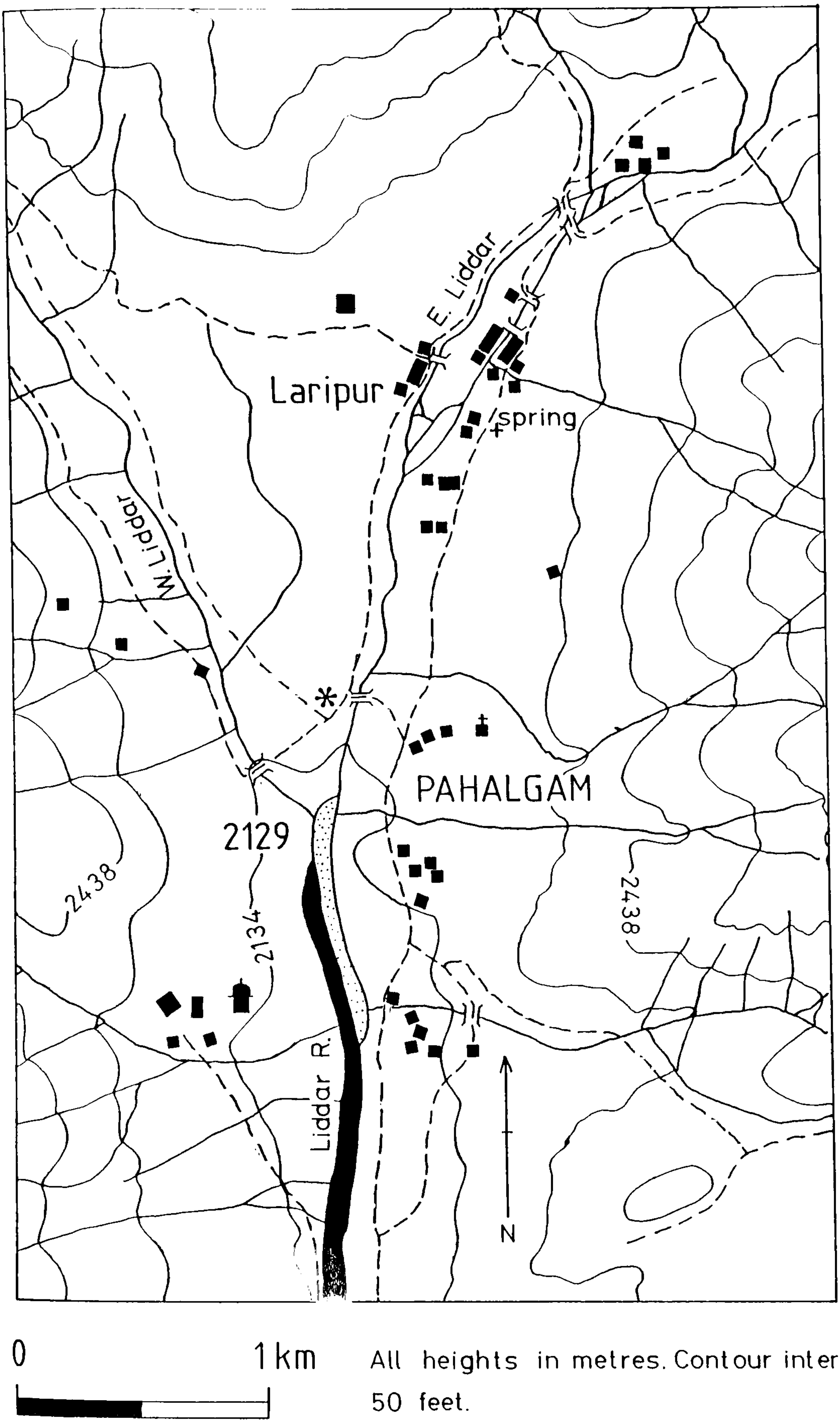
Because of the complex bedrock geology in Kashmir, adjacent moraines may have different sources and hence contrasting lithologies. When comparing sequences between valleys, this problem is heightened. A further problem, pertinent to all RD data, but especially to soils, is that of bioclimatic zoning with respect to altitude. This effect may alter the apparent relative age between older, lower-altitude moraines and younger and higher altitude features, depending on the nature of the climatic gradient. The significance of these problems for the glacial stratigraphy of Kashmir is dealt with in the results section.

Three other, commonly used, RD techniques have been omitted from this study. Lichenometry has a temporal span that limits its use to Holocene deposits. Because no attempt was made to subdivide the Holocene moraines in this study, lichenometry was not used. The development of desert varnish on boulder surfaces has been used as an RD measure in the Hunza Valley by Derbyshire et al. (1984). However, Kashmir is substantially more humid than the Karakoram Range, and desert varnish is not found. Finally, an attempt was made to use the point-compressive strength of boulders, measured using a Schmidt Hammer, as an indicator of weathering and, hence, relative age. Several studies have shown the precision and reproducibility of this technique (eg. Derbyshire et al., 1984; Mathews and Shakesby, 1984). In the present study, however, considerable variation was found in the Schmidt hammer values of hardness (R-values) both on a single boulder and between boulders on a single moraine. The reason for this is not clear. Testing of the Schmidt hammer both before and after making measurements ruled out instrument failure. However, there were large variations in the surface roughness of individual boulders, and it is known that roughness can affect R-values (Williams and Robinson, 1983). Another explanation could be the variable lichen cover found on the boulders. The extent of lichens on many older boulders made it impossible to test bare rock in many instances, and removal commonly disrupted the rock surface, despite the hardness of Panjal Trap. The lichens appeared to have a cushioning effect which would substantially reduce the R-value. Because of these problems encountered in using the Schmidt hammer, this technique was not included as a major part of the study.

e. Radiometric Dating Methods.

Since glacial deposits cannot be dated directly, it is necessary to date suitable material, which may lie stratigraphically above or below glacial deposits or, in certain circumstances, be intercalated between successive glacial sediment-units. In Kashmir, long exposures of glacial sediments are rare. This means that the chances of finding datable material were correspondingly low. However, datable material was found in two localities: at Pahalgam and Aru in the Liddar Valley. Dates on morainic bogs from Toshmaidan (Singh and Agrawal, 1976) and Butapathri (Dodia _et al._, 1984) in the Pir Panjal Range are also of use in the present study.

At Pahalgam (figure 69), the sediments are exposed in a road cutting. Up to 2.5 m of silty clay overlie calcareous till (figure 70). The fine material is interpreted here as loess although it has probably been reworked, since several small stone lines occur in the



All heights in metres. Contour interval 50 feet.

Figure 69

Sample location in the Pahalgam area.

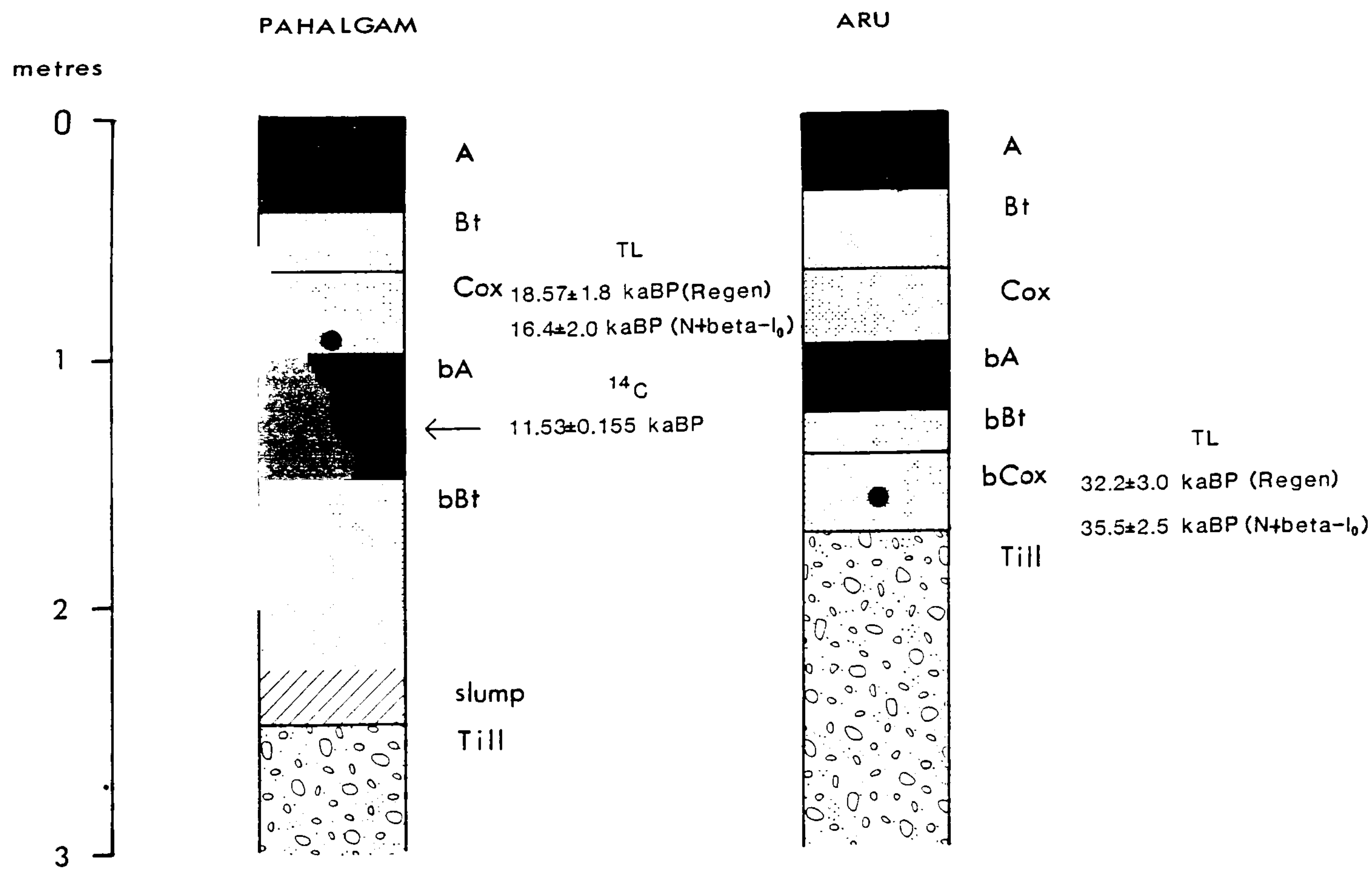


Figure 70
Stratigraphy of dated sections at Pahalgam and Aru.

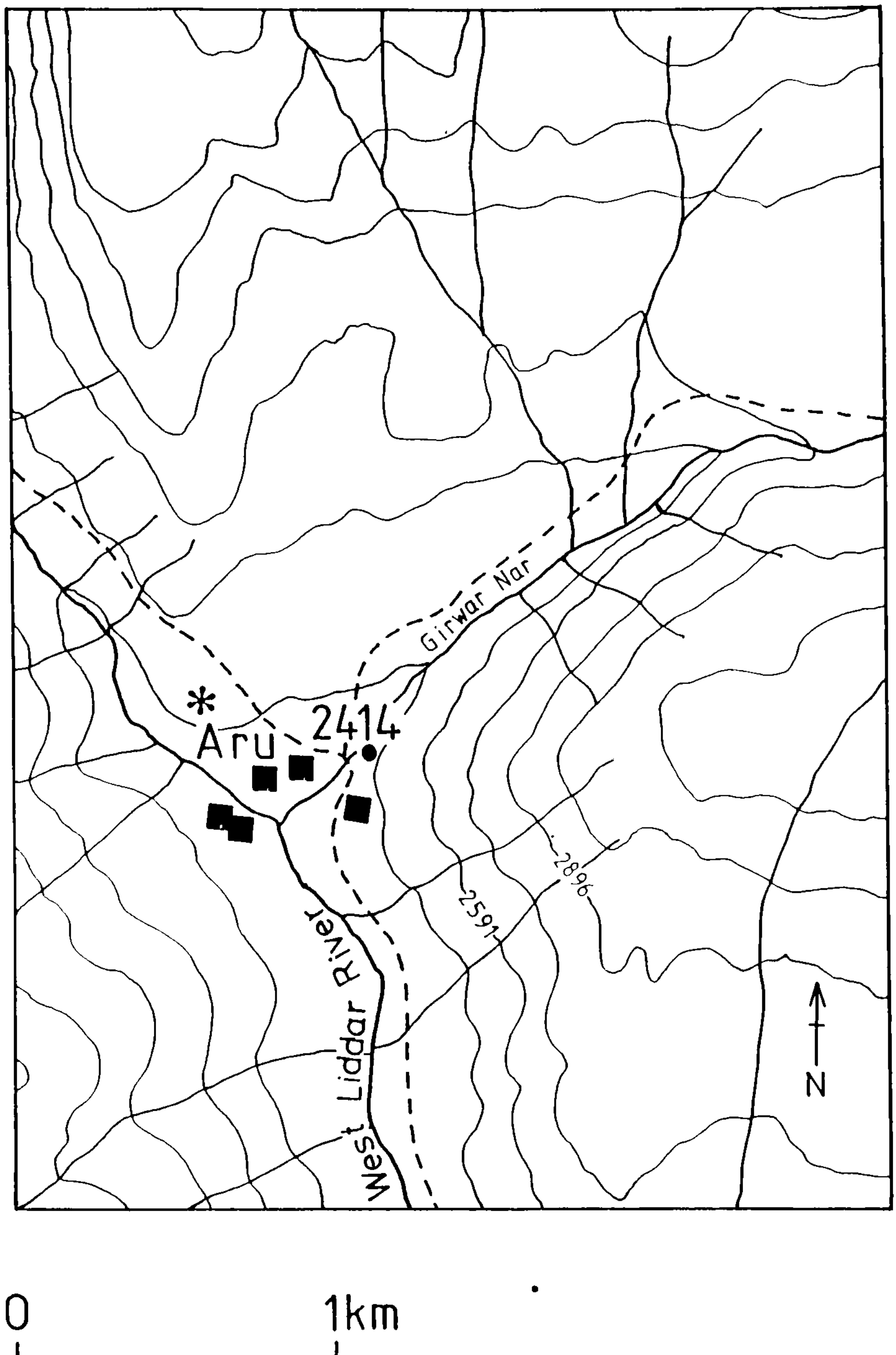


Figure 71

Sample location in the Aru area.

section. About midway in the profile, a distinctive dark horizon occurs, which is interpreted here as a fossil soil: above this is another unit of slightly weathered loess followed by the B and A horizons of the modern soil.

Two samples were taken from the morainic loess for dating purposes. Firstly, a bulk disturbed sample of the fossil Ah horizon was taken for radiocarbon dating. Secondly, an undisturbed block of the slightly weathered loess was taken for TL dating. The TL sample was taken from above, rather than below the fossil soil horizon, since the stratigraphically lower material was highly weathered and would have been less suitable for TL dating.

At Aru (figure 71), loess overlies till and outwash. Signs of soil development are much less clear in the loess at Aru than at Pahalgam and the loess is less well exposed. However, it was possble to establish in hand-dug pits that about 2 metres of loess overlies the glacial sediments, and there is a very weak palaeosol in the profile. Since the palaeosol was poorly developed, no sample was taken for radiocarbon dating. However, a sample of the loess was taken for TL dating (figure 70).

Samples for geochronometric dating purposes are generally collected by standardized procedures. For all dating samples, all slumped material is first removed from the face of the section prior to the removal of about 0.3 m thickness of undisturbed sediment. For radiocarbon samples, about 0.5 kg of material was placed in a thick polythene sample bag using clean tools. The sample was then double-bagged and labelled, prior to shipment to the UK by air. The sample was shipped to the UK within one week of collection and kept in cold storage prior to analysis. The radiocarbon measurements, undertaken at the Godwin Laboratory (University of Cambridge), were carried out on the total organic content of the palaeosol.

For TL samples, the same section-clearing procedures were employed as for radiocarbon-dating samples. However, because of the principles involved in TL dating, it is necessary to shield the sample from sunlight. This was achieved by shading the section while the sample was excavated and by removing an undisturbed block of material. The exposed outer surface of this material was then removed prior to laboratory measurements. The excavated blocks were wrapped in aluminium foil and then sealed in black polythene. All of the samples were returned promptly to the UK for analysis.

f. Results.

Field observations, mapping procedures and sedimentological analyses were used to produce maps of the glacial geology of the Sind, Liddar and Ningle Valleys shown in figures 72 to 74. The up-valley sequence of glacial deposits will be outlined briefly for each valley.

In the Liddar Valley (figure 73), the lowermost occurrence of till is at Pahalgam (plate 34). Terraces of outwash are found close to the valley floor downvalley from this moraine and the village of Pahalgam itself is built on one such terrace. The moraine at Pahalgam is a complex feature. It is composed of compact, calcareous lodgement till on the eastern side and outwash on the western side. The lithologies of these deposits suggest that the source for the till was the East Liddar Valley and that for the outwash the West Liddar Valley, as discussed earlier in this chapter. Moraines were also mapped at Aru (plate 35), Liddarwat (plate 36)and close to the current snout of the Kolahoi glacier (plate 38). The remains of a termino-lateral moraine complex was also seen at Basmai (plate 37), between Liddarwat and the Kolahoi Glacier. Outwash occurs discontinuously in much of the west Liddar valley and around Pahalgam. Where moraines occur, the outwash is generally banked up against them forming terraces. Between Pahalgam and Aru, the outwash is perched on benches which occur up to to 100 m above the floor of the valley. This suggests that incision of the West Liddar Valley has occurred since the deposition of the outwash.

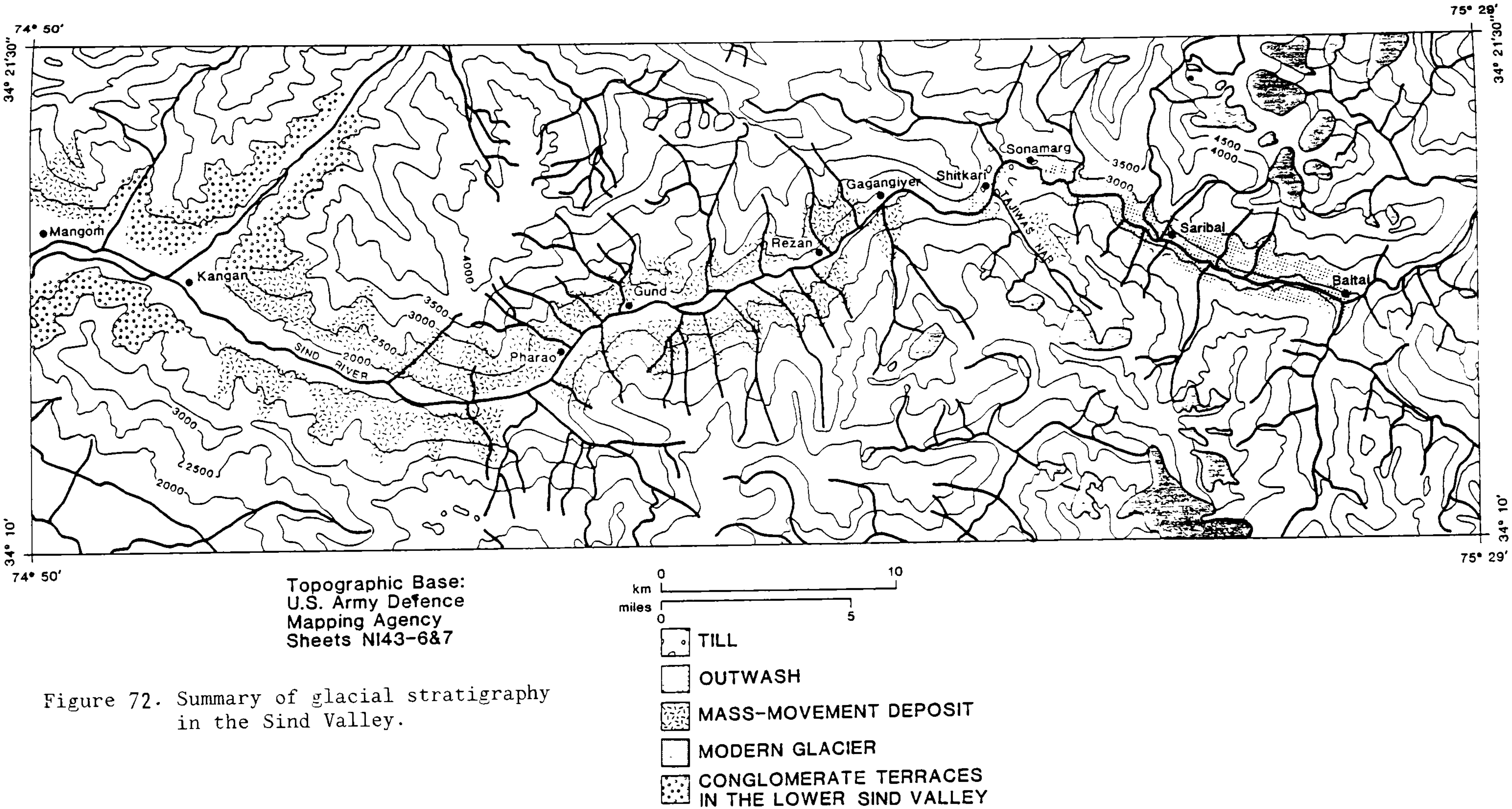


Figure 72. Summary of glacial stratigraphy in the Sind Valley.

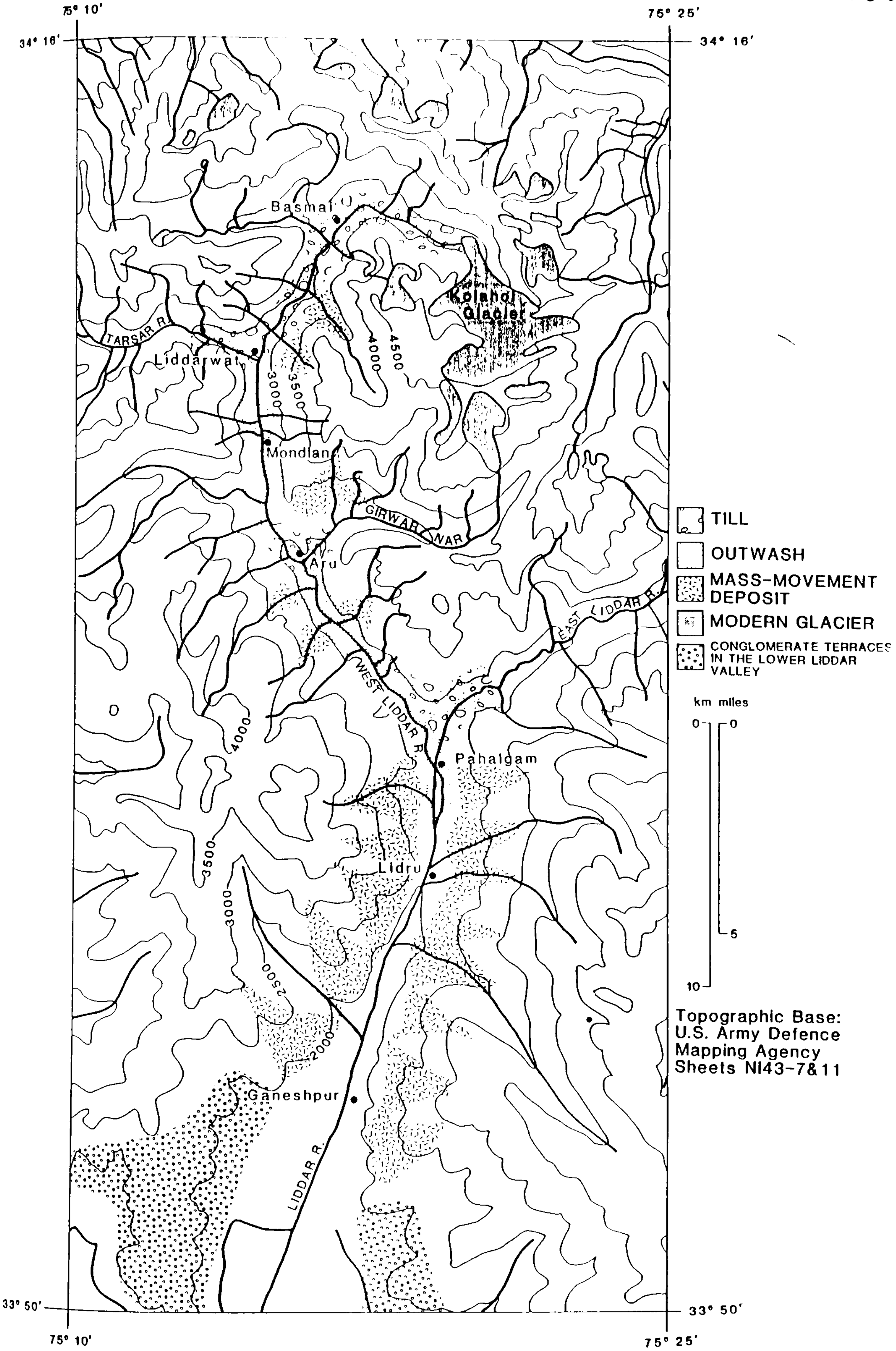


Figure 73. Summary glacial stratigraphy in the Liddar Valley.

Figure 74. Summary glacial stratigraphy in the Ningle Valley.

In the Sind Valley, observations were concentrated in the Sonamarg Basin. The large, hummocky moraines at Shitkari are the lowest occurrences of till in this valley (plate 32). A pair of nested moraines is found at the mouth of Lashmimarg Valley (plate 33). Extending upvalley from the large, hummocky moraines is a series of outwash terraces. These terraces rest against the large moraine where this has been incised by the Sind River. The outwash terraces extend upvalley as far as Baltal. A second hummocky moraine was mapped at Saribal and a series of unvegetated moraine rides, interpreted as products of Holocene glacial advances, occurs close to the snout of the Neh Nar glacier, at the head of the Sind Valley and around other modern glaciers in the catchment.

In the Ningle Valley, the lowest moraines occur at Gulmarg (plate 39). A higher set of moraines occur at Butapthri (plate 40). Outwash terraces occur in the mouth of Gulmarg Basin. Burbank (1982) also identified two sets of moraines in the Pir Panjal Range generally, from satellite imagery. These included a higher set, between 2600 and 2800 m a.s.l. and a lower set some 200 to 300 m below. The moraines at Gulmarg seem to correspond to Burbank's lower moraines and those at Butapathri to his higher moraines. A further set of morainic ridges, interpreted as products of Holocene ice advances, occurs around Alpather Lake, at the foot of Mount Apherwat.

The glacial sequence was most fully described in the West Liddar Valley. Deposits were traced from the inferred limit of glaciation, at Pahalgam, to the snout of the modern Kolahoi Glacier. RD measures were

TABLE 10. SUMMARY OF RD DATA.

VALLEY	SITE NO.	SBF	% WEATH-ERED	% SPLIT	% PITTED	PIT DEPTH MAX. (mm)	PIT DEPTH MEAN (mm)	MAX HT. RESIST. INTRU-SIONS (mm)	R-VALUE MIN. (cm)	R-VALUE MEAN (cm)	SOIL DEPTH (m)
SIND	S1	135	100	70	78	28	9	6	.25	.31	0.34+
	S2	99	100	76	69	29	12	7	.25	.31	0.36+
	S3	0	0	0	0	0	0	0	.25	.29	0.38+
	S4	0	0	0	0	0	0	0	.25	.32	0.28+
LIDDAR	L5	4	100	96	92	51	19	8	.25	.31	3.08
	L6	37	100	71	65	43	8	6	.25	.30	0.59+
	L7	41	100	77	60	48	12	7	.25	.31	0.81+
	L8	147	100	80	98	44	13	6	.25	.31	0.87+
	L9	197	100	80	100	35	12	7	.25	.28	0.85+
	L10	117	100	98	90	31	13	6	.25	.27	0.65+
	L11	>300	16	48	58	12	4	0	.25	.27	0.0
	L12	>300	20	34	78	13	5	0	.25	.28	0.0
NINGLE	N13	12	100	76	81	42	10	0	.25	.38	0.65+
	N14	8	100	80	79	36	12	0	.25	.36	0.57+
	N15	14	100	64	66	29	7	0	.25	.36	0.48+

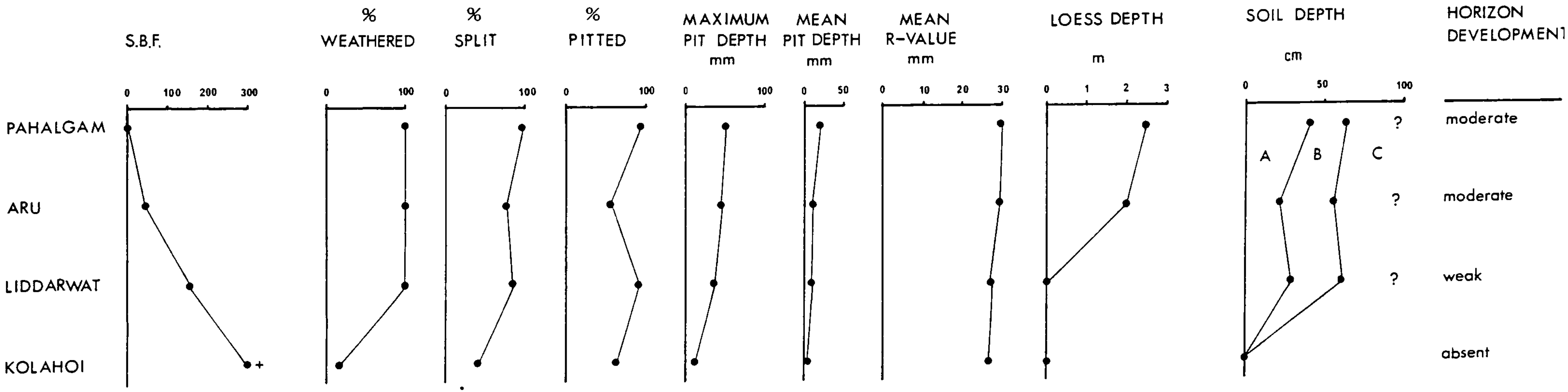


Figure 75

Summary graphs of RD data for the West Liddar Valley.

TABLE 11.
SOIL PROFILE NO. S1

HORIZON	DEPTH (cm)	DESCRIPTION
A	0 - 10	Black (5YR 1.7/1) loamy sand with occasional small clasts less than 5cm long; weakly developed; weak, fine granular structure; non-sticky, very slightly plastic.
B	10 - 16	Brownish black (10YR 2/2) loamy sand with abundant clasts less than 10 cm long; weakly developed; weak, fine granular structure; non-sticky, non-plastic.
Cox	16 - 24	Dull yellowish brown (10YR 4/3) loamy sand with abundant clasts less than 20 cm long; weakly developed; weak, fine granular structure; non-sticky, non-plastic.
Cu	24+	Yellowish brown (10YR 5/6 to 10YR 5/4) till with large (> 15 cm long) clasts.

TABLE 12.
SOIL PROFILE NO. S2

HORIZON	DEPTH (cm)	DESCRIPTION
A	0 - 20	Black(10YR 1.7/1) sandy clay loam with occasional small clasts less than 1 cm long; moderately developed; weak, fine granular structure; slightly sticky, slightly plastic.
Bt	20 - 26	Dark brown (10YR 3/3) sandy clay loam with common small clasts less than 1 cm long; moderately developed; weak, fine granular structure, weakly developed cutans on peds; slightly sticky, slightly plastic.
Cox	26+	Yellowish brown (2.5Y 5/6) sandy clay to sandy clay loam with common small clasts less than 1 cm long; moderately developed; weak, fine granular structure; weakly developed cutans, slightly sticky, moderately plastic.

TABLE 13.
SOIL PROFILE NO. S3

HORIZON	DEPTH (cm)	DESCRIPTION
A	0 - 5	Dark brown (10YR 3/3) silt loam; moderately developed; weak, fine granular structure; slightly sticky, slightly plastic.
Bt	5 - 23	Yellowish brown (2.5Y 5/4) silty clay loam; moderately developed; moderately firm, fine sub-angular blocky structure; weakly developed olive brown (2.5Y 4/4) cutans on peds; moderately sticky,moderately plastic.
Cox	23+	Olive brown (2.5Y 4/3) silty clay loam with abundant clasts less than 10 cm long; moderately developed; moderately firm, fine sub-angular blocky structure with oliver brown (2.5Y 4/4) cutans on peds; moderately sticky, moderately plastic.

TABLE 14.
SOIL PROFILE NO. S4

HORIZON	DEPTH (cm)	DESCRIPTION
A	0 - 8	Brownish black (10YR 3/2) silt loam; weakly developed; weak, fine granular structure; slightly sticky, slightly plastic.
Bt	8 - 18	Brown (10YR 4/6) silty clay loam; weakly developed; moderately firm, fine granular structure; moderately sticky, moderately plastic.
Cox	18+	Dull brown (7.5YR 5/.4) silty clay loam with abundant clasts less than 10 cm long; weakly developed; moderately firm, fine sub-angular blocky structure; weakly developed bright brown 7.5YR 5/6) cutans on peds; moderately sticky, moderately plastic.

TABLE 15.
SOIL PROFILE NO. L5

HORIZON	DEPTH (cm)	DESCRIPTION
A	0 - 40	Dark brown (10YR 3/4) silt loam with common clasts less than 5 cm long; moderately developed; weak, fine, sub-angular blocky structure; slightly sticky, slightly plastic.
Bt	40 - 63	Bright brown (7.5YR 5/6) silt loam with common clasts less than 5 cm long, occasional clast less than 10 cm but more than 5 cm long; moderately developed; moderately firm, fine to medium sub-angular; blocky structure with moderately well developed brown (5YR 4/6) cutans on peds; moderately sticky, moderately plastic.
Cox	63 - 98	Dark brown (10YR 3/4) silt loam with common clasts less than 10 cm long; weakly developed; moderately firm, medium sub-angular blocky structure with dark yellowish brown (10YR 4/6) cutans on peds; moderately sticky, moderately plastic.
bA	98 -148	Very dark brown (7.5YR 2/3) silt loam with rare clasts less than 5 cm long; apedal; rich in organic humus; moderately sticky, moderately plastic.
bBt	148 -248	Brown to dark brown (5Y 6/6) silty clay loam with common clasts less than 10 cm long; apedal; moderately sticky, moderately plastic.
Cu	248+	Olive (5Y 6/6) till; 10-20% $CaCO_3$.

TABLE 16.
SOIL PROFILE NO. L7

HORIZON	DEPTH (cm)	DESCRIPTION
A	0 - 20	Brownish black (5YR 2/2) silt loam lightening to dark brown (7.5YR 3/3) in the basal 5 cm; rare clasts less than 3 cm long; moderately developed; weak, fine sub-angular blocky structure, slightly sticky, slightly plastic.
Bt	20 - 56	Brown (7.5YR 4/4) silty clay loam with common clasts less than 5 cm long; moderately developed; weak, fine sub-angular blocky structure; moderately sticky, moderately plastic.
Cox	56+	Brown (10YR 4/4) loess with occasional to common clasts less than 5 cm long; moderately developed; weak, fine sub-angular blocky structure; moderately sticky, moderately plastic.

TABLE 17.
SOIL PROFILE NO. L8

HORIZON	DEPTH (cm)	DESCRIPTION
A	0 - 27	Dark brown (7.5YR 3/3) sandy clay loam with abundant clasts less than 10 cm long; weakly developed; weak, fine granular structure; slightly sticky, slightly plastic.
Bt	27 - 57	Brown (10YR 4/6) sandy clay with abundant clasts less than 10 cm long; weakly developed, weak, fine granular structure with weakly developed cutans on peds; moderately sticky, moderately plastic.
Cox	57+	Dark brown (10YR 3/3) sandy till with clasts less than 30 cm long.

TABLE 18.
SOIL PROFILE NO. L9

HORIZON	DEPTH (cm)	DESCRIPTION
A	0 - 34	Brownish black (7.5YR 2/2) sandy clay loam with abundant clasts less than 25 cm long; weakly developed; weak, fine granular structure; moderately sticky, moderately plastic.
Bt	34 - 68	Brown (7.5YR 4/6) sandy clay loam with abundant clasts less than 25 cm long; weakly developed; weak, fine granular structure; moderately sticky, moderately plastic.
Cox	68+	Dark brown (10YR 3/3) sandy till with abundant clasts less than 30 cm long.

TABLE 19.
SOIL PROFILE NO. N13

HORIZON	DEPTH (cm)	DESCRIPTION
A	0 - 15	Brownish black (10YR 2/3) sandy clay loam with occasional clasts less than 5 cm long; moderately developed; fine to medium granular structure; moderately sticky, moderately plastic
Bt	15 - 35	Dark brown (7.5YR 3/3) sandy clay loam with common clasts less than 10 cm long; moderately developed; weak, fine to medium granular structure; moderately sticky, moderately plastic.
Cox	35+	Yellowish brown (10YR 5/8) oxidised till with common clasts less than 100 cm long.

TABLE 20.
SOIL PROFILE NO. N14

HORIZON	DEPTH (cm)	DESCRIPTION
A	0 - 12	Brown (7.5YR 4/4) sandy clay loam; weakly developed; weak, fine granular structure; slightly sticky, slightly plastic.
Bt	12 - 27	Yellowish brown (10YR 5/6) sandy clay loam; weakly developed; weak, fine granular structure with common and moderately well developed cutans on peds; iron oxide staining present; Moderately sticky, moderately plastic.
Cox	27 - 57	Bright brown (7.5YR 5/8) silty clay; apedal and massive, except where vertical fissures have developed; very sticky, very plastic.
Cox	57+	Yellowish brown (10YR 5/8) oxidised till with common clasts less than 100 cm long.

used to subdivide the sequence and chronological control was provided by two TL dates and one radiocarbon date. The less well-described sequences from the Sind and Ningle Valleys were then correlated with the sequence from the West Liddar Valley. Rock-weathering measures for the three valleys are summarized in table 10. The soil profiles are descibed in tables 11 to 20. For the Sind Valley, RD and soil description sites S1 and S2 were on the cross-valley moraine; sites S3 and S4 were on the nested moraines. In the Liddar Valley, site L5 was at Pahalgam, sites L6 and L7 at Aru, sites L8 to L10 at Liddarwat and sites L11 and L12 at the front of the modern Kolahoi Glacier. In the Ningle Valley, sites N13 and N14 were at Gulmarg, site N15 was at Butapathri. All of these sites were on moraines. For the West Liddar Valley, summary RD measures are portrayed graphically in figure 75.

On the basis of RD measures, the only unambiguous age distinction is between the Kolahoi moraines and all of the others. The distinction between the Liddarwat, Aru and Pahalgam features is less clearly defined. There is a broad trend of increasing weathering from the Liddarwat moraines down to the Pahalgam moraines shown in many of the measures. However, in the case of the percentage of split and pitted boulders, there is a reversal in values between Aru and Liddarwat.All three sets of pre-Holocene moraines tend to have well weathered surface boulders which suggests considerable age and/or rapid rates of weathering. There is some evidence to support the view that the Liddarwat moraines are substantially younger than those at Aru and Pahalgam. This is indicated by the sharp change in surface boulder frequency between Aru and Liddarwat and the absence of loess from the surface of the Liddarwat moraines. The soils are also rather less well-developed on the Liddarwat moraines than at Pahalgam and Aru. However, this may simply be a reflection of the fact that the parent material is till at Liddarwat and loess at Pahalgam and Aru.

Radiometric dates from Pahalgam and Aru are shown on figure 70. The radiocarbon and TL dates from Pahalgam are inconsistent, since the maximum TL age of 18.57 kaBP is underlain by a palaeosol dated to 11.5 kaBP. The loess provided dates which were in good agreement for both additive dose and regeneration methods. This strongly suggests that the palaeosol has been contaminated by young carbon and has yielded an anomolously young radiocarbon date.

The loess sample from Aru was taken from a point stratigraphically lower in the loess from the Pahalgam sample, as shown in figure 70. As such, the TL date of 35.5 kaBP is not surprising. Taken together, these dates suggest a period of loess formation in the upper Liddar valley between at least 35.5 and 18.57 kaBP, with a phase of soil formation sometime in between. The advance of ice and deposition of moraines at both Pahalgam and Aru must therefore have occurred before 35.5 kaBP. If the loess deposition accompanied glacial advance and, by analogy with the present, soil formation occurred during interglacial periods, loess deposition at Pahalgam and Aru may have been contemporaneous with the advance of ice to Liddarwat. However, the Liddarwat moraine is undated. Between Liddarwat and the current snout of the Kolahoi Glacier, remnants of termino-lateral moraines are found on the valley-side near Basmai. However, since there are no cross-valley moraines in this part of the valley, it was not possible to establish the relative age of the moraines. However, from their position in the valley, they must be older than the Kolahoi moraines but younger than those at Liddarwat. For the west Liddar Valley then, a glacial sequence consisting of at least three major advances is suggested. The Pahalgam and Aru moraines were deposited in the first advance, the Liddarwat and possibly the Basmai moraines in the next oldest advance, and the Kolahoi moraines during the youngest advance.

The loess sequence in the Liddar Valley appears to correlate with that in Kashmir Basin. Comparison of the Liddar Valley chronology with that from one of the 'long' sections, such as Dilpur or Karpora, shows that the Pahalgam soil may correlate with either of the two upper soils in these sites, both of which are dated to less than 31.7 or 57.8 kaBP, depending on which chronology is used (see figure 49). This observation suggests that the soil forming phase in Kashmir was not restricted to the basin, but affected the montane valleys as well.

The next task is to attempt correlation between the West Liddar glacial sequence and those in the other two valleys. The moraines mapped in the Sind Valley occur in a small geographical area. The most obvious distinction is between the hummocky, cross-valley moraine at Shitkari and the pair of nested moraines in the mouth of Lashimarg Valley. On the basis of surface boulder frequency, the cross-valley moraine would appear to be younger than either of the nested moraines. However, the absence of boulders from the nested moraines may be a reflection of their morphology. They consist of very sharp, steep ridges and it is possible that surface boulders simply rolled off the surface following deposition. This idea is supported by the abundance of boulders around the base of the ridges. In fact, the steepness of the ridges suggests,if anything, a younger, rather than older age compared to the cross-valley moraines. The soil data (see table 11 to 14) provide no further basis for subdivision of the glacial sequence. In the absence of evidence to the contrary, the nested moraines and the cross-valley moraines will be regarded as similar in age.

Since no radiometric dates are available for the Sind Valley sequence, correlation with the west Liddar Valley is on the basis of RD data alone. Both sets of moraines in the Sonamarg Basin show a similar degree of soil development to any of the pre-Holocene moraines in the West Liddar Valley. However, on the basis of surface boulder weathering, the cross-valley moraine correlates better with younger, Liddarwat moraines rather than the older features at Aru and Pahalgam. If this correlation is correct, then the older part of the glacial sequence appears to be missing in the Sind Valley. There are several reasons why this may be so. One reason could be differential uplift of the Sind catchment compared to the Liddar. If greater uplift had been experienced in the Sind Valley, ice of a younger advance may have overrun, and so destroyed evidence of, an earlier advance that was equivalent in age to the Pahalgam advance in the West Liddar Valley. Another possible explanation is that the young age of the moraines in the Sind Valley is an apparent effect resulting from slower rates of weathering. The first of these explanations seems unlikely since the upper reaches of the Sind and West Liddar Valleys are adjacent and very close together. It is difficult to determine the possible role of different rates of weathering, although since the degree of soil development does not vary markedly on any of the pre-Holocene moraines in the two valleys, it seems unlikely that this explanation is plausible. The moraine at Saribal, further up the Sind valley, could not be studied in details since a military encampment was present on the surface. From the RD data, it is difficult to draw firm conclusions regarding the relative age of the moraines in the Sind Valley.

Burbank (1982) has previously suggested that the two sets of moraines in the Pir Panjal Range are of different ages. He tentatively proposed that the younger, higher moraines might have been deposited during the last glacial maximum and the older, lower moraines deposited either in the early part of the last glacial or during the penultimate glaciation. The RD data from the Ningle Valley in this study show that both the higher moraines at Butapathri and the lower moraines at Gulmarg are well weathered, have few surface boulders and thick, well-developed soil covers. The data do suggest that there might be an age difference between the two sets of moraines, but this distinction has been blurred, presumably by higher rates of weathering and pedogenesis on the Pir Panjal Range, which is more humid than the Great Himalaya.

Basal dates from bog sediments at Butapathri (Dodia et al., 1984) and Toshmaidan (Singh and Agrawal, 1976) suggest that the younger, higher moraines date from at least the last glacial maximum. At

Toshmaidan, the basal sediments in the bog were dated at 14760 aBP. This is overlain by an organic mud dated at 15250 aBP and underlain by a lacustrine clay dated at 13850 aBP (see appendix 2 for full details of the dates). Clearly, there is some error in these dates, since there are two inversions in the sequence. At Butapathri, the basal date on peats is 17110 aBP. This date is taken by Dodia et al.to indicate the onset of deglacierization at Butapathri by about 17000 aBP. However, some doubt must be cast over the accuracy of this date, since 39 cm thickness of sediment was used in the radiocarbon assay. Dodia et al. (1984) argued that the pollen zone in the basal part of the Buapathri sequence contains broad-leaved arboreal elements and therefore indicates climatic amelioration. However, the dated part of the sequence also covers part of the overlying pollen zone, which is dominated by blue pine (Pinus willichiana) and regarded as a cool phase. In view of these dating problems both at Butapathri and Toshmaidan, the dates for the onset of climatic amelioration at both sites must be regarded with caution.

Correlation between the Ningle Valley and the West Liddar Valley is problematical, since the two catchments are under contrasting climatic regimes. However, the advanced state of weathering of the Gulmarg moraines suggests that they may correlate with Pahalgam/Aru and possibly the Sonamarg moraines on the Himalayan flank. Dates on the higher moraines at Butapathri suggest that they postdate the maximum advance of ice in Kashmir. It is possible that they may correlate with the Liddarwat moraines in the west Liddar Valley. Fresh, unweathered moraines around Alpather Lake are regarded as Holocene in age and

TABLE 21. CORRELATION OF GLACIAL SEQUENCES FROM THE SIND, LIDDAR AND NINGLE VALLEYS.

AGE (kaBP)	STAGE			
	LIDDAR VALLEY	SIND VALLEY		NINGLE VALLEY
<10	5 Kolahoi	3 Neh Nar		3 Apherwat
?	4 Basmai	2 Saribal	2 Inner nest	2 Butapathri
?35-16	3 Lidderwat	1 Sonamarg	1 Outer nest	1 Gulmarg
>35	2 Aru			
>35	1 Pahalgam			

correlative with similar moraines in the foreland zone of the Kolahoi Glacier.

RD data and radiometric dates are used to produce the correlation between glacial deposits in the three valleys studied in this research. This information is summarized in table 21.

6.5 Present and Past Patterns of Glaciation.

a. Introduction.

The spatial variations in the size and distribution of glaciers are often studied in currently-glacierized regions. Such information can provide valuable insights into the nature of controls on glaciation, especially if climatic data are available. If the extent of glaciers can be reconstructed for periods in the past, constraints can be placed on the magnitude of climatic change and any accompanying changes in climatic gradients identified. Such information, therefore, makes an important contribution to the study of past climates.

Although the head- and snout-altitudes of glaciers may have some climatic significance, the positions of snow lines, firn lines and equilibrium lines are generally much more sensitive to climate. The snow line is the lower altitudinal limit of perennial snow between glaciers. The reason for defining the line between glaciers is that bodies of glacier ice will generally lie at lower altitudes than patches of perennial snow. Aspect and local topography tend to produce widespread variations in the height of the snow line within a region. These local irregularities are termed the 'orographic snow line'. The integration of these irregularities into a potentially wide band constitutes the 'regional snow line.' The 'climatic snow line' is a theoretical snow line which refers to the lower altitudinal limit of perennial snow on fully exposed, flat surfaces, between glaciers (Flint, 1971, pp.64-65). Osmaston (1975) argues that this is a rather misleading concept, since the distribution of snow bodies, and even climates themselves, is largely a function of topography. The firn line on a glacier separates ice from snow at the end of the ablation season. On temperate glaciers, it corresponds with the equilibrium line. The equilibrium line is defined as the line or zone on a glacier where annual ablation equals annual accumulation. It separates the accumulation area from the ablation area and has zero mass-balance. Changes in the position of the equilibrium line result from changes in mass balance which, in turn, reflect climate. In many studies of present and past glaciation patterns, the position of the firn line is mapped as a surrogate for the equilibrium line on the assumption that the two coincide.

An indirect indication of the climatic snowline is given by the glaciation level or limit (Andrews, 1975). The summit method of calculating an index of glacierization was devised by Bruckner (1886) and has subsequently been used in North America, Scandinavia and New Zealand (eg. Østrem, 1966; Andrews and Miller, 1972; Porter, 1975b). Porter (1977) prefers the term 'glaciation threshold' since it avoids confusion with the outer limits of glaciation implied by 'limit' and the occurrence of a flat surface implied by 'level'. Porter's term, abbreviated to GT, will be used here. The GT is defined as the arithmetic mean of the altitudes of the lowest glacier-clad summit and the highest ice-free summit, but excluding summits of unsuitable shape to support glaciers.

There are a number of methods by which patterns of modern glaciation can be determined. The best method of determining the ELA (equilibrium-line altitude) is to carry out mass-balance measurements on a glacier. Because mass balance fluctuates annually, it is necessary to carry out measuremants for several years in order to determine the ELA accurately. The equilibrium line determined in this way will relate only to the glacier upon which the measurements were made (Andrews, 1975). Porter (1975a) stresses the importance of studying spatial trends in ELAs within an area, since these will reflect climatic gradients and topographic control. As such, direct measurements of ELAs for a glacierized region are prohibitive in terms of time and cost.

Where high-quality topographic maps or aerial photographs are available, it may be possible to determine regional trends in the snow line. However, it is not possible to determine the equilibrium line directly from aerial photographs because of superimposed ice. Because of this, the equilibrium line will always lie below the limit of snow at the end of the summer season (Andrews, 1975). Topographic maps are often used to determine the position of equilibrium lines on glaciers. The transition from the accumulation to the ablation area is marked by a change from convex to concave contours on the maps. However, as Andrews (1975) points out, high quality topographic maps are required for this method and the relationship is only valid in simple topographic situations.

Many studies have used the ratio of accumulation area to ablation area (the accumulation area ratio, or AAR) in order to define the position of the equilibrium line. Studies on many glaciers have shown that the AAR typically lies between 0.5 and 0.8 when the glacier is in steady-state equilibrium (Meier and Post, 1962). The equilibrium line is defined using an assumed AAR and an area-altitude curve for each glacier.

In the absence of reliable area data, it is still possible to compute a value for the ELA using a height method (Osmaston, 1975). As a first approximation, it is possible to use the glacier mid-height (ie.the median altitude) for the ELA. In practice, however, some fraction of the height range other than 0.5 is usually used. The chosen value, termed the toe-to-headwall-altitude-ratio (THAR) is derived from actual observations of ELAs, ideally from within the study region, although this is frequently not possible, and the THAR is derived from glaciers elsewhere. In a comparative study of glaciers in the Colorado Front Range, Meierding (1982) used a THAR value of 0.40 in order to calculate ELAs. The height method has the obvious advantage that area information, often unreliable when topographic maps or aerial photographs are unavailable, is not required. A refinement of the height method is Höfer's 'general height method' (see Osmaston, 1975), in which ELAs are determined by plotting toe (T) and headwall (H) altitudes, for a homogeneous group of glaciers, on bivariate axes. The data points should lie close to a straight line. The ELA for the group of glaciers is defined by the line T = H (ie. X = Y).

Once values of ELAs and GTs have been calculated for a region, it is very useful to show variations by drawing contour lines of equal value (isoglacihypses). Such contours allow not only the trends in glacierization within a region to be identified, but also show areas with anomalously high or low values.

Several methods have been used to determine snow lines for past glaciers. A number of authors have used cirque altitudes as indicators of past glaciation. Notwithstanding the fact that cirque altitudes and ELAs are only concident for small cirque glaciers, trends in cirque alitudes can provide useful information about past glaciation gradients and, hence, climate (eg. Peterson and Robinson, 1969; Flint, 1971). In areas where glacial deposits have been well mapped, it may be possible to calculate values of the past GT using similar information as that for present GTs. (eg. Andersen, 1968; Wahraftig and Birman, 1965; Porter, 1977).

Although ELAs cannot be measured directly for former glaciers, there are a number of methods available for their estimation. If the geometry of former glaciers can be determined by detailed mapping of moraines, and a former AAR assumed, the past ELA on a glacier can be estimated using area methods. If the former geometries of glaciers cannot be determined, but toe (T) and headwall (H) altitudes are available for former glacier, the past ELA can be estimated using an assumed THAR value. Examples of use of the area and height methods include Porter (1970) and Burbank and Fort (1985) respectively. Osmaston (1975) has also used Höfer's general height method in the study of past glaciers.

If lateral moraines are preserved within a valley, their location can be used to locate the ELA, since lateral moraines only occur in the ablation zone of a glacier. Furthermore, lateral moraines also delimit the former extent of the glacier up the valley side and so provide important information about past glacier geometries.

Once values of former GTs, ELAs and cirque altitudes have been calculated for a region, they can be contoured in a similar fashion to the values for modern glaciers. Once this has been done, various palaeoclimatic conclusions can be drawn. Although glaciers are controlled by relatively simple equilibria, particularly so compared to biological evidence (Osmaston, 1975), the interpretation of past glaciation patterns should be undertaken with care. Although ELA and GT depressions for former glaciations have often been quoted in terms of temperature change, relative variations in values within a region are often more informative than comparison of absolute values between areas. Nevertheless, Porter (1975a) cites numerous examples of maximum depressions of ELAs during the Late Quaternary of 900±50 m. He stated that any significant deviation from this value would be of palaeoclimatic significance.

The size and altitudinal range of a glacier are essentially a function of temperature, precipitation, topography and aspect. If all other things remain equal, decrease in temperature, or increase in precipitation, will tend to cause a glacier to advance. However, anomalous values for past ELAs or GTs may be calculated if an area has undergone tectonic uplift since the glacial advance. Uplift will have the effect of decreasing the measured depression of the ELA and GT compared to actual values. In areas such as Kashmir, where rapid uplift has occurred during the Late Quaternary, it is important to recognise the possibility of modification in ELAs and GTs in this way and to make appropriate adjustments to any palaeoclimatic inferences drawn. However, uplift is of considerable importance in its own right and anomalous ELA or GT relationships within a region may indicate zones of rapid uplift.

b. Methods Used in This Study.

Any study of Himalayan glaciers is usually hampered by the lack of map and aerial photograph coverage and Kashmir is no exception to this. However, the 1:63 360 maps of Kashmir are generally adequate for this part of the study over most of the basin. Although the 100 ft. contour interval is rather large, there are numerous other spot heights shown that have been established by more accurate surveying methods. In addition, most of the lakes within cirques have spot heights on their shorelines and many mountain summits have surveyed heights. In general, the altitues reported in this study are regarded as accurate to ± 20 m.

In order to determine the pattern of present glaciation in Kashmir, each individual glacier depicted on the 1:63 630 maps was traced onto an overlay and numbered. Wherever possible, a note of the toe (T) and headwall (H) altitudes was made. The area of each glacier was determined using a digital planimeter and a note of orientation made. Although the maps use imperial units throughout, these have been converted to metric for the purpose of this study. Because T and H are often the only altitudinal information available for each glacier and because the outlines of glaciers were quite generalized, it was thought inappropriate to use an area method to locate the ELA. Instead, the height method was used.

The choice of a THAR value was governed by several considerations. The Neh Nar Glacier, in the Sind catchment, has been mapped and studied by Bhandari et al. (1983). Their detailed study reveals T and H values of 3920 and 4925 m a.s.l. respectively. However, it is difficult to determine a meaningful value for the ELA, since avalanching of snow is an important source of accumulation. The accumulation zone mapped by Bhandari et al. (1983) extends a little below 4200 m a.s.l. on the southern limb of the glacier and to 4475 m a.s.l. on the northern limb.The lower accumulation zone on the south-facing slopes is due to greater avalanching, a result of greater solar-radiation receipt. The ELA of the Neh Nar glacier may be around 4475 m a.s.l. if avalanching is discounted. However, this is not an adequate basis on which to determine a THAR value for use in Kashmir. Two other Himalayan glaciers have been studied by Bhandari et al. (1983). These are the Gara Glacier in Himachal Pradesh and the Changme

Khangpu Glacier in Sikkim. On neither of these two glaciers is avalanching an important source of accumulation. The THAR values for these glaciers are 0.38 and 0.47 repsectively (see table 22). A similar result is abtained for the Mir Samir Glacier in Afghanistan, which has a THAR value of 0.57. Meierding (1982) has shown, albeit for the Colorado Front Range, that a THAR value of 0.4 gives the best estimate of the ELA when using the height method. Therefore, a THAR value of 0.4 was used to calculate the ELAs of glaciers in Kashmir.

Values of the modern GT were also calculated for Kashmir. The glacierized area shown on the 1:63 630 maps was divided into 5' east by 5' west quadrangles. Within each quadrangle, the GT was determined using the method of Østrem (1966). In quadrangles where all the summits were too low to support glaciers, it was necessary to extrapolate from adjacent glacierized quadrangles. This procedure was only carried out if the adjacent quadrangle contained a glacier.

A large number of non-glacierized cirques was identified in Kashmir on the 1:63 630 maps. The height and orientation of each cirque was taken from the maps where ever possible. The cirque altitude was generally taken as the spot height on the lake shoreline. In the absence of a spot height, the height of the break of slope at the cirque backwall was used. Some authors use all the cirques in a region to indicate the former glaciation gradient whilst others use only the lowest individuals within a sampling scheme. Under both approaches, however, the cirque altitudes are strongly influenced by

TABLE 22. THAR VALUES OF HIMALAYAN AND TRANS-HIMALAYAN GLACIERS.

GLACIER	TOE ALT. (M)	HEADWALL ALT. (M)	OBSERVED ELA (M)	THAR
Neh Nar, Kashmir[1]	3920	4925	4475	0.55
Gara, Himachal Pradesh[2]	4710	5600	5050	0.38
Changme Khangphu, Sikkim[2]	4850	5800	5300	0.47
Mir Samir West, Hindu Kush[3]	4660	5080	4900	0.57

References
1. Nijampurkar et al. (1982).
2. Bhandari et al. (1983).
3. Gilbert et al. (1969).

topography (Meierding, 1982). In this study, all of the north-facing ($N \pm 30^0$) cirques were mapped in Kashmir.

The computed values for ELAs, GTs and cirque altitudes were then contoured. The mean ELA of north-facing glaciers ($N \pm 30^0$) was calculated for each 5' x 5' quadrangle and the data point placed at the quadrangle centre. This averaging procedure introduces some smoothing into the data, but was felt to be justified since it reduces the amount of error due to uncertainties in the topographic maps. For GTs, the value for each quadrangle was plotted at the quadrangle centre and the data points were contoured. For cirque altitudes, a similar procedure was adopted as for ELAs, whereby the mean height of cirques was calculated for each quadrangle. Changes in ELAs for Quaternary glacial advances were determined using the mapped former ice limits. Cirque altitudes were also used as a generalized indicator of past glaciation. Calculation of Former GTs was not attempted since glacial deposits have not been mapped extensively throughout Kashmir.

C. Results.

The present and past glaciation patterns in Kashmir are summarized in a number of figures, tables and appendices. The raw data upon which the patterns are based is summarized in appendix 15. The modern distribution of glaciers and their inferred maximum extensions are shown in figure 76. Isoglacihypses of modern ELAs and GTs are shown in figures 77 and 78 respectively. A similar diagram showing isoglacihypses of cirque altitudes is shown in figure 79. A cross-section, illustrating trends in the GT and cirque altitudes on a south-west to north-east transect across Kashmir, is shown in figure 80. The inferred total glacierized areas in Kashmir, for both the present and various stages in the past, are listed in table 23. Finally, an area - altitude curve, in which glacierized area is plotted against ELA depression is shown for the West Liddar Valley in figure 81.

Present-day glaciers in Kashmir fall into a number of distinct groups. On the Pir Panjal flank, glaciers are currently found only in the south-western extremity, in the headwaters of the Rembiara and Vishav Valleys. On the Himalayan flank, glaciers are more widespread. The upper reaches of the Liddar Valley are quite heavily glacierized and distinct groups of individual glaciers occur in the West and East Liddar Valleys. A limited number of glaciers is also found in the lower part of the Liddar Valley, particularly in tributary valleys facing north-west. In the Sind Valley, glaciers are found extensively in the upper reaches of the valley, close to the watershed. A second group occurs in the Tajiwas Valley, a tributary which joins the Sind close to Sonamarg. A third group occurs in the headwater region of Wangat Nala, a tributary of the Lower Sind. These major groups include most of the individual glaciers in Kashmir. A further group of glaciers which occurs beyond the Kashmir watershed, in Ladakh, is also considered. The area beyond Kashmir is only partially mapped at 1:63630 scale, but is important since it presents such a marked climatic contrast to Kashmir.

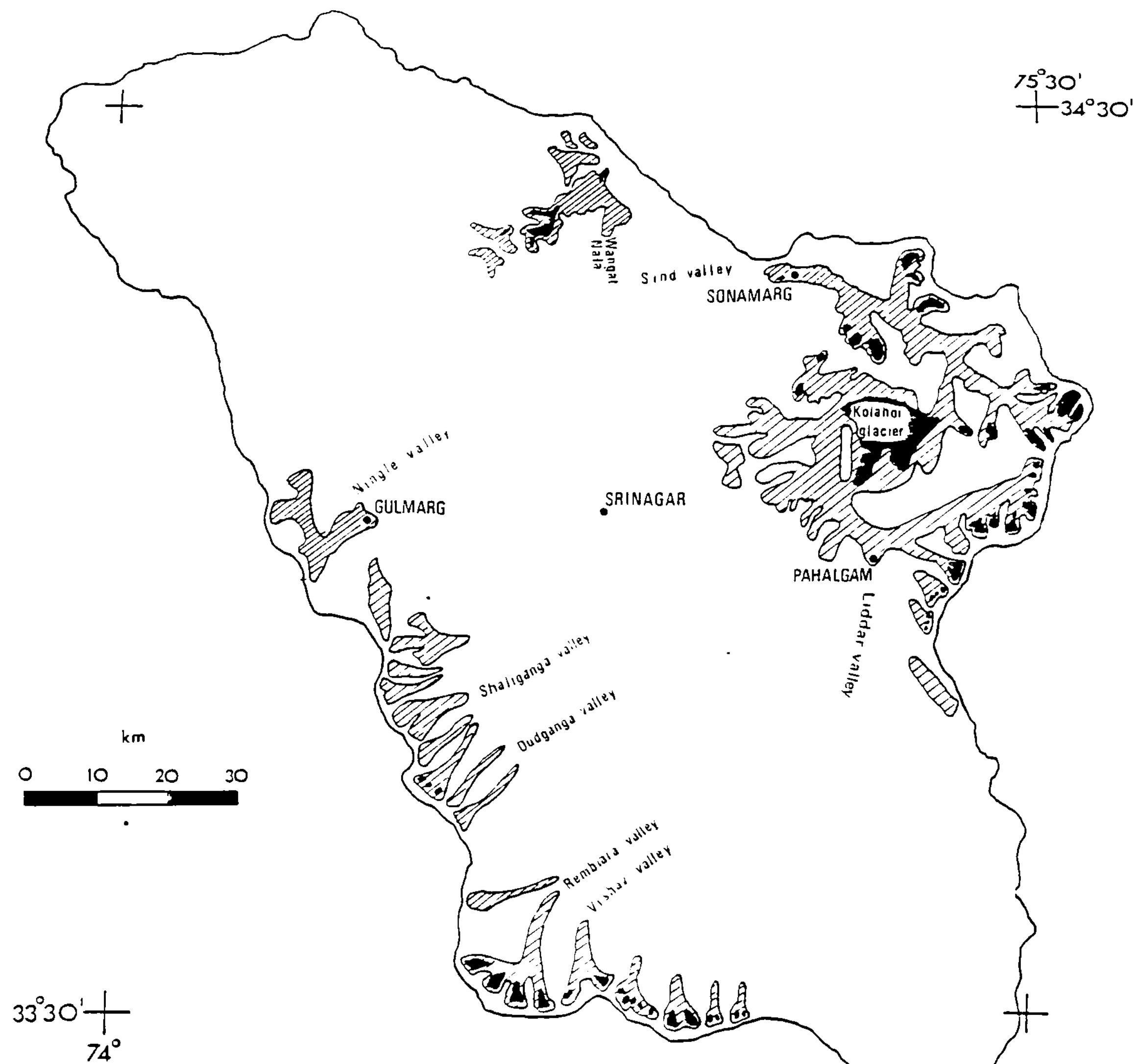


Figure 76
Map of the past and present glacier cover in Kashmir.
(Modern glacier cover shown by black shading. Maximum extent of glaciers during the Quaternary shown by cross-hatching).

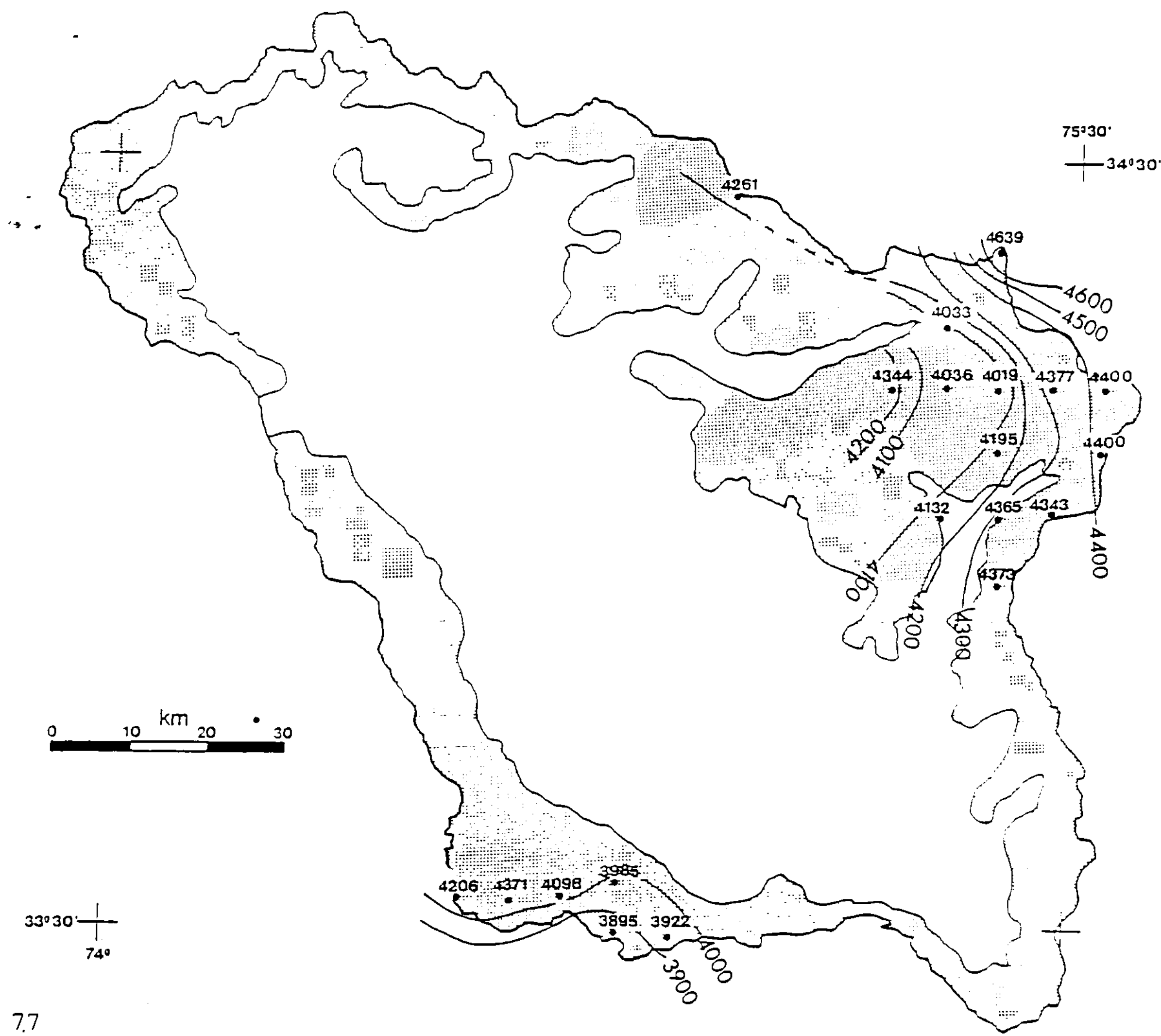


Figure 7.7
Isoglacihypses of ELAS of north-facing glaciers in Kashmir.

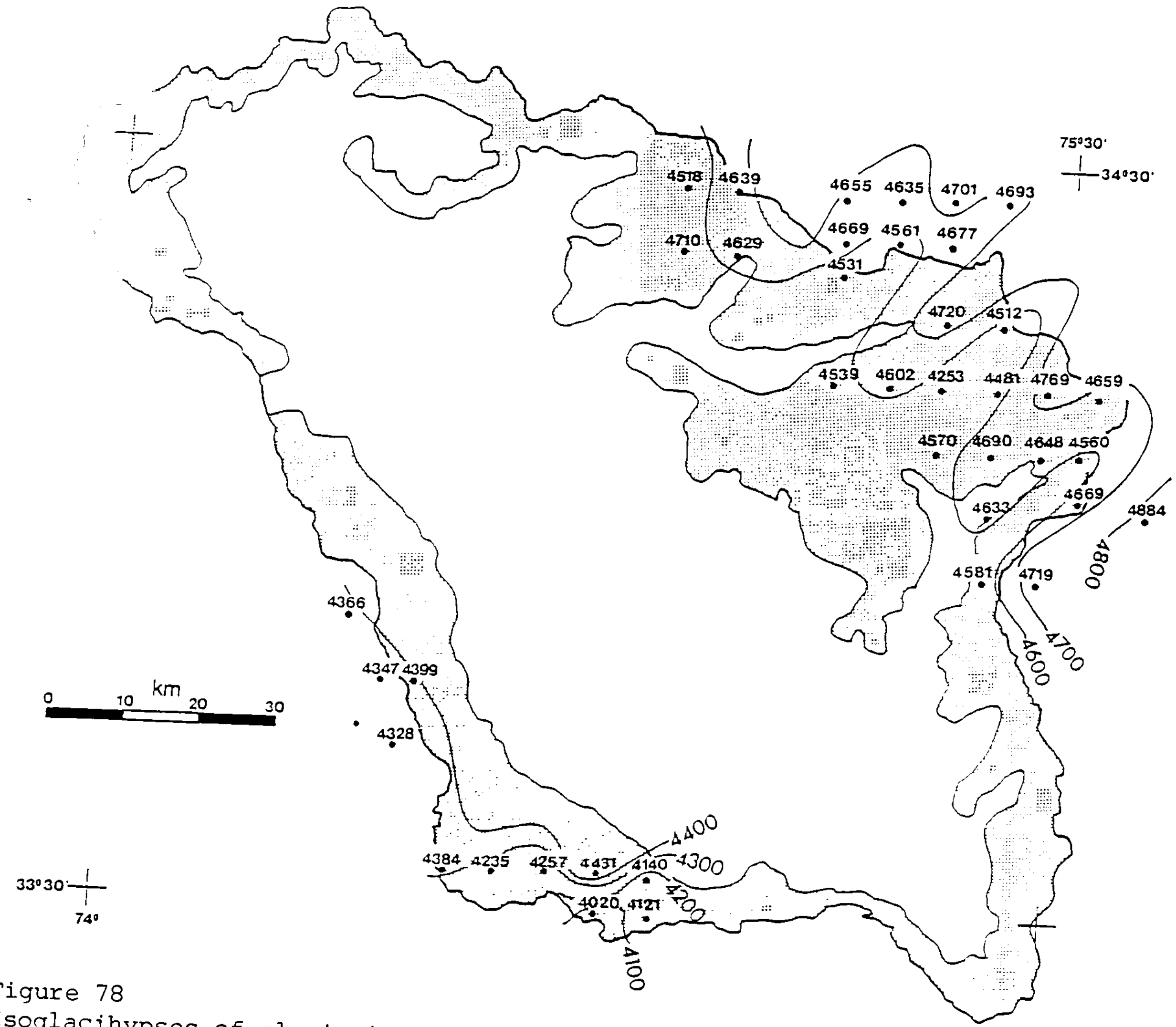


Figure 78
Isoglacihypses of glaciation threshold in Kashmir.

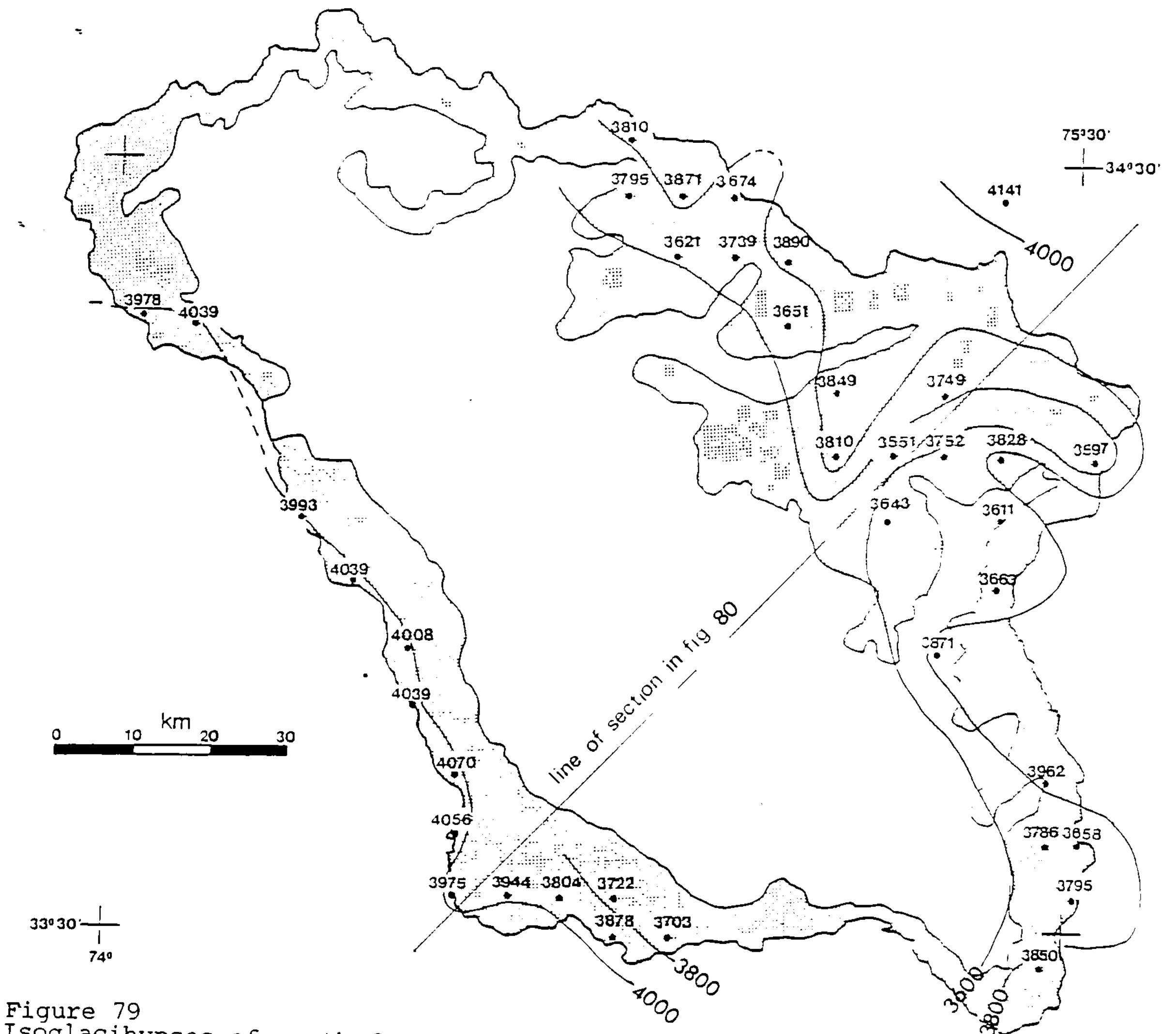


Figure 79
Isoglacihypses of north-facing cirques in Kashmir.

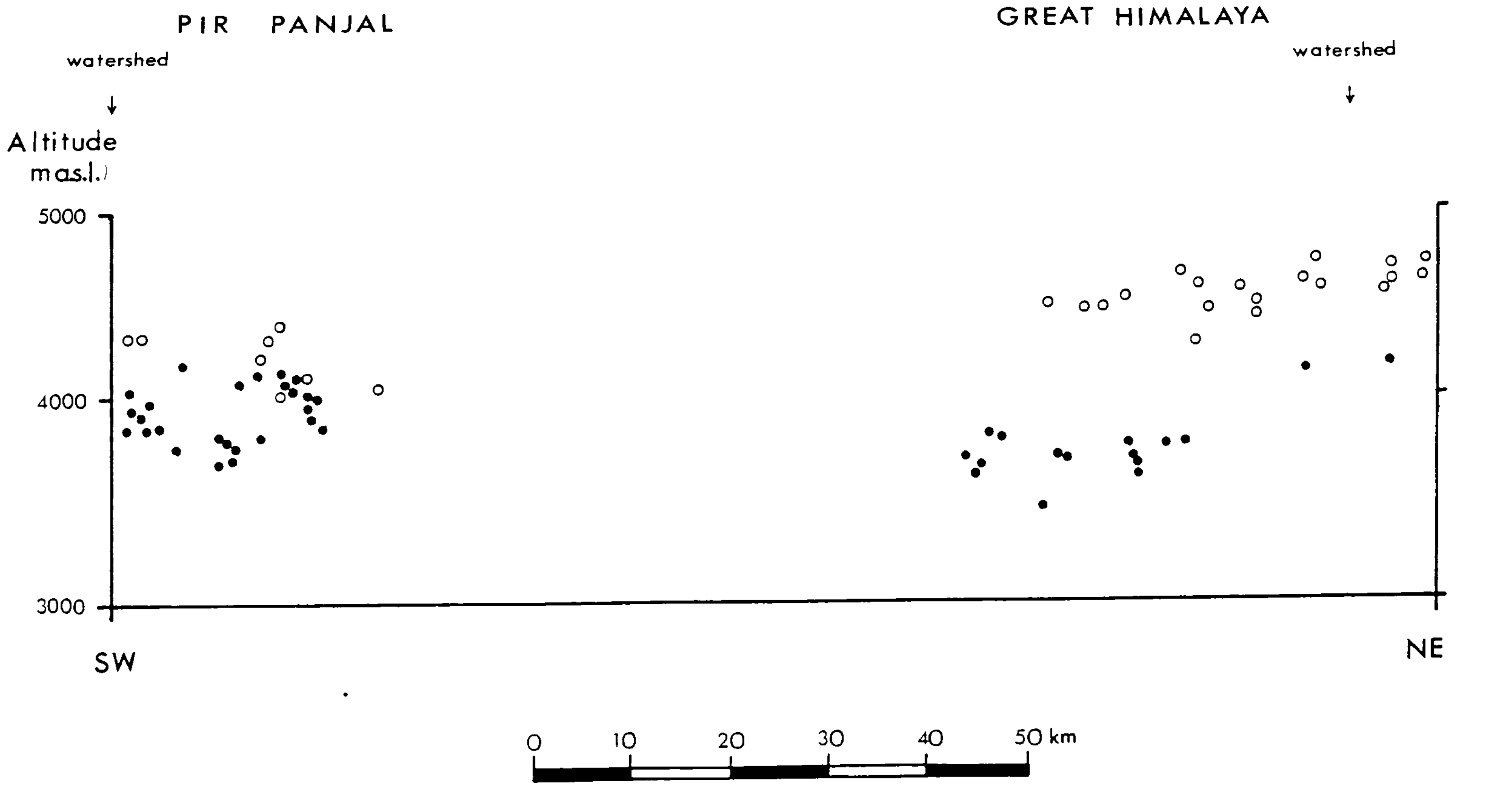


Figure 80

South-west to north-east transect across Kashmir showing the height of the glaciation threshold and cirque altitutes.
(Open circles ≙ glaciation threshold. Closed circles = cirque altitudes).

Topography appears to be the most important factor controlling the occurrence of glaciers in Kashmir. Although 6 major groups of glaciers can be identified, the actual distribution is rather limited. The glacierized parts of both the Himalayan and Pir Panjal flanks coincide with the areas of highest altitude. Within each glacierized zone, the distribution of glaciers is further controlled by valley orientation and glaciers tend to occur in north- rather than south-facing valleys.

Further information concerning the controls on glacier distribution is revealed by the patterns of GTs and ELAs. Determination of GTs for Kashmir yielded 40 data points. The isoglacihypses plotted on figure 78 reveal a number of important features. Firstly, distinct differences are found in GTs between the Pir Panjal flank, the Himalayan flank and the area to the east of Kashmir in Ladakh. Generally, the GT is low in the Pir Panjal Range (4100 to 4500 m a.s.l.), intermediate in the Himalayan Range (4600 to 4700 m a.s.l.) and high in Ladakh (>4700 m a.s.l.). In the Pir Panjal Range, the isoglacihypses show a south-west to north-east increase in the height of the GT. In the Himalayan flank, the pattern of the isoglacihypses is more complex. Although there is a discernable north-easterly rise in the GT, this is complicated by 3 or 4 deep bulges, oriented south-west to north-east, in the isoglacihypses.

The increase in the height of the GT from the Pir Panjal to the Himalaya and beyond is best explained by a precipitation gradient. The description of the climatology of Kashmir and north-west Himalaya in chapter 2 indicated two major sources of precipitation, one from the west in the form of winter westerly depressions, and one from the east, associated with summer monsoon disturbances. In either case, the glacierized part of the Pir Panjal Range would receive precipitation before that in the Himalaya. The aridity of Ladakh is also illustrated by the abrupt change in vegetation seen when crossing the watershed from Kashmir.

A potentially more accurate illustration of the precipitation control on the GT gradient lies in the available meteorologcal data for Kashmir. However, although this is hampered by the dearth of modern meteorological data, a number of tentative comments can be made. Firstly, the pattern of precipitation within Kashmir Basin (that is, between the mountain flanks) does not show any spatial pattern. Raza et al. (1979) suggested that total annual precipitation was lowest in the very centre of the basin and increased outwards towards the mountain flanks. This pattern is, however, the product of data from one or two stations. In any case, it does not bear much relevance to the glacierized region. Precipitation data for the mountain flanks would be of greater relevance to understanding the GT patterns. However, there are only two stations of possible value, at Gulmarg Research Station and Sonamarg. At first glance, these data appear to contradict the hypothesis of precipitation control on the GT patterns since the mean annual total precipitation is much greater at Sonamarg, on the Himalayan flank (1815.5 mm) than at Gulmarg (1308.6 mm). However, these data cannot be taken at face value for a number of reasons. Firstly, the precipitation records are of unequal length. Whereas the preipitation data for Sonamarg are based on a 49 year record, those from Gulmarg come from just 2 years record. Although a much longer record is available from the Indian Meteorological Department's (IMD) station at Gulmarg, this only records for the months of June to September. For the Research Station, complete data for 1968 and 1969 only could be found. When the summer months of these years are compared with mean summer precipitation from the longer, IMD records, there is a strong suggestion that 1968 and 1969 were dry years. Furthermore, when the summer months' precipitation from Sonamarg is compared with the longer record of summer precipitation from Gulmarg, precipitation totals are much greater for the latter site. Therefore, at least some of the apparent disparity between GT gradients and meteorological data may be a function of differences in the length of the records. The second problem relates to the proximity of the rain gauges to the glacierized zone itself. The Sind Valley contains active glaciers and Sonamarg is only a few kilometres to the nearest of these. However, Gulmarg is some way from the nearest of the Pir Panjal glaciers, which are concentrated in the south west part of the Pir Panjal. Therefore, the precipitation at Gulmarg has uncertain relevance to the glaciers in the Pir Panjal. Thirdly, Gulmarg is located in the lee of the Pir Panjal crest; and may be in a rainshadow. Fourthly, the ranfall data themselves are of doubtful reliability. The dubious quality of the data is indicated by the long lists of errata that accompany each meteorological summary and the not infrequent omission of data from stations on the grounds that the data are regarded as 'unreliable' by the IMD. Considering the mechanisms of precipitation affecting Kashmir and the reservations surrounding the meteorological data, the best explanation of the trend in GTs is precipitation control.

The isoglacihypses of GTs show a distinct rise from south-west to north-east across Kashmir. This is best explained by topography since the axis of high mountains also runs in this direction. Thus, although other areas in Kashmir may have the potential, in climatic terms, to support extensive glaciers, the topography is too low. The steady rise in GTs in a north-easterly direction across Kashmir is therefore a function of both precipitation gradient and topography.

The gradient of the GTs is indicated by the spacing of the isoglacihypses. Values of the gradient clearly vary quite markedly. The gradient in the GT is consistently high on the Pir Panjal flank, ranging from 20 to 68 mkm^{-1}. On the Himalayan flank, the gradient ranges from 5 to 68 mkm^{-1}. Whereas the lower values of the GT gradient are well within the range of figures quoted for other mountain ranges, the upper gradients exceed published values by a factor of 2 (see, for example, Porter, 1977). It is possible that the values for Kashmir are incorrect, due to errors in the topographic information on the maps. However, the highest gradients are, in most cases, constrained by several data points and are a feature of both mountain flanks. So, some reason other than data error is required to explain the high gradient in the GT.

The gradients quoted by Porter (1977) refer to mid- to high-latitude, maritime mountain ranges. The highest gradient, about 25

mkm^{-1}, occurred in mid-latitude ranges between about 40 and 60^0N, in areas of high precipitation gradient. In these maritime mountain ranges, precipitation was found to be highest adjacent to the coast and declined sharply inland. In higher latitude mountain ranges, overall totals of precipitation were lower, leading to more gentle gradients in precipitation and, hence, in the GT.

However, precipitation frequently interacts with topography to influence the GT gradient, especially on a more local scale. In a study of GTs in the Southern Alps of New Zealand, Porter (1975b) found that GT gradients were more gentle in areas of lowest topography, particularly where broad cols occur along the trend of the mountain crest. The gentle GT gradients in such situations were explained by the greater ease of penetration of moist maritime air masses across the cols and into the adjacent landmass. However, topography also has direct effects on glaciation, since relief of a given altitude is required for glaciers of certain dimensions to develop. The relative importance of precipitation and topography, together with the other important climatic variable affecting glaciers, ablation-season temperature, can be evaluated with some success (eg. Porter, 1977) if appropriate meteorological data are available. For Kashmir, the explanation of the GT gradients is necessarily rather conjectural due to the lack of such data. However, further information is provided by the more detailed patterns of the isoglacihypses.

On the Pir Panjal flank, the isoglacihypses follow the topography quite closely, although they show two quite pronounced bulges which are associated with slight topographic lows in the headwater region of the Vishav and Rembiara Rivers. Apart from this, the isoglacihypses follow the trend of the Pir Panjal crest, indicating an increase in the value of the GT from the watershed towards the centre of the valley. This pattern indicates both topographic and precipitation control on the GT. The topographic lows could allow the greater penetration of moist air, either from westerly depressions during winter or monsoonal disturbances during summer. The steepness of the GT gradient suggests that if precipitation is the main control, there must be a very rapid distance-decay in precipitation north and north-eastwards from the watershed. If this is the case, the rainshadow effect of the Pir Panjal Range must be very strong.

The pattern of isoglacihypses on the Himalayan flank is rather more difficult to explain. The presence of pronounced bulges suggests that precipitation and topography are important controls, since the major bulges coincide with the main valleys draining the Great Himalaya. GT gradients are highly variable, although the greatest value is found close to the watershed with Ladakh. This is to be expected, since the watershed marks a narrow zone of precipitation decline. It is possible that the pattern of isoglacihypses on the Himalayan flank is distorted due to the lack of data points. Although there are more data points on the Himalayan flank than the Pir Panjal, they are more widely spaced.

The isoglacihypses of current mean ELAs of glaciers are shown in figure 77. The procedure for calculating the mean ELAs, outlined in the previous section, has led to smoothing of the data. The pattern of ELAs shows some similarities to that for the GTs and some important differences. ELAs are generally of the order of 100 to 200 m lower than GTs, which is within the range of values found for other localities (eg. the Southern Alps of New Zealand: Porter, 1975a). Since the GT is based on summit heights, which will always be higher than equilibrium lines on glaciers, this difference in the ELA and GT is to be expected.

The trend in isoglacihypses for ELAs is similar to that for GTs. However, the steep gradient found in the GT data are not apparent in the ELA data, although the maximum ELA gradient is still steep: about 45 mkm^{-1}. The pattern of isoglacihypses is less complicated than that for GTs. However, this fact, together with the differences in the steepness of the gradients, may simply be a reflection of the smoothing procedures used to calculate mean ELAs. On the Pir Panjal flank in particular, the ELA gradient is much more gentle than the GT gradient. However, this is probably a reflection of the small number of data points used to determine isoglacihypses of ELAs.

The patterns of present glaciation can be summarized as follows. Firstly, the isoglacihypses indicate a broad increase in the height of glacierization from the Pir Panjal Range to the Great Himlaya, and from the Great Himalaya to Ladakh. Secondly, this increase shows both northward and north-eastward components. These are best explained by precipitation and topographic controls on glacierization. Thirdly, there are more localized variations in glacierization that are more difficult to explain. Some of the variations may arise from the low quality of the data whereas others are undoubtedly related to local precipitation variations, themselves linked to topography. Finally, the hypothesis of precipitation control on regional patterns of glacierization appears to be at variance with the available meteorological data. However, this is because the meteorological data are inadequate for the present purpose.

The pattern of past glaciation is indicated by two sources of evidence: isoglacihypses of the altitudes of north-facing cirques, and depressions of ELAs of the major glaciers.

Isoglacihypses of cirque altitudes are shown in figure 79. The variations in altitude of cirques, together with the height of the GT, are shown for a transect running south-west to north-east across Kashmir in figure 80. Both of these diagrams show that cirque altitudes are higher on the Pir Panjal flank than on the Himalayan flank. The differences in the mean, maximum and minimum cirque height between the two flanks are about 250 m. The spread of values on each flank is related to the fact that geological structure, as well as palaeoclimate, controls the location of cirques. Furthermore, some of the features on the maps may have been wrongly identified as cirques. However, despite the spread of values, the overlap between the altitudes of the Pir Panjal and Himalayan cirques is only about 100 m, excluding one very high cirque close to the watershed with Ladakh. Because of this, it is thought likely that the greater altitude of cirques on the Pir Panjal flank is a real effect and not the result of errors in the data.

The pattern of changes in ELAs adds little to the above arguments, since only three valleys were considered. However, depressions in ELAs are of considerable interest in their own right. ELAs were calculated for the maximum ice advances in the Sind, Liddar and Ningle Valleys. In addition, ELA depressions were calculated for advances of lesser magnitude for the West Liddar Valley. The maximum glaciated area was calculated for the whole of Kashmir. This was based on mapped ice limits in the three valleys mentioned above. Elsewhere, the ice limit was inferred from the mapped limits, taking into account the altitude and aspect of the individual valley. The total glacierized areas in Kashmir are listed in table 23.

Maximum depressions of the ELA vary from 740 m in the Ningle valley to 800 m in the Sind Valley. The value for the Ningle Valley must be taken with some reservation, however, since there are no active glaciers within the catchment for which a modern ELA could be determined. In the absence of modern glaciers, a value for the ELA in the Pir Panjal Range was taken from the Rembiara and Vishav Valleys (table 24).

The glacierized area for each major ice advance in the West Liddar Valley is also listed on table 23. This information has been plotted against the depression of the ELA for each advance to produce the area-altitude curve in figure 81. The glacierized area within a valley is a function of the equilibrium line and basin topography (Porter,

TABLE 23. PRESENT AND PAST GLACIERIZED AREA IN KASHMIR.

VALLEY		GLACIATED AREA (km^2)				
	PRESENT	G1	G2	G3	G4	G5
Lower Sind	26	52	..	..	..	..
Upper Sind	6	179	..	..	..	..
Total Sind	26	231	..	..	..	..
Lower Liddar	3	23	..	..	..	..
East Liddar	29	108	..	..	..	..
West Liddar	26	239	197	168	139	86
Total Liddar	58	370	..	..	..	..
Other Himalaya	00	32	..	..	..	..
Gt.Himalaya	84	633	..	..	..	..
Rembiara/Vishav	17	92	..	..	..	..
Other Pir Panjal	0	147	..	..	..	..
Pir Panjal	17	239	..	..	..	..

G5 = HOLOCENE
G1...G4 = QUATERNARY ADVANCES BASED ON W. LIDDAR VALLEY MORAINES

TABLE 24. PRESENT AND PAST ELA DATA FOR KASHMIR (ALL HEIGHTS IN METRES).

	HIMALAYAN FLANK			PIR PANJAL FLANK		
	MEAN	SD	N	MEAN	SD	N
GLACIATION THRESHOLD	4595 ±	117	19	4262 ±	144	9
ELAs OF NORTH-FACING GLACIERS[1]	4287 ±	215	51	4052 ±	177	24
ALTITUDES OF NORTH-FACING CIRQUE	3744 ±	153	60	3904 ±	163	36

[1] THAR = 0.4

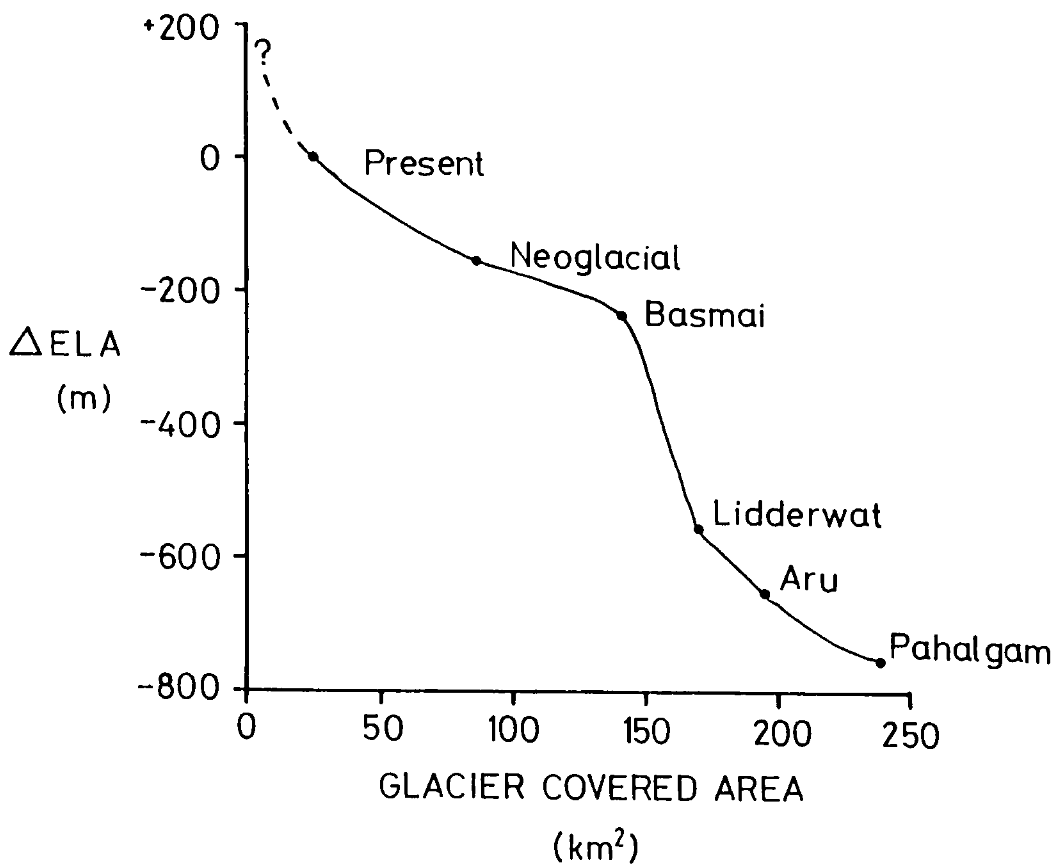


Figure 81

Area-altitude curve of glaciation in the West Liddar Valley.

1975a). The part of the curve above present ELA values is very steep, indicating that a rise in the equilibrium line of about 200 m would be necessary for the complete disappearance of glaciers. The steep slope of the curve is simply a reflection of the steep relief in the upper part of the basin. The gentle slope of the area - altitude curve between the present ELAs and those of the Basmai stage is a reflection of the topography in this part of the catchment, which is correspondingly gentle. Between the Basmai and Liddarwat stages, the area - altitude curve is once again steep. This is a reflection of the fact that many of the glaciers emerged from tributary valleys into the main West Liddar Valley during this stage. Since many of the tributary valleys are hanging above the main valley, a relatively large depression of the ELA would have brought about only a modest increase in glacierized area. For advances of greater magnitude than the Liddarwat stage, the area - latitude curve is gentle, a reflection of the gentle relief in the lower parts of the basin. Thus, a small change in the ELA would lead to a quite substantial change in glacierized area in the lower reaches of the valley.

6.6 Discussion: A Revised Model of the Late Quaternary Glacial History of Kashmir.

The preceding discussion can be synthesized to produce a revised model of the glacial history of Kashmir which is in stark contrast to previous work.

The maximum extent of glaciers during the Quaternary was much less than was previously thought. In this study, ice limits were mapped at Pahalgam in the Liddar Valley, Sonamarg in the Sind Valley and Gulmarg in the Ningle Valley. On the basis of mapping in these valleys and inference elsewhere, the maximum ice-cover in Kashmir during the Quaternary was about 870 km^2. This compares with an estimated present cover of 100 km^2. The maximum altitudinal descent of ice in Kashmir was to 2134 m a.s.l., in the Liddar Valley.

Evidence for only two major phases of glaciation, excluding Holocene advances, was found in Kashmir. In the Pir Panjal Range, such an incomplete record is probably largely due to uplift. Even if the range had been of sufficient altitude to support glaciers prior to the time when the oldest moraines were deposited, it is likely that any depositional evidence will have been destroyed by subsequent advances, which were of greater magnitude due to uplift. This hypothesis is well supported by independent estimates of rapid Quaternary uplift of the Pir Panjal Range (Burbank, 1982).

Radiometric dates on the West Liddar sequence suggest that the maximum ice advance in the Great Himalaya predates the last glacial maximum, if this is taken to be about 18 kaBP (Denton and Hughes, 1981). This is supported by the advanced state of weathering of the oldest moraines and the well-developed soils on their surfaces. The younger advances in the Pir Panjal Range have minimum ages that date them to about the last glacial maximum, although the actual deposition of the moraines may have been much earlier, given the problems surrounding the dates themselves. There is no evidence to suggest a major ice advance in Kashmir at the last glacial maximum, suggesting that the glaciation of Kashmir may have been out of phase with advances from other parts of the world.

The glacial succession of the West Liddar Valley suggests that advances were of progressively smaller magnitude. To some extent, this is a reflection of the nature of glacial systems, since overriding of deposits by a glacier tends to be destructive. However, there is evidence to suggest that the pattern of decreasing magnitude of ice advances in the West Liddar Valleys is a true reflection of changes in ice volume through time. It is likely that the ice advances represented by moraines in the West Liddar Valley took place whilst the Pir Panjal Range was undergoing rapid uplift. Thus, at around 400 kaBP, the Pir Panjal Range may have been between 4000 and 1400 m lower than present, using the uplift rate of 3.5 to 10.0 mma^{-1} for the Pir Panjal Range (Burbank, 1982). The reduced altitude of the Pir Panjal Range would have allowed the penetration of more moisture into the Great Himalaya. Under the assumption that the Great Himalaya has experienced relatively little uplift during the Middle and Late Quaternary, such an increase in precipitation would have enhanced the growth of glaciers. Thus, the progressive decrease in the size of glacial advances can be explained by uplift of the Pir Panjal Range and the associated decrease in precipitation in the Great Himalaya.

The pattern of cirque altitudes is the reverse of modern glaciation gradient as shown by GTs and ELAs. At first sight, this suggests a reversal of climatic gradients, since lower cirques would be expected with larger glaciers which would, in turn, occur in areas of higher precipitation. This suggests that precipitation was, some time in the past, greater on the Himalayn flank than the Pir Panjal. However, it is unlikely that the areas to the east of Kashmir would have been a source of moisture for Kashmir during a maximum glacial. Such areas include the Tibetan Plateau and the Tarim Basin, which are currently very arid. Evidence from oceanic sediments in the northern Indian Ocean (eg. Cullen, 1981; Duplessy, 1982) strongly suggests that glacial maxima were associated with aridity over southern Asia and that although the glacial north-east monsoon may have been strengthened, it was colder and drier. There is some recent evidence to suggest that winter westerly storm tracks over central Asia may have been stronger and, possibly associated with a slight increase in precipitation (see chapter 3 of this thesis). However, even if this was the case, precipitation gradients over Kashmir would not have been reversed.

In the absence of any plausible climatic explanation for the apparent change in glaciation gradients, it is necessary to look for some other cause. One possible explanation is that the pattern has been changed by differential tectonic uplift. If uplift followed cirque formation, the cirque altitudes observed today will obviously be higher than during glaciation. In reality, uplift prior to the last glaciation will have affected cirque altitudes, since cirques are preferential sites for the formation of glaciers and are likely to have been occupied more than once. In this case, differential uplift of a period extending back earlier than the last time the cirques were occupied will have influenced the trend in cirque altitudes.

However, in order to substantiate this hypothesis, independent evidence of uplift in Kashmir since the period of cirque formation is needed. Because the actual date of cirque formation is not known, this is difficult. However, there is evidence that the Pir Panjal Range has undergone rapid uplift over the past 400 kaBP at between 3.5 and 10 mma^{-1} (Burbank, 1982). During the same period, uplift of the Great Himalayan Range was limited. This mechanism could explain why the cirques in the Pir Panjal Range are higher than those in the Great Himalaya. A crude estimate of the amount of cirque uplift in the Pir Panjal range can be made under a number of assumptions. The first of these assumptions relates to the age of the cirques. The glacial stratigraphy of the Ningle Valley indicates that two major phases of glaciation have affected the Pir Panjal Range. From global chronology, this could span the last 200 ka at maximum. Assuming that the cirques were formed at the beginning of this period, it is possible to calculate their original altitudes from the uplift rates of Burbank (1982). If 3.5 mma^{-1} is chosen as the uplift rate, the highest cirque would have been 3500 m a.s.l. prior to tectonic uplift. If 10 mma^{-1} is chosen as the uplift rate, the highest cirque would have been at 2200 m a.s.l. In both of these cases, the range of altitudes of the Pir Panjal Cirques would have been lower than that for the Great Himalaya. There are three further points that can be made regarding these figures. Firstly, cirque development may have been initiated in the Pir Panjal Range before major glaciation, in which case the cirques would have been even lower when they were formed. Secondly, rates of uplift for the Pir Panjal may have been even higher than 10 mma^{-1} for the latter part of the Quaternary. Thirdly, the relatively low altitude of cirques on the Himlayan flank may reflect greater precipitation during the time when they were initiated, as a result of the less lofty Pir Panjal Range.

Radiometric dates indicate that glacial advances in the mountain flanks were accompanied by loess deposition in Kashmir Basin (see chapter 5). All evidence for glaciation postdates the drainage of the upper Karewa lake, although there is no implication in this statement that Kashmir was unglaciated prior to this time.

CHAPTER 7. REGIONAL AND GLOBAL CORRELATIONS.

7.1 Introduction.

One of the major aims of Quaternary research is to reconstruct spatial patterns of global climate for periods in the past in order to calibrate models of past climate and to test theories regarding the causes of climatic change. Although many reconstructions of Quaternary climate have been made, the density of global coverage is very low (see Bradley, 1985, pp.9-10). Furthermore, the spatial distribution of studies is very uneven. Whereas Europe, North America and parts of Australasia have been well studied, there is a lack of data for most other areas, including Asia. Additionally, most work has concentrated on the Late Quaternary. There are progressively fewer studies for the earlier parts of the Late Cainozoic. Thus, in order to achieve the aims of Quaternary science, further work into those areas and parts of the Late Cainozoic that are understudied, is necessary.

Individual records of Quaternary environmental change are limited in applicability to the region in which they were constructed. Thus, in order to illustrate purely local or regional effects within the sequence, it is necessary to correlate between one record and another. Continental records of Quaternary climatic change vary enormously in their completeness. However, they tend to be less complete than records from deep-ocean sediments. For this reason, it is most useful if individual Quaternary sequences can be correlated with the oceanic record. Ocean sediments, in many ways, provide the complete record of

Late Cainozoic climate. Many studies of deep-ocean sediments have been undertaken and a well-dated sequence of palaeotemperatures and, paticularly, palaeo-ice volumes, has been reconstructed.

Terrestrial sites that have good potential for correlation with the oceanic record include sequences from large tectonic basins, long lacustrine sequences and thick accumulations of aeolian loess. Quaternary glacial sequences present problems for correlation, since they are fragmentary and often difficult to date. However, variations in glaciers are of considerable interest in their own right, since they can illustrate local and regional changes in climate. Therefore, it is important to correlate individual glacial sequences with those from other areas and also with global records of Quaternary climate.

The three sedimentary units studied in this thesis, the lower Karewa formation, the upper Karewa formation and the glacial record, have been examined at very different levels of temporal resolution. The lower Karewa formation is a semi-continuous sequence that is well dated. However, the temporal resolution of the work carried out for this thesis is very coarse, due to the large sampling interval. Furthermore, clay minerals provide only low-order palaeoclimatic information. The work on the lower Karewa formation thus provides a generalized picture of climatic change over a relatively long time-period. An additional complication arises from the tectonic control of major sedimentological features of the lower Karewa formation. The work on the upper Karewa employed a finer sampling interval than that on the lower Karewa. Furthermore, ostracods are potentially more sensitive palaeoenvironmental indicators than clay minerals. The upper Karewa period was one of less active tectonism in Kashmir than the lower Karewa period. This means that the interpretation of the record has potentially fewer ambiguities. However, evidence from this study has shown that tectonic movements have affected the upper Karewa sediments as well. Although a continuous sequence, the upper Karewa has not been well dated, therefore chronological uncertainties remain. The glacial sequence proposed for Kashmir is, like many glacial sequences elsewhere, poorly dated and fragmentary. However, the respon of glaciers to climatic change and tectonic uplift is of major interest. In this chapter, the sequences reconstructed in this thesis will be correlated with data from both within Asia and from other parts of the globe.

7.2 The Lower Karewa Formation.

The work carried out on the clay mineralogy of the lower Karewa mudstones suggests a gradual change in clay-mineral assemblages during the Late Pliocene, between about 2.5 and 2.3 MaBP. There is a considerable body of evidence to suggest that important changes in the earth's climate took place during the Late Pliocene. This challenges the view that the Tertiary was a time of warm, 'preglacial' climate and the Quaternary a time of glacial-interglacial oscillations. In order to examine the significance of the change in lower Karewa clay mineralogy during the Late Pliocene, it is necessary to adopt a global view.

Major ~~transitions in~~ global climate are well documented from the geological record, although the evidence becomes less clear further back in geological time. Recent work indicates that important stages in the evolution of the Quaternary glacial-interglacial cycles can be traced back to the end of the Cretaceous Period, about 65 MaBP (eg. Buchardt, 1978; Crowley, 1983). However, there are also numerous studies that have focused on the critical period of climatic deterioration that occurred during the Late Pliocene. Much of this evidence comes from deep-ocean sediments. However, there are also terrestrial records, from both glacial and non-glacial environments, that span this period. This evidence is reviewed below and the possible causes considered.

Shackleton and Opdyke (1977) first extended the Northern Hemisphere oxygen-isotope record into the Pliocene with core V28-179, taken from the equatorial north Pacific. Despite problems of low resolution resulting from the low sedimentation rate, they were able to draw two important conclusions. Firstly, in the lower part of the record between 3.5 and 3.2 MaBP, the Northern Hemisphere ocean was in an isotopically interglacial state. Secondly, at about 2.5 MaBP, Northern Hemisphere glaciation began, with fluctuations of about two thirds of the magnitude of those in the Late Quaternary.

A considerable amount of information has been derived from the study of ice-rafted debris in ocean sediments. Shackleton et al. (1984) undertook isotopic, lithological and faunal analyses on cores from DSDP (Deep Sea Drilling Project) site 5528, in the northern North

Atlantic. The cores extended to before 3.5 MaBP. The onset of significant ice rafting occurred about 2.37 MaBP. This was shown by the influx of angular, sand-sized rock fragments, low calcium-carbonate values, isotopically-heavier ocean water and cold-climate nannofossils. It was argued that all of the above changes indicated Northern Hemisphere glaciation, since although Southern Hemisphere ice build-up could cause the isotopic changes, it would be unlikely to lead to significant ice-rafting in the northern North Atlantic. Some climatic variability was detected in the record prior to 2.37 MaBP, but this was of limited magnitude.

There is further evidence for the onset of ice rafting in the Late Pliocene. Loubere and Moss (1986) undertook faunal and isotopic analyses of cores from DSDP site 548, in the north-east Atlantic. Overall, they found that the surface waters of the north-east Atlantic had changed, in a stepwise fashion, from warm-temperate to subpolar during the period between 3.4 and 2.0 MaBP. Initially, between 3.4 and 3.1 MaBP, there was a cooling of ocean surface waters, followed by the reestablishment of temperate conditions. Between 3.0 and 2.6 MaBP, the faunal assemblages showed cyclical change, which led Loubere and Moss (1986) to suggest that this might be a response to orbital forcing. After 2.6 MaBP, the north-east Atlantic underwent progressive cooling, with the onset of Northern Hemisphere glaciation occuring about 2.4 MaBP.

Loubere (1988) carried out a high-resolution study of planktonic foraminiferal assemblages and oxygen-isotope ratios from benthic foraminifera on cores from DSDP site 548. In contrast to other studies, Loubere found that the onset of Northern Hemisphere glaciation was a response to rather gradual climatic forcing rather than a sudden event.

Sarnthein et al (1982) found a number of environmental changes at about 2.5 MaBP recorded in sediments from DSDP sites 366 and 397, off the north-west African continental margin. Their studies involved a combination of isotopic and lithological indicators. Their data revealed a period of climatic stability lasting from the Mio-Pliocene boundary, about 5.1 MaBP, until about 3.0 MaBP, after which a number of changes took place. During the period of change, the oxygen-isotope ratios of benthic foraminifera became progressively more negative and the positive extremes were less marked. However, the amplitude of isotope variations increased and this was accompanied by influxes of dust particles from the adjacent landmass.

Suc (1984) analyzed pollen assemblages in offshore sediments in the Western Mediterranean and was able to relate changes in the sequence to the structure of Mediterranean vegetation. Prior to about 3.2 MaBP, Suc inferred that the north-west Mediterranean supported a dense forest vegetation and he suggested that the climate was relatively warm and humid. A change in the pollen record occurred about 3.2 MaBP, with an increase in the number of xerophytic taxa present. It was suggested that, by this time, the summer-dry climate of the Mediterranean had evolved. After about 2.3 MaBP, the summer-dry climate and the characteristic Mediterranean vegetation became permanent features of this region.

Rea and Schrader (1985) examined the diatom biostratigraphy and ice-rafting history of cores from DSDP sites in the north Pacific. They found a pattern of change in climate similar to that for the north Atlantic ocean and Mediterranean basin. A major climatic threshold, indicated by both the diatom flora and ice-rafting events, was crossed between 2.5 and 2.4 MaBP. This was associated with Northern Hemisphere cooling and the accumulation of continental ice.

Blanc _et al_. (1983) suggested that the accumulation of ice in the Northern Hemisphere continents was time transgressive. They examined records of ice rafting and faunal assemblages for DSDP site 116, in the Hatton-Rockall Basin. They found no evidence for cooling or ice rafting at 3.1 MaBP when the initial phase of Late Pliocene climatic decline occurred and when boreal ice caps had become established in the western North Atlantic. The first influx of ice-rafted grains occurred around 2.7 MaBP at DSDP site 116. However, planktonic foraminifera indicated that there was still little change in surface-water temperature by this time. Both ice rafting and temperature decline only became significant about 2.3 MaBP. Blanc _et al_. (1983) suggested that this evidence pointed to a lag in the onset of Northern Hemisphere glaciation between the western and eastern sides of the North Atlantic. This lag was of the order of 0.8 MaBP, with the initiation of ice caps taking place, firstly, on the western side of the North Atlantic, about 3.1 MaBP. Blanc _et al_. (1983) suggested that the late development of ice caps in the eastern part of the North Atlantic could be explained by the persistence of the North Atlantic Drift ocean current in this area, between 3.1 and 2.3 MaBP. This hypothesis reconciles the apparent disparity between the age of the onset of glaciation in the western and eastern parts of the North Atlantic.

In the Southern Hemisphere, the chronology of glaciation is somewhat different and ice caps developed rather earlier than in the north (eg . Kennet, 1977). The onset of substantial sea-ice formation occurred as early as 38 MaBP,although it was not until the Middle Miocene (14 to 11 MaBP) that the Antarctic ice cap formed.

The abundant oceanic evidence of Pliocene climatic change is largely supported by the terrestrial evidence. For example, magnetostratigraphical dating of the Chinese loess suggests that the onset of deposition occurred around 2.5 MaBP (eg. Heller and Liu, 1982, Kukla, 1987). Smith (1984) found a major dry phase in sediments from Searles Lake in California, USA, around 2.5 MaBP. Bonnefille (1983) found evidence for a cold, arid climate in the polled record of lacustrine deposits in the Ethiopian highlands around 2.5 MaBP. Zagwijn (1974) identified a major transition in the north-west European vegetation, shown in the pollen record, together with the onset of a major cold phase, around 2.5 MaBP.

The above studies suggest that global cooling led to terrestrial glaciation prior to the beginning of the Quaternary. There is also direct evidence for terrestrial glaciation in some areas, where undoubted tillites have been found and dated. For example, Curry (1966) found a tillite in the Sierra Nevada, California. This tillite overlies andesite dated to 3.1 MaBP and underlies latite which was dated to 2.7 MaBP. Thus, the tillite provides direct evidence of Pliocene glaciation in the Northern Hemisphere. Mercer and Sutter (1982) found glacial tillite interbedded with volcanic deposits in southern Argentine (45^{0}S). The volcanic deposits, and hence the till, were dated to between 7 and 4.6 MaBP. In southern Alaska, Denton and Armstrong (1969) found extensive evidence of glacial deposits dating from the Mio-Pliocene. In Iceland, McDougall and Wensink (1966) found a tillite overlying basalt that was dated to 3.1 MaBP. From this, they suggested that Iceland was glaciated in the Late Pliocene.

In summary, global evidence suggests that a major climatic threshold was crossed in the Pliocene. Although the Southern Hemisphere was glaciated before this time, a significant climatic event, related to the onset of Northern Hemisphere glaciation, seems to have occurred between 2.5 and 2.3 MaBP.

A number of theories have been put forward to explain the Late Pliocene climatic threshold. Most authors agree that it cannot have resulted from orbital forcing, of the type responsible for the Quaternary glacial-interglacial cycles since orbital variations do not occur on an appropriate timescale. More usually, palaeogeographical changes are advocated. Two particular changes often regarded as important are the closing of the Panama seaway and the opening of the

Bering Straits. Both of these events occurred between about 4.0 amd 3.0 MaBP. Rea and Schrader (1985) outlined the classic palaeogeographical explanation of Late Pliocene climatic change. They argued that the closing of the Panama seaway deflected the equatorial current into the Gulf stream, transporting large amounts of relatively warm water to higher latitudes. This led to an increase in evaporation in the high latitudes of the Northern Hemisphere, and so to the development of ice caps. However, as Rea and Schrader (1985) point out, the onset of major Northern Hemisphere glaciation postdated the closure of the Panama seaway by about 1 Ma. Furthermore, work by Ruddiman and McIntyre (1984) has cast doubt upon the role of moisture influx in the initiation of ice sheets. They argued that ice-sheet changes are better explained by orbitally-controlled insolation variations amplified by the ocean, rather than direct responses to moisture supply. Volcanic activity cannot be used to explain the Late Pliocene climatic decline, since this was at a minimum at about 2.5 MaBP.

It is clear from the previous discussion that the causes of the Late Pliocene climatic decline are not fully understood. If the model of Blanc et al. (1983) is accepted as the cause of the lag between ice build-up on the western and eastern sides of the North Atlantic, it is possible that the closure of the Panama strait was an important control. However, the interactions between atmosphere, ocean and land that caused simultaneous glaciation, loess deposition and vegetation change must have been complex.

The changes in Northern Hemisphere climate during the late Pliocene occurred against a background of tectonic uplift in the Himalayan Range. This uplift may itself provide an explanation of why the earth's climate changed around 2.5 MaBP. As outlined in chapter 3 of this thesis, there is now evidence for rapid uplift of the Himalaya - Tibetan Plateau during the Late Tertiary. The upward projection of an obstacle as high as 4 km must have had a considerable effect on atmospheric circulation beyond the immediate area of south-central Asia. Such effects would have included diversion of the westerly airflow by the mountain barrier and the establishment of the South Asian monsoonal flow due to summer heating and winter cooling of the uplifted areas. The importance of major mountain-building processes in changing atmospheric circulation has already been suggested (eg. Rea and Schrader, 1985). In Kashmir, the response in the clay mineral record was a change from conditions favouring the formation and preservation of kaolinite to those favouring smectite at about 2.5 to 2.3 MaBP. This result suggests that Kashmir became more arid during the late Pliocene.

The final question remaining in this section relates to the onset of glaciation in Kashmir. As was stated in chapter 4, there is no direct evidence for glaciation in the Karewa deposits. Furthermore, there are no known examples of pre-Quaternary tills in the Himalayan Range as a whole. However, this is likely to be due to poor preservation and does not, in itself, prove that the range was unglaciated prior to the late Quaternary. In the examples of pre-Quaternary tills cited earlier in this section, preservation resulted from burial beneath volcanic deposits. Since Late Cainozoic volcanic activity was absent from the Himalaya, the means for similar preservation was unavailable.

The fundamental question as to the age of the first Himalayan glaciation rests, therefore, not on the acquisition of direct depositional evidence, but on the timing of Northern Hemisphere climatic deterioration and the attainment of sufficient altitude in the range to support glaciers. Evidence of the Late Pliocene climate of Kashmir presented in this thesis, together with examples of pre-quaternary glaciation known from other mid- to low-latitude mountain ranges, suggests that glaciers may have developed in the Great Himalayan Range in Kashmir during the Late Pliocene had the mountains been sufficiently high. Unfortunately, rates of Himlayan uplift are insufficiently well resolved to determine when sufficient altit de may have been attained. The onset of glaciation in the Pir Panjal Range must have occured much later, with the rapid uplift of the range during the Middle and Late Quaternary.

7.3 Upper Karewa Formation.

The upper Karewa spans the time-period from 400 to about 5 kaBP. The major facies changes, from conglomerate to fine-grained lacustrine sediments to loess, reflect basin-margin tectonic activity. However, the individual units themselves, especially the lacustrine beds and the loess, have considerable palaeoclimatic potential.

The whole of the upper Karewa therefore falls into the Middle- and Late-Quaternary. With the exception of ocean-core sequences, the Middle Quaternary is relatively poorly studied. Although the evidence may not be as well preserved compared with the Late Quaternary record, this is often not the case. A more important problem is one of dating. Whereas Late Quaternary sediments can be dated by a number of means, those of suspected Middle Quaternary age fall into an unfortunate gap in the age ranges of available techniques. However, TL dating has been succesfully applied to the loess in Kashmir, and also to the upper part of the lacustrine beds. The Middle to Late Quaternary is of considerable palaeoclimatic interest since it spans more than one glacial-interglacial cycle. The most promising sequences for correlation with the upper Karewa are the oceanic record of global ice volumes (eg. Shackleton and Opdyke, 1973); the Chinese loess (reviewed by Kukla, 1987); long lacustrine sequences such as that from Lake Biwa in Japan (eg. Horie, 1987); and the lower parts of the ocean cores from the northern Indian Ocean which indicate the changing intensity of the South Asian monsoon (Van Campo et al., 1982; Prell and Van Campo, 1986; Fontugne and Duplessy, 1986).

In figure 82, the main upper Karewa events are compared with the standard oxygen-isotope stratigraphy of Shackleton and Opdyke (1973). The whole of the upper Karewa sequence spans part of four glacial-interglacial cycles and includes terminations IV, III and II. Rapid uplift of the Pir Panjal, which commenced about 400 kaBP (Burbank, 1982) occurred in isotope stage 10 or 11. Soon after the onset of Pir Panjal uplift, the Karewa lake shifted away from the south-west margin

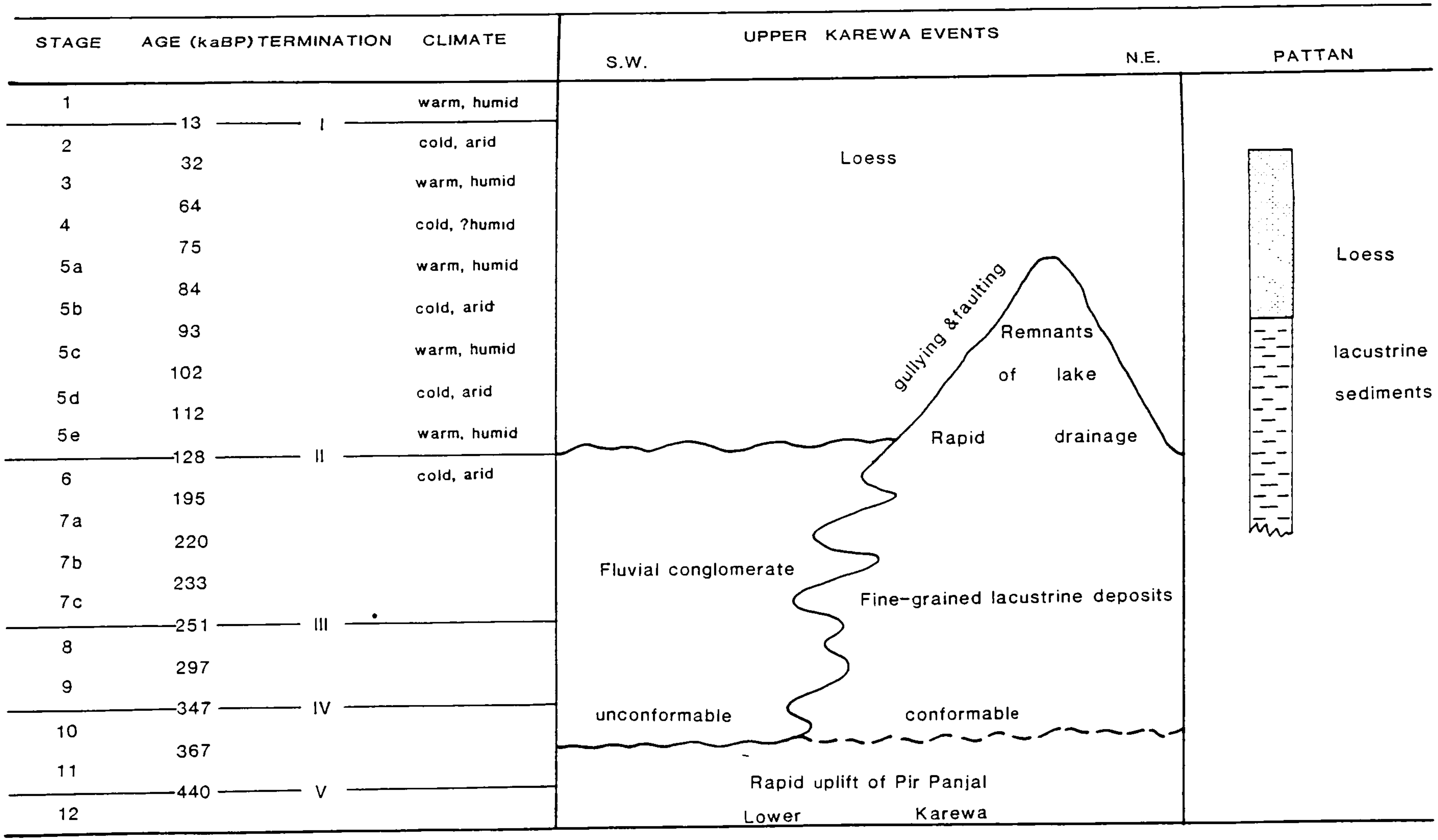


Figure 82 Summary of Upper Karewa events.
(Ages of isotopic boundaries from Shackleton & Opdyke, 1973; Molfino et. al. , 1984; and Kukla, 1987.
. Information on South Asian monsoon climate from Fontugne & Duplessy, 1986 and Van Campo et. al., 1982.)

of the basin. This event was accompanied by the deposition of tectonic alluvial conglomerates, originating from the Pir Panjal flank. Thus between 400 and 100 kaBP, deposition in Kashmir was characterized by alluvial fan and braided stream conglomerates in the south-western part of the basin, and fine-grained lacustrine deposition in the centre and on the north-eastern flank.

Conglomerate deposition continued until about 300 or 128 kaBP, depending on which chronology of the loess is accepted. By this time, the alluvial fan - braided river systems had become at least partially stabilized, and loess deposition had begun on their surfaces. However, neither TL chronology suggests that the cessation of conglomerate deposition occurred simultaneously in all parts of the basin. Localized braided-stream and alluvial-fan activity may have occurred alongside loess deposition. Dating suggests that conglomerate deposition had largely terminated by isotope stage 5d or 5e under Rendell's chronology or as early as stage 9 or 10 using the chronology of Singhvi et al. (1987).

Rapid drainage of the upper Karewa lake occurred just after 115 to 100 kaBP, in isotope stage 5b or 5c in both chronologies. However, small remanants of the lake persisted until about 80 kaBP,or stage 5a. Abundant evidence suggests that the drainage of the lake was not a response to global climatic forcing as reflected in the oceanic oxygen-isotope record.Firstly, although the change from lacustrine sediments to aeolian loess suggests that a major environmental change accompanied lake drainage, sedimentological analyses of the lacustrine deposits strongly suggest that aeolian loess was being deposited while the lake was still in existence. Secondly, there is no gradual change in the ostracod fauna of the Pattan section which might suggest progressive desiccation. If aridity had caused the shrinkage of the lake, it is likely that the ostracod fauna would have indicated increasing salinities. Thirdly, the upper part of the lacustrine beds show signs of minor tectonic disturbance, which could have accompanied tectonically-induced drainage of the lake. Although small remnants of the upper Karewa lake seem to have persisted following rapid drainage, this does not correspond to progressive desiccation. From this evidence, it is concluded that the major facies changes in the upper Karewa were responses to tectonic activity rather than climatic forcing.

Changes within each of the three upper Karewa sedimentary units may, however, be responses to climatic forcing. This is more likely to be true for the loess and the lacustrine beds than the conglomerate, since fluvial strata commonly show complex and undecipherable climatic responses.

The occurence of sand units within the upper Karewa lacustrine sediments probably reflects changes in the loci of fluviodeltaic deposition as well as to climatic changes. Changes in the ostracod assemblages are more likely to reveal palaeoclimatic information. Comparisons with the oxygen-isotope record show that ostracod productivity was low during isotope stage 6, the penultimate glacial. Productivity was low during stage 7a, but increased dramatically in the latter part of stage 6. If global climatic amelioration is equated with increases in ostracod productivity, there is a discord between the global oxygen-isotope record and the ostracod sequence from Pattan. This may be a result of uncertainties in dating. However, if, as is likely, the upper Karewa lake had an outlet, it would not have been very sensitive to climatic changes in terms of lake-level fluctuations or water chemistry.

Loess sequences are often regarded as excellent indicators of continental palaeoclimate. However, in the case of the Kashmir loess, there is no unambiguous relationship with global climatic change. According to the dates of Singhvi et al. (1987) on the 'long' loess sequences from Karpora and Dilpur, palaeosols in the upper, dated parts of the sections formed at aproximately 20 ka intervals. This suggests that palaeosol development in Kashmir was forced by orbital variations on the 21 ka precessional cycle, which dominates in the lower latitudes (Berger, 1978). However, if this relationship between palaeosol formation and orbital forcing is correct, the base of the loess at Karpora and Dilpur should date to about 180 kaBP, on the basis of the number of palaeosols present (see figure 49).This is much younger than the estimate of 300 kaBP from Singhvi et al. (1987) but older than Rendell's dates suggest. However, Rendell's dates do not suggest that the palaeosols developed every 21 ka. Clearly, more TL dates are required in order to solve this problem. In the Chinese loess, palaeosol fomation appears to follow the 21 ka precessional cycle quite clearly (eg. Burbank and Li, 1985; Kukla, 1987). The reasons why this may be so are best discussed by examining the Chinese loess in some detail.

The Chinese loess Plateau provides the best example of loess on earth. Kukla (1987) examined the loess sequence at Xifeng, on the Chinese loess Plateau. The age of the loess was determined by magnetic polarity stratigraphy and the ages of horizons between the major geomagnetic boundaries were established by interpolation. However, it is well known that palaeosols accumulate at a much slower rate than unaltered loess. Therefore, a unit thickness of palaeosol represents a much greater amount of geological time than the same thickness of loess. Using low-field magnetic-susceptibility data, Kukla (1987) was able to weight the sequence so as to allow for this. For the Chinese loess, this was possible since the palaeosols are characterized by low susceptibility values. The result of this exercise is a weighted time-series of low-field susceptibility data. Both the amplitude and timing of susceptibility peaks were found to correspond very well to the peaks in the oxygen-isotope data from deep-sea cores. Both time series showed distinct 100 and 40 ka components in addition to the 21 ka periodicity previously found.

Kukla (1987) found that differences in the low-field susceptibility between the loess and the palaeosols were due to differences in the concentration, but not composition, of the magnetic components. Kukla argued that these differences resulted from source-area changes rather than <u>in situ</u> weathering. A model was put forward to explain these differences. Kukla argued that the high susceptibility components of the sediment were derived from distant sources, such as the Mongolian desert or Pacific Volcanoes. Such sources would be transported at high level in the troposphere and not be affected by changes in climate. Low-susceptibility components, on the other hand, were thought to have been derived from more localized sources. Their accumulation would have been more closely controlled by climate. Thus, changes in susceptibility in the Chinese loess sequences reflect variable inputs of local silt against a constant background input of far-travelled components. Kukla (1987) argued that, during cold and dry stages, deposition from local sources would be at a maximum since the loess Plateau would have been sparsely vegetated. Thus, the loess would have low magnetic susceptibility. By contrast, the plateau would have been well vegetated during warm, humid stages and localized sources of dust would have been supressed. Therefore, the resulting sediments would have high susceptibilty. Kukla therefore argued that low-field magnetic-susceptibility provides a sensitive indicator of vegetation change.

In Kashmir, the palaeosols have higher magnetic susceptibility than the loess (Kusumgar et al., 1986). Loess-palaeosol magnetic susceptibility models fall into two categories. The first and more familiar of these assumes that magnetic minerals are altered by pedogenesis from haematite to magnetite, thus enhancing the susceptibility relative to unaltered loess. This can be termed the '*in situ* weathering model.' The second model incorporates the ideas of Kukla (1987) and may be called the 'variable source model.' The applicabiltiy of either of these models to the Kashmir loess remains unclear. Hoewever, detailed stratigraphic work and TL dating in the loess shows that it is unclear whether the loess-palaeosol record in Kashmir responded to orbital forcing in the same way as the Chinese loess, because of the existence of two conflicting chronologies.The applicability of the variable source model depends on a knowledge of the magnetic components of the loess and palaeosols. Such information is currently lacking for the Kashmir loess. It is possible, however, that variable sources of silt in Kashmir may have arisen due to tectonic changes in depositional systems, such as braided streams, within the basin. This might explain why the loess-palaesol record of Kashmir does not appear to reflect orbital forcing in the same way as the Chinese loess. Alternatively, the apparent lack of orbital forcing may be due to problems with the chronologies

Additional Middle to late Quaternary sequences from outside Kashmir do little to aid the interpretation of the upper Karewa record. The long sequence from Lake Biwa in Japan covers the time-period of the upper Karewa. However, like the Chinese loess, the sedimentological, palaeontological and geochemical changes in the sediments appear to reflect orbital climatic· forcing (eg. Horie, 1987).

With the exception of the tectonically-induced conglomerate deposition and the subsequent restriction of the lake, the upper Karewa has been regarded as a period of relative tectonic stability. However, it is now clear that this was not the case and that the whole of the upper Karewa period has recorded the delicate interplay between climatic and tectonic events. All of the major facies changes within the upper Karewa formation are better explained by tectonics than climatic change. This includes the onset of terrestrial, rather than waterlain loess deposition, as well as the onset of conglomerate deposition. The upper Karewa lacustrine beds show some climatic control, in variations in the ostracod assemblages. However, facies changes within this unit, particularly alternations from silt-clay to sand probably reflect migration in fluviodeltaic systems as well as climatically-induced lake-level fluctuations. Such migrations could, in turn, be responses to tectonic uplift or simply results of the tendency of deltaic systems to migrate laterally without external forcing. Finally, the loess-palaeosol sequences may have been out of phase with the 21 ka precessional cycle of orbital variations that predominates at lower latitudes. This may be because the sequence responded to subtle interactions between climatic change and tectonic uplift within the basin. However, this cannot be stated confidently without additional chronological information.

7.4 Quaternary Glaciation.

At the beginning of chapter 6, the lack of a review of Himalayan glaciation was noted. There have, however, been a number of partial reviews of the topic. For example, Mercer reviewed the occurrence and characteristics of modern glaciers in the Himalaya (Mercer, 1975a) and Karakoram Range (Mercer, 1975b). The International Association for Hydrological Science Publication 'World Glacier Inventory' (IAHS-AISH, 1980) contains several chapters on this region (eg. Müller on the Mount Everest Region, Higuchi et al. on eastern Nepal, Kick on the Nanaga Parbat Massif, and Shroder on Afghanistan). However, none of these studies attempts a regional synthesis of both present and past glaciation of the Himalayan Range. Kalvoda (1980) produced a paper titled 'A review of the Quaternary glaciation in the Himalayas.' However, this is inadequate since it fails to consider many recent studies.

In this section, a regional synthesis of Himalayan and Trans-Himalayan glaciation is attempted. Firstly, the present distribution of glaciers will be considered. Secondly, the timing and magnitude of Quaternary glacial advances will be discussed. Thirdly, the climatic background to Quaternary glaciation is considered by examining non-glacial evidence for the Late Quaternary in Southern and Central Asia. In addition, selected global data on glaciation are considered, both from mid- and low-latitude areas. This is important since it allows the Himalayan sequences to be placed in a global framework.

A considerable number of observations have been made of modern glaciers in the Himalaya. Although not all of these were by earth scientists, most of the data are of some value for the present purpose. A synthesis of the distribtuion of glaciers in High Asia was undertaken by von Wissman (1959), who plotted isoglacihypses of the height of the modern snow line. Although the isoglacihypses are based on observations in some areas, in others they have clearly been extrapolated from measurements elsewhere. For example Porter (1970) noted a discrepancy between ELAs measured on glaciers in the field in Swat Kohistan during 1968, and those on von Wissman's map of nearly 300 m. Similarly , there is a discrepancy between the pattern of snowlines for Kashmir shown on von Wissman's map and the estimated ELAs in this study. Von Wissman's map shows the pattern of isoglacihypses for Kashmir quite clearly. Snow-line altitudes are shown as 4400 m a.s.l. for both the Himalayan and Pir Panjal flanks. A trend of increasing snow-line altitude runs south-west to north-east in the Himalayan Range as a whole. However, no increase in the snow line is shown within Kashmir basin itself. Therefore, although the increasing snow line between Kashmir and Ladakh is shown clearly on von Wissman's map, the spatial variation in snow lines within Kashmir was not recognised. For the Pir Panjal Range, von Wissman's etimates of the snow line are too high by between 400 and 500 m. For the Himalayan Range, the overestimate is smaller: up to 300 m. This is consistent with over estimation of snow line altitudes in Swat Kohistan noted by Porter (1970).

Although von Wissman's absolute values for modern snow lines may be too high, the pattern of isoglacihypses provides an interesting insight into the spatial pattern of Himalayan glaciation. The most apparent pattern on the map is the increase in snow-line altitude from the Himalayan front to the inner Himalaya and Tibet. Thus, the gradient runs south to north in the Swat Kohistan area; south-west to north-east in the Karakoram Range, Kashmir and Ladakh; and south to north in Nepal. Snow-line altitudes are lowest in north-west Himalaya: the 4400 m isoglacihypse runs from Lahaul, to the south-west of Kashmir through to Swat Kophistan. The lowest mapped isoglacihypse in Nepal and central Himalaya is 4800 m The height of the snow line rises progressively towards the north and north-east, reaching altitudes of 6400 m a.s.l.

The pattern of glacierization depicted on von Wissman's map strongly suggests that precipitation is an important control on snow-line altitude throughout the Himalayan Range. A distance-decay effect of winter and summer precipitation was proposed for Kashmir in chapter 6 of this thesis. A similar situation appears to exist in the rest of the Himalaya. Values for observations of ELAs and snow-line altitudes are listed in table 25 and depicted graphically in figure 83. These data support the overall pattern of isoglacihypses shown on von Wissman's map. Some of the older data were actually used by von Wissman in the construction of his map,therefore there is no discrepancy between the two sources.

TABLE 25. PRESENT ELAs FOR HIMALAYAN AND TRANS-HIMALAYAN GLACIERS.

LOCATION	PRESENT ELA (m)	REFERENCE	ELA FROM VON WISSMAN (1959) (m)
Kashmir Pir Panjal	3900-4000	This study	4400
Kashmir Gt. Himalaya	4100-4400	This study	4400
Samir Valley Hindu Kush	~4900*	Gilbert _et al._ (1969)	4800
Ladakh Range (South side)	5200-5400	Burbank & Fort (1985)	5600
Zanskar Range (North side)	5200-5400	Burbank & Fort (1985)	5600
Nanga Parbat	4500-4700	Finsterwalder (1935)	4800
Garwhal-Dhauli & Lissar Ranges	5200-5700	Grinlington (1914)	4800-5800
Swat Kohistan	4100-4400	Porter (1970)	4400-4800
Gara Glacier Himachal Pradesh	5050*	Bhandari _et al._ (1983)	~5200
Changme-Khangpu Glacier, Sikkim	5300*	Bhandari _et al._ (1983)	~5500
Nangpo Basin Mt. Everest	5540±150	Müller (1980)	5000-5600
Dudh Kosi Basin Mt. Everest	5530±120	Müller (1980)	5000-5600
Imja Kola Basin Mt. Everest	5580±180	Müller (1980)	5000-5600
Mt. Everest (North side)	5800	Williams (1983)	~6000
Mt. Everest (South side)	5200	Williams (1983)	~5400-5600
Rakaposhi Range	4700	Kick (1964)	5000
Karakoram (North side)	~5500	Visser & Visser-Hooft (1938)	5000
Karakoram (South side)	5000-	Visser & Visser-Hooft (1938)	4800

* Denotes observation on single glacier.

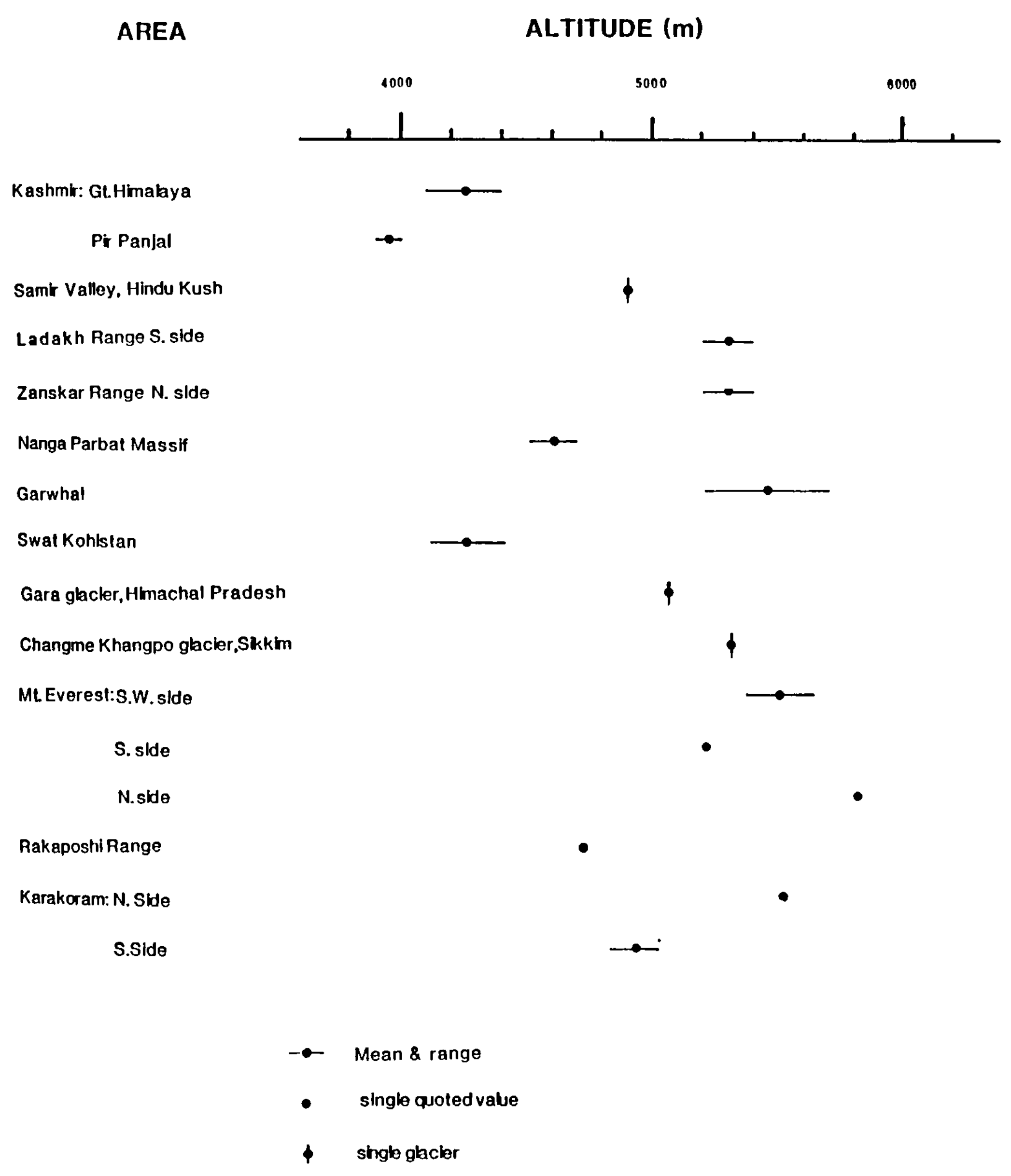


Figure 83

Summary of modern ELAS for Himalayan glaciers.

Work by Mayewski and Jesche (1979) and Mayewski et al. (1980) attempted to link fluctuations in Himalayan and Trans-Himalayan glaciers to changes in the monsoon circulation. This work is important for the understanding glacier-climate relationships for this area, since both variables were determined accurately. Information on the former positions of glacier snouts was derived from field observations made in the 19th and 20th centuries. From a study of 112 individiual glaciers, Mayewski and Jesche (1979) concluded that Himalayan and Trans-Himalayan glaciers had been in a state of net retreat since AD 1850. However, there had been phases of advance, stillstand and retreat throughout the region during this time.

Mayewski et al. (1980) argued that the Himalayan and Trans-Himlayan glacier fluctuations were largely a response to changes in the heat and moisture influx onto the Asian landmass during the period of the summer monsoon. Their comparison of the glacier fluctuations and meteorological records showed that Trans-Himalayan glaciers advanced during a period of reduced summer rainfall, between 1890 and 1910. The authors suggested that, whereas glaciers would tend to retreat rather than advance during periods of reduced rainfall, a lag between high rainfall and glacier advance would be expected because of the response time of a glacier to environmental changes. In contrast, Himalayan, as opposed to Trans-Himalayan, glaciers, did not advance between 1890 and 1910. Mayewski et al. (1980) suggested that this might be because Himlayan glaciers occur at lower altitudes than Trans-Himlayan glaciers. As a result, Himlayan glaciers have higher ablation rates and therefore lower response times to environmental change.

These data suggest that there may be a relationship between summer monsoon precipitation and glacier fluctuations during the recent historical period. This suggests that the summer monsoon may be an important source of moisture for Himalayan glaciers. However, no indication of the changes in winter precipitation is included in the analysis by Mayewski et al. (1980). There is evidence to suggest that Himalayan glaciers may undergo simultaneous accumulation and ablation throughout the summer (eg. Ono, 1985; Collins, 1988). Despite this, winter precipitation must be an important source of moisture for glaciers. This is partc-ularly so for western Himlaya, since most of the winter precipitation is associated with westerly depressions.

Extending this sort of analysis back into the Late Quaternary presents additional problems. It is difficult to map and date the former positions of glaciers, and independent palaeoclimatic data are often lacking. The lack of well-dated glacial sequences makes anything but a generalized review of the Late Quaternary glaciation of the Himalayan Range impossible. With these limitations in mind, the glacial history of the Himalayan Range will be reviewed below and evaluated in the light of independent palaeoclimatic evidence.

Depressions of ELAs for Himalayan glaciers during the Late Quaternary are listed in table 26. The values range from 300 to 1200 m below present values. This clearly represents quite wide variation in

TABLE 26. PRESENT AND PAST ELAs FOR HIMALAYAN GLACIERS AND TRANS-HIMALAYAN GLACIERS.

LOCALITY	PRESENT ELA (M)	QUATERNARY ELA (M)	ELA DEPRESSION (M)	REFERENCE
Kashmir				
Sind Valley	4300	3500	800	this study
Liddar Valley	4100	3300	760	this study
Ningle Valley	3900	3200	700	this study
Swat Kohistan	4100-4400	3000-3200	~1000	Porter (1970)
Ladakh Range (South side)	5200-5400	4300	~1000	Burbank & Fort (1985)
Zanskar Range (North side)	5200-5400	4700	~600	Burbank & Fort (1985)
Mt. Everest				
S. side	5200	4300	900	Williams (1983)
N. side	5800	5500	300	Williams (1983)
S.W. side	5550	4950	600	Müller (1980)
Pamirs	Not given	Not given	800-1000	Agakanyantz *et al.* (1981)
Tibetan Plateau	5900	4700	1200	Kuhle (1987)
Afghanistan Hindu Kush	3000-3600	4000-4600	1000	Porter (1985)

Himlayan glaciation during the Quaternary. The data on ELA depressions for the Himalaya fall into four groups. The first of these includes the Tibetan Plateau to the north of Mount Everest, for which Kuhle (1986) claims a maximum ELA depression of 1200 m. The second area includes glaciers with ELA depressions of about 900±100 m. It includes the Sind Valley in Kashmir, Swat Kohistan, the southern side of the Ladakh range, the southern side of Mount Everest and the Pamir Range. ELA depressions in this group do not differ markedly from values encountered elsewhere in the world (Porter, 1975a). The third group had ELA depressions of 700±100 m. Areas in this group include the Liddar and Ningle Valleys in Kashmir, the northern side of the Zanskar Range and the south-western side of mount Everest. Only one area falls into the fourth group. This is the area to the north of Mount Everest, which had a maximum ELA depression of only 300 m.

The ELA depressions for Kashmir are only slightly lower than the 'typical' global figure of 900±100 suggested by Porter (1975a). It has already been suggested, in chapter 6 of this thesis, that ELA depressions in Kashmir may have been modified by tectonic uplift. Significantly, the maximum ELA depression for Swat Kohistan was greater than that for Kashmir and Swat Kohistan has been tectonically stable during much of the Quaternary (Zeitler et al., 1982b). The difference between the ELA depressions on the south side of the Ladakh Range and the northern side of the Zanskar Range reported by Burbank and Fort (1985) has been explained by bedrock topography. The authors argued that glacier advance in the Zanskar Range had been severely constrained by valley topography and that this, rather than any climatic differences between the two ranges, explained the varying ELA depressions. However, this argument is not universally accepted (H. Osmaston, personal communication). Williams (1983) argued that the difference in Late Quaternary ELA depressions between the north and south sides of Mount Everest could be explained by Quaternary aridity. He argued that, during Quaternary glacial advances, the Tibetan side of Mount Everest became hyperarid, which restricted the development of glaciers. However, Williams' information for the Tibetan side of Mount Everest was derived from Landsat imagery and may be subject to a considerable degree of error. Müller (1980) found that ELA depressions on the south-west side of Mount Everest were 600 m below present values, and so intermediate between depressions for the northern and southern sides.

The paucity of studies on Himalayan ELA changes during the Quaternary precludes firm conclusions about the controls on the magnitude of former glaciation. However, a number of observations can be made. Firstly, the two areas with relatively low modern ELAs, Kashmir and Swat Kohistan, also had relatively large maximum ELA depressions during the Quaternary. Secondly, it is difficult to reconcile the large ELA depression in Ladakh with the small depression to the north of Mount Everest. Both of these areas are currently arid and the evidence discussed above suggests that Quaternary aridity during ice advances greatly retarded the development of glaciers on the northern side of Mount Everest. However, this apparently did not limit the development of glaciers in Ladakh, even though it is in a similarly interior position to the northern flank of Mount Everest.

Assuming the observations on ELA depressions, and the interpretations made from them, are correct, climatic conditions in the western part of the Tibetan Plateau, as shown by ELA depressions in the Ladakh Range, must have been different during ice advances to those to the north of Mount Everest. The inferred ELA depressions obtained for the southern side of the Ladakh Range (Burbank and Fort, 1985) are supported by the figures obtained for Kashmir in this study. The difference between the modern ELA in Kashmir and Ladakh is about 1200 m. This compares with a difference of about 1000 m. for the maximum ELA depression in each area. The depression calculated for Kashmir can be taken as a minimum value, since the actual depression will have been modified by uplift, at least for the Pir Panjal Range. The fact that there is some degree of similarity between maximum ELA depresions in Kashmir and Ladakh suggests that lack of moisture cannot have retarded glacier development in the latter area. This contrasts with the situation on the northern flank of Mount Everest. One possible explanation is that the source of moisture feeding Kashmir and Ladakh did not feed the Mount Everest region. If the winter westerly depressions were the main source of moisture to north-west Himalaya during periods of ice advance, this may 'explain the limited development of glaciers in the northern side of Mount Everest, since this may have been beyond the limit penetration of westerly depressions. The isolation of the northern flank of Mount Everest with repect to moisture is reflected in modern ELAs which are higher than any other area of the Himalaya or Trans-Himalaya (table 25).

The estimates of maximum ELA depression for the Tibetan Plateau by Kuhle (1986, 1987) are clearly at variance with the pattern discussed above, especially to the north of Mount Everest (see also table 26). In contrast to much of the recent work on glaciation in High Asia, Kuhle has suggested that the Tibetan Plateau was covered by a vast ice cap during the Late Quaternary. Kuhle (1987) argued that many of the features of glacial deposition in the Tibetan Plateau are unique to that region and bear little resemblance to glacial moraines found elsewhere. Hence, he argues that the limit of glaciation in High Asia has previously been placed too high. Recent research in the Himalaya and Trans-Himalaya has tended to contract, rather than extend, the limits of glaciation. This is true of Kashmir and in the Indus Valley (Derbyshire et al., 1984). The fact that there is so much evidence to the contrary must cast doubt over Kuhle's ideas. Furthermore, some of the evidence for the former position of ice margins is controversial. For example, in his 1987 paper, Kuhle argued that 'ice marginal ramps' are unambiguous indicators of glaciation. However, these features look very similar to alluvial fans. Therefore, until more evidence is collected to suggest otherwise, the idea of a large ice cap over the Tibetan Plateau with ELA depressions well in excess of 1000 m must remain in doubt.

One of the major shortcomings of the preceding discussion of Himalayan palaeoglaciation is the doubt over contemporaneity of maximum ELA depressions. Although dated Himlayan glacial sequences are rare, an attempt has been made to correlate sequences for which dates are available in table 27. Sequences that have been erected purely on the basis of relative-dating criteria have been omitted.

The data in table 27, although sparse, indicate that Himlayan and Trans-Himlayan ice advances have been out of phase with fluctuations of the Northern Hemisphere ice sheets. In particular, there is no evidence for an extensive advance at 18 kaBP, which was the last maximum glaciation in many parts of the world (Denton and Hughes, 1981). If the last glacial, maximum advance in the Himalaya had been substantial, it would most likely have overrun depositional evidence of previous advances due to the effects of progressive tectonic uplift on the magnitude of glaciation. In Kashmir, where Late Quaternary uplift has been considerable, the two glacial advances in the sequence were of progressively lesser magnitude. In Swat Kohistan, which has undergone limited uplift during the middle and Late Quaternary, there is still evidence for three glaciations of progressively smaller magnitude. Similar reduction in the magnitude of glacial advances through time was found for Zanskar (Osmaston, in press) and the Hunza Valley in the Karakoram Range (Derbyshire _et_ _al_., 1984).

Two conclusions can be drawn from the data on Himalayan glaciation. Firstly, glacial advances underwent a progressive decrease in magnitude from an undetermined time, possibly in the Middle Quaternary, up until the Late Quaternary. Secondly, advances seem to have been out of phase with the major global changes in ice volume. Progressive reduction in the extent of glacial advances may have been the indirect result of tectonic uplift. Although tectonic uplift

TABLE 27. CORRELATION OF HIMALAYAN GLACIAL SEQUENCES.

Kashmir Liddar Valley	Kashmir Ningle Valley & Toshmaidan Massif	Karakoram[1] Hunza Valley	Pamir Range[2]
<10kaBP Holocene advances	<10kaBP Holocene advances	'historical' Pasu II 830aBP Pasu I <10kaBP Batura	
35-16kaBP Loess in lower West Liddar. Advance to (?) Liddarwat	>15-17kaBP Advances to Toshmaidan and Butapathri	'Late glacial' Ghulkin II <47kaBP Ghulkin I	30-44kaBP Advance
>35.5kaBP Advance to Pahalgam	? Advance to Gulmarg	>50-60kaBP Borit Jheel >139kaBP Yunz ? Shanoz	120-300kaBP Advance 1-1.5MaBP Advance

1. Data from Derbyshire et al. (1984).
2. Data from Agakhanyantz et al. (1981).

places a greater portion of a mountain range within the theoretical accumulation-area of the glacier, it also has a blocking effect on rain-bearing monsoon circulation, leading to long-term decreases in precipitation. Secondly, uplift may lead to the progressive incision of glaciers into their troughs, at rates greater than those expected in stable tectonic situations. This would, in effect, reduce the potential for accumulation. The combined negative effects of tectonic uplift on accumulation must have exceeded the positive effect, to produce a reduction in glacier extent through time. Alternatively, the reduction in the magnitude of glacial advances may be apparent and simply an artefact of glacial erosion.

In order to explain why Himalayan glaciers seem to be out of phase with changes in global ice volumes, it is necessary to examine the independent evidence of Asian palaeoclimate that was reviewed in some detail in chapter 3. During the last glacial maximum, at 18 kaBP, the summer monsoon was greatly reduced in intensity. Although the winter monsoonal winds were stronger, conditions were generally drier and colder. Since glaciers respond to temperature and precipitation, conditions at 18 kaBP must have involved a combination of temperature and precipitation that supressed the advance of glaciers.

Data on palaeotemperatures for the Himalayan Range are lacking. However, some indication of temperature depressions at the last glacial maximum can be gained from numerical simulation models of past climate. In one such model, Gates (1976) found that the 18 kaBP temperature for High Asia was severely depressed, up to 20^{o}C below present values. However, rather different results were obtained in a more recent simulation, in which Kutzbach and Guetter (1986) found 18 kaBP temperatures only 2.5 to 5^{0}C below present values. The large depression in Gates' (1976) simulation could explain advances of glaciers at 18 kaBP, even under very arid conditions. However, the more limited depression of temperature found by Kutzbach and Guetter (1986) agrees better with the geological evidence of glacier chronology.

For the period before 18 kaBP, independent palaeoclimatic evidence is much more sparse. However, several studies of deep-ocean sediments from the northern Indian Ocean extend to before 18 kaBP (eg. Van Campo et al., 1982; Prell and Van Campo, 1986; Fontugne and Duplessy, 1986).

The pollen sequence of Van Campo et al. (1986) extends back to 150 kaBP and indicates changes in South Asian precipitation connected with the summer monsoon circulation. The work by Fontugne and Duplessy (1986), on palaeoproductivity in the northern Indian Ocean, provides a direct indication of changes in the winter, north-east, monsoon. A summary of the information provided by these studies together with the behaviour of Himalayan glaciers, is shown in table 28. Essentially, these data show that the supression of the south-west monsoon during periods of maximum glaciation was accompanied by cold and arid conditions over Southern Asia, together with an intensified winter monsoon. Periods of warm, humid climate corresponded with a summer-monsoon circulation of equal of greater strength to that of today. The climatic variation seems to have been forced by orbital changes on a

TABLE 28. ISOTOPE STRATIGRAPHY AND HIMALAYAN GLACIATION.

ISOTOPE STAGE	AGE (kaBP)	TERMINATION	GLACIAL ADVANCES Kashmir	Hunza Valley	Pamir Range
1			minor Holocene advances		
	13	I			
2			Liddarwat Butapathri Toshmaidan	?Ghulkin II	
	32				
3				?Ghulkin I	Advance
	64				
4			Pahlgam	?Borit Jheel	
	75				
5a					
	84				
5b					
	93				
5c					
	102				
5d					
	112				
5e					
	128	II			
6					
	195			Yunz	Advance

21 ka timescale related to the precession of the equinoxes. Dominanace of the precessional effect over the tilt effect would be expected at lower latitudes (Berger, 1978). The data in table 28 show quite clearly that the cold and arid periods in southern Asia are associated with isotope stage 6 and 2. Stage 2 was the LGM (last glacial maximum) when global ice volumes last reached a peak. The question remains, however, as to when major ice advances occurred in the Himalayas.

Evidence from Indian Ocean cores suggests that isotope stages 3, 5e, 5c and 5a were similar climatically to stage 1, the Holocene. Therefore, major ice advances would be unlikely to have occurred during these stages. During the cold stages 5b, 5d, and especially 6, conditions were similar to those in stage 2, when extensive glaciation did not occur in the Himalayan Range. During cold stage 4, however, palaeoproductivity of the oceans was low compared to the other cold stages. Fontugne and Duplessy (1986) concluded that this may have been due to increased precipitation over the Andaman and eastern Arabian Seas. If the increased humidity during the cold stage had extended onto the Asian landmass, stage 4 may have been ideal for the development and large-scale advance of glaciers.

Unfortunately, the Himalayan glacial chronologies are insufficiently detailed to confirm this hypothesis. In both Kashmir and the Hunza Valley, dates on glacial advances suggest that they could have occurred during isotope stage 4. Moreover, in both areas, these advances were extensive. However, the dates only place minimum age constraints on the age of the respective advances. In Kashmir, the advanced state of weathering of the oldest moraines indicates considerable age. If the most extensive advance in Kashmir dates from the last glacial cycle, then it could well have occurred during stage 4. However, if it dates from the penultimate cycle, then it probably occurred during an earlier stage that was similar, climatically, to stage 4. Unfortunately, available deep-ocean sediment studies do not yet extend below stage 6, the penultimate glacial maximum.

The conclusions drawn here essentially provide a hypothesis which can be tested once further data are collected. Clearly, much more information relating to the glacial chronology, and palaeotemperatures and palaeoprecipitation of Asia is required before a coherent picture of Himalayan palaeoclimate can emerge.

CHAPTER 8. CONCLUSIONS.

8.1 The Need for Further Research.

Kashmir is clearly an important area for the study of Late Cainozoic environmental change. However, as this study has shown, problems relating to chronology and the interpretation of the evidence, remain. Despite this research and previous work in Kashmir, there is still a need for further studies in some key areas. The most important of these are as follows.

1. A more detailed study of the clay mineralogy of the lower Karewa mudstones could provide important information regarding cyclical changes in weathering regimes in Kashmir. The work on the clay mineralogy in this thesis has shown it to be a useful indicator of environmental change. A much finer sampling interval would be required in order to show cyclical changes related to orbital forcing on 21, 40 and 100 ka wavelengths. The adoption of such a sampling interval would, however, require refinements in the chronological framework of the lower Karewa formation.
2. Further research could concentrate on the problems of interpreting the pollen and diatoms of the lower Karewa formation, the potential of which as palaeoclimatic indicators in Kashmir has not been realised.
3. Information regarding the rate and timing of uplift in the Great Himalayan Range in Kashmir is lacking, and additional work on uplift rates is required to rectify this.
4. A redefinition of the upper Karewa - lower Karewa boundary is

necessary. The fact that lithology is an inadequate means of distinguishing between the two units has been shown in this study. In order to correlate unambiguously between sections, an approach not involving lithology is required. One possible method might involve the use of aminostratigraphy. This is hampered by the absence of molluscs from many horizons. However, ostracods are more ubiquitous than molluscs and new developments in amino-acid racemization studies may eventually allow the routine application of the method to ostracod valves. Technical developments may also make TL dating applicable to older sediments. Additional chronological information is required for the loess, in order to assess the response of loess - palaeosol sequences to orbital forcing.

5. Further dates on the glacial sequences are required and mapping should be extended to valleys not examined in this research if and when restrictions on access are lifted by the Indian authorities. A better understanding of the distribution of modern glaciers is hampered by the current lack of detailed, high-resolution topographic maps, high-quality remote-sensing imagery and reliable meteorological data. A better understanding of modern glaciation in Kashmir would aid the interpretation of patterns of past glaciation. This statement holds true for much of the Himalayan Range as well as Kashmir. The relative importance of temperature and precipitation in glacier advances are poorly understood. Considerable insight into this problem could be gained by further studies of lake and bog sediments in the Kashmir Basin itself and the mountain flanks.

8.2 Conclusions From This Study.

1. Studies on the clay mineralogy suggest a change in the weathering regimes in Kashmir during the Late Pliocene at about 2.5 to 2.3 MaBP. This involved a change from conditions favouring the formation and preservation of kaolinite to those favouring smectite. This suggests an increase in aridity in Kashmir related to global climatic deterioration that has been documented in many other areas. This deterioration in climate may be partly related to the uplift of the Himalayan Range and Tibetan Plateau during the late Pliocene. The changes in clay mineralogy, however, were also responses to uplift. This is suggested by comparing the clay-mineral record with independent evidence for the Late Cainozoic uplift of Kashmir. By analogy with other mid- to low-latitude mountain ranges, the Himalayan range was probably glaciated during the Late Pliocene to Early Quaternary. However, glaciation of the Pir Panjal Range probably only began in the Middle to Late Quaternary, following rapid uplift of that range about 400 kaBP. There is no direct evidence for glaciation in the Karewa beds. This view accords with most recent research in Kashmir.

2. The rapid uplift of the Pir Panjal Range about 400 kaBP led to a restriction of the lake in Kashmir Basin. This was accompanied by fluvial conglomerate deposition on the south-western margin together with fine-grained lacustrine sedimentation in the basin centre and on the north-east flank (Bhatt, 1982; Burbank, 1982). In this research, it has been shown that the lake was restricted to below 1680 m a.s.l. and that there was substantial aeolian dust input into the lake. The

lake was cool, alkaline and shallow over much of its area and during most of the period of its existence. The lake underwent rapid drainage between 120 and 80 kaBP. The best explananation of this drainage is tectonic activity that led to a breach in the Braramulla Gorge. However, tectonic activity must have been of low intensity, since the structural disruption of the upper Karewa strata was limited. Following the drainage of the lake, aeolian input into the basin continued, but in the absence of a water body, it accumulated as true aeolian, rather than waterlain, loess. Changes in climate, as reflectd in lake-level fluctuations and changes in water chemistry, are not shown in either the ostracod assemblages or the sedimentology of the lacustrine deposits. This strongly suggests that the upper Karewa lake had an outlet throughout its existence.

3. Glaciers extended down to about 2600 m a.s.l. in the Sind Valley, 2150 m a.s.l. in the Liddar Valley and 2600 m a.s.l. in the Ningle Valley. Reconnaissance in the accessible lower parts of the Rembiara and Vishav Valleys, below 2300 m a.s.l. revealed no evidence for former glaciation in these parts of the valleys. Evidence was found for only two pre-Holocene advances in Kashmir. TL dates on morainic loess suggests that there was no major ice advance in Kashmir at the last glacial maximum, around 18 kaBP. It is suggested that glacial aridity in South Asia, associated with the suppression of the summer monsoon circulation, may be the reason for this. Present patterns of glaciation suggest a south-west to north-east gradient in Kashmir. This is explained by topographic and precipitation control. Gradients for past glaciers were apparently reversed. It is suggested that differential Quaternary uplift of the Pir Panjal Range, rather than

the reversal of climatic gradients, was responsible for this. TL dates on the morainic loess suggest that the palaeolithic artefacts from Pahalgam are more than 18 ka old, if they were found _in situ_.

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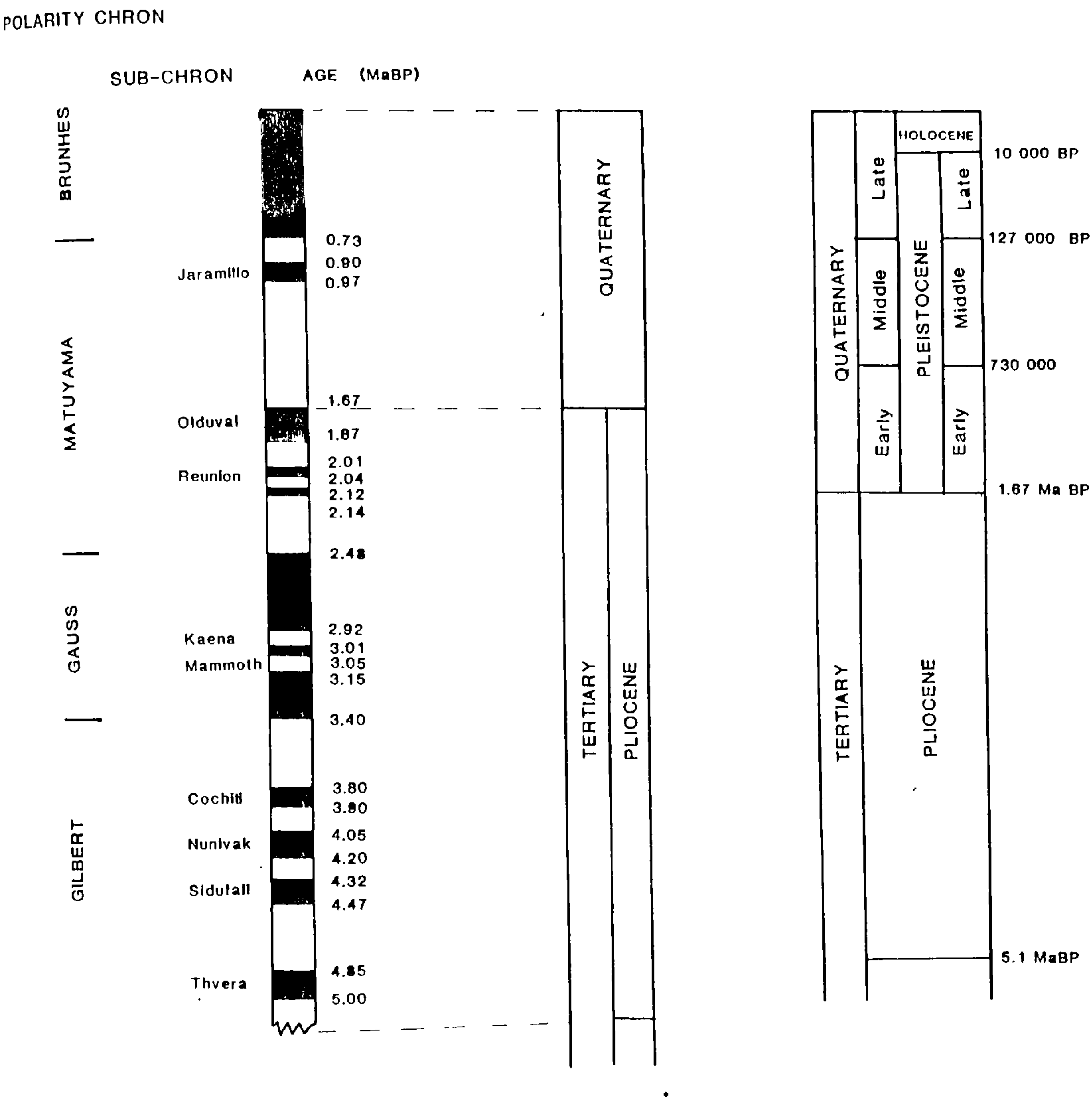
APPENDICES



Appendix 1

Chronological framework

APPENDIX 2. LIST OF DATES RELEVANT TO THIS STUDY.

A. RADIOCARBON DATES ON THE LOESS.

SITE	MATERIAL	LAB. NO.	HORIZON	AGE (a BP)
Burzahom	organic carbon	PRL 593	Palaeosol A	15700^{+370}_{-360}
	carbonate	PRL 611	Palaeosol A	20190^{+570}_{-530}
	carbonate	PRL 591	Palaeosol A	20340^{+1320}_{-1130}
	organic carbon	PRL 590	Palaeosol A	18460^{+820}_{-740}
	carbonate	PRL 590	Palaeosol A	17060^{+350}_{-340}
	carbonate	PRL 594	Palaeosol A	20430^{+920}_{-820}
	organic carbon	PRL 495	Palaeosol A	10320^{+170}_{-160}
	carbonate	PRL 500	Loess A	26450^{+890}_{-800}
	organic carbon	PRL 585	Palaeosol B	>31000
	organic carbon	PRL 586	Palaeosol C	>31000
Garhi	organic carbon	PRL 592	Palaeosol C	26340^{+2010}_{-1610}
Olchibag	carbonate	PRL 597	Palaeosol A	24960^{+1780}_{-1460}
	organic carbon	PRL 598	Palaeosol B	12560^{+450}_{-430}
	carbonate	PRL 598	Palaeosol B	21200^{+630}_{-580}
Seki Paparian	carbonate	PRL 595	Palaeosol A	33230^{+2500}_{-1900}
	organic carbon	PRL 596	Palaeosol B	>31000
	carbonate	PRL 596	Palaeosol B	>31000

A. RADIOCARBON DATES ON THE LOESS CONTINUED.

SITE	MATERIAL	LAB. NO.	HORIZON	AGE (a BP)
Puthkah	organic carbon	PRL 618	Palaeosol A	18550^{+600}_{-550}
	carbonate	PRL 618	Palaeosol A	>31000
	organic carbon	PRL 617	Palaeosol B	25800^{+1100}_{-960}
	carbonate	PRL 617	Palaeosol B	28560^{+1560}_{-1300}
Pakharpura	organic carbon	PRL 627	Palaeosol A	27630^{+1350}_{-1160}
Tsrar Sharif	organic carbon	PRL 624	Palaeosol A	>35000
	organic carbon	PRL 625	Palaeosol B	>35000
	organic carbon	PRL 626	Palaeosol C	>35000
Dilpur	organic carbon	PRL- 825	Palaeosol 1	14490±310
	carbonate	PRL 824	Palaeosol 1	10340±220
	organic carbon	PRL 830	Palaeosol 1	17740^{+630}_{-580}
	carbonate	PRL 826	Palaeosol 2	21840^{+1150}_{-1000}
	organic carbon	PRL 760	Palaeosol 2	>31000
Karpora	organic carbon	PRL 848	Top soil	5930±170
	organic carbon	PRL 849	Palaeosol	25190^{+1740}_{-1430}
Tilsur	organic carbon	PRL 850	Palaeosol	20740^{+1050}_{-980}
Malapur	organic carbon	PRL 851	Top soil	6500±190

Source: Kusumgar and Agrawal (1985).

B. TL DATES ON THE LOESS AND UPPER KAREWA LACUSTRINE BEDS.

SITE	MATERIAL	DEPTH BELOW SURFACE (m)	AGE (aBP)	METHOD	SOURCE
Karpora	Loess	4.5	31773±3724	Regen	1
	Loess	5.0	53975±6444	Regen	1
	Loess	5.2	43371±5898	Regen	1
	Loess	11.2	82883±7756	Regen	1
	Loess	20.2	95124±8573	Regen	1
	Loess	20.8	94108±8589	Regen	1
	Sandy silt	21.6	144436±15398	Regen	1
Wogahoma	Loess	3.8	46743±5593	Regen	1
	Loess	12.4	104769±9820	Regen	1
Dilpur	Loess	3.7	48950±5773	Regen	1
	Loess	8.6	85407±8139	Regen	1
	Loess	12.2	100111±9174	Regen	1
	Loess	18.2	124707±12829	Regen	1
Pehru	Loess	5.5	87985±8029	Regen	1
	Loess	8.7	97532±8937	Regen	1
	Loess	14.8	111112±10450	Regen	1
	Loess	19.2	112430±10938	Regen	1
	Lacustrine silt	20.9	115667±10938	Regen	1
Humhama	Loess	4.2	49515±5455	Regen	1
	Lacustrine silt	6.1	110887±10354	Regen	1
Burzahom	Loess	4.4	52767±5060	Regen	1
	Lacustrine silt	7.0	79532±7493	Regen	1

B. TL DATES ON THE LOESS AND UPPER KAREWA LACUSTRINE BEDS CONTINUED.

SITE	MATERIAL	DEPTH BELOW SURFACE (m)	AGE (aBP)	METHOD	SOURCE
Pattan	Lacustrine silt	8.5	82400±8000	Regen	2
	Lacustrine silt	8.5	116000±12000	N+beta-I_o	2
Pahalgam	Morainic Loess	0.9	18570±1800	Regen	2
	Morainic Loess	0.9	16400±2000	N+beta-I_o	2
Aru	Morainic Loess	1.5	32200±3000	Regen	2
	Morainic Loess	1.5	35500±2500	N+beta-I_o	2
Karpora	Loess	1.5	16500	Regen	3
	Loess	4.6	51000	Regen	3
	Loess	5.7	52000	Regen	3
	Loess	6.3	62000	Regen	3
	Loess	8.0	82000	Regen	3
	Loess	11.1	130000	Regen	3
Dilpur	Loess	3.1	58000	Regen	3
	Loess	6.1	110000	Regen	3
Burzahom	Loess	0.3	49000	Regen	3
	Loess	1.3	65000 .	Regen	3
	Loess	2.8	101000	Regen	3

SOURCES
1. Dr H. Rendell, pers. comm.
2. This study. Dates undertaken by Dr H. Rendell
3. Bronger _et al_. (1987). (Error regarded as ±15%).

C. RADIOCARBON DATES ON GLACIAL SEQUENCES.

SITE	MATERIAL	DEPTH (cm)	LAB NO.	AGE (aBP)	REF.
Toshmaidan	Peat	15 - 35	PRL 2(B)	2790±160	1
	Peat	50 - 70	PRL 3	9650±254	1
	Peat	75 - 90	PRL 4(B)	10005^{+340}_{-380}	1
	Peat	125 -140	PRL 5	11360^{+585}_{-600}	1
	Fine organic mud	205 -220	PRL 7	13980^{+520}_{-565}	1
	Clay mud	280 -295	PRL 9	15520^{+760}_{-820}	1
	Clay mud	317 -327	PRL 10	14760^{+1015}_{-925}	1
	Lacustrine clay	337 -350	PRL 12	13850^{+900}_{-785}	1
Butapathri bog 1	Peat	15 - 65	PRL 845	1120±130	2
	Peat	65 - 90	PRL 846	1020±140	2
	Peat	90 -115	PRL 847	2200±250	2
	Peat	165 -215	PRL 844	9190±210	2
	Peat	245 -284	PRL 843	17110^{+580}_{-540}	2
Butapathri Bog 2	Peat	140 -190	PRL 842	10950^{+340}_{-320}	2
Pahalgam	Palaeosol	120	Q 3084	11530±155	3

1. Singh and Agrawal (1976).
2. Dodia et al. (1984.
3. This study.

APPENDIX 3. LOCATION OF SITES AND SECTIONS IN THIS STUDY.

SITE	AREA (IF ANY)	LATITUDE (N)			LONGITUDE (E)			ALTITUDE
		D	M	S	D	M	S	(m)
Burzahom		34	10	14	74	52	50	1615
Pattan		34	08	32	74	32	28	1676
Pampur		34	00	04	74	55	05	1615
Sambur 1		33	59	08	74	55	23	1615
Sambur Village		33	57	29	74	55	45	1615
Wogahoma		33	49	46	75	06	38	1676
Pehru		33	44	13	75	11	00	1676
Karpora	Kanchi Kol	33	47	13	74	50	07	1980
Singhpora Rd.		34	14	08	74	26	23	1645
Dilpur		33	55	16	74	47	09	1737
Humhama		34	01	06	74	45	58	1615
Hardvail		33	59	25	74	40	13	1829
Watrehail		33	59	00	74	39	28	1890
Badgam		31	01	30	74	41	30	1615
Zadora		33	54	17	74	51	45	1630
Romu	Romushi	33	49	52	74	47	50	1890
Pakharpura	Romushi	33	48	04	74	46	17	1890
Khaigam	Romushi	33	48	46	74	00	25	1890
Aglar	Romushi	33	48	52	74	47	25	1859
Baltal	Romushi	33	50	00	74	48	58	1828
Birnai Nala	Romushi	33	48	41	74	47	30	1859
Qasba Nagum		33	55	37	74	47	35	1668
Anantnag Town		33	43	52	75	09	25	1585
Ganderbal		34	13	10	74	46	48	1600
Balpora	Rembiara	33	00	10	74	50	20	1920
Shupian	Rembiara	34	42	14	74	49	09	2072
Hirpur	Rembiara	33	40	27	74	44	05	2299
Nunar	Sind	34	00	25	74	47	33	1646
Mangom	Sind	34	16	58	74	48	10	1676
Kangan	Sind	34	15	46	74	54	05	1768
Pharao	Sind	34	14	27	75	03	48	2011
Rezan	Sind	34	00	27	75	00	25	2286
Shitkari	Sind	34	18	33	75	00	55	2682
Thjiwas	Sind	34	16	56	75	16	45	2860
Sonamarg	Sind	34	18	14	75	17	50	2667
Woyil	Sind	34	46	21	74	48	25	1706
Gund	Sind	34	16	35	75	00	53	2134
Saribal	Sind	34	16	12	75	31	25	2773
Neh Nar	Sind	34	09	52	75	30	25	3719
Kahalwan	Liddar	33	02	48	75	09	55	2073

APPENDIX 3. CONTINUED.

SITE	AREA (IF ANY)	LATITUDE (N) D	M	S	LONGITUDE (E) D	M	S	ALTITUDE (m)
Ganeshpur	Liddar	33	53	37	75	17	00	1828
Lidru	Liddar	33	58	10	75	19	05	2150
Nunawan	Liddar	34	00	00	75	19	15	2073
Pahalgam	Liddar	34	16	02	75	18	48	2134
Aru	Liddar	34	05	10	75	15	48	2414
Lidderwat	Liddar	34	09	27	75	13	23	2729
Basmai	Liddar	34	12	04	75	16	25	3133
Kolahoi	Liddar	34	11	56	75	20	00	3536
Gulmarg Club	Ningle Nala	34	03	25	74	23	20	2639
Killenmarg	Ningle Nala	34	02	48	74	21	23	3094
Butapathri	Ningle Nala	34	04	43	74	18	30	2850
Durhom	Ningle Nala	34	10	23	74	25	55	1661
Dandamuh	Ningle Nala	34	07	02	74	23	45	1859
Alpather	Ningle Nala	34	01	55	74	19	42	3978

APPENDIX 4. CLAY MINERAL DATA FROM LOWER KAREWA MUDSTONES

Sample	Age (Ma)	% S	I	K+C	K	C	I_r	I_c	$S_{v/p}$	$S_{l/h}$
H16a	3.664	35	30	35	22	13	1.10	0.37	0.93	1.23
H17	3.624	31	23	46	23	23	1.31	0.83	0.95	1.03
H18	3.560	26	26	48	23	25	1.31	0.54	0.96	1.21
H19	3.504	1	22	77	23	54	1.07	0.82	0.96	0.86
H20	3.456	28	22	50	24	26	1.22	0.59	0.91	0.90
H21	3.404	41	23	36	20	16	1.40	0.46	0.98	1.09
H22	3.384	4	24	34	15	19	0.99	0.53	0.81	1.09
H23	3.364	36	26	38	27	11	1.00	0.47	0.96	1.21
H24	3.320	11	28	61	33	28	1.32	0.60	0.93	0.85
H25	3.284	34	21	45	36	9	1.11	0.32	0.97	1.06
H26	3.264	20	27	53	20	33	1.01	0.44	0.92	0.99
H27	3.244	33	29	38	19	19	1.39	0.61	0.96	1.19
H28	3.212	30	16	54	24	30	1.31	0.53	0.94	1.01
H29	3.164	24	21	55	33	22	1.18	0.59	0.93	1.03
H30	3.152	37	24	39	30	9	1.17	0.54	0.97	1.25
H31	3.068	22	14	64	27	37	2.35	0.06	0.93	1.00
H32	3.016	0	17	83	50	33	0.70	0.47	0.69	0.49
H33	2.920	51	19	30	17	13	1.26	0.63	0.98	1.15
H 1	2.792	26	23	51	20	31	1.13	0.73	0.93	0.93
H 2	2.752	13	26	61	38	23	1.19	0.62	0.92	0.92
H 3	2.704	25	16	59	32	27	1.52	0.61	0.94	0.95
H 4	2.664	32	26	42	27	15	1.28	0.55	0.95	1.09
H 5	2.624	31	26	43	18	25	0.74	0.33	0.96	1.07
H 6	2.608	27	18	55	29	26	1.04	0.53	0.95	1.02
H 7	2.584	31	26	43	22	21	1.12	0.48	0.94	0.98
H 8	2.548	32	28	40	21	19	0.78	0.45	0.95	1.03
H 9	2.512	23	16	61	25	36	1.12	0.60	0.92	0.85
H10	2.468	36	20	44	17	27	1.16	0.53	0.96	0.99
H11	2.430	34	19	47	18	28	1.27	0.42	0.96	1.03
H12	2.399	27	17	56	12	44	1.22	0.38	0.95	0.97
H13	2.390	35	23	42	20	22	1.23	0.45	0.96	0.99
H14	2.374	29	16	55	29	26	1.20	0.42	0.85	0.89
H15	2.349	43	19	38	19	19	1.00	0.59	0.97	1.08
H16b	2.318	40	21	39	15	24	0.96	0.44	0.96	1.08
R 5	2.280	51	23	26	6	20	1.37	1.36	0.97	1.19
R 6	2.271	26	47	27	12	15	1.32	0.44	0.95	1.12
R 7	2.249	36	32	32	17	15	1.41	0.38	0.97	1.06
R 8	2.230	37	22	41	14	27	1.31	0.66	0.97	1.14
R 9	2.199	27	19	54	32	22	1.38	0.50	0.93	1.32
R10	2.168	48	27	25	25	0	1.73	0.88	0.96	1.17
R11	2.143	54	20	26	1	25	2.11	2.15	0.96	1.13
R12	2.121	24	18	58	50	8	0.74	2.17	0.81	0.68
R13	2.096	44	18	38	3	35	1.88	1.25	0.95	1.01
R14	2.068	23	28	49	28	21	1.42	0.59	0.96	1.10
R15	2.043	50	26	24	0	24	1.52	1.98	0.96	1.16
R16	2.030	23	17	60	24	36	1.29	0.91	0.95	1.12
R17	1.996	53	19	28	18	10	1.94	1.76	0.94	1.06
R18	1.961	48	18	34	10	24	1.77	1.14	0.95	1.08

APPENDIX 4. CONTINUED.

Sample	Age (Ma)	S	I	% K+C	K	C	I_r	I_c	$S_{v/p}$	$S_{1/h}$
R19	1.918	32	21	47	22	25	1.36	0.84	0.94	1.05
R20a	1.880	32	19	49	4	45	1.28	0.83	0.97	1.11
R20b	1.830	17	21	62	18	44	1.15	0.39	0.92	0.93
R21	1.793	41	23	36	14	22	1.34	0.73	0.95	1.00
R22	1.752	46	18	36	21	15	1.37	0.54	0.96	1.05
R23	1.740	47	23	30	18	12	1.25	0.78	0.96	1.15
R24	1.514	56	27	17	17	0	1.58	0.79	0.98	1.25
R25	1.420	71	21	8	8	0	1.55	1.20	0.90	1.05
R26	1.333	47	27	26	12	14	1.21	0.45	0.97	1.17
R27	1.251	67	22	11	11	0	1.57	0.82	0.98	1.02
R28	1.189	47	22	31	31	0	1.48	1.15	0.97	1.15
R29	1.108	43	20	37	20	17	1.34	0.77	0.96	1.06
R30	1.026	39	42	19	19	0	1.44	1.36	0.97	1.20
R31	0.976	25	20	55	39	16	0.80	0.41	0.94	1.08
R32	0.789	42	20	38	19	19	1.25	0.44	0.95	1.11
R33	0.720	43	15	42	16	26	1.41	0.46	0.95	0.97
R34	0.690	53	36	11	11	0	2.19	1.29	0.91	0.79
R35	0.520	42	30	28	18	10	1.30	0.76	0.95	1.22
R36	0.500	33	18	49	15	34	1.23	0.73	0.95	1.08
R37	0.480	60	22	18	12	6	1.42	0.93	0.96	1.11
R38	0.450	32	22	46	13	33	1.04	0.70	0.89	1.00
R39	0.420	61	32	7	7	0	1.39	1.08	0.93	1.12

S=smectite, I=illite, K=kaolinite, C=chlorite; I_r, Ic, $S_{v/p}$ and $S_{1/h}$ are indices ofthe composition and crystallinity of illite (I) and smectite (S) as described in the text.

APPENDIX 5. SUMMARY SEDIMENTOLOGICAL DATA FOR UPPER KAREWA SAMPLES.

Sample No.	Section	Bed No.	Moist Colour	$\%CaCO_3$
24	Pampur	14	5Y 4/3	1.7
26	Wogahoma	5	5Y 5/2	15.3
29	Burzahom	9b	2.5Y 5/2	0.0
30	Pampur	9	5Y 5/2	2.9
31	Burzahom	3a	5Y 5/2	43.5
37	Wogahoma	5	5Y 5/2	7.2
39	Wogahoma	3	5Y 8/4	11.7
42	Pampur	14	5Y 4/3	2.4
45	Pampur	5	2.5Y 5/6	5.8
49	Burzahom	3a	5Y 5/2	49.3
51	Pampur	9	5Y 5/2	3.1
66	Sambur	1	5Y 5/4	4.3
67	Sambur	2	5Y 5/4	7.6
68	Sambur	3	5Y 5/4	6.9
69	Sambur	5	5Y 5/4	8.9
70	Sambur	12	5Y 5/4	4.6
71	Sambur	14	5Y 5/4	5.7
72	Sambur	16	5Y 5/4	3.9
73	Sambur	17	2.5Y 7/4	6.3
74	Pattan	2	5Y 4/4	0.4
75	Pattan	11	5Y 4/4	0.0
76	Pattan	11	5Y 4/4	0.0
77	Pattan	11	5Y 4/4	0.0
78	Pattan	14	5Y 4/4	0.2
79	Pehru	3	5Y 5/2	47.0
80	Pehru	5	5Y 5/2	18.9
81	Pehru	5	5Y 5/2	11.7
82	Pehru	6	10YR 5/8	10.7
83	Burzahom	12b	2.5Y 5/4	0.0
84	Burzahom	10b	2.5Y 5/4	0.0
85	Pampur	11	5Y 5/3	2.4
86	Pampur	11	5Y 5/3	5.3
87	Pattan	18	2.5Y 6/4	10.0
88	Pattan	18	2.5Y 6/6	9.8
89	Burzahom	1a	5Y 6/3	2.6
90	Burzahom	2a	5Y 6/4	29.6
91	Burzahom	10b	2.5Y 5/4	0.0
92	Burzahom	5b	2.5Y 6/3	0.0
93	Wogahoma	1	2.5Y 4/4	8.7
94	Wogahoma	3	2.5Y 4/4	15.9
95	Wogahoma	3	10YR 3/3	1.1
96	Wogahoma	4	2.5Y 4/4	8.6
97	Pampur	4	2.5Y 5/6	11.2
99	Pampur	7	10YR 5/1	8.6
101	Pampur	11	5Y 5/3	5.6
103	Pampur	15	5Y 6/3	13.8
107	Sambur	4	5YR 5/8	2.3
109	Sambur	9	5Y 5/4	6.5
110	Sambur	9	5Y 5/4	7.0

APPENDIX 5. CONTINUED.

Sample No.	Section	Bed No.	Moist Colour	$\%CaCO_3$
111	Sambur	13	5Y 5/4	5.3
112	Sambur	15	5Y 5/4	11.1
113	Sambur	17	2.5Y 7/4	5.1
117	Burzahom	4b	2.5Y 6/4	0.7
118	Burzahom	8b	2.5Y 7/4	1.5
119	Burzahom	8b	2.5Y 7/4	0.0
200	Pattan	3	5Y 5/4	5.4
201	Pattan	4	5Y 5/4	2.5
202	Pattan	5	2.5Y 5/4	2.5
203	Pattan	7	5Y 5/2	2.5
204	Pattan	7	5Y 5/2	5.0
205	Pattan	7	5Y 5/2	4.3
206	Pattan	15	5Y 5/3	2.6
207	Pattan	15	5Y 5/3	2.8
208	Pattan	15	5Y 5/3	10.4
209	Pattan	16	2.5Y 6/4	7.6
210	Pattan	16	2.5Y 6/4	7.8
211	Pattan	16	2.5Y 6/4	7.4
212	Pattan	16	2.5Y 6/4	12.7
213	Pattan	18	2.5Y 6/4	3.9
214	Pattan	18	2.5Y 7/6	0.0

APPENDIX 6. PARTICLE-SIZE DATA.

SAMPLE NUMBER	MEAN	MEDIAN	SORTING	CLASS	SKEWNESS	CLASS	KURTOSIS	CLASS	PERCENT SAND	PERCENT SILT	PERCENT CLAY	INFERRED ORIGIN
3	3.01	1.80	3.71	VPS	0.53	SFS	0.94	M	61.9	29.6	8.5	3
4	7.62	7.43	1.71	PS	0.17	FS	0.77	P	0.0	77.1	22.9	2
5	3.67	0.80	4.86	EPS	0.78	SFS	0.62	VP	60.9	15.4	23.7	2
6	2.21	0.93	3.25	VPS	0.67	SFS	1.23	L	74.4	18.3	7.3	3
7	1.66	0.60	2.75	VPS	0.63	SFS	1.71	VL	82.1	15.3	2.6	7
8	3.43	2.71	3.78	VPS	0.34	SFS	0.76	P	52.2	39.3	8.5	7
9	1.81	0.79	3.26	VPS	0.65	SFS	2.38	VL	81.0	8.9	10.1	4
10	6.55	6.79	4.74	EPS	-0.06	NS	1.00	M	21.8	40.7	37.5	7
11	6.11	6.58	3.78	VPS	-0.10	NS	1.39	L	22.2	59.2	18.6	7
12	4.16	1.78	4.79	EPS	0.65	SFS	0.64	P	56.0	19.0	25.0	2
13	3.71	1.26	4.45	EPS	0.74	SFS	1.04	VP	73.0	8.1	18.9	4
14	1.32	0.16	2.84	VPS	0.73	SFS	1.84	VL	79.9	14.9	5.2	3
15	3.54	2.46	3.61	VPS	0.46	SFS	0.82	P	54.0	36.5	9.5	3
17	6.12	5.89	1.92	PS	0.39	SFS	1.53	VL	4.7	83.6	11.7	2
19	3.25	3.15	3.11	VPS	0.11	FS	0.75	VL	56.8	39.3	3.9	2
20	3.56	3.34	3.51	VPS	0.20	FS	0.83	P	50.8	41.2	8.0	3
23	3.14	2.26	2.23	VPS	0.57	SFS	0.83	P	66.0	30.8	3.2	8
24	6.22	5.70	2.29	VPS	0.41	SFS	1.33	L	8.5	80.0	11.5	5
25	2.09	0.48	3.64	VPS	0.74	SFS	1.06	M	74.0	21.6	4.4	7
26	1.75	1.02	2.01	VPS	0.66	SFS	2.97	VL	85.0	12.7	2.3	6
27	3.20	1.30	4.09	EPS	0.66	SFS	1.04	M	62.2	22.3	15.5	2
29	7.13	5.78	2.90	VPS	0.65	SFS	0.81	P	0.4	72.4	27.2	5
30	5.58	4.91	1.86	PS	0.72	SFS	1.90	VL	0.1	90.3	9.6	5
31	3.08	2.31	1.76	PS	0.78	SFS	4.64	EL	81.0	15.8	3.2	6
37	1.39	1.37	1.09	PS	0.35	SFS	3.21	EL	96.0	2.4	1.6	6
39	7.04	5.88	2.59	VPS	0.65	SFS	1.92	VL	0.1	79.8	20.1	5
42	5.67	4.58	2.48	VPS	0.69	SFS	1.35	L	13.9	72.7	13.4	5

APPENDIX 6. CONTINUED.

SAMPLE NUMBER	MEAN	MEDIAN	SORTING	CLASS	SKEWNESS	CLASS	KURTOSIS	CLASS	PERCENT SAND	PERCENT SILT	PERCENT CLAY	INFERRED ORIGIN
45	3.39	2.54	2.07	VPS	0.75	SFS	1.36	L	72.0	22.5	5.5	6
47	2.27	0.45	3.34	VPS	0.72	SFS	0.58	VP	65.4	31.7	2.9	3
49	7.09	5.84	2.84	VPS	0.61	SFS	1.39	L	0.1	82.6	17.3	5
51	6.84	6.61	1.59	PS	0.39	SFS	1.47	L	0.0	90.4	9.6	5
52	5.80	5.57	1.38	PS	0.55	SFS	2.06	VL	0.3	92.3	7.4	8
53	3.40	3.90	3.41	VPS	0.00	NS	1.24	L	50.4	40.0	9.6	7
54	3.78	3.93	3.16	VPS	0.08	NS	1.49	L	51.7	40.7	7.6	7
55	4.65	3.90	4.39	EPS	0.28	FS	0.75	P	51.0	30.0	19.0	7
56	5.72	5.14	4.38	EPS	0.19	FS	0.81	P	43.9	31.4	24.7	7
57	5.35	5.27	4.58	EPS	0.08	NS	0.72	P	42.2	35.0	22.8	7
58	3.71	1.70	4.61	EPS	0.61	SFS	0.74	P	61.0	20.1	18.9	2
59	2.72	1.30	3.40	VPS	0.70	SFS	1.10	M	70.9	18.4	10.7	8
60	2.72	2.80	0.64	MWS	-0.24	CS	1.56	VL	97.6	2.4	0.0	8
61	3.48	3.16	3.90	VPS	0.27	FS	0.96	M	56.8	27.3	15.9	1
62	1.46	0.85	2.37	VPS	0.45	SFS	1.02	M	83.0	15.4	1.6	1
63	-0.02	-0.12	0.83	MS	0.24	FS	1.19	L	97.7	2.3	0.0	1
64	3.27	3.05	2.06	VPS	0.31	SFS	1.69	VL	73.2	22.1	4.7	8
65	4.85	4.57	2.06	VPS	0.32	SFS	1.14	L	44.2	50.0	5.8	8
66	3.34	3.21	0.77	MS	0.50	SFS	1.62	VL	84.4	12.8	2.8	6
67	6.53	5.98	2.53	VPS	0.42	SFS	1.17	L	9.1	74.6	16.3	6
68	3.72	3.71	0.76	MS	0.30	FS	2.48	VL	68.0	29.6	2.4	6
69	4.29	3.85	1.42	PS	0.71	SFS	2.49	VL	61.7	33.1	5.2	6
70	2.22	2.28	0.66	MS	-0.12	CS	1.29	L	97.2	2.8	0.0	6
71	1.83	1.79	0.83	MS	0.19	FS	1.94	VL	98.4	1.6	0.0	6
72	2.76	2.70	0.38	WS	0.35	SFS	1.49	L	96.5	3.5	0.0	6
73	2.62	2.61	0.51	MWS	0.10	NS	1.38	L	97.0	3.0	0.0	6
74	1.61	1.70	1.08	PS	-0.14	CS	0.96	M	99.4	0.6	0.0	6
75	1.99	2.20	1.13	PS	-0.22	CS	0.94	M	95.5	4.5	0.0	6
76	1.47	1.60	1.12	PS	-0.17	CS	0.99	M	97.3	2.7	0.0	6
77	0.68	0.60	1.06	PS	0.13	FS	1.02	M	98.0	2.0	0.0	6
78	1.52	0.38	2.26	VPS	0.47	SFS	1.94	VL	86.0	9.2	4.8	6

APPENDIX 6. CONTINUED.

SAMPLE NUMBER	MEAN	MEDIAN	SORTING	CLASS	SKEWNESS	CLASS	KURTOSIS	CLASS	PERCENT SAND	PERCENT SILT	PERCENT CLAY	INFERRED ORIGIN
79	1.69	1.37	1.50	PS	0.37	SFS	1.17	L	94.2	4.2	1.6	6
80	0.27	0.18	1.00	MS	0.20	FS	0.93	M	97.8	2.2	0.0	6
81	1.41	1.41	1.06	PS	0.04	NS	1.07	M	95.9	4.1	0.0	6
82	1.99	1.95	0.62	MS	0.25	FS	1.41	L	95.8	4.2	0.0	6
83	8.81	8.07	2.48	VPS	0.40	SFS	0.85	P	2.7	60.3	37.0	5
84	7.13	6.10	2.70	VPS	0.57	SFS	1.13	L	0.7	79.5	19.8	5
85	8.84	8.08	2.69	VPS	0.31	SFS	1.01	M	0.3	62.1	37.6	5
86	8.56	7.55	2.77	VPS	0.50	SFS	0.71	P	0.2	58.6	41.2	5
87	6.79	6.48	1.48	PS	0.26	FS	1.44	L	3.8	93.4	2.8	5
88	7.27	5.95	3.10	VPS	0.53	SFS	0.98	M	5.2	68.1	26.7	5
89	5.86	5.46	2.13	VPS	0.45	SFS	1.96	VL	9.4	78.0	12.6	5
90	8.56	7.88	2.82	VPS	0.31	SFS	0.78	P	1.8	61.4	36.8	5
91	8.13	7.03	2.54	VPS	0.60	SFS	0.84	P	0.3	68.3	31.4	5
92	8.17	7.14	2.58	VPS	0.56	SFS	0.88	P	0.0	68.8	31.2	5
94	7.25	5.90	2.90	VPS	0.63	SFS	0.89	P	4.6	69.0	26.4	5
95	8.45	8.70	0.68	MWS	-0.59	SCS	3.83	EL	0.0	93.5	6.5	5
96	8.90	8.31	2.90	VPS	0.22	FS	0.79	P	0.1	55.2	44.7	5
97	7.44	6.05	3.14	VPS	0.58	SFS	0.71	P	4.2	63.2	32.6	5
99	8.30	7.48	2.88	VPS	0.36	SFS	0.80	P	2.6	62.3	35.1	5
101	7.37	7.57	2.66	VPS	0.05	NS	1.07	M	0.4	79.7	19.9	5
103	7.49	6.40	3.25	VPS	0.40	SFS	0.90	M	11.3	58.7	30.0	5
112	8.00	7.50	2.36	VPS	0.29	FS	1.06	M	0.0	70.0	30.0	5
107	9.67	9.10	2.33	VPS	0.31	SFS	0.77	P	0.0	48.8	51.2	5
109	7.52	6.76	2.62	VPS	0.45	SFS	1.09	M	3.2	73.7	23.1	5
110	8.45	8.10	3.14	VPS	0.14	FS	0.79	P	5.2	54.6	40.2	5
111	8.55	8.13	2.88	VPS	0.18	FS	0.84	P	1.6	60.9	37.5	5
113	7.00	5.72	2.96	VPS	0.61	SFS	0.89	P	5.4	68.9	25.7	5
116	7.83	7.06	2.76	VPS	0.40	SFS	0.98	M	1.3	71.3	27.4	5
117	6.00	5.30	2.44	VPS	0.50	SFS	2.29	VL	10.4	77.4	12.2	5
118	7.16	6.47	2.74	VPS	0.39	SFS	1.63	VL	10.1	72.9	17.0	5
119	6.19	5.30	2.40	VPS	0.62	SFS	2.16	VL	7.7	79.7	12.6	5

APPENDIX 6. CONTINUED.

SAMPLE NUMBER	MEAN	MEDIAN	SORTING	CLASS	SKEWNESS	CLASS	KURTOSIS	CLASS	PERCENT SAND	PERCENT SILT	PERCENT CLAY	INFERRED ORIGIN
120	5.10	4.35	4.66	EPS	0.25	FS	0.76	P	47.9	27.9	24.2	2
121	0.82	0.67	2.57	VPS	0.47	SFS	3.29	EL	88.0	1.9	10.1	4
122	5.78	6.61	4.94	EPS	-0.17	CS	0.73	P	42.1	22.7	35.2	7
123	4.23	2.85	4.51	EPS	0.45	SFS	0.82	P	54.0	26.8	19.2	7
124	1.61	0.03	3.32	VPS	0.79	SFS	2.20	VL	81.4	11.2	7.4	1
125	5.68	5.36	4.74	EPS	0.12	FS	0.67	P	43.0	28.9	28.1	2
126	6.03	5.88	4.74	EPS	0.06	NS	0.69	P	46.9	22.5	30.6	7
127	0.43	0.05	2.37	VPS	0.65	SFS	3.24	EL	87.4	4.9	7.7	2
128	4.57	2.15	5.13	EPS	0.60	SFS	0.54	P	53.9	8.8	37.3	3
129	1.08	0.18	2.66	VPS	0.72	SFS	2.14	VL	84.4	9.8	5.8	1
130	3.68	0.83	4.96	EPS	0.76	SFS	0.61	P	61.8	13.2	25.0	3
131	4.10	1.65	4.80	EPS	0.67	SFS	0.64	VP	53.7	21.2	25.1	2
132	2.64	0.96	3.82	VPS	0.69	SFS	0.90	M	68.9	21.0	10.1	1
133	5.17	4.89	4.57	EPS	0.14	FS	0.76	P	46.2	31.8	22.0	3
134	4.96	4.26	4.46	EPS	0.26	FS	0.73	P	48.8	30.3	20.9	7
135	4.59	4.05	4.14	EPS	0.25	FS	0.85	P	49.2	34.2	16.6	2
136	2.59	1.14	3.72	VPS	0.66	SFS	1.09	M	72.3	15.9	11.8	2
137	4.31	2.85	4.58	EPS	0.46	SFS	0.67	P	52.2	26.6	21.2	7
138	2.23	1.22	3.21	VPS	0.61	SFS	1.67	VL	76.2	15.1	8.7	3
139	8.75	8.46	3.12	VPS	0.09	NS	0.76	P	5.7	48.3	46.0	7
140	6.44	6.55	4.81	EPS	-0.01	NS	0.63	VP	34.9	27.2	37.9	2
141	5.91	6.55	4.91	EPS	-0.10	CS	0.61	VP	48.2	22.3	29.5	2
142	6.03	6.49	4.83	EPS	-0.07	NS	0.66	VP	33.6	37.3	29.1	2
143	3.24	0.97	4.28	EPS	0.75	SFS	0.74	P	58.1	29.9	12.0	3
144	4.99	4.75	4.19	EPS	0.14	FS	0.86	P	44.6	39.6	15.8	3
145	0.97	0.65	2.13	VPS	0.50	SFS	1.80	VL	88.0	8.1	3.9	1
146	5.73	5.85	4.82	EPS	0.01	NS	0.65	VP	40.0	32.0	28.0	2
147	3.18	1.79	3.81	VPS	0.58	SFS	0.88	P	68.1	20.6	11.3	2
148	5.53	5.60	4.33	EPS	0.03	NS	0.91	M	35.8	43.3	20.9	7
149	5.75	6.10	4.82	EPS	-0.04	NS	0.68	P	33.6	39.4	27.0	2
150	4.18	2.73	4.58	EPS	0.46	SFS	0.78	P	50.4	30.5	19.1	2
151	2.30	1.18	3.21	VPS	0.60	SFS	1.29	L	77.9	15.8	6.3	1

APPENDIX 6. CONTINUED.

SAMPLE NUMBER	MEAN	MEDIAN	SORTING	CLASS	SKEWNESS	CLASS	KURTOSIS	CLASS	PERCENT SAND	PERCENT SILT	PERCENT CLAY	INFERRED ORIGIN
152	2.97	0.95	4.12	EPS	0.72	SFS	0.80	P	68.8	18.3	12.9	3
153	3.56	1.45	4.56	EPS	0.65	SFS	0.72	P	58.9	22.7	18.4	3
154	5.38	5.56	4.70	EPS	0.02	NS	0.69	P	41.3	35.3	23.4	7
155	5.93	6.49	4.94	EPS	-0.09	NS	0.62	VP	40.1	30.9	29.0	7
156	3.17	1.32	4.10	EPS	0.67	SFS	0.91	M	70.0	15.4	14.6	3
157	6.08	6.24	4.50	EPS	-0.01	NS	0.83	P	31.0	43.2	25.8	7
158	3.64	0.58	4.92	EPS	0.81	SFS	0.62	VP	68.0	6.9	25.1	3
159	5.84	6.22	4.52	EPS	-0.05	NS	0.83	P	32.4	43.6	24.0	7
160	4.93	4.65	4.32	EPS	0.15	FS	0.80	P	44.2	36.9	18.9	2
161	4.98	4.40	4.62	EPS	0.21	FS	0.72	P	45.7	32.1	22.2	7
162	6.51	6.81	4.88	EPS	-0.06	NS	0.59	VP	30.4	31.7	37.9	7
163	4.81	2.71	4.88	EPS	0.56	SFS	0.58	VP	53.2	14.6	32.2	3
164	4.70	4.02	4.58	EPS	0.26	FS	0.76	P	50.0	28.6	21.4	3
165	6.31	6.57	4.82	EPS	-0.04	NS	0.62	VP	32.9	31.4	35.7	7
166	3.00	1.14	4.05	EPS	0.72	SFS	0.75	P	66.0	19.1	14.9	3
167	3.53	1.12	4.56	EPS	0.72	SFS	0.70	P	58.2	23.2	19.5	3
168	6.73	6.80	4.44	EPS	-0.02	NS	0.82	P	26.2	41.1	32.7	7
169	3.37	1.62	4.28	EPS	0.61	SFS	0.76	P	63.1	21.3	15.6	3
170	6.51	6.30	4.41	EPS	0.05	NS	0.85	P	26.9	43.0	30.1	7
171	7.04	6.55	3.81	VPS	0.17	FS	0.81	P	23.9	45.2	30.9	2
172	4.91	4.65	4.37	EPS	0.15	FS	0.80	P	44.9	36.2	18.9	2
173	6.36	6.35	4.40	EPS	0.00	NS	0.94	M	26.5	45.9	27.6	7
174	4.60	3.97	4.40	EPS	0.26	FS	0.83	P	50.0	31.3	18.7	7
175	3.23	1.10	4.26	EPS	0.72	SFS	0.83	P	69.1	15.1	15.8	3
176	5.41	5.80	4.88	EPS	-0.03	NS	0.64	VP	45.9	30.1	25.0	7
177	1.93	0.75	3.17	VPS	0.66	SFS	1.83	VL	81.0	12.2	6.8	3
178	8.35	7.93	3.14	VPS	0.17	FS	0.77	P	5.0	55.8	39.2	9
179	7.87	7.25	2.80	VPS	0.34	SFS	0.92	M	3.2	65.0	31.8	9
180	8.48	8.15	2.76	VPS	0.19	FS	0.83	P	0.9	65.2	33.9	9
181	8.87	8.66	2.93	VPS	0.12	FS	0.68	P	0.6	52.2	47.2	9
182	8.92	8.63	2.95	VPS	0.12	FS	0.71	P	1.0	51.8	47.2	9

APPENDIX 6. CONTINUED.

SAMPLE NUMBER	MEAN	MEDIAN	SORTING	CLASS	SKEWNESS	CLASS	KURTOSIS	CLASS	PERCENT SAND	PERCENT SILT	PERCENT CLAY	INFERRED ORIGIN
183	8.82	8.36	2.93	VPS	0.20	FS	0.69	P	1.8	51.9	46.3	9
184	7.98	7.25	3.30	VPS	0.27	FS	0.75	P	7.7	54.2	38.1	9
185	7.49	6.55	3.00	VPS	0.44	SFS	0.80	P	4.7	66.0	29.3	9
186	1.11	1.04	1.79	PS	0.15	FS	1.20	L	92.8	7.2	0.0	8
187	2.20	2.13	2.00	PS	0.06	NS	1.03	M	82.0	17.4	0.6	8
188	1.83	1.97	1.91	PS	0.01	NS	1.21	L	89.0	10.0	1.0	8
189	0.61	0.45	1.76	MS	0.30	FS	1.33	L	94.3	5.7	0.0	8
190	8.46	7.77	2.92	VPS	0.31	SFS	0.73	P	0.6	59.4	40.0	9
191	5.16	5.17	4.33	EPS	0.06	NS	0.87	P	49.0	31.4	19.6	9
192	7.98	7.35	3.08	VPS	0.25	FS	0.87	P	9.3	57.1	33.6	9
193	8.61	7.97	2.97	VPS	0.31	SFS	0.76	P	2.6	60.5	37.0	9
194	6.85	6.71	2.50	VPS	0.14	FS	1.68	VL	14.2	70.3	15.5	9
195	9.00	9.45	3.63	VPS	-0.20	CS	0.81	P	14.8	30.2	55.0	9
196	8.55	8.30	3.19	VPS	0.10	NS	0.69	P	3.5	50.5	46.0	9
197	8.83	8.70	2.91	VPS	0.06	NS	0.73	P	4.8	47.2	48.0	9
198	8.17	7.33	3.06	VPS	0.31	SFS	0.88	P	9.4	55.6	35.0	9
199	9.00	8.55	2.72	VPS	0.22	FS	0.71	P	0.5	54.5	45.0	9
200	7.40	6.50	3.02	VPS	0.41	SFS	1.09	M	12.7	62.3	25.0	5
201	6.49	5.50	3.42	VPS	0.39	SFS	0.96	M	34.6	44.9	20.5	5
202	7.82	7.15	2.60	VPS	0.44	SFS	1.14	L	0.6	74.4	25.0	5
203	6.98	6.06	2.38	VPS	0.62	SFS	1.32	L	3.0	78.5	18.5	5
204	7.60	6.30	2.71	VPS	0.68	SFS	0.91	M	1.4	69.6	29.0	5
205	7.71	6.82	2.99	VPS	0.38	SFS	0.93	M	7.8	63.2	29.0	5
206	6.46	5.70	3.13	VPS	0.40	SFS	0.98	M	26.6	52.9	20.5	5
207	8.86	8.07	2.89	VPS	0.26	FS	0.97	M	5.9	53.6	40.5	5
208	7.57	6.86	2.34	VPS	0.48	SFS	1.48	L	3.4	76.1	20.5	5
209	7.97	6.95	2.61	VPS	0.55	SFS	0.91	M	1.7	71.2	27.1	5
210	8.39	7.54	2.60	VPS	0.45	SFS	0.84	P	2.2	66.3	31.5	5
211	6.89	5.92	2.56	VPS	0.57	SFS	1.29	L	5.5	75.2	19.3	5
212	6.49	5.80	2.38	VPS	0.49	SFS	1.62	VL	10.1	74.4	15.5	5
213	7.08	5.89	2.61	VPS	0.70	SFS	1.17	L	3.9	75.6	20.5	5
214	7.02	5.89	2.76	VPS	0.61	SFS	1.27	L	5.2	74.4	20.4	5

KEY TO INFERRED ORIGIN

1 MODERN TILL
2 QUATERNARY TILL
3 OUTWASH
4 KAREWA GRAVEL
5 UPPER KAREWA SILTS
6 UPPER KAREWA SANDS
7 NON-GLACIAL DIAMICTONS
8 MODERN FLUVIAL SANDS
9 AEOLIAN LOESS

SIZE INTERVALS

Sand: -1 to 4 phi
Silt: 4 to 9 phi
Clay: 9 to 14 phi

FORMULAE FOR THE CALCULATION OF MEAN, SORTING, SKEWNESS AND KURTOSIS FOLLOW FOLK AND WARD (1957).

CATEGORIES FOR SORTING, SKEWNESS AND KUSTOSIS.

SORTING:	<0.35 phi	Very well sorted (VWS)
	0.35-0.50 phi	Well sorted (WS)
	0.50-0.71 phi	moderately well sorted (MWS)
	0.71-1.00 phi	Moderately sorted (MS)
	1.00-2.00 phi	Poorly sorted (PS)
	2.00-4.00 phi	Very poorly sorted (VPS)
	>4.00 phi	Extremely poorly sorted (EPS)
SKEWNESS:	+1.00 to +0.30	Strongly fine skewed (SFS)
	+0.30 to +0.10	Fine skewed (FS)
	+0.10 to -0.10	Near symmetrical (NS)
	-0.10 to -0.30	Coarse skewed (CS)
	-0.30 to -1.00	Strongly coarse skewed (SCS)
KURTOSIS:	<0.67	Very platykurtic (VP)
	0.67-0.90	Platykurtic (P)
	0.90-1.11	Mesokurtic (M)
	1.11-1.50	Leptokurtic (L)
	1.50-3.00	Very leptokurtic (VL)
	>3.00	Extremely leptokurtic (EL)

APPENDIX 7. OSTRACOD SPECIES COUNTS FOR THE PATTAN SECTION.

SAMPLE	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18
1	46	49	3	21	0	1	977	0	0	54	2	0	4	0	33	5	6	0
2	32	34	2	40	34	0	214	4	0	6	0	0	0	0	5	0	4	2
3	1	4	0	9	5	0	42	0	0	3	0	0	0	0	2	1	1	2
4	31	38	4	37	10	1	152	8	0	6	2	0	0	0	11	0	6	4
5	36	35	8	34	18	0	183	1	0	8	0	0	0	0	9	0	2	0
6	0	30	0	7	5	0	130	2	0	0	0	0	0	0	0	0	2	2
7	0	0	0	0	0	0	12	0	0	0	0	0	0	0	0	0	.5	8
8	55	0	0	0	0	0	134	0	0	0	0	0	0	0	0	0	4	4
9	56	7	4	9	0	1	221	0	0	0	0	0	0	0	0	0	6	4
10	124	19	0	4	0	2	351	4	0	34	0	0	0	0	1	0	5	2
11	10	0	0	0	0	0	9	5	0	0	0	0	0	0	0	0	<.5	0
12	6	1	0	0	1	2	20	0	0	0	0	0	0	0	0	0	<.5	0
13	29	11	4	33	13	1	69	1	0	0	0	0	0	0	0	0	1	0
14	31	8	0	17	7	0	41	0	0	0	0	0	0	0	0	0	.5	0
15	26	7	0	20	5	2	15	0	0	0	0	0	0	0	0	0	<.5	0
16	37	22	10	44	7	0	14	0	1	1	0	0	0	0	0	0	1	0
17	24	4	0	14	6	1	13	0	0	4	0	0	0	0	0	0	1	4
18	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
19	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
20	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
21	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
22	173	7	0	0	10	0	311	11	7	2	1	0	0	0	0	0	5	2
23	148	29	0	0	0	0	271	0	8	47	0	0	3	0	1	0	10	4
24	104	5	0	4	11	0	328	15	4	69	0	1	0	0	0	0	5	2
25	83	9	0	0	4	0	147	3	0	26	0	0	0	0	1	1	6	4
26	63	14	0	2	14	0	67	16	9	0	0	0	0	0	0	0	4	4
27	78	9	0	11	32	0	112	5	0	2	2	1	0	0	0	0	3	2
28	87	12	1	2	4	0	232	6	0	2	0	0	0	0	0	0	7	4
29	25	1	1	0	13	0	21	0	0	0	0	0	0	0	0	0	1	2
30	109	10	2	27	25	0	190	14	2	2	1	0	0	0	0	0	8	4

APPENDIX 7. OSTRACOD SPECIES COUNTS FOR THE PATTAN SECTION.

SAMPLE	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18
31	40	4	1	6	15	0	41	13	0	0	0	0	0	0	0	0	1	0
32	55	7	0	0	21	0	3	0	0	0	0	0	0	0	1	0	2	4
33	19	2	0	2	5	0	5	0	0	0	0	0	0	0	0	0	1	4
34	0	0	0	0	0	0	3	0	0	0	0	0	0	0	0	0	<.5	4
35	6	0	0	0	1	0	5	0	0	0	0	0	0	0	0	0	<.5	4
36	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
37	42	0	0	3	1	0	11	0	1	0	0	0	0	0	0	0	1	2
38	28	3	1	0	21	0	20	1	0	0	0	0	0	0	0	0	1	4
39	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
40	148	15	0	35	74	0	125	5	0	7	0	0	0	0	0	0	3	2
41	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
42	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
43	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
44	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
45	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
46	115	4	0	22	0	8	69	1	0	0	0	0	0	0	2	1	2	2
47	297	3	0	30	8	0	157	2	0	2	0	1	0	0	0	0	10	4
48	238	4	0	43	12	0	229	0	0	1	0	1	0	0	1	1	5	2
49	178	8	0	9	8	0	94	4	0	2	0	0	2	0	0	2	15	0
50	20	4	0	0	7	0	234	12	1	58	0	1	0	0	0	3	13	8
51	353	4	0	72	11	0	306	20	4	25	5	0	0	0	54	5	17	4
52	181	11	0	16	5	0	220	1	0	39	5	9	0	0	55	0	11	4

APPENDIX 7. OSTRACOD SPECIES COUNTS FOR THE PATTAN SECTION.

SAMPLE	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18
53	78	4	0	6	7	0	154	7	2	15	0	2	0	0	38	0	4	4
54	56	6	0	1	3	0	341	0	12	45	0	1	0	1	11	7	19	8
55	119	5	0	0	10	0	208	11	3	25	0	0	0	0	5	1	15	8
56	189	5	0	3	16	0	373	9	0	18	0	0	0	1	12	0	25	8
57	65	7	0	5	1	1	87	0	0	3	0	0	1	0	6	0	2	2
58	104	24	0	32	0	0	46	0	0	4	0	0	2	0	10	1	4	4
59	160	14	0	29	10	2	113	2	2	36	0	0	0	2	25	0	9	8
60	87	6	0	79	0	3	56	1	0	3	0	0	2	0	17	2	1	0
61	335	17	0	95	0	1	79	5	0	8	0	0	0	0	280	9	23	8
62	355	31	0	120	0	0	194	7	1	55	0	1	0	7	37	6	33	8
63	225	15	0	75	2	1	116	2	0	16	0	0	0	0	25	0	19	8
64	547	26	0	62	0	0	147	6	0	52	0	0	0	1	61	0	36	8
65	271	14	0	33	0	3	176	4	0	30	0	2	0	2	35	2	23	8
66	168	11	0	14	3	0	165	1	2	38	0	5	0	0	87	0	20	8
67	5	0	0	0	1	0	23	2	0	0	0	0	0	0	0	0	<.5	4
68	121	5	0	48	25	0	215	9	0	1	0	0	0	0	2	0	8	4
69	93	7	0	17	12	0	76	6	0	1	0	0	0	0	5	0	4	4
70	130	11	1	15.	2	0	203	7	0	1	0	0	0	0	1	1	7	4
71	68	2	1	2	0	0	110	10	2	0	0	0	0	1	0	0	2	2
72	84	9	0	3	1	0	169	12	0	0	0	0	0	0	0	0	6	4
73	172	5	0	4	0	0	416	9	0	33	5	0	0	1	19	0	27	8
74	42	4	0	0	0	0	93	1	4	81	0	0	0	0	22	0	10	8
75	102	11	0	2	0	0	182	6	1	92	0	0	0	0	21	0	8	4
76	68	22	0	0	0	0	186	8	0	55	0	3	0	2	38	2	8	4
77	29	5	0	0	0	0	232	6	1	89	0	0	0	0	33	5	16	8
78	82	7	0	0	0	0	83	5	0	28	0	0	2	2	30	0	2	2
79	88	3	0	1	0	0	237	0	7	70	10	0	5	0	16	4	35	16
80	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
81	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
82	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0

1 Ilyocypris bradyi
2 Ilyocypris gibba
3 Ilyocypris kashmirensis
4 Limnocythere franki
5 Parastenocypris delormei
6 Cytherissa lacustris
7 Candona spp.
8 Candona neglecta
9 Candona candida
10 Cyclocypris spp.
11 Cypria opthalmica
12 Cypria aculeata
13 Cypridopsis vidua
14 Eucypris spp.
15 Darwinula stevonsoni
16 Potamocypris spp.
17 No. valves g^{-1}
18 Splitting Factor

APPENDIX 8. LIST OF OSTRACOD SPECIES FOUND IN KASHMIR.

SPECIES	LOWER[1] KAREWA	UPPER[1] KAREWA	PATTAN (UPPER KAREWA)	RECENT[1]
Cypris pubera		*		*
Cypris subglobosa		*		
Dolerocypris sp. cf *D. fasciata*		*		
Eucypris awantipurensis		*		
Eucypris sp. cf *E. clavata*		*	*	
Eucypris sp. cf *E. moguntiensis*	*	*		
Eucypris zenkeri		*		
Eucypris sp.			*	
Herpetocypris reptans		*		
Hetrocypris incongruens		*		*
Isocypris priomena		*		
Parastenocypris delormei		*	*	*
Stenocypris major				*
Zonocypris costata	*			
Cypridopsis aculeata		*		
Cypridopsis diebeli		*		
Cypridopsis vidua	*	*	*	*
Cypretta mckenziei		*		
Potamocypris minuta patriciae		*		*
Potamocypris smaragdina		*		*

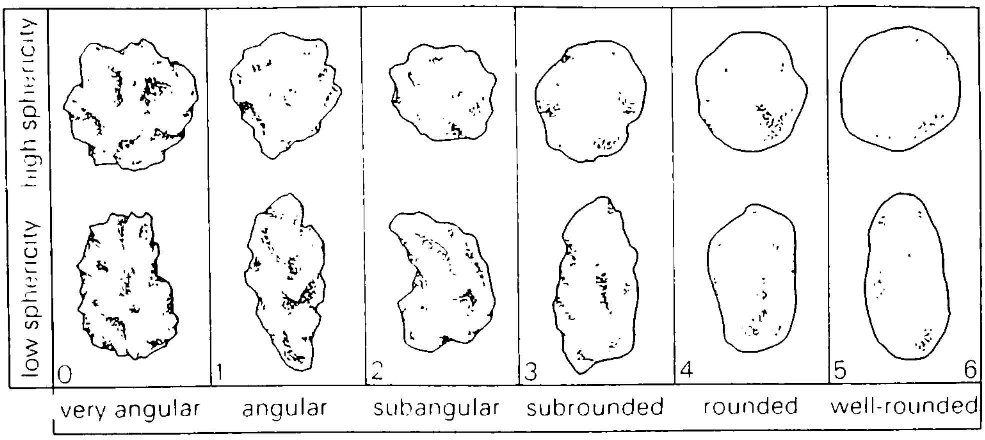
APPENDIX 8. CONTINUED.

SPECIES	LOWER[1] KAREWA	UPPER[1] KAREWA	PATTAN (UPPER KAREWA)	RECENT[1]
Potamocypris (Cyprilla) arcuata		*		
Potamocypris (Cyprilla) humilis		*		
Potamocypris (Cyprilla) pandei		*		
Potamocypris (Cyprilla) staplini		*		*
Potamocypris sp.			*	
Candona albicans		*		
Candona sp. cf _C. angulata_		*		
Candona candida	*	*	*	*
Candona compressa	*	*		*
Candona detecta		*		
Candona fabaeformis	*	*		
Candona sp. cf _C. havanaensis_		*		
Candona kashmirensis	*	*		
Candona lactea	*	*		*
Candona marengoensis	*	*		*
Candona neglecta	*	*	*	*
Candona sp. cf _C. pearlensis_		*		
Candona rawsoni		*		
Candona sp. cf _C. stammeri_		*		

APPENDIX 8. CONTINUED.

SPECIES	LOWER[1] KAREWA	UPPER[1] KAREWA	PATTAN (UPPER KAREWA)	RECENT[1]
Candona spp.			*	
Candonopsis kingsleii		*		*
Cyclocypris laevis	*	*	(*)	*
Cyclocypris ovum		*	(*)	
Cyclocypris spp.			*	
Cypria opthalmica		*	*	*
Ilyocypris bensoni	*			
Ilyocypris bhatiai		*		
Ilyocypris bradyi		*	*	*
Ilyocypris gibba	*	*	*	
Ilyocypris kashmirensis		*	*	
Ilyocypris shawneetownensis		*		*
Darwinula stevonsoni	*	*	*	*
Cytherissa lacustris	*	*	*	
Limnocythere blankenbergensis		*		
Limnocythere franki	*	*	*	
Limnocythere sancti patricii	*			

[1] After Singh (1974). Species from Pattan section recorded in this study



Appendix 9

Visual comparison chart for particle roundness (from Powers, 1953)

APPENDIX 10. SUMMARY OF SEDIMENTOLOGICAL ANALYSES UNDERTAKEN ON DIAMICTON SAMPLES.

SITE	SAMPLE	PARTICLE SIZE ANALYSIS	CLAST SHAPE	GRAIN SHAPE	MOIST COLOUR	% $CaCO_3$
Mangom	57	*	*	*	10YR 3/3	0.0
Mangom	122	*		*	7.5YR 4/4	0.0
Mangom	162	*		*	10YR 3/4	0.0
Mangom	165	*		*	10YR 3/4	0.0
Kangan	8	*	*	*	10YR 5/6	0.0
Gund	53	*	*	*	2.5Y 4/2	5.1
Gund	137	*	*	*	10YR 3/4	0.0
Gund	174	*	*	*	2.5Y 4/4	0.0
Rezan	55	*	*	*	5Y 5/3	0.0
Rezan	154	*	*	*	10YR 4/4	0.0
Rezan	155	*	*	*	2.5Y 3/2	0.0
Rezan	176	*	*	*	10YR 4/4	0.0
Gagangiyer	56	*	*	*	2.5Y 4/2	0.0
Sonamarg M.	120	*	*	*	10YR 3/4	0.0
Sonamarg M.	136	*	*	*	10YR 3/3	19.1
Sonamarg M.	140	*	*	*	10YR 3/4	0.0
Sonamarg O.	3	*	*	*	5Y 6/3	59.8
Sonamarg O.	6	*		*	5Y 4/3	27.2
Sonamarg O.	14	*	*	*	5Y 6/3	44.6
Sonamarg O.	15	*	*	*	5Y 6/3	59.6
Sonamarg O.	20	*	*	*	5Y 4/3	0.5
Sonamarg O.	47	*	*	*	5Y 6/2	45.3
Sonamarg O.	138	*	*	*	2.5Y 3/2	42.6
Sonamarg O.	164	*	*	*	2.5Y 3/2	nd
Sonamarg O.	177	*	*	*	2.5Y 4/2	49.3
Sonamarg F.	25	*	*	*	2.5Y 4/4	51.7
Sonamarg F.	148	*	*	*	2.5Y 3/2	nd
Sonamarg F.	157	*	*	*	2.5Y 3/2	25.0
Sonamarg F.	159	*	*	*	10YR 3/4	24.8
Sonamarg F.	173	*	*	*	2.5Y 3/2	nd
Tajiwas	7	*	*	*	2.5Y 4/2	4.4
Tajiwas	54	*	*	*	2.5Y 4/2	0.0
Tajiwas	123	*	*	*	2.5Y 4/2	0.0
Tajiwas	134	*		*	2.5Y 4/2	nd
Ganeshpur	126	*	*	*	10YR 4/3	0.0
Ganeshpur	168	*	*	*	7.5YR 4/4	0.7
Ganeshpur	170	*	*	*	10YR 3/4	nd
Lidru	11	*		*	10YR 4/3	0.0
Pahalgam O2.	133	*		*	2.5Y 4/4	0.0
Pahalgam O2.	144	*	*	*	2.5Y 5/4	0.0
Pahalgam O2.	152	*	*	*	2.5Y 4/4	0.0

APPENDIX 10. CONTINUED.

SITE	SAMPLE	PARTICLE SIZE ANALYSIS	CLAST SHAPE	GRAIN SHAPE	MOIST COLOUR	% $CaCO_3$
Pahalgam M.	4	*		*	2.5Y 6/4	nd
Pahalgam M.	12	*	*	*	7.5YR 4/4	0.0
Pahalgam M.	17	*			5Y 6/4	31.6
Pahalgam M.	19	*	*	*	2.5Y 6/4	17.4
Pahalgam M.	135	*	*	*	5Y 5/3	nd
Pahalgam M.	160	*	*	*	5Y 6/6	11.8
Pahalgam M.	172	*	*	*	2.5Y 5/4	17.2
Pahalgam Ol.	5	*	*	*	10YR 3/4	0.0
Pahalgam Ol.	128	*	*	*	7.5YR 3/2	0.0
Pahalgam Ol.	158	*	*	*	10YR 4/6	0.0
Pahalgam Ol.	163	*	*	*	10YR 3/4	0.0
E. Liddar F.	10	*	*	*	10YR 4/4	20.3
Mondlan	130	*	*	*	2.5Y 6/6	0.0
Mondlan	156	*	*	*	2.5Y 5/6	0.0
Mondlan	166	*	*	*	2.5Y 5/6	0.0
Aru O.	153	*	*	*	2.5Y 4/4	0.0
Aru O.	169	*	*	*	10YR 4/4	0.0
Aru O.	175	*	*	*	10YR 4/4	0.0
Aru F.	139	*	*	*	10YR 4/3	0.0
Aru O.	161	*		*	7.5YR 3/2	0.0
Liddarwat	58	*	*	*	10YR 3/3	0.0
Liddarwat	125	*	*	*	2.5Y 5/4	0.0
Liddarwat	127	*	*	*	10YR 3/3	0.0
Liddarwat	147	*	*	*	2.5Y 4/4	0.0
Liddarwat	171	*		*	2.5Y 4/4	0.0
Kolahoi	61	*	*	*	2.5Y 2/0	0.0
Kolahoi	62	*	*	*	2.5Y 4/0	0.0
Kolahoi	63	*	*	*	2.5Y 4/0	0.0
Kolahoi	124	*	*	*	5Y 5/4	0.0
Kolahoi	129	*	*	*	2.5Y 5/2	0.0
Kolahoi	132	*	*	*	5Y 5/2	0.0
Kolahoi	145	*	*	*	5Y 5/1	0.0
Kolahoi	151	*	*	*	2.5Y 5/2	0.0
Gulmarg M.	27	*	*	*	nd	0.0
Gulmarg M.	131	*	*	*	10YR 4/4	0.0
Gulmarg M.	141	*	*	*	2.5Y 4/4	0.0
Gulmarg M.	142	*	*	*	2.5Y 5/4	0.0
Gulmarg M.	146	*	*	*	2.5Y 4/4	0.0
Gulmarg M.	149	*	*	*	10YR 3/4	0.0
Gulmarg M.	150	*	*	*	10YR 3/4	0.0
Gulmarg O.	143	*	*	*	10YR 3/3	0.0
Gulmarg O.	167	*	*	*	10YR 3/6	0.0

APPENDIX 11. ROUNDNESS OF CLASTS AND MEDIUM SAND GRAINS FROM DAIMICTONS 1. CLASTS

SITE	SAMPLE	CLAST ROUNDNESS							LITHOLOGY			
		0	1	2	3	4	5	6	P	L	Q	O
Mangom	57	3	27	13	7	0	0	0	38	0	1	11
Kangan	8	3	40	5	2	0	0	0	27	0	1	22
Gund	53	3	17	24	6	0	0	0	25	3	9	13
Gund	137	2	29	19	0	0	0	0	25	0	2	23
Gund	174	6	36	7	1	0	0	0	25	0	0	25
Rezan	55	22	23	5	0	0	0	0	15	0	0	35
Rezan	154	15	35	0	0	0	0	0	0	0	0	50
Rezan	155	11	38	1	0	0	0	0	0	0	14	36
Rezan	176	15	33	2	0	0	0	0	28	0	0	22
Gagangiyer	56	19	31	0	0	0	0	0	39	0	5	6
Sonamarg M.	120	13	36	1	0	0	0	0	31	0	1	18
Sonamarg M.	136	0	24	17	6	2	1	0	9	25	4	12
Sonamarg M.	140	2	38	7	3	0	0	0	35	0	2	13
Sonamarg O.	3	0	16	21	6	4	2	1	2	32	4	12
Sonamarg O.	14	0	1	25	17	7	0	0	6	35	1	8
Sonamarg O.	15	1	10	12	9	12	4	2	6	36	0	8
Sonamarg O.	20	11	39	0	0	0	0	0	33	0	4	13
Sonamarg O.	47	0	32	8	7	3	0	0	5	39	0	6
Sonamarg O.	138	0	5	3	4	14	18	6	0	50	0	0
Sonamarg O.	164	0	4	4	5	21	10	3	4	29	0	7
Sonamarg O.	177	0	1	17	12	11	5	45	0	50	0	0
Sonamarg F.	25	4	42	4	0	0	0	0	0	45	0	5
Sonamarg F.	148	16	32	1	1	0	0	0	0	46	0	4
Sonamarg F.	157	8	42	0	0	0	0	0	8	42	0	0
Sonamarg F.	159	10	35	5	0	0	0	0	0	47	0	3
Sonamarg F.	173	16	33	1	0	0	0	0	10	40	0	0
Tajiwas	7	0	26	16	7	1	0	0	27	16	3	4
Tajiwas	54	6	38	6	0	0	0	0	39	0	1	10
Tajiwas	123	0	26	22	1	1	0	0	14	15	4	7
Ganeshpur	126	13	34	3	0	0	0	0	31	0	0	19
Ganeshpur	168	2	48	0	0	0	0	0	34	0	0	16
Ganeshpur	170	0	4	35	11	0	0	0	41	0	1	8
Pahalgam O2.	144	0	0	29	13	7	1	0	16	0	0	4
Pahalgam O2.	152	11	30	6	3	0	0	0	45	0	0	5
Pahalgam M.	5	4	38	7	1	0	0	0	34	0	0	16
Pahalgam M.	12	0	36	9	2	1	0	0	36	0	0	14
Pahalgam M.	19	12	28	8	2	0	0	0	17	22	4	7
Pahalgam M.	135	3	43	4	0	0	0	0	29	19	2	0
Pahalgam M.	160	0	8	24	15	3	0	0	35	7	2	6
Pahalgam M.	172	4	43	3	0	0	0	0	29	13	0	8
Pahalgam O1.	128	0	39	10	1	0	0	0	46	0	0	4

APPENDIX 11. CONTINUED.

SITE	SAMPLE	CLAST ROUNDNESS							LITHOLOGY			
		0	1	2	3	4	5	6	P	L	Q	O
Pahalgam Ol.	158	2	27	14	6	1	0	0	31	0	1	18
Pahalgam Ol.	163	0	13	20	10	3	4	0	22	0	2	26
E. Liddar F.	10	7	39	4	0	0	0	0	21	18	0	11
Mondlan	130	2	23	18	7	0	0	0	38	0	1	11
Mondlan	156	0	33	14	2	1	0	0	35	0	1	14
Mondlan	166	0	20	16	6	8	0	0	40	0	1	9
Aru O.	153	0	0	0	27	21	2	0	38	0	0	12
Aru O.	169	0	26	20	3	0	1	0	38	0	1	11
Aru O.	175	0	7	30	13	0	0	0	34	0	3	13
Aru F.	161	4	45	1	0	0	0	0	0	0	0	50
Liddarwat	58	4	26	11	5	4	0	0	36	0	2	12
Liddarwat	125	1	35	11	3	0	0	0	41	0	4	5
Liddarwat	127	1	38	7	4	0	0	0	38	0	1	11
Liddarwat	147	0	34	14	1	1	0	0	41	0	1	8
Kolahoi	61	16	32	1	1	0	0	0	15	0	30	5
Kolahoi	62	21	40	7	1	0	0	0	30	0	5	15
Kolahoi	63	4	31	13	2	0	0	0	35	0	3	12
Kolahoi	124	17	33	0	0	0	0	0	49	0	0	1
Kolahoi	129	25	24	1	0	0	0	0	45	0	2	3
Kolahoi	132	23	26	1	0	0	0	0	48	0	2	0
Kolahoi	145	4	27	7	11	1	0	0	37	0	2	11
Kolahoi	151	8	41	1	0	0	0	0	48	0	2	0
Gulmarg M.	27	2	27	17	8	3	0	0	45	0	0	5
Gulmarg M.	131	1	28	13	7	0	1	0	35	0	4	11
Gulmarg M.	141	0	23	21	5	1	0	0	30	0	2	18
Gulmarg M.	142	0	36	13	1	0	0	0	30	0	1	19
Gulmarg M.	146	0	10	19	13	6	1	1	36	0	1	13
Gulmarg M.	149	1	28	15	5	1	0	0	37	0	0	13
Gulmarg M.	150	0	0	27	17	4	2	0	44	0	3	3
Gulmarg O.	143	0	13	29	6	2	0	0	20	0	1	29
Gulmarg O.	167	0	24	15	8	3	0	0	27	0	2	21

APPENDIX 11. CONTINUED.
2. GRAINS.

SITE	SAMPLE	GRAIN ROUNDNESS 0	1	2	3	4	5	6
Mangom	57	40	9	1	0	0	0	0
Mangom	122	31	16	3	0	0	0	0
Mangom	162	0	33	17	0	0	0	0
Mangom	165	11	35	4	0	0	0	0
Kangan	8	9	28	8	5	0	0	0
Gund	53	1	30	11	8	0	0	0
Gund	137	29	19	2	0	0	0	0
Gund	174	29	18	3	0	0	0	0
Rezan	55	26	24	0	0	0	0	0
Rezan	155	16	31	3	0	0	0	0
Rezan	176	21	23	5	1	0	0	0
Gagangiyer	56	41	7	0	0	0	0	0
Sonamarg M.	120	19	23	8	0	0	0	0
Sonamarg M.	136	8	32	9	1	0	0	0
Sonamarg M.	140	28	19	3	0	0	0	0
Sonamarg O.	3	17	14	11	5	3	0	0
Sonamarg O.	6	14	22	11	3	0	0	0
Sonamarg O.	14	2	29	13	6	0	0	0
Sonamarg O.	15	4	31	9	5	1	0	0
Sonamarg O.	20	6	33	11	0	0	0	0
Sonamarg O.	47	25	16	8	1	0	0	0
Sonamarg O.	138	31	12	7	0	0	0	0
Sonamarg O.	164	23	16	9	2	0	0	0
Sonamarg O.	177	2	33	13	2	0	0	0
Sonamarg F.	25	19	19	8	4	0	0	0
Sonamarg F.	148	32	13	5	0	0	0	0
Sonamarg F.	157	18	26	6	0	0	0	0
Sonamarg F.	159	18	27	5	0	0	0	0
Sonamarg F.	173	25	18	5	2	0	0	0
Tajiwas	7	1	26	21	2	0	0	0
Tajiwas	54	25	22	3	0	0	0	0
Tajiwas	123	32	17	1	0	0	0	0
Tajiwas	134	30	17	3	0	0	0	0
Ganeshpur	126	1	36	11	3	0	0	0
Ganeshpur	168	11	25	9	5	0	0	0
Ganeshpur	170	14	22	9	4	0	0	0
Lidru	11	5	40	5	0	0	0	0
Pahalgam O2.	133	17	23	10	0	0	0	0
Pahalgam O2.	144	9	38	3	0	0	0	0
Pahalgam O2.	152	31	17	2	0	0	0	0
Pahalgam M.	4	19	23	7	1	0	0	0
Pahalgam M.	12	15	27	8	0	0	0	0
Pahalgam M.	19	0	28	17	3	2	0	0

SITE	SAMPLE	GRAIN ROUNDNESS 0	1	2	3	4	5	6
Pahalgam M.	135	0	10	38	8	2	0	0
Pahalgam M.	160	18	20	11	1	0	0	0
Pahalgam M.	172	8	28	11	3	0	0	0
Pahalgam O1.	5	20	20	10	0	0	0	0
Pahalgam O1.	128	3	30	12	4	1	0	0
Pahalgam O1.	158	2	30	12	6	0	0	0
Pahalgam O1.	163	28	16	5	0	1	0	0
E. Liddar F.	10	29	16	3	1	1	0	0
Mondlan	130	14	29	7	0	0	0	0
Mondlan	156	1	35	12	2	0	0	0
Mondlan	166	1	35	12	1	1	0	0
Aru O.	153	12	25	10	3	0	0	0
Aru O.	169	27	19	4	0	0	0	0
Aru O.	175	17	21	12	0	0	0	0
Aru F.	139	20	10	11	9	0	0	0
Aru F.	161	28	20	2	0	0	0	0
Liddarwat	58	12	20	11	4	2	1	0
Liddarwat	125	27	20	3	0	0	0	0
Liddarwat	127	15	19	15	1	0	0	0
Liddarwat	147	23	20	7	0	0	0	0
Liddarwat	171	7	37	6	0	0	0	0
Kolahoi	61	16	18	4	2	0	0	0
Kolahoi	62	22	23	5	0	0	0	0
Kolahoi	63	14	26	8	2	0	0	0
Kolahoi	124	6	37	7	0	0	0	0
Kolahoi	129	21	22	7	0	0	0	0
Kolahoi	132	30	16	4	0	0	0	0
Kolahoi	145	25	21	4	0	0	0	0
Kolahoi	151	20	25	5	0	0	0	0
Gulmarg M.	27	8	36	6	0	0	0	0
Gulmarg M.	131	29	19	1	0	1	0	0
Gulmarg M.	141	24	20	6	0	0	0	0
Gulmarg M.	142	14	34	2	0	0	0	0
Gulmarg M.	146	22	25	3	0	0	0	0
Gulmarg M.	149	14	29	7	0	0	0	0
Gulmarg M.	150	29	21	0	0	0	0	0
Gulmarg O.	143	21	19	10	0	0	0	0
Gulmarg O.	167	23	22	5	0	0	0	0

Key
M = Moraine
O = Outwash
F = Fan

APPENDIX 12. TECHNIQUES USED IN THE ANALYSIS OF SEDIMENTOLOGICAL DATA USING PRINCIPAL COMPONENTS ANALYSIS.

Analyses carried out using spida (Statistical package for interactive data analyses).

1. Data were standardized to give each variable a mean of zero and standard deviation of 1. From each observation, the mean of the variable was subtracted and the result divided by the standard deviation.
2. A variance-covariance matrix was calculated. Eigenvectors and eigenvalues of this matrix were then calculated.
3. The eigenvalues which contributed the greatest amount of variance were selected.
4. The linear combination of those associated eigenvectors multiplied by the observed observations and summed provided the PCA scores that there then plotted.

Reference: DAVIS, J.C. (1973) Statistics and data analysis in Geology. New York, 550pp.

APPENDIX 13. THE EFFECT OF CLAST LITHOLOGY ON SHAPE. (clast samples from lower Karewa conglomerates. Statistical analysis carried out using pair-wise Kolmogorov-Smirnov tests of association

Hirpur 1.

Sample vs	Sample	Dmax	Significance
Limestone	Panjal Trap	0.4	**
Limestone	Other	0.38	**
Limestone	quartzite	0.72	**
Panjal Trap	Other	2.000004e-02	
Panjal Trap	Quartzite	0.32	*
Other	Quartzite	0.34	**

Hirpur 2

Sample vs	Sample	Dmax	Significance
Limestone	Panjal Trap	0.34	**
Limestone	Other	0.38	**
Limestone	quartzite	0.68	**
Panjal Trap	Other	3.999999e-02	
Panjal Trap	Quartzite	0.34	**
Other	Quartzite	0.3	*

* = Significant at 5%, ** = Significant at 1%
50 clasts counted in each sample.

APPENDIX 14. CONVENTIONS USED FOR RD STUDIES.

1. SURFACE BOULDER FEATURES

SURFACE BOULDER FREQUENCY (SBF). The number of boulders greater than 30cm diameter found within a 30 by 6m quadrangle placed along the moraine crest. The quadrangle is placed in an area where the visible concentration of boulders is at a maximum (Burke & Birkeland, 1979).

PERCENTAGE OF SPLIT BOULDERS. For 50 boulders selected randomly, and of the same lithology, the number of split boulders was determined. A split boulder is one which shows cracking or breakage along a planar crack, which has occurred since emplacement of the boulder, by a mechanism other than spalling (Burke & Birkeland, 1979).

PERCENTAGE OF PITTED BOULDERS. For 50 boulders selected randomly and of the same lithology (not the same 50 as observed in the above measure), the number of pitted boulders was determined. A pitted boulder has one or more approximately circular and concave depressions, apparently caused by granular disintigration (Burke & Birkeland, 1979).

PIT DEPTH. The depth of the pit from the surface of the rock to the pit bottom. Depth of the deepest pit is measured on each boulder (Burke & Birkeland, 1979).

PERCENTAGE OF WEATHERED BOULDERS. For 50 boulders selected randomly and of the same lithology (not the same 50 as observed in either of the above measures), the number of weathered boulders was determined. A boulder was considered weatherd if more than 50 % of the exposed surface was rough to the touch due to single-grain relief (Burke & Birkeland, 1979).

POINT COMPRESSIVE STRENGTH. On each of the 50 boulders counted for weathering, 3 individual determinations of point compressive strength were made, using a Schmidt Hammer (Perrott & Goudie, 1984).

HEIGHT OF RESISTANT QUARTZITIC INTRUSIONS. On each of the boulders measured for weathering, the maximum height of resistant intrusions above the rock surface was measured (Burke and Birkeland, 1979).

2.SUBSURFACE BOULDER FEATURES

MAXIMUM CORNER ANGULARITY. 50 clasts of the same lithology retrieved from the soil pit. Angularity of the sharpest corner was determined using radius charts (Perrott & Goudie, 1984).

APPENDIX 14. CONTINUED.

3. SURFICIAL COVER

LOESS THICKNESS. Thickness of non-pedogenically altered loess on the moraine.

SOIL PROPERTIES. Routine field assessments of horizonation, colour, consistence, structure, organic content, presence of films and coats, pH. (Birkeland, 1984).

APPENDIX 15. DATA FOR THE PRESENT GLACIER COVER IN KASHMIR.

GLACIER NO.	MAP NO.& SECTOR		T (m)	B (m)	ORIENT-ATION	AREA (km^2)	ELA (m)	GT (m)	VALLEY
1	K10	7	4298	4145	N	0.13	4206	4384	1
2	K10	9	UC	UC	E	0.13	ND	4257	1
3	K10	9	UC	UC	NE	0.28	ND	4257	1
4	K10	9	4237	3993	N	0.31	4090	4257	1
5	K10	9	4237	3993	N	0.16	4090	4257	1
6	K10	9	4206	4054	N	0.05	4225	4257	1
7	K10	9	4420	3871	NW	1.45	4090	4257	1
8	K10	9	4176	3901	NE	0.36	4011	4257	1
9	K10	9	4176	3871	E	0.23	3993	4257	1
10	K10	9	4206	3993	E	0.21	4078	4257	1
11	K10	9	3993	3780	NE	0.28	3865	4257	1
12	K10	9	4023	3901	NE	0.31	3950	4257	1
13A	K10	9	3993	3932	NE	0.16	3956	4257	1
13B	N3	5	UC	UC	N	0.23	ND	4531	5
14	K5	9	4450	4389	NW	0.13	4414	4399	1
15	K5	9	4572	4115	NE	0.31	4298	4399	1
16	K5	9	UC	UC	E	0.13	ND	4399	1
17	K5	9	4663	4084	NE	0.85	4316	4399	1
18	K5	9	4536	4084	NE	0.14	4265	4399	1
19	N4	3	UC	UC	N	0.13	ND	4602	2
20	N4	3	4633	4450	N	0.44	4523	4602	2
21	N4	3	UC	UC	NW	0.16	ND	4602	2
22	N4	3	4724	4145	NW	1.81	4377	4602	2
23	N4	3	4663	4542	E	0.13	4590	4602	2
24	N4	3	4420	3993	N	0.28	4164	4602	2
25	N7	1	4542	4176	E	0.16	4316	4701	7
26	N7	1	4572	3955	N	1.76	4202	4701	7
27	N7	1	4572	4359	W	0.08	4444	4701	7
28	N7	1	UC	UC	E	0.52	ND	4701	7
29	N7	1	UC	UC	NE	0.08	ND	4701	7
30	N7	1	4572	4176	W	0.21	4334	4701	7
31	N7	1	UC	UC	NW	0.13	ND	4701	7
32	N7	1	4694	4359	N	0.16	4493	4701	7
33	N7	1	4694	4420	N	0.88	4529	4701	7
34	N7	1	UC	UC	N	0.80	ND	4701	7
35	N7	1	UC	UC	N	0.85	ND	4701	7
36	N7	1	4694	3962	N	3.94	4255	4701	7
37	N7	1	UC	UC	NW	0.08	ND	4701	7
38	N7	1	4755	4237	NW	0.57	4443	4701	7
39	N7	1	UC	UC	SW	0.21	ND	4701	7
40	N7	1	UC	UC	SW	0.18	ND	4701	7
41	N7	2	UC	UC	W	0.85	ND	<4763	7
42	N7	5	4663	4481	W	0.21	4554	<4763	7

APPENDIX 15. CONTINUED.

GLACIER NO.	MAP NO.& SECTOR		T (m)	B (m)	ORIENT-ATION	AREA (km^2)	ELA (m)	GT (m)	VALLEY
43	N7	5	4663	4084	SW	0.21	4304	<4763	7
44	N7	5	UC	UC	W	0.28	ND	<4763	7
45	N7	5	UC	UC	W	0.49	ND	<4763	7
46	N7	5	4877	4481	N	0.36	4639	<4763	7
47	N7	5	4267	4084	N	0.16	4157	<4763	7
48	N7	5	4877	4206	NW	4.25	4474	<4763	4
49	N7	5	UC	UC	NW	0.13	ND	<4763	7
50	N7	4	UC	UC	N	1.17	ND	4677	7
51	N7	4	UC	UC	W	0.08	ND	4677	7
52	N7	4	UC	UC	NE	0.08	ND	4677	7
53	N7	4	4633	4176	NW	0.93	4359	4677	7
54	N7	4	UC	UC	NW	0.18	ND	4677	7
55	N7	4	UC	UC	NE	1.37	ND	4677	7
56	N7	4	4420	4267	N	0.31	4328	4677	7
57	N7	8	UC	UC	NW	2.64	ND	4512	7
58	N7	8	UC	UC	SW	0.31	ND	4512	7
59	N7	7	4572	3627	N	0.73	4005	4720	7
60	N7	7	4572	3719	N	0.96	4060	4720	7
61	N8	1	4663	3444	N	1.97	3932	4253	2
62	N8	1	4389	4176	N	0.05	4261	4253	2
63	N8	1	4206	4115	S	0.03	4151	4253	2
64	N8	1	4206	4054	SW	0.08	4115	4253	2
65	N8	1	UC	UC	N	0.60	ND	4253	2
66	N8	1	UC	UC	NE	0.44	ND	4253	2
67	N8	2	4389	3993	N	0.75	4029	4481	2
68	N8	2	4389	3901	N	0.67	4097	4481	2
69	N8	3	4724	4145	N	1.61	4377	4769	4
70	N8	3	UC	UC	N	0.13	ND	4709	4
71	N8	2	4816	4267	NW	0.60	4487	4481	2
72	N8	2	UC	UC	E	0.28	ND	4481	2
73	N8	2	UC	UC	E	0.13	ND	4481	2
74	N8	2	UC	UC	NW	1.40	ND	4481	2
75	N8	2	UC	UC	NW	0.08	ND	4481	2
76	N8	2	4663	3444	N	16.37	3932	4481	2
77	N8	1	4298	3658	N	1.48	3914	4253	2
78	N8	1	3962	3444	NW	0.36	3652	4253	2
79	N8	1	4328	4225	W	0.13	4200	4253	2
80	N8	1	4602	4237	W	0.23	4383	4253	2
81	N8	4	4511	4267	W	0.16	4365	4570	2
82	N8	4	4511	4267	S	0.13	4365	4570	1
83	N8	4	UC	UC	SW	0.36	ND	4570	1
84	N8	4	UC	UC	SW	0.67	ND	4570	1
85	N8	5	4724	3993	S	2.62	4285	4690	1
86	N8	5	4511	4389	SE	0.05	4365	4690	1
87	N8	5	4663	4145	E	0.73	4353	4690	1
88	N8	4	UC	UC	N	0.16	ND	4570	1
89	N8	4	UC	UC	N	0.21	ND	4570	1
90	N8	4	UC	UC	NW	0.13	ND	4570	1
91	N8	5	4542	3932	N	0.44	4176	4690	1

APPENDIX 15. CONTINUED.

GLACIER NO.	MAP NO.& SECTOR		T (m)	B (m)	ORIENT-ATION	AREA (km^2)	ELA (m)	GT (m)	VALLEY
92	N8	5	4633	3932	N	0.65	4212	4690	1
93	N8	6	4420	4145	W	0.65	4255	4648	1
94	N8	6	UC	UC	NW	0.60	ND	4648	1
95	N8	6	UC	UC	NE	0.31	ND	4648	1
96	N8	6	UC	UC	NE	0.44	ND	4648	1
97	N8	8	4572	4359	E	0.52	4444	4633	1
98	N8	8	4755	4054	N	2.54	4334	>4579	1
99	N8	9	4755	4054	N	0.88	4334	>4579	1
100	N8	9	4206	4054	N	0.13	4115	>4579	1
101	N8	9	4785	3962	N	1.40	4346	>4579	1
102	N8	9	4785	3962	NW	0.49	4346	>4579	1
103	N8	9	5029	4267	N	0.28	4572	>4579	1
104	N8	9	4816	4054	N	2.28	4359	>4579	1
105	N8	9	UC	UC	N	0.80	ND	>4579	1
106	N8	9	UC	UC	N	1.24	ND	>4579	1
107	N8	9	UC	UC	N	0.44	ND	>4579	1
108	N8	7	4846	3657	N	6.50	4133	ND	1
109	N8	8	4511	4267	N	0.41	4365	4633	1
110	O5	2	4572	4359	W	0.08	4444	4581	3
111	O5	2	4298	3993	N	0.36	4298	4581	3
112	O5	2	4846	4054	N	0.57	4371	4581	3
113	O5	2	4846	3962	N	0.44	4316	4581	3
114	O5	2	4907	4237	N	0.03	4505	4581	3
115	O5	2	4877	4724	W	0.08	4785	4581	3
116	O5	2	4907	4663	W	0.16	4761	4581	3
117	O5	2	4816	4511	NW	0.13	4633	4581	3
118	O5	2	4633	4511	NW	0.05	4560	4581	3
119	O5	2	4663	4420	NW	0.08	4517	4581	3
120	O5	2	4663	4450	NW	0.21	4535	4581	3
121	K15	1	3962	3810	N	0.31	3871	4020	6
2	K15	1	3962	3810	N	0.05	3871	4020	6
123	K15	1	3962	3780	N	0.28	3853	4020	6
124	K15	1	4267	3712	N	2.93	4033	4020	6
125	K15	1	UC	UC	N	1.09	ND	4020	6
126	K15	1	4267	3566	N	2.18	3847	4020	6
127	K15	2	4267	3658	N	3.08	3901	4121	6
128	K15	2	4084	3932	N	0.28	3993	4121	6
129	K15	2	3962	3810	N	0.41	3871	4121	6
130	K14	7	4267	3993	NE	0.31	3920	4431	6
131	J15	3	4389	4176	N	0.28	4261	4679	5
132	J15	2	4602	4145	NW	0.52	4328	4518	5
133	J15	5	4999	4212	NE	3.50	4212	4740	5
134	J15	5	4877	3840	W	0.36	4255	4740	5
135	J15	5	4328	4115	NE	0.13	4200	4740	5
136	J15	5	4999	4377	E	1.32	4377	4740	5
137	J15	2	3962	3840	NE	0.08	3889	4518	5
138	K14	7	4267	3901	NE	0.60	4048	4431	6
139	K10	8	4481	4298	N	0.28	4371	4235	6

APPENDIX 15. CONTINUED.

GLACIER NO.	MAP NO.& SECTOR	T (m)	B (m)	ORIENT-ATION	AREA (km^2)	ELA (m)	GT (m)	VALLEY
140	N12 1	5014	4481	NW	0.52	4694	4659	4
141	N12 1	4572	4328	NW	0.16	4426	4659	4
142	N12 1	UC	UC	SW	0.41	ND	4659	4
143	N12 1	4877	4420	SW	2.41	4602	4659	4
144	N12 2	4907	4576	SW	0.85	4708	4659	4
145	N12 1	4694	4145	SW	0.65	4365	4659	4
146	N12 1	UC	UC	NW	0.08	ND	4659	4
147	N12 1	UC	UC	NW	0.08	ND	4659	4
148	N12 4	4873	3764	N	2.12	4209	4560	4
149	N12 4	4878	3719	N	2.31	4182	4560	4
150	N12 4	4878	3856	NW	2.18	4264	4560	4
151	N12 4	UC	UC	W	0.28	ND	4560	1
152	N12 4	UC	UC	W	0.28	ND	4560	1
153	N12 4	4694	4450	W	0.18	4548	4560	1
154	N12 4	4694	4298	NW	0.28	4456	4560	1
155	N12 4	4694	4267	NW	0.41	4438	4560	1
156	N12 4	4755	4267	NW	0.44	4462	4560	1
157	N12 7	UC	UC	N	0.60	ND	4669	1

EXPLANATION:
T equals the altitude of the glacier 'top' or headwall
B equals the height of the glacier 'bottom' or snout

GT equals Glaciation threshold

ELA equals equilibrium line of the glacier calculated using the toe-to-headwall-altitude-ratio method with a THAR of 0.4

'Sectors' are 5 x 5 minute quadrangles on the Survey of India 1:63 360 s maps and are numbered 1 to 9 starting from the top left-hand sector and moving to the bottom right-hand sector of the map.

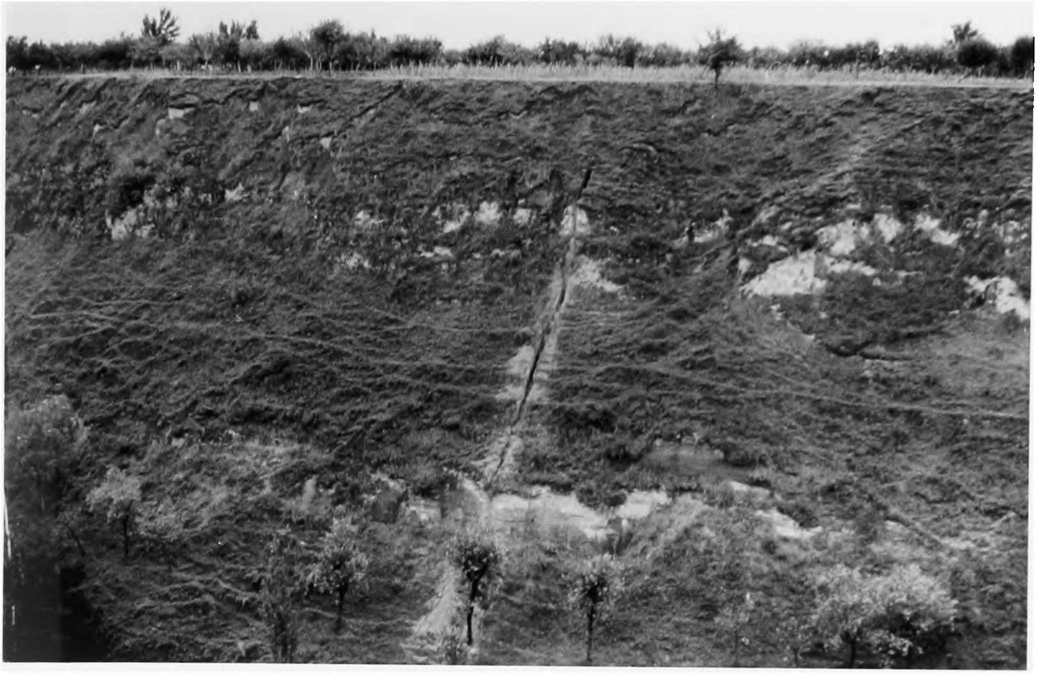
UC Denotes that the particular height was unclear on the map
ND Denotes that the particular value was not determined

VALLEYS ARE DENOTED AS FOLLOWS:
1 East Liddar
2 West Liddar
3 Lower Liddar and tributary valleys
4 Upper Sind and Tajiwas valleys
5 Lower Sind and Wangat Nala
6 Rembiara and Vishav Valleys
7 Ladakh (beyond the eastern extremity of Kashmir basin)

PLATES

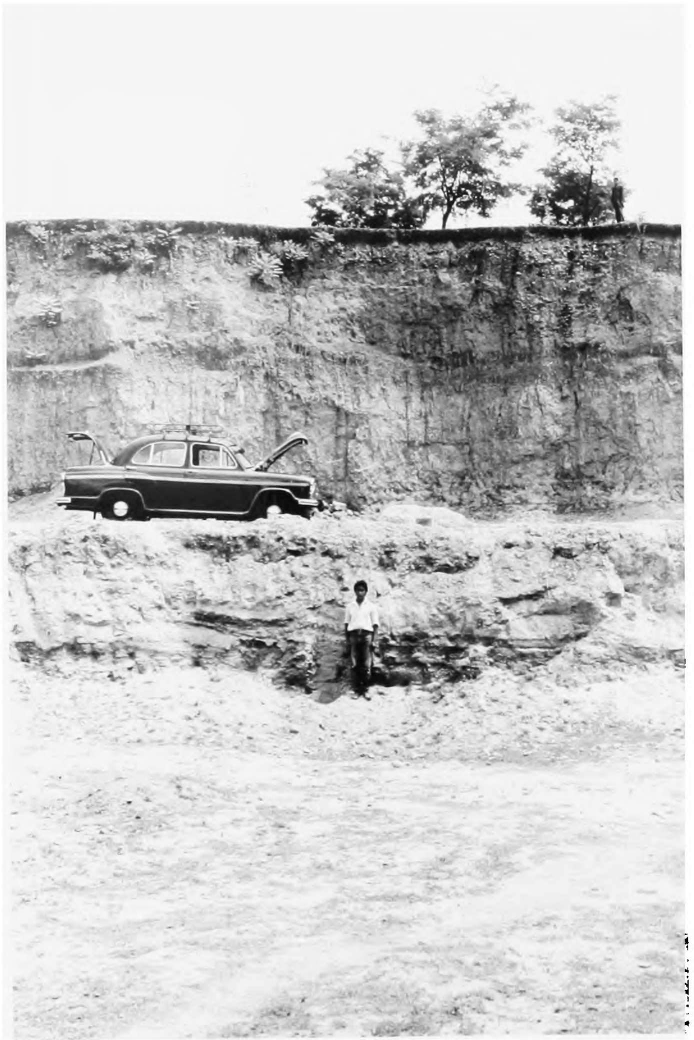
1. Tilted lower Karewa sediments exposed in the Rembiara Valley, near Hirpur. The exposed section is about 100 m. thick and the basal age about 2.8 MaBP.

2. Near-horizontal lower Karewa sediments exposed in the Romushi Valley, near Pakharpura. The exposed section is about 100 m. thick and the basal age about 2.0 MaBP.

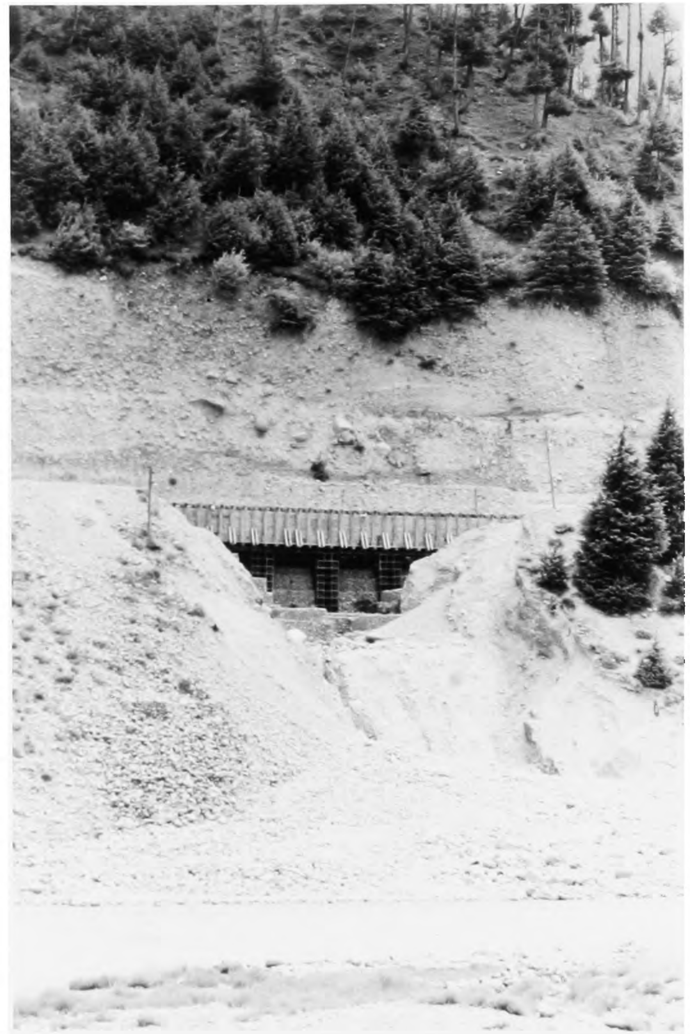


3. 25 metre-thick section exposed in gullied upper Karewa lacustrine sediments and loess at Pattan.

4. Upper Karewa sediments and loess exposed in a quarry at Sambur. The dark beds in the photograph are sand units and the light beds lacustrine silts.

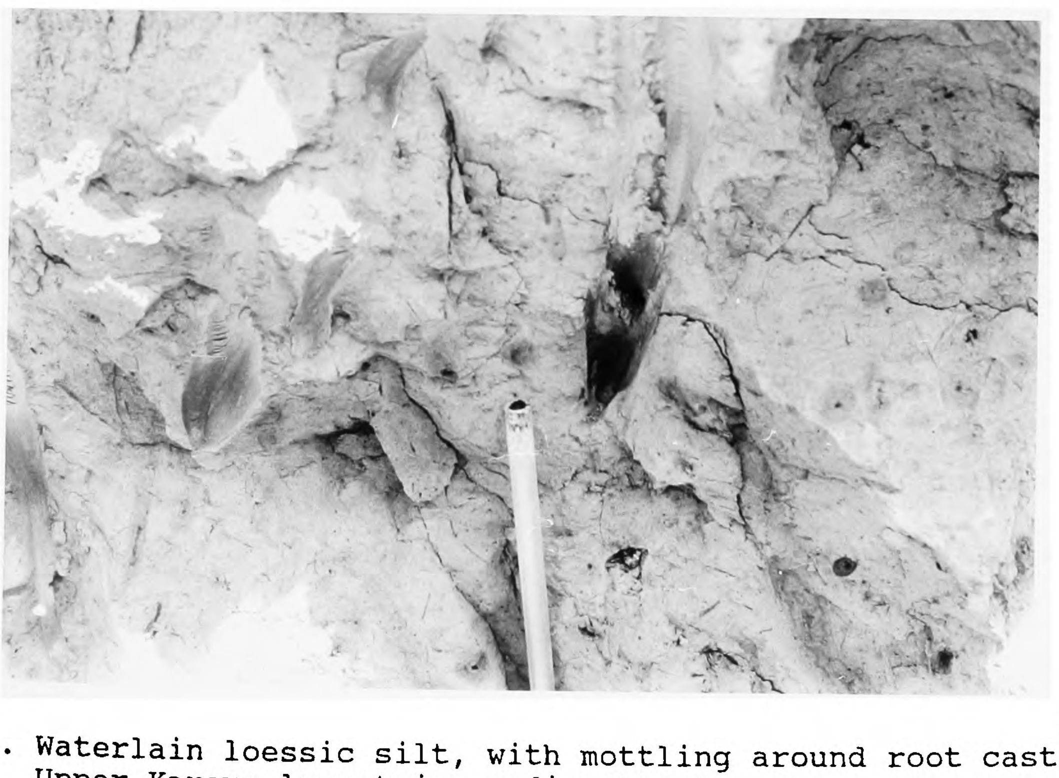


5. Upper Karewa lacustrine sediments and loess exposed in a quarry at Burzahom.

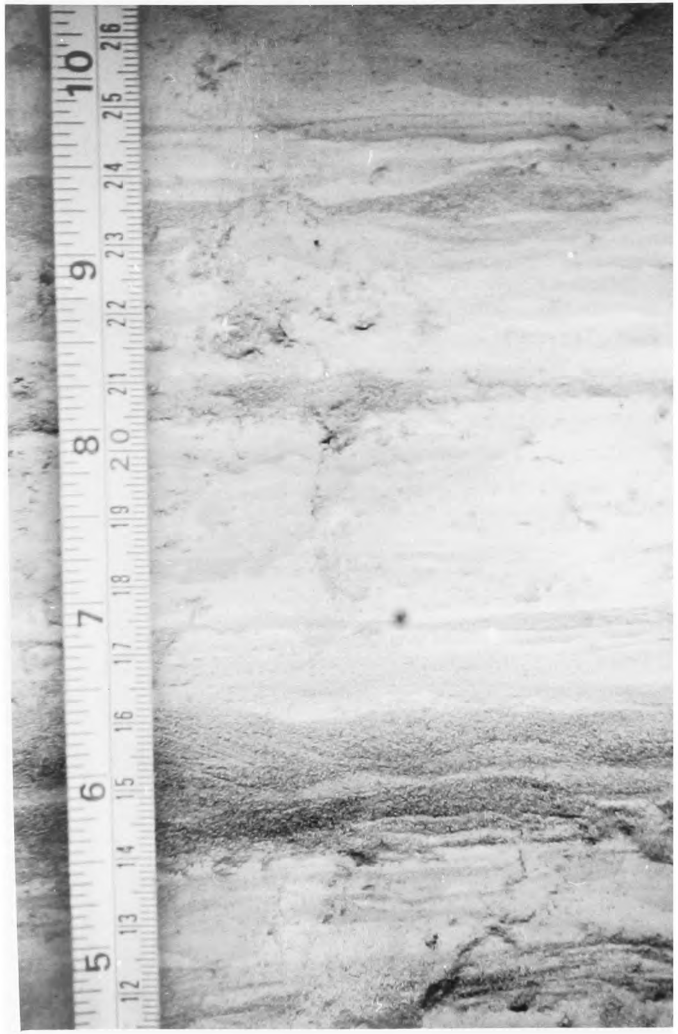


6. Conglomerate terraces in the lower Sind Valley, near Nunar. Note the large boulders above the aquaduct.

7. Conglomerate terraces in the lower Rembiara Valley near Balpora.

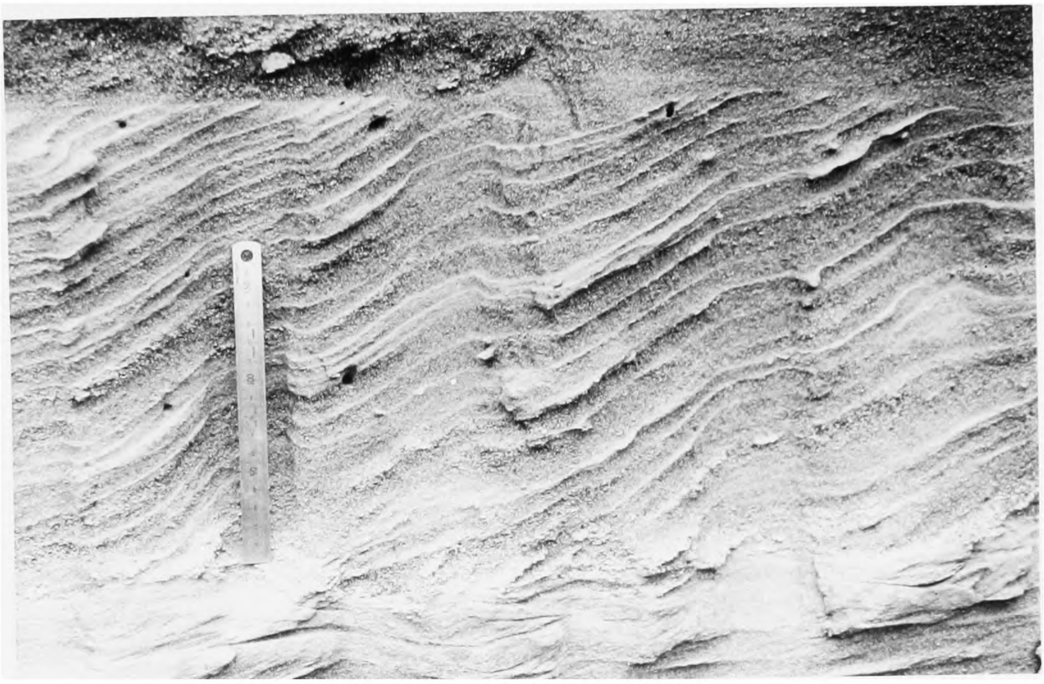


8. Waterlain loessic silt, with mottling around root casts. Upper Karewa lacustrine sediments, Burzahom.

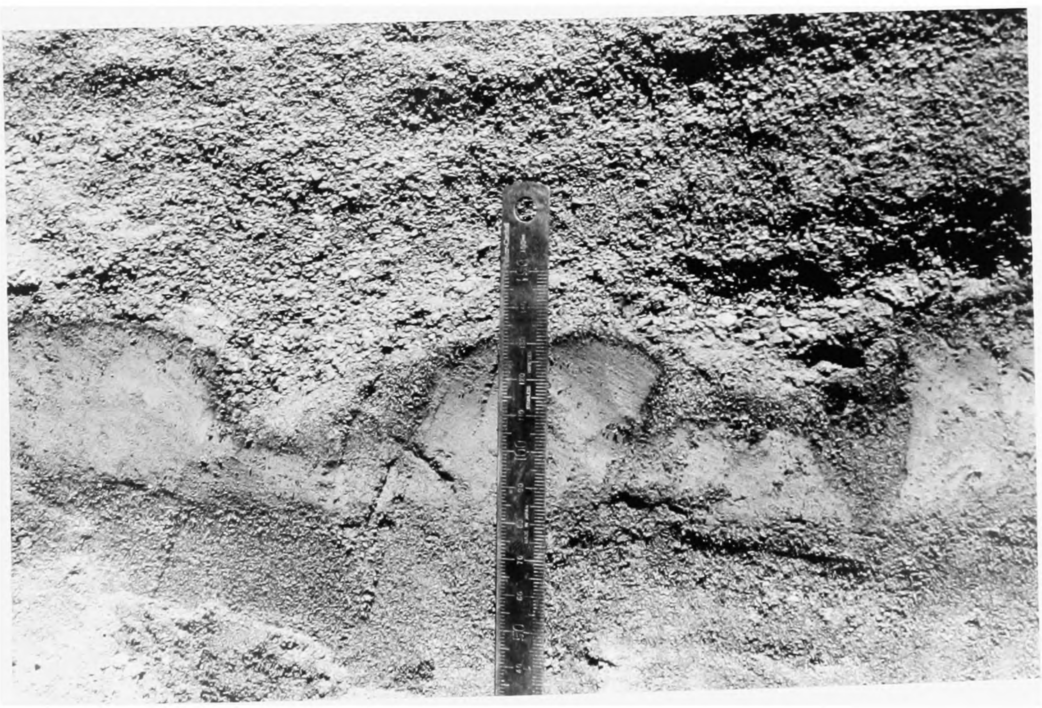


9. Ripple-laminated sand and clay-silt in upper Karewa lacustrine sediments from Pampur.

10. Mottled and finely-laminated clay silt in upper Karewa lacustrine sediments from Burzahom. Note the calcareous concretions in the lowest part of the photograph.



11.Type B ripple-drift cross-lamination in medium sands from upper Karewa sediments, Sambur.



12.Load structures in upper Karewa clay silt at Kahalwan. The overlying unit is coarse sand.

13. Small-scale faults in upper Karewa lacustrine sediments exposed in Badgam Quarry number 6.

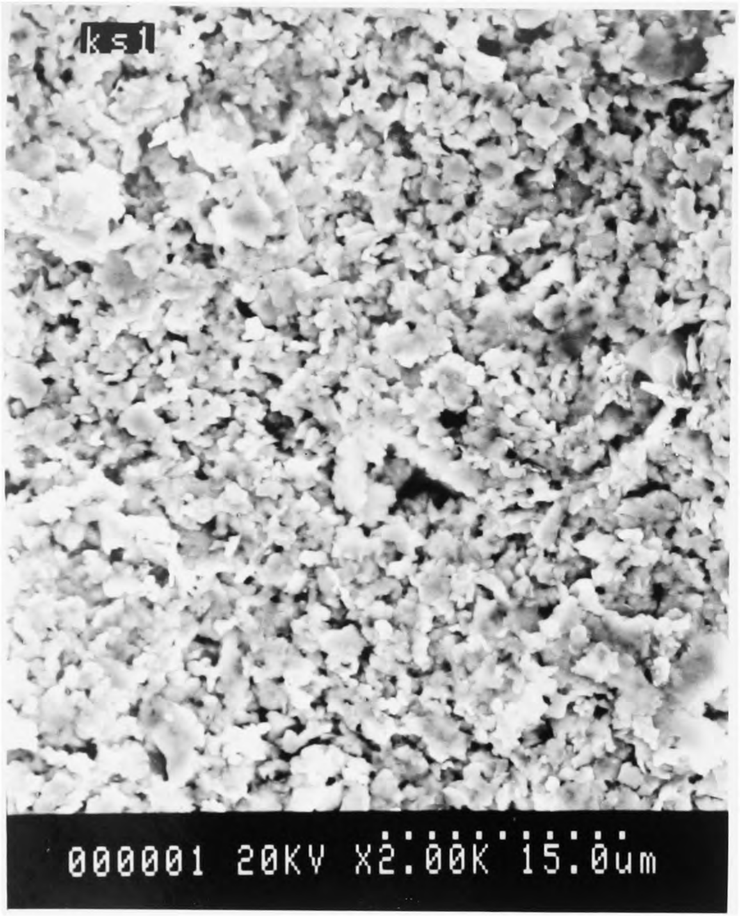


Plate 14. SEM micrograph of weathered loess from Nunar in the Sind Valley (sample KS1). Shows a dispersed fabric of very fine clay. Vertical face.

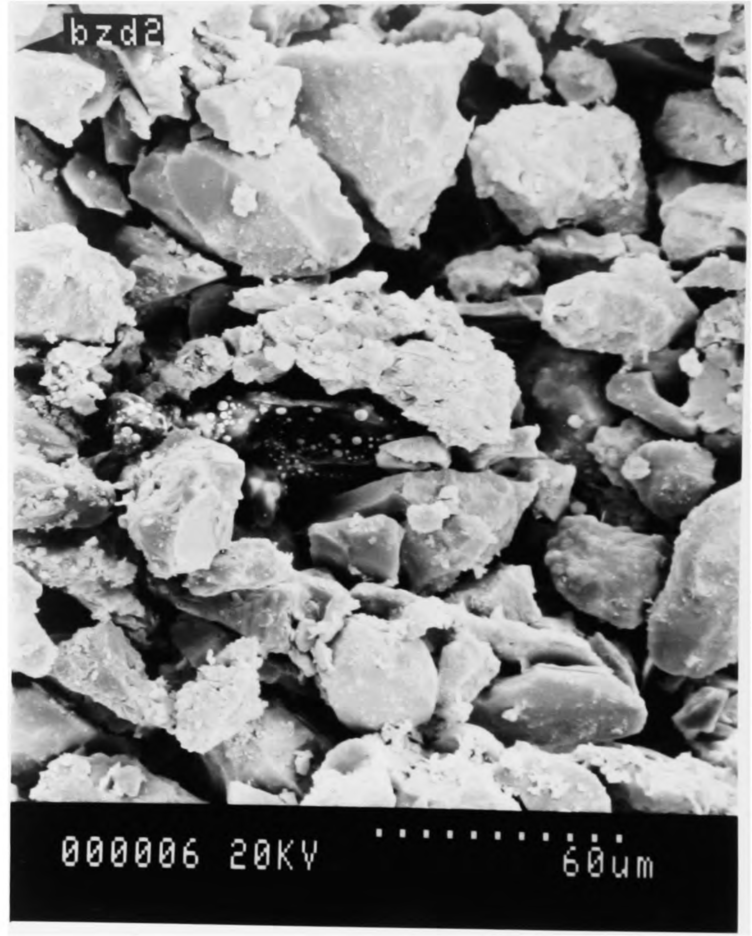


Plate 15. SEM micrograph of upper Karewa lacustrine silts from Burzahom (sample BZ-D2). Shows very clean, angular grains of quartz silt. The fine material is calcium carbonate and clay. Vertical face.

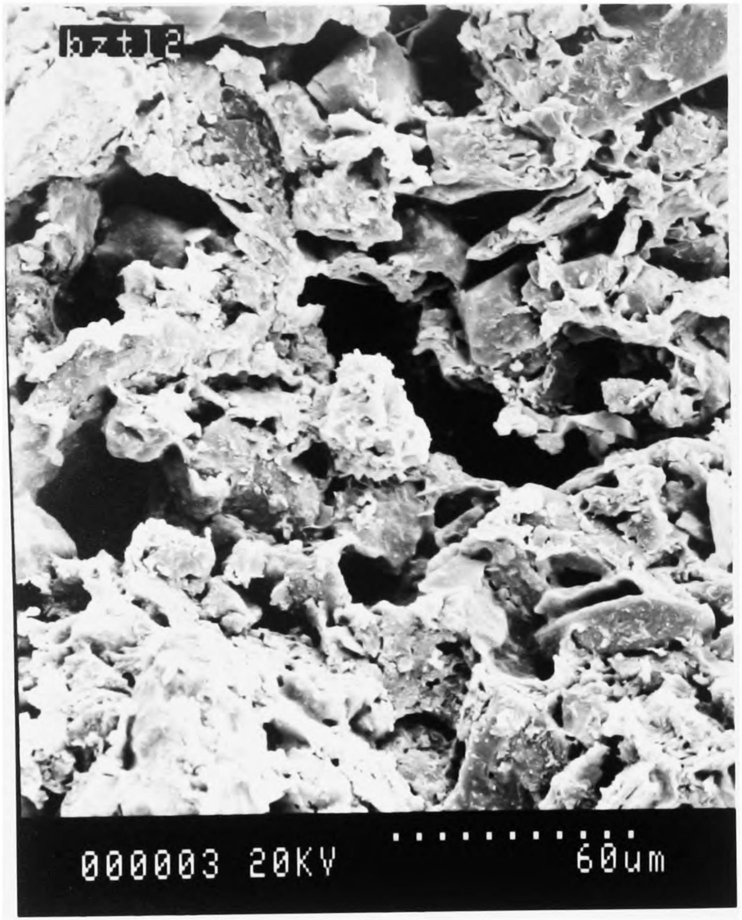


Plate 16. SEM micrograph of upper Karewa lacustrine silts from Burzahom (sample BZ-TL2). Shows angular grains of quartz silt with aggregates of calcium carbonate and clay minerals.

Vertical face.

Plate 17. SEM micrograph of upper Karewa lacustrine silts from Burzahom (sample Bz-TL2). Shows clay particles wrapped around a small tubule.

Vertical face.

Plate 18.

1. Eucypris sp. External lateral view of RV. L=0.77mm (x108).
2. Parastenocypris delormei. Internal lateral view of RV. L=1.01mm (x82).
3. Cypria opthalmica. External lateral view of carapace. L=0.43mm. (x197).
4. Cypridopsis vidua. External lateral view of carapace. L=0.60mm. (x135).
5. Cytherissa lacustris. Internal lateral view of RV. L=0.42mm. (x175).
6. Cytherissa lacustris. External lateral view of RV. L=0.58mm. (x143).
7. Cypridopsis aculeata. External lateral view of RV. L=0.60mm. (x125).
8. Cyclocypris sp. Internal lateral view of RV. L=0.43mm. (x178).
9. Candona sp. External lateral view of LV. L=0.78mm. (x93).

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Plate 19.

1. Ilyocypris gibba. External lateral view of LV. L=0.60mm. (x130).

2. I. gibba. Internal lateral view of RV. L=0.67mm. (x123).

3. I. bradyi. External lateral view of LV. L=0.71mm. (x105).

4. I. bradyi. Internal lateral view of LV. L=0.84mm. (x98).

5. I. kashmirensis. External lateral view of LV. L=0.40mm. (x215).

6. Limnocythere franki. External lateral view of RV. L=0.56mm. (x135).

7. L. franki. Internal lateral view of LV. L=0.57mm. (x130).

8. Darwinula stevonsoni. External lateral view of LV. L=0.64mm. (x128).

9. D. stevonsoni. Internal lateral view of LV. L=0.63mm. (x130).

Plate 20.

1. Candona neglecta. External lateral view of LV. L=1.16mm. (x70).
2. C. neglecta. Internal lateral view of RV. L=1.29mm. (x62).
3. Potamocypris sp. External lateral view of RV. L=0.47mm. (x170).
4. Candona candida. Internal lateral view of RV. L=0.87mm. (x93).
5. C. candida. External lateral view of RV. L=0.66mm. (x108).


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21.Exposure of a debris-flow deposit from the Sind Valley, near to the mouth of Wangat Nala.

22.Exposure of a debris-flow deposit at Mangom, in the Sind Valley.

23. Exposure of a debris-flow deposit at Gund, in the Sind Valley.

24. Exposure of Late Quaternary outwash near Sonamarg, in the

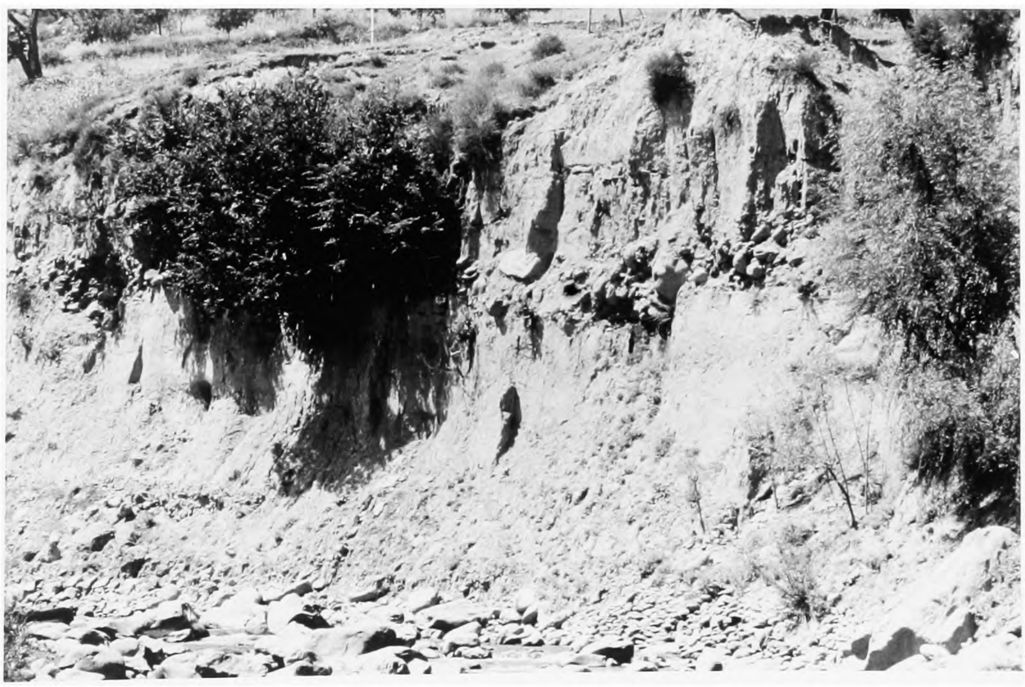
25.Exposure of debris-flow deposits interbedded with loessic colluvium at Ganeshpur, in the Liddar Valley.

26.Exposure of crudely-bedded outwash near Pahalgam, in the Liddar Valley.

27. Exposure of calcareous till at Pahalgam, in the Liddar Valley.

28. Exposure of outwash in the West Liddar Valley, near Pahalgam.

29. Exposure of outwash between Pahalgam and Aru, in the West Liddar Valley.

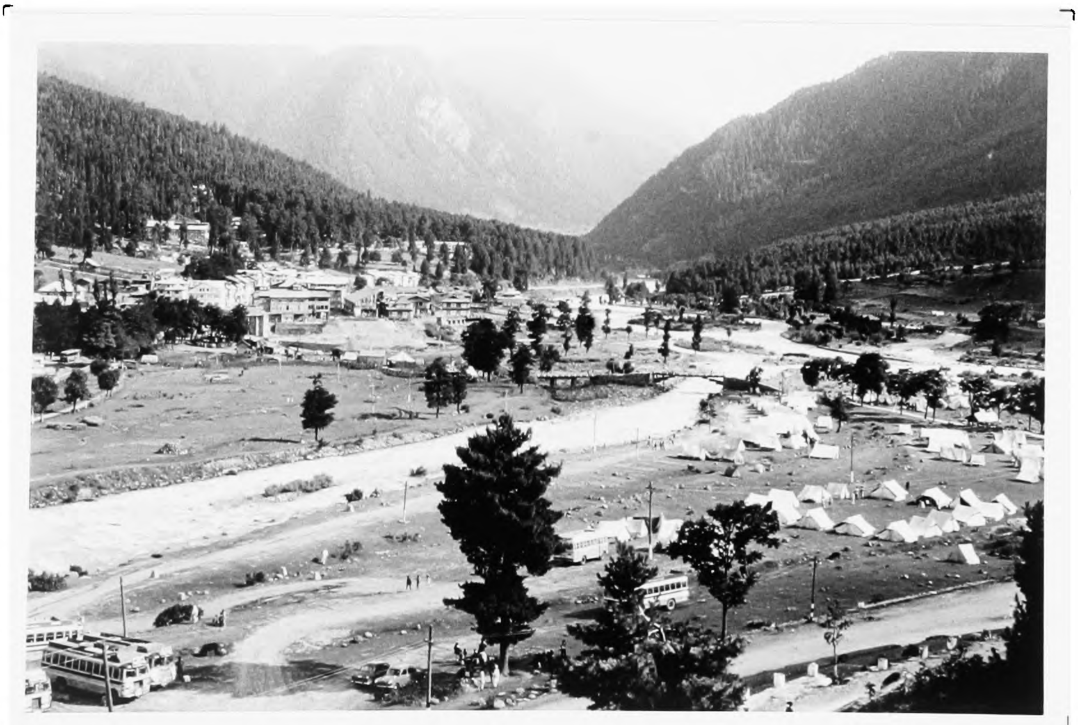


30. Exposure of fluvial conglomerate interbedded with fine-grained lacustrine sediments at Durhom, in the lower Ningle Valley.

31.Exposure of till at Gulmarg.

32.View of the Sonamarg Basin in the Sind Valley, looking east. Outwash terraces above the present floodplain of the river are visible in the foreground. The moraine at Shitkari is visible in the left background.

33.Nested moraines near Shitkari, in the Sind Valley.



34.View of the Pahalgam basin, in the Liddar Valley, looking south.

35.View of Aru, in the West Liddar Valley, looking south-east.

36.Moraine crest at Liddarwat, in the West Liddar valley.

37.Dissected lateral moraines at Basmai, in the West Liddar valley, on the northern flank of the valley.

38.Holocene moraines near the snout of the Kolahoi Glacier, in the west Liddar Valley.

39.Morainic ridges in the Gulmarg Basin.

40.Morainic ridges at Butapathri, in the upper Ningle Valley.