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# Evolution of Miocene fluvial environments, eastern Potwar plateau, northern Pakistan

IMRAN A. KHAN,* JOHN S. BRIDGE,* JOHN KAPPELMAN† and REGAN WILSON†
**Department of Geological Sciences, Binghamton University, Binghamton, New York 13902–6000, USA*
*†Department of Anthropology, The University of Texas at Austin, Austin, Texas 78712–1086, USA*

## ABSTRACT

The Miocene-Pliocene Siwalik Group records changing fluvial environments in the Himalayan foreland basin. The Nagri and Dhok Pathan Formations of this Group in the eastern Potwar Plateau, northern Pakistan, comprise relatively thick (tens of metres) sandstone bodies and mudstones that contain thinner sandstone bodies (metres thick) and palaeosols. Thick sandstone bodies extend for kilometres normal to palaeoflow, and are composed of large-scale stratasets (storeys) stacked laterally and vertically adjacent to each other. Sandstone bodies represent single or superimposed braided-channel belts, and large-scale stratasets represent channel bars and fills. Channel belts had widths of km, bankfull discharges on the order of $10^3$ cumecs and braiding parameter up to about 3. Individual channel segments had bankfull widths, maximum depths, and slopes on the order of $10^2$ m, $10^1$ m and $10^{-4}$ respectively, and sinuosities around 1·1. These rivers are comparable to many of those flowing over the megafans of the modern Indo-Gangetic basin, and a similar depositional setting is likely. Thin sandstone bodies within mudstone sequences extend laterally for on the order of $10^2$ m and have lobe, wedge, sheet and channel-form geometries: they represent crevasse splays, levees and floodplain channels. Mudstones are relatively bioturbated/disrupted and represent mainly floodbasin and lacustrine deposition. Mudstones and sandstones are extremely disrupted in places, showing evidence of prolonged pedogenesis. These 'mature' palaeosols are m thick and extend laterally for km. Lateral and vertical variations in the nature of their horizons apparently depend mainly on deposition rate.

The 500 m-thick Nagri Formation has a greater proportion and thicker sandstone bodies than the overlying 700 m-thick Dhok Pathan Formation. The thick sandstone bodies and their large-scale stratasets thicken and coarsen through the Nagri Formation, then thin and fine at the base of the Dhok Pathan Formation. Compacted deposition rates increase with sandstone proportion (0·53 mm/year for Nagri, 0·24 mm/year for Dhok Pathan), and palaeosols are not as well developed where deposition rates are high. Within both formations there are 100 m-scale variations (representing on the order of $10^5$ years) in the proportion and thickness of thick sandstone bodies, and tens-of-m-scale alternations of thick sandstone bodies and mudstone-sandstone strata that represent on the order of $10^4$ years. Formation-scale stratal variations extend across the Potwar Plateau for at least 100 km, although they may be diachronous: however, 100-m and smaller scale variations can only be traced laterally for up to tens of km.

*Present address: Imran A. Khan, Geological Survey of Pakistan, Palaeontology and Stratigraphy Branch, 84, H-8/1, Islamabad, Pakistan.

Alluvial architecture models indicate that increases in the proportion and thickness of thick sandstone bodies can be explained by increasing channel-belt sizes (mainly), average deposition rate and avulsion frequency on a megafan comparable in size to modern examples. 100-m-scale variations in thick sandstone-body proportion and thickness could result from 'regional' shifts in the position of major channels, possibly associated with 'fan lobes' on a single megafan or with separate megafans. However, such variations could also be related to local changes in subsidence rate or changes in sediment supply to the megafan system.

Formation-scale and 100-m-scale stratal variations are probably associated with interelated changes in tectonic uplift, sediment supply and basin subsidence. Increased rates of hinterland uplift, sediment supply and basin subsidence, recorded by the Nagri Formation, may have resulted in diversion of a relatively large river to the area. Alternatively, changing river sizes and sediment supply rates may be related to climate changes affecting the hinterland (possibly linked to tectonic uplift). Climate during deposition of the Siwalik Group was monsoonal. Although the deposits contain no *direct* evidence for climate change, independent evidence indicates global cooling throughout the Miocene, and the possibility of glacial periods (e.g. around 10·8 Ma, corresponding to base of Nagri Formation). If the higher Himalayas were periodically glaciated, a mechanism would exist for varying sediment supply to megafans on time scales of $10^4$–$10^5$ years. Although eustatic sea-level changes are related to global climatic change, they are not directly related to Siwalik stratigraphic changes, because the shoreline was many 100 km away during the Miocene.

## INTRODUCTION

The Siwaliks are fluvial deposits that accumulated in the foreland basin on the southern side of the Himalayas. This foreland basin formed in association with the collision of the Indian and Eurasian plates and the uplift of the Himalayas, initiated ≈60 million years ago (Patriat & Achache, 1984; Besse & Courtillot, 1988; Klootwijk *et al.*, 1992; Beck *et al.*, 1995). The Siwaliks provide a record of changing depositional environments, faunas and floras that is significant for three main reasons. First, the hydrologic, sedimentologic and tectonic processes and forms of the modern Indo-Gangetic fluvial basin are probably very similar to those in the past (at least the last 20 million years) and therefore provide a valuable analogue for interpreting the Siwaliks. Second, it is possible to evaluate the effects on the changing nature of the deposits of extrinsic factors such as regional tectonism, eustatic sea-level changes and climatic changes, for which there is a growing body of information from outside the depositional basin. Third, terrestrial vertebrate animals and plants evolved dramatically during deposition of the Siwaliks, and the reasons for this evolution can be framed in terms of changing environments.

The Miocene to Pliocene Siwaliks of the Potwar plateau in northern Pakistan are particularly well exposed and have been studied extensively. The earlier studies were mainly concerned with lithostratigraphic and biostratigraphic subdivision and correlation (e.g. Pilgrim, 1913, 1934; Lewis, 1937; Gill, 1951; Fatmi, 1973; Shah, 1977). Fossil vertebrates are abundant and biostratigraphic zonation based on vertebrate fauna is now well developed (Pilbeam *et al.*, 1977, 1979, 1980; Barry *et al.*, 1980, 1982). Palaeomagnetic polarity studies in many parts of the Potwar plateau have facilitated absolute dating of deposits and biozones, chronostratigraphic correlation, and determination of compacted sediment accumulation rates (e.g. Barndt *et al.*, 1978; Opdyke *et al.*, 1979, 1982; Pilbeam *et al.*, 1979; N.M. Johnson *et al.*, 1982, 1985, 1988; Tauxe & Opdyke, 1982; Badgley *et al.*, 1986; Tauxe & Badgley, 1988; Badgley & Tauxe, 1990; McRae, 1990; Flynn *et al.*, 1990). Changes in sedimentary characteristics have been related to tectonism on a relatively broad scale (e.g. Visser & Johnson, 1978; Raynolds & Johnson, 1985; N.M. Johnson *et al.*, 1985; G.D. Johnson *et al.*, 1986; Burbank & Raynolds, 1984, 1988; Burbank *et al.*, 1986, 1988; Cerveny *et al.*, 1988, 1989; Beck & Burbank, 1990; Burbank & Beck, 1989, 1991; Burbank, 1992; Mulder & Burbank, 1993; Meigs *et al.*, 1995), and attempts have been made to relate mammalian faunal change to global climatic and tectonic events (Barry *et al.*, 1985; Opdyke, 1990). More detailed sedimentological studies in specific parts of the Potwar plateau have involved reconstruction of depositional environments in combination with palaeoecological-taphonomic studies of vertebrate faunas (e.g. Badgley & Behrensmeyer, 1980; Behrensmeyer & Tauxe, 1982; Behrensmeyer, 1987, 1989). However, despite this impressive body of previous work, a detailed knowledge of Siwalik deposits throughout the Potwar Plateau (and beyond) is lacking.

In the late 1980s, a collaborative research project was initiated that had the following objectives: (1) detailed description and palaeomagnetic dating of Miocene-Pliocene fluvial deposits in different parts of the Potwar Plateau in order to establish the temporal and spatial variation of the deposits; (2) interpretation of the evolution of palaeoenvironments (e.g. palaeochannel and floodplain geometry, flow and sediment transport; nature of palaeosols; organic activity) and deposition rates across the area, including identification of local vs. regional trends; (3) examination of the relationship between palaeoenvironmental and faunal evolution, and; (4) analysis of changes in palaeoenvironments, faunas and deposition rates in relation to factors such as river diversions, variations in sediment supply, tectonism, climate and base-level change. Results of this project that have been published to date include detailed sedimentological analyses from already well-studied areas of the Potwar plateau (Willis, 1993a,b; Willis & Behrensmeyer, 1994, 1995; Behrensmeyer *et al.*, 1995), and analysis of oxygen and carbon isotopes of palaeosol carbonates that has implications for changing climates and biotas (Quade *et al.*, 1989, 1992, 1993; Cerling *et al.*, 1993; Harrison *et al.*, 1993). Detailed sedimentological studies have now been completed in two other areas within the Potwar Plateau (Khan, 1993; Zaleha, 1994). The objective of this paper is to report on the sedimentologic-palaeomagnetic study undertaken by Khan and associates in the eastern Potwar plateau, and to compare this with the results of Willis's study (1993a,b). Subsequent papers by Zaleha will continue this process of describing and interpreting new sections and comparison with previous work.

## STUDY AREA AND METHODS

The study area is in the eastern Potwar plateau near the villages of Chakoha and Chak Mehun, Tehsil Dina, District Jhelum (Fig. 1). The strata are exposed on the southeastern flank of the Mahesian anticline, where they dip at 13° to 35° to the east-south-east. A detailed, unbroken sedimentological log, encompassing 1·2 km of strata, was measured along and adjacent to Chakwala Kas (Fig. 1C). In addition, lateral (along strike) variation in the strata was documented for three stratigraphic intervals within this section using closely spaced (tens of m) sedimentological logs tied to photomosaics and field tracing of individual strata. Each interval was up to 75 m thick and extended laterally for over 1 km (Fig. 1C). Scale diagrams of these stratigraphic intervals were constructed by assuming that the tops of major sandstone bodies and palaeosols are approximately horizontal along strike.

### Palaeomagnetic materials and methods

Rock samples for palaeomagnetic analysis were collected from 100 sites following the procedure outlined by Johnson *et al.* (1975), with a mean stratigraphic spacing of 11·7 m (standard deviation 10·0 m, range 1·0–51·9 m). Site spacing was somewhat constrained by the presence of thick sandstone bodies, since only siltstone or claystone were sampled. Holes up to 0·7 m deep were dug at each site to collect a minimum of three unweathered samples. Sample locations were photographed and marked on a map and the section log. The sedimentological log with the positions of palaeomagnetic samples could not be published here: however, this information is available in Khan (1993) or can be obtained directly from Khan or Bridge. Samples were orientated on the outcrops using a hand rasp and a Brunton compass, and each sample was milled into a 2·5 $cm^3$ block using a cut-off saw and vertical platter grinder.

Each sample was measured for its natural remanence (NRM). Measurements were made on a 2G SCT superconducting magnetometres in a magnetically shielded room at the University of Texas. Previous studies of Siwalik sediments have shown that both magnetite (Opdyke *et al.*, 1979) and specular haematite (Tauxe *et al.*, 1980) can serve as carriers of primary magnetic remanence. It is also common to find varying contributions of pigmentary haematite in these rocks (Tauxe *et al.*, 1980; Tauxe & Badgley, 1984). Given

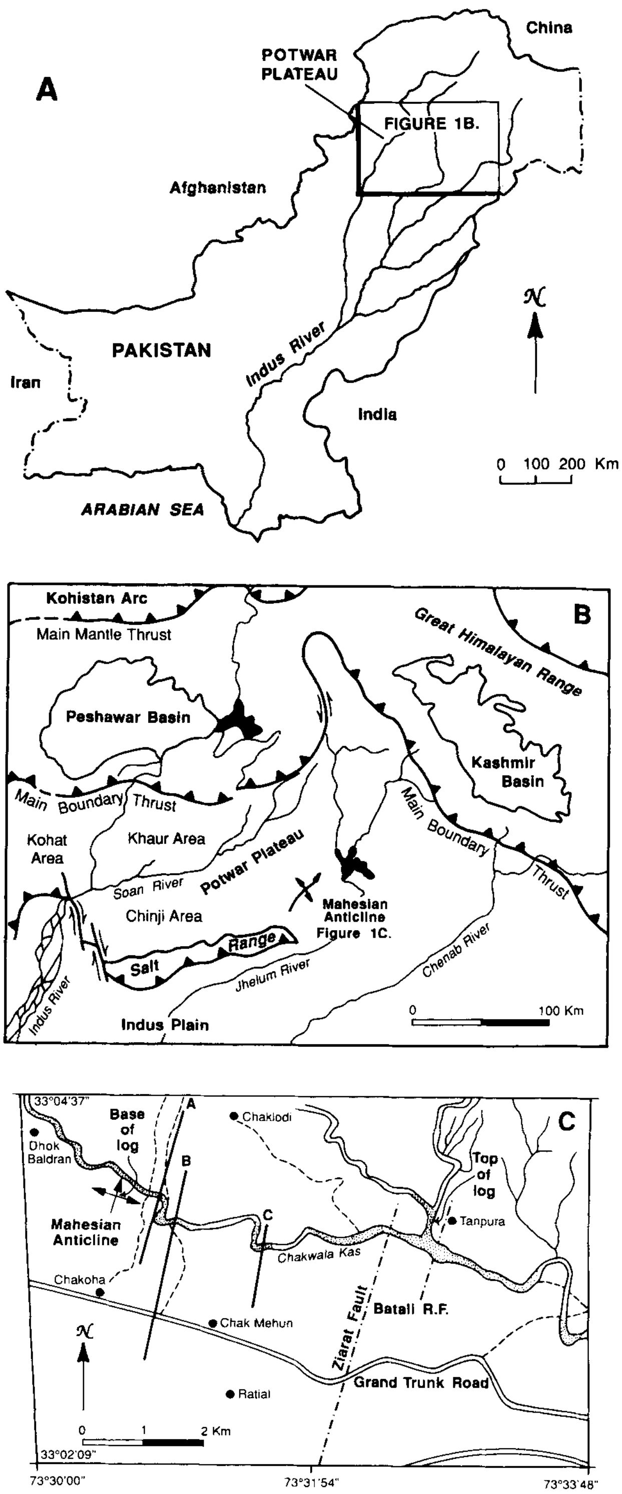



**Fig. 1.** Location maps of study area. Detailed location map of Mahesian area drawn from the Survey of Pakistan topographic map 43 G/12 (scale 1: 50,000). The sedimentological log was measured along Chakwala Kas. Locations of three stratigraphic intervals studied in detail are indicated by letters A, B, C.

this potential mix of magnetic minerals, thermal demagnetization is preferred over alternating-field demagnetization. A program of step-wise thermal demagnetization was used to remove secondary magnetic components acquired from either the present or past fields or oxidation reactions. Temperature stepping was carried out with a low-field Schonstedt Thermal Demagnetizer (TSD-1) in a shielded room. This involved seven and in some cases eight temperature steps, with measurements taken after each step. Calculations were made using PalaeoMagCalc© (a palaeomagnetics analysis program for Windows (ASC Scientific, Carlsbad, CA).

There are several sources of error in the palaeomagnetic age dating of stratigraphic sections (see also Zaleha, 1994): (1) insufficient sampling density; (2) reliability of individual sample-site means; (3) the possibility of erosional removal of parts of the section containing polarity reversals; (4) mismatching of the observed pattern of magnetic polarities with the magnetic time scale; (5) errors associated with the absolute age of magnetic polarity transitions, taken to be about 0·1 My (Cande & Kent, 1992). In addition, when performing chronostratigraphic correlations with data from other areas, it is necessary to determine how closely coordinated the sedimentological logging and palaeomagnetic sampling was, and whether the same palaeomagnetic time scale was used. In order to compare our age data with those used by Willis (1993a,b), it was necessary to convert the data he used to the Cande & Kent scale (see also Zaleha, 1994). In view of these potential errors, it is unlikely that resolution of palaeomagnetic age dating is better than 0·1 My.

## Palaeomagnetic results

The thermal demagnetization results for two typical samples are shown in Fig. 2. Sample M5A (27·5 m) is of normal polarity (Fig. 2A). It shows a low-temperature normal overprint that is removed by 400°C, with a linear decay towards origin. Sample 3A(10 m) is of reversed polarity (Fig. 2B). It demonstrates a nearly linear decay towards origin, with a slightly more random direction at 675°C. Some other reversed-polarity samples had a normal overprint that was removed by 400°C.

The magnetic vector for each sample was analysed by calculating a best-fit line using the principal component analysis (PCA) routine in PalaeoMagCalc©. In many cases, the low

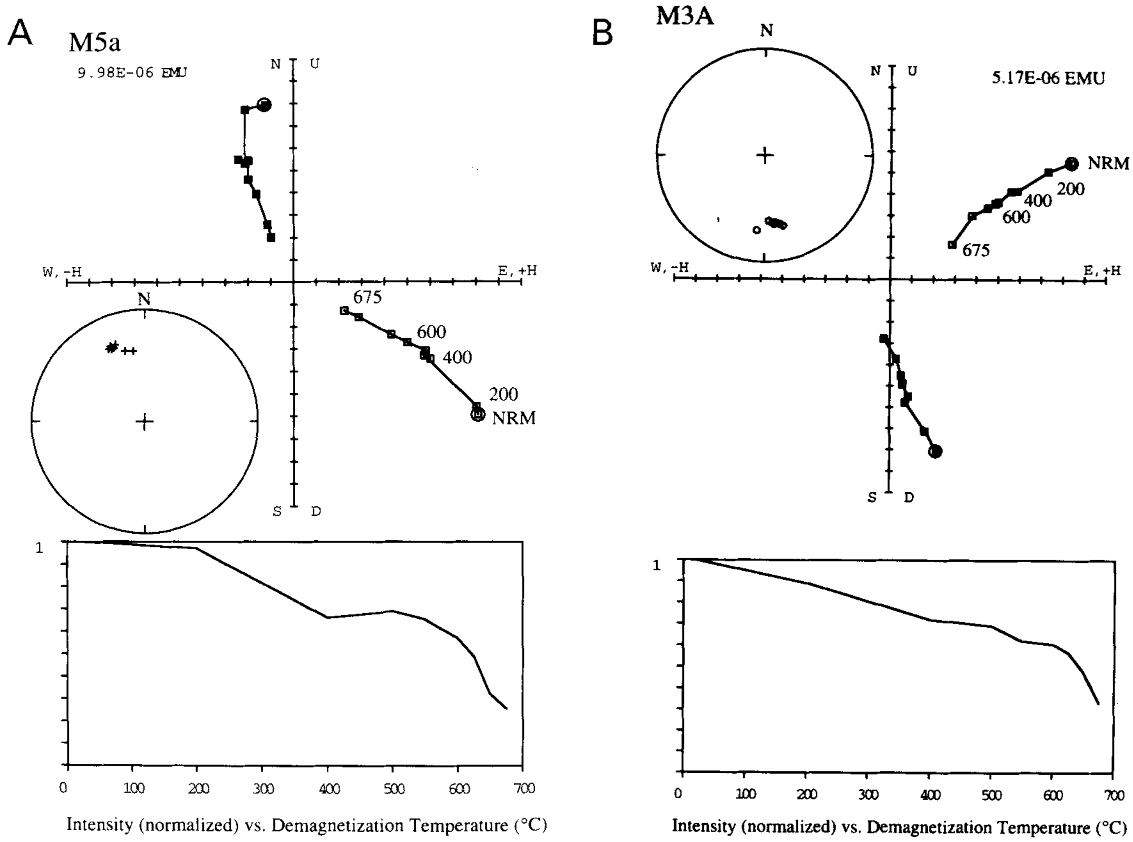
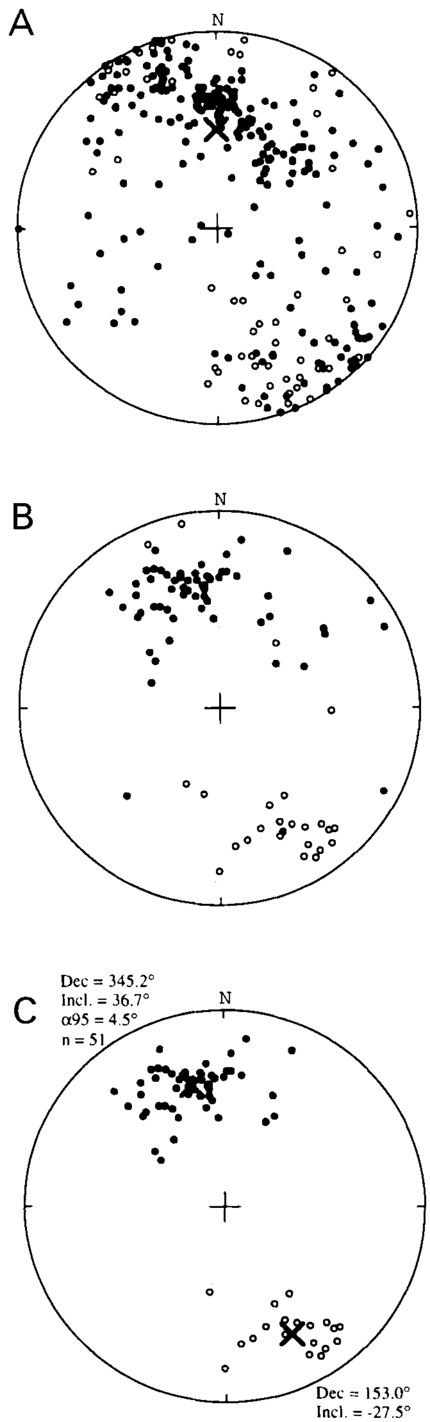



**Fig. 2.** Results from thermal demagnetization treatments of representative samples shown by vector end-point diagrams, stereoplots, and normalized intensity and temperature plots. (A) Sample M5A (27·5 m) demonstrates a normal overprint at NRM that is removed by 400°C, with the isolation of a stable normal vector with a decay to origin. (B) Sample M3A (11·0 m) shows a stable reversed vector from NRM to 650°C with a somewhat random component at 675°C. The plot of normalized intensity vs. temperature suggests the presence of a combination of haematite and magnetite or maghemite. Solid squares=declination, hollow squares=inclination, crosses=lower hemisphere, hollow circles=upper hemisphere.

temperature steps isolated a normal overprint, probably of recent origin, and these steps were not included in the analysis. Values were accepted as statistically significant if the maximum angular deviation (MAD) was less than 15°. In cases where the MAD value exceeded 15° but the polarity of the sample was not in question, a mean vector was calculated using the various temperature steps with Fisher's (1953) statistics, and statistical significance was evaluated with Watson's (1956) test for randomness. Site means and virtual geomagnetic pole latitudes were next calculated using Fisher statistics for those sites with three or more statistically significant samples. Sites meeting these criteria were classified as type A, whereas those that either did not meet the criteria or had only two samples of certain polarity were classified as type B.

Of the 100 sites in this study, 89 were class A (84 PCA, 5 Fisher fit), and 5 were class B (4 PCA, 1 Fisher fit). Six sites did not satisfy the statistical criteria. A stereoplot of the NRM measurements (Fig. 3A) shows the effect of the normal overprint discussed above. Stereoplots of the statistically significant class A PCA site means after thermal demagnetization are given in Fig. 3B. The normal and reversed populations are roughly antipodal to one another, and demonstrate a counter-clockwise rotation that has been noted previously for the Siwaliks (Opdyke *et al.*, 1982). The Mahesian section is relatively uniform in its direction of dip, and so a fold test could not be carried out with these data. A reversal test was conducted with the class A PCA site means, and those sites with an angular deviation exceeding 1·5 were deleted, resulting in a sample of 70 sites (Fig. 3C). An *F*-test shows that although the site means have minor overlap at the $\alpha 95\%$ level, the two populations differ at $P<0{\cdot}05$. This result suggests that the thermal demagnetization treatments may not have been completely successful in removing the normal overprint. These data do show, however, that stable magnetic vectors of both normal and reversed polarity are preserved in the strata of the Mahesion section. It is very likely that these directions reflect the primary detrital remanence.




The palaeomagnetic reversal stratigraphy of the Mahesian section is given in Fig. 4. Virtual geomagnetic pole latitudes are plotted for each sample, and class A and B site means obtained by either PCA or Fisher fits. Most of the intervals of normal or reversed polarity are defined by several sites. However, ten intervals are defined by one site only, but in all but two cases these are defined by class A sites. Both of the two class B sites (three samples at 1074·6 m, and two samples at 1122·7 m) are PCA best fits with MAD values for each sample of less than 7° and 3°, respectively. So the polarities are without question and these samples can be used to define polarity of these two intervals with a high degree of confidence. Half of the single-site intervals are defined by closely spaced samples of opposite polarity (R2, R6, R7, N7, R9), suggesting that these intervals are well defined. The wider spacing of samples around the other five single-site intervals (R1, R3, N10, N11, R12) suggests that the limits of these intervals can be better defined by future resampling.

Correlation with the geomagnetic time scale of Cande and Kent (1995), given in Figs 5 and 6, ties the lower intervals and the distinctive long normal interval between 112·5 m and 604·6 m to chron 5, and the upper polarity intervals to chrons 4Ar-4n and 3Br. Compacted sediment accumulation rates are calculated for each interval and averaged over the Nagri and Dhok Pathan Formations in Fig. 6. The mean rate of 0·39 mm/year for the 21 polarity intervals is within the range of rates for the Siwaliks published by others (e.g. Meigs *et al.*, 1995; Badgley *et al.*, 1986; Johnson *et al.*, 1982; Kappelman *et al.*, 1991).

There are, unfortunately, no independent means for establishing the age of the Mahesian section, as there are for the Chita-Parwala-Gabir Kas section to the southwest (Johnson *et al.*, 1982;

**Fig. 3.** Plots of Schmidt equal area projections for samples and site means. (A) Plot of all samples at NRM with the present field direction given by the large X, showing the effect of the present field overprint. Many of the samples are of reversed polarity at NRM. (B) Plot of PCA best-fit site means after thermal demagnetization with MAD values of less than 15°, demonstrating a separation into normal and reversed populations that are largely antipodal to one another. (C) Plot of data in B with the removal of sites with angular standard deviations in excess of 1·5, showing better definition of the two populations. Symbols as in Fig. 2, with large × representing present field (A) and population means (C).

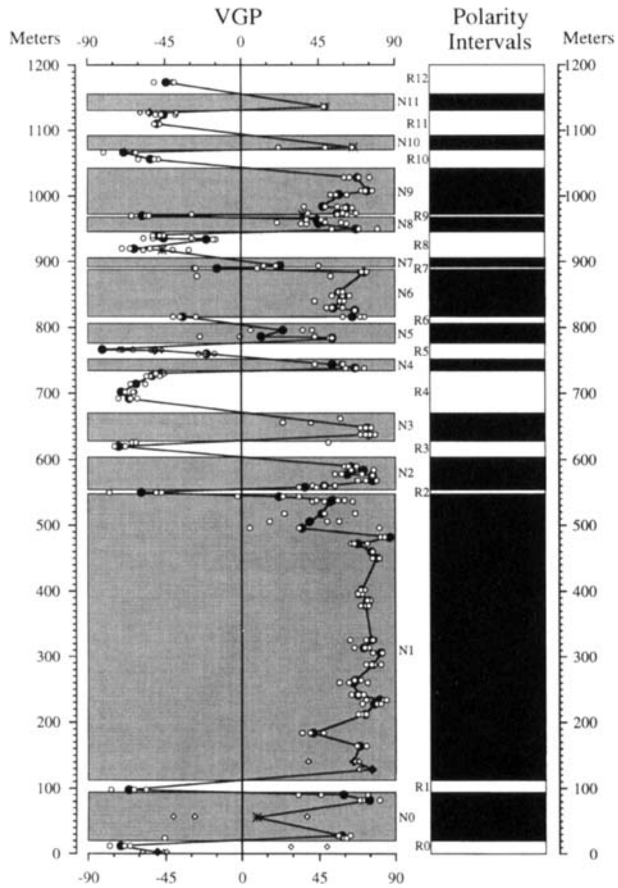



**Fig. 4.** Reversal stratigraphy of the Mahesian section. Palaeomagnetic class A samples are plotted in stratigraphic superposition by their virtual geomagnetic pole latitude based on either a PCA calculation (hollow circles) or a Fisher fit (hollow diamonds), with statistically significant class A site means based on three or more samples shown by solid circles (PCA calculations) or solid diamonds (Fisher fit). Class B site means are shown by stars. The interpreation of the palaeomagnetic reversal stratigraphy is given by the light stippling, with the transition between polarity zones plotted as the midpoint between class A or B site means of opposite polarity. This interpretation is repeated at right, with black intervals representing normal polarity, and white intervals representing reversed polarity. The majority of polarity intervals are defined by multiple sites, and many of the single-site intervals are tightly constrained by closely sampled sites.

Johnson *et al.*, 1985) or younger sections to the north (Opdyke *et al.*, 1979) that contain dated volcanic tuffs. One very firm biostratigraphic correlation is available, however. *Hipparion* is a distinctive three-toed perissodactyl that entered the Old World during the late Miocene, but the exact dating of this event has been plagued with uncertainty, thereby bringing into question its utility as a datum (Sen, 1989). However, more recent work (Kappelman *et al.*, 1996; Pilbeam *et al.*, 1996) demonstrates that the *Hipparion* dispersal event was largely synchronous and occurred near the beginning of the long normal interval of chron 5 (i.e. about 10·7 Ma). A *Hipparion* tooth found at 656·5 m in the Mahesian section supports the argument that this level is younger than 10·7 Ma.'

## STRATIGRAPHY

Within the Siwalik Group of the Potwar Plateau, it is normally possible to distinguish relatively thick (tens of metres) sandstone bodies from mudstones that contain relatively thinner sandstones (metres thick) and palaeosols. The nature and proportion of these different rock types varies in different parts of the Potwar plateau. Stratigraphic subdivision of the Siwalik Group has been based historically on a combination of lithological characteristics and vertebrate faunas, thus confusing the proper definition of formations (review in Khan, 1993). Modern lithostratigraphic subdivision in the Potwar Plateau is based, among other things, on the proportion of sandstone plus stratigraphic age (e.g. Fatmi, 1973; Shah, 1977; Willis, 1993b). However, the defining criteria are not used consistently by different workers. For instance, it is uncertain whether only the thick sandstone bodies are included in the calculation of sandstone proportion, and over what stratigraphic interval this calculation is made. Placement of formation boundaries on measured sections is also somewhat arbitrary. This uncertainty confounds lithostratigraphic correlation across the region.

In this study, lithostratigraphic subdivision is based primarily on: the proportion of thick sandstone bodies measured over hundreds of metres of strata; on the thickness, lateral extent and mean grain size of individual sandstone bodies; and on the colour of the mudstone strata. The Mahesian section is subdivided into two 100 m scale sequences (Figs 6–8, 9). The lower, 500 m-thick sequence has a greater proportion (57%) and thicker sandstone bodies than the overlying sequence, and palaeosol horizons are red to orange. The overlying 700 m-thick sequence has 40% sandstone bodies, that are relatively thinner on average, and orange to yellow palaeosols. These lithological characteristics are *generally* similar to those of the Nagri and Dhok Pathan Formations in their stratotype areas, and therefore these terms will be used here without

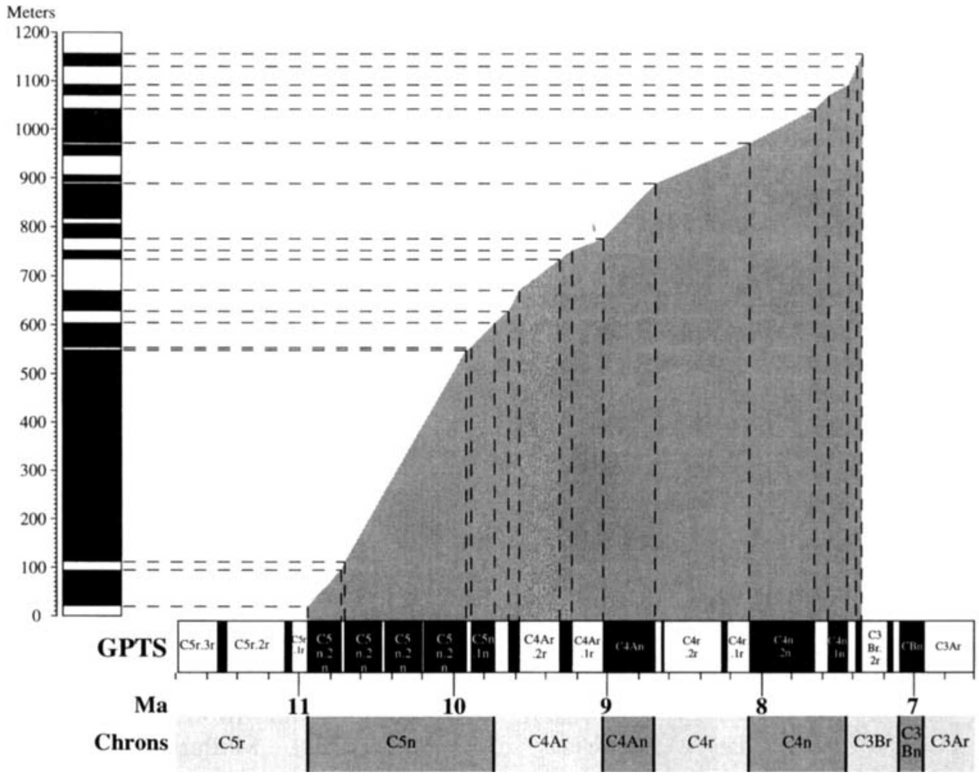



**Fig. 5.** Correlation between the Mahesian section and the geomagnetic polarity time scale (Cande & Kent, 1995) ties the long normal in the lower half of the Mahesian section to chron 5, with the upper intervals to chrons 4Ar, 4r, 4n and 3Br. Compacted sediment accumulation rates decrease from the lower to upper parts of the section.

implying any specific correlation with other areas.

Palaeomagnetic dating indicates that the Mahesian section ranges in age from $\approx$11 Ma to 7·2 Ma, with the Nagri-Dhok Pathan Formation boundary at about 10 Ma. These ages are broadly similar to those of the Nagri and Dhok Pathan Formations in their stratotype areas, but the Formation boundary is diachronous (Johnson *et al.*, 1985; Zaleha, 1994); It is, however, very difficult to assess the degree of diachroneity of the Formation boundary because the boundaries are defined on different criteria in different areas, and because of uncertainties in accuracy of age dates as discussed above.

## DESCRIPTION OF THICK SANDSTONE BODIES

The general character and organization of the strata are illustrated in Fig. 10 along with the descriptive terminology used (according to guidelines proposed by Bridge, 1993a). The strata will be described here mainly with reference to the three detailed lateral sections (Figs 11–13).

The sandstones are olive grey to yellow litharenites and felspathic litharenites (Krynine, 1937; Khan, 1993, 1994; and others) that are mainly medium to very-fine grained. These also include some extraformational conglomerate with pebbles of quartzite, chert and phyllite, and intraformational breccias with clasts up to boulder size of sandstone, mudstone, calcareous nodules, oncolites and vertebrate fossils.

Large-scale inclined strata (Fig. 10) are decimetres to metres thick, extend laterally for up to hundreds of metres, and are bounded by relatively minor erosion surfaces. Each stratum commonly fines upwards. Medium-scale trough cross strata and planar strata (Fig. 10) are the most common internal structures, with small-scale cross strata limited to the finer grain sizes. Soft-sediment deformation is common in the medium-scale cross strata.

Sets of large-scale inclined strata (storeys of other workers) are metres to tens of metres thick (varying laterally) and extend along strike for hundreds of metres. Bases of sets are defined by major, planar to concave-upwards erosion surfaces (that include flute and gutter casts) overlain by breccia or conglomerate. In most cases,

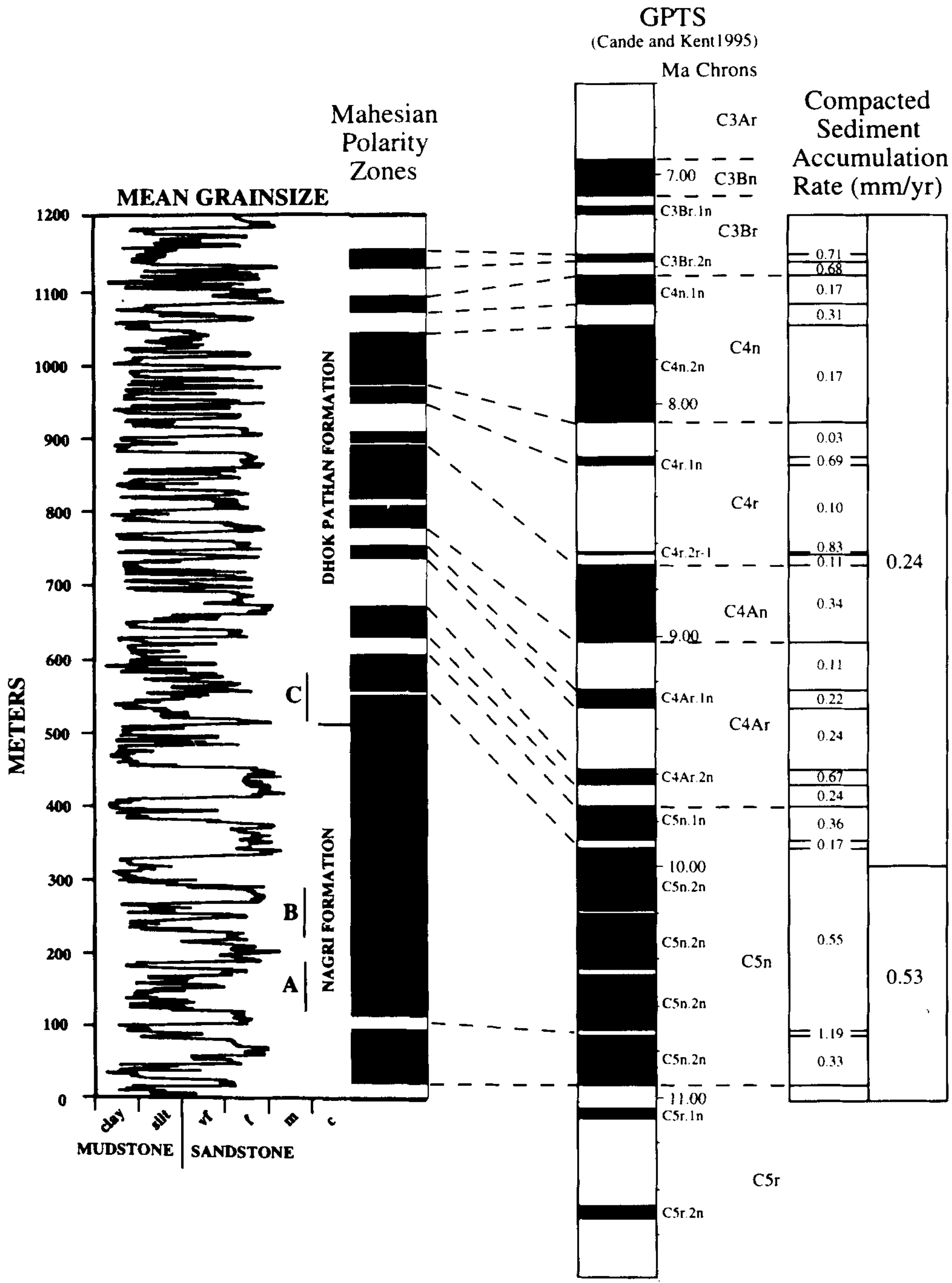



**Fig. 6.** Stratigraphy of the Mahesian section, eastern Potwar Plateau. Mean grain size log is based on samples at 1 m intervals. The Nagri Formation is defined by an upward increase in the thickness and mean grain size of thick (tens of metres) sandstone bodies. The Dhok Pathan Formation has a lower proportion and thinner sandstone bodies. Positions of stratigraphic intervals A, B, and C are marked on the log. The magnetic polarity log is correlated with the magnetic polarity time scale of Cande and Kent (1995), and calculated compacted sediment accumulation rates are shown.

large-scale strata in the lower parts of sets are relatively coarse grained (i.e. medium to fine sand) with medium-scale trough cross strata and planar strata. Higher up, the strata are generally finer grained with more small-scale cross strata, and rare medium-scale planar cross strata. These vertical trends are not present in all cases.

Large-scale strata are inclined at up to 12° relative to the basal erosion surface of the set, with the largest angles occurring where

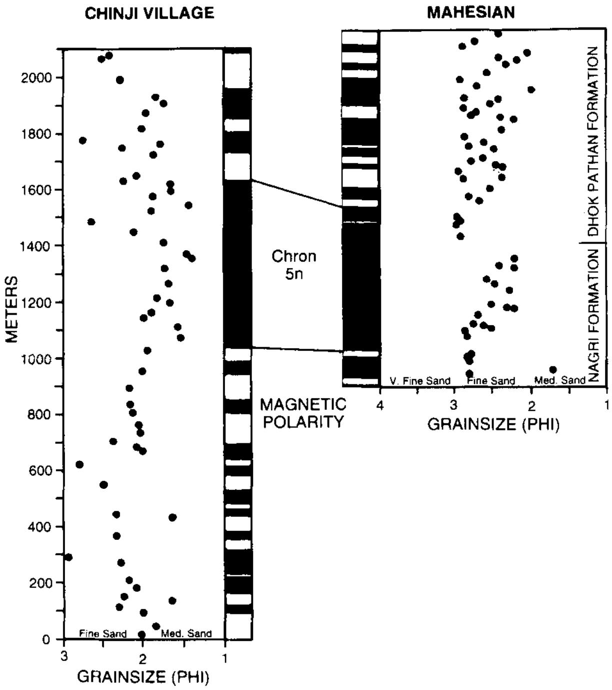



**Fig. 7.** Stratigraphic variation of mean grain size of thick sandstone bodies through the Mahesian section and comparison with that in equivalent age rocks in the Chinji area (Fig. 1B; data from Willis, 1993b).

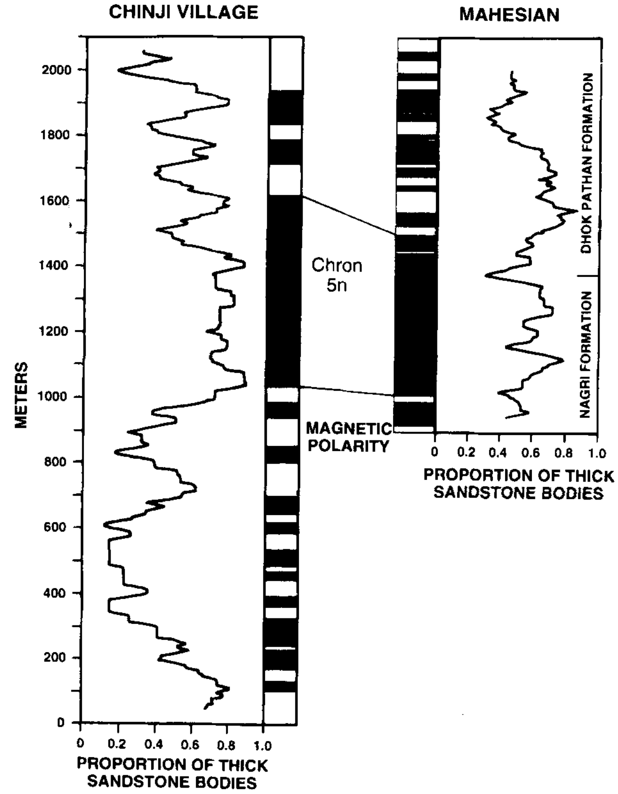



**Fig. 8.** Stratigraphic variation of proportion of thick sandstone bodies through the Mahesian section (based on averaging 50 m of section in successive 5 m steps) and comparison with that in equivalent age rocks in the Chinji area (Fig. 1B; data from Willis, 1993b).

palaeocurrents are perpendicular to the outcrop faces. These inclinations vary both vertically and laterally within a large-scale set, and the inclined surfaces may be planar, concave upwards or convex upwards. Some inclined strata clearly fill channels that have varying degrees of cross-sectional asymmetry. Strata in channel fills tend to be relatively fine grained (at least at the top) with more planar strata and small-scale cross strata than other parts of sandstone bodies. Some parts of channel fills have symmetrical-rippled very-fine sandstone and laminated mudstone.

The thick sandstone bodies are generally 10 m thick, and extend laterally for kilometres (Figs 11–13). They are composed of large-scale inclined stratasets stacked vertically and laterally adjacent to each other. Only the uppermost large-scale sets in these sandstone bodies have untruncated tops, but even these may have relatively small channel fills cutting into their upper parts (e.g. Fig. 11, set D). In general, palaeocurrents vary by less than about 30° throughout a given sandstone body. The differences between the sandstone bodies in different parts of the sequence will be addressed in detail below.

Biogenic structures are generally found in the upper few metres of large-scale stratasets, especially in channel fills, where there may be complete disruption. They also occur in fine grained caps of large-scale strata lower down in sets. A range of different burrow types is present (details in Khan, 1993). Most burrows are single unbranched tubes, circular to oval in cross section, up to 70 mm in diameter, up to 200 mm long, and oriented from vertical to horizontal. Burrow fillings are similar in texture to the host sediment, and may be structureless or meniscate. Root casts that branch and thin downwards, generally have clay linings, and are filled with red-grey siltstone. Elongate calcareous concretions are common around and within root casts (i.e. rhizo-concretions). Compressed wood logs occur at the base of some sandstone bodies. Vertebrate fossils are common in the middle to top of channel fills, and rare in breccias above basal erosion surfaces.

## INTERPRETATION OF THICK SANDSTONE BODIES

By way of an overview, the thick sandstone bodies are interpreted as deposits of sinuous, braided

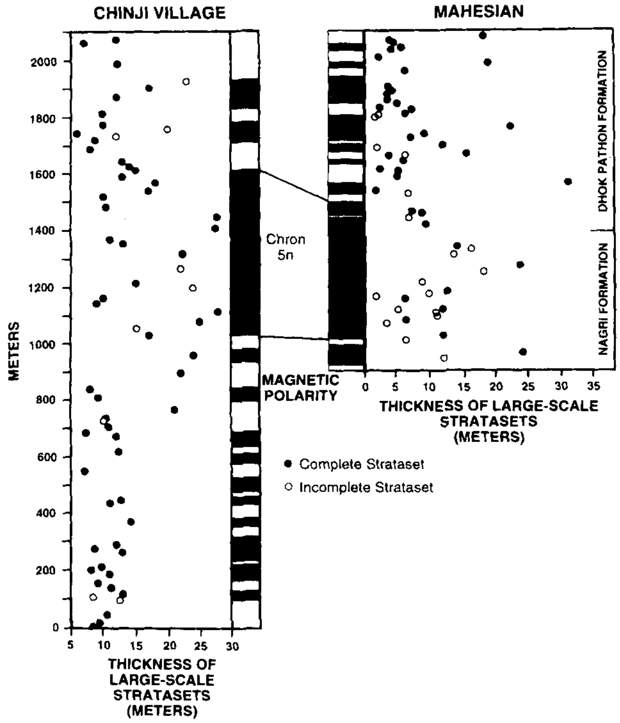



**Fig. 9.** Stratigraphic variation of thickness of large-scale stratasets in thick sandstone bodies through the Mahesian section and comparison with that in equivalent age rocks in the Chinji area (Fig. 1B; data from Willis, 1993b).

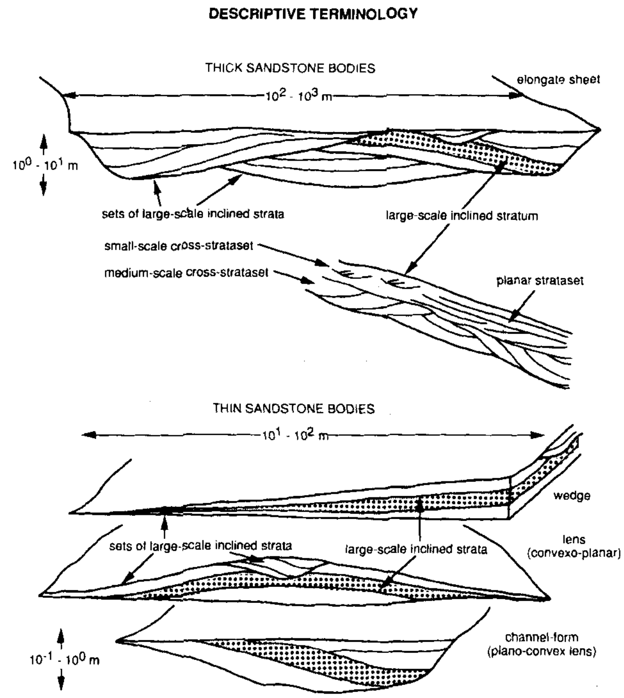



**Fig. 10.** Terminology used for describing the geometry and internal structure of the sandstone bodies. Sections are drawn normal to palaeoflow direction purely for illustrative purposes.

channels that migrated laterally across an alluvial plain or megafan. Large-scale inclined strata represent deposits of single floods, and their occurrence in sets with basal erosion surfaces represents lateral migration of channel bars and the filling of adjacent channels. The stacking pattern of large-scale inclined stratasets represents the movement of channels within single or multiple channel belts.

The medium-scale trough cross strata, planar strata and small-scale cross strata that constitute the large-scale inclined strata record deposition associated with, respectively, dunes, upper-stage plane beds, and ripples. The upward fining and change in interpreted bedforms within the large-scale strata represent a decrease in flow velocity and/or depth as flood waters waned. Soft-sediment deformation indicates rapid deposition, and the bioturbated/disrupted tops of some strata indicate post-flood organic activity.

Basal erosion surfaces of large-scale inclined stratasets represent migration of the deepest, erosional parts or the cut banks of channels. The flute and gutter casts indicate local erosion of cohesive mud exposed in the base of the channels, and the intraformational breccia overlying these erosion surfaces probably represents material eroded from cut banks. Lateral variation in the inclination of basal erosion surfaces and the thickness of large-scale inclined stratasets reflect changes in local flow depth, channel filling or net deposition during channel migration (Bridge *et al.*, 1986; Bridge, 1993b; Willis, 1989, 1993a,c). As the basal erosion surfaces and tops of large-scale sets rise only a few metres over along-strike distances of hundreds of metres, net vertical deposition is much less than lateral deposition (Fig. 11, sets E, F and G).

The vertical trends in sediment texture and internal structure within most large-scale stratasets suggest the presence of sinuous-crested dunes and upper-stage plane beds in relatively coarse sediment in the lower parts of channels during floods, whereas straight-crested dunes and ripples were more common on the upper, finer grained channel-bar surfaces and in-channel fills. The fine-grained, laminated material in some channel fills suggests intermittent lacustrine conditions. The abundance of bioturbation/disruption in the upper parts of large-scale sets, and especially in channel fills, indicates relatively long periods of slow moving water or exposure. Trace fossils indicate the presence of arthropods (including insects), worms and bivalves, and that the river banks and adjacent floodplain were vegetated by trees, bushes and grasses (Khan, 1993).

The dominance of fining-upward large-scale stratasets indicates preferential preservation of the downstream parts of the bars due to a major component of downstream bar migration. Coarsening-upward sets, or those showing little vertical variation in grainsize, represent upstream parts of the channel bars (Jackson, 1976; Bridge, 1993b; Willis, 1993a). Palaeocurrent variation in such bar deposits indicates a relatively low channel sinuosity (Bridge, 1993b). Fining upward trends in channel fills record gradual reduction in flood flow velocity and depth as the channels filled.

Lateral variations in the inclinations of large-scale strata and in lithofacies result from cross sections through channel bars and fills with specific geometries and modes of migration (Bridge, 1993b; Willis, 1993a). Sets A, D and G in Fig. 11 represent sections of bars passing laterally into coarse-grained channel fills, as seen in a section cut approximately perpendicular to flow direction. The convex-upward surfaces dipping in opposite directions in sets B and E indicate cross sections cut perpendicular to flow direction (Bridge 1993b; Willis, 1993a). Sets C and F probably represent small channels cutting across the tops of the bars represented by sets B and E, respectively. The predominance of relatively coarse-grained channel fills suggests that flow was generally maintained in these channels as they filled. This and the clear evidence of mid-channel bars supports a braided river interpretation.

Vertical stacking of large-scale stratasets within the sandstone body in Fig. 11 could represent the superposition during lateral migration of one channel bar over an adjacent bar within a channel belt, or stacking of bars associated with channel switching and cut-off within a channel belt, or superposition of separate channel belts (Bridge, 1985, 1993b; Willis, 1993a). The stacking of set B on set A may be due primarily to superposition of different channel-belt deposits, because of their similarity in thickness, and the almost complete preservation of the lower set. In contrast, the minor sets C and F (representing chute channels) formed within the same channel belt as the major sets they cut. The lateral truncation of set B by set D may represent the downstream migration of a side bar (D) over a downstream adjacent braid bar (B) (Fig. 14). If this is the case, the width of the channel belt containing sets B, D and E is at least 1·3 km, and the numbers of channels in a transect across the channel belt (i.e. braiding index) is at least 2 or 3.

The thick sandstone body in Fig. 12 is exposed normal to slightly oblique to the palaeocurrent direction. The variable but low-angle inclinations of the basal erosion surface and large-scale strata of set A define three laterally adjacent channels with intervening braid bars (B) and (E) (Bridge, 1993b). The maximum depths and widths of the channels are 12 m and 800 m, respectively. The low-angle dips of the large-scale strata and the small variation in palaeocurrent directions indicate relatively low sinuosity channels. The braid bars are represented by eroded remnants of sets B and E. Set F may represent a relatively minor cross-bar (chute) channel. This distribution of channels and bars suggests a channel belt at least 1·3 km across, with a braiding parameter of at least 2 or 3. It is not certain whether the thick upper sets C and D are part of the same channel belt as the lower sets, or whether one or both are part of a separate one. Fine-grained deposits between sets A and C suggest that they might be deposits of separate channel belts. The bedding geometry of set D indicates a coarse-grained channel fill with channel widths and maximum depths of at least 550 m and 7·5 m, respectively.

Large-scale strataset A in Fig. 13 is a section through a single channel bar that migrated to the right and then the channel started to fill. Lateral change in the dip of the large-scale strata is partly due to changing outcrop orientation but may also be related to changes in channel orientation during lateral migration. The width and maximum depth of the channel represented by set B are 114 m and 7 m, respectively. The upward coarsening within this channel fill may be due to either reoccupation of the channel before final filling or progradation of an upstream coarse part of the fill over a downstream finer part (e.g. Bridge *et al.*, 1986). Sets A and B may be part of the same channel belt, because they are not separated by a palaeosol horizon.

Further quantitative evaluation of palaeochannel geometry and hydraulics was accomplished by comparing well-exposed large-scale stratasets within the thick sandstone bodies shown in Figs 11–13 with a physical model that predicts the equilibrium flow pattern, bed topography and mean grainsize within channel bends of regular planform at channel-forming (bankfull) discharge (Bridge, 1977, 1978, 1982). The model has been tested successfully with data from modern single-channel rivers (Bridge

**Table 1.** Quantitative reconstructions of palaeochannel geometry and hydraulics.*

| Large-scale strataset (Figure) | A (Fig. 11) | E (Fig. 11) | A (Fig. 13) |
|---|---|---|---|
| Bar width (m) | 215 | 167 | 156 |
| Bankfull channel width (m) | 323 | 250 | 234 |
| Mean channel depth (m) | 4·1 | 7·6 | 6·7 |
| Maximum channel depth (m) | 9 | 17 | 14 |
| Bend wavelength along channel (m) | 3586 | 2886 | 2664 |
| Bend sinuosity | 1·1 | 1·13 | 1·13 |
| Darcy–Weisbach friction coefficient | 0·065 | 0·068 | 0·06 |
| Dynamic friction coefficient | 0·75 | 0·75 | 0·65 |
| Mean friction slope | 0·000055 | 0·00004 | 0·000047 |
| Mean flow velocity ($ms^{-1}$) | 0·51 | 0·55 | 0·61 |
| Bankfull discharge (cumecs) | 674 | 1058 | 967 |

*Estimated for bankfull flow conditions in single curved channel segments of a braided river.

& Jarvis, 1982; Bridge, 1984b, 1992) and single channels on either side of braid bars (Bridge & Gabel, 1992). It has been used for palaeohydraulic interpretations of many ancient fluvial deposits (Bridge & Diemer, 1983; Bridge & Gordon, 1985; Gordon & Bridge, 1987; Willis, 1989; 1993a,c). Here, the model is applied to the curved channel segments adjacent to braid or side bars. Methodology is explained in the publications cited above. Representative palaeochannel reconstructions are given in Table 1. Individual major channel segments have mean bankfull depths on the order of metres and maximum depths a little more than twice the mean depths. Channel widths are on the order of hundreds of metres, mean bankfull flow velocities are 0·5–0·6 m $s^{-1}$, and bed slopes are on the order of $10^{-4}$. The low sinuosities ($\approx$1·1) are consistent with low palaeocurrent variability and the coarse-grained nature of most channel fills. In view of the observation that 2 or 3 channels may occur side-by-side in these braided channel belts, full bankfull river discharge is reconstructed to be on the order of $10^3$ $m^3s^{-1}$.

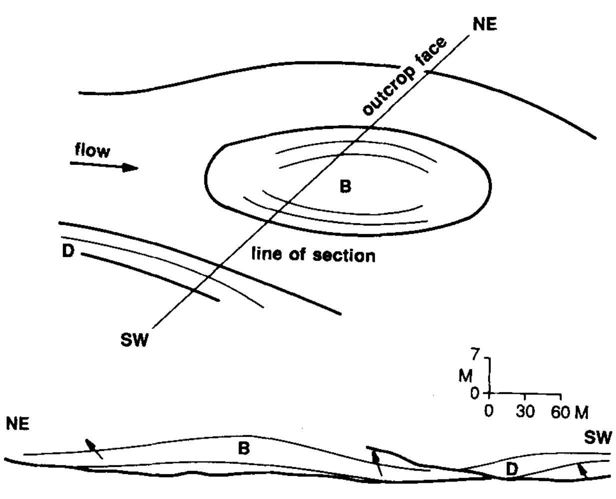



**Fig. 14.** Interpretation of the formation of large-scale stratasets B and D of the thick sandstone body in Fig. 11 in terms of palaeochannel geometry and migration.

## COMPARISON WITH MODERN HIMALAYAN RIVERS

There are broad similarities between the channel geometry, discharge and sediment types of the Siwalik rivers and those of the modern Indo-Gangetic basin (Willis, 1993b; Khan, 1993; Zaleha, 1994; and others). The major rivers of the Indus basin (Indus, Jhelum, Chenab, Ravi, Beas, Sutlej) flow more-or-less parallel to the basin axis (SSW). In their upper reaches, before they converge to become a single channel, they are spaced on the alluvial plain at intervals of 60–100 km (Geddes, 1960). Where they emerge from the mountain belt, these rivers have deposited coalescing sediment fans on the order of 100 km wide and hundreds of km long. The rivers are mainly of low sinuosity and braided, with slopes generally on the order of $10^{-4}$ and bankfull discharges on the order of $10^3$ cumecs (Zaleha, 1994). Recent avulsions of the lower Indus have occurred every few hundred years and have moved its course over 100 km (Geddes, 1960; Holmes, 1968; Schumm, 1986; McDougall, 1989; Jorgensen *et al.*, 1993).

The major rivers that enter the northern side of the modern Ganges basin (Yamuna, Ganges, Ramganga, Gogra, Gandak, Kosi, Mahananda, Brahmaputra, Meghna) emerge from the mountain belt at intervals of 100–200 km, and gradually turn from their basin-transverse orientation to become generally parallel to the basin axis. Nearly all of the rivers form large sediment fans (on the order of 100 km in width and in length) that coalesce downstream. There are also many smaller rivers that may emerge from the mountain belt or originate on the fans, fed by groundwater (Geddes, 1960; Gole & Chitale, 1966; Mukherji, 1976; Parkash *et al.*, 1983; Wells & Dorr, 1987a,b; Gohain & Parkash, 1990; Mohindra *et al.*, 1992; Singh *et al.*, 1993; Sinha & Friend, 1994). Most rivers are of low sinuosity and braided, with braiding index commonly up to 2 or 3. Some of the small groundwater-fed rivers are single-channel and meandering. Slopes are generally on the order of $10^{-4}$, and bankfull discharges of the major rivers are commonly on the order of $10^3$ cumecs, except for their lower reaches where bankfull discharges are on the order of $10^4$ cumecs. Mean grain size of bed sediment decreases down-fan from pebbly sand to medium to fine sand. The smaller, groundwater-fed rivers tend to carry finer sediment. Recent avulsion periods of the larger rivers range from decades to thousands of years. Avulsing channels may occupy the courses of pre-existing channels. The role of tectonic tilting in determining the direction and timing of avulsion is uncertain (e.g. Coleman, 1969; Gole & Chitale, 1966; Wells & Dorr, 1987a,b; Gohain & Parkash, 1990; Mohindra *et al.*, 1992; Singh *et al.*, 1993; Mackey & Bridge, 1995).

The surficial deposits of some of the Ganges basin rivers have been studied (e.g. Sarkar & Basumallick, 1968; Coleman, 1969; Singh, 1972; Singh & Kumar, 1974; Parkash *et al.*, 1983; Wells & Dorr, 1987a; Bristow, 1987, 1993; Singh & Bhardwaj, 1991; Singh *et al.*, 1993), but details are only available for deposits above the water table, a small fraction of the channel deposits. Thus direct comparison of modern deposits with the extensive 2-D sections presented here is futile. Nevertheless, the upper parts of the modern deposits that could be studied are broadly similar to the upper parts of the thick sandstones.

Channel geometries, slopes, discharges and mean grain sizes of the Siwalik rivers are comparable to those of the modern Indo-Gangetic rivers on and between fans. The Siwalik rivers are also taken to be associated with broad sediment fans because of their comparable reconstructed geographical setting, because fans occupy most of the area of the modern Indo-Gangetic plain, and because Siwalik strata generally thin and fine away from the mountain belt (Rao, 1973; Parkash *et al.*, 1980; Cerveny *et al.*, 1989; Burbank, 1992; Willis, 1993b; Zaleha, 1994). The Siwalik rivers of the Potwar plateau flowed generally south-east, but their direction downstream is not known. It is also not known whether they were transverse or parallel to the basin axis, because of the location in the Miocene of the Potwar plateau close to the change in orientation of the mountain belt and foreland basin. As a result of this, it is virtually impossible to test the various theories for the way the orientation of the river systems respond to changes in the nature of tectonic uplift in the mountain belt and subsidence in the foreland basin (see Burbank, 1992; Willis, 1993b).

## DESCRIPTION OF MUDSTONE-SANDSTONE STRATA

Within the mudstone strata, there are relatively thin sandstone bodies as well as distinctive palaeosol horizons. The sandstones are grey to yellow and fine to very-fine grained. They occur in large-scale strata (centimetres to decimetres thick: Fig. 10) that have erosional bases and generally fine upwards to red-brown-yellow mudstone. Intraformational breccia occurs in coarser grained strata or parts of strata. Medium-scale trough cross strata occur in the coarser sandstones, with planar strata and small-scale cross strata in the finer ones. Palaeocurrents derived from these structures are approximately parallel to those in the thick sandstone bodies. The mudstone caps to the large-scale strata are highly disrupted/bioturbated with burrows, root casts and desiccation cracks.

The thin sandstone bodies appear in 2-D sections to have plano-convex, convexo-planar, triangular or parallel-sided shapes (Fig. 10); however, exposures are good enough to establish that in 3-D these bodies are, respectively, channel forms, lobes, wedges, and sheets. Channel-form sandstone bodies are up to 7 m thick and up to hundreds of metres wide normal to palaeocurrent direction (Figs 11–13). Width/thickness ratios range from 20 to 32. These bodies are generally composed of a single set of large-scale strata that are inclined at up to 5° relative to the top of the body. The bodies generally fine upwards and towards their steep outer margins. Associated with this fining is an increased degree of disruption by burrows, root casts and desiccation cracks. Rare impressions of plant axes are concentrated near the tops of these bodies, and their flat tops may be capped by a palaeosol (Fig. 11, sets I and J).

Lobate to wedge-shaped sandstone bodies are decimetres to metres thick and extend along strike for up to hundreds of metres. Large-scale strata within these bodies may be inclined at up to a few degrees relative to the flat basal erosion surface. The bodies generally fine upwards and in the direction of thinning, in concert with an increase in the proportion of small-scale cross strata and degree of disruption/bioturbation. Rarely, these sandstone bodies occur thinning away from the cut bank of a thick sandstone body (Fig. 11, sets E and H; Fig. 12, base of section), and some coarsen upwards.

Mudstone is mainly red-brown-yellow claystone to coarse siltstone. Strata are decimetres to metres thick and extend laterally for up to hundreds of metres: they are generally sheets, but also lenticular. The strata are recognizable by subtle changes in colour, texture and degree of bioturbation/disruption. They generally fine upwards but some show little vertical variation in grain size or coarsen upwards. Where visible, internal structure is mainly planar lamination, with symmetrical and asymmetrical ripple marks and associated cross lamination in coarser parts. Intense disruption associated with burrows (similar to those in sandstone bodies), root casts and different sizes (generations) of desiccation cracks occurs near the tops of strata. Some of the strata are calcareous, and plant axis impressions are common.

The disruption of nearly all mudstone strata by desiccation cracks, roots and burrows indicates a degree of pedogenesis. However, some laterally extensive, metres-thick mudstone layers have distinctive red and orange colors, are intensely disrupted, and contain pedogenic features that indicate that they are mature palaeosols (Figs 11–13; Behrensmeyer & Tauxe, 1982; Johnson *et al.*, 1982, 1985; Behrensmeyer, 1987, 1989; Behrensmeyer *et al.*, 1995; Retallack, 1985, 1986, 1991, 1992; Willis, 1993a; Willis & Behrensmeyer, 1994; Zaleha, 1994). The relatively undisrupted deposits between these distinctive palaeosols were referred to as *palaeosol-bounded sequences* by Willis (1993a) and Willis & Behrensmeyer (1994). Many palaeosols can be divided into two horizons. Upper horizons range in thickness from decimetres to several metres; have abundant desiccation cracks that impart a blocky texture; have mm-diameter burrow tubes and root casts; have mm-scale patches of different colour (mottling) and texture related to burrows, roots and desiccation cracks; have slickensided clay-films (cutans) along areas of clay concentrations that are rarely associated with pseudoanticlines; lack finely disseminated calcium carbonate (no effervescence in dilute HCl) but contain mm-size calcium carbonate and Fe-oxide nodules (some with concentric internal laminae) and calcium carbonate crystals concentrated along cracks, burrows and root casts.

Lower horizons have similar thickness and are dominated by calcium carbonate nodules, but also contain iron oxide concretions, burrows and roots. Isolated soft, diffuse, whitish calcium carbonate nodules occur along mudcracks, burrows and root casts (rhizocretions). Hard carbonate nodules of varying sizes (few cm in diameter), shapes and composition form loosely interconnected networks, and they commonly have distinct borders with the surrounding sediments. The boundaries of smaller nodules within larger nodules are generally defined by thin clay films.

There are also palaeosols that do not have well developed horizons (specifically lacking noncalcareous upper horizons) but have many of the other features mentioned above, such as intense disruption/bioturbation, mottles, slickensided clay-lined surfaces, and concretions (e.g. Fig. 11, logs 1–15, metres 15–25; Fig. 12, logs 1–4, metres 5–15). Zaleha (1994) recognized these palaeosols as being distinctive, referring to them as *calcareous palaeosols*, and presented detailed descriptions.

Palaeosols show significant variation both vertically and laterally in terms of colour, development of cutans and slickensides, concentration of carbonate and Fe concretions, thicknesses and along-strike lateral extent of both the types of palaeosol horizons (Fig. 11; logs 1–15, metres 31–38; logs 25–30, metres 25–30). Some palaeosols converge along strike and form a compound palaeosol (e.g. Fig. 11, logs 13–14, metres 34–35; Fig. 12, logs 1–6, 17–20, metres 25–45). Others gradually thin laterally (e.g. Fig. 12, logs 1–5, metres 14–18). There is no obvious correlation between lateral changes in palaeosols and parent sediment type. In the upper parts of the succession, the palaeosols are less well defined with smaller or absent carbonate and iron oxide concretions (Fig. 13).

## INTERPRETATION OF MUDSTONE-SANDSTONE STRATA

The mudstone-sandstone strata are interpreted as floodplain (overbank) deposits. The channel-form sandstone bodies are interpreted as deposits of either crevasse channels or floodplain drainage channels. The low width/thickness ratio (<35) and large-scale stratal geometry of these sandstone bodies suggest that they represent single-channel rivers with limited lateral migration, that is typical of low-powered, low-sinuosity streams flowing through muddy sediments (e.g. Smith *et al.*, 1989). The limited lateral migration of the channels is possibly because of a short life span, but is more likely due to limited erodibility of muddy banks protected by vegetation and/or soil concretions. The overbank channels are typically 5 m deep and 100 m wide; their sinuosity is difficult to estimate because of the scarcity of exposures.

The distribution of grain size and internal structure within channel-form sandstone bodies indicates episodic flood deposition from migrating dunes and ripples, and on upper-stage plane beds. The predominant bedforms on active channel bars at flood stage were dunes and upper-stage plane beds. Current ripples represent relatively low flow velocities on shallow areas of the bars (particularly during falling discharge) and in channel fills. Mud was deposited from suspension on bars and in channel fills during low flow stages and as the channels became filled. The abundance of desiccation cracks, burrows and root casts in the upper parts of these sandstone bodies suggests major fluctuations in water level, but that water flow was probably perennial. Relative abundance of vertebrate fossils and intense disruption might be related to the presence of congregation areas for large animals (e.g. waterholes; Behrensmeyer, 1988; Behrensmeyer *et al.*, 1995; Wells & Dorr, 1987a).

Lenticular and wedge-shaped sandstone bodies are interpreted as crevasse-splay and levee deposits by virtue of their large-scale stratal geometry and (in places) their relationship to major channels (e.g. Fisk, 1947; Coleman, 1969; Bridge, 1984a; O'Brien & Wells, 1986; Farrell, 1987; Tye & Coleman, 1989a,b; Smith *et al.*, 1989). Large-scale strata record episodic deposition during floods of sand moving as ripples (mainly), as dunes, and on upper-stage plane beds, followed by deposition of mud from suspension and subsequent desiccation and bioturbation as floodwaters receded. Lateral thinning and fining of individual large-scale strata suggest a correlation between deposition rate and flow velocity, both of which appear to decrease away from the major channels. Erosional bases and fining upward trends of these sandstone bodies are evidence for initial erosion and deposition by strong currents and subsequent reduction in deposition rate and flow velocity in successive floods. In contrast, coarsening upwards sandstone bodies with slightly inclined large-scale strata indicate progradation of crevasse splays or levees onto the flood-basin areas over several flooding events. General lack of channels cutting into crevasse splays indicates their deposition mainly from sheet floods.

Mudstones record episodic, suspended-load deposition from slow moving flood flows on broad, low-relief floodplains (i.e. megafans; Parkash *et al.*, 1980, 1983; Wells & Dorr, 1987a), although cross-lamination and ripple marks indicate periodic movement by stronger currents and waves. Floods were followed by intense desiccation, evaporation of carbonate-rich lake- and ground-water, and bioturbation. Well preserved laminae indicate rapid deposition relative to disruption processes and/or inhospitable subaqueous conditions (e.g. hypersaline perennial and ephemeral lakes). Calcareous laminae indicate low siliciclastic deposition rate and possible cyanobacterial mats (Rogers & Astin, 1991). Metres-thick coarsening and fining upwards sequences in mudstone-sandstone strata may be related to progradation or abandonment of adjacent levees and crevasse splays or due to regional changes in overbank flood hydraulics and sediment supply (Bridge 1984a). The upward increase in mudcracks and bioturbation within laminated mudstone sequences may indicate shallowing of the local ephemeral lakes.

The mudstones with blocky texture, abundant root traces and burrows, mottles, slickensided surfaces, upper parts devoid of groundmass carbonate, and lower parts with abundant calcium carbonate and iron oxide concretions are interpreted as mature palaeosols, as stated above. The distinctive reddish and yellowish coloration is due to the presence of ferric iron compounds. Mottling is produced by locally reduced iron oxide associated with root casts, burrows and desiccation crack fills. Development of blocky texture and clay films (cutans) along walls of the blocks reflect formation of peds. Slickensided cutans result from swelling and shrinking during wetting and drying cycles (stress cutans). Some of the thicker cutans may represent downward illuviation of clays. Pedogenic calcite nodules indicate precipitation of calcium carbonate due to evaporation of groundwater, primarily within the capillary fringe. The concentration of carbonate nodules in lower palaeosol horizons is probably related to proximity to the base-flow groundwater table. The relatively larger carbonate nodules that extend vertically show evidence for carbonate precipitation along pre-existing mudcracks and as rhizocretions (e.g. Sehgal & Stoops, 1972; Ahmed *et al.*, 1977; Blodgett, 1988; Joeckel, 1991). The hard discrete nodules with sharp, clay-coated boundaries suggest movement of nodules within the palaeosol due to swelling and shrinking of clays and bioturbation. Iron concretions were formed where water remained ponded after flood events (Sehgal & Stoops, 1972; Brinkman, 1977; Sobecki & Wilding, 1983; Wright *et al.*, 1991). The co-existence of calcite and iron oxide concretions in these palaeosols indicates marked dry and wet seasons, which is consistent with independent evidence of a monsoonal climate during the Miocene.

Non-calcareous upper horizons of palaeosols have been interpreted as being the result of leaching of a calcareous mud (Quade *et al.*, 1993; Willis, 1993a; Willis & Behrensmeyer, 1994; and others) and/or the result of non-precipitation of carbonate (from groundwater or lake water) in mud that originally had negligible amounts of carbonate (Zaleha, 1994). Thus palaeosols with non-calcareous upper horizons were interpreted by Zaleha (1994) as having formed in relatively well drained areas where the upper horizon lay above the capillary fringe of the base-flow groundwater table, whereas the calcareous palaeosols were interpreted as having formed mainly within the capillary fringe in less well-drained areas. However, conclusive evidence that these different types of palaeosol are related to palaeotopography and/or parent sediment type is lacking.

The palaeosols are similar to modern alluvial soils of the Indo-Gangetic basin (Sehgal *et al.*, 1968; Sehgal & Stoops, 1972; Ahmed *et al.*, 1977; Brammer & Brinkman, 1977; Brinkman, 1977; Sidhu & Gilkes, 1977; Sidhu *et al.*, 1977; Beg, 1993; Srivastava *et al.*, 1994). Similar palaeosols within the Siwalik Group in other parts of the basin have been explained as oxisols to alfisols because of their iron concretions and red colouration (Retallack, 1985, 1986, 1992; Johnson *et al.*, 1981). These palaeosols developed under tropical conditions with vegetation ranging from wooded grasslands to savanna vegetation. A humid, woodland setting for the alfisols was supported by the presence of blocky peds, deep root traces and large desiccation cracks (Retallack, 1985).

The thick nodular palaeosols are commonly taken to represent long depositional hiatuses or periods of very low deposition rate (Leeder, 1975; Retallack, 1984; Kraus & Bown, 1986; Willis, 1993a; Willis & Behrensmeyer, 1994; Zaleha, 1994). The average time separating stratigraphically adjacent palaeosols is on the order of $10^4$ years, and Willis & Behrensmeyer (1994) assume that most of this time is taken up in palaeosol formation. Areas where palaeosols are combined must represent even longer hiatuses. It is very difficult to assess, based on the limited lateral extent of the measured sections (up to a few km), whether these palaeosols represent regional (say, megafan-wide or basin-wide) reductions in sediment supply and deposition rate, or whether they represent local areas distant from a sediment supply (i.e. a river channel) or elevated relative to floodwaters. Although there is no conclusive evidence that lateral and vertical variations in the palaeosols are related to local variations in palaeotopography, parent sediment type, or biota, there is some evidence (discussed later) that deposition rate influences the character of the palaeosols.

## STRATIGRAPHIC VARIATIONS THROUGHOUT THE MAHESIAN SECTION

The Nagri Formation has a greater proportion and thicker sandstone bodies than the Dhok Pathan Formation, and the sandstone bodies coarsen through the Nagri Formation from fine to medium sand (Fig. 7). Each formation represents on the order of $10^6$ years of deposition. Figures 6 and 8 reveal at least two smaller scales of stratigraphic sequence: 100-m-scale variations in the proportion and thickness of thick sandstone bodies that represent on the order of $10^5$ years; tens-of-metres scale alternations of thick sandstone bodies and mudstone-sandstone strata that represent on the order of $10^4$ years. The parts of the section with highest proportion of thick sandstone bodies correspond to relatively thick large-scale stratasets (deep channels; Fig. 9) and/or where these stratasets are superimposed (superimposed channel belts). On average, large-scale stratasets are thickest in the Nagri Formation. There is a correlation between the thickness of large-scale stratasets, their mean grain size and amount of conglomerate and breccia (e.g. upper Nagri Formation); however, there is little correlation between the proportion of thick sandstone bodies and mean grain size (e.g. Dhok Pathan Formation). Mean palaeocurrent orientations of large-scale stratasets (derived from gutter and flute casts, medium-scale cross strata and parting lineation in the lower parts of the large-scale sets) are between north-east and south-east throughout most of the section: deviations from this general direction do not seem to be related to any other property of the rocks (Fig. 15). There are

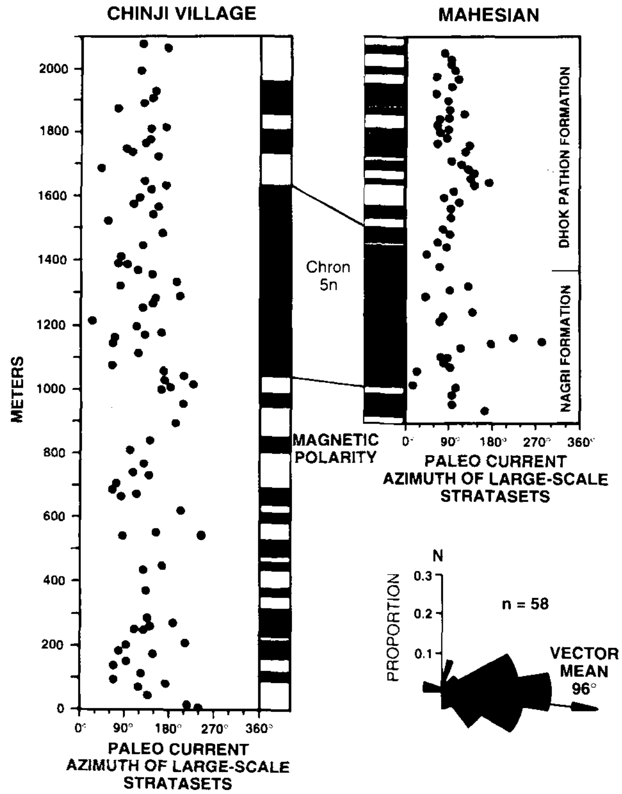



**Fig. 15.** Stratigraphic variation of mean palaeocurrent direction of large-scale stratasets in thick sandstone bodies through the Mahesian section and comparison with that in equivalent age rocks in the Chinji area (Fig. 1B; data from Willis, 1993b).

no statistically significant variations in the mudstone-sandstone strata, except for lower proportions of mature palaeosols, less palaeosol concretions and less well-defined horizons in parts of the section where the proportion of thick sandstone bodies is high (e.g. upper Nagri and middle Dhok Pathan Formations). Compacted deposition rates are 0·53 mm/year for the Nagri Formation and 0·24 mm/year for the Dhok Pathan Formation (Fig. 6). Deposition rates within formations (calculated for the strata between individual magnetic polarity transitions) are quite variable and greatest where sandstone proportion is high (Figs 6, 8). Sandstone and mudstone were apparently both compacted by ≈20%, there being no evidence of differential compaction (Willis, 1993b; Zaleha, 1994).

## COMPARISON WITH STRATIGRAPHIC VARIATIONS IN OTHER AREAS

These stratigraphic variations can be compared directly with those in the Chinji village section studied by Willis (1993b). The upward increase in grain size of the thick sandstone bodies through the Nagri Formation, and the decrease at the top, occur at approximately the same time in both the Mahesian and Chinji sections, although the overall grain size is finer at Mahesian (Fig. 7). Within the lower Dhok Pathan Formation, grain sizes are also finer at Mahesian. The overall decrease in proportion of thick sandstone bodies from the Nagri to Dhok Pathan Formations is similar in both areas (Fig. 8): however, sandstone body proportions are generally greater in the Chinji section (at least for the Nagri Formation) and 100-m-scale sequences do not correlate lithostratigraphically or in time. Large-scale stratasets have a similar range of thickness in both sections and generally decrease in average thickness from the Nagri to Dhok Pathan Formations, but the thickest sets (25–30 m) at Chinji are not represented at Mahesian (Fig. 9). Palaeocurrents at Chinji show a similar variation to those at Mahesian (Fig. 15), bearing in mind that Willis rotated his palaeocurrents by 10° clockwise to correct for the effects of counterclockwise rotation of the Potwar Plateau. Willis reported similar trends in the nature of palaeosols to those seen at Mahesian. Compacted deposition rate for the Nagri Formation in Chinji is ≈0·55 mm/year (based on revised time scale), which is similar to that at Mahesian.

The Nagri and Dhok Pathan Formations are exposed all over the Potwar plateau and beyond (Lewis, 1937; Gill, 1951; Fatmi, 1973; Johnson *et al.*, 1981). Unfortunately, none of the other sections studied in the Potwar Plateau or elsewhere in Pakistan and India are sufficiently detailed to allow close comparison with the Mahesian and Chinji sections (but see forthcoming work of Zaleha, 1994). However, some of the information from other areas is relevant. The Siwalik formations of the Potwar Plateau apparently thin and fine to the south and east, and the base of the Nagri Formation gets younger in the direction of thinning (Gill, 1952; Barry *et al.*, 1985; Beck & Burbank, 1990; Burbank, 1992). Mean palaeocurrent directions are to the south-east throughout this region at this time (Burbank & Beck, 1991; Parkash *et al.*, 1974, 1980). The proportion of blue-green hornblende in heavy-mineral assemblages from sandstones is reported to increase through the Nagri and Dhok Pathan Formations in the Chinji area (Johnson *et al.*, 1985; Cerveny *et al.*, 1989). However, this trend is absent in the Kotal Kund area (about 30 km west of Mahesian), and the appearance of these distinctive amphiboles occurred more than a million years earlier in the

Kohat area west of the Indus (Fig. 1B; Cerveny *et al.*, 1989). It is also note-worthy that there are important increases in sedimentation rate and the proportion of calcic amphiboles in the deposits of the Bengal deep-sea fan between 10·9 and 7·5 Ma (Amano & Taira, 1992).

## EVALUATION OF CONTROLS OF STRATIGRAPHIC VARIATIONS

Interpretation of the stratigraphic variations throughout this sequence is made using knowledge of the general geographical and tectonic setting, our reconstructions of the depositional environments, and evidence from outside the basin of changing climate, sea-level and tectonism. By comparison with the modern Indo-Gangetic foreland basin, the Siwalik basin was probably at least 300 km wide normal to the mountain belt, with deposition dominated by 100-km-scale megafans of major braided rivers. River avulsions must have been an important control of sediment supply to different parts of fans. The area was ≈100–200 km from the mountain belt in the Miocene (Burbank, 1983; and others), that is on the middle to lower parts of megafans, and possibly as much as 1000 km from a coastline (Willis, 1993b; Zaleha, 1994).

Throughout the Nagri and Dhok Pathan Formations, increases in mean grain size of channel-belt deposits are associated with increases in channel size, and to a lesser extent with the preferential preservation of the lower, coarser-grained parts of truncated channel bars and fills. There is no conclusive evidence that changes in grain size are related to changes in channel slope, implying that channel slopes do not change discernibly throughout the sequence (see also Willis, 1993b; Zaleha, 1994). The lack of correspondence between grain size and channel size in some cases is probably a reflection of lateral variations in the grain sizes of channels that experienced periodic changes in discharge. The overall decrease in sandstone grain size from the Chinji and Khaur areas to the Mahesian area in coeval deposits represents a downstream or lateral decrease, as is observed on modern megafans.

In view of the depositional setting of the Siwaliks, it is appropriate to use a certain class of alluvial stratigraphy model (e.g. Bridge & Leeder, 1979; Bridge & Mackey, 1993; Mackey & Bridge, 1995) to help explain the alternations of channel-belt deposits and overbank deposits, and the occurrence of mature palaeosols within the overbank deposits. The proportion of channel-belt deposits increases with the thickness and lateral extent of the channel bar and fill deposits (i.e. with channel and channel-belt dimensions) and with the degree of horizontal and vertical superposition of channel belts. This correlation between channel-deposit proportion and connectedness of channel belts is predicted by the alluvial stratigraphy models, isolated channel belts only occurring where channel deposit proportion is less than about 40%. Using the reconstructed dimensions of channel belts and time-averaged deposition rates, it is possible to predict the range of observed channel-deposit proportions using floodplain widths of 100 km or less and time-averaged avulsion periods of 1000 years (Table 2). Under these conditions, the zones of higher channel-deposit proportion are associated with relatively large channels belts and probably also relatively high average deposition rates and avulsion frequencies (as found by Zaleha, 1994). Clearly it is not possible to obtain unique solutions because floodplain widths and mean avulsion periods are only informed guesses and short-term (less than $10^5$ years) variations in average deposition rate are excluded, as discussed further below.

The general decrease in channel-deposit proportion from the Nagri to the Dhok Pathan Formation can be explained by a general decrease in channel belt dimensions (mainly), deposition rate and avulsion frequency (Table 2). These Formation-scale changes in the proportion of channel-belt deposits, deposition rate and mean sandstone grain size can be seen in other regions of the Potwar plateau and beyond (Willis, 1993b; Zaleha, 1994), although the changes are not exactly coeval. This suggests a regional control. Willis (1993b) explained Formation-scale changes in channel-deposit proportion in the Chinji area as due to the changing interaction between two different river systems. Intervals of high channel-deposit proportion (e.g. Nagri Formation) were interpreted as due to attraction of a relatively large river system and megafan to the area that was previously occupied by a smaller river system on a smaller fan or interfan area. Willis (1993b; Fig. 15) specifically ruled out the possibility that the two different-sized rivers occupied the same fan. Behrensmeyer & Tauxe (1982; & Pilbeam *et al.*, 1979) also interpreted the interaction between two river systems of different geometry and sediment composition from part of the Dhok Pathan Formation in the Khaur area. However, the lack of a bimodal distribution of

**Table 2.** Comparison between observed and simulated alluvial architecture parameters*.

| | | Nagri Formation | | Dhok Pathan Formation | |
|---|---|---|---|---|---|
| | | Observed | Simulated | Observed | Simulated |
| Large-scale strataset thickness (m) | mean | 14 | 15 | 10 | 10 |
| (maximum channel depth) | range | 7–24 | 7–24 | 4–22 | 4–16 |
| Channel-belt width (km) | mean | >1·3 | 4 | >1·3 | 2 |
| | range | — | 2–6 | — | — |
| Floodplain (fan) width (km) | | — | 100 | — | 100 |
| Floodplain (fan) length (km) | | — | 200 | — | 200 |
| Compacted floodplain deposition | mean | 0·53 | 0·58 | 0·24 | 0·21 |
| rate (mm year$^{-1}$) | range | 0·33–1·2 | 0·24–0·93 | 0·03–0·8 | 0·14–0·29 |
| Channel-deposit proportion | mean | 0·55 | 0·55 | 0·4 | 0·4 |
| | range | — | 0·4–0·7 | — | 0·31–0·63 |
| Avulsion period (years) | mean | — | 800 | — | 1300 |
| | range | — | 70–7,000 | — | 100–10,000 |

*Simulated results are based on approximately 50 000 years of simulated time (involving 50 avulsions) using the model of Mackey & Bridge (1995). Mean values for simulated parameters are based on time and spatial averages for the whole floodplain. Ranges of values for simulated parameters are based on average values calculated for several cross-floodplain transects. Means and ranges of values of observed parameters are determined from data in the single Mahesian sedimentological log.

major-channel depths and widths in either the Mahesian, Chinji or western Khaur areas (Fig. 9; Zaleha, 1994) does not support an interpretation of two separate and coexisting river systems of different size in the Nagri and Dhok Pathan Formations. Thus, it is possible that variations in major-channel size in these formations reflect the degree of activity (in terms of the supply of water and sediment) of different rivers on a single megafan. Furthermore, different sediment compositions of different rivers on a single megafan may be related to different sediment sources and/or to different degrees of post-depositional chemical weathering of the deposits. For example, the smaller rivers may be analogous to the groundwater fed rivers of the Kosi fan (Singh *et al.*, 1993; Sinha & Friend, 1994).

Mackey & Bridge's (1995) model predicts marked variations in avulsion period that are related to changes in floodplain (megafan) relief (Table 2). During periods of new alluvial ridge growth in an area, the time separating avulsions may be on the order of $10^3$ years. However, once some threshold floodplain relief is reached, a series of avulsions occurs with periods on the order of $10^2$ years or less. This behaviour may help to explain the large variation in avulsion periods observed in modern Himalayan rivers. This simulated mode of behaviour may explain the 1–10 m-thick palaeosol bounded sequences described by Willis & Behrensmeyer (1994). The well-stratified deposits between the mature palaeosols may represent periods when major channels were actively avulsing in the region, with avulsion periods on the order of $10^2$ years. The mature palaeosols may represent periods of $10^3$–$10^4$ years when the major channel belts occupied a distant region of the megafan and/or when floodplain relief was transiently not conducive to avulsions. As discussed below, this latter stage might also have been related to factors such as local tectonism and/or extrinsic controls on water and sediment supply. If the development of the m- to 10 m-thick sequences is related to the intrinsic processes mentioned above, it is unlikely that they could be correlated across palaeovalleys for distances greater than a few tens of km.

The 100 m-scale sequences also apparently cannot be correlated laterally for distances greater than a few tens of km (Zaleha, 1994), and are difficult to explain. The sandstone-rich parts of these sequences apparently represent times when relatively large channel belts were moving around the area, and avulsion frequencies and deposition rates were relatively high. Such conditions may have been a result of growth and progradation of either a whole megafan or a lobe of a megafan. Zaleha (1994) explained the 100 m-scale changes in channel-deposit proportion in the Khaur area as due to switching of the positions of megafans while the entry point of the major river from the mountain belt remained fixed. Relatively large rivers and high channel-deposit proportions were formed by the large megafans, whereas smaller rivers and low proportions of channel deposits were associated with interfan areas or small fans when the larger fans had switched out of the area.

The basis of the above hypotheses for the explanation of palaeosol-bounded overbank sequences and within-formation variation in channel-deposit proportion is that channel-belt size, avulsion frequency and deposition rate vary in time in a particular area of a megafan due to intrinsic shifting of channel belts and/or megafan lobes. This shifting of the main loci of deposition arises from the evolution of depositional topography and random variations in flood discharges. However, there are a number of alternative hypotheses (see also Zaleha, 1994): (1) temporal changes in water and sediment input to the basin due to factors influencing the hinterland such as climate change, tectonism and geomorphic evolution of the landscape (e.g. Steel *et al.*, 1977; Fraser & Decelles, 1992; Schumm, 1993; Wescott, 1993; Blum, 1994; Olsen, 1994); (2) regional changes in accomodation space due to changes in base level and/or subsidence; (3) local changes in subsidence or uplift due to intrabasinal tectonism (e.g. Bridge & Leeder, 1979; Alexander & Leeder, 1987; Leeder, 1993; Alexander *et al.*, 1994; Mackey & Bridge, 1995). The stratigraphic effects of these changes could be local or regional, as discussed below when considering evidence for changes in climate, base-level and tectonic activity.

Increases in the proportions of certain heavy minerals in parts of the Potwar Plateau and further afield have been interpreted as due to a major phase of tectonic uplift and exposure of metamorphic rocks of the Kohistan island arc terrain in northern Pakistan (Johnson *et al.*, 1985; Cerveny *et al.*, 1989; Amano & Taira, 1992). However, the appearance of sediment derived from a particular source rock does not necessarily imply uplift, just that it was exposed and transported by a particular river system (see also Copeland, 1993). The fact that different heavy-mineral assemblages apparently occur in different regions of Pakistan and India at different times suggests that they are related to the evolving catchment areas of different river systems (Parkash *et al.*, 1974, 1980; Willis, 1993b).

The consistent south-east direction of mean palaeocurrents across the Potwar Plateau and further afield suggests little regional variation in palaeoslope direction over a time span of at least 4 million years, even though there are important changes in deposition rates and facies.

## EFFECTS OF CHANGES IN CLIMATE, BASE LEVEL, AND TECTONISM

### Climate

There is much evidence that the climate in this region during deposition of the Siwalik Group was warm, humid, sub-tropical to tropical, and monsoonal. This evidence comes from the nature of the palaeosols (Quade *et al.*, 1989, 1992, 1993; Cerling *et al.*, 1993; Willis, 1993b; Zaleha, 1994), isotopic studies of marine microfossils (e.g. Wright & Thunell, 1988; Miller *et al.*, 1991; and many others), plant material (Nandi, 1975; Sahni & Mitra, 1980; Singh, 1982; papers in Whyte, 1984; Phadtare, 1989), and climate modelling (Ruddiman & Kutzbach, 1989; Ruddiman *et al.*, 1989; Iacobellis & Somerville, 1991a,b; Raymo & Ruddiman, 1992; Kutzbach *et al.*, 1993; Prell & Kutzbach, 1993). The relatively constant thickness of the horizons of mature palaeosols in different formations (age 15–8 Ma) in the Chinji area was taken to imply constant mean annual rainfall by Willis (1993b). However, studies of the isotopic compositions of palaeosol carbonate nodules and fossil teeth (Quade *et al.*, 1989, 1992; Cerling *et al.*, 1993) suggest a major change in vegetation (from dominantly trees and shrubs to dominantly grasslands) at *c.* 7·4 Ma (but Morgan *et al.*, 1994; put the beginning of the change at 9·4 Ma). This change in vegetation was also associated with major changes in the fauna, with less woodland-dependent fauna and more grazing fauna (Barry *et al.*, 1985; Potts & Behrensmeyer, 1992; Morgan *et al.*, 1994), and a cooler and drier climate. The postulated climate change around 7·4 Ma is not reflected in changes in alluvial architecture, and the consensus appears to be that climatic changes were not important enough during deposition of the Nagri and Dhok Pathan Formations to have a discernible effect on deposition (Amano & Taira, 1992; Willis, 1993b; Zaleha, 1994). However, there appears to be some disagreement now on exactly when the change in vegetation occurred, and perhaps sedimentological evidence for changes in climate is being missed.

There is evidence for accelerated formation of the Antarctic ice cap since *c.* 15 Ma, associated with episodically falling sea level, decrease in atmospheric $CO_2$, and general global cooling (Klootwijk *et al.*, 1992; Potts & Behrensmeyer, 1992; Zaleha, 1994). There is a particularly major eustatic sea-level fall at c.10·8 Ma, near the base of the Nagri Formation (Fig. 16) and a dramatic decrease in atmospheric $CO_2$ from 11 to 8 Ma

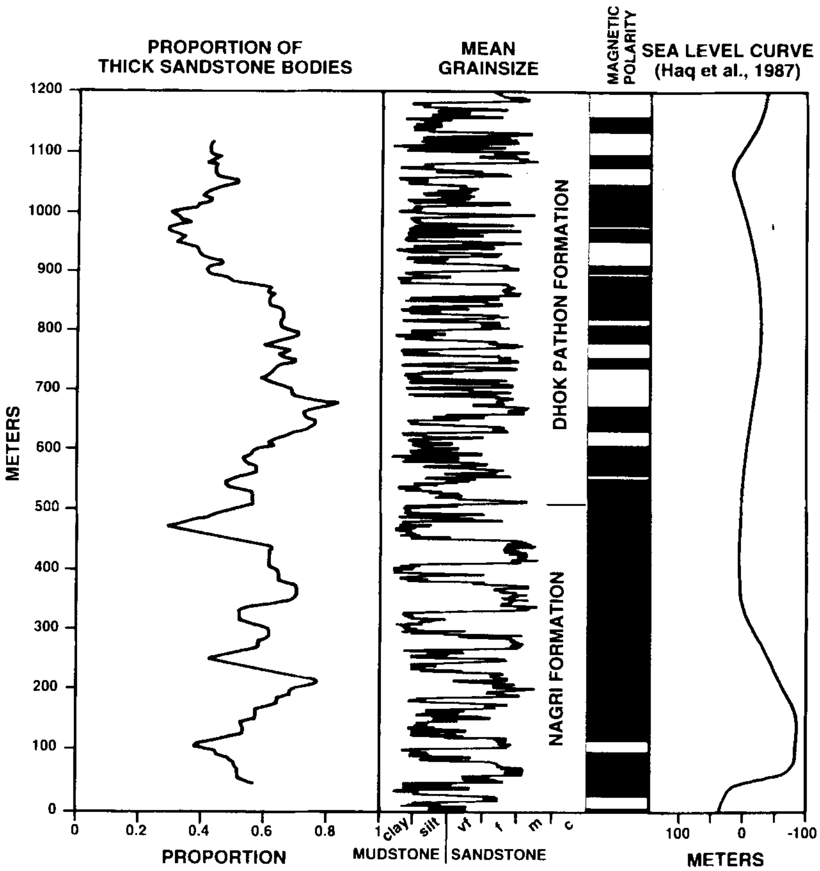



**Fig. 16.** Time variation in relative sea level (from Haq *et al.*, 1987) in comparison with stratigraphic variations in the Mahesian section.

(Freeman & Hays, 1992). It is therefore probable that there was a glacial period during deposition of the Nagri Formation at least, and it is a distinct possibility that the higher Himalayas were glaciated. Such global climate change would not necessarily have a major effect on the climate of the Indo-Gangetic foreland due to its low latitude and elevation. Zachos *et al.* (1993) have suggested similar transient climates in the Oligocene.

Evidence from the modern Indus valley near the Himalayas indicates that aggradation rates increased by an order of magnitude during the last (Pleistocene) glacial advance, and have progressively decreased up to now (Jorgensen *et al.*, 1993). Thus, increasing deposition rates in the Nagri Formation may be linked to increased erosion rates and sediment supply from a partially glaciated hinterland. If this was the case, it is also possible that water and sediment supply could have been linked to Milankovitch climatic cycles (periods of $10^4$–$10^5$ years). Reduced river discharges and deposition rates in interglacial periods may give rise to smaller rivers and better developed palaeosols, and *vice versa* for glacial periods. There may also be a tendency for enhanced uplift of the mountain belt during times of increased glacial erosion and/or melting ice (Molnar & England, 1990; Burbank, 1992). Although such climatic changes should affect the entire Himalayan region, the changing rates of water and sediment supply would not necessarily appear simultaneously in all areas of the foreland basin, depending on tectonism in the hinterland, and changes in the catchment areas of rivers and their positions on megafans. Furthermore, as the best resolution of palaeomagnetic dating is on the order of $10^5$ years, it is doubtful if stratal variations associated with Milankovitch climatic cycles could be correlated in time.

## Eustasy

The eustatic sea-level curve of Haq *et al.* (1987; Fig. 16) for the time period of interest shows My-scale fluctuations that are generally accepted as glacio-eustatic (refs. in Zaleha, 1994). There is much independent evidence for the major, 140 m sea-level fall around 10·8 Ma (Zaleha, 1994); however, sea-level fluctuations at time scales of 0·1 My or less cannot be recognized or verified.

The major fall c.10·8 Ma corresponds *approximately* with the base of the Nagri Formation in the Potwar plateau, and the subsequent sea-level rise and stability correspond *approximately* to the Nagri and Dhok Pathan Formations, respectively (Fig. 16; Willis, 1993b; Zaleha, 1994). There is thus a suggestion that low and rising sea level is loosely associated with high deposition rate, at least in some sections of the Potwar Plateau. Barry *et al.* (1985) have correlated the sea level lowstand at 10·8 Ma with a major faunal turnover (*Hipparion* datum) and immigration of species from other regions facilitated by new land bridges.

Despite the approximate correlation between sea-level changes and Siwalik deposition rate, it is unlikely that sea-level changes directly caused any changes in the architecture of the deposits because the Potwar plateau was up to 1000 km from the shoreline. There are many models now for the effects of relative sea-level changes on alluvial architecture and deposition rate, and substantial disagreement amongst them (e.g. Jervey, 1988; Posamentier & Vail, 1988; Posamentier *et al.*, 1988; Butcher, 1990; Ross, 1990; Miall, 1991; Shanley & McCabe, 1993; Schumm, 1993; Leeder & Stewart, 1995; and others). It is becoming recognized that relative sea-level changes probably do not have as great an effect on the alluvial system as has been assumed in the past (Schumm, 1993; Blum, 1994; Koss *et al.*, 1994). For instance, whether or not a river valley is incised during sea-level fall depends (among other things) on the slope of the exposed shelf relative to that of the river valley (Schumm, 1993; Leeder & Stewart, 1995). In general, effects of sea-level change are expected to be negligible beyond $\approx$300 km from the shoreline in a large river like the Mississippi (Saucier, 1968, 1981; Autin *et al.*, 1991; Schumm, 1993). There is no evidence for an incised valley in Siwalik strata associated with the enormous sea-level fall around 10·8 Ma. However, increased deposition rates on the Bengal fan starting around 10·8 Ma may be associated with increased sediment supply from incision of the exposed shelf. If so, there would be no direct link between uplift of the Himalayas and increased deposition on the fan (*contra* Amano & Taira, 1992).

## Tectonism

The regional, formation-scale variations in deposition rate and alluvial architecture of the Siwaliks are commonly related to tectonism (Parkash *et al.*, 1980; N.M. Johnson *et al.*, 1985; G.D. Johnson *et al.*, 1986; Beck & Burbank, 1990; Burbank, 1992; Amano & Taira, 1992; Meigs *et al.*, 1995 and others). For instance, the change in lithofacies at the base of the Nagri Formation has been attributed to increased regional tectonic loading and uplift of the orogenic belt around 10–11 Ma, and the increase in deposition rate around this time has been related to formation of the Main Boundary Thrust in the western Himalaya (Meigs *et al.*, 1995). This period of tectonic activity was also postulated to have affected sedimentation on the Bengal fan and beyond (Amano & Taira, 1992; Klootwijk *et al.*, 1992). Others argue for episodes of rapid uplift and erosion in different areas at different times and that the overall altitude and relief of the Himalayas has not changed much for the past 18–20 My (Cerveny *et al.*, 1988; Copeland Harrison, 1990; Copeland, 1993). The relative timing of changes in tectonic uplift and erosion of the mountain belt and subsidence and deposition in the foreland basin are very difficult to ascertain, and there is much debate on the issue (Beaumont, 1981; Jordan, 1981; Quinlan & Beaumont, 1984; Tankard, 1986; Gordon & Bridge, 1987; Beaumont *et al.*, 1988; Blair & Bilodeau, 1988; Burbank *et al.*, 1988; Heller *et al.*, 1988; Jordan *et al.*, 1988; Flemings & Jordan, 1989, 1990; Jordan & Flemings, 1991; Sinclair *et al.*, 1991; Fraser & Decelles, 1992; Heller & Paola, 1992; Paola *et al.*, 1992). Zaleha (1994) applied the model of Paola *et al.* (1992) to Siwalik strata and concluded that changes in deposition rate and sediment size at the Formation scale could be explained mainly by changes in rate of sediment supply (and diffusivity to a lesser extent) arising from tectonic uplift and erosion; that is, changes in the nature of deposition are syntectonic. Changes in sediment diffusivity could be related to changes in major channel size.

Details of the timing and areal extent of uplift are difficult to obtain. Episodic regional uplift cannot be explained by variations in the rate of sea-floor spreading and continental collision because these apparently did not vary appreciably over the time interval of interest (Cande & Kent, 1992). Episodic regional uplift (or at least variations in sediment supply) could occur, however, with a glacial-interglacial mechanism. Growth of local deformation structures (faults, anticlines) in the thrust belt may cause river diversions and captures (Alexander & Leeder, 1987; Gupta, 1993) that in turn could result in changes in the size and position of megafans, and in the amount and composition of sediment supplied to them. This is another possible explanation of the diachroneity of changes in channel-deposit proportion across the area.

Episodic tectonism has been invoked by a number of workers to explain 100-m-scale facies variations (Steel *et al.*, 1977; Heward, 1978; Rust & Koster, 1984; Fraser & Decelles, 1992; and others). Quaternary surface deformation in the Himalayas could apparently cause diversions of rivers and their fans with periods of $10^3$–$10^5$ years (Zaleha, 1994). However, the relationship between earthquakes and river avulsions is equivocal (compare Coleman, 1969; Wells & Dorr, 1987b; Schumm, 1986; Mackey & Bridge, 1995).

## CONCLUSIONS

The Nagri and Dhok Pathan Formations of the Miocene Siwalik Group, eastern Potwar Plateau, northern Pakistan, comprise relatively thick (tens of metres) sandstone bodies and mudstones that contain thinner (metres thick) sandstone bodies and palaeosols. Thick sandstone bodies extend for kilometres normal to palaeoflow, and are composed of large-scale stratasets (storeys) stacked laterally and vertically adjacent to eachother. Sandstone bodies represent single or superimposed braided-channel belts, and large-scale stratasets represent channel bars and fills. Channel belts had widths of kilometres, bankfull discharges on the order of $10^3$ cumecs and braiding parameters up to about 3. Individual channel segments had bankfull widths, maximum depths and slopes on the order of $10^2$ m, $10^1$ m and $10^{-4}$, respectively, and sinuosities around 1·1. These rivers are comparable to many of those flowing over the megafans of the modern Indo-Gangetic basin, and a similar depositional and tectonic setting is likely.

Thin sandstone bodies within mudstone sequences extend laterally for on the order of $10^2$ m and have lobe, wedge, sheet and channel-form geometries: they represent crevasse splays, levees and floodplain channels. Mudstones are more-or-less bioturbated/disrupted and represent mainly floodbasin and lacustrine deposition. Mudstones and sandstones are extremely disrupted in places, showing evidence of prolonged pedogenesis. These 'mature' palaeosols are m-thick and extend laterally for kilometres. Lateral and vertical variations in the nature of their horizons depend mainly on deposition rate (but possibly also parent material and floodplain topography).

The 500 m-thick Nagri Formation has a greater proportion and thicker sandstone bodies than the 700 m-thick Dhok Pathan Formation. The thick sandstone bodies and large-scale stratasets thicken and coarsen through the Nagri Formation, then thin and fine at the base of the Dhok Pathan Formation. Compacted deposition rates increase with sandstone proportion (0·53 mm/year for Nagri, 0·24 mm/year for Dhok Pathan), and palaeosols are not as well developed where deposition rates are high. Within both formations there are 100 m-scale variations in the proportion and thickness of thick sandstone bodies (representing on the order of $10^5$ years) and tens-of-m-scale alternations of thick sandstone bodies and mudstone-sandstone strata that represent on the order of $10^4$ years. Formation-scale stratal variations extend across the Potwar Plateau for at least 100 km, although they may be diachronous. However, 100 m and smaller scale variations can only be traced laterally for up to tens of km.

Alluvial stratigraphy models indicate that increases in the proportion and thickness of thick sandstone bodies can be explained by increasing channel-belt sizes (mainly), average deposition rate and avulsion frequency on a megafan comparable in size to those in the modern Indo-Gangetic basin. 100 m-scale variations in thick sandstone-body proportion and thickness could result from 'regional' shifts in the position of major channels, possibly associated with 'fan lobes' on a single megafan or with separate megafans. However, such variations could also be related to local changes in subsidence rate or temporal changes in sediment input to the megafan system.

Climate during deposition of the Siwalik Group was monsoonal, with marked wet and dry seasons. The deposits indicate no *direct* evidence for climate change: however, independent evidence indicates global cooling throughout the Miocene, and the possibility of glacial periods (e.g. around 10·8 Ma). If the higher Himalayas were periodically glaciated, a mechanism would exist for varying sediment supply to megafans on time scales of $10^4$–$10^5$ years. Although eustatic sea-level changes are related to episodic global climatic change, they are not directly related to Siwalik stratigraphic changes (Potwar Plateau was *c.* 1000 km from the shoreline in the Miocene). Formation-scale and 100 m-scale stratal variations are probably also associated with regional and local changes in tectonic uplift, sediment supply and basin subsidence. Increased rates of hinterland uplift, sediment supply and subsidence recorded by the Nagri Formation may have resulted in diversion of a relatively large river to the area. But changing river sizes and sediment supply rates may also be related to climate changes affecting the hinterland (possibly linked to tectonic uplift). In order to sort out these various controls on alluvial architecture, it is necessary to have more detailed sedimentologic-palaeomagnetic studies from other parts of the basin, and independent evidence of tectonic-climatic activity in the hinterland.

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