III. HYDROLOGY AND SEDIMENTATION

III.1 ANNUAL WATER YIELD

In analyzing highland-lowland interactions in the Ganges-Brahmaputra river basin, it is obviously of interest to examine what proportion of the flow carried annually by the two main rivers originates in the two major upland units, viz. the old plateaus of relatively low relief in the south and the much steeper and more elevated Himalaya in the north.

Rao (1975) presented annual streamflow totals for India's major rivers, suggesting that 30% of the flow arriving at the Farakka barrage in the Ganges is contributed by the southern tributaries, two-thirds of which is delivered by the Yamuna alone (Figure 1).

An inspection of streamflow records for the major northern tributaries flowing through Nepal (summarized by Alford, 1987) indicates that of the remaining 70% of the flow at Farakka, ca. 45% (i.e. 30% of the total) comes from Nepal.

Similarly, the Himalayan tributaries supply about 63% of the amount of water discharged annually by the Brahmaputra at Bahadurabad in Bangladesh (Figure 1). About half the Himalayan contribution is made up by the Brahmaputra (then called Dihang) itself at the point where it emerges from the mountain range (Sharma, 1985). It can be safely assumed, therefore, that about two-thirds of the total flow of the combined river system originates in the Himalayas and the adjacent plains.

The vast amounts of water that are discharged each year into the Bay of Bengal ultimately derive from parts of the river basin showing important climatic contrasts. As such, major spatial variations in streamflow amounts and timing are to be expected.

At the bottom end of the range one finds such rivers as the Chambal in the far western corner of the basin (140,000 km^2), and the upper Brahmaputra (Tsangpo, 153,200 km^2) in western Tibet, which show runoff totals of only 225 and 200 mm/yr respectively (Rao, 1975; Guan & Chen, 1981), reflecting the arid conditions prevailing over much of their drainage basins.

At the other end of the scale, there are rivers such as the Burhi-Dihing (6000 km^2) in Assam, or the Seti, Chepe and Balephi rivers (300-600 km^2) in the High Himalaya of Central Nepal, all of which have streamflow totals over 2500 mm/yr (Sarma, 1986; Alford, 1988a).

In contrast to the Burhi-Dihing, which is entirely rainfed and of relatively low elevation, these high-altitude Nepalese rivers receive yet unspecified contributions of snowmelt.

Rivers draining the hot southern plateau area, such as the Betwa, Ken and the Son, typically exhibit streamflow values of 350-450 mm/yr (Rao, 1975).

As already indicated, the Himalayan tributaries show a wide variation, reflecting an equally wide range of hydrometeorological conditions. For example, within the Sapt Kosi river basin in eastern Nepal, the discharges of the major tributaries range from 580 mm/yr for the Arun to 1920 mm/yr for the adjacent Tamur (Alford, 1988a).

This can be explained by the fact that the Tamur river originates on the south-facing slopes of the High Himalaya, which experience relatively high precipitation totals (Figure 11). In addition, the Tamur must receive important contributions in the form of meltwater from the glaciers of Mt. Kanchenjunga.

The Arun, on the other hand, has about 90% of its drainage area located north of the main range in the Trans-Himalaya, where both precipitation and snowmelt contributions are modest (Guan & Chen, 1981).

In general, the larger a watershed's proportion in the rainshadow zone north of the Great Himalaya, the lower will be the annual water yield

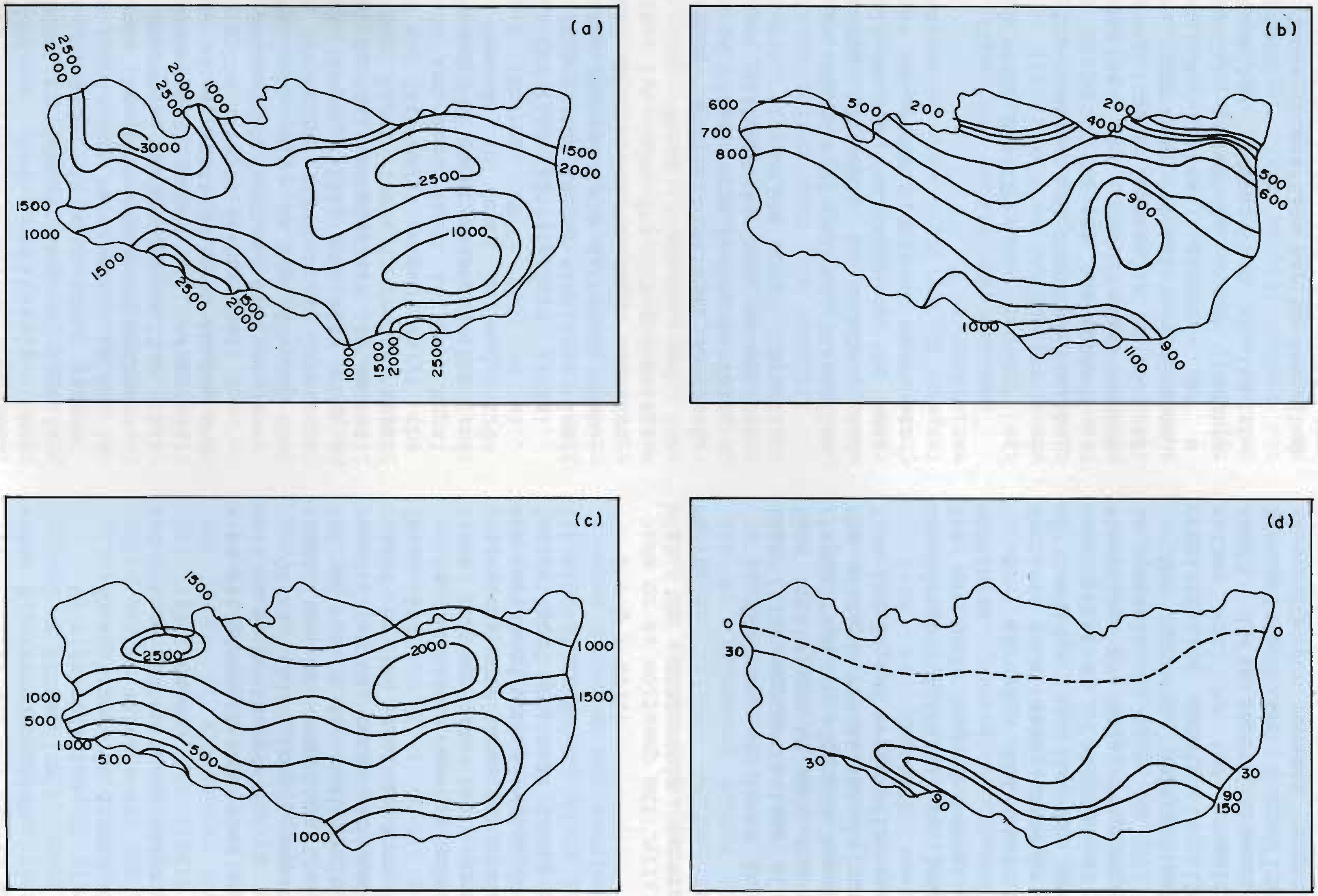

Figure 19. (a) Average precipitation, (b) actual evapotranspiration, (c) surplus, and (d) water deficiency (mm/yr) in the Nepalese part of the Sapt Kosi River Basin (after Subramanyam & Upadhyay, 1983).

(Alford, 1988a).

Another group of rivers originate in the Middle Himalaya. Found at intermediate elevations (mostly between 500 and 2500 m a.s.l.), these rivers experience a strong monsoonal influence and relatively high evaporative demands and no significant snowmelt. As such their streamflow totals may be expected to lie in between the above-mentioned extremes.

However, variations within this physiographic zone are especially large as a result of rain-shadow and orographic effects (Figure 11).

Geological factors, such as leaky river beds (e.g. in the Siwalik-Dun zone (Chyurlia, 1984) or in limestone areas (Rawat, 1985) may also play a role. Typical examples are the Bagmati and East Rapti rivers in Central Nepal (850 and 1640 mm/yr respectively), the Sarada and West Rapti in West Nepal (750 and 690 mm/yr respectively), and the Bemunda in Tehri Garhwal (1200 mm/yr) (Sharma, 1977; Puri et al., 1982; Chyurlia, 1984; Alford, 1988a).

Each of the larger Himalayan river systems represents a different combination of the three major environments, i.e. the dry Trans-Himalaya, the icy Great Himalaya and the monsoonal Middle Himalaya. As such, each of these large basins exhibits a unique, although to some extent predictable, streamflow total (Chyurlia, 1984).

An analysis of regional streamflow patterns in Nepal is underway and should become available in the near future (D. Judge, personal communication).

III.2 WATER BUDGETS

The spatial and temporal variability of water in any environment is determined by the relationships between inputs (generally as rain or snow), storages (in soils, geological deposits, lakes, glaciers or a seasonal snowpack), and outputs (streamflow, evaporation to the atmosphere).

Their interaction can be described conveniently with the help of the water budget equation, which is essentially a continuity equation (Ward, 1975):

$$P = Q + ET \pm \delta S \pm L \quad (1)$$

when P = precipitation
Q = streamflow
ET = evaporation
δS = changes in stored amounts of water/snow
and L = subterranean gains or losses,

with all elements expressed as mm/time unit (e.g. month, year).

Often this basic equation is simplified by computing the balance for an entire seasonal cycle ("water year"), which tends to reduce the change in the storage term to zero (Ward, 1975).

In addition, careful selection of river gauging sites with respect to geological conditions may minimize the sub-terranean term as well. For such a "watertight" drainage basin equation (1) simplifies to

$$P = Q + ET \quad (2)$$

with the components expressed as mm/yr.

It follows, at least in theory, that first approximations of streamflow totals can be made for ungauged rivers from an adequate knowledge of the spatial variations in precipitation inputs and evaporation outputs.

Such an approach meets with considerable difficulties, however, in the extremely rugged terrain of the Himalayas.

As already indicated in the sections on climate, there is a dearth of knowledge regarding the amounts of precipitation falling at higher elevations and about evaporation rates in general.

The general underestimation of precipitation inputs is illustrated rather well by the unrealistically small differences between annual precipitation and streamflow totals (i.e. ET) for several large Nepalese river basins as computed by Chyurlia

(1984).

According to these figures, the areally weighted average evaporation over the Sapt Kosi basin (57,200 km^2) would amount to only 105 mm/yr. Corresponding values for the Kali Gandaki (34,440 km^2) and Karnali (19,260 km^2) basins read 135 and 190 mm/yr respectively (Chyurlia, 1984).

More plausible, but essentially untested, are the results of the water balance computations by Subramanyam & Upadhyay (1983) and Alford (1987) for mountainous catchments in Nepal. The former applied the classical Thornthwaite & Mather (1957) approach to compute evaporation rates from temperature data and compared these with precipitation data to define zones of water surplus and deficiency in the Sapt Kosi basin (Figure 19).

Alford (1987) took the procedure one step further by manipulating the precipitation and evaporation relationships with altitude in such a way, that the computed streamflow outputs (cf. equation 2) for his study basin - the Seti Khola near Pokhara in West Nepal - matched the measured amounts of flow.

Naturally, the question as to what extent this apparent success in simulating streamflow totals, using a relatively simple procedure, is the result of a fortuitous combination of variables, remains unanswered. This would involve measurements of the various components of the water budget for carefully selected catchment areas in the respective physiographic zones.

One Himalayan catchment, for which such data have been published, is the 1754-ha Bemunda watershed, situated between 800 and 2200 m in the "Middle Hills" of Tehri Garhwal (Puri et al., 1982). Over the period June 1981 to May 1982, some 1210 mm of streamflow were recorded, against a rainfall total of 2235 mm. This would imply an evaporation figure of 1025 mm/yr (including δS) for this largely forested catchment.

Unfortunately, Puri et al. (1982) considered their streamflow figures underestimates, because of yet unknown amount of groundwater flow underneath their measuring flume. Clearly, the study of water budgets in the Himalayan context has only been touched upon until now and much more work is urgently needed if we are to increase our understanding of the hydrological behaviour of this environment (Kattelmann, 1987).

More information in this respect is available for the rivers draining the southern plateau area. Annual totals of precipitation and streamflow, as well as evaporation increase as one goes from west (Chambal river) to east (Son, Godavari).

Typical values for ET range from 530 mm/yr in the west (mean rainfall 755 mm/yr; Jha et al., 1988) to 890 mm/yr in the south-east (mean rainfall 1185 mm/yr; Biksham & Subramanian, 1988).

III.3 SEASONAL VARIATIONS IN STREAMFLOW

In a strongly seasonal climate like that of the Ganges-Brahmaputra river basin, monthly variations in streamflow assume great importance. Mean monthly flows for several basins of intermediate size, considered representative for the major physiographic zones within the Himalaya described earlier, are given in Figure 20.

Although all four basins exhibit maximum flows during the summer months, when both rainfall inputs and rates of snowmelt are at a maximum, there are marked differences with respect to the amplitudes in seasonal variations.

Absolute variations are small in the Tibetan example (Figure 20a), where streamflow is the result of a combination of groundwater outflow from alluvial deposits (50%) and modest contributions of meltwater (20%) and rainfall (30%) in the summer months (Guan & Chen, 1981).

The groundwater contribution typically decreases in spring, when neither melt nor rainfall replenishes the system (Figure 20a). Incidently, no such decline in flow occurs in larger Tibetan basins, reflecting the

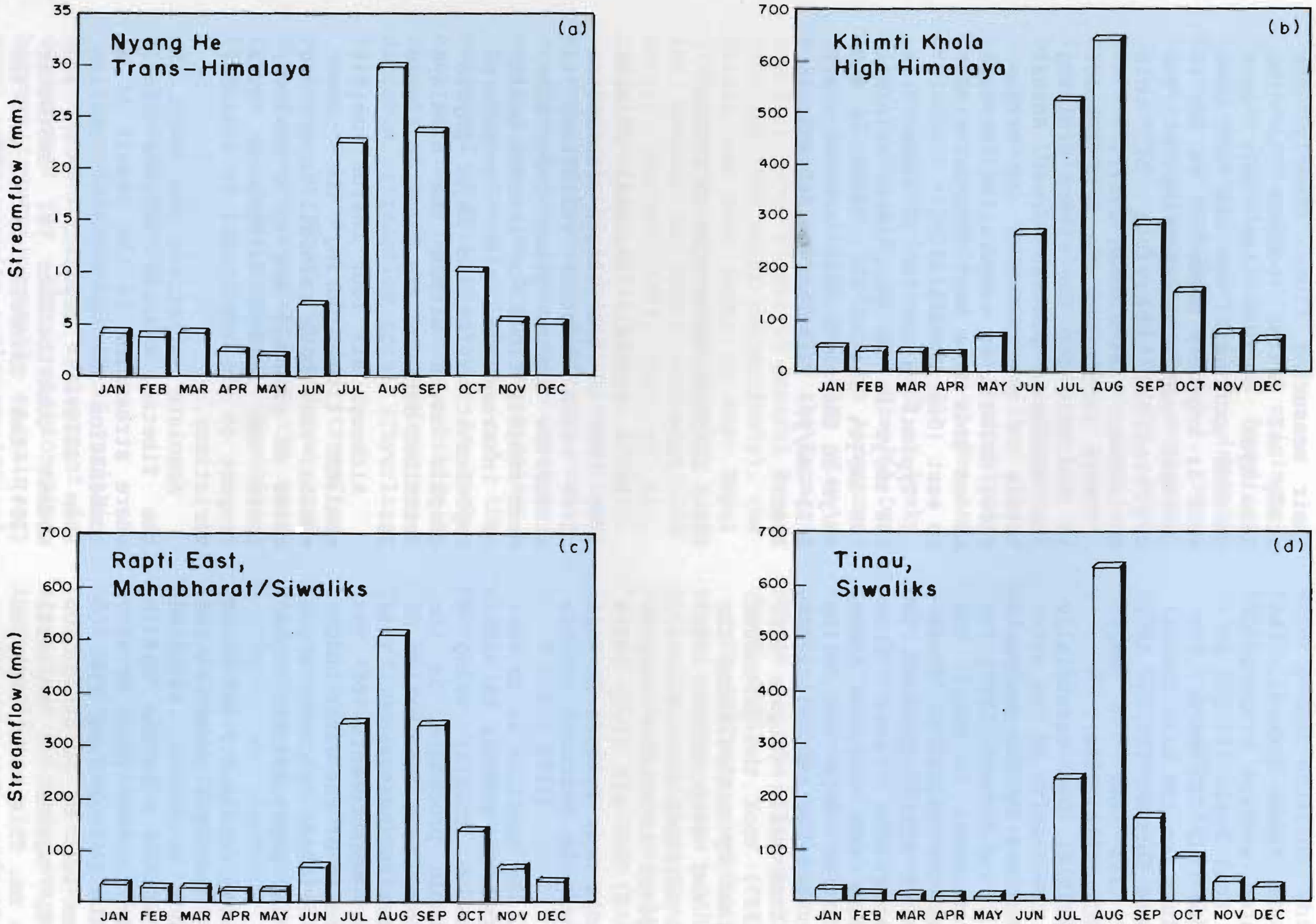

Figure 20 **Monthly streamflow (mm) for selected Himalayan rivers (based on data presented by Guan & Chen, 1981; Surface Water Records of Nepal; Sharma, 1977).**

much higher groundwater reserves in large river valleys (Guan & Chen, 1981).

Since the Nyang He watershed is about ten times as large as the other catchments in Figure 20, strictly speaking its flow regime cannot be compared directly. Presumably, springtime recession would almost certainly have been steeper and the ratio between maximum and minimum flows correspondingly larger.

The flow regime of the Khimti Khola, which rises not far from the source of the Nyang He on the south-facing slopes of the main range, seems much more extreme at first sight (Figure 20b).

This is mainly a matter of the amounts of water involved as the ratios between maximum and minimum flows are quite comparable for the two basins (18 vs. 16 respectively).

Indeed, the High Himalayan zone appears to be the richest area of water (Alford, 1987) as a result of a combination of factors. These include the full exposure to moist air of the summer monsoon, relatively low evaporation rates associated with high elevations (Subrahmanyam & Upadhyay, 1983), and sustained meltwater contributions in spring (Figure 20b). This moderating effect of snowmelt contributions on low flows in the pre-monsoon months is lacking in the case of streams originating in the Middle Hills and the Siwaliks (Figure 20c, d).

In addition, rainfall is more intense here as compared to the High Himalaya (Chyurlia, 1984). Therefore, the ratios between maximum (Q_{max}) and minimum (Q_{min}) monthly flows tend to increase substantially under these conditions, i.e. the streams become more "flashy". This is especially so for streams rising in the southern Mahabharat-Siwalik zone, where strong rainfall leads to high summer flows, whilst the highly permeable nature of the valley fills favours leakage into the river beds, thus further reducing baseflows in winter and spring (Figure 20d; Plate 7 & 16).

Ratios between Q_{max} and Q_{min} are also seen to increase as one goes from east to west within a physiographic zone, reflecting the corresponding trend in rainfall seasonality (Sharma, 1977).

A semi-quantitative interpretation of the flow regimes of several large Nepalese river basins with respect to the origin of the water was undertaken by Chyurlia (1984).

He distinguished four major flow components, viz. inflow of groundwater from valley deposits (baseflow Q_b), meltwater (effectively between February and August only, and peaking in June: Q_s), "interflow" (Q_i) and "direct flow" (Q_d) (Figure 21).

In this view, "interflow" is considered to represent the portion of the precipitation, that is temporarily held in storage in the soil mantle before being released to the streams. "Direct flow", on the other hand, enters the stream directly (presumably during rainstorms) without being stored.

Although Chyurlia (1984) himself considered the separation between Q_i and Q_d "a rather arbitrary affair in the absence of detailed basin studies", it would seem as though this contention applies to the rest of the separation lines in Figure 21 as well.

For example, a glance at Figure 22, depicting the seasonal course of meltwater discharge from a glaciated basin in the Khumbu area of East Nepal, reveals that melt is peaking in July and August, rather than becoming negligible in August, as suggested by Figure 21.

This difference in timing of the snowmelt season (February-August vs. May-November) may partly be due to the fact that the Khumbu study reflects meltwater from a glacier rather than from a seasonal snowpack at lesser elevations. A considerable portion of the flow now assigned to "interflow" in Figure 21 may thus in reality be due to meltwater contributions.

Also, it is difficult to envisage how hillside vegetation would survive a four-to-five month long dry period after "the interflow storage is exhausted in December and only baseflow accounts for runoff (Chyurlia, 1984).

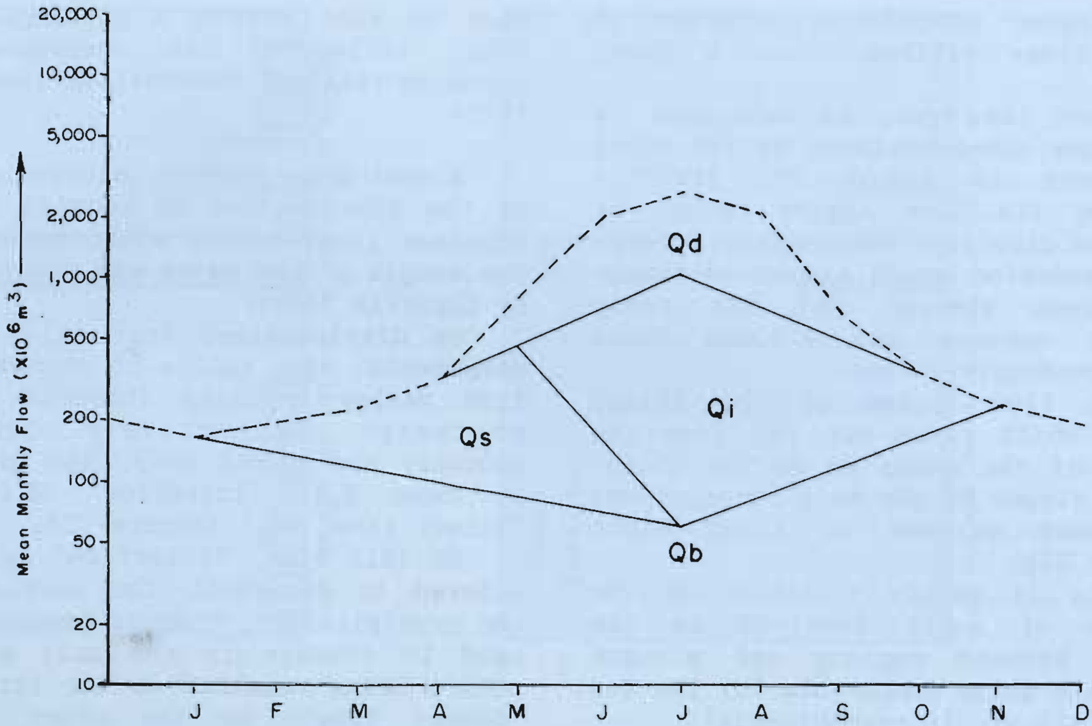

Figure 21 **Hydrograph separation for the Seti river, West Nepal (after Chyurlia, 1984).**

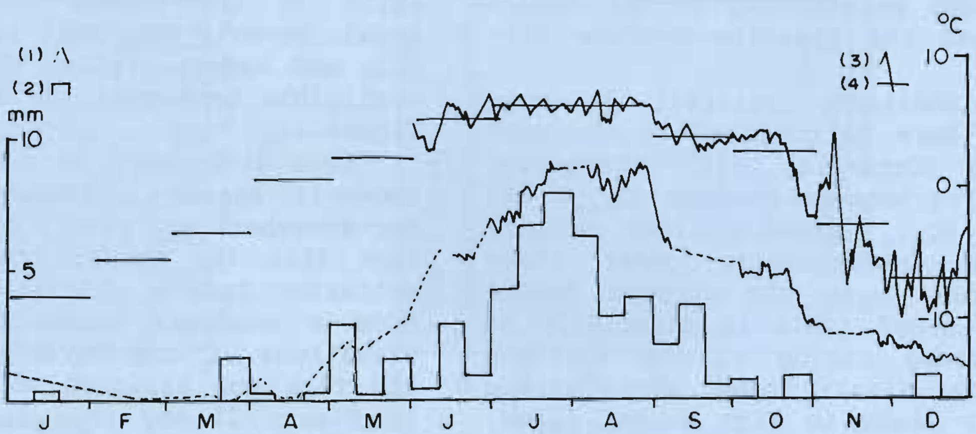

Figure 22 **Water discharge (1) of the Imja Khola at Dingboche, precipitation (2) and air temperature (3) and (4) at Lhajung (eastern Nepal), 1974/75 (after Higuchi et al., 1976).**

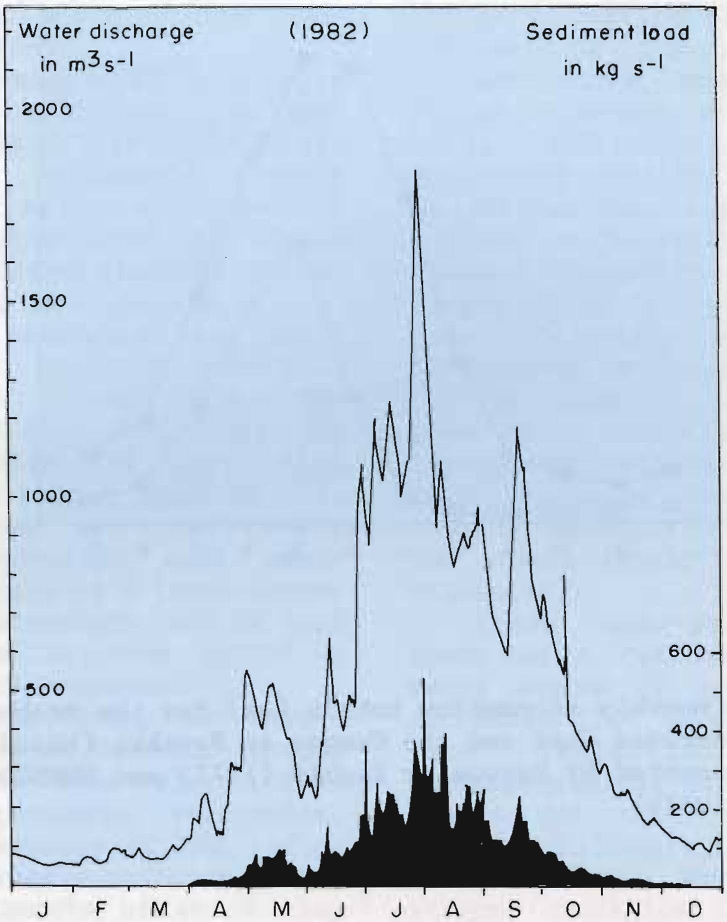

Figure 23 **Water discharge (m^3/sec) of the Burhi Dihing river, Assam, in 1982 (after Sarma, 1986).**

Rather, the saturated zones around (smaller) streams in the valleys, which supply the baseflow, should be seen as being fed by continuous (although steadily diminishing after the monsoon) "interflow" from the hillslopes (Hewlett & Hibbert, 1963; Figure 26).

The subject of flood-generating "direct flow" will be discussed in more detail in the next Section.

Whilst the Himalayan rivers, whose regimes were illustrated in Figure 20, carry 69% (Nyang He) to 83% (Tinau) of their annual flow during the summer monsoon months (June through September), this proportion is even higher for the southern rivers.

The Yamuna (346,000 km^2), for example, discharges 91% of its annual streamflow total into the Ganges during the monsoon. The rest is more or less equally distributed over the winter months (October-February) and the pre-monsoonal period (March-May).

Even more extreme in this respect are the Sind (93% flowing during the monsoon), the Chambal (95%) and the Betwa (97%) rivers (Jha et al., 1988).

A less extreme picture emerges for the Burhi-Dihing in Assam (Figure 23). Here, the contribution to total flow by streamflow in the spring months is comparatively greater as a result of an earlier start of the rains (Sarma, 1986; Figure 7).

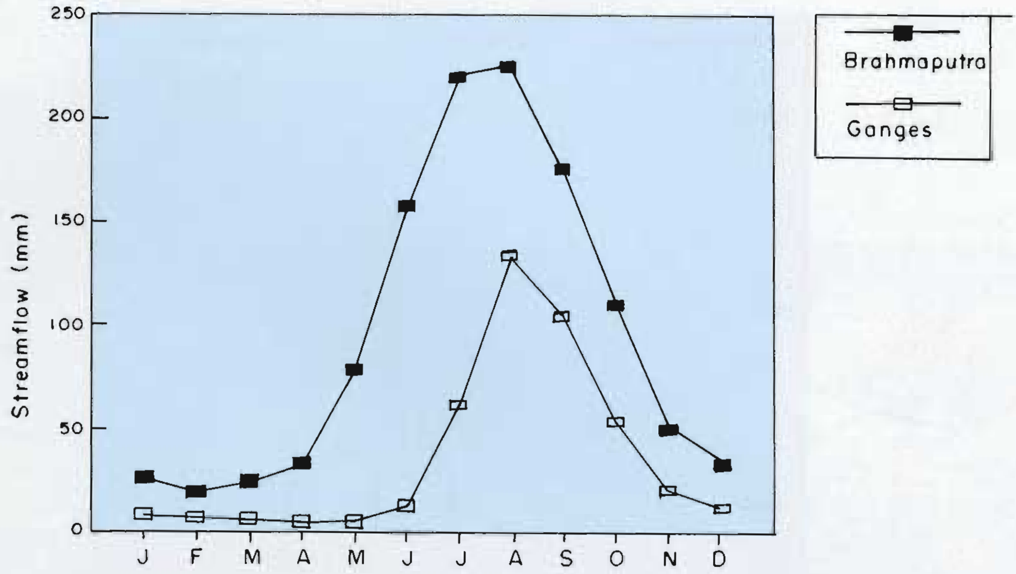

Figure 24 **Average monthly streamflow totals (mm) for the Brahmaputra at Bahadurabad Ghat and the Ganges at Farakka (based on data presented by Haroun er Rashid (1977) and UNESCO (1971) respectively).**

Naturally, all this has a bearing on the flow regimes of the Ganges and Brahmaputra rivers themselves (Figure 24). The Brahmaputra is seen to carry much more water per area unit than the Ganges (1155 vs. 430 mm/yr), reflecting the wetter conditions prevailing in the eastern part of the combined river basin (Figures 8 and 9).

Furthermore, the Ganges exhibits minimum flow rates in May, just before the onset of the monsoon, and peaks in August and September.

Conversely, the Brahmaputra has a minimum in February and starts to rise in March in response to the arrival of the pre-monsoon rains, culminating in July and August (Figure 24).

As such, August has always been the month in which widespread flooding in Bangladesh is most likely to occur (Haroun-er-Rashid, 1977). According to the same author (referring to the situation prior to the building of the barrage at Farakka), floods in the delta occurring between May and July are usually due to the Brahmaputra-Jamuna itself, and between August and October due to the combined flow with the Ganges.

The issue of peakflows will be considered further in the next section on runoff response to rainfall.

III.4 HILLSLOPE HYDROLOGICAL RESPONSE TO RAINFALL

When examining how peakflow rates in the Ganges-Brahmaputra River basin are generated, it is helpful to distinguish three scales, viz. a macro-, a meso- and a micro-scale (Ives et al., 1987). At the macro-scale one looks at the interaction between Highland and Lowland systems, e.g. the Himalayan block vs. the Indo-Gangetic depression.

At the intermediate scale a major tributary river basin, e.g. the Gandaki or the Tista, is considered. Such meso-scale catchments usually represent an array of environmental conditions, ranging from the alpine heights of the Great Himalaya to the tropical hills of the Siwaliks.

Micro-scale watersheds are small enough to fall entirely within a single specific zone, e.g. the Siwaliks, the Middle Hills, or the arid Trans-Himalaya. As such, they lack significant biotic zonation (cf. Figure 17) and are relatively homogeneous. Obviously, each meso-scale basin consists of a large number of micro-scale watersheds, and in turn the macro-system is made up of a number of meso-scale basins.

Similarly, the timing and magnitude of streamflow of the macro-rivers will reflect the flow distribution between the meso-scale watersheds, just like the response of the latter reflects the aggregate response of the micro-scale catchments of which they are comprised.

Rain falling on a hillslope may reach the adjacent stream channel in several ways (Figure 25a). If rainfall intensities exceed the capacity of the top-soil to absorb all rainfall, then the surplus runs off as "infiltration excess" or "Hortonian" overland flow (Horton, 1933; flow path Q(o) in Figure 25a).

This type of flow is rarely observed under forested conditions, but may occur after disturbance and degradation of the soil, as well as in sparsely vegetated areas (Dunne, 1978).

The remainder infiltrates into the soil profile and, depending on vertical and lateral hydraulic conductivities, soil moisture patterns and slope steepness, may take one of several routes to the stream (Figure 25a).

In the (relatively rare) case of deep, permeable and uniform deposits, the water will tend to travel vertically downward to the zone of saturation and from there onwards will follow a curving path to the stream (flowpath Qg in Figure 25a).

More often, however, soil permeability decreases with depth (Gilmour et al., 1987). Part of the water then percolates vertically until it meets an obstruction, such as a clayey B-horizon or bedrock. It is then deflected laterally and usually referred to as "interflow" or "throughflow" (Kirkby & Chorley, 1967; flowpath Qt in Figure 25a).

The bulk of this "throughflow" generally travels rather slowly through the soil matrix, feeding saturated sections around the streams, thereby maintaining the baseflow of the stream (Hewlett & Hibbert, 1963; Figure 26).

These near-saturated riparian zones in a catchment may act as a major source of stormflow, with the mechanism that produces this "quickflow" reflecting the prevailing geomorphological setting (Ward, 1984).

In the case of the situation depicted in Figure 25, i.e. a concave-shaped slope bottom bordering a stream, quickflow is often generated through the formation of what has been called a "riparian groundwater ridge" (Ward, 1984). During rainstorms the groundwater table, already relatively close to the surface in these riparian zones, rises rapidly due to throughflow contributions from upslope and direct rainfall on the spot, and may reach the surface (Figure 25b). Any rain falling on the now saturated area is unable to infiltrate and runs off along the surface together with the emerging throughflow ("return flow") as "saturation overland flow" (Flowpath Qo(s) in Figure 25).

If the process continues long enough, the entire foot slope may become saturated with water (Figure 25c) and become a source area for storm runoff. Since in such situations a considerable part of the quickflow is due to saturation overland flow, stormflow patterns in the stream will immediately reflect differences in rainfall intensity as they occur (Dunne, 1978).

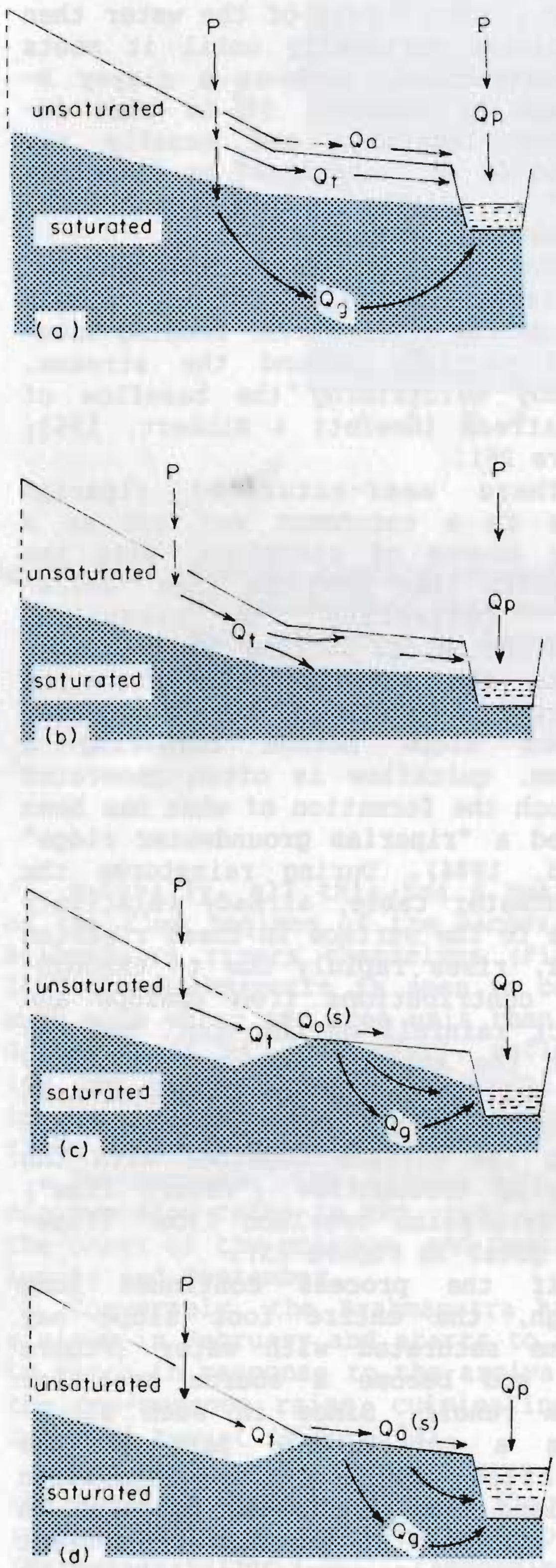

Figure 25.

(a) Flow paths of the sources of streamflow: Q_p is direct precipitation onto the water, Q_o is overland flow, Q_t is throughflow and Q_g is groundwater flow.

(b)-(d) The response of streamflow to precipitation in humid headwater areas: an integrated view (after Ward, 1984).

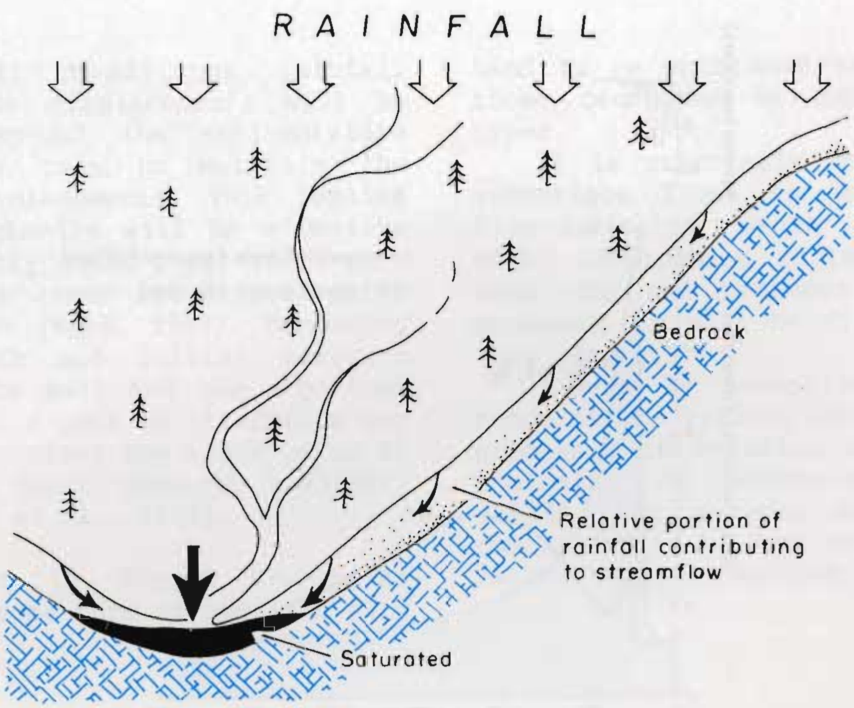

Figure 26 The relative contributions of rainfall to streamflow (after Ward, 1984; based on an original diagram by J.D. Hewlett, 1961).

However, where deep permeable soils overlie impermeable bedrock and where steep hillslopes border a narrow flood plain, there will be little scope for the generation of saturation overland flow, either in the valley bottoms or on the hillsides themselves.

To explain the immediacy of streamflow response in areas without appreciable overland flow of any type, it has been suggested that part of the throughflow travels through the upper soil horizons at a speed high enough to reach the stream channel during the storm (hence the frequently used term "subsurface stormflow").

Decayed root channels, animal burrows and other "macropores" have been advocated as conducts for such rapid throughflow (Whipkey, 1965; Mosley, 1982a). Whereas this may be true in certain cases, e.g. where subsurface "pipes" have developed (Jones, 1981), rates of water movement through soils are generally far too slow to enable "new" rainfall to reach the stream during a storm event (Dunne, 1978).

Therefore, Hewlett & Hibbert (1967) introduced the concept of "translatory flow", a "push-through" mechanism whereby each new addition of rainfall to a hillside displaces an approximately equivalent amount of "old" water, thus causing the oldest water to exit from the bottom end of the system and into the stream (cf. Figure 26).

Recent work by Pearce et al. (1986) and Sklash et al. (1986) has shown the overriding inportance of the "push through" mechanism in areas of steep slopes, permeable soils and high rainfall.

Naturally, an equivalent displacement of stored soil water by new rainfall can only be expected if the moisture storage capacity of the hillslope soil mantle is already filled or near full.

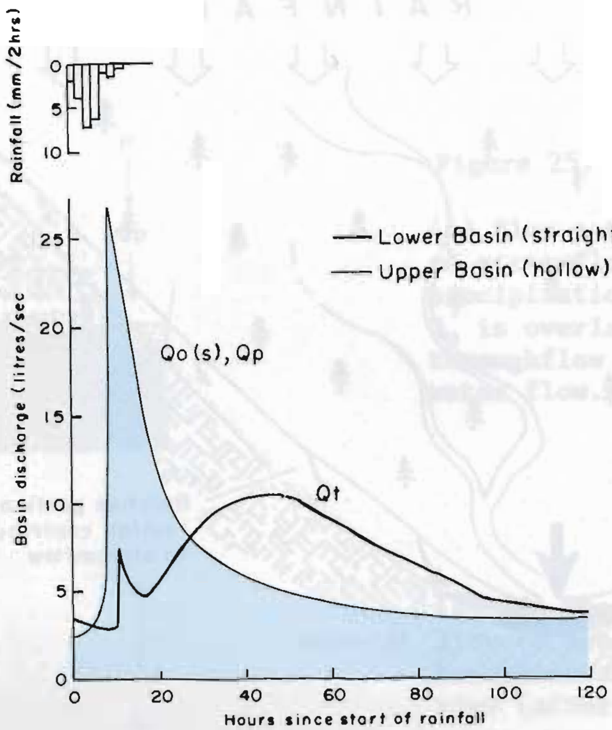

Figure 27 **Storm-runoff hydrographs from two areas with contrasting topography within the East-Twin basin (0.2 km^2), United Kingdom (after Calver et al., 1972).**

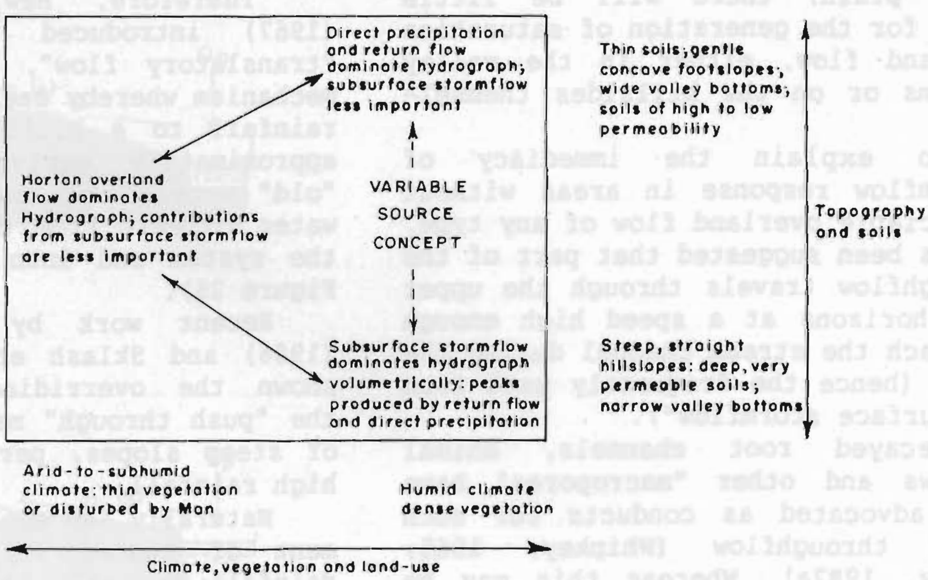

Figure 28 **Schematic representation of the occurrence of various runoff processes in relation to their major controls (after Dunne, 1978). "Direct precipitation and return flow" equivalent to saturation overland flow.**

In drier conditions rainfall inputs and/or displacements will be used to "top-up" the soil-moisture store rather than to maintain the chain of displacements. This implies that the mechanism will be effective most frequently after a period of rain and/or on the lower and moister parts of the slopes (Ward, 1984). Depending on the depth and initial moisture status of the soil and the magnitude of the storm, a peak in streamflow may occur shortly after the storm or up to several days later (Hewlett & Nutter, 1970; Sklash et al., 1986).

As shown in Figure 27, peaks produced by some form of overland flow tend to be much more pronounced than those generated by sub-surface flow types.

It is especially this shift from subsurface flow- to Horton overland flow-dominated storm runoff, that often accompanies certain changes in land use and produces an array of problems, as will be discussed in the next chapter.

Figure 28 summarizes the occurrence of the various runoff generating processes in relation to their major controls. As emphasized by Dunne (1978), the various modes of storm runoff should be seen as complementary rather than contradictory.

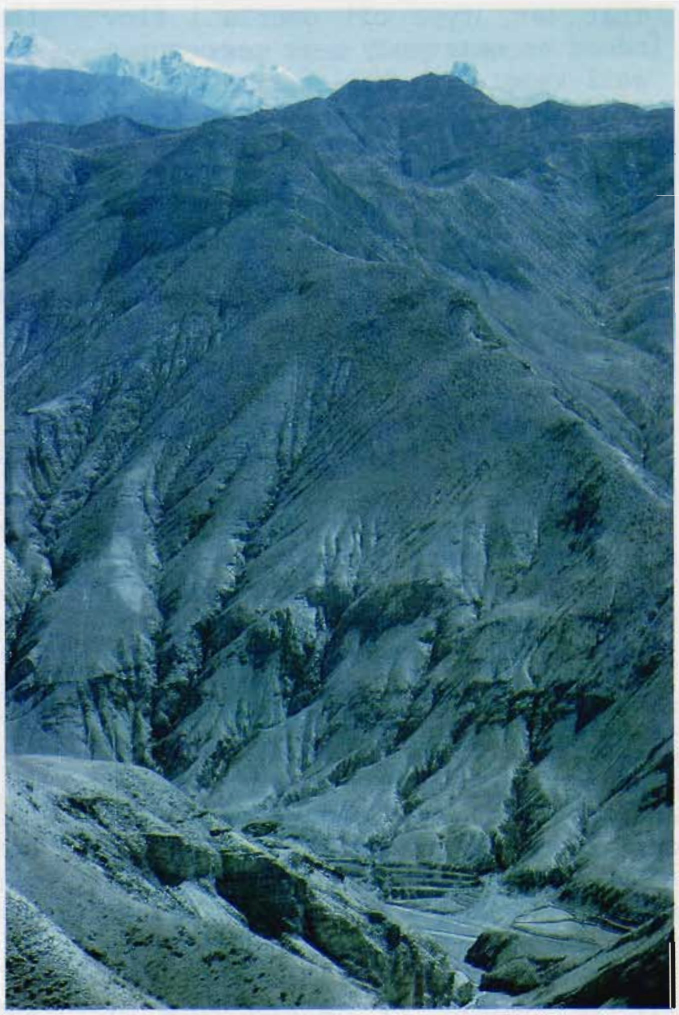

Plate 13

Rill and gully erosion on slopes underlain by shales in Dolpo, West Nepal (photograph by P. Laban).

How does all this translate to the Ganges-Brahmaputra region, and to the Himalaya in particular?

Leaving disturbed situations for the moment aside, one can make a general distinction between the dry Trans-Himalayan zone and the more humid parts of the mountains.

In the former, infiltration excess overland flow will dominate runoff peaks (Plate 13), whereas in the latter sub-surface stormflow (translatory flow) will be the major component of storm runoff (Figure 28; Pandey et al., 1983/4).

In view of the often very long and steep slopes and the relatively narrow valley bottoms (Plate 14), the contributions made by saturation overland flow from the riparian zone will be modest at best, with the exception of some poorly drained parts of the terai and perhaps the Dun valleys.

That (any type of) overland flow is indeed an extremely rare phenomenon on well-vegetated hillsides in the Middle Himalaya, was demonstrated convincingly by the work of Pandey et al. (1984) and Gilmour et al. (1987).

The former measured very low amounts of overland flow during two consecutive rainy seasons in a range of forest types in the Kumaon Himalayas, whilst the latter predicted low frequencies of occurrence of overland flow on the basis of rainfall intensity and soil hydraulic conductivity data for several sites in Central Nepal.

Interestingly, both groups of investigators observed relatively low rainfall intensities for most of the time (Table 3). In addition, the majority of storms was reported to be of relatively small magnitude (< 20 mm). Therefore, soil infiltration characteristics must have degraded considerably before appreciable quantities of overland flow could occur regularly under the rainfall regime prevailing in the Middle Himalaya (Pandey et al., 1983/4; Gilmour et al., 1987).

This aspect will be discussed more fully in Chapter IV on the role of vegetation and land-use.

Rainfall intensities are generally higher at lower elevations (Chyurlia, 1984) and as such the frequency of overland flow may be expected to increase for such areas as the Siwaliks.

Figure 29 illustrates the average intensity-duration curves for rainfall in Kathmandu (Middle Hills) and Dehradun (Indian foothills zone). Rainfall intensities are consistently higher at Dehradun, also for return intervals longer than a year (not shown).

Near-saturated top-soil permeabilities (Ksat) for forest soils from the two areas (Gilmour et al., 1987; Patnaik & Virdi, 1962) have been added for comparison (Figure 29). Based on these limited data, chances for the occurrence of overland flow would seem higher in the Indian Dun and Siwalik zone than in the Middle Hills of Nepal.

Grazing is a common practice in these forests and the herb/grass layer is often absent (Sastry & Narayana, 1984). According to Patnaik & Virdi (1962), top-soil infiltration capacities of these poorly structured soils were easily reduced by 50 % in case of compaction by grazing animals. Yet the value quoted in Figure 29 pertained to forest "with good leaf litter". Narayana & Sastry (1983) reported even lower values for grazed forests in the valley-bottom part of Doon valley, where soils are somewhat heavier. Carson (personal communication) observed widespread overland flow on Siwalik hillsides in Nepal during the monsoon.

Subba Rao et al. (1985) presented storm hydrographs for a densely forested (but grazed) headwater catchment in the Siwalik hills, which definitely suggest stormflow to be dominated by some sort of overland flow (Figure 30; cf. Figure 27).

These investigators also reported, that runoff would not be produced before the area had become thoroughly wetted up thoroughly, whilst flows would stop soon after the withdrawal of the monsoon in September.

In the absence of information on sub-soil permeabilities and groundwater levels in these catchments, it

Plate 14 Tributary valley of the Marsyangdi river near Tal, West-Central Nepal, with steep slopes and narrow valley bottom. Note the enormous boulders in the centre.

(a) Annual number of days with certain 5-minute rainfall intensities as recorded at Kathmandu, June - September, 1971-1979 (after Gilmour et al., 1987).
(b) Frequency distribution of rainfall intensity in different vegetation zones in the Kumaon Himalaya, June - September, 1981-1982 (after Pandey et al., 1984).

Rainfall intensity class ($mm\ hr^{-1}$)	No. of days
0- 9.99	26.2
10- 19.99	9.8
20- 29.99	6.7
30- 39.99	4.9
40- 49.99	2.3
50- 59.99	0.8
60- 69.99	1.6
70- 79.99	0.7
80- 89.99	0.3
90- 99.99	0.6
100-109.99	0.1
110-119.99	0.0
120-129.99	0.4

Intensity classes (mm/ 30 min.)	Sites (% of events)		
	sal forest	pine forest	mixed-oak, Rianj-dominated forest
< 2	28.0	34.7	38.5
2—4	36.0	27.5	29.2
4.1—6	27.0	29.0	18.7
6.1—8	4.9	4.1	10.4
8.1—10	2.5	4.1	3.1
> 10	1.2	0	0
Total number of timed events	81	98	96

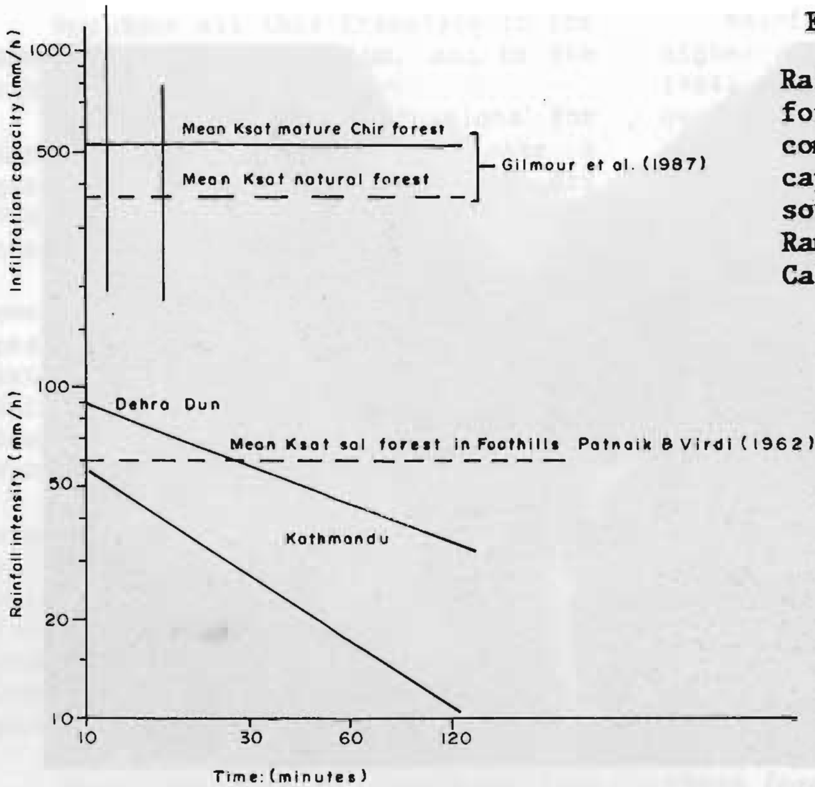

Figure 29.

Rainfall-intensity-durations for Dehradun and Kathmandu as compared with infiltration capacities of local forest soils. Rainfall data after Ram Babu et al., (1978) and Caine & Mool (1982).

is difficult to decide whether the overland flow involved is of the saturation or infiltrationexcess type. More detailed process-oriented work is needed to shed more light on these matters.

Other investigations of the hydrological response to rainfall of small catchment areas in the Himalayas are virtually limited to the studies carried out by the Central Soil & Water Research & Training Institute, Dehradun, in the valleys and foothills around Dehradun and Chandigarh (Narayana, 1987). Limited data have also been reported for the 1754-ha Bemunda catchment in the "Middle Hills" of Tehri Garhwal (Puri et al., 1982).

No information has been published for Nepal in this respect, although a catchment study has been initiated recently by the Topographic Survey Branch in the Dhulikhel area (P.B. Shah, personal communication; cf. Plate 4).

Average monsoonal runoff percentages (June-September 1976-1983) for two forested catchments underlain by deep and well-drained alluvial deposits in the Dun area ranged from 12% of incident rainfall (70-ha basin) to 19% (4.4 ha) according to data extracted from Annual Reports of CSWCRTI. Maximum (but still modest) runoff percentages were recorded for the month of August at the height of the monsoon (17 and 27% respectively).

The streams draining these flat basins are not perennial, suggesting that considerable amounts of infiltrated rainfall may contribute to deep drainage rather than to any type of lateral flow. In view of the very low relief of these basins (2-6%), however, it can not be excluded that valley-bottom saturation overland flow contributions to stormflow (Figure 25c) are significant as well.

These catchments are part of a larger experiment in which the hydrological effects of several conservation techniques are investigated (Ram Babu & Narayana, 1982; Sastry & Narayana, 1984).

Since the results of these experiments are often automatically assumed to be applicable to "the"

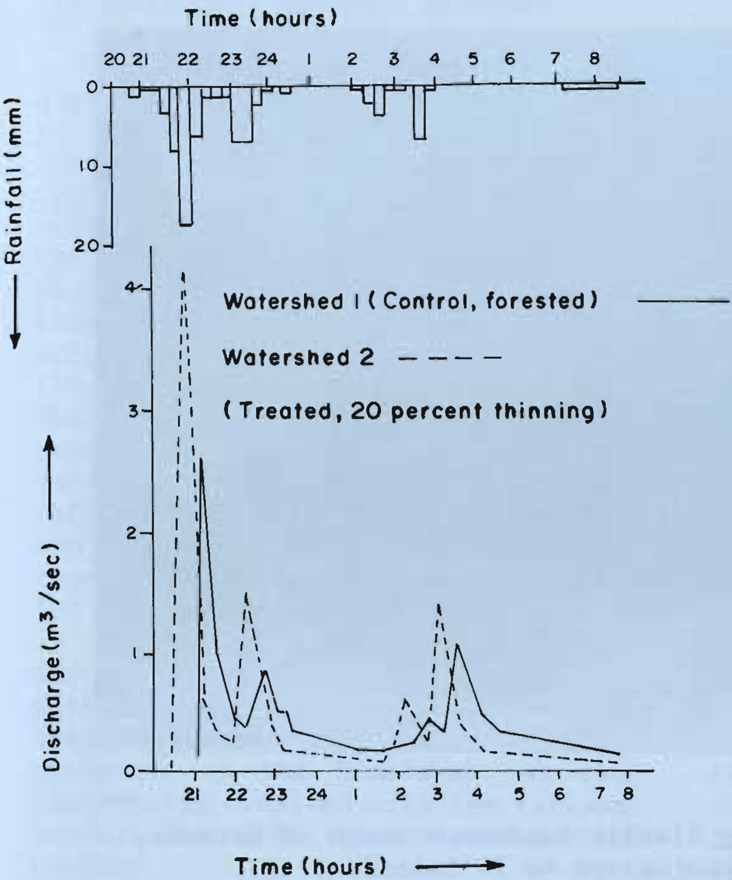

Figure 30.

Runoff response of two small forested headwater catchments in the Indian Siwaliks to 78 mm of rain (after Subba Rao et al., 1985).

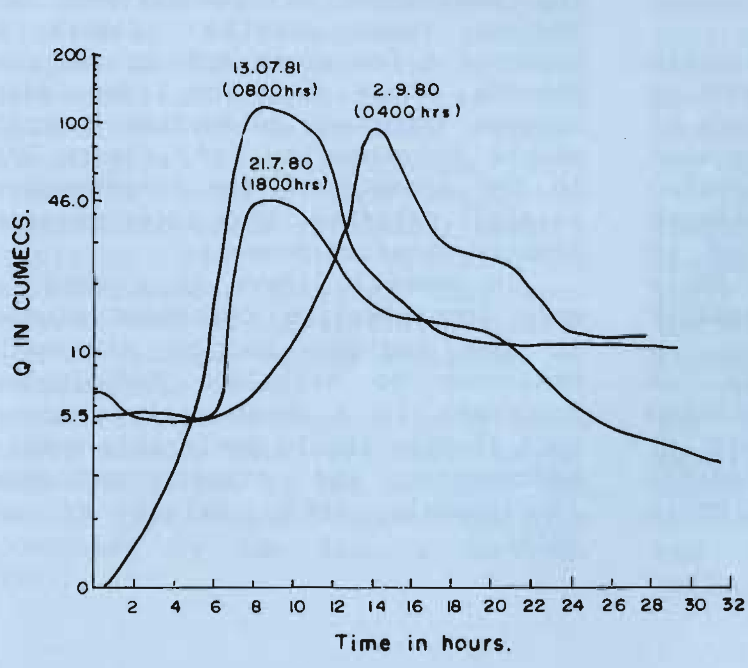

Figure 31.

Typical stormflow hydrographs for the pine-forested Bemunda catchment, Tehri Garhwal (after Puri et al., 1982).

Plate 15 Actively eroding Siwalik landscape south of Hetauda, Central Nepal (photograph by P. Laban).

Himalaya (e.g. Ghosh & Subba Rao, 1979; Bahadur et al., 1980; Narayana, 1987), it is important to remember that these catchments are by no means representative for any part of the Himalaya other than Doon Valley and similar areas of low relief (e.g. the Chitwan valley, etc.).

Limited data are also available for a very small (1.3 ha) catchment in the Siwaliks near Chandigarh (Gupta et al., 1974, 1975). In this somewhat steeper (25 %) area, which was covered with scrub, grasses and scattered trees, stormflow totals amounted to only 6 % of incident rainfall. Soils were deep and sandy, and presumably quite permeable (Plate 15).

As for the Middle Himalaya, it would seem as though the forested Bemunda catchment could be more or less representative of the overall hydrological conditions prevailing in the steeper parts of this zone.

According to Puri et al. (1982), streamflow rates during the rainiest months constituted 70-80% of incident rainfall, and on average 44% during the non-monsoonal months. Such high runoff coefficients for the rainy season could be seen as support for the importance of translatory flow during these months. Figure 31 presents a few storm hydrographs from Bemunda, whose recession limbs indeed suggest important subsurface contributions to stormflow (cf. Figure 27). In the absence of the corresponding rainfall patterns, this interpretation remains untested however.

In general, there is a need for more comprehensive catchment studies in the region, linking streamflow behaviour to hillslope hydrological processes in a quantitative manner. Such studies should preferably combine hydrometric and tracer techniques (Bruijnzeel, 1983; Sklash et al. 1986).

III.5 FLOODS: REGIONAL PATTERNS

In the absence of in-depth studies, several attempts have been made to predict catchment response (peakflow rate, stormflow volume) to rainfall in the (Lesser) Himalaya using empirical methods such as the rational formula (Meijerink, 1974; Chyurlia, 1984). The latter is essentially based on the concept that storm runoff is generated by infiltration excess overland flow (cf. Figure 25a). The formula consists of the product of catchment area, rainfall intensity and a "runoff coefficient" reflecting the lumped effects of soil and topographic characteristics, vegetation and anything not accounted for (Chow, 1964).

It will be clear from the above discussion of hillslope hydrological patterns that reality is more complex than envisaged by Horton (1933), certainly in the Himalayas. As such the results obtained with the rational formula should be treated with extreme caution.

Meijerink (1974) claimed to have obtained good results with the method for catchments up to 20 km^2 in the homogeneous Siwalik area around Dehradun. However, his method for assigning numerical values to the runoff coefficients associated with various types of soils, topography and land-use remained rather obscure. Also, when catchments situated outside the Siwaliks were included, results were less satisfactory, possibly because of variations in rainfall intensities. Nevertheless the calibration of Meijerink's "photo-hydrological" interpretation with streamflow results from small- to intermediate sized catchments seems a promising route to hydrological regionalization, provided that more reliable data on the spatial distribution of rainfall become available for the Himalayas.

An attempt at regionalizing storm runoff amounts from very small catchments (< 100 ha) in the foothill- and Siwalik area around Dehradun was also undertaken by Ram Babu & Narayana (1982).

Using long-term rainfall-runoff records for five catchments from Doon valley, ranging in size between 4 and 83 hectares, they calibrated a multiple regression model having as variables catchment area, channel length, storm duration and magnitude, maximum 30-minute rainfall intensity and soil wetness. The resulting coefficient of determination (R^2) was a modest 0.52, perhaps reflecting the fact that differences in landuse types had not explicitly been taken into account. Also, application of the model by the present writers to a small catchment in the Siwalik zone near Chandigarh (Gupta et al., 1974) produced a serious overestimation. This suggests that this "regional" model should not be used outside the context of the flat Doon valley, even though it was developed to predict stormflow characteristics for the region at large (Ram Babu & Narayana, 1982).

Chyurlia (1984) encountered similar difficulties when analyzing streamflow behaviour of larger (> 1000 km^2) catchments in Nepal. Although there were clear differences in amounts of "direct runoff" (cf. Figure 21) produced by the various basins, it was not possible for him to interpret these directly in terms of geological and rainfall factors.

Chyurlia circumvented the problem by examining the seasonal variation in runoff coefficients, which are largely determined by rainfall patterns. Statistically significant relationships were found to exist between monthly runoff coefficients and amounts of rain in the preceding month for each basin (Figure 32).

As expected, the highest runoff coefficients were observed for the months of August and September, when soils have been wetted thoroughly by the monsoon rains, thereby diminishing storage opportunities.

Interestingly, the constant and the coefficient of the regression equations appeared to correlate significantly with the percentages of catchment area occupied by the Middle and High Mountains physiographic regions (cf. Figure 4), at least for the 42.890 km^2 Karnali basin. Equally interesting was Chyurlia's finding that the maximum observed direct runoff coefficients only correlated significantly with percentage area in the Middle Mountains region (full data set).

In other words, the Middle Mountains physiographic region seems to have response characteristics to rainfall favouring greater quantities of direct runoff than do other physiographic regions, notably the High Mountains and, to a lesser extent, the Siwaliks (Chyurlia, 1984).

Whether this reflects higher rainfall totals or reduced infiltration characteristics is difficult to tell, since the analysis concerned river basins of considerable size (1000-57,000 km^2), with wide variations in geology and rainfall (Chapter II).

Chyurlia himself concluded from a principal component analysis of the entire data set that in the absence of more detailed catchment studies the total annual rainfall and percentage catchment area in the Middle Mountains were equally important and apparently inseparable.

Summarizing, our (quantitative) knowledge of storm-runoff generation in small and intermediate-sized catchments in the Himalaya is limited. As pointed out by Chyurlia (1984), the study of spatial variations in runoff coefficients and their dependence on rainfall and other factors, constitutes an area of research that should be given high priority.

The role of vegetation and land use in this respect will be discussed in Section IV.1.3 and should certainly not be underestimated at this scale.

The timing and distribution of rainfall becomes even more important with respect to the occurrence of peakflows at the meso- and macro-scale levels than in smaller catchments.

Clearly, stormflow generated by heavy rainfall in one part of a large basin will be "diluted" by lower flows from other parts of the basin receiving less or no rainfall at that time. Therefore, although the ***average*** seasonal flow patterns for the major Himalayan rivers are quite similar (Sharma, 1977), the timing of ***actual*** peak discharges on one river does not necessarily coincide with that on other tributaries.

Alford (1988b) conducted a preliminary peakflow analysis for the Ganges at Farakka and all major Nepalese tributaries as recorded between 1967 and 1976. He was unable to detect any correlation between the occurrence of maximum flows at Farakka and those for the Himalayan rivers. More importantly, there was also no correlation in this respect between the Himalayan rivers themselves. Alford (1988b) concluded that "there are no uniform 'wet' and 'dry' years dominating the entire region", or even Nepal. Although his analysis was preliminary and covered only one decade, it reached essentially the same conclusion as that of Mooley & Parthasarathy (1983), who studied regional rainfall records over the period 1871-1980 (see Section II.3.2).

In other words, spatial variations in rainfall patterns seem to override any other factor in the generation of floods at this scale, both directly and indirectly (see below). As such, extreme streamflow events must reflect the occurrence of too much rain falling persistently over a large area.

As shown earlier in Figure 13, large parts of the Ganges-Brahmaputra River basin may receive substantial amounts of rain in a single day. Figure 33 shows an example from Assam, where an area of almost 4000 km^2 received at least 300 mm of rain in two days, with the central 500 km^2 of the storm receiving more than 600 mm. The total area enclosed by the 100 mm isohyet amounted to more than 22.000 km^2 (Raghavendra, 1982). Similar examples, both for the upper Brahmaputra and Ganges areas, can be found in Pal & Bagchi (1975), Pant et al. (1970), and in Dhar et al. (1982b).

It will be clear from the above that the source of flood generation is not restricted to extreme rainfall in

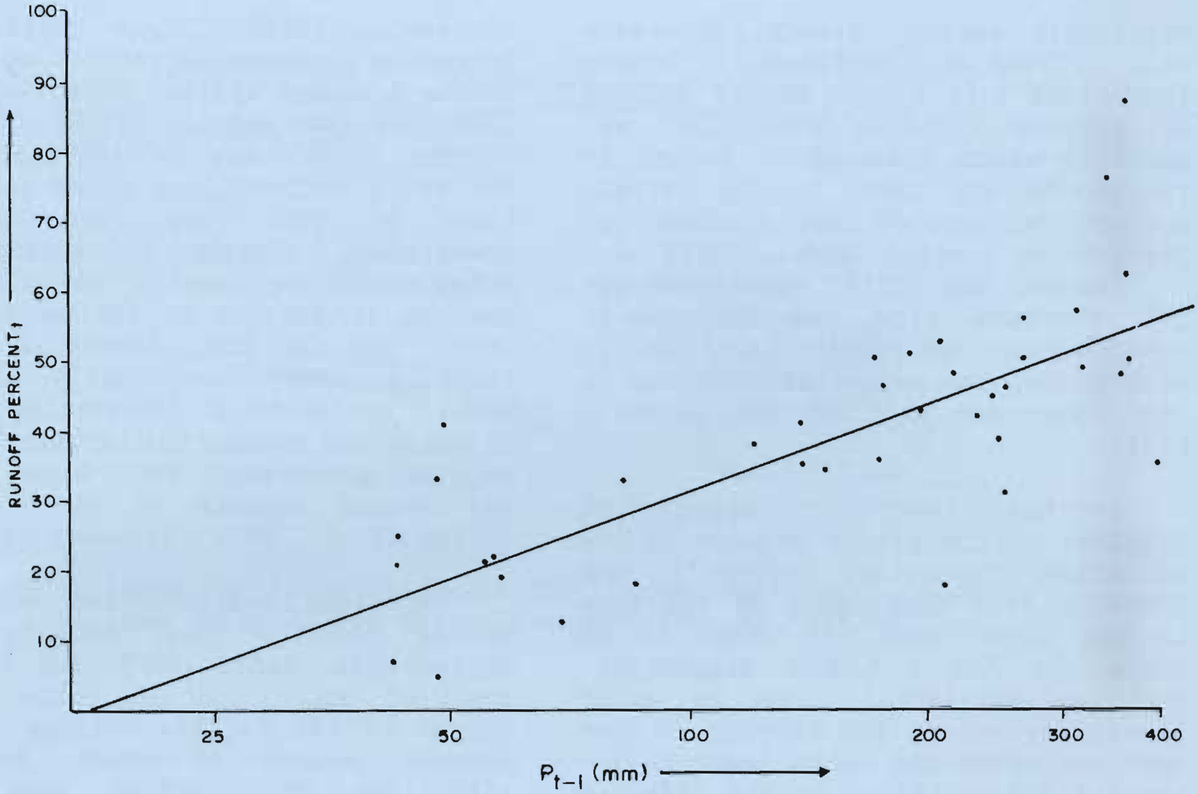

Figure 32 **Semi-logarithmic relationship between percentage of direct runoff and antecedent monthly rainfall in Nepal (after Chyurlia, 1984).**

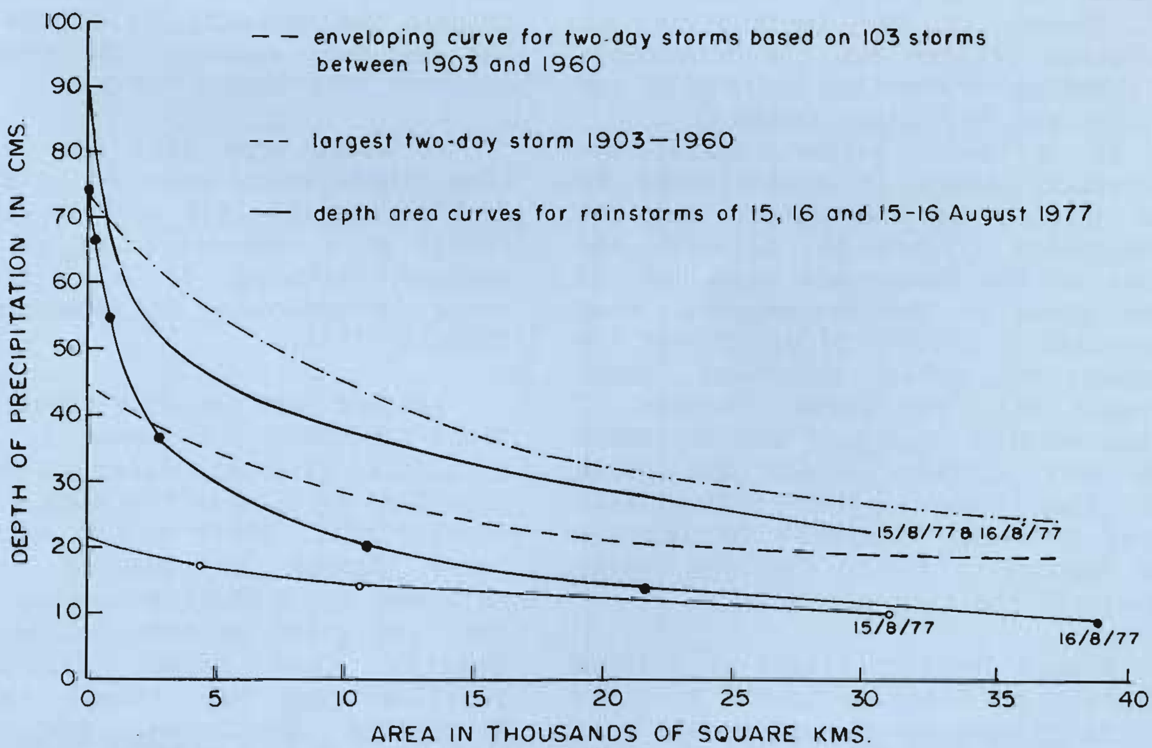

Figure 33 **Depth-area curves for extreme rainfalls in the Assam valley (after Raghavendra, 1982).**

previously wetted uplands. Excessive rain falling on the flood (!) plains themselves (cf. Figure 13) is capable of causing flooding overnight, especially since groundwater tables in the plains are close to the surface during the monsoon and storage opportunities limited (Rudra, 1979).

Indeed, Sen (1979) considered local overland flow and groundwater contributions extremely important in determining the height of flooding in the downstream part of the Gangetic plain.

Another important aspect of flooding in the plains relates to the so-called "backwater effect": the blocking at a confluence of the flow in one river when the other is in spate (or has a higher discharge). Such an artificial rise in water levels may extend for 100-150 km from the confluence and often leads to increased sedimentation in the affected river stretch (Figure 34; Anonymous, 1983).

The effect is illustrated in Figure 35 for a part of the Ganges near Patna, where water levels are raised by 6 to 8 metres due to the joining of the mighty Gandaki river (cf. Figure 1). The implications of backwater effects for the occurrence of flooding in the flat terrain of the plains need no further comment.

At a much larger scale the backwater effect is demonstrated by the joining of the Ganges and the Brahmaputra (Figure 1). Although the flows on the Ganges are much smaller than those on the Brahmaputra, when expressed as a layer of water over the respective total catchment areas (Figure 24), the actual volumes of water meeting in August and September are very similar (Figure 36; Plate 16). Also, there is the effect of rain water becoming "trapped" on land in the floodplain due to the high water levels in the main rivers.

A more indirect effect of extreme rainfall on flooding levels consists of the triggering of large-scale mass movement, especially when earth tremors occur during such wet spells (Goswami, 1985). These landslides introduce enormous amounts of sediment to the drainage system (Carson et al., 1986; see also section III.6).

The effect may be two-fold: (1) the extra sediment may cause aggradation of the river bed further downstream, thereby increasing the water level in general and flooding hazards in particular (Pal & Bagchi, 1975), and (2) the largest of these landslides may temporarily dam a river, producing a devastating surge of water and sediment after the debris barrier gives way. Such flood waves may attain heights of 15 to 20 m (Singh et al., 1974; Mahmood, 1987).

Increased sedimentation may also be the result of river training works or barrages. Rudra (1979) for example reported deposition of sediment upstream of the Farakka barrage in the Ganges, because of which the possibilities of flooding have been increased in various parts of Bihar.

Similarly, the relatively silt-free water released from the barrage tries to recover its lost load by eroding the river bed downstream of Farakka. This material is in turn deposited further downstream and hampers the navigability of the river and may even endanger the future of Calcutta Port (Rudra, 1979).

It would seem that the devastating floods which occurred in Bangladesh during the 1988 monsoon were the result of a combination of the above factors, although it will probably never be possible to separate them (Rogers, 1988).

Another and equally destructive flood-producing phenomenon is formed by glacial lake outbursts, which occur from time to time in the High Himalaya (Ives, 1986). These sudden bursts of lakes dammed by glacier ice or moraines may produce volumes of water that are often an order of magnitude greater than normal rain-derived peakflows and may travel tens of kilometres downstream, transporting colossal amounts of debris (Galay, 1987; Plate 17). Unfortunately, their

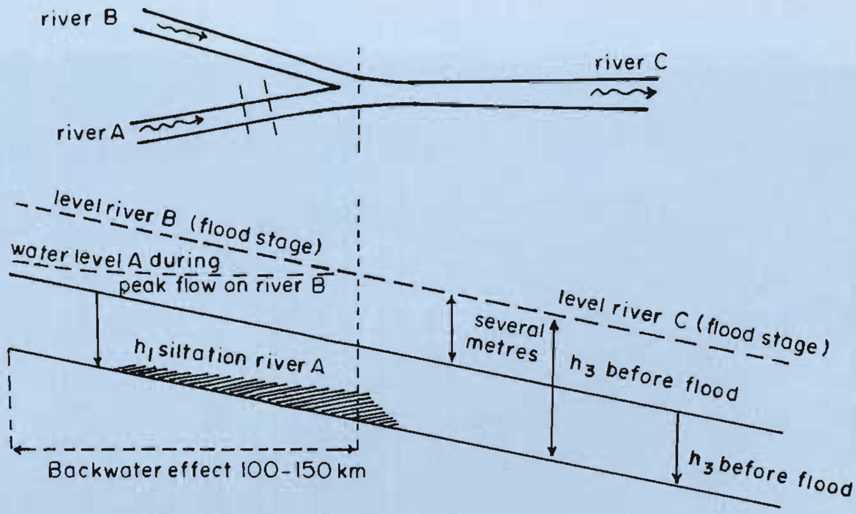

Figure 34 Schematic illustration of the "backwater effect" (modified from Dutch Inland Water Transport Mission, 1982-83).

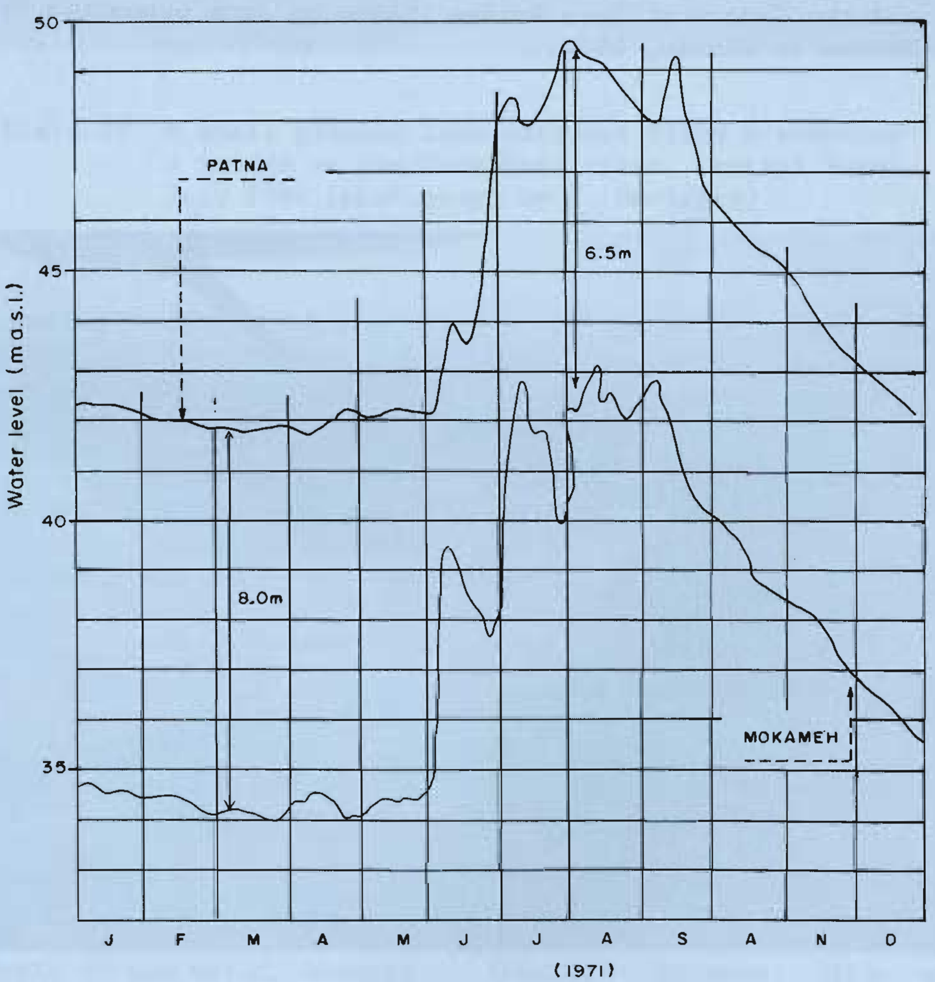

Figure 35 Example of the "back water effect" on the Ganges near Patna (after Dutch Inland Water Transport Mission, 1982-83).

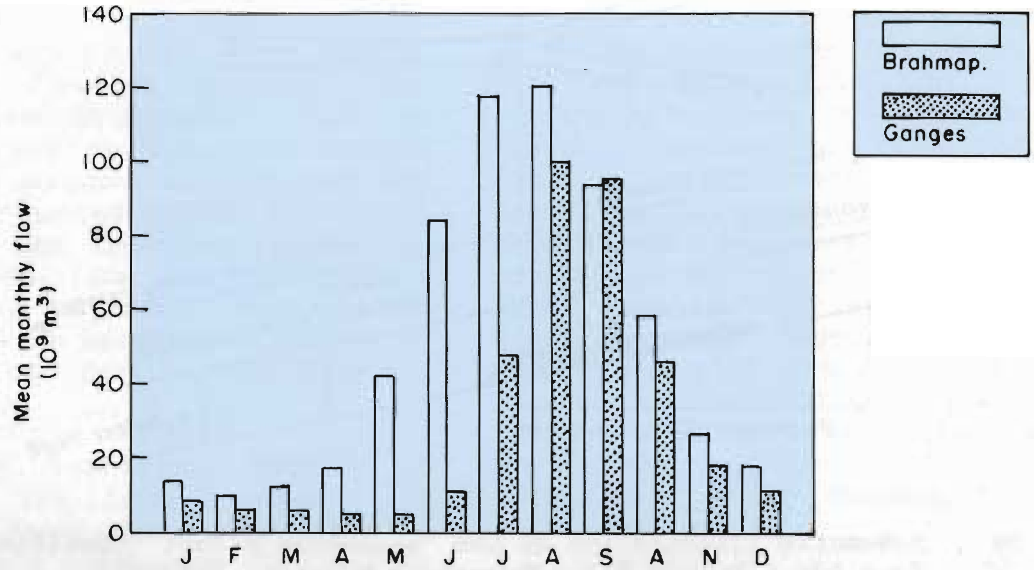

Figure 36. Average monthly discharge of the Brahmaputra at Bahadurabad and the Ganges at Sara Bridge (based on data presented by Haroun er Rashid, 1977).

Plate 16 The confluence of the rivers Ganges (left) and Jamuna (right) in Bangladesh, June 1986 (photograph by G.J.Klaassen).

Plate 17 A small glacial lake outburst flood disrupting a bridge on the Dudh Kosi river, Central Nepal, July 1984 (photograph by J. Desloges).

very nature makes "GLOFs" extremely difficult to predict (Ives, 1986).

All in all, there seem to be enough reasons why the lower Ganges-Brahmaputra plains should be liable to regular and extensive flooding, even when the neighbouring uplands were fully forested. The question to what extent recent deforestation in the mountains has exacerbated the flood problem will be addressed in Section IV.1.3.

III.6 RIVER SEDIMENT LOADS

In his reconnaissance survey of amounts of sediment transported in suspension (as opposed to rolling along the river bed) to the seas by the major rivers of the world, Holeman (1968) concluded that Asia's rivers were by far the greatest contributors, possibly supplying up to 80% of the world total.

Of the Asian rivers, the combined Ganges-Brahmaputra basin ranked as the first, with an estimated annual sediment yield of 2.4 billion tons (15 t/ha), of which the Ganges alone contributed about two thirds (Holeman, 1968).

A more modest estimate (1.67 billion tons, or 11.3 t/ha) was published more recently by Milliman & Meade (1983). Although sediment loads transported by such very large rivers can never be determined with great precision, it will be clear that the amounts of sediment carried by the two rivers must be enormous.

This is well illustrated by a comparison with the sediment loads for such tropical rivers as the Orinoco (0.9-2.1 t/ha/yr) or the Mekong (2.1 t/ha/yr) (Holeman, 1968; Milliman & Meade, 1983). Since both streamflow totals and basin size for these rivers are comparable to those of the Ganges, such differences in erosion must

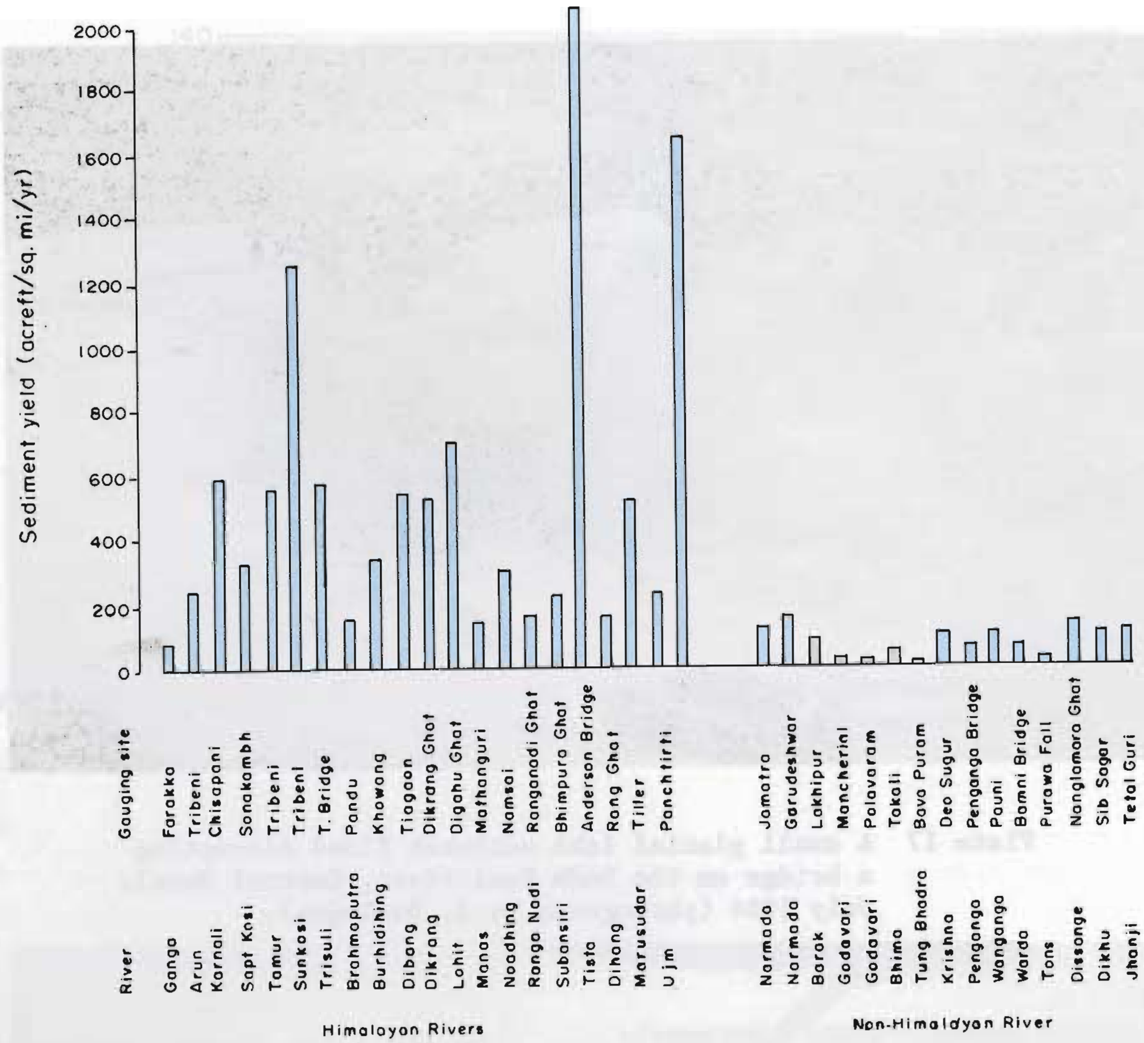

Figure 37 Sediment yields for Himalayan and Peninsular rivers in India (after Gupta (1975) in Tejwani, 1985).

reflect differences in basin relief and geology (Gregory & Walling, 1973).

As such, major contrasts in sediment transport may be expected between "Himalayan" and "Peninsular" rivers within the Ganges-Brahmaputra river basin (Figure 37).

A glance at Figure 37 not only reveals much lower sediment loads for non-Himalayan rivers, but also rather dramatic differences between the Himalayan streams themselves. In some cases such differences will reflect contrasts in basin size, since sediment yields (per km^2) generally tend to decrease with basin area; larger basins usually have more opportunities to store sediment, e.g. on their floodplains (Dunne, 1977).

Furthermore, as was observed for annual streamflow totals (Section III.1), the larger the portion of a catchment in the dry Trans-Himalaya the lower will be its sediment load (Das, 1968). Finally, certain rock types (e.g. phyllites, shales or unconsolidated sandstones) are much more susceptible to erosion and/or mass movements than others (see section IV.2), resulting in widely different stream sediment loads under otherwise similar conditions (Rawat, 1985; Biksham & Subramanian, 1988).

As such, each "meso-scale" basin naturally represents a more or less unique combination of the above variables. A good example of such local contrasts in sediment yield is exhibited by the three main branches of the Kosi above Tribeni in East Nepal (Table 4).

In the Tamur basin, the combination of high streamflow rates (reflecting high rainfall and meltwater contributions) and steep and unstable terrain with relatively few opportunities to store sediment, results in an impressive rate of sediment transport (cf. Figure 37). However, the influence of human activities in this area should not be underestimated (Brunsden et al. 1981). This aspect will be discussed more fully in Chapter IV.2.

Sediment yields in the Arun basin on the other hand are quite low, reflecting the fact that 90% of its area receives very little precipitation. In addition, a significant part of this dry headwater area consists of flat and broad valley bottoms which act as effective traps for sediment coming from the slopes.

However, if we take the 10 m^3/ha/yr quoted for the Arun in Table 4 at face value, the contributions to the overall sediment load made by the High and Middle Himalaya physiographic regions in the Nepalese part of the basin must be considerable.

Taking 1.5 t/ha/yr as a representative value for the amounts of sediment transported by South Tibetan rivers of this magnitude (Guan & Chen, 1981), it follows that the remaining 10% of the basin should supply about 120 t/ha/yr. Although the latter figure may seem excessive, it certainly points to the importance of the High and Middle Himalaya regions as contributors of stream sediment (Carson, 1985).

The Sun Kosi occupies an intermediate position between the Arun and Tamur, as could be expected on the basis of its annual streamflow total and percentage catchment area in the dry Himalaya (Table 4).

Table 5 explores a little further the spatial variations in suspended sediment loads between the various physiographic units (cf. Figure 3).

It would seem from these limited (semi-long term) data that the Middle Himalaya supplies the greatest quantities of (fine) sediment to the streams. This accords with the observation of Chyurlia (1984) that the zone also produces the largest amounts of stormflow (see section III.4).

Virtually no reliable data are available for streams in the High Himalaya and it may well be that suspended sediment loads in this highly glaciated and high-energy environment are larger than suggested in Table 5.

Alternatively, it could be argued that much of the sediment released by physical weathering in this zone is initially too coarse to be transported in suspension (Plates 2 and 3) and will move along the river bed until sufficiently broken for suspended transport (Galay, 1987). Further work on this matter is desirable.

As already indicated, rivers in the Trans-Himalayan zone and streams draining the old plateau in the south carry very little sediment (Table 5).

Not all of a river's sediment load is transported in suspension. A significant, but often unmeasured part moves along the channel as bedload. Although separate measurements of bedload transport in Himalayan rivers are rarely available (Galay, 1987), there is a growing body of information regarding total (i.e. suspended plus bed-) loads deposited behind dams erected on several major rivers in the area (Table 6).

Recorded rates of sedimentation in the Himalaya again vary widely. The low value found for the Bhakra reservoir must reflect the fact that a considerable part of the upper Sutlej catchment is situated in the dry Trans-Himalaya. In addition, the lower parts of the basin do not receive much rainfall either, due to their position in the far west (Figure 8). Furthermore, there may be a scale effect as well, the Bhakra basin being the largest Himalayan catchment quoted in Table 6.

The Kalagarh reservoir is fed by streams that mainly drain the Middle

Table 4. Mean annual streamflow and suspended sediment loads (1948-1959) for three major tributaries of the Kosi river, eastern Nepal (Das, 1968)

Rivers	Catchment area (km^2)	Water yield (mm/yr)	Sediment yield (m^3/ha/hr)
Sun Kosi+	18.985	570	28.6
Arun*	34.525	445	10.0
Tamur	5.770	1405	51.3

(+20% in Trans-Himalaya
(*90% in Trans-Himalaya

Table 5. Suspended sediment loads for selected rivers of intermediate size in the Ganges-Brahmaputra river basin

Physiographic Zone	River	Suspended sediment (t/ha/yr)	Period of measurements	Basin area (km^2)
Trans-Himalaya	Nyang He[1]	1.5	1965-1975	6215
	Lhasa He[1]	0.4	1963-1975	6225
High Himalaya	Trisuli[2]	18	1973	4100
Middle Himalaya	Bagmati[2]	45	1973	585
M + H Himalaya	Tehri[3]	30*	1973-1981	7510
	Tamur[4]	70*	1948-1959	5700
Siwaliks	Rapti[2]	ca 15	1973?	5150
	Kansal) 5 Sukhetri)	98*+	1958-1979	44
Eastern Plateau (Assam)	Burhi Dihing[6]	7	1972-1982	5180
Southern Plateau	Godavari[7]	5	1969-1980	54000

*original data in $m^3/km^2/yr$ converted to t/ha/yr by applying a density factor of 1.4 kg/m^3 (Singha & Gupta, 1982)

+including bedload; disturbed
[1]Guan & Chen (1981)
[2]Sharma (1977)
[3]Singha & Gupta (1982)
[4]Das (1968)
[5]Gupta (1983)
[6]Sarma (1986)
[7]Biksham & Subramanian, 1988)

Table 6. Rates of sedimentation for major reservoirs in and around the Ganges river basin

Name of reservoir	Siltation rate ($m^3/ha/yr$)	Period of measurement	Catchment area (km^2)
Himalaya			
Bhakra (Sutlej)[1]	6.9	1959-1965	56.875
	6.0	1965-1970	
	5.9	1970-1981	
Kalagarh (Ramganga)[2]	57.5	1975-1978	3.105
	52.5	1978-1981	
Southern Plateau			
Mayurakshi (Mayurala)[3]	16.5	1955+	1.790
Maithon (Baskar)[3]	14.3	1956+	5.205
Panchet (Damodar)[3]	10.8	1956+	9.815
Matatila (Betwa)[3]	4.4	1958+	20.750
Gandhisagar (Chambal)	9.6	1960+	21.875

[1]Singha & Gupta (1982); [2]Anonymous (1987); [3]Gupta (1983)

Himalayan and Siwalik zones. The high siltation rate is in agreement with other observations for this zone (Table 5).

As for the southern plateau, there may again be a tendency for larger basins to have somewhat lower siltation rates (i.e. per km^2). However, this may also be related to differences in rainfall as the Maithon and Panchet reservoir areas receive more rainfall than those situated further west (cf. Figure 8; Lal et al., 1977).

The very high amounts of sediment carried by most major Himalayan rivers cause them to adopt a braided pattern upon reaching the piedmont zone (Figure 6; Plates 7 and 9).

A good example is the Kosi river in eastern Nepal, which has not been able to cut a deep and stable bed for itself after leaving the Chatra gorge.

During times of high flow, the river easily overtops its shallow banks and spreads over a vast expanse that may in places be up to 30-40 km wide (Pal & Bagchi, 1975). Due to the continued deposition of sediment, the river bed has risen to several metres above the surrounding plain, thereby creating a highly dangerous situation (Zollinger, 1979).

However, as shown in Figure 38, the Kosi has been notoriously unstable for more than 250 years (Gole & Chitale, 1966; updated by Galay, 1987). Interestingly, the shift has been progressively westwards (Pal & Bagchi, 1975).

Also, according to Sharma (1977), the rivers lying east of Butwal (Nepal's Western Development Region) are shifting towards the west, and those west of Butwal towards the east, suggesting a tectonic origin for such movements. Morgan & McIntire (1959) have described several other cases of major river shifts in their analysis of the Quaternary geology of the Bengal basin (see also Coleman (1969) and Klaassen & Vermeer, 1988).

Tejwani (1985) discussed the work

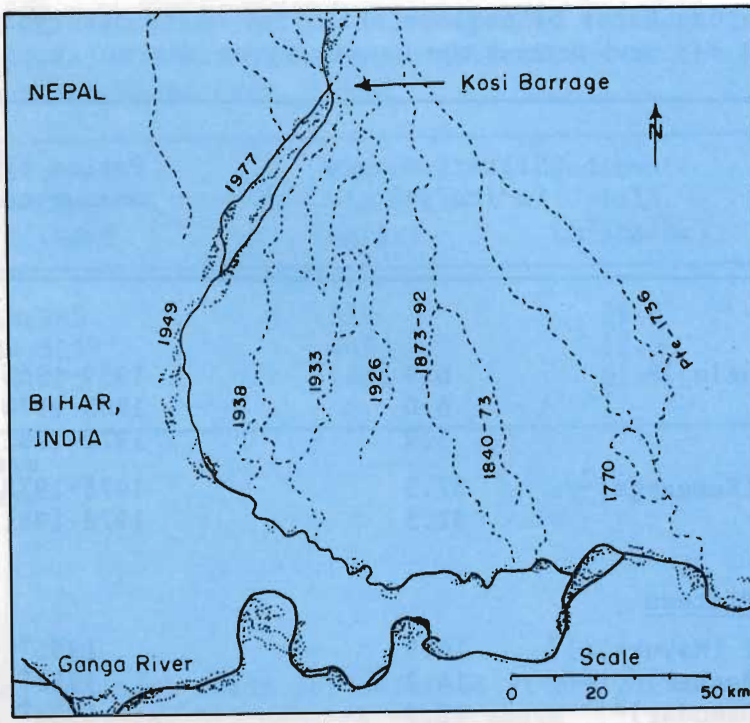

Figure 38 **Shifting of the Kosi River over its own alluvial fan, 1736-1977 (after Gole & Chitale, 1966; updated by Galay, 1987).**

of Gupta, who analysed sequential satellite imagery and observed substantial widening of river beds in the Terai zone between 1972 and 1979. The effect was ascribed to upland deforestation. As such, it is most important to monitor such changes in river morphology closely to reveal any further trends.

In view of the strongly seasonal character of most of the rivers in the region (Figure 20), the bulk of the annual sediment transport will take place during the summer monsoon. Also, major variations between years are to be expected as a result of inter-annual variations in rainfall distribution (Singha & Gupta, 1982; Figure 39).

It follows that long-term observations of sediment discharge are needed in order to arrive at meaningful estimates of average values. For example, suspended sediment transport in the Bhagirathi river (Tehri Garhwal, catchment area 7510 km^2) between 1973 and 1981 varied by a factor of 5.4 (Figure 39).

Therefore, the results of short-term studies, such as the comparative investigations of Subramanian (1979), Abbas & Subramanian (1984) and Jha et al. (1988), should be seen as representing an order of magnitude at best.

The inadequacy of a limited sampling program was demonstrated rather dramatically in a study of sediment transport on the Godavari, one of the rivers draining the southern plateau area (Biksham & Subramanian, 1988).

Based on occasional sampling during the various seasons over a period of three years, the authors computed a mean suspended sediment concentration

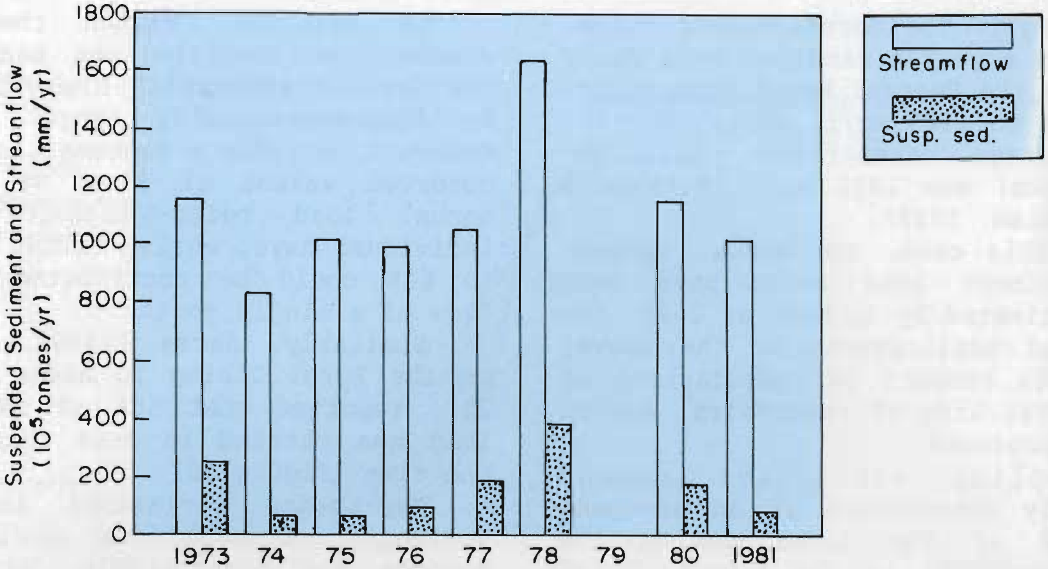

Figure 39 (a) Annual streamflow and suspended sediment load for the Bhagirathi river, Tehri Garhwal, 1973-1981.

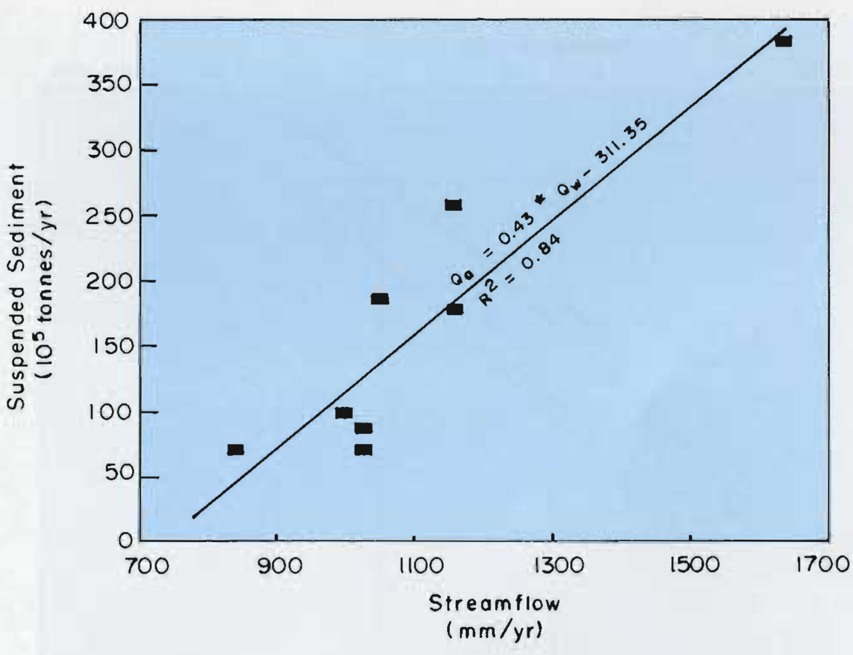

Figure 39 (b) Relationship between annual streamflow and amount of sediment carried in suspension by the Bhagirathi river (based on original data of Singha & Gupta, 1982).

of 770 mg/l. The corresponding value, based on a year's sampling on a daily basis by the Central Water Commission, amounted to 1525 mg/l, whilst the 10-year average concentration (daily observations) was 1845 mg/l (Biksham & Subramanian, 1988).

In this case, the annual suspended sediment load would have been underestimated by as much as 240%. The practical implications of the above, e.g. with respect to computations of the useful life of reservoirs, are of course profound.

Sampling strategies should obviously concentrate at an adequate coverage of the flows during the summer monsoon. In the already cited study of Singha & Gupta (1982) on the Bhagirathi river, as much as 92 to 99% of the annual suspended load was carried between June and September. Similarly high values have been reported for several rivers in Nepal (Sharma, 1977) and on the southern Plateau (Biksham & Subramaman, 1988), as well as for the Brahmaputra itself (Goswami, 1985).

In addition, within the rainy season, the contributions made by a few days of abnormally high flows can be disproportionally large. On the Godavari, Biksham & Subramarian (1988) observed values of 5 to 9% of the annual load being transported on individual days, whilst values of up to 64% could be contributed by the flow of a single month.

Similarly, Sarma (1986), working on the Burhi Dihing in Assam (Figure 23), reported that 50% of the total load was carried in less than 7% of the time (Figure 40).

Day-to-day variations in river sediment concentrations during the monsoon are considerable, with peak values being especially frequent in the second half of the rainy season (Figure 41b; Carson, 1985; Biksham & Subramanian, 1988).

Since surface erosion rates in the area are highest when vegetation cover is still sparse and rainfall intensities high, i.e. at the beginning of the rains in June (Figure 41a; Impat, 1981), some other mechanism of

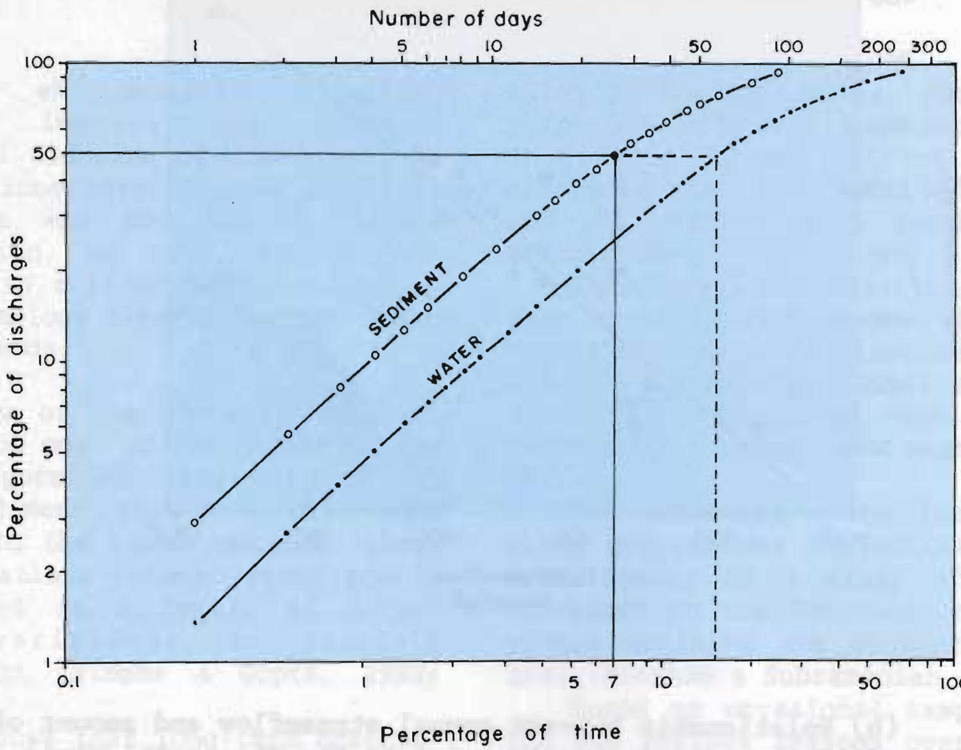

Figure 40 **Cumulative percentages of suspended sediment- and water discharges on the Burhi Dihing river against cumulative percentage of time for the year 1974 (after Sarma, 1986).**

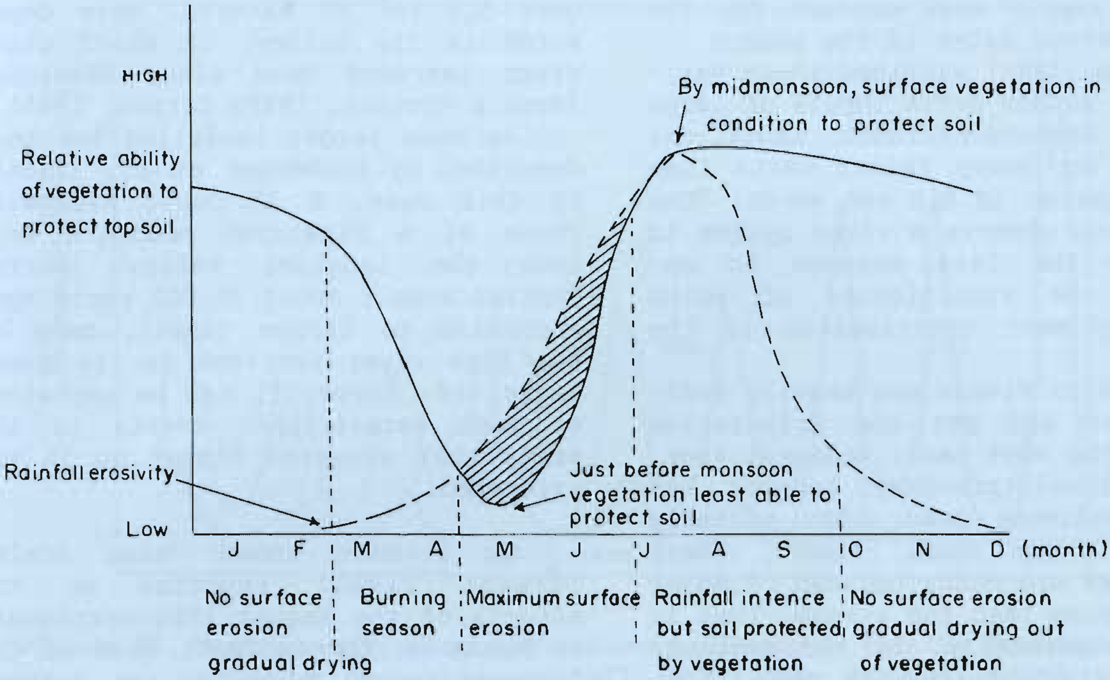

Figure 41 (a) Hypothetical course of rainfall erosive power and state of surface vegetation throughout the year in the Middle Mountains of Nepal (after Carson, 1985).

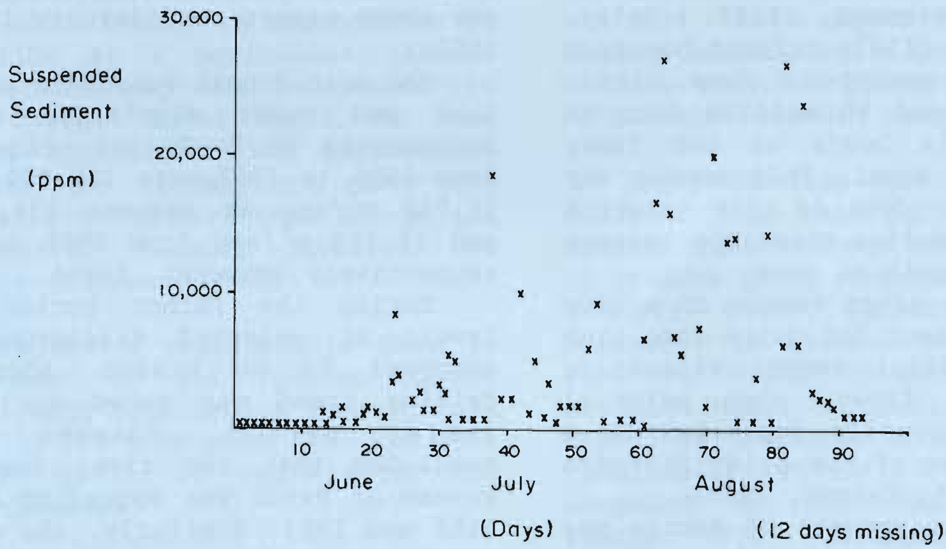

Figure 41 (b) Instantaneous suspended sediment concentrations for the Narayani river, Narayanghat, June-August, 1979 (after Carson, 1985).

sediment supply must account for the peaks observed later in the season.

Carson (1985) ascribed these variations to sudden extra inputs of large sediment amounts through landslides triggered by heavy rains, earth tremors, or both. In his own words: "One has only to observe a river system in Nepal in the late monsoon to appreciate the significance of point source sediment contribution in the Himalaya.

The major rivers are heavily sediment laden and yet the tributaries are, for the most part, sediment free. The occasional tributary, however, has extreme sediment loads that markedly discolour the main river. These tributaries are muddy because of major mass slumping into the stream, that is supplying virtually all the sediment to that river system at that time. Point source sediment contributions, caused by mass wasting, are the major contributors of sediment for many Himalayan rivers" (Carson, 1985).

As discussed earlier in Section III.5 the largest of such slides (cf. Plate 6) may temporarily block a river. When such dams burst, the result is a devastating downstream surge of water and sediment (Singh et al., 1974; Mahmood, 1987; Galay, 1987). Mahmood (1987) related how such an event, occurring in June, 1980, transported about 60 million tons of sediment in 14 hours on the Tamur river in East Nepal. This amount was equivalent to 36% of the river's annual load and five times the average load for the month of June.

Such sites often remain an active source of sediment for quite some time after the initial event, especially during higher flows, when material that was temporarily deposited as a fan at the base of the slide is again eroded (Plate 6; Carson, 1985).

Even larger amounts of debris may be carried by glacial lake outburst floods (Galay, 1987; Plate 16). One of the most spectacular of such events in historic times was the GLOF on the Seti Khola in the Pokhara valley, which occurred about 600 years ago (Yamanaka, 1982). During the event, over 5.5 km^3 of material were deposited in the valley, in which giant river terraces have since developed (Fort & Freitel, 1982; Carson, 1985).

An even larger landslide has been described by Heuberger et al. (1984). In this case, a 15 cubic kilometre chunk of a Himalayan mountain fell into the Langtang valley (North-Central Nepal) about 30,000 years ago. According to Carson (1985), many of the high river terraces in the Himalayas (cf. Plate 21) can be explained by such catastrophic events in the past, that occurred higher up in the watershed.

On a more recent time scale, Goswami (1985) reported on the effects of the August 1950 earthquake in Assam on the sediment load of the Brahmaputra in Assam Valley between 1955 and 1979. During this earthquake, apparently one of the most severe ever recorded, massive landslides occurred which temporarily blocked the Subansiri, Dibang and Dihang rivers (Figure 1).

Bursting of these dams after several days (!) not only produced devastating floods downstream, but also brought down enormous volumes of sediment, thereby raising the beds of the above rivers considerably (Poddar, 1952).

The mean annual suspended sediment load and water discharges of the Brahmaputra at Pandu (cf. Figure 2b) from 1955 to 1963 were 750.000 m^3 and 16,530 m^3/sec as against 130,000 m^3 and 14,850 m^3/sec from 1969 to 1976, respectively (Goswami, 1985).

During the former period water levels at selected discharges were observed to be rising, whereas a falling trend was noted during the latter period. Goswami (1985) concluded that the river reach upstream of Pandu was aggrading between 1955 and 1963. Similarly, the channel was degrading between 1969 and 1976.

However, one should be extremely careful with the interpretation of changes in sediment load with time. The increased sediment discharges observed in the late 1970s at Pandu, for example (Figure 42), appeared to

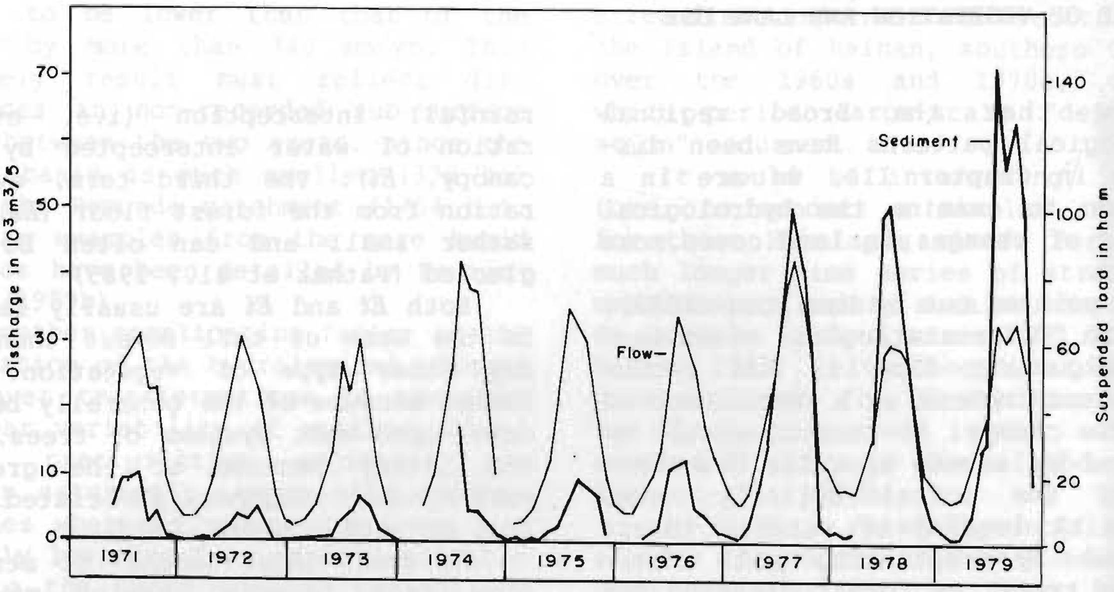

Figure 42. Mean monthly flow and sediment discharge for the Brahmaputra river at Pandu, Assam, 1971-1979 (after Goswami, 1985).

reflect temporary degradation of the reach immediately upstream of the gauging site.

The reach immediately downstream experienced significant aggradation during that time (Goswami, 1985).

As such, trends in aggradation or degradation at a particular channel cross section may not be indicative for an overall river reach. An analysis of aggradation and degradation rates for the Brahmaputra river bed over a stretch of more than 600 km and over several decades revealed the following (Goswami, 1985):

- There was a considerable gain in sediment in all reaches except one, with aggradation ranging from 0.5 to 2.4 m between 1957 and 1971 (i.e. up to 20 years after the earthquake).

- Between 1971 and 1977 an average degradation of about 20 cm was determined; i.e. only a small fraction of the material deposited earlier was again removed.

- Since streamflow amounts did not differ appreciably between the two periods, the recent removal of sediment from the river bed reflected decreased rates of sediment inputs to the system.

Thus, there appear to be phases of rapid aggradation associated with extreme events followed by periods of relatively slower removal (Goswami, 1985).

Any predictive equations of stream sediment load that fail to take into account such extreme events are bound to produce gross underestimates (Singh & Gupta, 1982). Therefore, it is no surprise to learn that rates of reservoir siltation predicted by the standard equation developed by Khosla (1953) on the basis of observations on "peninsular rivers", were severely exceeded in the Indian Himalayas (Gupta, 1983). As such, there is no need to ascribe the discrepancies between "predicted" and observed sediment loads to "deforestation" as is frequently proposed (e.g. Murty, 1985; Tejwani, 1987, and many others). This aspect will be discussed in more detail in Section IV.2.3.