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Fluvial systems in the Siwalik Miocene and Wyoming Paleogene

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Abstract

The 3 km thick Miocene Siwalik Group (Himalayan foredeep in northern Pakistan) and the 2 km thick Paleogene Fort Union/Willwood formations (Bighorn Basin in Wyoming) both preserve long records of fluvial deposition adjacent to rising mountain belts. Depositional environments and associated habitats change across large basins along with changing physiography and with the location of different river systems that may have varied greatly in patterns of channel deposition and the drainage of adjacent floodplain areas. Deposits exposed in these two basins provide very different records of shifting paleoenvironments and patterns of basin filling. These differences reflect distinct patterns and scales of depositional environments, the nature of the exposures, and the types of sedimentologic studies that have been carried out in each basin.

The Siwalik Group fills a basin that extended at least 1000 km along its axis and 150–250 km away from the mountain front. Comparison of Siwalik deposits and modern drainages in the Himalayan foredeep suggests the ancient Siwalik basin was filled by large rivers that deposited low gradient sediment fans covering areas on the order of 1000 km^2, and by smaller interfan rivers with more poorly drained floodplains. Despite the scale of these river systems relative to Siwalik exposures in Pakistan, transitions between different systems have been recognized. Deposits of coeval river systems in the Siwalik Basin show pronounced differences in alluvial architecture, the character of overbank deposits, and the abundance and taphonomy of organic remains.

In contrast, the Bighorn Basin in Wyoming is a relatively small intermontane foreland basin extending 200 km along its axis and about 80 km across. Bighorn Basin strata were deposited by a river that flowed south to north along the basin axis and by smaller rivers that flowed transverse to the basin axis. Much of this basin is exposed and thus it is possible to reconstruct changing patterns of deposition and environments through time in more detail than in the Siwalik Basin. These patterns indicate changes in basin-wide drainage conditions and environments through time, but there are also important differences among coeval strata.

Upsection shifts in environments and vertebrate faunas within both the Siwalik and Bighorn Basins may reflect tectonic or climatic forcing, but this comparison emphasizes the importance of recognizing deposits from different contemporaneous river systems before inferring such large-scale controls on paleoecological change through time.

1. Introduction

The Neogene Siwalik Group of the Himalayan foredeep in northern Pakistan, and the Paleogene Fort Union/Willwood Formations of the Bighorn Basin in Wyoming (USA) both preserve long records of fluvial deposition adjacent to rising mountain belts (Fig. 1). These two stratigraphic successions are compared in this symposium volume because each contains abundant fossil remains, and each provides unusually continuous

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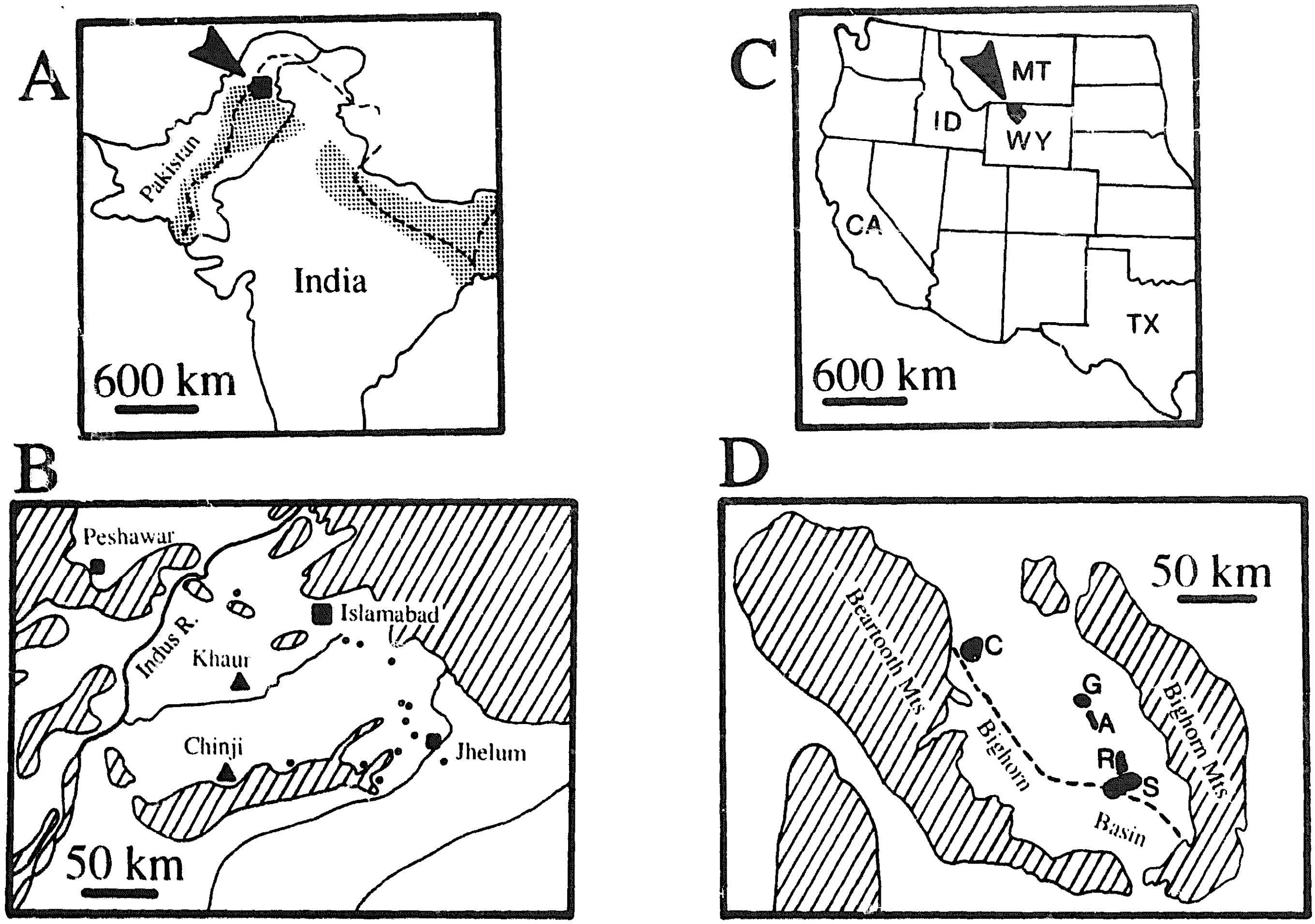

Fig. 1. Location maps of the Himalayan and Bighorn Basins. A. Map of India and Pakistan showing the modern Indus and Ganges Basins (stippled) and the location of the Potwar Plateau (arrowed box). B. Map of the Potwar Plateau showing locations of major cities (boxes) and field areas studied. Most paleontological and sedimentological studies have documented deposits near Chinji and Khaur Villages (triangles), but stratigraphic studies and more limited prospecting for fossils have also documented areas along the eastern Potwar Plateau (dots). C. Map of western USA showing location of Bighorn Basin (arrowed). D. Map of the Bighorn Basin showing location of the basin axis exposed in earliest Eocene deposits (dashed line), and locations of exposed strata mentioned in the text: C= Clark's Fork area; G= Gould Butte area; A= Antelope Creek area; R= Rairden area; and S= Sand Creek area.

and well-documented records of terrestrial mammalian evolution. Paleontologic records within each of these basins have been related to sedimentologic records of environmental change. This paper reviews these sedimentologic records.

Pronounced sedimentologic changes have been observed within both the Siwalik Group and the Fort Union/Willwood Formations over scales of kilometers to many tens of kilometers across each basin, and tens to hundreds of meters vertically through each stratigraphic succession. In some cases, these lithologic variations are associated with changes in biotic composition, taxonomic abundances, and the nature of organic preservation. Sedimentologic trends record basin-wide variations in paleoenvironmental conditions related to tectonically controlled rates of sediment supply and basin subsidence, and to regional climatic change. However, these variations also reflect more local drainage patterns and the positions of river systems that varied greatly in size, sediment load, channel pattern, migration and avulsion frequencies, and style of overbank deposition. Documentation of the scales at which these deposits vary is important to comparisons with their faunal record, because relatively local patterns of deposition and basin drainage may bias our view of basin-wide temporal changes over the long time intervals for which paleoecological change is reconstructed.

In this paper we compare large-scale sedimentological variations observed in strata of the Bighorn Basin of Wyoming and in strata of the Himalayan foredeep exposed on the Potwar Plateau of north Pakistan. For each area we describe the basin geometry, the definition of formations, the character of exposures, and the nature of sedimentologic studies completed. Sedimentologic records of environmental change are then described in terms of the patterns and scales over which depositional environments change, and interpretations of the processes that controlled sedimentologic variations within well-exposed parts of each stratigraphic succession. Finally we examine factors that bias interpretation of environmental change from the sedimentologic record. Despite similar overall tectonic settings, sedimentologic records exposed in these two basins are quite different. These differences reflect distinct distributions of paleoenvironments within each basin, but they also reflect dissimilar patterns of basin filling, the scale of depositional systems, and the nature of exposures. These large-scale differences influence comparisons of the environmental and biologic records contained within these two important successions.

2. Himalayan foredeep of northern Pakistan

2.1. The basin and its strata

The Himalayan Basin exposed in northern Pakistan formed due to the collision of India and Eurasia, which began at about 40–50 Ma and continues today (Patriat and Achache, 1984; Besse and Courtillot, 1988; and others: Figs. 1, 2). The Basin is structurally asymmetrical, with greater subsidence rates and thicker deposits near the mountain belt (i.e., towards the north; Barry et al., 1980; Johnson et al., 1982, 1986; Johnson et al., 1985; Burbank and Beck, 1989; Beck and Burbank, 1990). In northern Pakistan the orogenic belt has a sharp inflection, from the NW–SE orientated Himalayan Range to the NE–SW orientated zone of deformation that generally parallels the Chaman transform zone (i.e., essentially parallel to the Pakistan–Afghanistan border). The Ganges Basin is a NW–SE depression that parallels the trend of the Himalayan range, and extends nearly 2000 km from the Punjab of India to the Brahmaputra River in Bangladesh, and finally to the Bay of Bengal. The Indus Basin parallels the NE–SW trend of the Chaman transform zone for over 1000 km from the Punjab of Pakistan to the Arabian Sea. Each of these basins is 200–300 km across and reflects structural downwarping due to crustal loading in an adjacent mountain belt.

The Siwalik Group is 3–4 km thick and records fluvial deposition within the Himalayan foredeep spanning the middle Miocene through the Pliocene, about 18–2 Ma (Fig. 3; Barry et al., 1980; Johnson et al., 1982, 1986; Johnson et al., 1985). Formations within the Siwalik Group are defined by pronounced upsection variations in the proportion of channel sandstone bodies relative to mudstone-dominated overbank deposits (Fatmi, 1973; Shaw, 1977; Pilbeam et al., 1979), and by the color of channel sandstones and characteristics of paleosols. The Siwalik Group is divided into the Lower, Middle and Upper subgroups, each subgroup composed of a sandstone-rich formation and an overlying relatively sandstone-poor formation. The Siwalik Group outcrops in locations scattered along a thin strip uplifted at the southern foot of the Himalayan mountain belt across most of northern Pakistan and India (Fig. 2), but the best exposures occur on the Potwar Plateau in northern Pakistan. Here uplift exposed Siwalik strata along a number of broad synclines and anticlines over an area that extends nearly 100 km perpendicular to the mountain belt (Fig. 2; Yeats et al., 1984). Although stratigraphic dips can vary from horizontal to vertical across these structural folds, strata in areas extending tens of kilometers along strike can be otherwise undeformed. Steep badlands valleys cut approximately perpendicular to structural strike allow the thickness of sandstone- and mudstone-dominated intervals and the magnetic polarity of sediments to be documented through continuously exposed, kilometers-thick stratigraphic successions (Fig. 1; see also summaries in Johnson et al., 1982, 1986; Burbank et al., 1986), and areas of extensive badlands

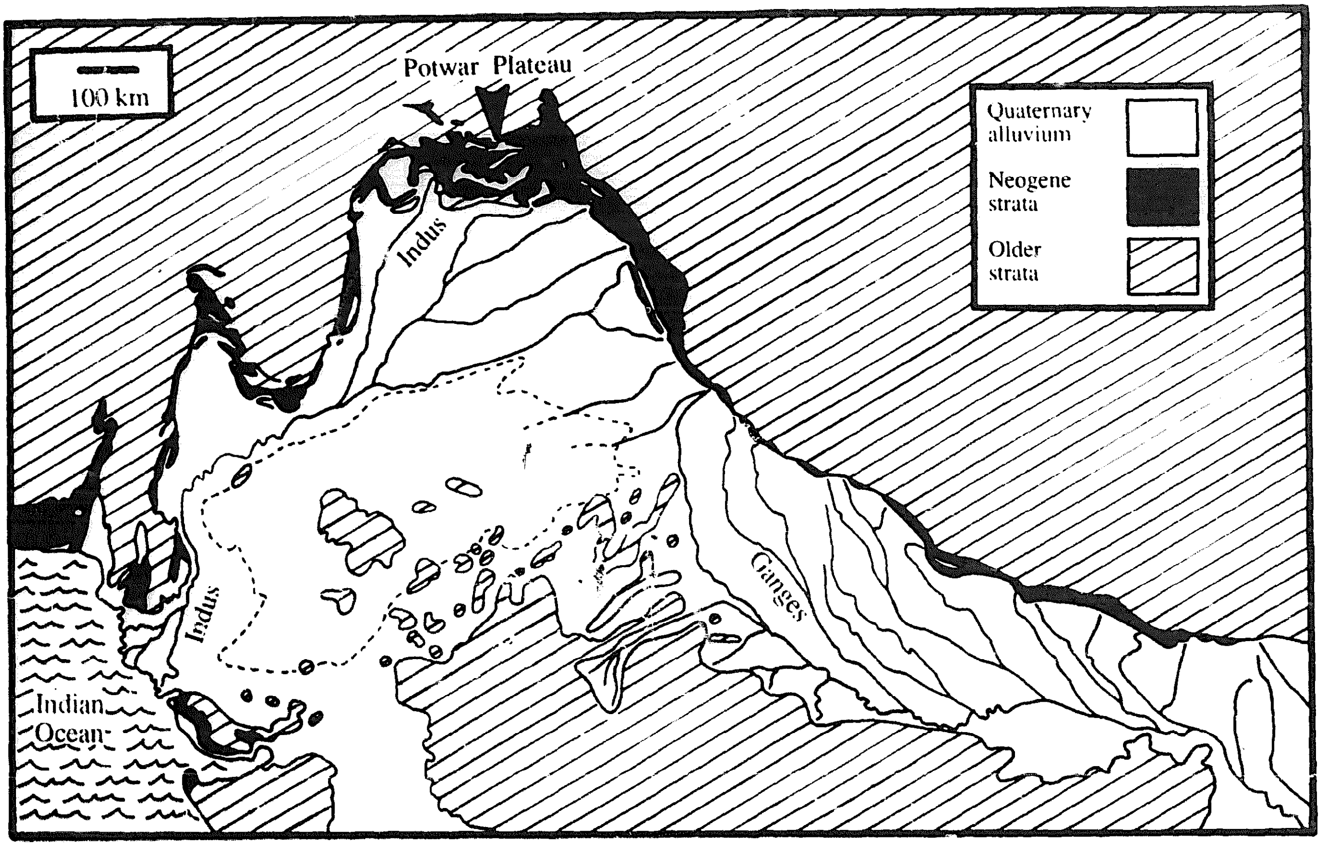

Fig. 2. Map of the Indus and Ganges Basins showing modern drainages (solid lines), areas of exposed Neogene strata including the Siwalik Group (black), areas of exposed strata older than Neogene (diagonal lines), areas of exposed strata younger than Neogene (unshaded), and the location of the Potwar Plateau (arrow). The boundaries of the Thar desert, an area with no through flowing drainage in southern Pakistan, is marked by a dashed line. Modified from Willis (1993b).

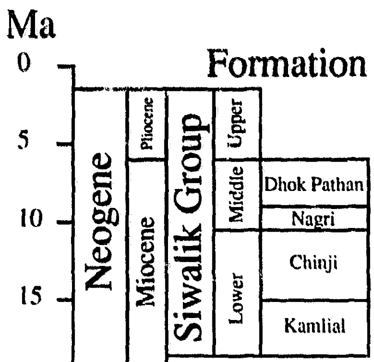

Fig. 3. Age of Siwalik Group Formations exposed along type sections of the Chinji and Nagri formations in Chinji Village area of northern Pakistan. Age of formation boundaries based in the paleomagnetic data of Johnson et al. (1985) and the paleomagnetic time scale of Berggren et al. (1985).

exposure allow strata to be examined for tens of kilometers parallel to strike (e.g., Badgley and Behrensmeyer, 1980; Behrensmeyer and Tauxe, 1982; Behrensmeyer, 1987; Willis, 1993a; Willis and Behrensmeyer, 1994). Structural folds and broad areas covered by Quaternary sediments do not allow stratigraphic variations to be mapped in detail across the Potwar Plateau. Instead, correlations of vertical sections are based on magnetostratigraphic studies, and to a lesser extent on biostratigraphic and lithostratigraphic data. Upsection lithologic trends and dated stratigraphic transitions vary greatly in different areas, but formations of the Siwalik Group are generally assumed to be continuous over the many tens to hundreds of kilometers that separate areas where long stratigraphic sections are exposed. In many cases, however, it is difficult to define formation boundaries independently of chronostratigraphic and biostratigraphic correlations, particularly for strata exposed further away from type sections (Johnson et al., 1982; Khan, 1993). Therefore, large-scale patterns of deposition and environments across the ancient Siwalik Basin may have been more complex than that suggested by the current lithostratigraphy of the Siwalik Group in Pakistan (e.g., Gill, 1951; Fatmi, 1973; Barry et al., 1980; Johnson et al., 1982, 1986).

Vertical sections through Siwalik Group strata exposed on the Potwar Plateau have produced fairly detailed records of sediment progradation due to tectonism (Johnson et al., 1985; Johnson et al., 1986; Burbank et al., 1986, 1988. Willis, 1993b). However, this record has not yet been extended across the entire Siwalik Basin. Siwalik strata exposed on the Kohat Plateau to the west are currently being examined (Beck and Burbank, 1990); however it is not yet clear how variations in these strata are related to upsection sedimentologic variations documented along type sections on the Potwar Plateau. Similarly, Siwalik Group strata in India are difficult to compare in detail with strata exposed on the Potwar Plateau because upsection lithologic variations have not been described in detail, age dating is generally lacking, and formations in India are defined by vertical variations in heavy minerals (Parkash et al., 1980) and not necessarily by pronounced changes in the proportion of sandstone bodies as in Pakistan.

Sedimentological studies on the Potwar Plateau have generally progressed in tandem with fossil prospecting, and thus strata examined in detail sedimentologicly also typically contain the best known paleontologic records. The greatest concentration of work has focused on two areas near the villages of Khaur and Chinji, areas separated by about 40 km north–south (Fig. 1B). In these areas studies of sedimentologic variations over E–W distances of tens of kilometers have documented lithofacies distributions, sediment body geometries, paleocurrent orientations, traces and characteristics of paleosols, and the taphonomy of fossil concentrations along specific intervals at different levels within the stratigraphic succession (e.g., Badgley and Behrensmeyer, 1980; Behrensmeyer and Tauxe, 1982; Behrensmeyer, 1987; Willis, 1993a; Willis and Behrensmeyer, 1994). These more detailed records of sedimentologic variation span a very small area relative to the scale of the modern and ancient basins adjacent to the Himalayan mountains, and thus represent a relatively local record of environmental change. An improved understanding of basin-scale environmental change will require detailed sedimentologic studies in areas that span a greater region of the basin, and a more critical assessment of the continuity and correlation of lithostratigraphic units between widely spaced stratigraphic sections.

2.2. Depositional systems

Although basin-scale patterns of deposition across the ancient Siwalik Basin have yet to be documented in detail, patterns of deposition within the modern Himalayan basin provide a close analog (see also Willis, 1993b). Deposition within the modern Himalayan foredeep is controlled by large river systems that emerge from the Himalayan mountains at intervals of about 100–200 km (Fig. 2). These rivers have well-defined antecedent channels through the adjacent mountain belt such that their entrance positions into the basin have been fixed over millions of years (Parkash et al., 1980). Where such rivers enter the basin, they deposit large low-gradient sediment fans that extend up to 200 km into the basin and expand to widths up to 100–200 km (Geddes, 1960; Gole and Chitale, 1966; Wells and Dorr, 1987a,b; Parkash et al., 1980, 1991; Friend, 1989). Sediment grainsize, floodplain gradients, depositional rates, and the braiding indices of channels decrease away from fan apices (Geddes, 1960; Wells and Dorr, 1987a,b). Between sediment fans deposited by these major rivers, smaller rivers of varying size drain more local foothill areas of the mountain belt and local areas within the basin (i.e., rivers smaller in scale than those shown in Fig. 2; see Geddes, 1960). These smaller rivers carry finer sediment loads, traverse relatively low gradient and poorly drained floodplains, and have generally lower depositional rates (Geddes, 1960; Parkash et al., 1983; Parkash, 1991).

Large rivers entering the basin from the mountain belt initially flow south, transverse to the structural axis of the basin. In northern Pakistan these rivers gradually converge across the Punjab region into the Indus River, which then flows southward across a 100–150 km wide floodplain toward the Arabian Sea. In northern India these rivers gradually turn eastward and join the Ganges, which flows along a relatively narrow (tens of kilometers wide) floodplain parallel to the mountain belt and adjacent to the far southern edge of the basin. By analogy with these modern drainage systems, the proportion of channel sandstones and the character of overbank sequences in older deposits of the Himalayan Basin are expected to vary both parallel to and away from the mountain belt near major alluvial fans. Also, the dimensions of channel bodies should be distinct for deposits of rivers flowing transverse to the basin axis and those of the larger basin-axial rivers.

Siwalik strata in northern Pakistan are exposed for about 150 km along the basin and 50 km across it, an area similar in scale to that occupied by a sediment fan deposited by an individual major river entering the modern basin. Despite the relatively ·all area documented by sedimentologic studies, different river systems have been recognized within Siwalik strata of the Potwar Plateau. Two examples are summarized below.

The boundary between the Chinji and Nagri Formations marks a pronounced coarsening within the Siwalik alluvial succession that has been recognized throughout northern Pakistan and India (Johnson et al., 1982; Sahni and Mitra, 1980; Parkash et al., 1980) and by coeval increases in sediment accumulation rates within the Indian Ocean (Amano and Taira, 1992). This stratigraphic transition has been documented in detail within the Chinji Village area of the Potwar Plateau (Figs. 1, 4; Johnson et al., 1985; McRea, 1990; Willis, 1993a,b). Both formations consist of sandstone bodies that are tens of meters thick and are interpreted as deposits of large rivers, separated by mudstone dominated intervals interpreted as overbank deposits. Details of sediment variations within these major sandstone bodies and within mudstone-dominated sequences have been presented previously (Raza, 1983; Behrensmeyer, 1987; McRae, 1990; Willis, 1993a,b; Willis and Behrensmeyer, 1994), so they are discussed only briefly here.

Major channel sandstone bodies are continuous along strike for at least tens of kilometers in outcrops oriented somewhat oblique to the mean paleoflow direction (Fig. 5). Deposits within major sandstones can be divided into different storeys, each of which record deposition within an individual migrating channel segment. Individual storeys are generally 5–10 m thick within the Chinji Formation, but can be up to 30 m thick within the Nagri Formation (Fig. 5). Storeys are defined by a set of inclined beds, each bed representing depos-

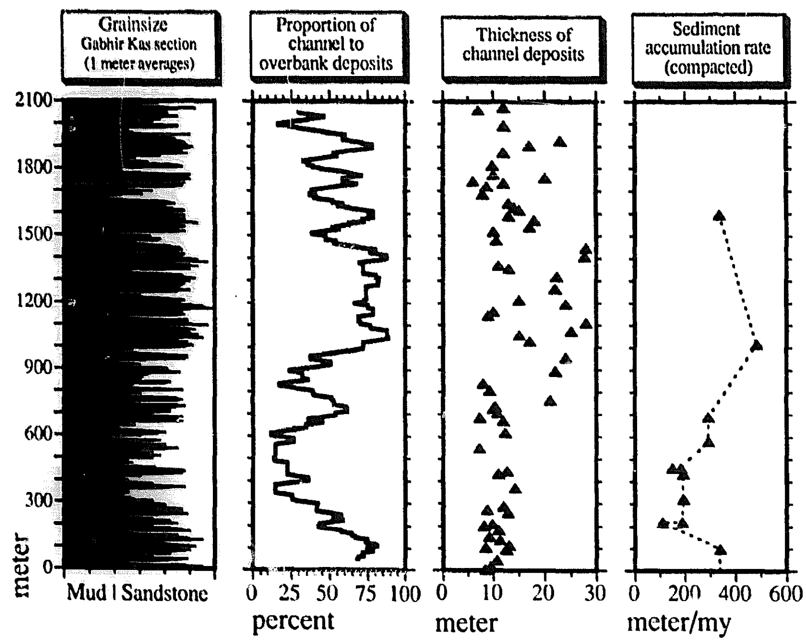

Fig. 4. Vertical deposit variations measured along Gabhir Kas near Chinji Village (modified from Willis, 1993b).

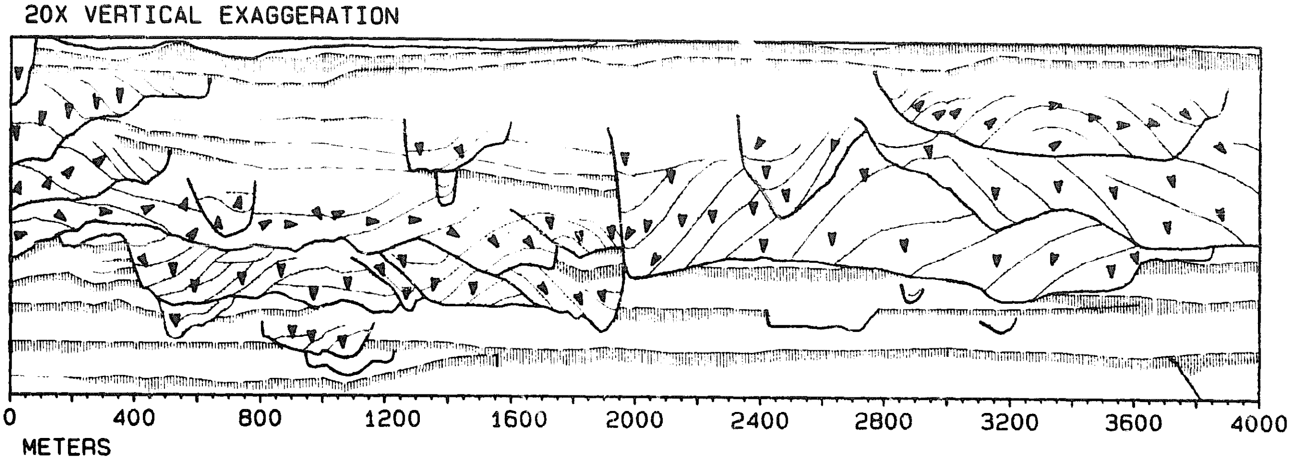

Fig. 5. Bedding diagram showing a major sandstone body (stippled) and vertically adjacent overbank deposits (unstippled) at the Chinji-Nagri Formation boundary in the Chinji Village area. Traces of carbonate-leached horizons that mark the upper parts of paleosols are shown by vertical line shaded layers in overbank deposits. Storey boundaries, trends in bedding dip within storeys and paleocurrent orientations relative to the outcrop plane (up is into the outcrop) are shown within the major sandstone body. Note smaller size of storeys exposed to the left relative to those exposed to the right. These storeys reflect the deposits of rivers of different scale superimposed along the same stratigraphic level. See detailed discussion in Willis (1993a).

ition during an individual flood and channel migration event. Bounding surfaces of inclined beds within storeys record cross-sectional profiles of the ancient channel, and together the set of inclined beds comprising a storey provides a record of changing paleochannel geometry and migration as viewed within the outcrop plane (see detailed discussion in Willis, 1993a).

Major sandstone bodies contain a complex stacking of storeys, both vertically and along strike. Storeys within sandstone bodies can be grouped into different channel belts (i.e., groups of channel and bar deposits separated by avulsion events) by their alignment along specific horizons and the local preservation of overbank mudstone beds and paleosols within sandstone bodies. Within the Chinji Formation, individual channel belt deposits are laterally continuous for only 1–2 km in exposures oriented perpendicular to paleoflow and contain several (4–8) storeys stacked along strike. Within the Nagri Formation such channel belt deposits extend for at least 5 km perpendicular to the paleoflow. Generally braided channel patterns within channel belts are indicated by: (1) the relatively large number of storeys within individual channel belt deposits exposed perpendicular to paleoflow; (2) the limited lateral extent of individual storeys exposed perpendicular to paleoflow relative to estimates of bar wavelength; (3) the relatively low proportion of lateral-accretion to channel-filling beds; (4) the dominance of coarse-grained channel fills; (5) the low paleocurrent variation within individual storeys (commonly less than 45°); (6) the abundant evidence for channel-bar superposition due to channel switching; and 7) local unambiguous evidence for mid-channel bars (Willis, 1993a).

Channel belt deposits are not randomly distributed within formations, as is predicted by alluvial stratigraphy models (e.g., Bridge and Leeder, 1979), but instead they tend to be clustered vertically on a scale of a 100 m into intervals of connected channel belt deposits separated by intervals that are dominated by overbank deposits with a few unconnected channel belt deposits (Fig. 4). Clusters of channel belts continue along strike for the tens of kilometers of exposure in the Chinji Village area (Raza, 1983), and reflect short term (0.3–0.5 m.y.) variations in net sediment aggradation rates or temporary restriction of avulsing channels to particular areas on the floodplain (Willis, 1993b).

Paleochannel reconstructions from storeys exposed within the Chinji Formation (Willis, 1993a) indicate that individual channel-segments typically had widths of 80–200 m, depths of 4–13 m, and discharges of 400–800 m^3/s. Full channel belt widths of 1–2 km and braiding indices of 2–3, estimated from exposures perpendicular to paleoflow, indicate a full-river discharge of 1500–2000 m^3/s. Larger channel segments reconstructed from Nagri Formation deposits had widths of 200–400 m, depths of 15–30 m, and discharges of 3000–5000 m^3/s. As Nagri Formation channel systems were also clearly braided, and channel belts were at least 5 km wide in exposures perpendicular to paleoflow, full river discharges were probably at least a factor of two greater than for individual channel segments (i.e., on the order of 10,000 m^3/s).

Mudstone dominated deposits between major sandstone bodies are composed of: minor channel-form sandstone bodies recording deposition in crevasse, tributary, and floodplain drainage channels; lobate and wedge-shaped sandstone or mudstone dominated-bodies recording deposition on crevasse splays and levees; thin laminated claystone beds recording deposition in floodplain lakes; and paleosols recording depositional hiatus and the disruption of previously deposited sediments. Most overbank sediments comprise splay/levee deposits, whereas minor channel deposits constitute only 14% and 12% of overbank deposits within the Chinji and Nagri Formations, respectively.

Across the Chinji-Nagri Formation boundary: (1) the ratio of major channel sandstone bodies relative to overbank deposits increases; (2) the average thickness of individual storeys increase; (3) mean sediment accumulation rates increase; and (4) there is a decrease in abundance of preserved faunal remains and changes in faunal composition (Fig. 4; Johnson et al., 1985; Barry et al., 1985; Willis, 1993b). Higher proportions of channel deposits in the Nagri Formation reflect the larger size of the channels and the more rapid return interval of channel belts to a given area on the floodplain (i.e., mean recurrence interval of 200,000 and 60,000 yr for the Chinji and Nagri formations, respectively).

Although distinct paleosols disrupt a greater proportion of the Chinji Formation overbank deposits (about 40%) relative to the Nagri Formation (20–30%), overbank sequences of the Chinji Formation contain a greater proportion of deposits with preserved primary stratification. This reflects a pronounced cyclicity over 4–10 m vertically within Chinji Formation overbank successions defined by the alternation of deposits with preserved primary stratification and completely disrupted deposits in paleosols (Willis, 1993a; Willis and Behrensmeyer, 1994). Such cyclicity is less pronounced in the Nagri Formation, where primary stratification in overbank deposits is more completely disrupted throughout. The Chinji Formation also contains a greater amount of lacustrine claystones (6.2% and 1.9% of overbank deposits, respectively) and paleosols with iron concretions (80% and 30% of paleosols, respectively). Minor floodplain channel bodies in the Chinji Formation commonly contain relatively undisrupted beds alternating between sandstone and mudstone, have laterally continuous sets of inclined beds relative to widths of channel fills, some contain oncolites along basal surfaces, and many contain concentrations of vertebrate fossil remains. These features suggest channels were long lived and contained some flow throughout the year. However, pedogenically disrupted horizons preserved in these bodies near their terminating margins indicate periods of subaerial exposure and more gradual filling during the final stages of channel abandonment (see Behrensmeyer et al., 1994). In contrast, typical minor channel deposits of the Nagri Formation are completely sandstone dominated and bioturbation disrupts all primary stratification, suggesting more seasonal flow in these channels and perhaps that deposition occurred only in association with major floods. All these features of overbank deposits suggest that Chinji floodplains were more poorly drained, at least seasonally, and local overbank deposition rates were more episodic (Willis, 1993a; Behrensmeyer et al., 1994).

Taken together, along-strike variations within individual sandstone bodies and upsection variations across the Chinji and Nagri formations suggest two river systems that differed in scale and associated overbank environments. Only the smaller river system is recorded in deposits of the Chinji Formation. This river was comparable in scale to relatively small rivers within the modern Indus and Ganges basins; such rivers drain local areas within the basin or the adjacent orogenic belt and have relatively poorly drained floodplains. Deposits of a larger river system dominate sandstone bodies within the Nagri Formation. This river was comparable in scale to larger rivers of the Himalayan foredeep that drain substantial areas within the orogenic belt, flow southward into the basin, and deposit large, low gradient sediment fans (i.e., upper Indus, Jhelum, Kosi, and other basin transverse rivers shown in Fig. 2). Very large river systems comparable to axial rivers within the modern Himalayan basin (i.e., lower Indus and lower Ganges Rivers; Fig. 2) are not recognized within these deposits.

A second example of variations within the Siwalik Group has been documented at the transition of the Nagri Formation to the overlying Dhok Pathan Formation, in the area of Khaur Village (Fig. 1B) This is the "U-sandstone" stratigraphic interval described by Badgley and Behrensmeyer (1980), Barry et al. (1980), and Behrensmeyer and Tauxe (1982). Here contemporaneous deposits were examined for 30 km along strike within a 40 m thick interval. Deposits within this interval change dramatically from east to west (Fig. 6). To the west, sediments are dominated by thick connected channel-belt deposits. These deposits are similar to those that occur within the Nagri Formation in the Chinji Village area (described above), and they have been similarly interpreted as deposits of large braided rivers. The margins of these thick channel deposits are associated with distinct overbank mudstones that gradually thin to the east. Deposits toward the eastern end of the interval examined are dominated by mudstone beds with smaller-scale channel deposits that are generally 2–5 m thick and continue along strike for less than a few kilometers. These smaller-scale channel deposits show a wider range of paleocurrent orientations relative to those within the larger-scale channel deposits to the west. On average, paleocurrent azimuths within these different deposits differ by about 20° (107° for the larger-scale deposits versus 126° for the smaller; Behrensmeyer and Tauxe, 1982). These smaller-scale channel deposits were interpreted as resulting from relatively minor, sinuous rivers that were tributaries to the larger braided river further down basin.

Fossils are abundant along this interval, and faunal composition and taphonomic features of fossil concentrations have been documented in detail (Badgley and Behrensmeyer, 1980; Barry et al., 1980; Badgley, 1986; Badgley et al., 1994). Of particular importance here is the scarcity of fossil localities within the large channel deposits to the west and within their associated overbank deposits. Thus there is a dramatic along-strike change in the abundance of preserved fossil remains, reflecting differences between deposits associated with the two distinct river systems.

The vertical change in strata across the Chinji-Nagri Formation boundary documented within the Chinji Village area is similar to changes documented along strike within the Khaur Village area. Such large-scale changes in Siwalik deposits and paleoenvironments (e.g., changes used to define formations) have normally been ascribed to periods of tectonism that greatly changed basin subsidence rates, sediment deposition rates, and the coarseness of the basin fill (e.g., Parkash et al., 1980; Johnson et al., 1985; Johnson et al., 1986). Our studies, however, emphasize the importance of recognizing the juxtaposition of deposits from different but contemporaneous river systems before inferring such large-scale controls on paleoenvironmental change through time. Because different river systems are being compared in long vertical sections measured on the Potwar Plateau, it is not yet clear to what extent stratigraphic variations reflect spatial changes across the basin versus basin-scale environmental changes through time.

2.3. Bighorn Basin

The basin and its strata

The Bighorn Basin in Wyoming is a N–S trending, Laramide intermontane basin, 200 km long and about 80 km wide (Fig. 1C). This basin forms the southern part of a 500 km NW–SE trending structural trough that includes the Clark's Fork, Stillwater, and Crazy Mountains basins progressively to the north (Fig. 7; Gingerich, 1989a). The Bighorn Basin is structurally asymmetrical, with greater subsidence rates and thicker deposits

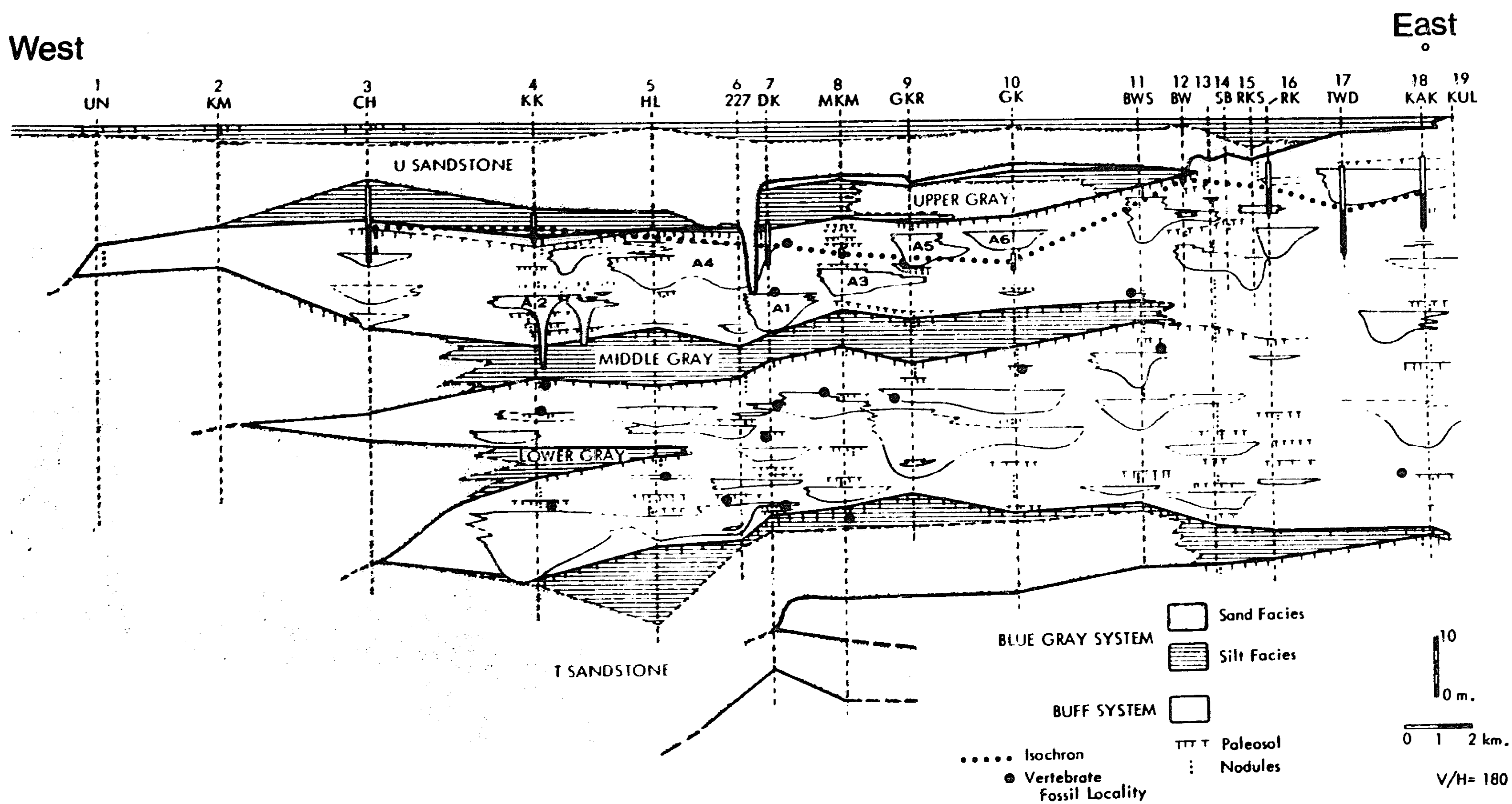

Fig. 6. Diagram showing lithologic variations and fossil sites (dots) in the Khaur area (from Behrensmeyer and Tauxe, 1982). Thick channel sandstones dominate to the west (left) whereas mudstone sequences with minor channel sandstones occur to the east (right). The interconnected channel sandstone bodies to the left pass laterally into distinctive mudstone sequences. These lithologic changes over tens of kilometers along strike are interpreted to reflect the intertonguing of deposits from two coeval river systems within the ancient Siwalik Basin.

towards the western margin (Bown, 1975, 1980; Gingerich, 1983).

Late Paleocene to Early Eocene Laramide uplift of the Beartooth Mountains, early Eocene uplift of the Owl Creek mountains, and latest Early Eocene development of the volcanic Absaroka Range formed structural highs along the western margin of the Bighorn Basin (Bown, 1975, 1980, 1982). Along the southwest margin of the basin a more subdued area of uplift connecting the Beartooth and Owl Creek Mountains (the Cody Arch) was episodically uplifted from Late Cretaceous through Middle Eocene time (Sundell, 1990). The Oregon Basin Fault (exposed at the surface along the Tertiary structural axis of the basin: Stone, 1985; Parker and James, 1986) parallels the Cody Arch and cuts the Fort Union but not the Willwood Formation, suggesting fault motion ceased by the end of the Paleocene (Kraus and Bown, 1993). The Tertiary structural axis of the basin is overridden by the Beartooth Mountains and there structural asymmetry of the basin is more pronounced, reflecting greater thrust loading along the north west margin of the basin (Bown, 1980; Bonini and Kinard, 1983; Gingerich, 1983). To the east and northeast the basin is bounded by the Bighorn and Pryor Mountains, which were also uplifted during eastward directed Laramide thrusting.

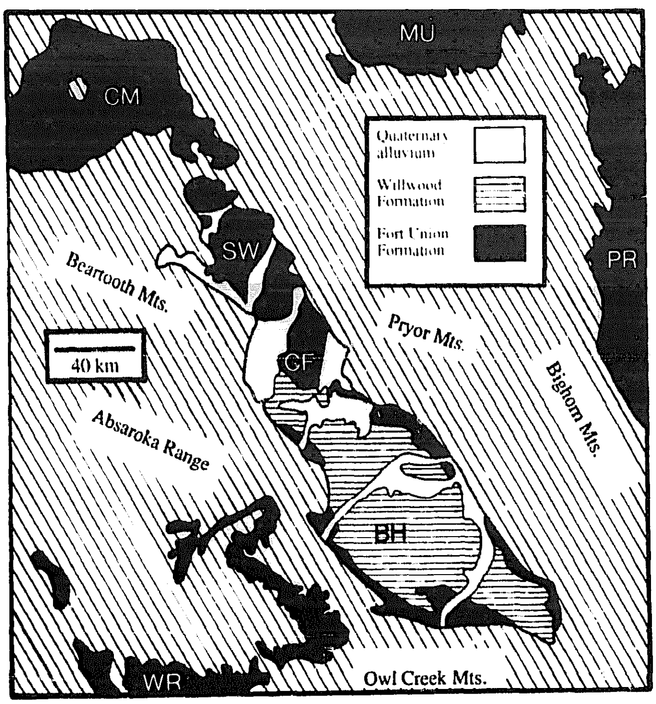

Fig. 7. Map of the Bighorn (*BH*), Clark's Fork (*CF*), Stillwater (*SW*), and Crazy Mountains (*CM*) Basins showing distribution of the Willwood Formation (horizontal lines), the Fort Union Formation (black), Recent alluvial sediments (unshaded), and older strata including mountainous areas (diagonal lines). Adjacent Laramide basins include the Musselshell basin (*MU*) the Power River Basin (*PR*) and the Wind River Basin (*WR*). Modified from Gingerich (1983).

The Fort Union Formation (600–3000 m thick) and the Willwood Formation (up to 850 m thick) record fluvial, palustrine and lacustrine deposition spanning the Paleocene and early Eocene, about 65–50 Ma (Fig. 8; Van Houten, 1944; Neasham and Vondra, 1972; Bown, 1980; Hickey, 1980; Gingerich, 1983, 1989a; Kraus and Middleton, 1987). The boundary between the Fort Union and Willwood formations is distinguished by a change from drab sediments with abundant lignite beds lower in the section (Fort Union Fm.) to beds with abundant bright variegated paleosols higher in the section (Willwood Fm.). Both formations are dominated by mudstone beds; however, there are pronounced lateral variations in lithofacies, the geometry of sediment bodies and the characteristics of paleosols across the basin within each formation (e.g., Bown, 1980; Hickey, 1980; Wing and Bown, 1985; Kraus, 1992).

Strata in the Bighorn Basin form a broad syncline, but are nearly flat-lying away from the margins of the basin. Exposures vary from deeply weathered, low-lying badlands, to badlands ridges

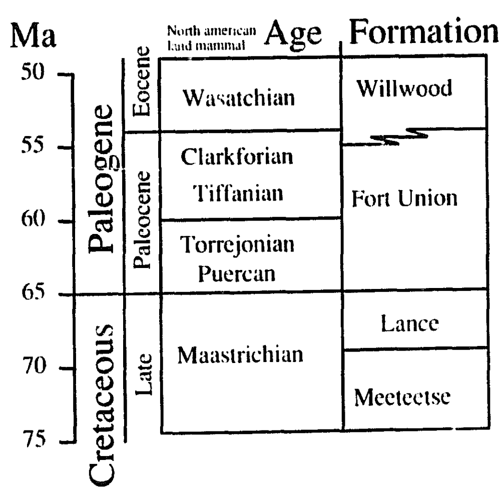

Fig. 8. Age of Bighorn Basin Formations.

and canyons that can be complexly dissected by ephemeral drainages. Chronostratigraphy is defined mainly by biostratigraphic zonation keyed to rare dated ash layers (Gingerich, 1983; Wing et al., 1991). The Fort Union Formation is better exposed along the edges of the basin, whereas exposures of the Willwood Formation dominate the central part of the basin (Fig. 7; Bown, 1979; Gingerich, 1989a; Wing et al., 1991). Biostratigraphic zones are elongated along the N-S axis of the basin, and in the Willwood Formation older deposits generally are better exposed on the east side of the basin, and progressively younger strata are better exposed westward towards the structural axis of the basin (Gingerich, 1989a; Wing et al., 1991).

In contrast to the Siwaliks of northern Pakistan, there are extensive exposures throughout the Bighorn Basin allowing sedimentologic and pedogenic features within formations to be mapped over large areas. However, because the strata are nearly horizontal across most of the basin, continuous sections in any one location typically extend vertically for only a few hundred meters. Records of long term change thus are not keyed to continuous vertical sections as in the Siwaliks, but to basinal traverses or more commonly to a comparison of deposits documented within successive biostratigraphic zones.

Deposits in most areas of the basin have been described in terms of: (1) qualitative estimates of the proportion of channel sandstones relative to overbank deposits, the shapes of channel sandstone bodies (sheet versus ribbon), and mean paleo-flow orientations; (2) the occurrence and lateral extent of lignites and carboniferous shales; and (3) the character of the paleosols. These observations provide a record of environmental change across the Bighorn Basin; however, detailed sedimentologic logs have generally not been published. The most continuous records of temporal change in the basin are documented by a series of parallel traverses through the Fort Union Formation in the northern part of the basin (Fig. 9; Hickey, 1980) and by a long traverse through deposits of the Willwood Formation in the southern part of the basin (Fig. 10; Wing et al., 1991; Bown and Kraus, 1993; Kraus and Bown, 1993).

More detailed sedimentologic studies in specific areas of the Bighorn Basin describe local channel-proximal to -distal gradients in sediments, paleosols, and fossil remains over kilometers (e.g., Bown and Kraus, 1987; Bown and Beard, 1990; Kraus and Aslan, 1993), estimate typical dimensions of channel deposits within sandstone bodies (Kraus, 1980; Kraus and Middleton, 1987; Kraus and Aslan, 1993), and address the lithology and geometry of beds containing well preserved plant fossils (Wing, 1984). Neasham and Vondra (1972) mapped paleoflow directions across the basin, but did not resolve the age of deposits beyond confining measurements to the Willwood Formation. DeCelles et al. (1991) documented the architecture of coarse-grained alluvial fan deposits preserved along the western margin of the northern Bighorn Basin (i.e., Clark's Fork sub-basin).

Additional sedimentological studies of channel dimensions, internal bedding geometry, and paleocurrent orientations within sandstone bodies, both across coeval strata and between different stratigraphic intervals, are needed before details of changing drainage patterns, boundaries of sediments deposited by different river systems, and directions of sediment dispersal within the basin can be delineated.

2.4. *Depositional systems*

The overall depositional history of the Bighorn Basin is fairly well known (see summaries in Bown, 1980; Gingerich, 1983, 1989a; Beck et al., 1988), and a brief account focused on the Fort Union and Willwood Formations is given below. The Bighorn Basin began to form in the late Cretaceous with the onset of Laramide deformation and withdrawal of the Lewis Sea. The Meeteetse Formation records sediments deposited in deltaic, near-shore fluvial, and coal swamp environments associated with retreat of the Lewis Sea from the basin. Fluvial deposits of the Bighorn Basin were sand-rich initially and include the thick, interconnected channel sandstone bodies of the latest Cretaceous Lance Formation (Bown, 1975, 1980). In the early Paleocene (lower Fort Union Fm.), less sand-rich alluvial deposits dominated the central part of the basin, and lacustrine and coal swamp environ-

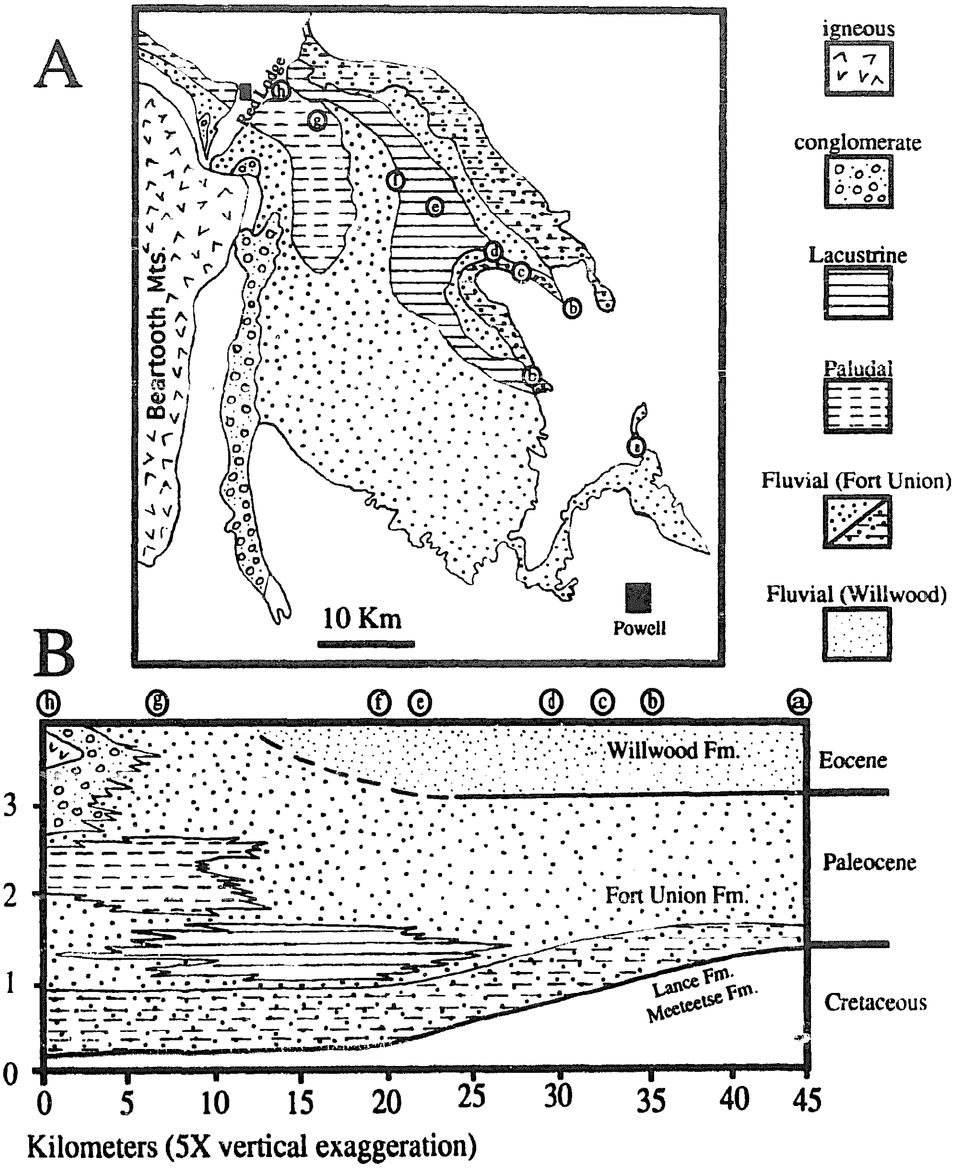

Fig. 9. Outcrop map (A) and cross-section (B) modified from Hickey (1980) showing lithologic units that comprise the Fort Union and lower Willwood formations (Clark's Fork sub-basin; see location *C* in Fig. 1D). The position of measured logs relative to the cross-sectional line of strike is shown on the map by letters *a–h*. Upsection lithologic units depicted in cross-section reflect up section basin traverses orthogonal to the strike of the cross-section. The lower and upper Paleogene fluvial deposits, and the lacustrine, paludal, and conglomerate units to the western side of the basin, are each recognized by Hickey (1980) to be members of the Fort Union Formation.

ments prevailed in more northerly and southerly areas (Rea and Barlow, 1975; Bown, 1980; Hickey, 1980). Channel sandstones commonly had sheet-like geometries (i.e., thickness to lateral extent greater than 1:20; Bown, 1979; Kraus, 1980). Sediment accumulation rates were relatively low (about 20 m/m.y.; Gingerich, 1983). The major basin drainage throughout the Paleogene probably flowed northward along the more rapidly subsiding western margin of the basin.

In the northern part of the basin, sediment accumulation rates increased progressively through the middle and late Paleocene (from 130 to 280 m/m.y.) and the basin became more asymmetrical, indicating a period of intensified tectonism and sediment supply (Bown, 1980; Gingerich,

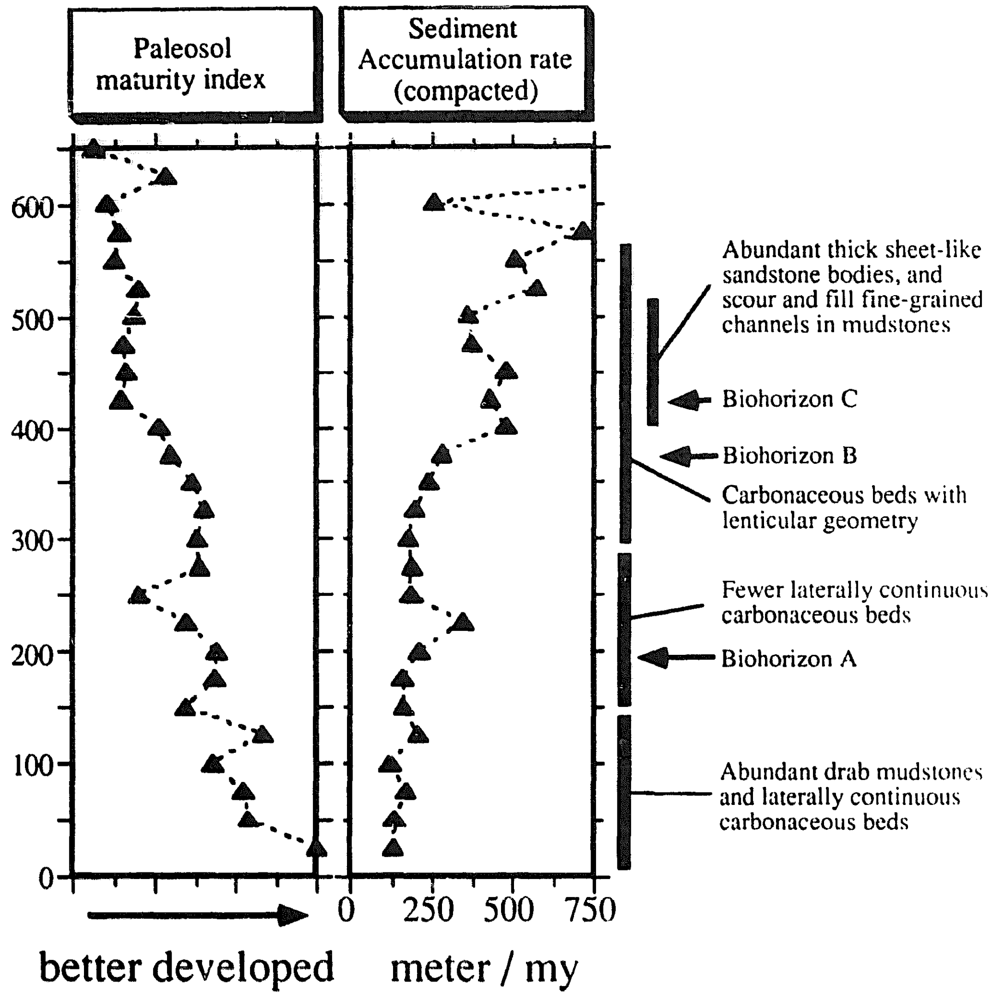

Fig. 10. Interpreted vertical change in paleosol maturity and compacted sediment accumulation rates upsection through the Willwood Formation exposed in the southern part of the Bighorn Basin (modified from Bown and Kraus, 1993). The section was measured along a 85 km long east to west traverse of the basin (i.e., basin axis distal to basin axis proximal). Overall deposition rate of this interval was estimated from dated horizons stratigraphically just above and below this section (Wing et al., 1991). Biohorizons mark important times of faunal change within the succession (see details in Wing et al., 1991; Bown and Kraus, 1993).

1983). Paleocene sediment accumulation rates in the southern part of the basin are not known. Thick lacustrine and palustrine deposits dominated the northern part of the Bighorn Basin (Fig. 9; Hickey, 1980; Gingerich, 1983), probably recording a ponding of drainages due to increased rates of basin subsidence relative to sediment supply. Relatively coarse grained sediments were deposited on multiple alluvial fans adjacent to the Beartooth Mountains, but these deposits extended only a few kilometers into the basin (Hickey, 1980; DeCelles et al., 1991). Elsewhere in the basin, drab alluvial sediments with carbonaceous shales and lignite beds were widespread (upper Fort Union).

Lower Eocene deposits in the basin are dominantly fluvial and these strata comprise the Willwood Formation. The transition from drab alluvial deposits in the upper Fort Union to the brightly variegated, paleosol-rich Willwood Formation probably records generally improved drainage across the basin related to decreased basin subsidence rates but continued high rates of sediment input from the adjacent mountains (Gingerich, 1983). However, this change may also reflect a regional warming and drying of climate (Hickey, 1980; Wing and Bown, 1985; Johnson and Middleton, 1990; Wing et al., 1991). In the northern part of the basin sediment accumulation rates remained high (250 m/m.y.), albeit slightly lower than during the uppermost Paleocene (Gingerich, 1983). In the southern part of the basin Willwood sediment accumulation rates are estimated to have averaged 220 m/m.y., but vary upsection (see below).

Strata directly overlying the Fort Union/Willwood Formation boundary (latest Paleocene to earliest Eocene) have been described across a wide area of the basin, and provide a record of the distribution of coeval paleoenvironments (Wing and Bown, 1985; Kraus, 1992). These coeval strata have been subdivided into broad areas containing similar deposits based on: the size, geometry and interval characteristics of channel sandstone bodies; the color, nodule content and degree of mottling of mudstones; and the abundance and geometry of carbonaceous mudstones. Channel sandstone bodies are generally defined as multi-storied sheets (width to thickness ratio greater than 20) recording relatively large river systems and more restricted ribbons (width to thickness ratio less than 15) recording minor floodplain channels (Kraus and Middleton, 1987; Kraus, 1992; Kraus and Aslan, 1993). Sandstone sheets typically are 6 to 30 m thick, but individual channel deposits within are generally between 4–7 m thick. All ribbon shaped-bodies are less than 7 m thick. Mudstones with dominantly red hues and carbonate nodules are interpreted to record well-drained floodplains, whereas those with predominantly grey hues, siderite nodules or laterally extensive carbonaceous beds are interpreted to record areas that were waterlogged for prolonged periods (Wing, 1984; Wing and Bown, 1985; Kraus, 1992).

Kraus (1992) builds on the work of Wing and Bown (1985), and recognizes four distinct areas within the lower Willwood Formation in the southern part of the Bighorn Basin. The Rairden area (*R* in Fig. 1D) contains abundant sheet sandstones with westward paleocurrent orientations reflecting deposition by a large river system flowing from the Bighorn Mountains (see also Neasham and Vondra, 1972), but also ribbon shaped channel sandstone probably recording deposition within crevasse channels. Adjacent mudstones are grey with orange mottles or pale orange with gray mottles, suggesting poorly to moderately drained floodplains, respectively. Paleosols here were interpreted as immature. The lower most part of the Willwood Formation in this area is different however. These deposits comprise reddish mudstone with carbonate nodules, and lack sheet channel sandstone bodies, suggesting the floodplain was initially better drained and was not traversed by the large river system recorded later.

To the south, deposits in the Sand Creek area (*S* in Fig. 1D) contain smaller ribbon-shaped sandstones oriented southwestward (Bown, 1979), cutting into mostly red and purple mudstones with intense mottling and layers of iron-manganese nodules. Red hues suggest relatively well drained paleosols, but iron-manganese nodules suggest repeated episodes of reduction and oxidization, and seasonal fluctuation in floodplain drainage conditions. Paleosols here were interpreted as mature. To the north in the Antelope Creek area (*A* in Fig. 1D) mudstones are mostly drab with pervasive mottling, but a few horizons are red, suggesting improved drainage conditions locally. Carbonaceous units are abundant and laterally extensive, whereas major channel sandstone bodies are absent. Further north in the Gould Butte area (*G* in Fig. 1D) mudstones are consistently drab and contain siderite nodules, carbonaceous units remain abundant, and major channel sandstones remain absent.

In the northern part of the basin, basal Willwood deposits are exposed along the axis of the basin (i.e., in Clark's Fork sub-basin; *C* in Fig. 1D; Kraus, 1980; Kraus and Middleton, 1987; Gingerich, 1989b). Kraus and Middleton (1987) recognized two different scales of channel sandstone bodies here. Larger sheet sandstones, generally 6–7 m thick and continuous for a few kilometers perpendicular to paleoflow, contain many storeys stacked both vertically and laterally. Individual storeys are defined by sets of inclined (6–12°) beds that are 3–7.5 m thick, suggesting that channels were 50–100 m wide and 3–7.5 m deep. Mean paleocurrents vary between different storeys by up to 140°. These bodies were interpreted as deposits of a sinuous, single-channel river that was stable enough in position to allow time for extensive bar migration and the formation of a well-defined channel belt (Kraus and Middleton, 1987). Mean paleocurrent orientations and the position of these deposits in the basin suggest a river that flowed along the basin axis. If the axial river contained a single active channel (as suggested by Kraus and Middleton, 1987), then this river probably carried a discharge several times lower than those reconstructed form deposits of the Siwalik Chinji Formation, and an order of magnitude lower discharge than those reconstructed from the Siwalik Nagri Formation (cf. Willis, 1993a). However, detailed estimates of channel discharge from deposits in the Bighorn Basin are lacking, and interpretations of channel pattern based only on the presence of inclined beds and estimates of moderately high channel sinuosities are speculative (see discussion in Bridge, 1985). Smaller-scale, ribbon-shaped bodies are less than 7 m thick and extend laterally for less than a few hundred meters perpendicular to paleochannel flow. Many of these bodies also contain multiple storeys, but storeys generally lack inclined beds and show restricted paleocurrent variations. These ribbon-shaped bodies are interpreted as recording smaller channel belts containing low sinuosity channels that did not migrate to any great extent before channel switching or river avulsion occurred (Kraus and Middleton, 1987).

Deposits along the northern basin axis show important changes upsection though the lower Willwood Formation. One stratigraphic interval contains a particularly thick and complex stacking of channel deposits (Kraus, 1980; Kraus and Middleton, 1987). This interval marks the Clarkforkian–Wasatchian boundary, which is an important interval of faunal change in the Bighorn Basin (Gingerich, 1989b). Here a single sandstone body is over 30 m thick and at least 12 km wide. It contains up to six vertically stacked channel deposits. This thick sandstone body records rapid return of the basin-axial river to this location on the floodplain relative to net floodplain aggradation rates, or a prolonged period when the channel belt stayed in one location and aggraded vertically. Overbank deposits adjacent to this sandstone body contain paleosols that are more mature than those observed in vertically adjacent intervals that contain fewer and thinner sheet sandstone bodies. This suggests an association between the concentration of channel deposits and periods of lower net sediment accumulation rates in this northern part of the basin (Kraus and Middleton, 1987). This association is contrary to the case modelled by Alexander and Leeder (1987) in which concentration of channel sandstones along the basin axis was suggested to reflect tectonic tilting of the floodplain towards the thrust belt, preferential avulsion of the channel to the down-tilted side of the basin, and higher deposition rates along the axis of the basin. Rather, Kraus and Middleton (1987) suggest that basin geometry and avulsion frequency of the axial river remained relatively constant while the net aggradation rates of the floodplain temporarily decreased.

Studies of the Lower Willwood Formation have documented important differences between coeval deposits. These differences reflect local basin physiography and floodplain drainage, varying sediment accumulation and basin subsidence rates, the positions of different river systems within the basin (as was also the case within the Siwalik succession). Kraus (1992) postulated that variations between coeval deposits in the southern part of the Bighorn Basin reflect subsidence of crustal blocks along preexisting normal faults that cut across the basin axis, and these faults may also have controlled the positions of rivers entering the basin from the adjacent mountain belt. The diversity of coeval deposits in the lower Willwood Formation clearly suggests there were important spatial changes in environments and habitats across the ancient Bighorn Basin.

The most complete record of Eocene stratal change in the Bighorn Basin occurs along a 80 km E–W traverse that documented a 650 m thick section through the Willwood Formation in the southern part of the basin (Fig. 10; Wing et al., 1991; Bown and Kraus, 1993; Kraus and Bown, 1993). Bown and Kraus (1993) estimated the relative deposition rates of each 25-m thick stratigraphic interval along this section, based on interpretations of the relative paleosol maturities (Fig. 10). They recognized an overall decrease in paleosol maturities upsection. Several stratigraphic intervals show paleosols that break from this general trend, some associated with times of important faunal change. The overall upsection trend may reflect the progression of the logged traverse towards the basin axis where deposition rate was higher, whereas the smaller-scale variations in paleosol maturity were interpreted to reflect temporal changes in sediment accumulation rates (Bown and Kraus, 1993). Sediment accumulation rates are lowest and carbonaceous mudstones are unusually abundant and laterally extensive in the bottom 140 m of this section (20% versus 2% for the Willwood Formation overall, Kraus and Bown, 1993), indicating that floodplains were broad and poorly drained (Wing, 1984). Higher in the section carbonaceous mudstones are lenticular, more spatially restricted, and probably reflect deposition within abandoned overbank channels cutting into better drained floodplains. In the upper part of the Willwood Fm. (Fig. 10, meters 400–500), one interval shows an abrupt decrease in estimated cumulative paleosol maturity, followed by a gradual increase. This interval contains an unusual concentration of thick multi-storied sandstone bodies, and erosional scour and fill deposits (up to tens of meters thick) within overbank successions. These deposits are interpreted to record a period of decreased accommodation space relative to sediment supply and the erosion and intensive reworking of deposits by channel systems. It is difficult to interpret variations observed along this long traverse because upsection changes clearly could reflect a juxtaposition of environments spatially across the basin, as well as temporal changes in environments associated with evolution of the basin.

3. Comparison of depositional systems

The Siwalik Group exposed in Pakistan and the Fort Union/Willwood formations of Wyoming record very different patterns of basin filling. Strata exposed in the Bighorn Basin record deposition in an underfilled foreland basin, where the axis of drainage was aligned along the rapidly subsiding structural axis of the basin (see also Gingerich, 1983; Beck et al., 1988). For the most part, sediments filling the basin appear to have been deposited by a large, low-gradient river (sheet sandstone bodies) flowing along the western margin of the basin, and to a lesser extent by smaller, low sinuosity rivers flowing transverse to the basin axis (ribbon sandstone bodies; eg. Kraus and Middleton, 1987). However, details of the size, depositional style, and flow orientations of river systems across most stratigraphic intervals remain to be documented. During periods of tectonism, rates of sediment input to the basin did not keep pace with rates of basin subsidence. Lacustrine and palustrine environments replaced fluvial environments in rapidly subsiding areas of the basin during deposition of the Fort Union Formation. Fluvial environments became restricted to areas associated with coarse-grained alluvial fans directly adjacent to the western mountain belt, areas to the far east side of the basin where subsidence rates were lower, and areas along the basin axis to the north where drainages exited the basin. As basin subsidence rates decreased relative to sediment supply, low areas within the basin filled and fluvial environments again became dominant in the basin during deposition of the Willwood Formation. Drainage conditions across the basin improved as low areas in the basin filled, there was widespread development of paleosols containing carbonate nodules, and perhaps there was a greater contribution of sediments deposited directly from smaller basin-transverse rivers (particularly these originating in the mountains to the west; i.e., see paleocurrent trends in the Willwood Fm. documented by Neasham and Vondra, 1972). Thus, as the basin evolved there were major changes in the spatial distribution of depositional environments and habitats over time. These environmental changes controlled lithologic variations used to define formations. Such changes also must have exerted a strong influence on the ecology of faunas and floras, and on taphonomic modes (e.g., Wing and Bown, 1985; Gingerich, 1989b; Behrensmeyer et al., 1994).

In contrast, Siwalik Group strata appear to record deposition of an overfilled basin, where rates of sediment input to the basin always exceeded rates of basin subsidence. This assumes that the ancient Siwalik Basin was similar to the modern Himalayan foredeep, where deposition across most of the basin is associated with basin-transverse river systems and the axis of drainage is confined to the mountain distal margin of the basin. This assumption is supported by constant continental convergence rates throughout the Miocene to Recent (Patriat and Achache, 1984; Besse and Courtillot, 1988), and the continuous supply of sediments that bypassed the basin and were deposited on the Indus and Bengal submarine fans in the northern Indian Ocean (e.g., Whitmarsh et al., 1974). During periods of tectonism and increased basin subsidence, sediments within the Siwalik Basin became coarser, the proportion of channel relative to overbank deposits increased dramatically, and net sediment aggradation rates doubled. In areas now uplifted and exposed in the Chinji Village area of the Potwar Plateau, the sediment fan deposited by a larger river (i.e., Nagri Fm.) appears to have expanded at the expense of floodplain areas traversed by a smaller river (i.e., Chinji Fm.). Floodplain environments on this large fan (i.e., associated with the larger river) appear to have been better drained, exhibited less episodic deposition with fewer long hiatuses, contained few minor overbank channels that maintained flow throughout the year, and generally preserved fewer fossil remains (Badgley and Behrensmeyer, 1980; Behrensmeyer, 1987; Willis, 1993a). It is less clear that environments and habitats of individual river systems changed markedly through time in association with tectonism, rather than simply shifting spatially.

4. Comparison of environmental change

Variations in the ratio of channel deposits to overbank sequences, the size and depositional style of river channels, the lithology of overbank sequences and the character of paleosols within the Bighorn Basin and the Himalayan foredeep of northern Pakistan may reflect basin-scale changes in the distribution of environments, and thus biota, through time. However, several factors bias these large-scale sedimentologic records of environmental change. Biases affecting interpretation of environmental change from the sedimentologic record include estimating how deposits vary across the basin and through time from the exposed strata, and relating the distribution and relative proportions of different types of sedimentary deposits within the basin fill to changing spatial distributions of actual environments across the ancient alluvial landscapes. Potential biases relating to interpretations of environmental change within each of these basins are discussed below.

Exposures across the Potwar Plateau provide a relatively small "window" through which to view changing paleoenvironments in the ancient Siwalik basin. This window is located within the upstream reaches of the depositional basin. Because deposits and the taphonomy of fossil remains clearly can differ between adjacent river systems within this basin, relatively minor shifts in the positions of these river systems may produce dramatic changes in the relatively small area that is exposed. For example, sediment variations across the Chinji–Nagri Fm. boundary were probably related to expansion of the sediment fan deposited by a major river into basin areas formerly occupied by a smaller river system (Willis, 1993b). Sedimentologic and taphonomic changes across this transition appear to reflect contrasts between these different river systems rather than temporal changes within an individual river system. Basin-scale processes, such as tectonically controlled rates of sediment supply and basin subsidence, may ultimately control the positions of the different rivers within the basin and thus the nature of deposits. However, we cannot assume that variations observed within one area are characteristic of the basin as a whole. Changes in deposits over time will be strongly influenced by the position of the exposed "window" in the basin, the characteristics of the particular river systems exposed in that window, and whether the view stays focused on a single river system or changes as different river systems shift in position within the evolving basin.

Exposures extend over the entire Bighorn Basin, and thus these strata provide a better record of across-basin environmental changes. Coeval strata are exposed along north-south elongated zones (Gingerich, 1983; Wing and Bown, 1985), and thus environmental changes parallel to the basin axis are better recorded than mountain proximal–distal variations. Further, it is not possible to measure more than a few 100m of vertical section without making significant traverses across the basin. Thus distinguishing between temporal and spatial changes in depositional environments is also problematic in the Bighorn Basin, but perhaps to a lesser extent than in the Siwalik succession.

The proportion of sediments deposited in different fluvial environments does not directly reflect aerial proportions of environments across the ancient landscape. In general, environments associated with higher sediment aggradation rates will be over-represented relative to those associated with lower aggradation rates, and some types of deposits would be expected to compact during burial to a greater extent. For example, most overbank deposits in the Siwaliks represent deposition on crevasse splays that were partially disrupted during subsequent episodes of pedogenesis (eg. Willis, 1993a; Willis and Behrensmeyer, 1994; Behrensmeyer et al., 1994). However, a thick sequence of overlapping crevasse splay deposits may represent a short and localized period of rapid deposition on the floodplain, whereas a single paleosol may represent a long period when broad areas of the floodplain were covered by mature forests. Thus the dominance of crevasse splay deposits in Siwalik overbank sequences does not indicate that most areas on ancient Siwalik floodplains were active sites of crevasse splay deposition at any one time, and in fact it is more likely that most areas of the floodplain were covered by developing soils. Similarly, the proportion of channel relative to overbank deposits is not controlled only by the proportion of the ancient floodplain occupied by active channel belts, but also reflects the avulsion frequencies of channel belts relative to net sediment aggradation rates on floodplains (Bridge and Leeder, 1979).

Although upsection changes in the lithologic features of sediments within these basins clearly reflect variations in depositional processes, in some cases lithologic changes may not reflect major changes in the spatial distribution of paleoenvironments and habitats across the basin. For example, pronounced ecological changes in paleoenvironments that are not sites of deposition may not be recorded by lithologic variations (e.g., environments characterized by soil formation or peat accumulation). In the Siwalik Dhok Pathan Formation, a distinct change of carbon isotopes in paleosol carbonate nodules and associated thickening of preserved organic layers in the upper parts of paleosols are interpreted to reflect a major change in climate and flora (Quade et al., 1989). However, this isotopic change does not occur at a formation boundary, and it is not associated with distinct changes in preserved depositional environments or the taphonomy of fossil remains.

Temporal or spatial changes in patterns of erosion may also cause pronounced variations in preserved deposits and associated organic remains that are unrelated to environmental change across an ancient basin. For example, in the Siwalik Group, decreases in laminated claystones and the proportion of paleosols with iron nodules across the Chinji–Nagri boundary may record a basin-wide improvement of floodplain drainage over time. Alternatively, these variations may reflect increased preferential removal of deposits representing channel-distal environments, caused by more rapid return of the channel to a given area of the floodplain (i.e., channel belt return frequencies are three times greater in the Nagri Formation relative to the Chinji Formation; Willis, 1993b). In the Bighorn Basin, biases associated with sediment preservation may not be as great because the stratigraphic succession is more mud-rich overall, and thus fewer deposits can be expected to have been removed by erosion.

Finally, in both areas some environments are far more important for the preservation of organic remains than others (Badley et al., 1994; Behrensmeyer et al., 1994). Changes that affect such taphonomically important environments will have a more pronounced effect on the quality of the fossil record. For example, there is a marked change in the deposits and the abundance of fossil remains observed across the Chinji–Nagri Formation boundary in the Siwalik Group. Some of the difference in fossil abundance across this formation boundary is clearly related to the greater proportion of overbank deposits in the Chinji Formation (70%) relative to the Nagri Formation (30%); however, even with this difference acknowledged, overbank sequences of the Nagri Formation typically contain fewer fossil remains than those in the Chinji Formation. The majority of fossil remains are observed in fills of minor floodplain channels (Behrensmeyer, 1987), which constitute a low proportion of overbank deposits within both formations (Willis, 1993a). However, the character of minor channel fills varies between these two formations. Floodplains of the Chinji Formation were more poorly drained and typical minor floodplain channels appear to have contained at least some flow throughout the year, in contrast such channels in the Nagri Formation appear to record more ephemeral flow (see above and Willis, 1993a). The change in the abundance of fossils across the Chinji–Nagri Formation boundary probably was influenced by this difference, despite the fact that minor channel environments covered only a small area of the ancient floodplain during any one time.

5. Conclusions

The Bighorn and Siwalik (Himalayan) Basins both preserve long records of alluvial deposition adjacent to rising mountain belts and contain abundant organic remains. The Bighorn Basin was generally underfilled with sediments, while the Siwalik Basin was overfilled. Periods of tectonism adjacent to the Bighorn Basin and more rapid rates of basin subsidence were characterized by more mudstone-rich strata overall, a ponding of drainages along the axis of the basin, and the restriction of coarser-grained clastics to basin margins. Deposit variations used to define formations in the Bighorn Basin reflect important changes in the spatial distribution of paleoenvironments through time related to varying rates of basin subsidence and sediment supply across the basin. Periods of tectonism adjacent to the Siwalik Basin were associated with dramatic increases in the proportion of channel sandstone bodies relative to overbank successions. It is less clear, however, that these periods of tectonism were associated with a dramatic change in the spatial distribution of paleoenvironments across the entire basin. Upsection variations observed within strata of the Potwar Plateau may reflect environmental differences between coeval river systems that become superimposed over time and/or changing patterns of deposition and erosion over time which controlled the amounts of sediments preserved from different environments. In the Himalayan foredeep of northern Pakistan, the small scale of exposures relative to the scale of the basin and relative to individual river systems within the basin, pronounced contrasts in the characteristics of different river systems, pronounced temporal changes in the proportion of channel to overbank deposits preserved, and the restriction of preserved organic remains to spatially limited depositional environments all contribute a bias to reconstruction of the fossil record in relation to the distribution of environments that made up the ancient alluvial ecosystems. In contrast, a relatively large area of the Bighorn Basin is exposed, deposits from floodplain environments are more completely preserved, and fossils occur in a wider range of deposit types throughout the succession (including paleosols).

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References

Alexander, J. and Leeder, M.R., 1987. Active tectonic control on alluvial architecture. In: F.G. Ethridge, R.M. Flores and M.D. Harvey (Editors), Recent Developments in Fluvial Sedimentology. Soc. Econ. Paleontol. Mineral. Spec. Publ., 39: 243-252.

Anano, K. and Taira, A., 1992. Two-phase uplift of higher Himalayas since 17 Ma. Geology, 5: 391-395.

Badgley, C., 1986. Taphonomy of Mammalian fossil remains from Siwalik rocks of Pakistan. Paleobiology, 12: 119-142.

Badgley, C. and Behrensmeyer, A.K., 1980. Paleoecology of Middle Siwalik sediments and faunas, northern Pakistan. Palaeogeogr. Palaeoclimatol. Palaeoecol., 30: 133-155.

Badgley, C., Bartels, W.S., Morgan, M.E., Behrensmeyer, A.K. and Raza, S.M., 1995. Taphonomy of vertebrate assemblages from the Paleogene of northwest Wyoming and the Neogene of northern Pakistan. Palaeogeogr. Palaeoclimatol. Palaeoecol., 115: 157-180.

Barry, J.C., Johnson, N.M., Raza, S.M. and Jacobs, L.L., 1985. Neogene mammalian faunal change in southern Asia: Correlations with climatic, tectonic and eustatic events. Geology, 13: 637-640.

Barry, J.C., Behrensmeyer, A.K. and Monaghan, M., 1980. A geologic and biostratigraphic framework for Miocene sediments near Khaur Village, northern Pakistan. Postilla, 183: 1-19.

Beck, R.A. and Burbank, D.W., 1990. Continental-scale diversion of rivers: A control of alluvial stratigraphy. Geol. Soc. Am. Abstr. Programs, 22: 238.

Beck, R.A., Vondra, C.F., Filkins, J.E. and Olander, J., 1988. Syntectonic sedimentation and Laramide basement-thrusting, Cordilleran foreland: Timing of deformation. In: C.J. Schmidt and W.J. Perry (Editors), Interaction of the Cordilleran Thrust Belt and Cordilleran Foreland. Geol. Soc. Am. Mem., 171: 465-487.

Behrensmeyer, A.K., 1987. Miocene fluvial facies and vertebrate taphonomy in northern Pakistan. In: F.G. Ethridge, R.M. Flores and M.D. Harvey (Editors), Recent Developments in Fluvial Sedimentology. Soc. Econ. Paleontol. Mineral. Spec. Publ., 39: 169-176.

Behrensmeyer, A.L. and Tauxe, L., 1982. Isochronous fluvial systems in Miocene deposits of Northern Pakistan. Sedimentology, 29: 331-352.

Behrensmeyer, A.K., Willis, B.J. and Quade, J., 1995. Floodplains and paleosols of Pakistan Neogene and Wyoming Paleogene deposits: Implications for the taphonomy and paleoecology of faunas. Palaeogeogr. Palaeoclimatol. Palaeoecol., 115: 37-60.

Berggren, W.A., Kent, D.V., Flynn, J.J. and Van Couvering, J.A., 1985. Cenozoic geochronology. Geol. Soc. Am. Bull., 96: 1407-1418.

Besse, J. and Courtillot, V., 1988. Paleogeographic maps of the continents bordering the Indian Ocean since the Early Jurassic. J. Geophys. Res., 93: 11,791-11,808.

Bonini, W.E. and Kinard, R.E., 1983. Gravity anomalies along the Beartooth Front, Montana. In: Annu. Field Conf. Wyo. Geol. Assoc. Guidebook, 34: 89-94.

Bown, T.M., 1982. Geology, paleontology, and correlation of Eocene volcaniclastic rocks, southeast Absaroka Range, Hot springs County, Wyoming. U. S. Geol. Surv. Prof. Pap., 1201-A: 1-75.

Bown, T.M., 1980. Summary of latest Cretaceous and Cenozoic sedimentary, tectonic, and erosional events, Bighorn Basin, Wyoming. In: P.D. Gingerich (Editor), Early Cenozoic Paleontology and Stratigraphy of the Bighorn Basin, Wyoming. Univ. Mich. Pap. Paleontol., 24: 87-94.

Bown, T.M., 1979. Geology and mammalian paleontology of the Sand Creek facies, lower Willwood Formation (lower Eocene), Washakie County, Wyoming. Geol. Surv. Wyo. Mem., 2: 1-151.

Bown, T.M., 1975. Paleocene and Lower Eocene rocks in the Sand Creek-No Water Creek area, Washakie County, Wyoming. Wyo. Geol. Assoc. Guidebook, 27: 55-61.

Bown, T.M. and Beard, K.C., 1990. Systematic lateral variation in the distribution of fossil mammals in alluvial paleosols, lower Eocene willwood Formation, Wyoming. In: T.M. Bown and K.D. Rose (Editors), Dawn of the Age of Mammals in the Northern Part of the Rocky Mountain Interior, North America. Geol. Soc. Am. Spec. Pap., 243: 135-151.

Bown, T.M. and Kraus, M.J., 1993. Time-stratigraphic reconstruction and integration of paleopedologic, sedimentologic, and biotic events (Willwood Formation, Lower Eocene, northwest Wyoming, U.S.A.). Palaios, 8: 68-80.

Bown, T.M. and Kraus, M.J., 1987. Integration of channel and floodplain suites, 1. Development sequence and lateral relations of alluvial paleosols. J. Sediment. Petrol., 57: 587-601.

Bridge, J.S., 1985. Paleochannel patterns inferred from alluvial deposits: a critical evaluation. J. Sediment. Petrol., 55: 579-589.

Bridge, J.S. and Leeder, M.R., 1979. A simulation model of alluvial stratigraphy. Sedimentology, 26: 617-644.

Burbank, D.W. and Beck, R.A., 1989. Early Pliocene uplift of the Salt Range; Temporal constraints on thrust wedge development, northwest Himalaya, Pakistan. Geol. Soc. Am. Spec. Pap., 232: 113-128.

Burbank, D.W., Beck, R.A., Raynolds, R.G.H., Hobbs, R. and Tahirkheli, R.A.K., 1988. Trust and gravel progradation in foreland basins: A test of post-thrusting gravel dispersal. Geology, 16: 1143-1146.

Burbank, D.W., Raynolds, R.G.H. and Johnson, G.D., 1986. Late Cenozoic tectonics and sedimentation in the northwestern Himalayan foredeep: II. Eastern limb of the northwest syntaxis and regional synthesis. In: P.A. Allen and P. Homewood (Editors), Foreland Basins. Int. Assoc. Sedimentol. Spec. Publ., 8: 293-306.

Decelles, P.G., Gray, M.B., Ridgway, K.D., Cole, R.B., Srivastava, P., Pequera, N. and Pivnik, D.A., 1991. Kinematic history of a foreland uplift from Paleocene synorogenic conglomerate, Beartooth Range, Wyoming and Montana. Geol. Soc. Am. Bull., 103: 1458-1475.

Fatmi, A.N., 1973. Lithostratigraphic units of the Kohart-Potwar Province, Indus Basin, Pakistan. Mem. Geol. Surv. Pakistan, 10: 1-80.

Friend, P., 1989. Hydrological controls of alluvial geometry in the Indo-Gangetic sub-basins. In: Progr. Abstr. 4th Int. Conf. Fluvial Sedimentology, Barcelona-Sitges, Int. Assoc. Sedimentol., p. 133.

Geddes, A., 1960. The alluvial morphology of the Indo-gangetic plains. Trans. Inst. Br. Geogr., 28: 253-276.

Gill, W.D., 1951. The stratigraphy of the Siwalik Series in the northern Potwar, Punjab, Pakistan. Geol. Soc London Q. J., 107: 375-394.

Gingerich, P.D., 1989a. Paleocene and early Eocene of the Bighorn Basin and Clark's Fork Basins, Wyoming. In: J. Flynn (Editor), Mesozoic/Cenozoic Vertebrate Paleontology; Classic Localities, Contemporary Approaches. Field Trips 28th Int. Geol. Am. Geophys. Union, Washington, DC, pp. 47-57.

Gingrich, P.D., 1989b. New Earliest Wasatchian mammalian fauna from the Eocene of northwestern Wyoming: Composition and diversity in a rarely sampled high-floodplain assemblage. Univ. Mich. Pap. Paleontol., 28: 1-97.

Gingerich, P.D., 1983. Paleocene-Eocene faunal zones and a preliminary analysis of Laramide structural deformation in the Clark's Fork Basin, Wyoming. Wyo. Geol. Assoc. Annu Field Conf. Guidebook, 34: 185-195.

Gole, C.V. and Chitale, S.V., 1966. Inland delta building activity of the Kosi River. Proc. Am. Soc. Civil Eng. J. Hydrol. Div., 92: 111-126.

Hinkey, L.J., 1980. Paleocene stratigraphy and flora of the Clark's Fork Basin: In: P.D. Gingerich (Editor), Early Cenozoic Paleontology and Stratigraphy of the Bighorn Basin, Wyoming. Univ. Mich. Pap. Paleontol., 24: 33-49.

Johnson, G.D., Raynolds, R.G. and Burbank, D.W., 1986. Late Cenozoic tectonics and sedimentation in the northwestern Himalayan foredeep: I. Thrust ramping and associated deformation in the Potwar region. In: P. Allen and P. Homewood (Editors), Foreland Basins. Int. Assoc. Sedimentol. Spec. Publ., 8: 273-291.

Johnson, N.M., Opdyke, N.D., Johnson, G.D., Lindsay, E.H. and Tahirkheli, R.A.K., 1982. Magnetic polarity stratigraphy and ages of Siwalik group rocks of the Potwar Plateau, Pakistan. Palaeogeogr. Palaeoclimatol. Palaeoecol., 37:17-42.

Johnson, N.M., Stix, J., Tauxe, L., Cerveny, P.F. and Tahirkheli, R.A.K., 1985. Palaeomagnetic chronology, fluvial processes and tectonic implications of the Siwalik deposits near Chinji Village, Pakistan. J. Geol., 93: 27-40.

Johnson, S.J. and Middleton, L.T., 1990. Tectonic significance of Paleocene alluvial sequence, Clark's Fork Basin, Wyoming-Montana. In: R.W. Specht (Editor), Wyoming Sedimentation and Tectonics. Wyo. Geol. Assoc. Guidebook, 41: 69-88.

Kraus, M.J., 1992. Alluvial response to differential subsidence: Sedimentologic analysis aided by remote sensing, Willwood Formation (Eocene), Bighorn Basin, Wyoming, U.S.A. Sedimentology, 39: 455-470.

Kraus, M.J., 1980. Genesis of a fluvial sheet sandstone, Willwood Formation, northwest Wyoming. In: P.D. Gingerich (Editor), Early Cenozoic Paleontology and Stratigraphy of the Bighorn Basin, Wyoming. Univ. Mich. Pap. Paleontol., 24: 87-94.

Kraus, M.J. and Aslan, A., 1993. Eocene hydromorphic paleosols: significance for interpreting ancient floodplain processes. J. Sediment. Petrol., 63: 453-463.

Kraus, M.J. and Bown, T.M., 1993. Short-term sediment accumulation rates determined from Eocene alluvial paleosols. Geology, 21: 743-746.

Kraus, M.J. and Middleton, L.T., 1987. Contrasting architecture of two alluvial suites in different structural settings. In: F.G. Ethridge, R.M. Flores and M.D. Harvey (Editors), Recent Developments in Fluvial Sedimentology. Soc. Econ. Paleontol. Mineral. Spec. Publ., 39: 253-262.

McRae, L.E., 1990. Paleomagnetic isochrons, unsteadiness, and non-uniformity of sedimentation in Miocene fluvial strata of the Siwalik Group, northern Pakistan. J. Geol., 98: 433-456.

Neasham, J.W. and Vondra, C.F., 1972. Stratigraphy and petrology of the lower Eocene Willwood Formation, Bighorn Basin, Wyoming. Geol. Soc. Am. Bull., 55: 165-210.

Parkash, B., 1991. Sedimentation in the middle Gangetic Plains and its implications for interpretation of the Siwalik sediments. Abstr. Birbal Sahni Birth Centenary Symp. Siwalik Basin, Wadia Inst. Himalayan Geol., Dehra Dun, p. 29-30.

Parkash, B., Awasthi, A.K. and Gohain, K., 1983. Lithofacies of the Markanda terminal fan, Kurukshetra district, Haryana, India. In: J.D. Collinson and J. Lewis (Editors), Modern and Ancient Fluvial Systems. Int. Assoc. Sedimentol. Spec. Publ., 6: 405-420.

Parkash, B., Sharma, R.P. and Roy, A.K., 1980. The Siwalik Group (molasse)—Sediments shed by collision of continental plates. Sediment. Geol., 25: 127-159.

Parker, S.E. and Jones, R.W., 1986. Influence of faulting on Upper Cretaceous-lower Tertiary deposition, Bighorn Basin, Wyoming. In: 50th Annu. Field Conf. Guidebook, Mont. Geol. Soc. Yellowstone Bighorn Res. Assoc., pp. 137-144.

Patriat, P. and Achache, J., 1984. India-Eurasia collision chronology has implications for crustal shortening and driving mechanism of plates. Nature, 311: 615-621.

Pilbeam, D.R., Behrensmeyer, A.K., Barry, J.C. and Shaw, S.M.I., 1979. Miocene sediments and faunas of Pakistan. Postilla, 179: 1-45.

Quade, J., Cerling, T.E. and Bowman, J.R., 1989. Development of Asian monsoon revealed by marked ecological shift during the latest Miocene in northern Pakistan. Nature, 342: 163-166.

Raza, S.M., 1983. Taphonomy and paleoecology of Middle Miocene vertebrate assemblages, Southern Potwar Plateau Pakistan. Thesis. Yale Univ., New Haven, CT (unpublished).

Rea, B.D. and Barlow, J.A., 1975. Upper Cretaceous and Tertiary rocks, northern part of Bighorn Basin, Wyoming and Montana. Wyo. Geol. Assoc. Annu. Field Conf. Guidebook, 27: 63-71.

Sahni, A. and Mitra, H.C., 1980. Neogene palaeobiogeography of the Indian subcontinent with special reference to fossil vertebrates. Palaeogeogr. Palaeoclimatol. Palaeoecol., 31: 39-62.

Shah, S.M.I., 1977. Stratigraphy of Pakistan. Mem. Geol. Surv. Pak., 12: 1-138.

Stone, D.S., 1985. Geological interpretation of seismic profiles, Bighorn Basin, Wyoming, Part II: West flank. In: R.R. Gries and R.C. Dyer (Editors), Seismic Exploration of the Rocky Mountain region. Rocky Mt. Assoc. Geol., Denver Geophys. Soc., Denver, CO, pp. 175-186.

Sundell, K.A., 1990. Sedimentation and tectonics of the

Absaroka Basin of northwestern Wyoming. In: R.W. Specht (Editor), Wyoming Sedimentation and Tectonics. Wyo. Geol. Guidebook, 41: 105–122.

Van Houten, F.B., 1944. Stratigraphy of the Willwood and Tatman Formations in northwestern Wyoming. Geol. Soc. Am. Bull., 55: 165–210.

Wells, N.A. and Dorr, J.A., 1987a. A reconnaissance of sedimentation on the Kosi alluvial fan of India. In: F.G. Ethridge, R.M. Flores and M.D. Harvey (Editors), Recent Developments in Fluvial Sedimentology. Soc. Econ. Paleontol. Mineral. Spec. Publ., 39: 51–61.

Wells, N.A. and Dorr, J.A., 1987b. Shifting of the Kosi River, northern India. Geology, 15: 204–207.

Whitmarsh, R.B., Weser, O.E., Ross, D.A. et al., 1974. Init. Rep. DSDP, 23: 33–527.

Willis, B.J., 1993a. Ancient river systems in the Himalayan Foredeep, Chinji Village area, northern Pakistan. Sediment. Geol., 88: 1–76.

Willis, B.J., 1993b. Evolution of Miocene fluvial systems in the Himalayan foredeep through a two kilometer-thick succession in northern Pakistan. Sediment. Geol., 88: 77–121.

Willis, B.J. and Behrensmeyer, A.K., 1994. Architecture of Miocene overbank deposits in northern Pakistan. J. Sediment. Res., B64: 60–67.

Wing, S.L., 1984. Relation of paleovegetation to geometry and cyclicity of some fluvial carbonaceous deposits. J. Sediment. Petrol., 54: 52–66.

Wing, S.L. and Bown, T.M., 1985. Fine scale reconstruction of Late Paleocene-Early Eocene paleogeography in the Bighorn Basin of northern Wyoming. In: R.M. Flores and S.S. Kaplan (Editors), Cenozoic Paleogeography of West-central United States. Rocky Mt. Sect., Soc. Econ. Paleontol. Mineral., pp. 93–105.

Wing, S.L., Bown, T.M. and Obradovich, J.D., 1991. Early Eocene biotic and climatic change in interior western North America. Geology, 19: 1189–1192.

Yeats, R.S., Khan S.H. and Akhtar, M., 1984. Late Quaternary deformation of the salt Range of Pakistan. Geol. Soc. Am. Bull., 95: 958–966.