9 The Himalayan foreland basin

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Abstract

Tectonic loading during the Cenozoic growth of the Himalaya created the Indo-Gangetic foreland basin, a flexural depression that is the largest terrestrial foreland basin on the earth's surface. Stratigraphic, chronologic, petrographic, sedimentologic, and structural data from across the foreland are synthesized here to produce an overview of the Cenozoic evolution of this basin. Within the foreland, variations in crustal rigidity along the strike and the presence of long-lived basement faults have modulated its width and large-scale subsidence patterns. During most of the Cenozoic, the rates of migration of the distal basin pinch-out and of the foreland depocenter suggest that about 20% of the Indo–Asian convergence has been accommodated by thin-skinned thrusting within the foreland, often related to detachments localized at the top of the basement or in incompetent foreland strata, and by thrusts with hanging walls more than 4–10 km thick along its proximal margin. Episodic southward relocation of these basin-margin thrusts has served to define large intermontane basins.

Although the early foreland deposition is poorly documented, ≥ 8 km of Paleogene strata have accumulated in proximal locations and show clear Himalayan derivation. Neogene deposition encompasses the Siwalik Group strata of Miocene and younger age. In northwestern Pakistan and parts of India and Nepal, excellent time controls derived from magnetostratigraphic studies permit detailed correlations, analyses of accumulation and subsidence patterns, and delineation of the interrelationships between deformation and deposition. During early and middle Miocene times, axial rivers, fed by numerous oblique tributaries, appear to have flowed southeastward from northwestern Pakistan into the Ganges drainage. In northwestern Pakistan, a major stratigraphic change occurred at about 11 Ma: Accumulation rates increased 30–200%, sandstone abundances doubled, fluvial discharges increased fivefold and a major influx of detritus occurred, attributed to unroofing of the Kohistan island-arc. Upward coarsening and accelerated accumulation also prevailed in India and Nepal. Those changes are interpreted as having resulted from initial motion and loading by the Main Boundary Thrust at about 11 Ma, approximately 5 m.y. earlier than previously documented.

Proximal portions of the foreland are more abundantly represented within the preserved Upper Neogene strata. As deformation encroached on the foreland, depositional environments and regimes of accumulation and erosion became increasingly heterogeneous. On the basis of scale, petrology, and dispersal patterns, often four river systems can be delineated within the foreland as it became structurally disrupted: an axial river, large transverse rivers with hinterland catchments, small transverse rivers with foreland catchments, and rivers sourced on newly created intra-foreland uplifts. Late Miocene–Pliocene growth of the Salt range at the southern limit of a 100-km-wide salt-lubricated detachment in northwestern Pakistan initially created a large piggyback basin that localized the ancestral Indus River. Continued shortening caused diversion of the Indus to the southwest, coinciding with rapid accumulation and southward pinch-out migration in the Trans-Indus region. Large-scale out-of-sequence thrusting occurred in northwestern Pakistan between 6 Ma and 1 Ma, coeval with deposition in Plio–Pleistocene intermontane basins ponded behind major basin-bounding thrusts. Across the foreland, records of initial conglomeratic strata indicate highly diachronous progradation (ranging from >8 to <1 Ma) that was strongly influenced by fluvial interactions with local folds and thrusts. Whereas tectonic loading appears to continue to control depositional patterns in the Indus foreland, changes in the position of the axial river and in cross-sectional stratal geometries suggest that erosional unloading during the past 4–5 m.y. has become increasingly important in the Himalaya adjacent to the Gangetic foreland. This may be a response to expanding glacial climates. In contrast, the inferred strengthening of the Asian monsoon at about 7–8 Ma does not appear to have led to an increased detrital-sediment flux to the foreland.

Major unresolved problems in the Himalayan foreland include the following: our poor understanding of its Paleogene history; sparse reliable chronologic control in the central and eastern foreland; absence of clear "fingerprinting" that could tie foreland strata to specific hinterland source areas; inadequate seismic analysis of foreland structure; poorly understood climatic influences on deposition and erosion; and uncertainties regarding strain partitioning and its variation along strike within the foreland.

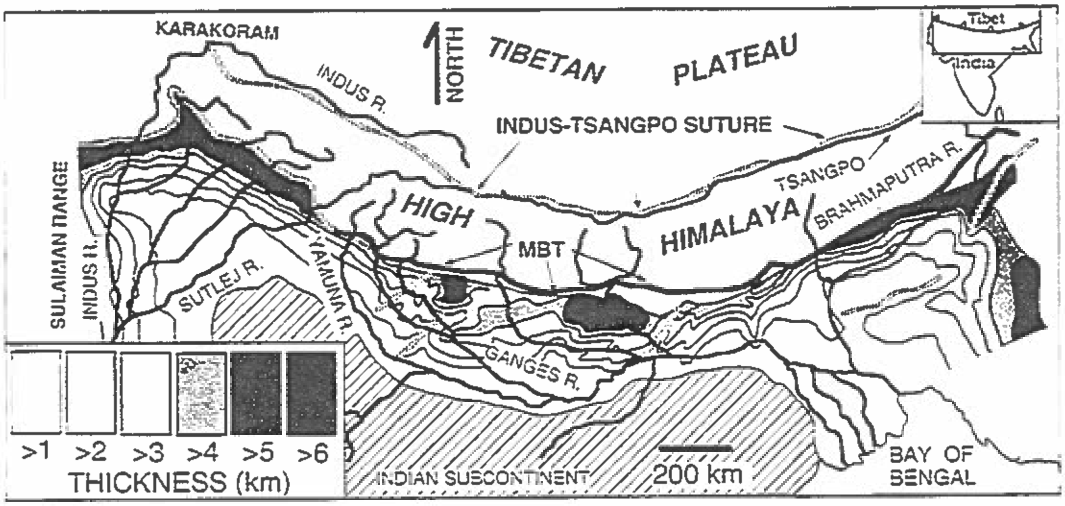


Figure 9.1. Isopach map of sediment thicknesses in the Himalayan foreland basin, showing major rivers, structures, depocenters, and basement faults, based on data from Karunakaran and Ranga Rao (1979) and Raiverman et al. (1983). MBT, Main Boundary Thrust.

Introduction

The Himalayan foreland basin is one of the largest and most dynamic terrestrial basins on the earth's surface. Like a huge garland draped across the northern margin of the Indian subcontinent, it stretches between the syntaxial kinks in the northwestern and northeastern Himalaya before arcing southward to the Arabian Sea in the west and the Indian Ocean in the east (Figure 9.1). Three of the world's largest rivers (Brahmaputra, Ganges, and Indus) traverse it, and during the Cenozoic the progenitors of these rivers and their tributaries deposited a thick succession of strata that record many aspects of the Himalayan collision. Where the Brahmaputra and its tributaries debouch into the Indian Ocean and release their sedimentary load, they have created the largest accumulation of Cenozoic strata known today: the Bengal Fan.

Since the time that the first sediments entered the Himalayan foreland, before 50 Ma, the Indian subcontinent has experienced more than 3,000 km of convergence with Asia (Le Pichon, Fournier, and Jolivet, 1992). This ongoing collision has caused either overthrusting or uplift and erosion of many of the foreland strata, such that only a fragmentary record exists today, particularly for Paleogene times. The Neogene depositional record is considerably closer to being complete, although it remains largely unstudied in many areas of the foreland. In this chapter, we focus on the Miocene-to-Quaternary record of terrestrial deposition in the Himalayan foreland basin. At present, the foreland can be geographically subdivided on the basis of the modern drainage patterns. The Gangetic foreland basin fronts the eastern 75% of the Himalaya, whereas the Indus foreland is situated south of the northwestern ranges of the Himalaya and east of the transpressional ranges that border the western edge of the Indian subcontinent. Herein, we emphasize the evolution of the northwestern Himalayan foreland, where the most nearly complete stratigraphic data set is presently available. Whenever practical, we integrate and compare these data with correlative data from other parts of the foreland. Our objective is to examine (1) the large-scale controls on and geometry of the foreland, (2) the character of deformation and structural styles within and along the margins of the foreland, and (3) the major characteristics of the depositional systems as they responded to and recorded the deformation in the Himalayan hinterland and foreland during the Neogene.

Controls on foreland-basin formation and geometry

Much as in other collisional foreland basins, the downward crustal flexure above which foreland sediments have accumulated was the result of loads emplaced on the crust during convergence. In the Himalaya, overthrusting of the southern margin of Asia, underthrusting of the leading edge of the Indian subcontinent, and imbrication of slices of Indian crust in the Lesser Himalaya have generated a large crustal load, as represented by the Himalaya. Flexure of the Indian plate due to that load and filling of the resultant depression by detrital sediments can account for some part of the large negative gravity anomaly that exists across the foreland (Molnar, 1984; Lyon-Caen and Molnar, 1985). The width and depth of this deflection, which is represented by the foreland, were dependent on the rigidity of the flexed crust and on the placement of the load with respect to the edge of the depositional basin and the loaded plate. In the central Himalaya, crustal rigidities are estimated to be about 0.7×10^{25} N · m (Molnar, 1988), yielding predicted widths of more than 200 km for the Gangetic basin. The rigidity of the Indian plate beneath the Indus foreland in northwestern Pakistan has been estimated to be as much as 10–15 times lower ($\sim 0.4 \times 10^{24}$ N · m) (Duroy, Farah, and Lillie, 1989) than that in central India. Such a contrast in rigidities predicts (Turcotte and Schubert, 1982) that the flexural wavelength of the Indus foreland in northern Pakistan should be about 50% shorter than that of the Gangetic foreland.

In addition to the lateral variations in crustal rigidity at present, it is important to consider the impact of changes in the

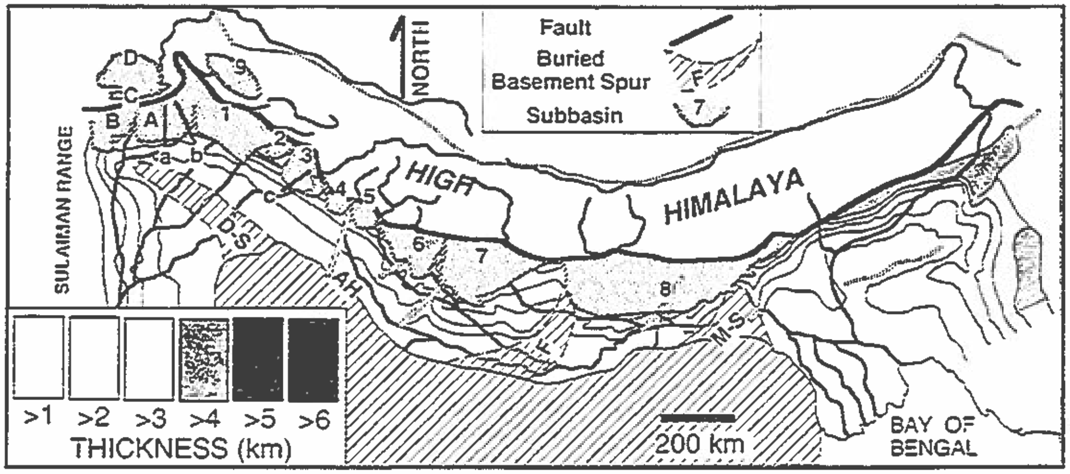
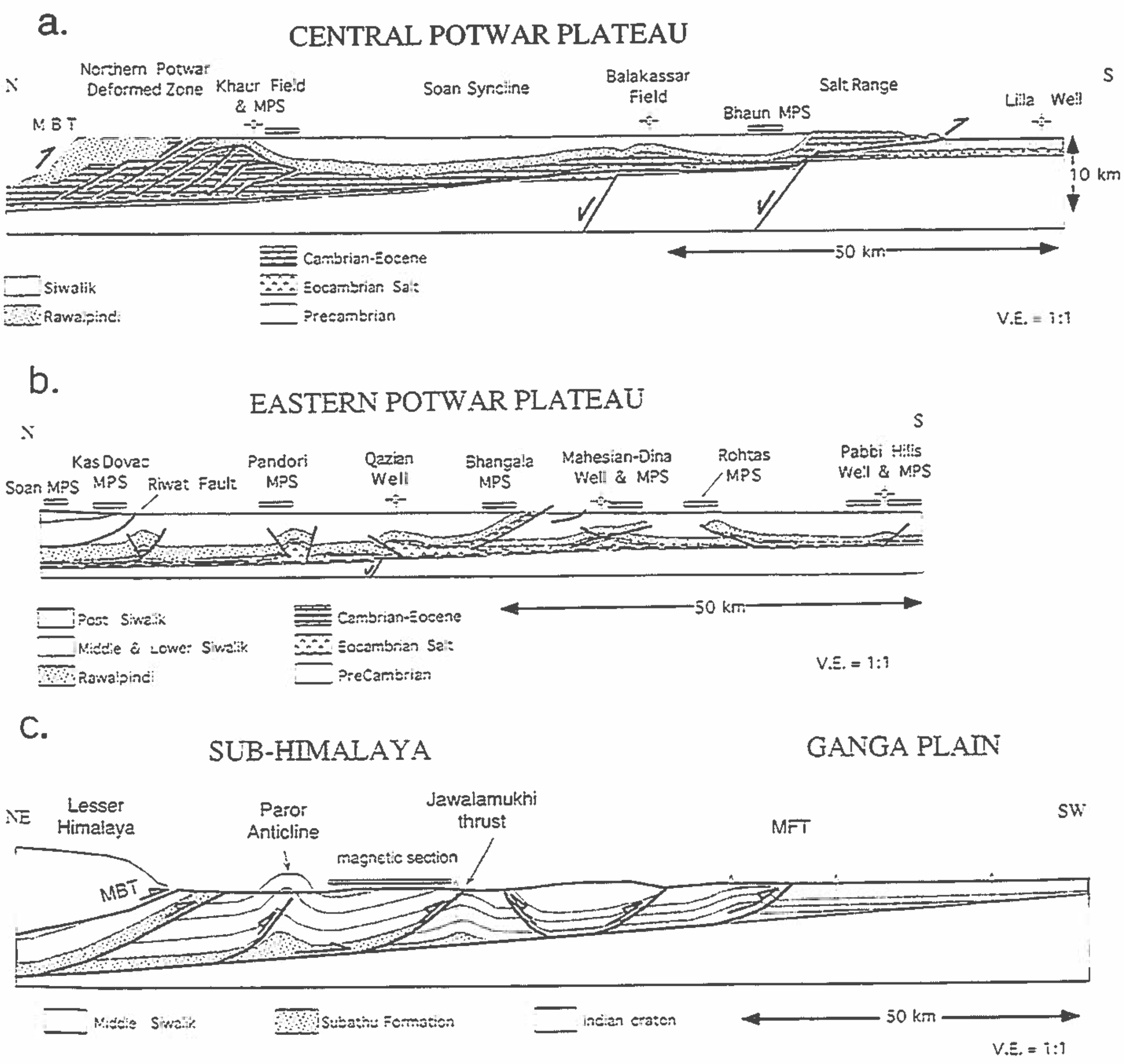


Figure 9.2. Map of Himalayan sub-basins and basement ridges in India and Pakistan, based on data from Raiverman et al. (1983). Sub-basins in Pakistan: A, Potwar Plateau; B, Kohat Plateau and Trans-Indus; C, Campbellpore basin; D, Peshawar Basin. Sub-basins in India: 1, Naoshera basin; 2, Ramnagar basin; 3, Kangra basin; 4, Subathu basin; 5, Dehra Dun basin; 6, Ramganga basin; 7, Sarda basin; 8, Gangak basin; 9, Kashmir basin. Basement ridges: AH, Aravalli horst; D-S, Delhi–Sargoda ridge; F, Faizabad Ridge; M-S: Monghyr–Saharsa Ridge. Locations of cross sections are given in Figure 9.3.

rigidity of the underthrust plate through time. For example, during the early stages of the Indo–Asian collision, the Indian crust that was being underthrust probably was younger than the Precambrian craton that is now being carried into the foreland. Moreover, that crust would have been a passive margin (Powell and Conaghan, 1973) and therefore more highly extended than the modern Indian shield. Both the young age of the crust and the weaknesses introduced by extension should have considerably reduced the rigidity of the underthrust Indian crust. Given a similar load on the plate, a deeper and narrower foreland basin should have resulted (Turcotte and Schubert, 1982). In fact, some of the oldest and most northerly of the preserved exposures of foreland strata (the Eocene Murree Formation in the Kashmir–Hazara Syntaxis) (Bossart and Ottiger, 1989) are also some of the thickest sedimentary successions in the foreland (~8 km).

At a spatial scale much smaller than that of the entire foreland, it is clear that the underlying Indian plate is not homogeneous and that there are variations in crustal strength and in the locations of long-lived basement fractures or weaknesses (Valdiya, 1976). In response to emplaced loads, the differential flexural responses of this variable crust should have produced an irregular basement configuration beneath the foreland basin on scales comparable to the width of the foreland (150–300 km). Additional irregularities in the basement topography would have resulted from transport of the residual erosional topography of the northward-migrating Indian subcontinent from south of the foreland into the region of active foreland deposition. Thus, former highlands in the Precambrian shield and overlying Phanerozoic rocks could become topographic basement highs that might subsequently be onlapped by strata of the migrating foreland. On the basis of geologic mapping, gravity anomalies, and information from reflection seismology and boreholes, the depths of the present foreland strata and the position of the top of the basement have been estimated for much of the Gangetic and northern Indus foreland (Karunakaran and Ranga Rao, 1979; Raiverman, Kunte, and Mukherjea, 1983). Such studies have depicted several basement arches, including the Delhi–Sargodha Ridge, the Aravalli horst, the Faizabad Ridge, and the Monghyr–Saharsa Ridge (Figure 9.2), several of which trend at high angles to the Himalayan foothills. The basement highs have served two functions. First, they have structurally defined the margins of several sub-basins with widths of 100–600 km along the present northern foreland (Figure 9.2) (Raiverman et al., 1983). Although interpretation of the isopach patterns is not simple, these maps depict enhanced sediment thicknesses within many of these sub-basins (Karunakaran and Ranga Rao, 1979). For some of the defined sub-basins (Raiverman et al., 1983), however, there are no mapped deflections in the sedimentary isopachs across the inferred basement highs that bound them (e.g., Dehra Dun and Kangra basins, Figure 9.2). These particular sub-basins seem to be defined on the basis of their association with structural re-entrants along the Main Boundary Thrust, and indeed some of these re-entrants have focused transverse rivers into the foreland during much of the Neogene. Second, during ongoing compression, rheologic contrasts between the basement highs and the adjacent sedimentary fill have apparently concentrated transverse faulting along the margins of the basement blocks (Talukdar and Sudhakar, 1971). Some of these faults extend into the Lesser Himalayan foothills, and they appear to have guided the courses of several major rivers, including the Chenab, Sutlej, Yamuna, and Ganges, as those rivers enter the foreland (Talukdar and Sudhakar, 1971).



Large-scale geometry of the basin fill

Like most foreland basins, the Himalayan foreland, in cross section, can be regarded as a wedge-shaped basin that thickens toward the hinterland. Within this succession, groupings of strata bounded by isochronous surfaces generally also thicken toward the hinterland. Two major stratigraphic subdivisions within the Himalayan foreland are the Rawalpindi Group, which includes the Murree Formation and ranges in age from latest Paleocene/early Eocene to early Miocene, and the Siwalik Group of Miocene and younger age (Figure 9.3). Hinterlandward thickening is not necessarily ubiquitous, however, because the proximal part of any stratal grouping may be either overthrust or uplifted. In the former case, the strata will continuously thicken toward the thrust front, as long as there is little footwall deformation. In the latter case, however, uplifting resulting either from buried thrusts propagating into the foreland (Raynolds and Johnson, 1985; Burbank and Raynolds, 1988) or from isostatic responses to hinterland unloading (Heller et al., 1988; Flemings and Jordan, 1990; Burbank, 1992) will decrease the available space in which sediments can accumulate, and erosion may also thin strata that were deposited earlier. Thus, strata deposited coevally with thrusting, as well as those deposited previously, can display thinning toward the hinterland.

If the Indian plate is used as a frame of reference, then a simplified model of the collision indicates that the mass of the d.

Stratigraphic Nomenclature for the Potwar Plateau Area Molasse Sediments

Age (Ma)	Group	Formation	Lithology	Thickness
0	Upper Siwalik	Soan	Highly variable. Varicolored sandstone, mudstone, and conglomerate. Several volcanic ashes.	0-2000m
4-7	Middle Siwalik	Dhok Pathan	Variable. White-grey to buff-brown sandstones with red-brown siltstones.	400-1600m
7-8		Nagri	White to blue-grey sandstone with subordinate red-brown siltstone.	400-1300m
9-11	Lower Siwalik	Chinji	Red-brown to bright red silts with subordinate white to grey sandstone.	500-1300m
12-14	Rawalpindi-Dharmsala	Kamlial	Brown resistant sandstone with subordinate red-purple siltstone.	400+m
17		Murree	Red-brown to bright red silts with subordinate white to gray sandstone.	0-6000m
≈23-55				

Figure 9.3 Cross sections across (a) the central Potwar Plateau (Baker et al., 1988), (b) the eastern Potwar Plateau and Jhelum re-entrant (Pennock et al., 1989), and (c) the outer part of the Himachal Pradesh re-entrant (Yeats and Lillie, 1991). (d) Stratigraphic nomenclature for the northwestern Himalayan foreland strata (Shah, 1977). Cross sections are based on borehole, seismic, and outcrop data. The locations of magnetostratigraphic sections projected onto the line of section with respect to the structures are shown by open boxes. Note the northward-thickening wedges of molasse strata and the geometric similarity between the Siwalik and Murree clastic wedges. For locations of cross sections, see Figure 9.2.

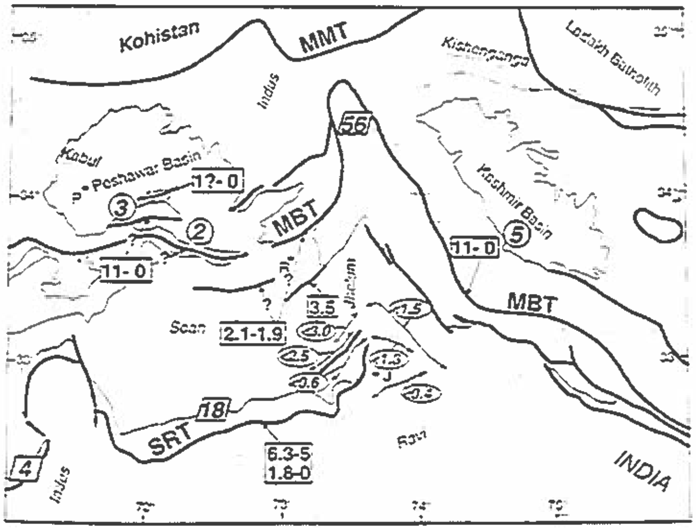


Figure 9.4. Summary of key stratigraphic and structural ages in the northwestern Himalayan foreland. Parallelograms enclose ages of basal strata (distal pinch-out) in the main foreland flexural depression; circles enclose ages of basal strata in intermontane basins; rectangles enclose ages of thrust motion on designated structures; ellipses enclose ages for the beginning of surface expression of folds (Raynolds, 1980). MBT, Main Boundary Thrust system; SRT, Salt-range thrust.

overthrust Himalaya and the flexural depression of the foreland can be envisioned as systematically encroaching on the northern margin of the Indian plate. As a consequence, both the distal pinch-out (Figure 9.4) and the depocenter of the foreland strata will migrate toward the Indian plate with time (Raynolds and Johnson, 1985). During the Eocene, the migration velocity of the depositional axis of the Murree foreland basin in the Jhelum re-entrant may have been 5–10 mm/a (maximum 30 mm/a) (Bossart and Ottiger, 1989), whereas during the Miocene the rate of pinch-out migration varied from about 4 to 12 mm/a in India (Raiverman et al., 1983). This represents only a small fraction of the total Indo–Asian convergence rate and emphasizes the fact that less than 30% of the convergence is likely to have been absorbed through shortening in the Himalaya and the foreland.

At present, two well-integrated river systems, the Ganges and the Indus, account for nearly the entire sediment and water flux across the Himalayan foreland west of the Brahmaputra River. Eastward from the Yamuna River, the foreland appears rather homogeneous. The Ganges is flowing axially to the southeast along the distal margin of the basin and is fed by large transverse rivers (Figure 9.1). Basement spurs or uplifts do not appear to perturb the course of the Ganges, except where it sweeps northwestward prior to crossing the Monghyr–Saharsa Ridge (Figure 9.2). Similarly, except in the proximal foreland, the tributary rivers seem to be uninfluenced by the basement configuration. Given the isopach data that serve to define significant sub-basins (Karunakaran and Ranga Rao, 1979), one might wonder if the Gangetic foreland was more clearly partitioned in the past. Although it is apparent that thinning occurred in the past across the basement arches, there is little evidence for truncation against basement highs. It seems likely that despite differential subsidence and the relative rise of basement spurs, the depositional surface was maintained above the top of the basement, as it is today. Full resolution of this issue, however, must await more seismic and stratigraphic data from transects running parallel to the trend of the range and across the transverse basement elements.

In contrast to the situation for the main arc of the Himalaya, strongly transpressional margins bound the east (Alam, 1989; Holt et al., 1991) and west (e.g., Pivnik and Sercombe, 1993) sides of the Indian subcontinent. The axis of the foreland rotates such that it is approximately north–south and parallel to the bounding ranges. Fold-and-thrust belts extend into and disrupt the proximal foreland basin. Whereas in most cases the trends of these structures are parallel to the foreland, festoon thrust belts such as the Sulaiman lobe occur along these transpressional margins and have no direct analogues in the more orthogonal parts of the collisional zone. In addition to major strike-slip faults, such as the Chaman fault (Lawrence et al., 1981), many of the thrust faults along the transpressional margins also accommodate considerable lateral motion (Nakata, 1989; Pivnik and Sercombe, 1993). One would expect a more complicated provenance history for sites in the transpressional foreland as it was continuously carried northward past different potential source areas.

Structural character of the deforming margin of the foreland

Thin-skinned thrusting in the foreland

Nearly any satellite image of the proximal Himalayan foreland will reveal youthful structures that disrupt the depositional surface and perturb the fluvial systems. Most of these folds and thrusts are nearly parallel to the mountain front, as defined by the southern margin of the Lesser Himalaya or the trend of the Main Boundary Thrust (MBT). Many of the folds form a succession of regularly spaced ridges that control both depositional and erosional patterns in the proximal foreland. Seismic-reflection data from the Potwar Plateau and adjacent areas of northern Pakistan (e.g., Lillie et al., 1987; Pennock et al., 1989) clearly reveal the thin-skinned geometry of thrusting within the foreland (Figure 9.3). The major detachment surface here is localized within the Salt Range Formation, an eo-Cambrian evaporitic unit that underlies the Potwar Plateau and extends eastward into the Jhelum re-entrant. Additional detachment levels here and in the Kohat Plateau occur within Eocene evaporites and within shale of the Murree and Siwalik strata. The large lateral extent and the low yield strength of the Salt Range Formation evaporite have permitted the basal detachment beneath the Potwar Plateau to step about 100 km to the south (Figure 9.3a,b). Very little internal deformation has occurred, and virtually no topography has developed within the hanging wall, which has been transported about 20 km to the south (Baker et al., 1988). Broad, flat synclines, narrow, boxlike anticlines, and doubly vergent thrusts suggest that evaporite also underlies the western part of the Jhelum re-entrant (Jaumé and Lillie, 1988; Pennock et al., 1989), where deformation has also extended more than 100 km south of the MBT (Figure 9.3b).

Whereas the foreland strata in northern Pakistan rest on a succession of Phanerozoic strata that become thinner toward the east, most of the foreland strata east of the Jhelum River pinch out directly against Precambrian basement (Figure 9.3c). In Pakistan, the angular taper between the hindward-dipping basement and the forelandward-dipping topographic surface is about 3° (Yeats and Lillie, 1991), and the width of the deformed belt is 100–150 km. In contrast, in India the deformed belt is 30–75 km wide and has a taper of 5–7°. The unusual width of the deformed zone in northwestern Pakistan indicates the important control exerted by the strength of detachment horizons on the width of deformation.

Several previous analyses of seismic and borehole data in the Indian foreland have interpreted thrust faults there to be basement-involved faults that cut the basement at high angles (Talukdar and Sudhakar, 1971; Karunakaran and Ranga Rao, 1979; Raiverman et al., 1983; Le Pichon et al., 1992). Recent field studies and reinterpretations of the seismic lines (Figure 9.3c), however, suggest that an important detachment occurs near the top of the Precambrian strata and that additional ones occur within the Dharmsala and Subathu strata (Srivastava and Mitra, 1994). If correct, such an interpretation would suggest that deformation across the entire Himalayan foreland is dominated by thin-skinned thrusting (Yeats and Lillie, 1991; Yeats et al., 1992).

Thrusting along the foreland margins

Whereas no thrusts involving crystalline basement appear to impinge on the foreland, we draw a distinction between the thin-skinned thrust sheets within the foreland and many of the thrusts along its margins. The hanging walls of thin-skinned thrust sheets in the foreland comprise pre-tectonic strata only a few kilometers thick and have generally experienced less than 10 km of shortening (Figure 9.3). Along the basin margins, in contrast, many thrust sheets have hanging walls comprising 4 km to more than 10 km of pre-tectonic strata (Fuchs and Sinha, 1978; Schelling, 1992; Srivastava and Mitra, 1994) and have experienced large displacements (>20 km). In general, the basal detachment surface steps upward to progressively shallower stratigraphic levels toward the south. Within the foreland, the detachment commonly lies near the base of the Phanerozoic section and steps up into the molassic units, whereas the basal detachment typically is situated deeper within the Precambrian strata in the marginal thrusts.

Interactions among at least four factors determine the impact of developing thrusts on depositional and erosional patterns within and adjacent to the foreland. First, the erodability of the hanging wall will determine the limiting rate at which thrust-generated topography can be lowered by prevailing surficial processes. Whereas foreland molasse strata can be eroded nearly as rapidly as they are elevated above the local base level (Burbank and Beck, 1991a), most pre-tectonic strata are considerably more resistant to erosion. Second, the magnitude of displacement and the geometry of ramps will determine whether or not resistant strata will be carried to the erosional surface. Magnitude of displacement is also generally proportional to fault length. Third, the rate of displacement on a thrust will determine how rapidly hanging-wall strata are uplifted above footwall ramps. In order to maintain a pre-thrusting river course or to bevel off newly emergent topography, stream power has to be sufficient to keep pace with the rate of bedrock uplift. Fourth, the thickness of the pre-tectonic hanging wall will influence the spacing of thrusts (Dixon and Liu, 1992; Marshak, Wilkerson, and Hsui, 1992), because of a tendency for the spacing of folds and faults to be proportionate to the thickness of the "beam" being deformed. Thus, thin-skinned foreland thrusts tend (1) to have significant thicknesses (0.5–4 km) of readily eroded, syntectonic molassic strata in their hanging walls, (2) to have insufficient displacements to bring resistant but buried strata to the surface, (3) to create less topography, because of erosion of uplifted syn-tectonic strata, and (4) to create narrower thrust-top basins, because of closer thrust spacing. In contrast, thrusts with thicker pre-tectonic hanging walls and greater displacements tend (1) to bring thick successions of resistant strata to the erosional surface, (2) to generate higher hanging-wall topography, and (3) to create wider intermontane or thrust-top basins. The topography resulting from thin-skinned thrusts can mimic that due to thicker-skinned thrusts if sufficient shortening and imbrication or duplexing occur or if there are resistant strata.

When encountering thin-skinned foreland thrusts and folds, rivers generally experience simple deflections around short-wavelength folds (5–15 km) or else are able to maintain their courses across the uplifting but easily eroded strata. In contrast, emergent and resistant hanging walls typically cause a major reorganization of preexisting drainage patterns. Whereas thin-skinned thrusts can create shallow depressions ("duns") that will fill during ongoing deposition from deflected rivers, thrusts with thicker hanging walls can create major intermontane basins, within which significant ponding of previously through-flowing fluvial systems can occur. When there is structural and topographic closure due to thrusting, such basins may fill with lacustrine strata that can accumulate up to several kilometers in thickness. The Kashmir and Peshawar basins in the northwestern Himalaya are examples of intermontane basins bounded by thrusts with relatively thick hanging walls and filled with fluvial and lacustrine strata (Burbank, 1983a; Burbank and Johnson, 1983).

In most cases, the MBT system is the most prominent example of a thrust with a thicker pre-tectonic hanging wall adjacent to the northern foreland. Although the hanging wall of the MBT itself generally comprises Mesozoic and/or Paleozoic strata, it is often in close proximity (1–10 km) to a hinterlandward thrust carrying a thick succession of Precambrian meta-sedimentary rocks in its hanging wall, such as the MBT and the Cherat fault in Pakistan (McDougall, Hussain, and Yeats, 1993) or the MBT and the Panjal thrust in Kashmir (Wadia, 1931). It is likely that these pairs of thrusts merge into a common MBT detachment at depth. In the northwestern Himalaya, the thick stack of imbricated, resistant thrust sheets associated with the MBT has generated considerable topographic relief, which defines large intermontane basins.

Paleogene foreland deposition

The deposition in the Himalayan foreland during the Paleogene illustrates some of the characteristics of the initial stages of the Himalayan collision. The Murree and Balekot formations in Pakistan (Critelli and Garzanti, 1994), the Dharmsala and Dagshai formations in the northwestern Indian foreland (Chaudhiri, 1975; Raiverman et al., 1983; Najman et al., 1993), and the Chulung La and Kong formations in Ladakh (Critelli and Garzanti, 1994) compose the oldest preserved record of predominantly detrital deposition. Unfortunately, the records of these early stages of foreland deposition are fragmentary and incomplete. In most of the Gangetic foreland basin there are no preserved foreland strata of Paleogene age. One interpretation of this absence would be that the collision had not proceeded sufficiently far in the east to create a flexural basin in which such sediments could accumulate. We prefer an explanation suggesting that these strata were overthrust during the extensive shortening along the MBT and other thrusts of the Lesser Himalaya. Such a contention gains support from those locations where early foreland strata are most extensively preserved in the western foreland: within structural re-entrants that permit views farther into the hinterland than are possible along much of the range front. Here, thick deposits comparable to the overlying Siwaliks of Neogene age are preserved. For example, near the apex of the Jhelum re-entrant, about 8 km of Murree strata are preserved in steeply plunging folds and thrust sheets (Bossart and Ottiger, 1989). As determined on the basis of intercalated shallow-marine fauna, deposition of the lower two-thirds of these detrital strata occurred during latest Paleocene to middle Eocene times (Bossart and Ottiger, 1989). In the Himachal Pradesh re-entrant (Kangri subbasin), about 2 km of Dharmsala strata are preserved in some of the innermost thrust sheets (Raiverman et al., 1983). Between these two re-entrants there are extensive regions of folded Murree strata up to several kilometers thick within the proximal foreland (Raiverman et al., 1983). The argument for extensive overthrusting of the Murrees is also supported by seismic travel-time data from northern Pakistan west of the Kashmir–Hazara Syntaxis (Ni, Aomar, and Roecker, 1991). These data are interpreted as showing that a thick succession of low-velocity rocks (possibly Murree strata) has been buried beneath an overthrust that has moved more than 100 km.

Early studies of the Murrees and related rocks suggested that they were derived from the south (Gansser, 1964) and essentially predated the formation of the Himalayan foreland. More recent studies, however, have demonstrated unequivocally that these strata thicken northward and have Himalayan provenance (Karunakaran and Ranga Rao, 1979; Raiverman et al., 1983). In the Jhelum re-entrant, Paleogene Murree strata comprise approximately 20-m-thick upward-fining cycles of meandering tidal channels and tidal sandflats and mudflats. Sub-aerial exposure of interchannel and supratidal areas is indicated by pedogenic carbonates and gypsiferous horizons (Bossart and Ottiger, 1989).

Petrographic studies of Murree sandstones from the Kashmir–Hazara Syntaxis (Critelli and Garzanti, 1994) indicate primary derivation from low-grade meta-sedimentary rocks, with lesser quantities coming from volcanic, sedimentary, and ophiolitic rocks. The absence of high-grade metamorphic rock fragments implies that (1) Murree strata could not have been derived from the metamorphosed rocks in cratonal India to the south, and (2) uplift within the youthful Himalaya was insufficient to expose abundant high-grade rocks to erosion. The absence of high-grade Himalayan sources is perhaps surprising, given the Eocene ages for peak metamorphism and magmatic generation in several areas of northern Pakistan (Treloar et al., 1989; Chamberlain, Zeitler, and Erickson, 1991; Smith, Chamberlain, and Zeitler, 1992; Spencer, 1993) and given the Eocene ages for cooling below approximately 200°C for several of these same areas, as interpreted from fission-track ages (Zeitler, 1985).

In northwestern India, Eocene Dharmsala strata are characterized by upward-fining and -coarsening cycles of finely laminated sandstone and by carbonaceous shale with abundant plant debris. These rocks are also interpreted as representing intertidal flats and tidal channels (Raiverman et al., 1983). The overlying Dharmsala succession of probable Oligocene age, contains multistoried, amalgamated, upward-fining sand bodies that suggest fluvial deposition. Grain sizes in these strata coarsen northward and contain pebble lenses (Raiverman et al., 1983) in the Ramnagar sub-basin (Figure 9.2). Paleo-currents from Oligocene strata in the several sub-basins indicate rivers entering the foreland from the north, northeast, and northwest (Raiverman et al., 1983; Srivastava and Casshyap, 1983). Strata of the pre-Miocene Dagshai Formation are about 350 m thick in Himachal Pradesh and are correlative with the Dharmsala rocks. These red, upward-fining sandstones are interpreted as having been deposited by meandering rivers on a wide floodplain (Najman et al., 1993). Paleo-currents indicate deposition from rivers flowing to the southwest and southeast. Similar to the Murrees in the Jhelum re-entrant (Critelli and Garzanti, 1994), the lithic grains and dense mineral contents of the Dharmsala strata in northwestern India (Chaudhiri, 1975) indicate northern derivation dominantly from low- and medium-grade metamorphic rocks, and locally from high-grade metamorphic rocks, in the pre-Miocene Himalaya (Raiverman et al., 1983; Najman et al., 1993). At the same time, Chulung La deltaic red beds were deposited in a "thrust-top basin" to the south of the suture zone (Critelli and Garzanti, 1994). These sandstone strata include abundant volcanic detritus, suggesting a provenance from an uplifted Andean arc–trench system.

Because the early foreland basin has been dismembered by subsequent thrusting, and commonly either overthrusted or eroded, relatively little is known about this interval. There have been few reliable chronostratigraphic studies, and age-diagnostic faunal or floral assemblages are uncommon. No reliable comparisons of coeval strata within the foreland at time scales of less than 3–5 m.y. are possible at present. This Eocene–Oligocene interval represents a fundamental gap in our present knowledge of the evolution of the Himalaya and presents a major opportunity for future research.

Neogene foreland deposition

Stratigraphy

Throughout much of the foreland basin, a three-fold subdivision of Siwalik strata has been utilized. Lower Siwalik strata usually contain thinner and less amalgamated sand bodies than do Middle Siwalik rocks (Willis, 1993a). The upper Siwalik strata typically coarsen upward and are less indurated than underlying strata. Preserved middle and lower Siwalik strata typically record deposition only in the medial and distal parts of the foreland. The correlative record of proximal deposition usually has been either overthrusted or eroded. Upper Siwalik strata are present across nearly the entire foreland, but they are best exposed in the thrust sheets and folds of the proximal foreland.

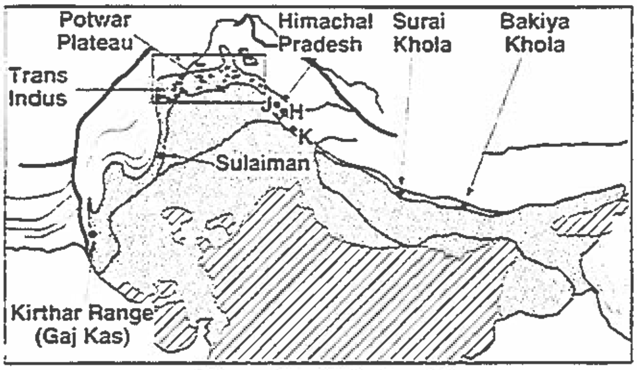
Further stratigraphic subdivisions of the Siwalik Group (Figure 9.3d) have been utilized in many parts of the foreland (Shah, 1977). We focus much of the following discussion on the Pakistani foreland, where the Kamlial Formation represents the upper part of the Rawalpindi Group, and the Chinji Formation represents the lower Siwaliks (Shah, 1977). The middle Siwalik strata comprise the Nagri (older) and Dhok Pathan (younger) formations, and the upper Siwaliks are represented by the Soan Formation or the "Boulder Conglomerate."

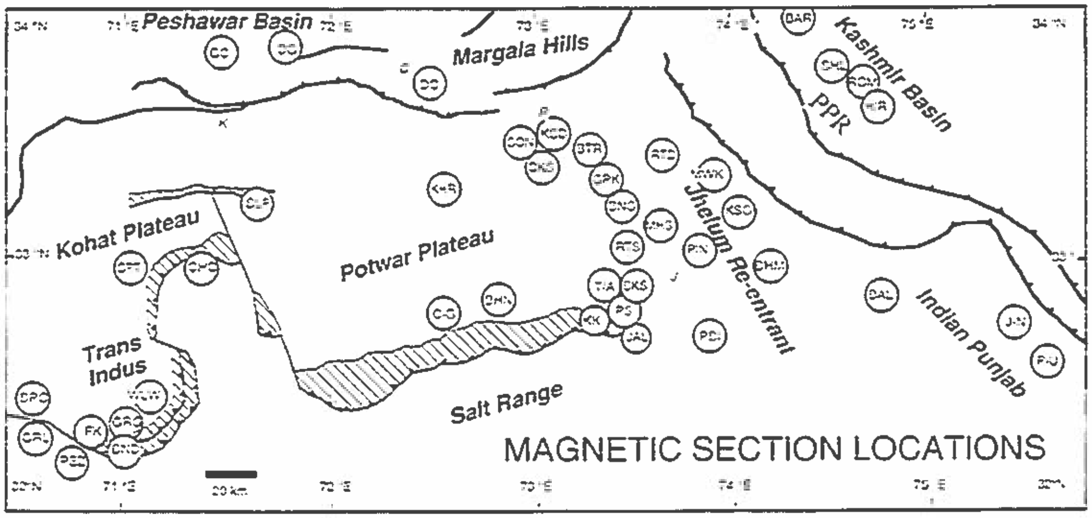
Chronology of deposition

In contrast to the Eocene–Oligocene strata in the foreland, considerable chronostratigraphic data exist for the Neogene strata of the foreland. Early discoveries of numerous vertebrate fossils, including hominids, in Siwalik strata led to intensive fossil prospecting and development of detailed biostratigraphies for several key localities (e.g., Pilbeam et al., 1977; Johnson et al., 1983; Barry et al., 1985). Subsequently radiometric dating of volcanic ashes (Johnson et al., 1982a) and an extensive and ongoing program of magnetostratigraphic sampling have created a broad array (Figures 9.5 and 9.6) of well-dated stratigraphic successions (Johnson et al., 1979, 1982b, 1983, 1985; Keller et al., 1979; Opdyke et al., 1979; Raynolds, 1980; Tauxe and Opdyke, 1982; Azzaroli and Napoleone, 1982; Burbank and Johnson 1983; Khan et al., 1984; Tandon and Kumar, 1984; Burbank and Tahirkheli 1985; Raynolds and Johnson, 1985; Tokuoka et al., 1986; Khan, Opdyke, and Tahirkheli, 1988; Ranga Rao et al., 1988 Appel, Rösler, and Corvinus, 1991; Friedman et al. 1992; Harrison et al., 1993). Although the density of dated sections is greatest in northern Pakistan, ongoing magnetostratigraphic studies are providing an expanding data base in parts of India and Nepal.

Several volcanic ashes that erupted from volcanic sources in Afghanistan between 2 Ma and 3 Ma and spread across the northwestern foreland between the Kohat Plateau and Kashmir (Johnson et al., 1982a; Burbank and Johnson, 1983) have helped to identify the Gauss-Matuyama magnetic chron boundary (~2.46 Ma) (Cande and Kent, 1992) within many Plio–Pleistocene sections of northern Pakistan (Visser and Johnson, 1978) and the Indian Punjab (Ranga Rao et al., 1988). In this same region, additional ashes occur in several Siwalik sections at about 10 Ma and 1.6 Ma (Johnson et al., 1982a).

Given that paleomagnetic data are extensively used to provide precise dates for the stratigraphic record, it is important to consider the uncertainties that are inherent in these data. First, because many of these rocks are terrestrial red beds, there is a possibility that the magnetic polarity recorded by a stratum will





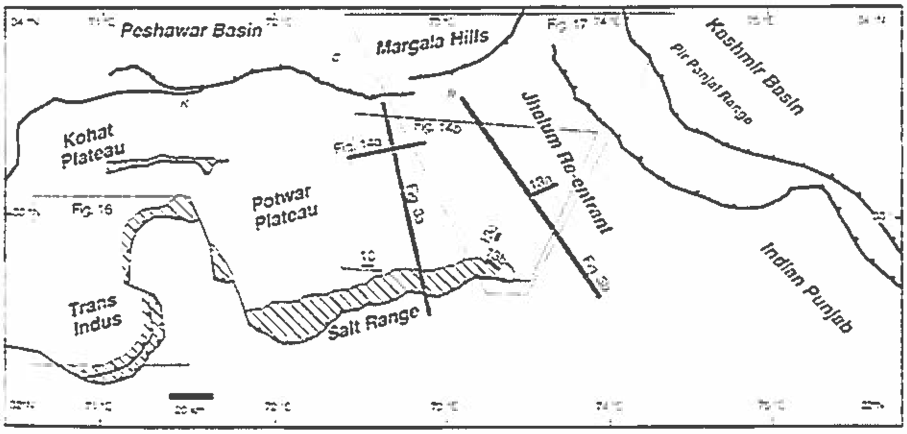


Figure 9.5. (a) Location map of magnetic sections in the Himalayan foreland. H, Haritalyangar; J, Jawalamukhi; K, Khetpurali. (b) Location map of magnetic sections in the northwestern Himalaya. Sections: BAL, Balli; BAR, Baramula; BHM, Bhimber; BHN, Bhaun; BKS, Basawa Kas; BNG, Banghala; BPC, Bain Pass composite; BND, Bhandara; BTR, Buttar; CHC, Chichali; C-G, Chinji–Gabir; DC, Dheri Choan; DG, Dag; DN, Dina; DKS, Dhok Saiyidan; FK, Faqir Killie; GC, Garhi Chandan; GPK, Ganda Paik; GRL, Garhi Landa; GRG, Gharangai; HIR, Hirpur; JAL, Jalalpur–Jamarghal; J-N, Jammu–Nagrota; KHR, Khaur; KK, Kotal Kund; KSD, Kas Dovac; KSG, Kas Guma; MHS, Mahesian; MWK, Mawa Kaneli; PBI, Pabbi hills; PEZ, Pezu; PIN, Pindori; PS, Pind Savikka; P-U, Parmandal; ROM, Romushi; RTD, Rata-Dadial; RTS, Rohtas; SHL, Shaliganga; SK, Sakrana; SLP, Sultan Pathan; SPT, Spalmai Tangi; SON, Soan; T/A, Tatrot–Andar; WLW, Walewal. Towns: C, Campbellpore; J, Jhelum; K, Kohat; R, Rawalpindi (c) Location map of figures in northwestern Himalaya.

not be syn-depositional. Several studies within the Siwaliks, however, have demonstrated that the magnetic signal revealed through thermal demagnetization usually was acquired at or very near the time of deposition (Tauxe and Badgley, 1988; Tauxe, Constable, and Stokking, 1990). Second, polarity zonations typically are based on calculations of virtual-geomagnetic-pole (VGP) positions or on declination data. Particularly for the earlier work, uncertainties in the polarity, such as that revealed by the confidence interval on the VGP latitudes, were not reported. Third, numerous magnetozones have been based on single sites, which increases the possibility that the calculated polarity does not indicate the polarity at the time of deposition. For example, incorrect sample orientations, mistaken compass readings, or lightning strikes all can result in miscalculations. Redundant sampling that duplicates these single-site reversals can serve to reinforce the polarity determination. Fourth, finite sample spacing, typically due to lack of exposure or inappropriate lithologies for sampling, dictates that the positions of reversal boundaries may not be well known or that polarity reversals may be missed (Johnson and McGee, 1983). Uncertainties in these

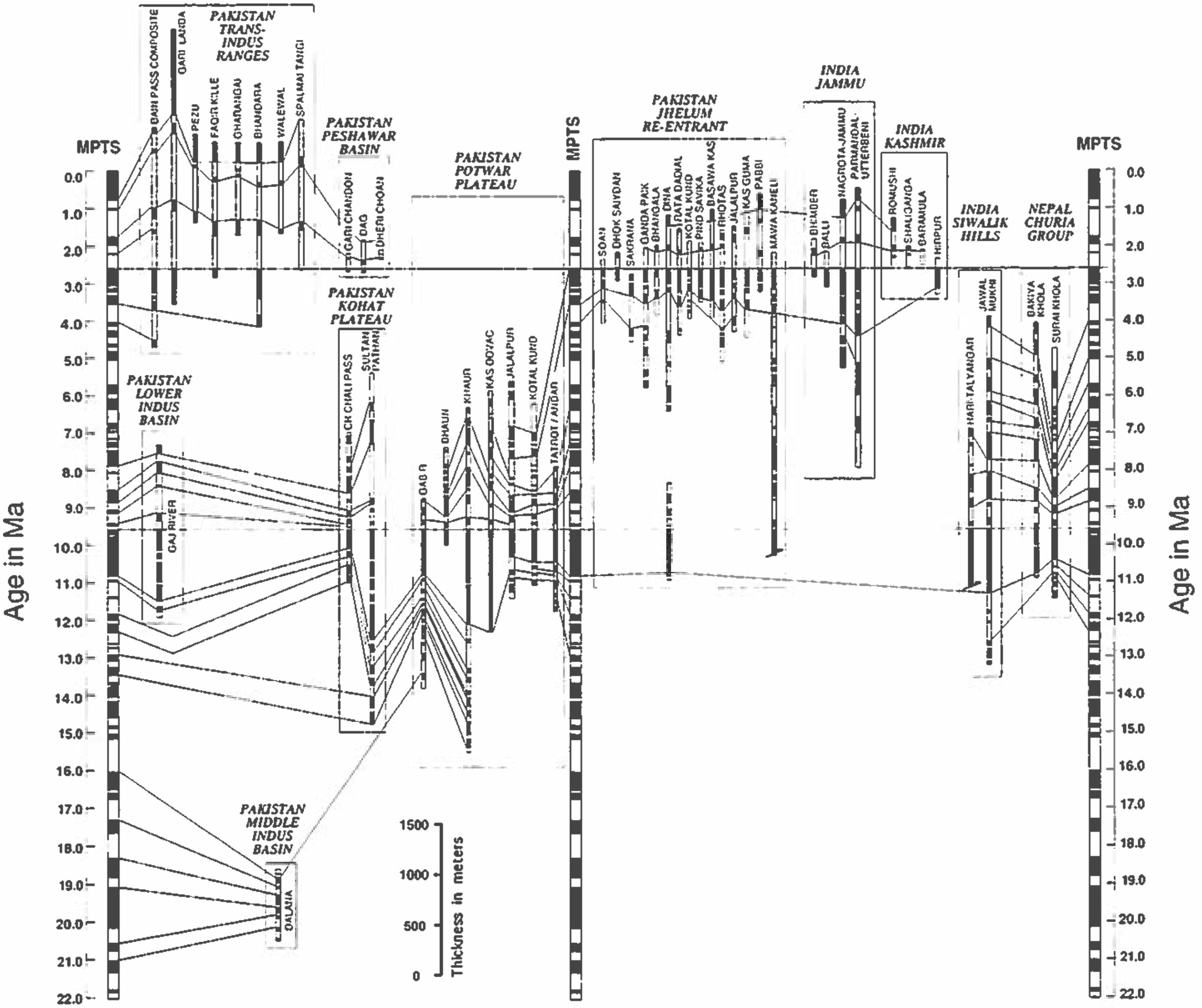


Figure 9.6. Correlation of local magnetic-polarity stratigraphies from the Indo-Gangetic foreland with the magnetic-polarity time scale of Cande and Kent (1992). Sources of magnetic data: Trans-Indus ranges and lower Indus basin, Khan et al. (1984, 1988); middle Indus basin, Friedman et al. (1992); Peshawar Basin, Burbank and Tahirkheli (1985); Kohat Plateau, unpublished data and Khan (1983); Potwar Plateau, unpublished data and Johnson et al. (1982b, 1985) and Raynolds (1980); Jhelum re-entrant, Raynolds (1980) and Raynolds and Johnson (1985); Jammu, Ranga Rao et al. (1988); Kashmir, Burbank and Johnson (1983); Siwalik hills, unpublished data and Johnson et al. (1983); Chura Group, Appel et al. (1991) and Harrison et al. (1993). In the context of the Cande–Kent time scale, especially because of the more complex reversal patterns between 6 Ma and 9 Ma, several of the published correlations have been reinterpreted, including Gaj River, Haritalyangar, and Bakiya Khola.

boundary positions propagate into any time or rate calculations. Fifth, preserved strata may contain an incomplete record of magnetic reversals, because of erosion or the presence of strata unsuitable for sampling. Sometimes reversals are discovered in sections that have no apparent counterpart in the magnetic-polarity time scale (MPTS), either because of the reasons listed earlier or because the MPTS is incomplete; for example, compare the magnetic time scales of Berggren et al. (1985) with those of Cande and Kent (1992). Sixth, field measurements of stratigraphic thicknesses introduce typical uncertainties of about 10%. Finally, correlations of local reversal sequences with the MPTS may be ambiguous (Talling and Burbank, 1993). Typically it is assumed that deposition is fairly constant, and in some cases in which there are radiometric controls high and low in the magnetic section (Johnson et al., 1982b) this can be demonstrated to be true (e.g., Kotal Kund, Figure 9.6). If sediment accumulation rates are assumed to vary widely and unpredictably, then almost any correlation with the MPTS is possible. Once a correlation has been made, however, many of the potential error sources described earlier are commonly ignored when precise ages derived from the MPTS are applied to stratigraphic analysis. It is important, therefore, whenever possible, to review the original magnetic data.

In the analysis presented here, we have re-correlated each section with the Cande and Kent (1992) magnetic time scale and revised the stratal ages accordingly. When compared with previously published magnetic time scales (Berggren et al., 1985; Harland et al., 1990), the Cande and Kent (1992) time scale commonly indicates a more complex reversal pattern, especially between about 6 Ma and 9 Ma. This suggests that some previously published correlations are likely to be incorrect. Whereas, in general, we have followed the correlations of the original authors (Figure 9.6), a few correlations have been significantly revised. Most notably, the top of the dated Haritalyangar section appears to be about 2 m.y. older than in the original correlation (Johnson et al., 1983). This revision is more nearly consistent with the sedimentation rates and lithofacies in the nearby longer and better-dated Jawalamukhi section (Meigs, Burbank, and Beck, 1995) and with the faunal data from Haritalyangar and Pakistan (Johnson et al., 1983; Barry et al., 1985). Similarly, we have re-correlated the Gaj River magnetostratigraphy (Khan et al., 1984) with the 11–7.5-Ma part of new magnetic time scale. This correlation yields steadier sediment accumulation rates and eliminates the anomalously old ages previously assigned to vertebrate fauna in the Manchar Formation (Khan et al., 1984).

Despite the potential problems described earlier, the presently available combination of faunal data with magnetic-polarity stratigraphies and radiometric dates on ashes permits classification of the northwestern Himalayan foreland as probably the best-dated terrestrial foreland in the world. There are, nonetheless, still significant gaps in both spatial and temporal coverage. Relatively few sections extend to times earlier than about 13 Ma, and only two, Gabir Kas (Johnson et al., 1985) and Zinda Pir dome in Pakistan (Friedman et al., 1992) extend to about 18 Ma. Thus, the early stages of Siwalik deposition are poorly documented. The spatial distribution of dated sections is also very irregular. For example, few sections spanning the Plio–Pleistocene interval are available from the Potwar Plateau of Pakistan, whereas within the adjacent Jhelum re-entrant, where there are numerous sections of this age, there are relatively few dated sections older than 5 Ma. In comparison with the published data from India and Nepal, where few volcanic ashes have been identified and where only a few magnetostratigraphic sections have been sampled, northern Pakistan has an abundance of well-dated sections. Recent and ongoing magnetostratigraphic studies between Jammu and Himachal Pradesh in India and within several areas of southern Nepal promise to broaden the available chronologic data base considerably over the next several years.

Sediment accumulation rates and subsidence histories

Dated sections provide a basis for calculating rates of sediment accumulation and basin subsidence. The slopes of simple plots of observed sediment thicknesses versus time represent compacted-sediment accumulation rates (Figure 9.7). Consequently, changes in slope can be interpreted as revealing variations in the rate. Whereas it would be preferable to calculate accumulation rates based on decompacted stratal thicknesses, there have been few detailed studies of compaction in terrestrial settings, and few generalized rules for compaction during pedogenesis, lithification, and burial of terrestrial sediments are available. Many Siwalik strata appear to have experienced early cementation and subsequently to have undergone relatively little compaction, but this condition is not ubiquitous. Given the coarse-grained nature of many sections and the absence of thick intervals of fine-grained strata, compacted-sediment accumulation rates are likely to underestimate true accumulation rates only moderately.

Under conditions of steady-state convergence, subsidence rates should steadily increase as a site is rafted closer to the thrust front. When an appropriate reference frame for defining the position of the depositional surface can be defined, the dated record of sediment accumulation can be used to reconstruct a subsidence history (Burbank and Beck, 1991b). Although sea level is commonly used to provide such a reference frame (e.g., Angevine, Heller, and Paola, 1990), this is impractical for the northwestern Himalayan foreland, because this was a terrestrial basin, where deposition occurred more than 1000 km from the sea during Neogene times, such that any sea-level effects would be strongly damped. A different strategy is needed for this region. We use two approaches to create a reference frame. First, in sections oriented transverse to the mountains, we argue that the depositional surface of the preserved foreland was essentially horizontal (Burbank and Beck, 1991b), because regional stratigraphic analyses, paleo-currents (Burbank and Beck, 1991b), and detrital-mineral studies (Cerveny et al., 1988) indicate that throughout the Miocene, a major axial river system switched freely across the medial-to-distal foreland. Second, by analogy with the modern foreland, topographic gradients along the depositional surface of the undeformed foreland, particularly parallel to the axial rivers, were very low (<0.001). Such surface slopes can be ignored or considered horizontal if the stratal thickness of the interval of interest is large and the distance between sections being compared is small compared with the depositional gradient (~100 m per 500 km). The uncertainties introduced by such assumptions typically are small compared with the uncertainties inherent in the dating and measuring of the sections. If both the gradients of axial rivers in the foreland and their altitudes as they crossed downstream bedrock thresholds that may have regulated the base level (e.g., the Monghyr–Saharsa Ridge for the modern Ganges River, Figure 9.2) are assumed to have remained approximately constant through time, then basin subsidence can be reconstructed from the inverted sediment-accumulation curves.

Neogene deposition prior to 11.5 Ma

The detailed character of early Neogene deposition is not well known. In the Indus foreland basin there are two sections with

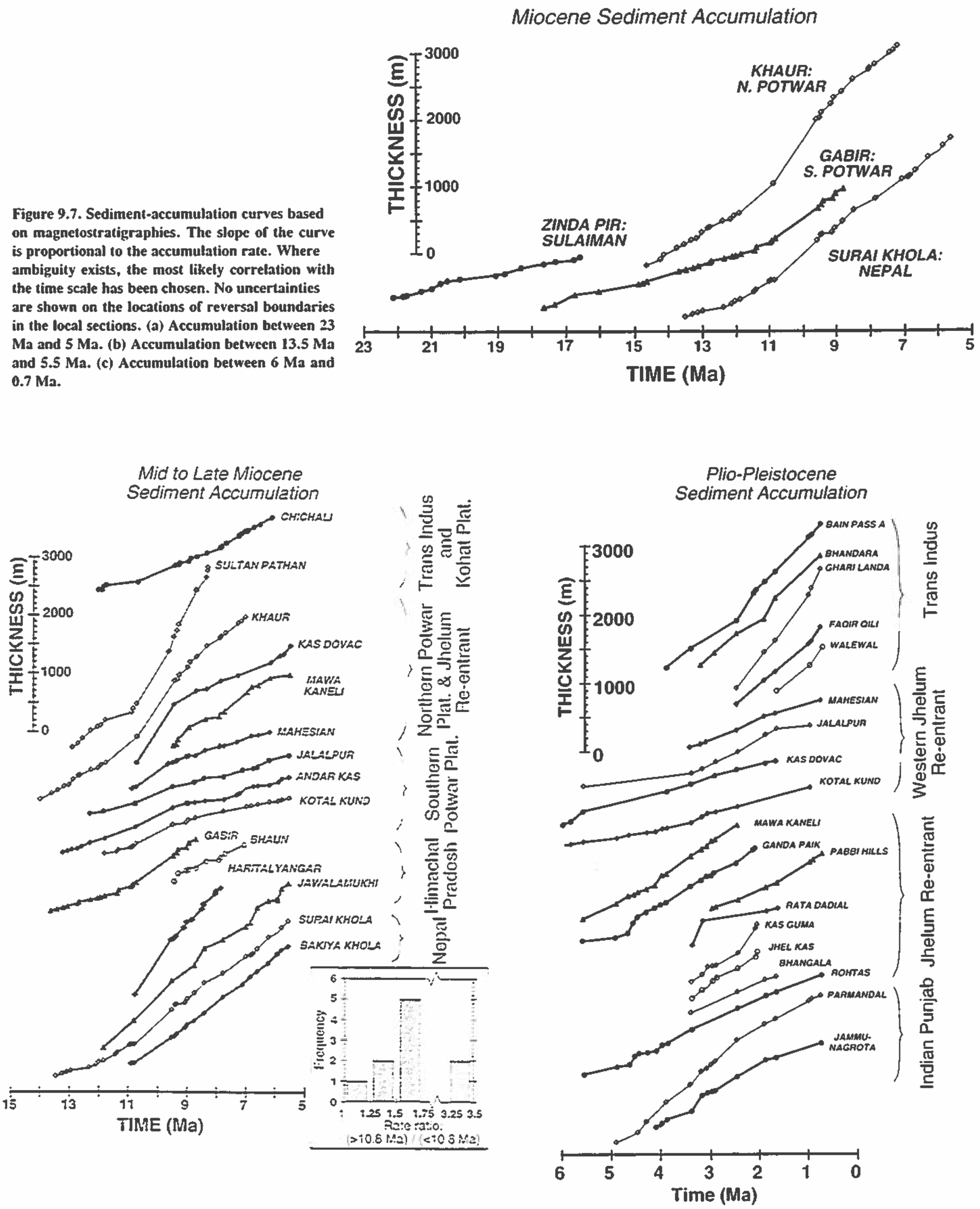


Figure 9.7. Sediment-accumulation curves based on magnetostratigraphies. The slope of the curve is proportional to the accumulation rate. Where ambiguity exists, the most likely correlation with the time scale has been chosen. No uncertainties are shown on the locations of reversal boundaries in the local sections. (a) Accumulation between 23 Ma and 5 Ma. (b) Accumulation between 13.5 Ma and 5.5 Ma. (c) Accumulation between 6 Ma and 0.7 Ma.

magnetically dated strata older than about 14 Ma: Chinji–Gabir (Johnson et al., 1985) and Zinda Pir dome (Friedman et al., 1992). No well-dated, detailed sections of this age have been published for India or Nepal, although there are several localities, such as the Himachal Pradesh re-entrant, where rocks of this age are exposed in continuous sections. At Chinji village in the southern Potwar Plateau (Figure 9.5) the basal Kamlial strata are dated at about 18 Ma (Figure 9.6) and directly onlap Eocene carbonates. These Eocene limestones are exposed along the northern edge of the Salt range, where the Potwar allochthon ramps southward (Lubetkin and Clark, 1988; Willis, 1989). Although this 18-Ma date has been cited as evidence for the beginning of significant foreland-basin deposition (e.g., Harrison et al., 1992), it should be understood simply as the distal pinch-out of the foreland at that time and place (Figure 9.4). Older foreland strata are clearly preserved to the north (Raiverman et al., 1983; Lillie et al., 1987). Along the distal (southern) margin of the early Miocene foreland, near Chinji village, deposition in the Kamlial Formation appears to have been dominated by rivers flowing to the southeast, parallel to the axis of the foreland basin. Channel sandstones, displaying average story thicknesses of about 10 m, constitute about 65% of the Kamlial succession (Willis, 1993a). Average sediment accumulation rates were about 0.11 km/m.y. (Figures 9.7a and 9.8). It is interesting to note that even in this very distal position, major contributions from cratonal rivers are not evident.

Southwest of the Salt range and adjacent to the transpressional Sulaiman ranges, foreland strata on Zinda Pir dome, recently dated as spanning about 22–16 Ma (Friedman et al., 1992), are preserved within this north–south-trending segment of the Indus foreland (Figures 9.5 and 9.6). The local sedimentology indicates that estuarine deposition (Chitarwata Formation) commenced at about 22 Ma (Downing et al., 1993). Subsequent tidal-channel and delta-plain deposition in the upper Chitarwata Formation yielded to southward-prograding (Waheed and Wells, 1990) fluvial deposition (Vihowa Formation) at about 18.6 Ma (Downing et al., 1993). Mean sediment accumulation rates during Chitarwata deposition were low: about 0.9–0.12 km/m.y. (Figure 9.7a, Table 9.1). The estuarine and tidal deposits represent the final marine incursion in this part of the foreland (Figure 9.9). The transition to fluvial deposition here coincided in age with the basal Kamlial strata in the southern Potwar Plateau (Johnson et al., 1985). This suggests that the geographic shape of the early Miocene foreland was similar to that of today, inasmuch as the distal edge of the foreland wrapped around the northwestern corner of exposed cratonal rocks. Further, this geometry suggests that, similar to the present, loading by both transpressional ranges and the Himalaya created a two-part flexural moat that had nearly orthogonal trends. Unlike today in the Indus foreland, where the rivers of the Punjab are tributary to the Indus River, paleo-current indicators (Waheed and Wells, 1990) suggest that during the early and middle Miocene the major fluvial systems flowed orthogonal to each other and parallel to the mountain front, such that the Potwar system flowed to the southeast (Willis, 1993a), where presumably it joined the Ganges system, and the Sulaiman system flowed to the south-southwest. It thus appears that the drainage divide separating the rivers draining to the Bay of Bengal from those draining to the Arabian Sea was considerably farther west of its present position and may have been situated west of the present-day Peshawar Basin.

Although the number of dated early Miocene sections is small, there are several foreland sections that include middle Miocene strata dated between 14 Ma and 11.5 Ma (Figure 9.6). In the Potwar Plateau, the Chinji Formation (Figure 9.3d) encompasses strata of approximately this age (Tauxe and Opdyke, 1982; Johnson et al., 1982b, 1985). Based on the ages of the dated sections, it is possible to discern differential subsidence along north–south and east–west sections across the Potwar Plateau. The subsidence has been greatest in the northern plateau and tends to be greater in the west than in the east (Figures 9.7b and 9.8, Table 9.1). In fact, regional subsidence patterns through time (Figure 9.8) suggest that the geographic boundary between the southeastern Potwar Plateau and the Jhelum re-entrant approximately coincides with a transition zone of slower subsidence. That reduced subsidence may have resulted from the presence of a basement promontory that generally delineated the eastern margin of the evaporite basin in which the Salt Range Formation accumulated (Leathers, 1987). In the southern Potwar, sediment-accumulation histories reveal that during the Kamlial and Chinji depositions, there was no significant difference in rates over the interval. Overall, the Chinji strata are characterized by an alternation of channel sandstone and overbank mudstone and sandstone. Channel sandstones averaging about 10 m in thickness constitute about 30% of the Chinji strata (Willis, 1993b). Storeys typically are separated by about 25 m of intervening overbank deposits, typically displaying extensive pedogenic alteration that has led to prominent color and textural zonations. Soil bands and intervening sand bodies 5–15 m thick can commonly be traced as far as 10 km across the middle Miocene floodplain (Johnson et al., 1988).

Detailed sedimentologic reconstructions of the Chinji Formation in the vicinity of Chinji village on the southern Potwar Plateau (Willis, 1993b) delineate the character of the middle Miocene fluvial system. Multistoried sandstones represent braided rivers with typical maximum depths of 4–13 m (Figure 9.10), channel widths of 80–200 m, and bank-full discharges across the entire channel belt of about 1,500–2,000 m^3/s (Willis, 1993a). Mean flow directions were toward the southeast (Figure 9.8). This orientation is nearly orthogonal to the modern Indus River and suggests that during the middle Miocene the major river in this region was flowing eastward toward the Gangetic foreland and the Bay of Bengal (Burbank and Beck, 1991b).

Foreland strata of similar ages (14–11.5 Ma) have been described in western Nepal (Figures 9.5–9.7) at Surai Khola (Appel et al., 1991), where they accumulated at mean rates of about 0.17 km/m.y. (Table 9.1, Figure 9.7), at Arung Khola in west-central Nepal (Tokuoka et al., 1986), and in the Himachal Pradesh re-entrant (Raiverman et al., 1983). Although the lower Siwalik strata in Himachal Pradesh quite closely resemble the Chinji lithofacies in Pakistan, coeval strata in Nepal are strikingly

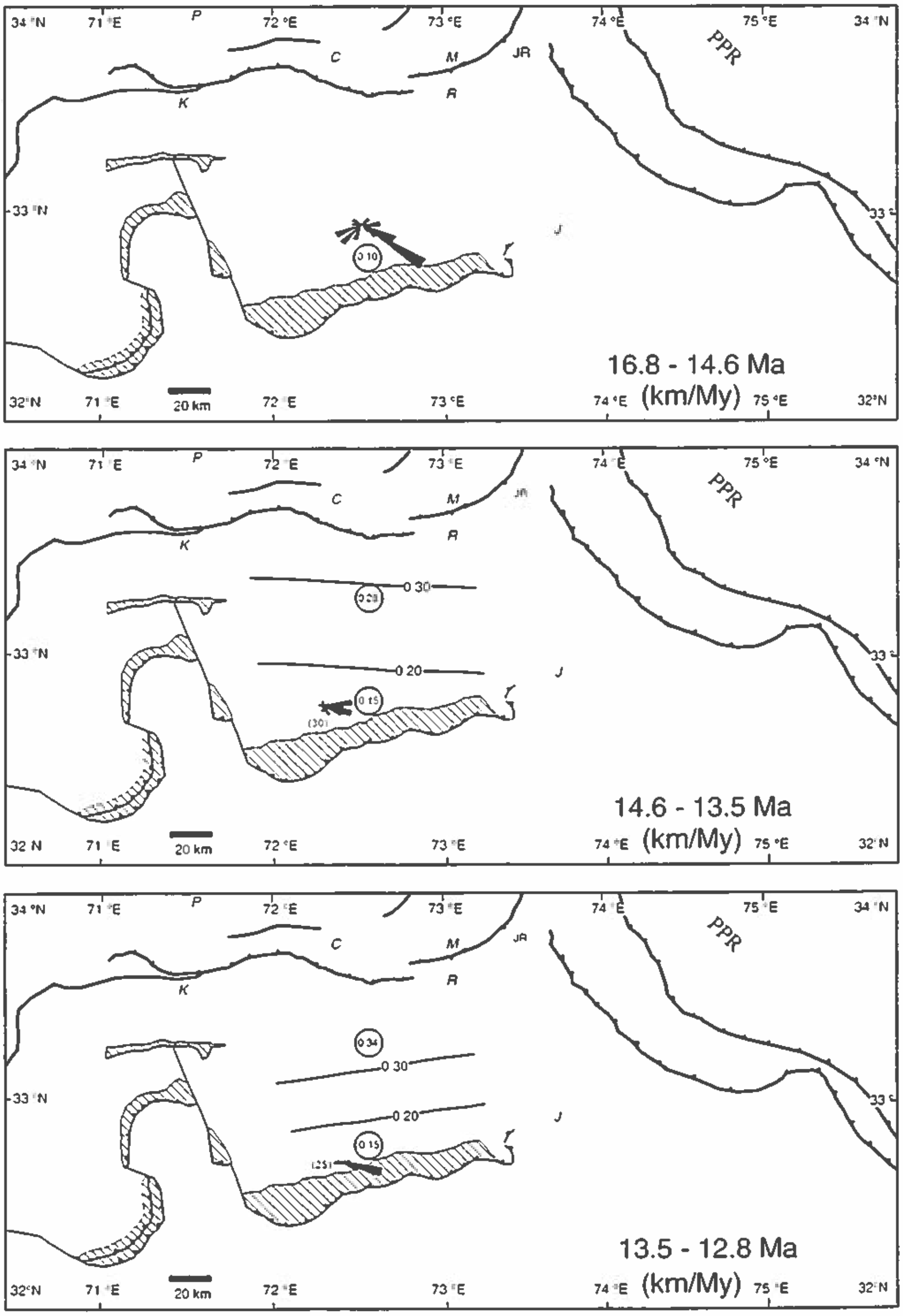
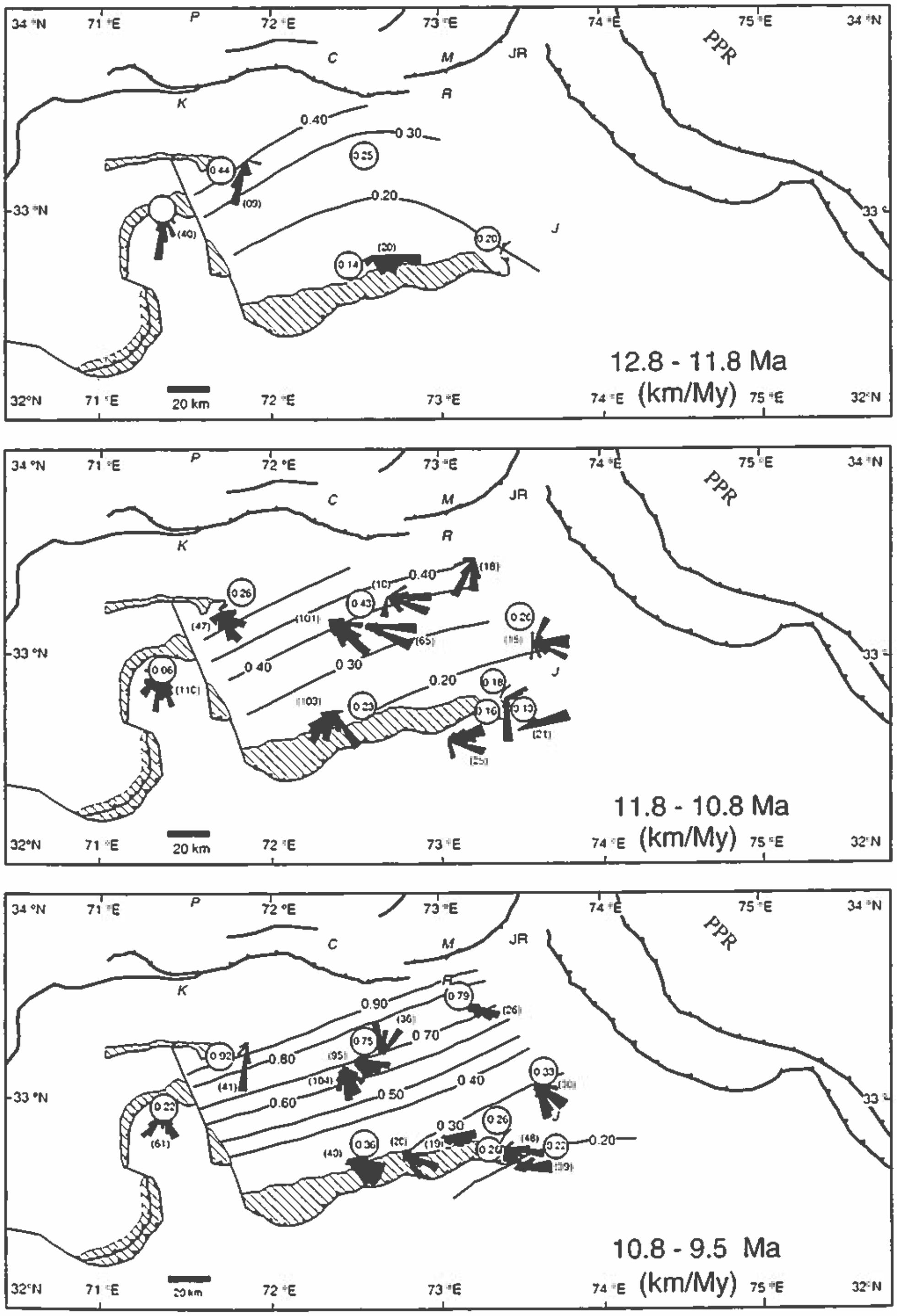


Figure 9.8 Time-slice maps of Miocene-to-Pleistocene subsidence and paleocurrents across the northwestern Himalayan foreland. Rates (in km/m.y.) are circled at each site and represent compacted-sediment accumulation. The numbers of measurements are shown in parentheses adjacent to the paleocurrent roses. Arrows in the 2.5–1.7-Ma reconstruction (Fig. 9.8) represent "brown sandstones" from Raynolds (1980). ACR, Attock–Cherat range; C, Campbellpore; J, Jhelum; JR, Jhelum re-entrant; K, Kohat; M, Murree; P, Peshawar; PPR, Pir Panjal range.



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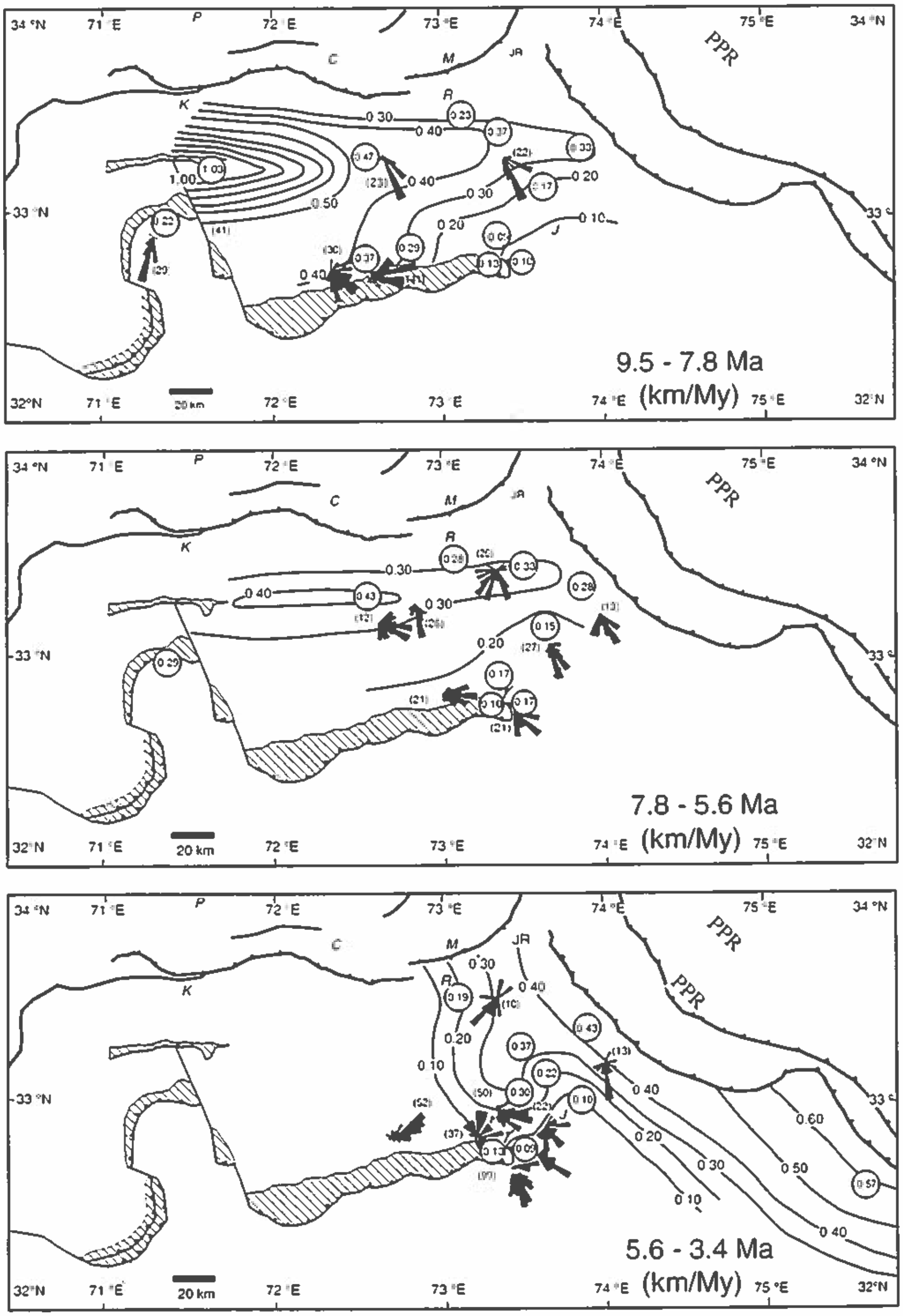

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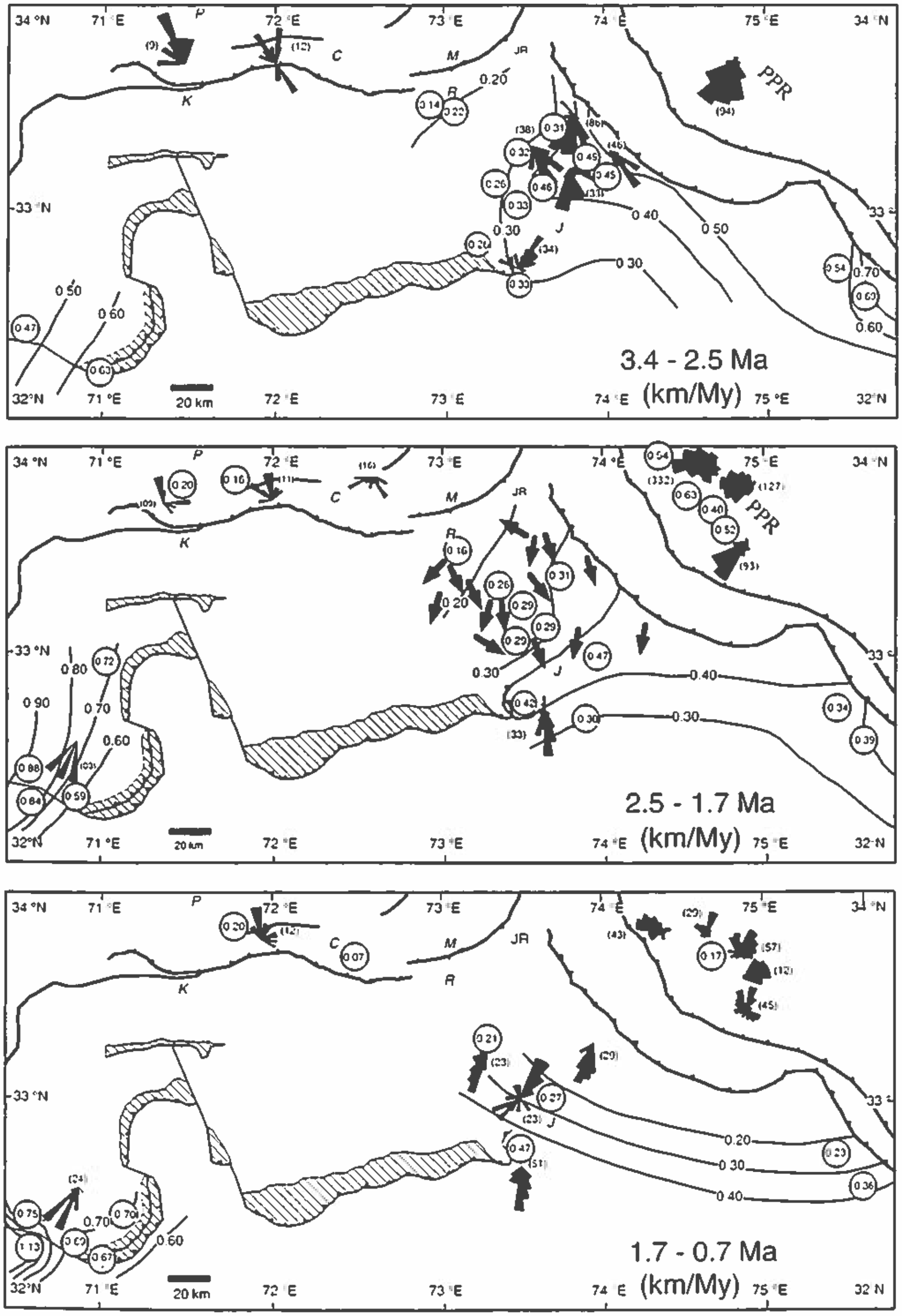

Figure 9.8 (*cont.*)

Table 9.1. *Mean sediment accumulation rates*

Section	Author	Area	Sediment accumulation rate (m/m.y.)											
			0.7–1.7	1.7–2.5	2.5–3.4	3.4–5.6	5.6–7.8	7.8–9.5	9.5–10.8	10.8–11.8	11.8–12.8	12.8–13.5	13.5–14.6	14.6–16.8
Andar Kas	Johnson et al. (1982b)	Potwar Plateau					168	89	257	183	199	140		
Bain Pass A	Khan (1983)	Trans-Indus	748	885										
Bain Pass B	Khan (1983)	Trans-Indus			470									
Bakiya Khola	Harrison et al. (1992)	Nepal					394	402	347					
Baramula	Burbank (1982)	Kashmir		536										
Bhandara	Khan (1983)	Trans-Indus	672		632									
Bhangala	Raynolds (1980)	Jhelum re-entrant		264	326									
Bhaun	Opdyke et al. (1979)	Potwar Plateau						293						
Buttar	Small (1980)	Jhelum re-entrant					327	372						
Chichali	Khan (1983)	Trans-Indus					293	218	221	58				
Dag	Burbank (1982)	Peshawar Basin	195	159										
Dheri Choan	Burbank (1982)	Campbellpore basin	72											
Faqir Qili	Khan (1983)	Trans-Indus	694	590										
Gabir	Johnson et al. (1985)	Potwar Plateau						374	363	231	140	148	146	99
Gaj Kas	Khan et al. (1984)	Kirthar						472	679					
Ganda Paik	Raynolds (1980)	Jhelum re-entrant				368								
Ghari Chandon	Burbank (1982)	Peshawar Basin		201										
Ghari Landa	Khan (1983)	Trans-Indus	1128	848										
Haritalyangar	Johnson et al. (1983)	Himachal Pradesh						528	≥497					
Hirpur	Burbank (1982)	Kashmir		516										
Jalalpur	Johnson et al. (1982b)	Potwar Plateau		418	331	88	161	103	221	135				
Jammu–Nagrota	Ranga Rao et al. (1988)	Jhelum re-entrant	228	344	544									

Jawalamukhi	Meigs et al. (1995)	Himachal Pradesh					417	439	474	529			
Jhel Kas	Visser and Johnson (1978)	Jhelum re-entrant		470	464								
Kas Dovac	Raynolds and Johnson (1985)	Potwar Plateau		160	224	185	284	229	785				
Kas Guma	Raynolds and Johnson (1985)	Jhelum re-entrant			448								
Khaur	Tauxe and Opdyke (1982)	Potwar Plateau					430	471	750	433	247	338	279
Kotal Kund	Johnson et al. (1982b)	Potwar Plateau			256	132	99	135	264	163			
Mahesian-Dina	Mulder (1991)	Jhelum re-entrant	206	295	267	225	148	166	331				
Mawa Kaneli	Mulder (1991)	Jhelum re-entrant			491	426	282	333					
Pabbi hills	Keller et al. (1979)	Jhelum re-entrant	466	301									
Parmandal	Ranga Rao et al. (1988)	Jhelum re-entrant	363	387	694	569							
Pezu	Khan (1983)	Trand-Indus											
Pindoori	Raynolds and Johnson (1985)	Jhelum re-entrant				102							
Rata Dadial	Raynolds and Johnson (1985)	Jhelum re-entrant		308	308								
Rohtas	Raynolds and Johnson (1985)	Jhelum re-entrant	271	295	331	306							
Romushi	Burbank (1982)	Kashmir	162	402									
Sakrana	Raynolds and Johnson (1985)	Jhelum re-entrant											
Shaliganga	Burbank (1982)	Kashmir		603									
Soan	Moragne (1979)	Potwar Plateau			135								
Spalmi Tangi	Khan (1983)	Trans-Indus		717									
Sultan Pathan	Beck et al (1994)	Kohat Plateau						1030	916	264	435		
Surai Khola	Appel et al. (1991)	Nepal					401	365	470	284	532	189	204
Walewal	Khan (1983)	Trans-Indus	705										
Zinda Pir	Friedman et al. (1992)	Sulaiman											

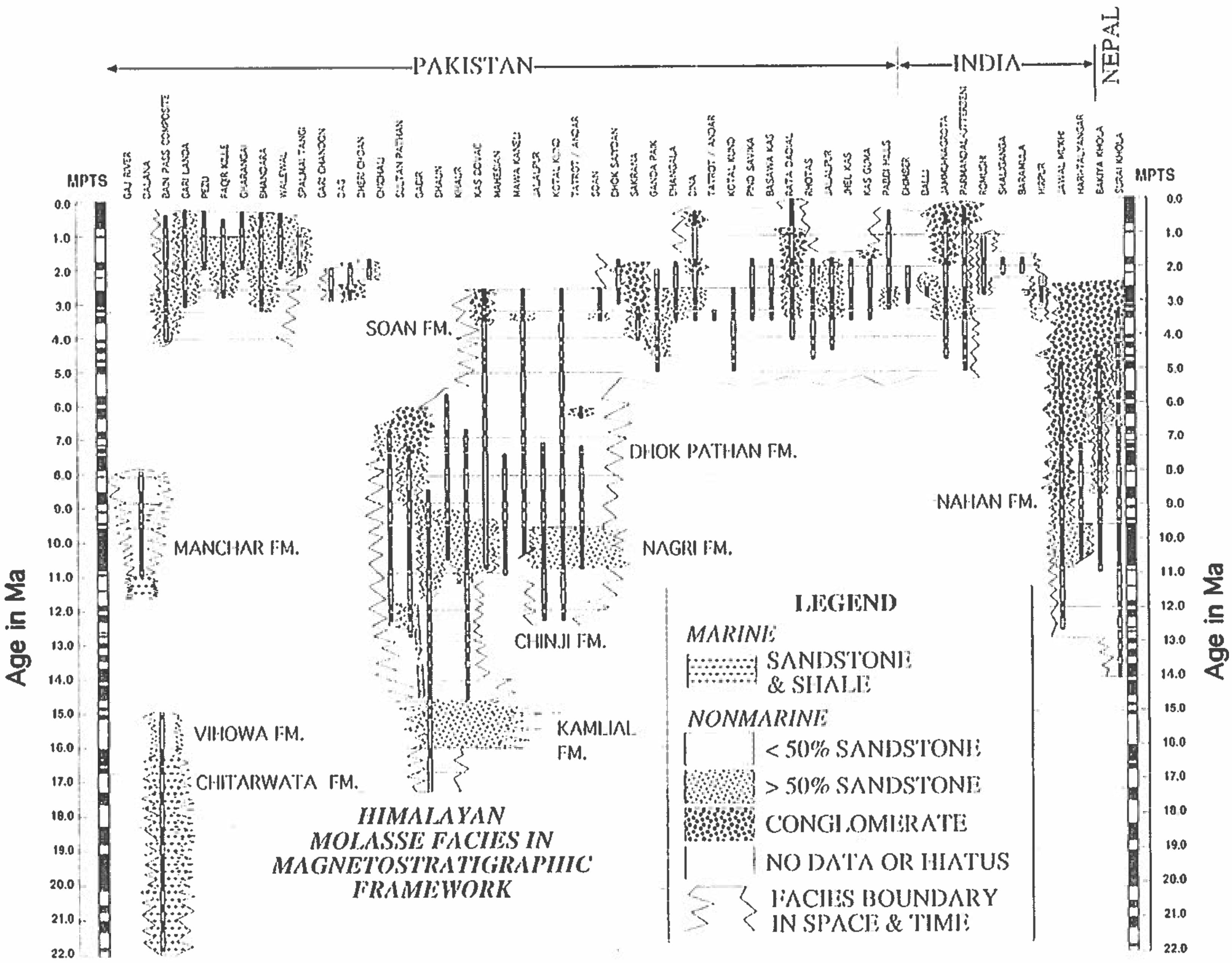

Figure 9.9 Lithofacies versus time for the Himalayan foreland. Lithofacies boundaries are plotted versus magnetically defined ages. Note the time equivalence of Nagri and Nahan sandstones and different ages of conglomerates within sections.

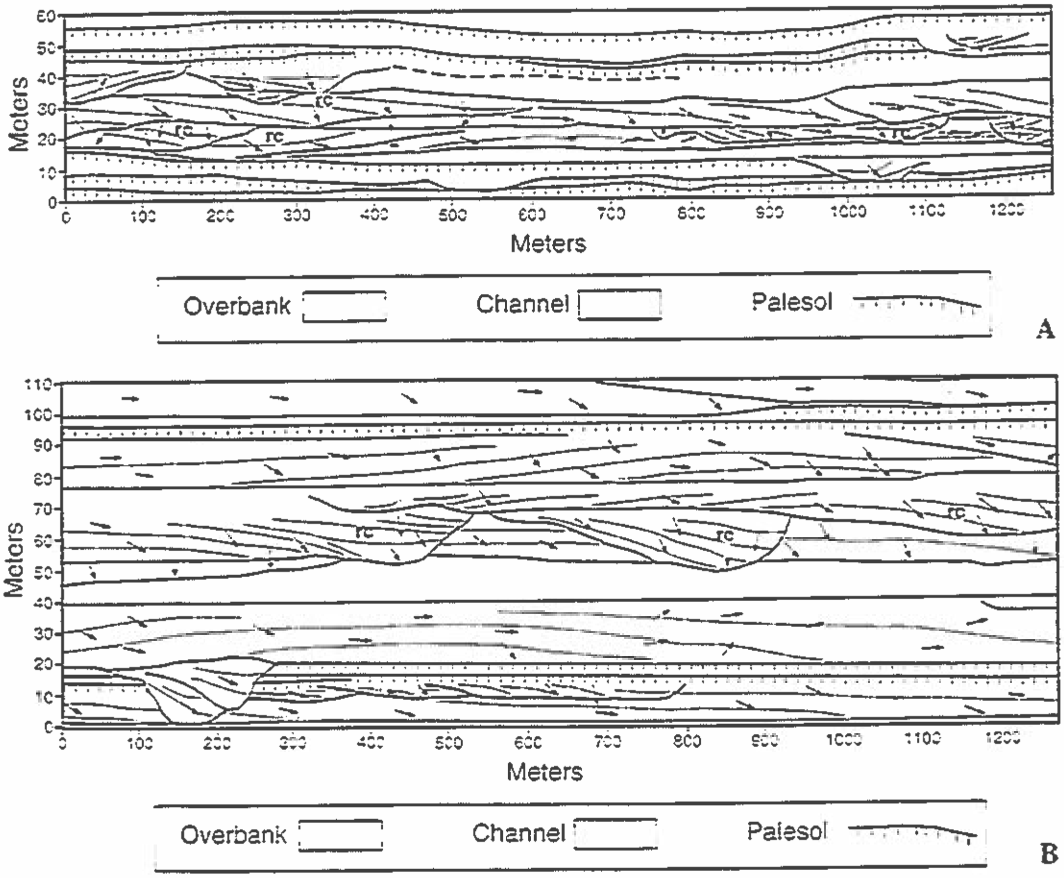

Figure 9.10. Sedimentology of Chinji and Nagri facies in Pakistan (Willis, 1993b). (a) Chinji strata at Gabir Kas, showing typical channels 4–10 m deep and 100–200 m wide. Small arrows show paleocurrents with respect to the orientation of the panel. Thin lines within individual channels represent major bedding surfaces. The least ambiguous reconstructions of channel cross sections can be made where flow is perpendicular to the panel, such as at sites labeled "rc." (b) Channels in Nagri sandstones in Gabir Kas. When compared with the Chinji strata, note wider (>200m) and deeper (>15 m) channels and higher overall abundance of sandstones.

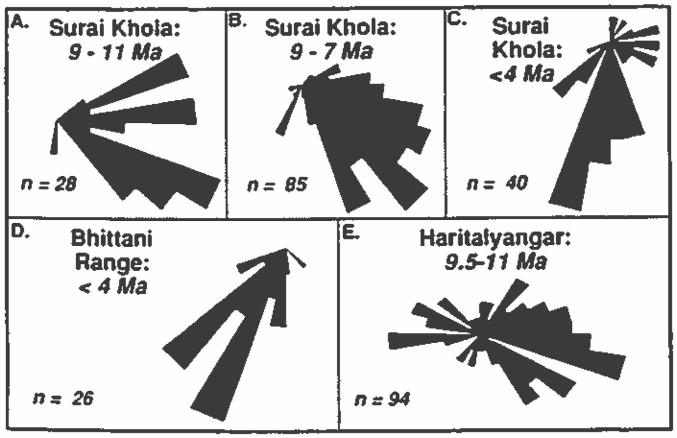

Figure 9.11. Paleo-current measurements from (A) 13–11 Ma at Surai Khola, (B) 11–7 Ma at Surai Khola, (C) <4 Ma at Surai Khola, (D) <4 Ma in the Bhittani range, and (E) ~10 Ma at Haritalyangar.

different (Figure 9.9). At Surai Khola, for example, 14–11.5-Ma strata are dominated by mudstone, with common calcareous paleosols and organic-rich horizons. Sandstone beds usually are less than 1 m thick, typically coarsen and thicken upward, and rarely attain thicknesses of 10–15 m. Abundant plant remains and aquatic vertebrate fossils (West, 1984) suggest depositional environments that were typified by swampy, poorly drained floodplains and shallow streams in a warm, wet climate. Although these fluvial strata have been interpreted as part of a distal alluvial fan that was prograding southward from the Himalayan front, our paleo-current measurements indicate primary flow toward the southeast (Figure 9.11). This suggests that these rivers were part of a tributary system oriented obliquely to the mountain front, as are many of the present tributaries to the modern Ganges (Figure 9.1). (Note that wherever sufficient magnetic data are available, paleo-currents have been corrected for post-depositional vertical-axis rotations.)

In comparison with coeval strata in Pakistan, the overbank lower Siwalik Nepali strata are considerably less oxidized and less pedogenically altered. Extensive mottling suggests hydromorphic soils indicative of episodically saturated environments. Widespread preservation of organic debris suggests a more moist, less well drained depositional setting, in comparison with Pakistan. Present-day climatic gradients would be expected to produce similar contrasts between the arid regions of the northern Pakistani foreland and the monsoon-dominated foreland of Nepal.

Neogene deposition between 11.5 Ma and 8.5 Ma

Beginning at about 11 Ma, throughout the Himalayan foreland, all of the dated sections of appropriate ages and sufficiently long records show accelerations in accumulation and subsidence (Table 9.1, Figures 9.7b and 9.8). Most commonly, rates increased by 35–75%. Some of the highest accumulation rates (~ 1 km/m.y.) documented within the Himalayan foreland occurred during that time interval, particularly at sections in the northern foreland. Where data are available from north–south transects, such as between Khaur and Gabir in the Potwar Plateau or at Sultan Pathan and Chichali in the Kohat Plateau, comparable relative rate changes are observed in the more rapidly subsiding northern sections and in the more slowly subsiding southern sections (Figure 9.8). No dated sections in the foreland display significant decreases in subsidence rates for that time, although one (Jawalamukhi) shows only a small (~ 10%) increase.

Increased accumulation rates coincide with several key changes in foreland deposition. In the lower Indus basin (Figure 9.5), a transition from marine to fluvial deposition (Khan et al., 1984) occurred at about 11 Ma (Figure 9.9). Across the Potwar Plateau, the Jhelum re-entrant, Himachal Pradesh, and western Nepal (Johnson et al., 1982b, 1983, 1985; Appel et al., 1991) there was an important and abrupt upward coarsening, such that channel sandstones generally compose about 70% or more of the stratigraphic succession. This change defines the base of the Nagri Formation and the Nahan sandstones (Johnson and Vondra, 1972) in the Potwar region and Himachal Pradesh, respectively. The base of the Nagri is slightly time-transgressive between sections (Johnson et al., 1982b) and ranges from about 11.4 Ma to about 10.8 Ma (Figure 9.9). On the basis of sedimentologic analysis of outstanding exposures in the southern Potwar Plateau (Willis, 1993b), the Nagri depositional system comprises multistoried sandstones in which individual storys average about 18 m in thickness (Willis, 1993b). Mean flow was to the southeast (Figure 9.8). Cross sections of channels orthogonal to the flow display typical channel widths of 200–400 m and maximum channel depths of 15–30 m (Figure 9.10). Reconstructed bank-full discharges across the entire channel belt are estimated to have been 10,000 m^3/s or more (Willis, 1993b). Thus, in comparison with the sandstone beds of the underlying Chinji lithofacies, Nagri sandstones are both strikingly thicker and more common, and paleo-discharges probably were 4 times greater, or more, during Nagri deposition. Nagri discharges were similar to those of the major transverse rivers flowing across the foreland today, but were considerably less than the discharges of major axial rivers, such as the Indus or Ganges, in their lower valleys (Willis, 1993b).

In India, the Nahan sandstone in the Himachal Pradesh re-entrant (Johnson et al., 1983) is of an age (~11.5–9.8 Ma) and sedimentologic character similar to those of the Nagri lithofacies in Pakistan (Figure 9.9). Paleo-flow directions, at least in the eastern part of the re-entrant, were toward the southeast (Figure 9.11e). Cycles of amalgamated sandstone attain thicknesses of 50–150 m (Johnson and Vondra, 1972) and are separated by thin overbank mudstone successions. These fluvial strata have been interpreted as the products of "loosely sinuous" rivers (Johnson and Vondra, 1972). Lithologically similar, although undated, sandstone-dominated strata occur below "Middle Siwalik" strata (Kumar, 1989; Kumar et al., 1991) in the Dehra Dun sub-basin (Figure 9.2). The upper Nahan Formation at the Nahan-type locality near Dehra Dun comprises coarse- to medium-grained, poorly sorted, multistoried sandstones with subordinate pedogenically reddened siltstone (Sen, 1981). These strata were deposited by southeastwardly flowing rivers (Sen, 1981) and are assumed to be time-correlative with Nahan strata in Himachal Pradesh. In western Nepal, at Surai Khola, clear upward coarsening and thickening began at about 11.5 Ma (Appel et al., 1991). Sandstone beds, typically 5–15 m thick, constitute more than 50% of the section. Paleo-current directions remained toward the southeast during that interval (Figure 9.11a,b). At Arung Khola in west-central Nepal, a similar upward coarsening began at about 10.5 Ma (Tokuoka et al., 1986). The sandstone-dominated succession of the upper Arung Khola Formation comprises 10–15 m thick, upward-fining cycles deposited by meandering river systems flowing toward the southeast (Tokuoka et al., 1986).

Most of the available data from Pakistan to Nepal indicate paleo-flow parallel to the axis of the foreland and to the southeast during the late middle Miocene (Figures 9.8 and 9.11). The comparable scales of sand bodies in the Potwar, Jhelum, and Himachal Pradesh regions suggest that these could be parts of a single large, paleo–Indus River system that was flowing southeastward into the Ganges drainage at that time. Although showing similar flow directions, the much smaller sandstone thicknesses in Nepal cannot be logical downstream continuations of the main Indo-Gangetic channel system, but must belong to a subsidiary Himalayan tributary system trending sub-parallel to the axial drainage.

In northern Pakistan, the dense-mineral assemblage changed abruptly with the beginning of Nagri deposition. The abundance of blue-green hornblende increased about threefold (Figure 9.12c) to more than 20% at that time in the southern and western Potwar Plateau (Cerveny et al., 1989; Mulder, 1991). Those changes can be interpreted in the context of the dense-mineral content of the modern rivers of the Punjab (Figure 9.12a). Whereas the modern Indus River averages about 33% blue-green hornblende in its dense-mineral assemblage, the Jhelum, Ravi, Chenab, and Sutlej all contain less that 15%, and more typically less than 10%, blue-green hornblende (Cerveny et al, 1989). It appears that the blue-green hornblende is derived from the Kohistan terrane in northern Pakistan (Tahirkheli, 1979; Cerveny et al., 1989). Whereas Kohistan is within the catchment of the Indus, it is not part of the catchments of the other major Punjabi rivers.

At least three hypotheses can be invoked to explain the dense-mineral changes at about 11 Ma: (1) Although the paleo–Indus River had previously transported this assemblage, it was suddenly diverted into the Potwar region of the foreland at that time because of either stream capture or tectonic diversion. (2) Bedrock uplift and associated erosion or tectonic denudation in Kohistan caused the blue-green hornblende source rocks to be extensively exposed for the first time at about 11 Ma. (3)

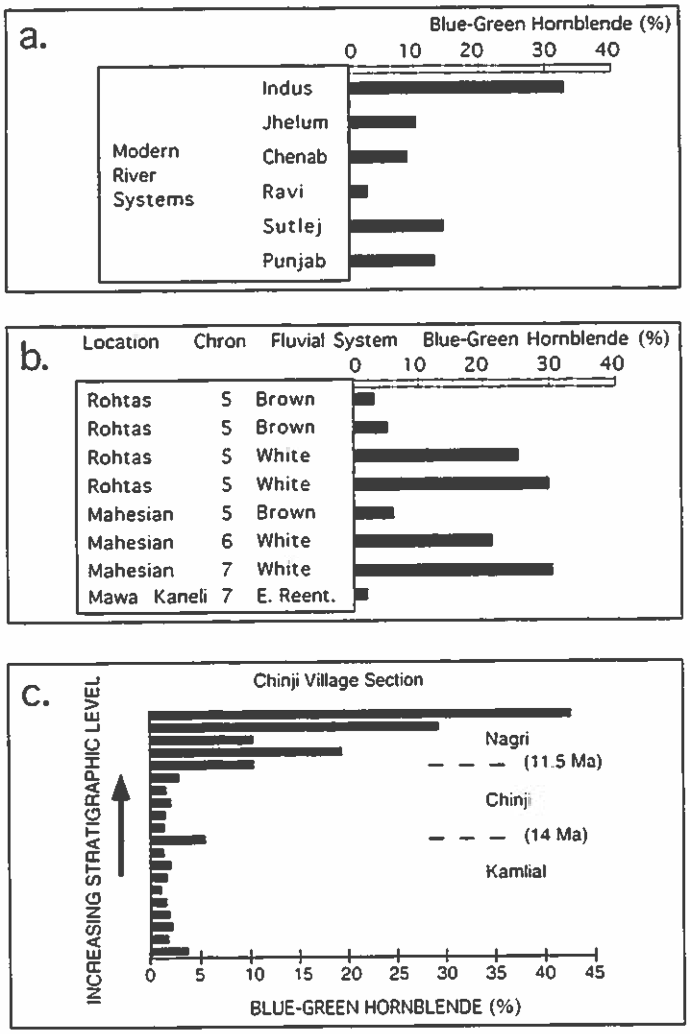

Figure 9.12. Abundances of blue-green hornblende in (a) modern Punjabi rivers (Cerveny et al., 1989), (b) sandstones of the Jhelum re-entrant (Mulder, 1991), and (c) sandstones in the southern Potwar Plateau at Gabir Kas (Cerveny et al., 1989).

Gradual erosion through the carapace of Kohistan rocks happened to expose the source rocks at about that time. Some combination of these hypotheses is also possible. For example, rapid uplift in Kohistan could have exposed the source rocks at the same time that related deformation could have diverted the paleo–Indus River.

Detailed cooling histories from Kohistan that could be used to evaluate hypothesis 2 are not abundant. The available data do not indicate rapid mid-Miocene cooling at a regional scale (Zeitler, 1985), but some areas have apatite fission-track ages (annealing temperatures of 100°C) as young as 6 Ma (Zeitler, 1985). We have no means of evaluating hypothesis 3. The large increase in paleo-discharge and the major changes in sandstone abundances that coincided with the initiation of Nagri deposition and with the influx of blue-green hornblende support the concept of diversion or capture of the Indus into the Potwar region at that time. The concurrent accelerations in accumulation and subsidence at about 11 Ma indicate that active tectonic deformation was also affecting the foreland.

We interpret that acceleration to have resulted from development of significant new thrust loads within the Himalaya. Flexural models (Flemings and Jordan, 1989; Sinclair et al., 1991) suggest that if an increased load were applied in the same location as the previous load, the new load would have to be 2–4 times greater than the previous one in order to generate the observed accelerated subsidence. It seems more likely that the locus of loading was shifted farther toward the foreland, rather than there having been a great increase in the size of the load. We suggest that this subsidence event represented the initial development of a new system of major, large-displacement thrusts across much of the Himalaya. The most likely candidate for this thrust system is the MBT system. Thus, we propose that the MBT was initiated during the middle Miocene at about 11 Ma. Interestingly, a 500-m-thick succession of middle Siwalik conglomerates at Jawalamukhi in the Himachal Pradesh re-entrant has recently been dated as spanning about 9–7 Ma (Meigs et al., 1995). These conglomerates are located more than 60 km from the MBT thrust (and thus are in the medial part of the basin) and comprise igneous and meta-sedimentary rocks, with likely source areas in the Lesser Himalaya. They appear to represent a coarse facies that was focused down the axis of the re-entrant in response to thrust loading along the flanks of the re-entrant. These are the oldest well-dated, extensive conglomerates in the Miocene record of the foreland and are particularly noteworthy on this account (Figure 9.9). Outside of the axis of the Himachal Pradesh re-entrant, however, nearly continuous conglomerate deposition did not begin until about 7.5 Ma, after which at least 2 km of conglomerates accumulated in many areas in the re-entrant. Additional evidence for development of the MBT, at least in the western Himalaya, derives from apatite fission-track ages in northern Waziristan, Pakistan (Meigs et al., 1995). *En echelon* hanging-wall imbricates of the MBT comprising Cretaceous and Paleocene shelfal strata have yielded apatite ages of about 8–10 Ma. Given the estimated maximum burial temperature (~175°C) for these samples and their large vertical trajectory (>6 km), it is likely that uplift due to MBT thrusting was initiated 1–2 m.y. prior to cooling through 100°C (the approximate apatite annealing isotherm). Therefore, these three independent lines of evidence (subsidence, conglomerates, and cooling ages) support an initiation of the MBT in the western Himalaya at about 11 Ma, or 5–6 m.y. earlier than previously suggested (Burbank et al., 1988).

Neogene deposition between 8.5 Ma and 5 Ma

In the easternmost Kohat Plateau (Sultan Pathan) and the Trans-Indus ranges (Chichali) sediment accumulation rates increased slightly for rocks younger than about 9 Ma (Figures 9.7b and 9.8). From the central Potwar Plateau eastward across the foreland to Nepal, the sediment-accumulation data indicate that rates generally began to decrease around 9–8 Ma (Figure 9.7b). In many sites, that rate variation was accompanied by significant

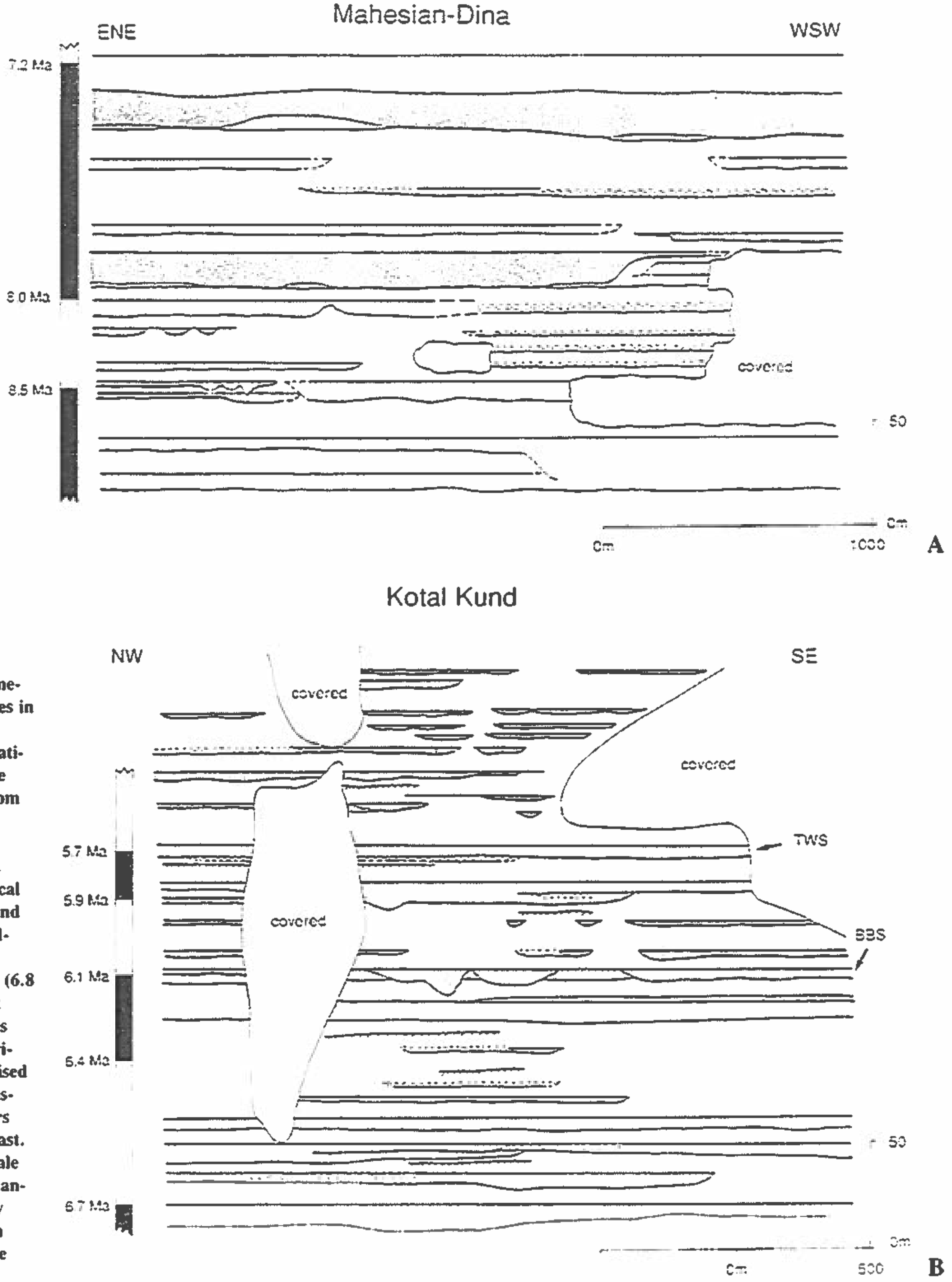

Figure 9.13. Outcrop geometries of channel sand bodies in the Potwar Plateau and Jhelum re-entrant. (a) Stratigraphic cross section at the Mahesian–Dina section from ~8.7 Ma to ~7.1 Ma. "Brown" sandstones are shown by darker patterns. Note the comparable vertical thicknesses of the brown and white sandstones. (b) Sand-body geometries at Kotal Kund in the late Miocene (6.8 to ~5.4 Ma). Beginning at ~6.1 Ma, white sandstones from a southeastwardly oriented paleo–Indus are incised by brown sandstones deposited by much smaller rivers flowing toward the northeast. As uplift continues, the scale of the brown-sandstone channels becomes progressively smaller. BBS, basal brown sandstone; TWS, top white sandstone.

lithofacies changes (Figure 9.9). In the Potwar Plateau, Jhelum re-entrant, and eastern Himachal Pradesh re-entrant, the abundance of coarse facies decreased abruptly at that time. That change defines the boundary between the Nagri and Dhok Pathan formations (Johnson et al., 1982b; Shah, 1977) in northern Pakistan and between the Nahan sandstone and the "Lower Alternations" (Johnson and Vondra, 1972) in Himachal Pradesh. In Pakistan, the major channel belts in the Dhok Pathan remained oriented (Figure 9.8) toward the southeast (Burbank and Beck, 1991b; Mulder, 1991). Limited paleocurrent data suggest a similar flow direction in the eastern Himachal Pradesh re-entrant.

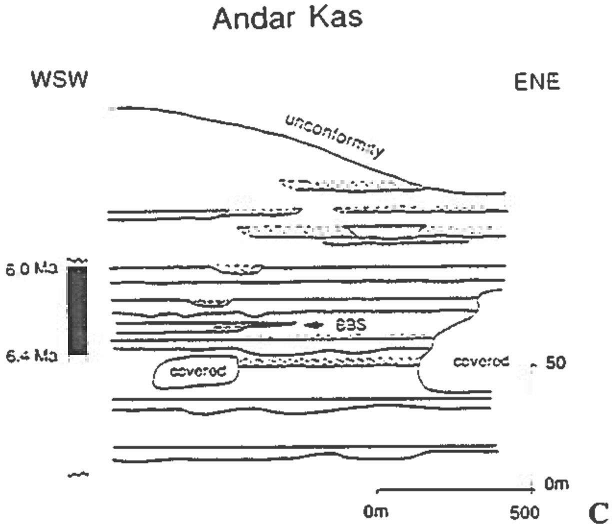

Figure 9.13. **(*cont.*) (c) Stratigraphic panel in Andar Kas, showing replacement of white sandstones by brown sandstones beginning at ~6.3 Ma. White sandstones disappear near the top of the exposure, and the remaining brown sandstones result from a much smaller, locally derived fluvial system.**

In contrast to the switch to less-sandstone-rich deposition in India and Pakistan, that interval in the Dehra Dun re-entrant (Figure 9.2) and in western Nepal brought increasingly coarse deposition. In the Dehra Dun area, middle Siwalik strata, deposited by rivers flowing toward the south and east, generally are both coarser-grained and more amalgamated than the underlying Nahan sandstone (Kumar, 1989; Kumar et al., 1991). At Surai Khola (Appel et al., 1991), coarse- to fine-grained sandstones up to 50 m thick dominate the section. The degree of channel amalgamation and mean grain size increased during that interval, providing evidence for considerably wider and deeper rivers than those that previously traversed the area. Very large scale trough cross-beds, gutters, scours, and channel margins provide clear evidence that flow was also to the east-southeast (Figure 9.11b). The scale of the bedding and orientation of the river suggest that a major axially oriented river dominated deposition in that region during the late Miocene. This could imply that the position of the axial river had shifted northward *toward* the hinterland in the late Miocene, given the scale and flow direction of the middle Miocene rivers in Nepal. If that occurred in response to flexural changes in the foreland, it would require either a more hinterlandward placement of the tectonic load (Sinclair et al., 1991) (a scenario consistent with the decreased rates of subsidence in that interval) or enhanced tectonic or erosional denudation in the forward part of the tectonic load. Alternatively, the greatly increased channel sizes and water discharges reconstructed for the Nepali foreland in the late Miocene might have resulted from the effects of increased monsoonal precipitation within the Himalaya. Several studies have suggested that the Asian monsoon significantly strengthened starting at about 8 Ma (Prell and Kutzbach, 1991; Harrison et al., 1992).

At present, the most detailed stratigraphic and sedimentologic descriptions of deposits of late Miocene age are from the Potwar Plateau and Jhelum re-entrant. Here, at least four different types of fluvial systems can be identified: (1) The major axial river was characterized by flow toward the east-southeast, thick (>10–15 m) channel sandstones, abundant blue-green hornblende in its dense-mineral assemblage (Figure 9.12), a "white" color to its sandstone facies, and conglomeratic clasts that, when compared with those of coeval fluvial systems, have higher relative proportions of intrusive, volcanic, and metamorphic clasts versus limestone and quartzite clasts (Mulder, 1991). Although likely to have been smaller in terms of discharge, such rivers appear analogous to the modern Indus River. (2) Major tributary rivers with considerable source areas within the hinterland may have been nearly as large as the axial river systems (Figures 9.1 and 9.13a). Those rivers typically were characterized by "brown" sandstones 10–20 m thick and more than 3 km wide (Raynolds, 1980); their flow tended to be transverse to the trend of the orogenic axis, with blue-green hornblendes being in low abundances (Figure 9.12), and intrusive clasts less common. Many of the brown sandstones in the Jhelum re-entrant deposited from southward-flowing rivers appear to represent an ancestral Jhelum River, carrying clasts distinctive of the Panjal Trap (Raynolds, 1980; Burbank and Raynolds, 1988). (3) River systems originating primarily within the foreland itself generally were represented by sandstone channel bodies of small dimensions (<10 m thick and 1–5 km in width). The sandstones typically are brown or "buff" in color (Behrensmeyer and Tauxe, 1982), have incised bases (Figure 9.14a), and have no preferred paleo-flow directions. High quartz abundances suggest recycling of previously deposited and partially weathered foreland strata. (4) Rivers originating primarily within the foreland, but having source areas controlled by tectonic deformation within the foreland itself, can commonly be distinguished from foreland rivers without tectonized catchments. The hanging walls of folds and thrusts, which disrupted preexisting drainage patterns and may have exposed new source areas to erosion, defined at least part of the catchments for these rivers. In comparison with the axial or major transverse rivers, the sand bodies deposited by these rivers were considerably smaller in vertical and lateral dimensions (Figure 9.13). When pre-molasse source rocks were exposed in folds, the derived sediments typically were texturally immature because of short travel distances and could contain unique lithologies that correlate with specific source areas. Coarse strata deposited from these rivers frequently include calcareous and iron soil concretions that were derived from erosion of uplifted foreland paleosols.

Of the strata produced by these four river types in the Potwar Plateau and Jhelum re-entrant, only the axial-river strata are immediately identifiable in the field, because of their distinctively light coloration. The other three systems produced varieties of "brown" sandstones that are distinguished on the basis of dimensions, orientation, and composition. Identification of the various fluvial systems, however, permits detailed reconstruction of the evolving foreland depositional system.

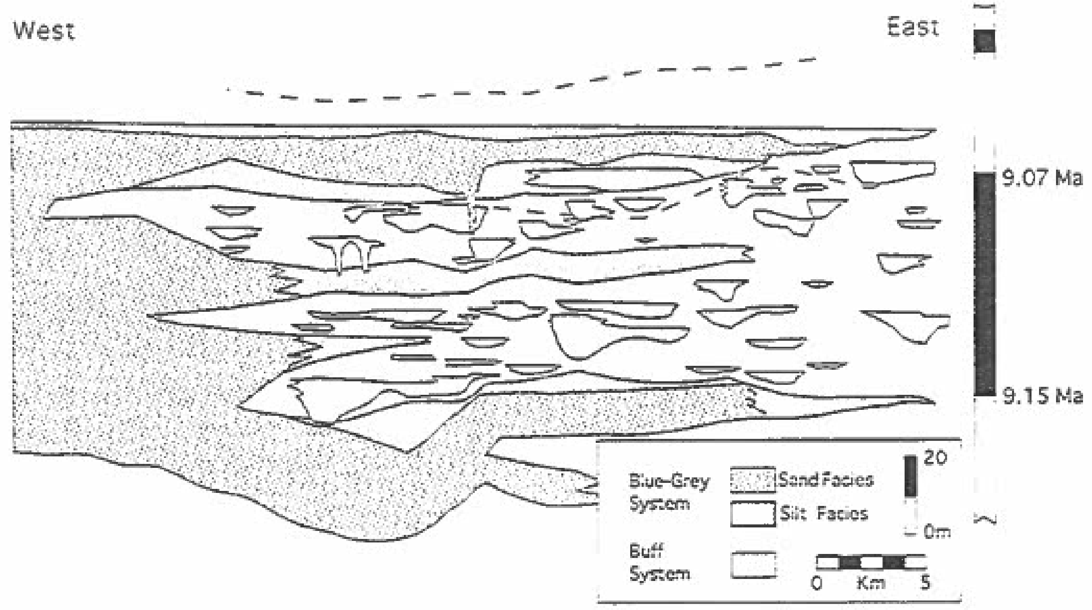

Figure 9.14. (a) Example of interfingering fluvial systems on the northern Potwar Plateau in the late Miocene (Behrensmeyer and Tauxe, 1982). The blue-gray ("white") sandstones represent the ancestral Indus River flowing toward the east; 5–20-m-thick sandstones can be traced laterally for >30 km. The much smaller "buff" sandstones represent southeasterly flowing rivers originating in the foreland and nearby foothills. The buff sandstones have higher abundances of quartz grains because of recycling of previously deposited foreland sediment, whereas the white sandstones have higher proportions of metamorphic lithic fragments.

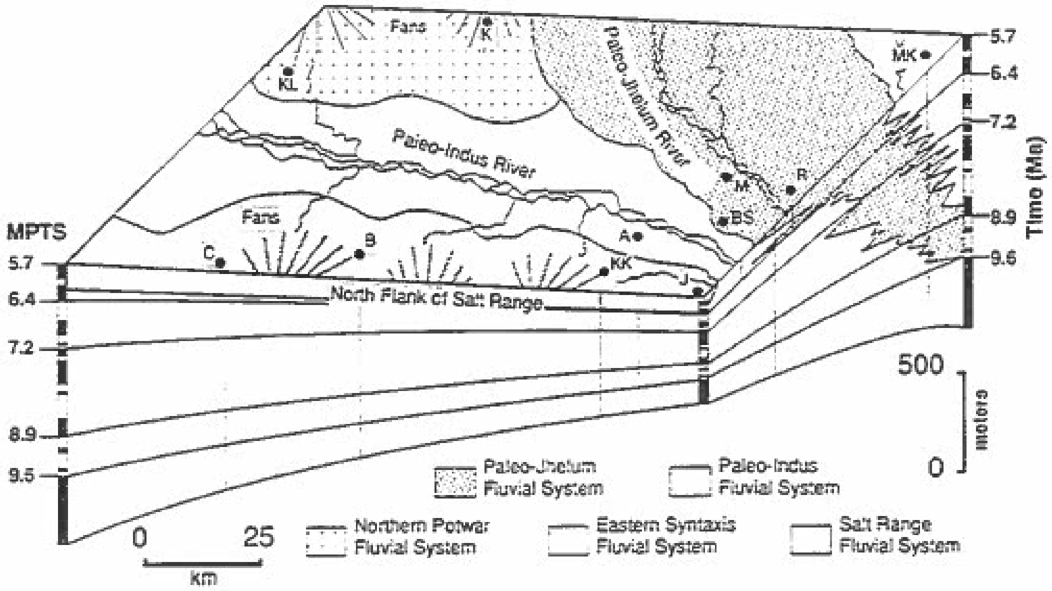

Figure 9.14 (b) Block diagram depicting basement subsidence, depositional facies, and the orientations of river systems in the Potwar Plateau during the late Miocene (Mulder, 1991). Magnetically defined isochrons show asymmetric deepening to both the north and the west. Salt range uplift creates a new source area at ~6.3 Ma and confines the Indus in the thrust-top basin formed above the Potwar detachment.

Late Miocene deposition in northern Pakistan can be subdivided into two intervals. Prior to about 6.4 Ma, the medial foreland was undeformed, whereas between 6.4 Ma and 4.5 Ma, the Salt range experienced its initial deformation (Burbank and Beck, 1989; Mulder and Burbank, 1993), and the nearby river systems were affected. During the earlier interval, "white" axial sandstones with "brown" transverse and locally derived sandstones were interfingered throughout the southern Potwar and across much of the northern Potwar and Jhelum re-entrant (Figures 9.13a and 9.14). Gradual southward displacement of the axial river system is suggested by successive disappearances of the white sandstones in the northern exposures: from Kas Dovac at about 9.5 Ma, from Khaur at about 7–8 Ma, and from Mahesian–Dina at about 5.5 Ma. A coeval southward displacement of the depocenter is seen in the subsidence patterns, in which the subsidence rates became lower in the northern Potwar, as compared with the central Potwar, for the first time after about 9.5 Ma (Figure 9.8). White sandstones predominated across the southern Potwar Plateau continuously until about 6.3 Ma.

In the northern part of the Potwar Plateau and Jhelum re-entrant, several contrasting fluvial systems are evident in strata younger than 9 Ma. Brown sandstones first occurred at Kas Dovac at about 9.5 Ma, and at Mahesian–Dina at about 8.5 Ma (Mulder, 1991). Whereas the brown sandstones at Mahesian appear to represent a major transverse river (Figures 9.13a and 9.14b) (probably a paleo-Jhelum River), smaller, locally derived, southwardly flowing systems (the northern Potwar fluvial system) predominated at Kas Dovac, Khaur, and Mawa Kaneli

(Behrensmeyer and Tauxe, 1982; Mulder, 1991) (Figure 9.14b). At Kas Dovac (Figure 9.15), the presence of both calcite nodules and weakly resistant gypsum clasts in brown sand bodies by about 7.5 Ma suggests short transport distances and the likely exposure of Eocene limestones and evaporites in the Margala or Kala Chitta ranges to the north. That deformation could be attributable to continued movement on the MBT or to development of the "Northern Potwar Deformed Zone" (Leathers, 1987; Baker et al., 1988) at that time. Relatively thin-bedded sandstones at both Kas Dovac and Mawa Kaneli in the late Miocene suggest that those areas were consistently dominated by local drainages. At Kas Dovac, the abundance of second-cycle, monocrystalline quartz in the sandstones (Critelli and Ingersoll, 1994) also suggests derivation from deformed molassic strata within the foreland. Although the ancestral Jhelum River apparently flowed between these sites (Figure 9.6), this major transverse river never or rarely flowed across them. This absence suggests that the structural control exerted by the Jhelum re-entrant on the position of the Jhelum River may have effectively focused the river down the axis of the re-entrant (Meigs et al., 1995) since early in the history of MBT deformation (Raynolds, 1980).

Although deformation along the proximal margin of the foreland is suggested by conglomeratic beds and locally sourced rivers in the northern foreland, no record of major deformation within the foreland itself is apparent until the Salt range began to form. The Salt range lies at the southern edge of the Potwar allochthon, which rides above a gently (2–3°) northward-inclined detachment localized primarily within the evaporites of the Salt Range Formation (Lillie et al., 1987). The unusual width of the allochthon (>100 km) appears to have resulted from very efficient decoupling at the evaporite-dominated detachment. The position of the footwall ramp in the southern Potwar coincides with a basement normal fault (Baker et al., 1988).

At the time that the Salt range thrust first propagated beneath the Potwar Plateau, it represented a major southward jump in the locus of deformation. Most significantly in terms of foreland deposition, a zone of uplift suddenly developed in the midst of the previously subsiding basin (Burbank and Beck, 1989). The impact of uplift on fluvial systems in the southern Potwar region was dramatic. Beginning at about 6.3 Ma, white sandstones deposited from a southeastwardly flowing axial (paleo–Indus) river were replaced by brown sandstones of much smaller dimensions (Figure 9.13b,c). Following a brief interval of interfingering between these systems, the white sandstones were completely displaced from the southern Potwar sections by about 5.7 Ma (Mulder, 1991).

The sandstones that supplanted the white sandstones were products of a local, northeastwardly flowing river system that carried distinctive clasts of red granites (Talchir clasts) derived from Paleozoic strata exposed in the Salt range (Burbank and Beck, 1989). In addition to Talchir clasts, the conglomeratic facies in the replacement strata generally have abundant clasts representing carbonate and iron concretions. Such clasts apparently resulted from stripping of Siwalik paleosols during uplift, and they are uncommon in the underlying Siwalik sandstones. Further evidence supporting a Salt-range source for these brown sandstones is provided by the increasing clast angularity, grain size, and abundance of conglomerates in sections more proximal to the Salt range (Mulder, 1991). The temporal gap between the first appearance of brown sandstone (in response to initial uplift and reorganization of the drainages) and the first appearance of identifiable Paleozoic clasts, which required erosional exposure of previously buried strata, sets some limits on the probable rates of erosion within the early Salt range. It seems likely that during the first several kilometers of shortening, and over a period of less than 1 m.y. (Burbank and Beck, 1991a), rivers originating in the Salt range incised about 3 km through both molassic strata and the underlying Mesozoic and Paleozoic strata that were carried in the Potwar hanging wall.

The Salt-range thrusting thus created a broad thrust-top basin extending into the medial foreland. Prior to the rise of the Salt range, the course of the paleo–Indus River flowed unconstrained toward the southeast across the area where the Salt range developed (Burbank and Beck, 1991b). Following structural partitioning of the foreland, the river could have been shifted to a more distal position south of the Salt range; instead, it was displaced northward toward the axis of the newly defined thrust-top basin (Figure 9.15). That position of the paleo–Indus was facilitated by the absence of structural closure along the eastern margin of the Potwar Plateau and by the very efficient detachment beneath the allochthon that permitted the region to be translated southward with negligible internal deformation. White sandstones were deposited along the axis of the thrust-top basin until about 5 Ma at Rohtas, Mahesian, Ganda Paik, and Basawa–Sanghoi Kas (Figure 9.15). Given the northward tilt of the basement beneath the detachment, however, overall subsidence could not persist indefinitely in the thrust-top basin. Southward translation of the allochthon along the 2–3° slope resulted in vertical uplift of the thrust-top basin. By 5.0 Ma, the Indus was unable to maintain its course across the uplifting Potwar Plateau and was shunted to the west, presumably to a location close to its present position (Figure 9.15). From that early Pliocene event onward, a southward-flowing paleo–Jhelum River dominated deposition in the eastern Potwar and the adjacent Jhelum re-entrant. At sites like Jamarghal near the eastern termination of the Salt range, this sequence of events is recorded by an initial replacement of white sandstones by beige-colored sandstone channels of small dimensions. These are interpreted to represent drainages sourced in the incipient Salt range. By about 5 Ma, the buff sandstones were supplanted by large-scale brown sandstones of the paleo–Jhelum system that are presently traceable across the landscape for many kilometers.

Deposition from 5.0 Ma to the present

There are several key differences between the Miocene record of the foreland basin and the Plio–Pleistocene record. First, although deformation presumably was occurring along the proximal foreland basin throughout its evolution, only the

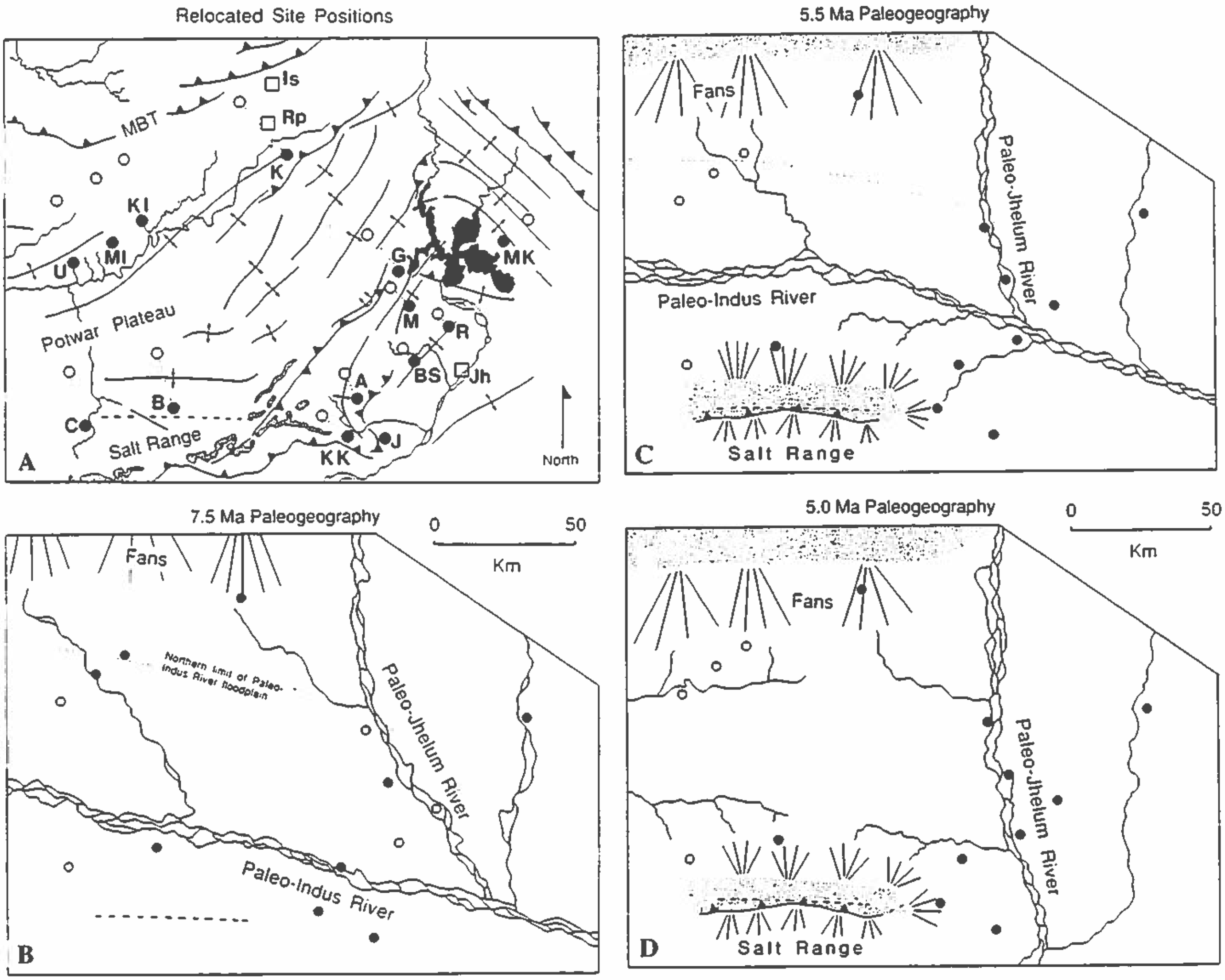

Figure 9.15. Late Miocene and early Pliocene paleogeographic reconstructions in the Potwar Plateau and Jhelum re-entrant. (a) Sites used in 7.5-, 5.5-, and 5.0-Ma reconstructions. Each site has been relocated (open circles) to account for shortening since 6.5 Ma. Site abbreviations: A, Andar Kas; BS, Basawa-Sanghoi Kas; B, Bhaun; C, Chinji; G, Ganda Paik; J, Jamarghal; K, Kas Dovac; KL, Kaulial Kas; KK, Kotal Kund; M, Mahesian-Dina; MK, Mawa Kaneli; ML, Malhuwalla Kas; R, Rohtas; U, Utran Kas. Geographic locations: Is, Islamabad; Jh, Jhelum; Rp, Rawalpindi. (b) The 7.5-Ma paleogeography. Filled circles represent sites for which there are data of that age. Note the axial trend of the paleo-Indus and the presence of both major (paleo-Jhelum) and lesser tributary rivers from the northern foreland. The northern limit of the Indus floodplain is based on the presence or absence of white sandstone.

Figure 9.15 (c) Initial uplift of the Salt range at ~6.3 Ma caused displacement of the paleo-Indus northward into the axis of the thrust-top basin. (d) Within 1.3 m.y., uplift and transport of the Potwar allochthon and nearby deformation shunted the paleo-Indus to the west of the Potwar Plateau. The transverse paleo-Jhelum dominated deposition in the eastern regions.

proximal deposits and associated folds and thrusts of Plio–Pleistocene age are extensively preserved and exposed at present. In fact, it is this tectonism that has caused uplift and erosion of many of the presently exposed strata of the foreland. In areas without deformation, ongoing deposition has buried the Plio–Pleistocene section. Second, major climatic changes in the late Cenozoic caused expanded Himalayan glaciation and changed the previous regimes of erosion and sediment discharge within the Himalaya and the adjacent foreland. Thus the youngest strata exposed in the foreland have commonly been affected both by proximal tectonism and by a climatic regime that contrasts strongly with that which prevailed during the Miocene.

Traditionally, the "Boulder Conglomerate" was thought to represent deposition during the Quaternary in many areas (Gansser, 1964). Improved chronologies, primarily due to magnetostratigraphy and some dating of volcanic ashes, have shown that the presence or absence of conglomerates is not a reliable temporal indicator in the foreland. At present, more than 30 magnetic sections (Opdyke et al., 1979; Azzaroli and

Napoleone, 1982; Johnson et al., 1982a,b; Burbank and Johnson, 1983; Tandon and Kumar, 1984; Raynolds and Johnson, 1985; Tokuoka et al., 1986; Khan et al., 1988; Ranga Rao et al., 1988) span portions of the past 5 m.y. of foreland deposition (Figure 9.6). Unfortunately, nearly all of these are concentrated in the Indus foreland of northwestern India and Pakistan (Figure 9.5). Detailed sedimentologic descriptions and location maps are available for about half of these dated sections.

Because of the propagation of deformation into the foreland, few systematic patterns are discernible in accumulation and subsidence rates (Figure 9.7). In the Jhelum re-entrant and the Indian Punjab, for example, three sections (Figure 9.7c) show increasing accumulation rates between 4 Ma and 5 Ma: Ganda Paik, Rohtas, and Parmandal (Raynolds and Johnson, 1985; Burbank et al., 1988; Ranga Rao et al., 1988). Several nearby sections, such as Mawa Kaneli and Kotal Kund, however, show no significant changes through that same interval. For many of the sections in the Jhelum re-entrant and the Indian Punjab, rate decreases occurred between 3 Ma and 2 Ma, whereas sections at Jalalpur, Pabbi hills, and Kas Guma apparently show no change or even increases for that same interval and subsequent intervals (Figure 9.7c). The irregular rate changes may in part be attributable to the locations of the sections of the flanks of folds that developed during the Plio–Pleistocene. Furthermore, seismic data from the Hazara region west of the re-entrant (Seeber, Armbruster, and Quittmeyer, 1981) and analysis of reflection seismic and outcrop data from the Kohat Plateau and Peshawar Basin (Burbank and Tahirkheli, 1985; Pivnik, 1992; Pivnik and Sercombe, 1993) indicate that transpressional deformation was prevalent in that part of the foreland. Dominantly strike-slip regimes generally produce very irregular subsidence and uplift patterns, and transpressional regimes are also likely to show closely spaced rate changes adjacent to faults.

An analysis of the geographic distribution of Plio–Pleistocene accumulation rates in the Jhelum re-entrant (Raynolds and Johnson, 1985) suggests both a spatial trend and a temporal trend. A northeast–southwest axis of persistent rapid accumulation was present from 3.4 Ma to about 2 Ma, and rates in the northern sites tended to diminish in the Pleistocene, whereas rates in the south increased. The axis of subsidence is oriented slightly obliquely to the center of the structural re-entrant. It is not clear whether this enhanced subsidence resulted from loading of an inherited weakness in the underlying basement, from constructive interference due to the imposition of two thrust loads (the Himalaya and Hindu Kush) in nearly orthogonal orientations, or simply from sediment loading. The shifting of more rapid rates toward the south in the latest Pliocene and early Pleistocene appears likely to have been a response to increasing deformation of the northern basin margin.

In contrast to the irregular accumulation patterns east of the Potwar Plateau, accumulation in the Trans-Indus region shows some distinctive trends (Khan et al., 1988). This region displays the most rapid, sustained accumulation rates found in the dated Neogene and Quaternary foreland record. Many of the accumulation curves, moreover, show increasing rates of accumulation for strata younger than 2.4–1.6 Ma (Figure 9.7c). A final unusual characteristic of the Trans-Indus region is that molasse deposition began in many of the sections only at about 4 Ma (Figures 9.4 and 9.6). Where the bases of the measured sections are exposed, the Pliocene foreland strata in the Marwat and Bhittani ranges rest on pre-Siwalik rocks. The Trans-Indus ranges, therefore, appear to represent an area of very rapid and young foreland deposition. Less than 50 km to the north at Chichali, or 50 km east in the Potwar Plateau, foreland deposition began more than 10 m.y. earlier. The young age of basal Trans-Indus strata may simply represent a slower-than-usual migration of the distal basin pinch-out, possibly due to local basement uplift or a preexisting basement high.

The Plio–Pleistocene deposition in the Bhittani range occurred in three phases (Figure 9.16). Between about 4.1 Ma and 3.6 Ma (Kargocha Formation) (Khan et al., 1988), mudstone with interbedded sheetlike and ribbonlike red sandstone and conglomerate dominated deposition and was the product of small, locally sourced fluvial systems (Nio and Hussain, 1984). As evidenced by channel sandstone complexes as much as 100 m thick and more than 10 km wide, and by south- to southwest-directed paleo-currents within the Marwat Formation (Nio and Hussain, 1984), the Indus River flowed across this region between about 3.6 Ma and 0.7 Ma (Khan et al., 1988). Until the Indus was shunted to its present position east of the Bhittani range after 0.7 Ma, accumulation rates actually accelerated, reaching as high as 1 km/m.y. during that interval. Presumably the diversion of the Indus resulted from incipient thrusting in the Trans-Indus ranges. As much as 200 m of upward-coarsening sandstone and conglomerate in the overlying Malaghan Formation accumulated as terminal fan complexes within the newly formed intermontane basin (Nio and Hussain, 1984) after about 0.7 Ma (Khan et al., 1988).

Whereas there is little or no record of deposition during Plio–Pleistocene times across most of the Potwar Plateau, the Peshawar and Campbellpore intermontane basins became depositional centers during that interval (Burbank and Tahirkheli, 1985). Alluvial fans that were shed northward off the Attock–Cherat ranges ring the southern margin of the Peshawar Basin, whereas fluvial and lacustrine strata dominate the central parts of the basin. The frequent occurrence of lacustrine strata suggests that uplift along the MBT and related faults may have episodically ponded the fluvial systems as they traversed water gaps in the bounding ranges. Whereas previously it was thought (Burbank and Tahirkheli, 1985) that the oldest preserved intermontane-basin strata might provide close limits for the timing of initial MBT thrusting, it appears more likely that much of the accumulation postdated major motion on the MBT and occurred in response to rates of regional subsidence overtaking rates of local uplift.

Deposition in the Peshawar Basin has been strongly affected by catastrophic flooding during the past 700,000 years (Burbank, 1983b; Burbank and Tahirkheli, 1985). At least two generations of upward-fining, graded rhythmites occur on the southern margin of the basin. These strata are sedimentologically similar

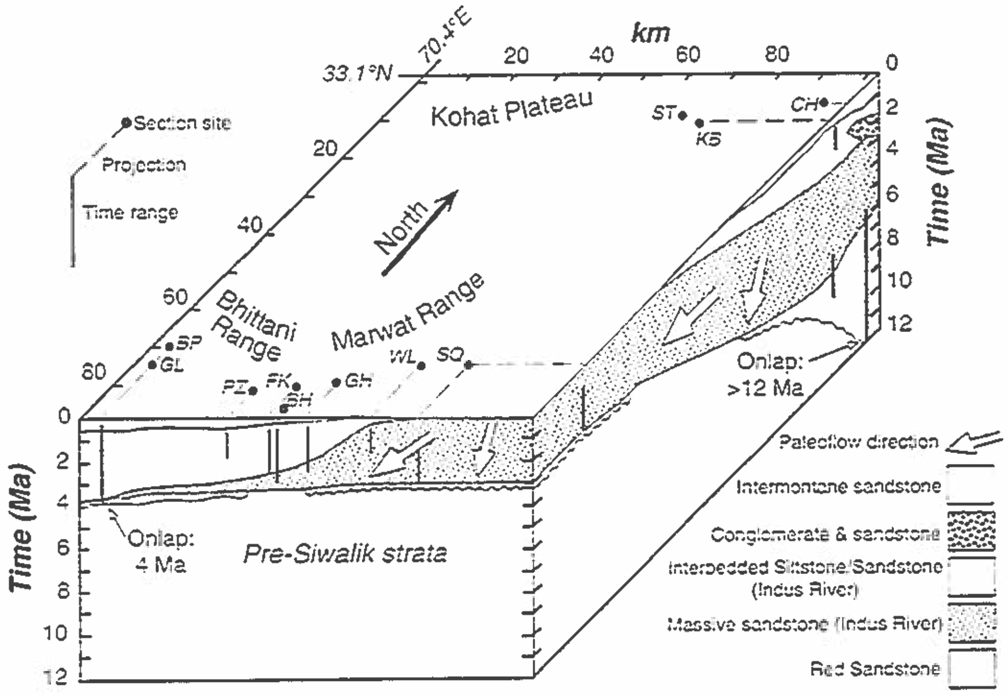

Figure 9.16. Block diagram of the Trans-Indus region during Neogene and Quaternary deposition (Khan, 1983). Onlap in the Bhittani range at ~4 Ma occurred shortly before thick sand bodies of the paleo–Indus River occupied the area. Onlap was much earlier near Chichali, 60 km north of the Bhittani range. Diversion of the Indus to the west and development of intermontane strata due to thrusting occurred after 0.7 Ma. Vertical lines show temporal extent of sections as projected on the sides of the block. BH, Bhandara; BP, Bain Pass; CH, Chichali; FK, Faqir Kille; GH, Charangai; GL, Garhi Landa; KB, Khora Baroch; PZ, Pezu; SQ, Spin Qammar; ST, Spalmai Tangi; WL, Walewal.

to slack-water deposits generated by the Missoula floods: giant, glacial-outburst floods in Montana and Washington (Waitt, 1980). It appears that on numerous occasions in the past, breakage of dams formed by landslides or glaciers along the Indus River and its tributaries within the hinterlands caused massive flooding in the intermontane basins.

The 1.8–1.6-Ma Lei Conglomerate, which caps the Kas Dovac and Soan sections in the northeastern Potwar, has received considerable attention because it unconformably overlies Siwalik strata as young as 2.1 Ma (Johnson et al., 1982b; Burbank and Raynolds, 1984; Raynolds and Johnson, 1985; Pivnik, 1992). Very rapid late Pliocene deformation is indicated by the overlap of a 6-km-thick panel of nearly vertically dipping strata in the northern limb of the Soan syncline by horizontal conglomerates that are only about 0.3 m.y. younger than the youngest strata in the vertical panel. It appears that folding of the presently vertical strata is attributable either to thrusts related to the MBT (Burbank and Raynolds, 1984) or to a partially buried strand of the Khair-i-Murat fault (Pivnik, 1992), which is an imbricate within the Northern Potwar Deformed Zone. Conglomerate in the southern limb of the Soan syncline also appears to have resulted from folding and thrusting. Within the Kas Dovac section, a northward-tilted unconformity dated at about 3.4 Ma is interpreted to be a response to southward-vergent thrusting along the Riwat fault that defines the southern limit of the syncline (Raynolds, 1980; Burbank and Raynolds, 1988). Strata below the unconformity have experienced a vertical-axis rotation that is about 20° greater than that found in the overlying strata (Burbank and Raynolds, 1988). At the unconformity, limestone-bearing conglomeratic strata are replaced by conglomerate dominated by quartzites and volcanic clasts. This change suggests that source areas in the Margala hills were replaced by sources in the Pir Panjal and areas feeding the northern part of the Jhelum re-entrant. In response to the activity along the Riwat fault and its propagation to the northeast, northward tilting of the hanging wall apparently caused proximal erosion of the hanging wall, displaced those drainages emanating from the Margala hills farther to the north, and diverted rivers from the axial part of the Jhelum syntaxis into an oblique trough that formed as a small thrust-top basin above the fault.

Unlike the situation for the Trans-Indus region, continuous Miocene-to-Pleistocene deposition was recorded at many sites in the Jhelum re-entrant (Raynolds, 1980; Raynolds and Johnson, 1985). Analysis of coeval strata among numerous sections has demonstrated that the axis of maximum subsidence corresponded geographically with the depositional axis (Raynolds, 1980). That fluvial axis was characterized by thick channel sand bodies that were flanked laterally by broad zones of thinner interbedded sandstones and overbank siltstones. That pattern suggests that the primary course of the paleo–Jhelum River was continuously localized along the zone of more rapid subsidence in the axis of the re-entrant.

Complex interfingering of local fluvial systems with the major, transverse Jhelum system occurred along the margins of the zone of maximum subsidence during the late Pliocene and Pleistocene. Various fluvial systems can be identified on the basis of their clast lithologies. For example, in Basawa Kas at about 1.8 Ma and thereafter, sandstones with quartzite and volcanic clasts of the paleo–Jhelum system were interbedded with conglomerate comprising Talchir (red granite) and limestone clasts (>10 cm in diameter) derived from the Salt range to the southeast. The clear provenance of these latter clasts suggests that rivers sourced more than 30 km away within intra-foreland uplifts were capable of transporting coarse sediment considerable distances into the foreland.

Although we have data from many magnetic sections located on the flanks of young folds in the Jhelum re-entrant and easternmost Potwar Plateau (Figures 9.3 and 9.5), precise ages for the development of the various structures are difficult to calculate. A decrease in the local accumulation rate at a site may reflect the initiation of relative uplift. The age of the youngest preserved strata on the flanks of the folds provides a maximum age for the surface expression of the fold. These limiting ages form temporal progressions that become younger toward the south on both sides of the re-entrant, but show no coherent pattern when compared across the axis of the re-entrant (Figure 9.4). The interference pattern of folds and thrusts in the re-entrant suggests that two separate thrust systems met there. Thus, the eastern folds and thrusts may have been controlled by encroachment of Himalayan deformation on the foreland, whereas the western half of the re-entrant likely was controlled by tectonism in Kohistan and the Hazara regions. The boundary between these two domains approximately coincides with the edge of the eo-Cambrian evaporitic basin.

Conglomerate became increasingly common in the Jhelum re-entrant after about 3 Ma (Raynolds, 1980). Extrabasinal conglomerate advanced from north to south at a rate of about 3 cm/a (Raynolds and Johnson, 1985; Burbank et al., 1988). Whereas deposition of conglomerates in the north (e.g., Sakrana or Rata–Dadial, Figure 9.5) was nearly uninterrupted following their first appearance (Figure 9.9), at many of the more southern sections the conglomerates were isolated and interbedded with sandstones and siltstones. Although these conglomeratic beds have been cited as examples of both syn-thrusting (Burbank et al., 1988) and post-thrusting (Heller et al., 1989) gravel progradation, it is likely that many of the isolated conglomerate bodies in the south represent nonequilibrium progradation (Paola, Heller, and Angevine, 1992) that occurred in response to short-lived, climatically induced changes in the regional hydrologic regime, following the initiation of widespread glaciation in the Northern Hemisphere after 2.6 Ma (Shackleton et al., 1984).

Conglomerate deposition also occurred in the Plio–Pleistocene sections of the Indian Punjab (Ranga Rao et al., 1988). At Jammu–Nagrota, conglomerate first appeared at about 2 Ma and dominated sedimentation from 1.6 Ma onward (Figure 9.9). Only 20 km to the southeast, conglomerate did not appear until about 1 Ma and was not prevalent until after 0.7 Ma (Ranga Rao et al., 1988). The possible relationship of these conglomerates to the growth of the Suruin–Mastgarh anticline is unknown. Farther east, in the Subathu sub-basin (Figure 9.2), conglomeratic deposition from southward-flowing braided rivers began at about 1.5 Ma (Tandon and Kumar, 1984). The underlying fluvial strata record large-scale drainage reversals that have been interpreted to have been responses to deformation within the Punjabi foreland between about 3 Ma and 1.5 Ma (Tandon and Kumar, 1984). Undated/Plio–Pleistocene boulder conglomerates in the Dehra Dun sub-basin both predate and postdate the growth of the Mohand anticline. The older conglomerates flowed across the present crest of the structure, whereas younger gravels were trapped within the newly defined thrust-top basin (Kumar et al., 1991).

At Surai Khola in Nepal (Appel et al., 1991), conglomerates became prevalent beginning at about 3 Ma. Whereas axially (southeasterly) oriented paleo-currents characterize the underlying Siwalik strata, south-, east-, and west-flowing rivers deposited the coarse upper Siwalik strata (Figure 9.11c). Progressively rotated syn-tectonic strata indicate that thrusts began to disrupt that part of the foreland during late Pliocene times. Although deposition continued within "duns" (valleys formed between adjacent anticlines) well into the Quaternary, eroding hanging walls above thrusts created highly variable source areas, and paleo-currents indicate flow off both the forelimbs and backlimbs of folds, as well as along the structural troughs that lay parallel to their crests.

Throughout much of this Plio–Pleistocene interval, lacustrine deposition (Karewa Formation) prevailed in the Kashmir intermontane basin to the northwest of the Jhelum re-entrant (Burbank and Johnson, 1983). The oldest dated strata in Kashmir are from about 4 Ma and are underlain by undated strata that probably date as far back as 5 Ma. Conglomerates initially prograded across the basin at about 3 Ma, and then in several pulses after 1.8 Ma. Whereas the older conglomerates appear to have been derived from the High Himalaya, most of the younger conglomerates apparently were derived from the uplifting Pir Panjal range to the southwest (Burbank and Johnson, 1983). Since about 0.5 Ma, 1.5–3 km of uplift of the Pir Panjal range has been inferred from the presence of Karewa strata of unknown age at altitudes higher than 3,500 m, more than 1.5 km above the Karewas along the northeastern flank of the range.

Modern depositional patterns

Along the margins of both the Indus and Gangetic forelands, growing anticlines interact with present-day fluvial systems. Actively aggrading duns are present behind many of these anticlines. Rivers either have been diverted around their plunging ends or else flow across them through water gaps. Where there are water gaps, there is a greater tendency to have low-gradient or "ponded" fluvial systems upstream of the folds. Many of the modern rivers enter the foreland either through structural re-entrants or along transverse structures, such as wrench faults.

There is an interesting contrast between the Indus and Gangetic forelands in the positions of the major rivers with respect to the active fold belts on the proximal margin of the foreland. In India, the Ganges and Yamuna rivers flow along the distal margin of the foreland, and transverse tributary rivers flow several hundred kilometers across the foreland to join them (Figure 9.1). In some areas, the Ganges and Yamuna appear to be flowing at the feathered edge of deposition and to have no significant accumulation of older molasse beneath them (Burbank, 1992). This geometry represents a strong contrast both with pre-Pliocene deposition in the Gangetic foreland, where axially flowing rivers dominated deposition in the *medial* and distal foreland basin (Burbank and Beck, 1991b), and with the modern Indus foreland, where the Indus River flows close to the active fold belt. In Pakistan, the transverse tributary rivers are generally short, and the Indus occupies the proximal-to-medial part of the foreland, where it flows above a thick succession of earlier Siwalik strata.

The Jhelum River also flows close to the frontal zone of the Salt range. A pattern of abandoned river bends that are open to the north is visible in satellite images, suggesting (Leeder and Gawthorpe, 1987) that the Jhelum has recently migrated northward to its present position. This hinterlandward migration may be a response to decreased post-glacial sediment discharges from the Salt range. Under a regime of constant foreland subsidence, enhanced sediment fluxes during glacial intervals would tend to push the axial river farther from the range front. Subsequent decreases in fluxes emanating from the Salt range would permit the Jhelum to migrate back toward the region of higher subsidence. Seismic and borehole data show that folding is occurring south of the Salt range at Lilla (Baker et al., 1988). Additional folding in the flood plain farther to the south could also be causing the migration of the Jhelum River. Such an interpretation would be consistent with the southeastward migration of the Chenab River, as is apparent from abandoned meanders seen on satellite images, which would situate the Chenab on the southeastern flank of an incipient fold.

Active uplifts across which the Indus flows appear to exert strong control on the morphology of the Indus channels (Jorgensen et al., 1993). Fluvial incision and elevated terrace remnants mark zones of uplift, whereas anastomosing channels occur upstream of the uplifts in reaches with lowered gradients. Increases in sinuosity and meandering tend to characterize the steepened reaches downstream of the axes of uplift (Jorgensen et al., 1993).

Discussion

Depositional patterns

At a large scale, the foreland depositional system can be conceptualized as comprising two types of rivers: axial rivers oriented parallel to the orogenic axis, and transverse rivers oriented perpendicular to the mountain front. Whereas the former typically flow across the medial and distal foreland basin, the latter commonly dominate the proximal foreland. The point at which these two systems merge is determined primarily by the pattern of subsidence across the basin and the sediment fluxes carried by the transverse rivers. High proximal subsidence rates and low transverse fluxes draw axial rivers into the proximal part of the foreland, whereas slow subsidence and high transverse sediment feeds push the axial rivers toward the distal part of the foreland.

The available paleo-current, isopach, and subsidence data indicate that during Miocene times, an axial river dominated the medial and distal foreland (Burbank and Beck, 1991b). In northwestern Pakistan, the paleo–Indus is interpreted to have flowed toward the southeast into the Ganges drainage. No definitive documentation, however, of the continuity of the Indus system with the Ganges system has yet been discovered. We need clearer fingerprinting of the Indus characteristics along its hypothesized course and a better understanding of the effects of downstream dilution by tributaries. Data from Nepal also indicate the presence of an axial river system in the medial foreland during the Miocene, but the data are local and fragmentary.

Within major structural re-entrants where there are available data, extensive and temporally persistent transverse fluvial systems are evident for the Miocene (e.g., the Himachal Pradesh re-entrant and the Jhelum re-entrant). Re-entrants along the Himalayan front effectively focus major rivers along their axes, typically above a zone of enhanced subsidence. For Plio–Pleistocene times, transverse systems become evident wherever they are studied along nearly the entire northern margin of the foreland. Superimposed on this overall transverse trend, there are complex local variations that resulted from the propagation of folding and faulting into the foreland. In the Trans-Indus area in the westernmost foreland, however, the Indus flows as a major axial river very close to the front of the transpressional ranges, and transverse rivers are limited in extent within this part of the foreland.

Conglomerate compositions and heavy-mineral assemblages provide useful data on provenance, source-area lithology, and dispersal patterns. Because there exist strong lithological contrasts in conglomerates and sandstones derived from uplifted areas, such as the Salt range, Kohistan, the Pir Panjal, or the Margala hills, it is possible to determine the patterns of interfingering and the fluctuating geometry of rivers in the foreland through time. A clearly dominant axial river in the Miocene record of northern Pakistan was replaced in the latest Miocene to Pleistocene by a complex pattern of interfingering fluvial systems. Development of thrust faults and folds within the foreland controlled changes in the surface topography and resulted in major rearrangements of the depositional system. Large-scale shifts in the positions of major rivers appear linked to major faults, such as those bounding the Salt range and the Trans-Indus ranges. Smaller-scale deformation, such as that associated with the Riwat fault or the folds in the Jhelum re-entrant, appears in part to have modulated the shifts in the transverse and small local rivers.

Thrust sequencing and timing

It is now well established that out-of-sequence thrusting has occurred on a large scale within the foreland (Burbank and Raynolds, 1988; Burbank and Beck, 1989). The Salt-range thrust represents a 100-km-wide southward step in deformation that was initiated at about 6.3 Ma (Figure 9.17). Subsequently, more

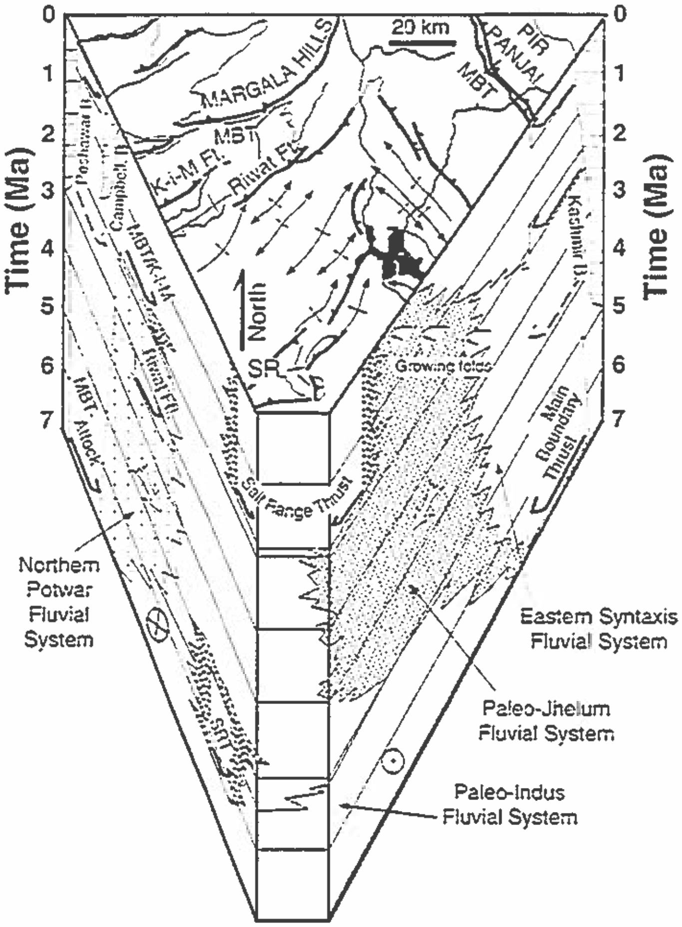

Figure 9.17. Block diagram of the Jhelum re-entrant and eastern Potwar Plateau showing timing of thrusting, spatial relationships among depositional systems, and growth of folds. Parallelograms filled with dark shading define intervals of thrusting. Intervals of possible thrusting are lightly shaded. Out-of-sequence thrusting is displayed by the northward progression from the Salt-range thrust (SRT) to the Riwat fault and then to the MBT or Khaur-i-Murat fault (K-i-M). Largely fine-grained, intermontane deposition was primarily associated with quiescent phases.

northerly thrusting of the Riwat fault (~3.4 Ma) and the Khair-i-Murat fault or MBT at about 2.0 Ma adjacent to the Soan syncline represents a hinterlandward thrust progression along the eastern margin of the Potwar allochthon. The most likely circumstances under which that hindward sequencing could be reinterpreted would be those in which such northern thrusts had actually been active for long intervals of time (prior to 6.5 Ma), with only the last motions being recorded and dated by nearby strata. This possibility is argued against by the apparent continuity of deposition within the tilted panels of strata that underlie the unconformities used to define deformation. They show no clear evidence of earlier tilting or deformation. It has also been suggested that the apparent out-of-sequence thrusting actually represents a continual forward progression of thrusting along a series of detachments located at progressively deeper stratigraphic levels. In this context, an apparent out-of-sequence thrust would actually be the expression of the outermost thrust that soled into a deeper detachment level and cut upward through a higher-level detachment that had itself been previously active. The seismic data from northern Pakistan (Pennock et al., 1989) appear to preclude this possibility: Both the Riwat fault and Salt-range thrust root in the same detachment (Figure 9.3).

One of the major conclusions arising from this synthesis is that the MBT began significant movement at about 11 Ma. Calculation of the timing of initiation of a major thrust fault is generally difficult, because erosion has removed those strata directly associated with the thrust that could unambiguously record its motion (Burbank and Raynolds, 1988; Burbank et al., 1992). A minimum age limit can be derived from sediment deposited and preserved in intermontane basins blocked by hanging-wall uplifts. For example, the ages of the Karewa strata in Kashmir (Figure 9.17) indicate that the MBT was active by 4–5 Ma (Burbank and Johnson, 1983). Similarly, the fill of the Peshawar intermontane basin indicates that MBT thrusting commenced prior to about 3 Ma (Burbank, Raynolds, and Johnson, 1986). It is however, likely that these ponded basins commonly represent the latter stages of motion, during which the rate of thrusting slowed and regional subsidence moved the basin from a regime of uplift and erosion as part of an active hanging wall to a regime of relative quiescence and deposition. Nonetheless, the uplift of the Karewas in Kashmir by 1.5–3 km indicates that several kilometers of shortening on the MBT occurred during the Pleistocene.

The acceleration of subsidence rates at about 11 Ma in every dated section of that age across the foreland (Figure 9.7b) provides strong evidence for enhanced thrust loads, most likely in a position closer to the foreland – presumably the MBT. Although the MBT zone exists along the entire length of the Himalaya, there is no requirement for its initiation to have been a synchronous event throughout the range. The evidence for enhanced subsidence is clearest in the northwestern foreland, where there are numerous dated sections with similar records. Nonetheless, the existing data from Nepal and to a lesser extent from Himachal Pradesh are also consistent with initial thrusting during that same interval and support a relatively synchronous initiation of thrusting along a system of thrusts, now termed the MBT. Very significantly, the thick deposits of conglomerates found in the central part of the Himachal Pradesh re-entrant and recently dated at about 8.8 Ma (Meigs et al., 1995) provide clear evidence for significant uplift and erosion within the Lesser Himalaya at that time. The presence of these conglomerates within the medial foreland is attributable both to syn-tectonic progradation and to focusing along the re-entrant.

The natural variation of accumulation rates due to autocyclic depositional processes at time scales of less than 0.5 m.y. (Johnson et al., 1988) dictates that tectonic interpretations of short-term rate variations must be made cautiously. When numerous closely spaced sections are examined, rate changes of more than 200% for intervals of 0.25–0.5 m.y. and rapid relative changes among the sections (Johnson et al., 1988) strongly suggest that only short-term subsidence changes that are coherently displayed by multiple sections, or changes in rates manifested over intervals sufficiently long to contain multiple control points (typically >1–2 m.y.), are likely to display a primarily tectonic signal. Rapid rate changes, especially when associated with short magnetic chrons, are most likely due to nontectonic causes. When such changes can be correlated consistently among numerous sections, they might represent a tectonic signal, but given the predicted diachrony of responses to loading in a transect across a foreland (Flemings and Jordan, 1990), synchronous short-term responses should be unlikely at time scales of 0.25 m.y.

Another approach to the use of foreland strata to reconstruct hinterland tectonism is to determine the difference between the time of deposition of individual detrital grains and the time that component minerals cooled below some critical isotherm within the crust. Assuming reliable mineral ages, the difference represents the time required for grains to be brought to the surface after cooling and then to be eroded, transported, and deposited in the foreland basin. The pioneering Himalayan study was undertaken by Cerveny et al. (1988) in the Potwar area using fission-track dating of detrital zircons. Whereas it was expected that their study would reveal rapid cooling and erosion associated with the rapidly subsiding foreland and thick Nagri sandstones at about 11 Ma, the data (Cerveny et al., 1988) indicate that throughout the 18-m.y. record, depositional ages and the youngest cooling ages differed by less than 3 m.y. Thus, rapid bedrock uplift and erosion must have been sustained at various sites within the Indus catchment throughout the Neogene. Based on ^{39}Ar/^{40}Ar dating, subsequent studies (Copeland and Harrison, 1990; Harrison et al., 1993) have demonstrated a similar coincidence of cooling and depositional ages, even at sites as remote from the Himalaya as the southern end of the Bengal Fan. Therefore, it appears that bedrock uplift, cooling, and denudation at rates (>3 mm/a) similar to those prevailing today at Nanga Parbat (Zeitler, 1985) have been present in the Himalaya throughout the presently studied Neogene record.

It remains to be seen at what time in the Paleogene slower rates of bedrock uplift and erosion can be detected. This probably will require better dating of pre-Siwalik strata than is presently available. It also remains to be seen if detailed studies of detrital ages at specific sites can be used to improve our understanding of the uplift/erosion history in the catchment. Not only does this await studies using many more dated grains, but also it will require an ability to reconstruct the evolving catchment that contributed to the study site – a problem that no one has yet solved in the Himalaya.

Impact of climatic changes

The strengthening of the monsoon at about 8 Ma (Prell and Kutzbach, 1991) would be expected to bring enhanced precipitation to the Himalaya and to increase the rates of mechanical erosion. Terrestrially based evidence, such as the carbon-isotope compositions of paleosols (Quade, Cerling, and Bowman, 1989), the timing of major faunal turnovers (Barry et al., 1985), and the inferred age of attainment of a maximum sustainable altitude for the Tibetan Plateau (Harrison et al., 1992), has previously been cited in support of climatic change at that time. Expansion of C4 grasslands across the Himalayan foreland has been attributed to lowered levels of atmospheric CO_2 starting at about 7 Ma (Cerling, Wang, and Quade, 1993) – a phenomenon that could have been caused by enhanced rates of chemical weathering (Raymo and Ruddiman, 1992). Recent studies, however, of the isotope compositions of tooth enamel from fossil herbivores in the Siwaliks indicate that C4 grasses were an important part of the diet at least as early as 9.4 Ma, such that their emergence cannot simply have been a response to a global atmospheric change after 8 Ma (Morgan, Kingston, and Marino, 1994).

Enhanced monsoonal precipitation would most strongly affect the southern slope of the Himalaya and would be expected to promote increased sediment fluxes out of the Himalaya and into the foreland. Contrary to that expectation, almost all of the rates of sediment accumulation across the foreland, from Pakistan to Nepal, indicate diminishing rates of accumulation at that time (Figure 9.7b). One possible explanation is that the increased sediment loads bypassed the foreland and were deposited in the deep-sea fans that flank the Indian subcontinent. Comparison of the accumulation data from the Bengal Fan, which is the largest repository of sediment derived from the Himalaya, with data from the foreland shows that accumulation rates decreased in both of these areas during late Miocene times (Burbank, Derry, and France-Lanord, 1993). There appears, therefore, to be little evidence for increased delivery of detrital sediment to the foreland at the time the monsoon is supposed to have been strengthening. It is improbable that enhanced chemical weathering could account for the missing detrital load. Most arguments regarding enhanced chemical weathering in the Himalaya have relied on rapid bedrock uplift and mechanical weathering to produce fresh mineral surfaces (e.g., Raymo and Ruddiman, 1992). The diminished delivery of Himalayan detrital sediment in the late Miocene does not support this model. Although the monsoon may have strengthened at about 8 Ma, there are no data that can clearly define how much it strengthened or the time span over which it achieved characteristics typical of the modern monsoon. In fact, in the absence of unambiguous supporting data from terrestrial sources, it is possible that the changes in the marine record from the Arabian Sea (Prell and Kutzbach, 1991) that have been attributed to the strengthening of the monsoon may have been caused by unrelated oceanographic processes.

The potential importance of climate in promoting bedrock uplift in mountains has recently been emphasized (Molnar and England, 1990). Two different end members for bedrock uplift can be delineated: In one, the tectonic shortening along basement-involved thrusts causes uplift of the earth's surface with respect to sea level and increases the magnitude of the tectonic load. In the second, enhanced erosion causes reduction of the load, lowers the mean height of the earth's surface, and induces isostatic uplift of the mountains. The cross-sectional geometry of the foreland and the patterns of fluvial deposition can be used to discriminate between these two end members (Burbank, 1992). Stratal successions that thicken toward the hinterland and short transverse rivers are indicative of subsidence that is driven by active thrust loading. On the other hand, tabular stratal successions that thicken away from the hinterland and long transverse rivers that push any axial river into the distal foreland are indicative of proximal basin uplift, potentially in response to erosional unloading in the hinterland.

During Miocene times in the Himalayan foreland, it appears that thrust loading dominated. There was an axial river that flowed across the medial and distal parts of the foreland. It flowed above a thick wedge of previously accumulated molasse that thickened toward the hinterland. During the later part of the Pliocene and throughout the Quaternary, it appears that transverse rivers have dominated the medial Gangetic foreland. Successions of upper Siwalik strata tend to become thicker toward the foreland (Acharyya and Ray, 1982; Raiverman et al., 1983; Burbank, 1992), the Ganges is presently at the far edge of the foreland, and there are essentially no older molasse strata beneath the present course of the Ganges. This geometry is consistent with uplift of the proximal foreland and with climatically induced unloading in the hinterland (Burbank, 1992). The present arrangement of the Gangetic foreland contrasts strongly with that of the Indus foreland. The Indus is in a medial-to-proximal position, flows above a thick succession of older molasse strata, and is fed by short transverse rivers. Apparently, sediment fluxes from transpressional ranges in western Pakistan are small relative to the magnitude of tectonically driven subsidence. As a consequence, the Indus lies close to the range front. We speculate that the absence of extensive glaciation in the transpressional ranges south of Kohistan has suppressed the transverse sediment flux and reduced the magnitude of erosional unloading.

At least two potential objections can be raised with respect to our contention that the Himalaya in the catchment of the Gangetic foreland have experienced a change from tectonically dominated loading to erosionally dominated unloading. First, the present, very distal position of the Ganges River may be only temporary. Perhaps large late-glacial to post-glacial sediment fluxes have pushed the Ganges toward the far side of the basin during the Holocene. In other words, outward migration of the Ganges could represent a nonequilibrium response to a short-term perturbation in sediment supply (Paola et al., 1992). As subsidence continues, if the transverse sediment supply were to diminish, perhaps the Ganges would migrate northward to the middle of the foreland. Second, the reason that young strata appear to taper toward the hinterland may be that tectonic deformation has propagated into the proximal foreland. In this case, active tectonism, rather than erosionally driven isostatic uplift, could cause displacement of the Ganges.

We can partially evaluate these objections through an examination of the geometry of the Plio–Pleistocene strata. If tectonic loading still dominates, then these strata should thicken toward the mountains, irrespective of short-term, nonequilibrium events within the record. Although only a few well-dated upper Siwalik sections in the Gangetic foreland have been reported (Karunakaran and Ranga Rao, 1979; Acharyya and Ray, 1982), these strata appear to thin toward the hinterland, providing support for the climatic-unloading hypothesis. Given the known Plio–Pleistocene deformation in the proximal foreland, however, tectonic uplift is also likely to cause thinning of the proximal strata. The density of data on sections of that age in India is not sufficient to determine whether or not the degree of taper is consistent with the expected flexural wavelength of the Gangetic foreland and, therefore, with isostatic uplift of the range. We need more appropriately dated sections before we can confidently choose among these possibilities.

Major unresolved problems

The largely unknown depositional history of the foreland between the Eocene and early Miocene represents one of the largest gaps in our present understanding of the sedimentary evolution of the Himalayan orogen. In northwestern India, sections that are continuous with and underlie the Siwalik Group are presently being studied and dated. These promise to give some insights into the late Oligocene and early Miocene depositional record. The 20-m.y. gap between the age of the biostratigraphically dated Murrees in the northern Jhelum re-entrant (Bossart and Ottiger, 1989) and the Burdigalian-aged (early Miocene) Murrees near to Rawalpindi needs to be spanned. The Oligocene record of deposition in Bangladesh may provide some insights, although at 30 Ma Bangladesh was situated at least 300 km farther to the south with respect to the axis of the Himalayan orogen.

Numerous radiometric studies of the cooling ages for detrital minerals that are now incorporated in dated sections in the foreland or the Bengal Fan have shown that the cooling ages nearly match the depositional ages and, consequently, that very rapid bedrock uplift and denudation have been prevalent in the Himalaya since about 18 Ma (Cerveny et al., 1988; Copeland and Harrison, 1990; Raymo, Green, and George, 1991; Harrison et al., 1993). These data have been cited as evidence for accelerated early Miocene Himalayan uplift (Harrison et al., 1992). In order to evaluate whether or not such rapid cooling and denudation began only in the early Miocene, similar studies of well-dated Oligocene- and Eocene-aged strata are needed. More may also be learned by detailed examination of the distribution of cooling ages through time as recorded at given sites. Interpretation of such data, however, relies heavily on the ability to reconstruct the catchment that was tributary to the depositional site. Given the complex geometry and largely unknown temporal evolution of Himalayan drainages (Seeber and Gornitz, 1983), definition of pre-Pleistocene catchments remains fraught with uncertainties.

It is clear that a great deal has been learned about the northwestern Himalayan foreland because of the high density of studied and dated sections there. Such extensive studies probably are not justified in all other areas of the foreland, but certainly there is a need for a broader distribution of well-dated sections in order to provide a more coherent and reliable reconstruction of potential differences and similarities along the length of the foreland. Seismic and borehole data have proved to be powerful tools in analyzing the structure of the northwestern Himalaya (e.g., Seeber et al., 1981; Pennock et al., 1989; Ni et al., 1991). Similar data from the Indian and Nepali forelands should provide needed information on subsurface structural geometries and should facilitate estimates of the amounts and sequences of shortening, in addition to clarifying the roles of basement structures in controlling basin geometries and localizing major faults.

Numerous petrographic studies of Siwalik and Murree strata have revealed changes in sandstone detrital modes through time and have defined the stratigraphic positions and relative abundances of dense minerals, metamorphic index minerals, and diagnostic lithic fragments (Johnson and Vondra, 1972; Chaudhiri, 1975; Raiverman et al., 1983; Critelli and Garzanti, 1994; Critelli and Ingersoll, 1994). Similar to the radiometric studies of detrital minerals, many of the petrographic studies suffer from the inability to identify specific source areas. Much would be gained by knowing that a particular lithology in a particular site within the orogen first became exposed to erosion at a given time. A more informative approach to the use of petrographic data to analyze events within the hinterland would be to combine these data with radiometric studies of the source area and analysis of detrital minerals attributed to it.

Numerous uncertainties persist with respect to the structural history exposed in the foreland. Most of the contractional structures within and bounding the foreland have been uncritically interpreted as responses to purely dip-slip shortening. Modern seismic data (Seeber et al., 1981) however, clearly indicate abundant strike-slip faulting within the northern Pakistani foreland. Active strike-slip faults cut modern fan surfaces in the Peshawar Basin (Burbank and Tahirkheli, 1985), and numerous high-angle reverse faults in the Kohat region have been interpreted as indicative of strong transpressional motion (Pivnik and Sercombe, 1993). Vertical-axis rotations of varying magnitudes are associated with most magnetic sections in Pakistan and suggest oblique shortening across nearby structures. Nonetheless, the relative amounts of strike-slip versus dip-slip shortening are poorly known for almost all faults. No coherent model has yet been proposed that could match the available magnetic data concerning the extent and timing of vertical-axis rotations.

In contrast to the prevalent interpretations of shallow thrusts that sole into an extensive detachment at the base of the Phanerozoic section in Pakistan, nearly all structural interpretations within the Indian foreland depict reverse faults that commonly steepen with depth and cut the basement at high angles (Acharyya and Ray, 1982; Raiverman et al., 1983). Such geometries are contrary to those interpreted for most other forelands. If such steep faults exist in abundance in the Indian foreland, the reasons for this unusual occurrence need to be explored. Alternatively, the geometry of faults and the role of detachments beneath the foreland need to be reexamined in the Gangetic foreland.

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