Indian Journal of Geology
Vol. 72, No. 2, p. 77-106, 2000

RECOGNITION AND ANALYSIS OF FLUVIAL DEPOSITS : A BRIEF OVERVIEW

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ABSTRACT

This article reviews the fundamentals of fluvial sedimentology and recent developments in the recognition and analysis of fluvial deposits. Different facies of the fluvial system, and the patterns of their organisation in a succession are discussed in relation to short- and long-term evolution of the fluvial system in response to autogenic and allogenic processes.

In accordance with the modern systems, ancient alluvial successions show two major components: 1) channel-belt coarser sedimentaries and 2) overbank finer sedimentaries. Intensive studies in the last two decades have developed a specific methodology for characterisation of channel-belt deposits - known as architectural element analysis that involves identification of different orders of fluvial bedforms and their bounding surfaces within a fluvial succession.

The overbank deposits record flood events, and provide evidence for paleoclimate in the soil profiles that develop on the sediments during the long interval between successive flood events. Moreover, overbank deposits include the sedimentary record of the process of avulsion - the geologically sudden change in the course of a river. The frequency of avulsion and net sedimentation rate largely control the proportion and connectedness of channel-belt sediment bodies within the fine-grained floodplain deposits, which can be utilised to build up a sequence stratigraphic framework for the alluvial succession.

Alluvial systems are generally associated with intracontinental rift, strike-slip and foreland basins. Sediment dispersal in these basins principally takes place through axial and transverse drainages. Tectonic movements along the basin bounding faults affect the fluvial style and largely control the stratigraphic distribution of the axial and transverse deposits in the basin fill. The effects of long-term changes in climate on the large-scale pattern of alluvial basin fill have also been discussed.

Key Words : Fluvial Deposits, ancient, modern, recognition.

INTRODUCTION

Fluvial deposits are an important component of the sedimentary rock record. They host a variety of economic mineral deposits including fossil fuels and placers. Sedimentary strata representing products of ancient alluvial systems are reliable documents of the past earth surface processes and are also extremely useful in analysing paleoclimates. Unraveling the geologic controls on the large-scale stratigraphic architecture of the basin-fill (sequence stratigraphy), although has been successful in marine or fluvio-marine basins, the continent-interior fluvial successions still remain fuzzy in this respect demanding more detailed introspection.

In India and elsewhere, the coal-bearing Gondwana strata consist dominantly of fluvial deposits (Casshyap, 1979; Veevers & Tewari, 1995).

Much of the Tertiary hydrocarbon deposits are also hosted in fluvial strata. Fluvial deposits have also been reported from a number of Proterozoic sedimentary basins of India (Bhattacharyya & Morad, 1993; Chakraborty, 1996; Chakraborty, 1999). In addition to these fluvial successions, a giant alluvial prism has accumulated in the foreland basin, in front of the Great Himalayan belt since Tertiary (Tandon et al. 1985; Zaleha, 1997a). The present-day alluvial systems in the Ganga plains also provide the opportunity to observe different fluvial processes in action (Sinha & Friend, 1994; Singh, 1972), offering scope to unravel the cause-and-effect relationships with regard to depositional dynamics and stratigraphic preservation of fluvial systems.

The generalised importance of the fluvial sedimentology and its enormous scope in the Indian context has provided the backdrop for this overview.

The article discusses the different geomorphologic features and processes of the alluvial plain, and reviews the distinguishing features of ancient fluvial deposits and their stratigraphic architectures in relation to the causal geologic factors.

GEOMORPHOLOGY OF THE ALLUVIAL PLAIN

An alluvial plain is the geographic domain bordered by topographic barriers within which processes related to fluvial sediment transport and deposition operate (Fig. 1). It comprises principally of two geomorphic components: (1) the channels, which are the main conduits through which water flows most of the time carrying the sediment-load, and (2) the overbank environment which surrounds the channels. Sediment diffusion in the overbank areas takes place only during flood events when the water-carrying capacity of the channels are exceeded and as a result water spills over to areas outside the channels (Fig. 1). In addition to periodic flooding, a variety of other processes particularly those related to soil formation and aeolian activity affect the overbank environment during the long interval between successive floods. Recent investigation of the Ganga River alluvial plain has revealed that the channel belts at places are flanked by extensive, elevated tracts that are not inundated even during the floods and have been referred to as "upland interfluve" or "Doab". Accumulation of fine-grained sediments in these tracts appears

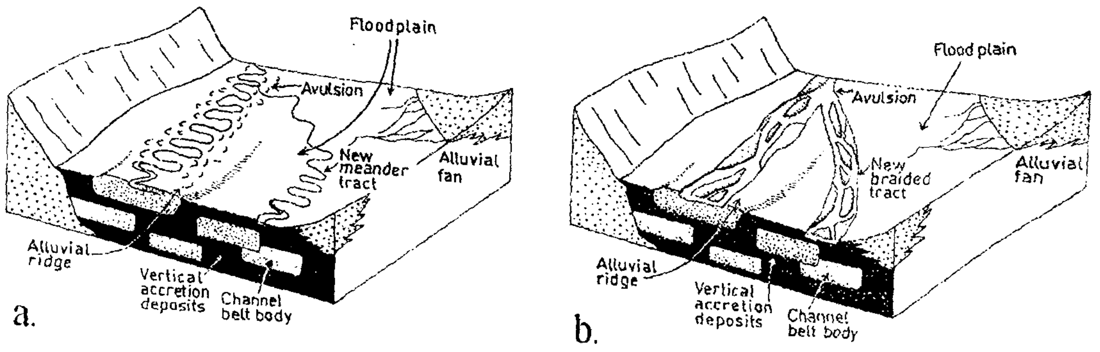

Fig. 1 Diagram illustrating an alluvial plain flanked by topographic barriers feeding alluvial fans. The alluvial plain is shown to consist of a river tract (meandering in a and braided in b) surrounded by floodplain areas. The frontal vertical section shows the sedimentary succession that develops over a geological time interval under net aggradational situation. Note that the alluvial succession comprises two kinds of deposits: 1. channel bodies (stippled) and 2. floodplain deposits (black). Multiple channel bodies in the succession result due to repeated changes in the river course (avulsion). Also note that the width of the channel sediment body is larger than that of the active river tract defining what is known as channel-belt width.

to be unrelated to present day processes operating in the main channel (Singh et al., 1999).

We shall be mainly dealing with alluvial rivers, i.e. those which flow over earlier deposits of channels and overbank, and exclude discussion of features related to bedrock channels.

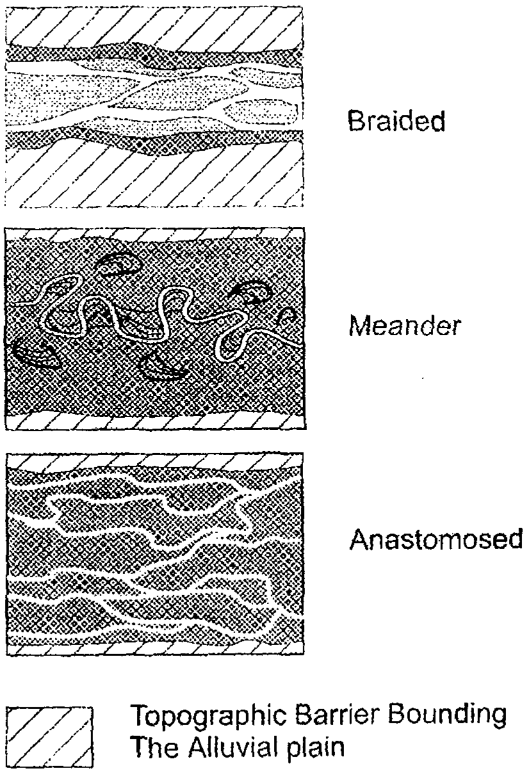

Fig. 2 Schematic representation of the principal types of natural river patterns in plan.

Channel system

Natural rivers generally show three distinct planform patterns of channels: a) meandering, b) braided and c) anastomosed (Fig. 2). Rivers characterised by single straight channels do occur in nature but are less common and they may eventually become sinuous (see later). Among the above three, meandering rivers are characterised by single channel whereas braided and anastomosed rivers are essentially multi-channel systems (Fig. 2). Within an alluvial plain, the channel/s of the river could shift their positions laterally within a specified width defining what is known as a channel-belt (Fig. 1; Leopold & Wolman, 1957, Bridge and Leeder, 1979, see later sections).

Natural channels may show innumerable planform patterns and some of the workers emphasise that, since the processes in all types of channels are dynamically similar they represent a continuum and so any attempt to classify them into subgroups is essentially artificial (e.g. Bridge, 1985, 1993). We, however, believe that classifying the natural fluvial systems into three important types mentioned above, is useful in denoting certain distinctive aspects of channel processes, and in spite of some overlaps in the processes associated with different channel patterns, such classification significantly aids characterising and interpreting ancient alluvial deposits.

Meandering rivers : Meandering river systems are characterized by a single, sinuous channel with bordering floodplains. Meandering rivers, when compared with the braided ones are usually characterised by gentler gradients, finer sediment-load and lesser discharge fluctuations (Knighton, 1984). It should be mentioned that exception to the above criteria is not uncommon and gravelly meandering streams (McGowen & Garner, 1970; Levey, 1978) or silty braided streams do exist in nature, although rarely.

All the natural channels show some amount of curvature in planform but the pronounced sinuous pattern in the channel-course, often resembling sinusoidal waves, is the hallmark of the meandering variety. The channel sinuosity is defined as:

$$\frac{\textit{Curvilinear channel length}}{\textit{Linear valley length}}$$

Meandering channels are considered to have sinuosity greater than an arbitrary critical value of 1.5. However, patterns of many natural meandering channels are highly irregular with little resemblance to the idealised sine curve and appear to be fractal in nature (Ghosh, 2000).

It has been shown both experimentally and from the study of natural streams that in straight channels preferential deposition occurs in the alternate banks in response to increased resistance to the flow near the banks. Sediment bodies formed in the alternate banks of the channel (alternate/side-bars) then force the water to take a sinuous path (Fig. 3). The incipiently sinuous flow path in turn, further enhances growth of the side-attached bars, eventually converting them to point bars typical of meandering rivers and the channel assumes a strongly sinuous course (Fig. 3; Leopold & Wolman, 1957, Bridge & Diemer, 1983).

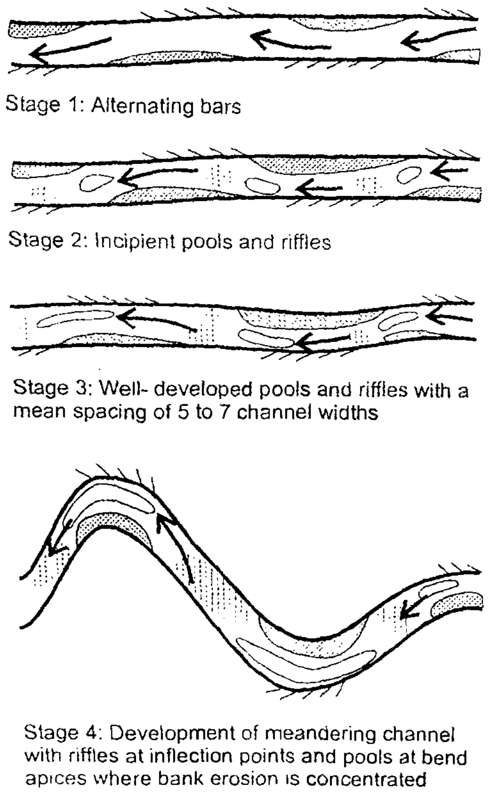

Fig. 3 Successive stages of evolution of a meandering river from a straight channel (after Kinghton, 1984).

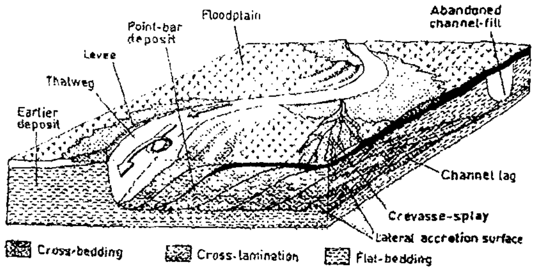

Fig. 4 Three-dimensional representation of the flow pattern, geomorphic features and deposits characterizing a meandering river. Modified from Allen (1986).

The fluid flow within a curve channel is essentially helical. Due to this helical motion water moves down the outer bank and erodes it whereas it deposits sediments on the point bar surface while moving along it. As a result, an asymmetric channel cross-section develops with a steep outer bank and a gentle inner bank (Fig. 4). From one bend to the next, the sense of asymmetry changes and the direction of helical flow also reverses. The phases of accretion of the point bars are preserved as accretionary beds commonly referred to as epsilon cross-bedding (Fig. 4).

Assuming that the meander loops are semicircular in plan, disposed symmetrically about the meander axis and growing in unison, the sinuosity values for a meandering stretch have been shown to range between 1.36 and 5.24 (le Roux, 1992; Ghosh, 2000). Simulation of a more realistic pattern of meander evolution indicates that the median sinuosity value is around 3.14 (Stolum, 1996).

It has been observed that in natural meandering and braided rivers, the depth of the channel also varies systematically along its length producing a series of shallows and deeps commonly referred to as riffles and pools respectively (Fig. 3). Riffle-pool structures probably arise due to some autogenic erosional and depositional processes (Clifford, 1993). The spacing between successive pools is found to be related to channel width and is on average 5 to 7 times the latter (Keller & Melhorn, 1978).

The channel floors of meandering rivers are usually covered with loose sandy material that responds to the overlying flow by developing bedforms, usually dunes. Higher up on the point bar surface the scale of the bedforms becomes smaller in response to decreasing depth and boundary shear stress and larger dunes are replaced with smaller dunes or ripples. The coarsest material is often left in the deepest parts of the channels (thalweg) as lag deposits. In case of mixed-load channels the pauses in point bar accretion is marked by muddy layers resulting in inclined sand-mud heterolithic stratification (Thomas et al., 1987). In a sinuous channel with well-developed point bars, the majority of the bedforms are oriented parallel to the flow along the curved channel path, but the cross-channel component of the helical flow produces bedforms that move transverse to local channel flow. These bedforms migrate up the point bars, producing ridges parallel to the contours on the point bar surfaces and are called scroll bar (Fig. 4). A vertical section through point bars, therefore, should ideally reveal a basal coarse-lag deposit, larger cross-stratification at the lower part, a thinning of strata-sets up the profile coupled with reduction in grain size (Fig. 4). Muddy accretionary surfaces oriented orthogonal to the mean azimuths of the cross-strata can also be recognised within the point bar deposit. However, in reality point bar deposits often show considerable departures from the idealized pattern described above depending upon the geometry, grain size and accretion style of the point bars (Jackson, 1975; Bluck, 1971; Bridge, 1996).

Braided rivers : These are characterised by multiple channels of different order with repeated branching and rejoining pattern and when looked from the air reveal the intertwining effect of a braid (Fig. 2; Rust, 1978; Bristow & Best, 1993). Usually braided rivers are characterised by higher discharge fluctuation, slope and bedload (Schumm, 1977). The main feature that typifies braided rivers is the presence of a large number of bars that divide the low stage flow in a number of smaller order channels. At high flow stage most of the bars are submerged and the river takes the form of a single channelised flow (Thorn et al. 1993). The first order channels usually have a low sinuosity. Unlike the point bars associated with meandering channels, the braid bars are not necessarily bank-attached and usually accrete along or slightly oblique to the flow. In curved channel segments braid bars may also develop lateral accretion surfaces that unlike point bars occur as paired sets on both sides of the braid bar denoting mid-channel position (Bridge et al, 1986; Bristow & Best, 1993).

A quantitative index of braiding has been developed which measures the number of braids per mean curved channel length. Recently Friend and Sinha (1993) have defined a new quantitative parameter for braiding index, the braid channel ratio (B), which is defined as :

$$B = \frac{\textit{Sum of mid-channel lengths of all channels in a reach}}{\textit{Mid-line length of the widest channel in the reach}}$$

Coarse-grained braided streams develop either plane beds or dunes within gravels whereas dunes are the dominant bedform within sandy braided rivers (Hein & Walker, 1977; Bluck, 1976; Collinson, 1970; Cant & Walker, 1978). Rapid flow stage fluctuation in both the settings is recorded by different types of erosion and reactivation surfaces and patches of fine-grained sediments.

Anastomosed rivers : Anastomosed river is a variety of multiple channel systems, and consists of channels of similar order that repeatedly branch and rejoin around stable, usually vegetated islands which are several times larger in dimension than that of the individual channels (Fig. 2; Collinson, 1976). The distinction of anastomosed rivers from the multiple-channel braided rivers lies in the fact that 1) the braided rivers are confined within a single pair of floodplain whereas in anastomosed rivers large stretches of islands separating the individual channels act as the floodplains; 2) the mid-channel bars in braided rivers are submerged during bankful discharge but the stable islands in anastomosed rivers are much larger in size and are not submerged during bankful discharge; 3) the individual channels and mid-channel bars in a braided river shift their position frequently in contrast to the stable channels and dividing islands of anastomosed rivers, and 4) in a braided system the channels are of different orders whereas that in an anastomosed system are essentially of the same order.

Anastomosed channels are characterised by low-gradient, stable bank and high, fine-grained sediment load and are typical of low-gradient muddy alluvial plains such as inland drainage basin of Australia, temperate wetlands of Canada, Gangetic delta plains etc.

Contrary to common belief, studies on modern anastomosed systems have revealed that vegetation, climate and grain size of the bedload are not the major controlling factors of river anastomosis (Nanson & Knighton, 1996). Stability of individual channels is rather attributed to the cohesion of fine-grained (usually clay-silt) bank material and moderate discharge that is unable to erode the cohesive bank (Nanson & Knighton, 1996; Gibling, 1998). Nanson & Knighton (1996) postulated that anastomsed river channels, being unable to increase gradient or widen their flow path by eroding the banks, tend to divide themselves into multiple anabranches to increase their ability to transport sediment and water.

Individual channels within the anastomosing river may be straight, sinuous or braided (Smith & Smith, 1980; Gibling et al., 1999). Thus, features typical of braided or meandering rivers may characterize individual channel-fills. However, as channels are laterally stable, the accretionary deposit that forms on the flanks of bars are of limited lateral extent (Gibling et al., 1999). Close association with well-developed vegetated floodplain, floodplain lakes and levees (see later section) appears typical of anastomosed river systems of humid region (Smith et al., 1989; Nadon, 1994). On the other hand, in arid regions, anastomosed channels are usually associated with dessicated, pedoturbated or aeolian dune-covered overbank areas (Gibling et al., 1999).

For modern rivers slope-vs-discharge diagram shows distinctive field for braided, meandering and anastomosed rivers (Leopold & Wolman, 1957; Nanson & Knighton, 1996). But such discriminant diagrams are unavailable for the analysis of the ancient alluvial deposits. In the preceeding discussion we have, therefore, emphasised that the major river types not only differ in their planform pattern but are also associated with distinctive geomorphic features. Records of these geomorphic features, if preserved, provides the clue to successful recognition of and discrimination between the deposits left by ancient braided, meandering and anastomosed rivers.

Overbank system

Sedimentation processes : Overbank areas, more commonly referred to as alluvial floodplain, are the domain bordering the channel-belt, suffering inundation during floods (Fig. 1). The floodplains may be narrow, nearly as wide as the channel-belt or may be wide, many times wider than the channel-belt. Maximum width of the Brahmaputra floodplain is about 20 Km. (Coleman, 1969). Broad low-lying areas of the flood plain may show extensive mudflats, marshes, and small lakes including ox-bow lakes.

Overbank flooding is the major process that introduces sediments in the floodplain. Fine-grained sand, silt, clay diffused from the main channel during flood events are the dominant sediment types in the floodplain including reworked mud-pellets, pedogenic nodules and vegetal matter. As the flow overtops the riverbank, a rapid deceleration of flow and reduction in turbulence forces the diffused sediments to be deposited in the overbank areas. Overbank sediments are deposited as levees on the edge of the channel-bank, from crevasse channels spurting out breaching levees, or as splays. During catastrophic floods, however, the entire alluvial plain is covered by sheet flow (McKee et al.1967). Such overbank sheet floods leave a record of dominantly parallel laminated sheet-like sediment body. Sediments also accrete in small channels and gutters, characteristic of larger floodplains (flood basin), which drain water within the floodplain. The sediments of arid region floodplains may be reworked into aeolian dunes and sand sheets. Growth of vegetation and soil forming processes are the principal natural agents that modify the primary sediments of the overbank areas (see later).

The average rate of floodplain sedimentation is of the order of a few mms to a few cms per major flood (Bridge & Leeder, 1979). Sedimentation rate and grain size decrease exponentially away from the channel (Marriot 1996). The coarsest sediments are deposited close to the riverbank as levees and the distal parts of the floodplain receive finer-grained silt-clay grade sediments.

The differential sedimentation between areas close to the channel and distal floodplain together with higher compactional ability of silt-clay deposits results into development of alluvial ridges comprising the channel and the adjacent levee-crevasse complex, that stand at a higher elevation (few decimeters to meters) than the surrounding floodplain (Fig. 1). As a consequence, the cross-channel gradient may eventually become higher than the along-channel gradient inducing diversion of the channel along the favourable gradient- a process called avulsion that is described in detail later.

Levees form as wedge-shaped bodies internally showing gently inclined strata that become finer-grained away from the channels (Brierly et al., 1997). The strata usually comprise parallel lamination overlain by ripple-lamination and is topped by mudstone (Singh, 1972). Levee deposits laterally intertongue with or grade into channel-deposits. Exposure of the levees following flood leads to development of desiccation and bioturbation features within the deposit.

Crevasse splay deposit usually forms wedge-shaped, convex-up or sheet-like sand bodies showing evidences of decelerating flow (Fig. 4), and are frequently associated with small channels (Fig. 4).

Extensive parallel laminated sandstone with parting lineation has been recognised from the alluvial plains and has been inferred to record deposition from catastrophic sheetflood that covered the entire alluvial plain (Tunbridge, 1981).

Soil forming processes : The floodplain areas of a river remain exposed and escape active sedimentation or erosion for considerably long periods. During these periods exposed sediments are subjected to the soil forming processes. The soil forming process is a collection of physical, chemical and biologic processes that operate usually in the vadose zone. Prolonged period of pedogenesis olter the original sediments into soil which is structurally, texturally, mineralogically distinct from the original material.

The downward percolation of precipitated meteoric water from the surface downward acts as a major agent for the soil formation. The passage of meteoric water dissolves away soluble components and washes away finer clastics from the upper part of the host sediments. Components that are chemically and physically relatively immobile are left behind. Continued uptake of solutes along with the evaporative loss of water in the vadose zone may cause supersaturation at a certain depth. In such a situation, solutes are precipitated usually as authigenic carbonate, silicate and iron-rich minerals, in the pores of the host sediment or as grain coats. Suspended clastics carried down from the upper parts are also deposited similarly. This mechanism produces three distinct horizons in a soil. An upper horizon, from which material has been removed (known as the horizon of illuviation and is often called A-horizon). A middle horizon, where new materials have accumulated (horizon of alluviation or B-horizon) and a lower horizon of unmodified or barely modified host sediment(s) (C-horizon). Thickness and character of the individual horizons are largely dependent on the nature of the host sediment, prevailing climate and geomorphology of the terrain.

A-horizon is characterised by dissolution features and light/drab colours produced due to the removal of clay/silt grade materials from these horizons. The fines removed from the A-horizon are usually translocated to the B-horizon which causes B-horizons to be usually darker than A. The translocated material initially fill up the available pore spaces in the host sediment. As the pedogenic process continues, the porosity of B-horizon gradually decreases. After a time the original sedimentary fabric of the host is disrupted to accommodate the additional pedogenic material. At this stage the framework grains loose contact with the adjacent grains allowing the formation of grain coats by authigenic minerals and translocated fines brought from the A-horizon. With further pedogenesis the pedogenic materials starts to cluster within the B-horizon. These clots are commonly called pedogenic concretions or nodules or glaebules. Glaebules are prolate, discoid or equant bodies with highly irregular surfaces. The size of the glaebule is dependent on the intensity and the duration of the pedogenic modification and commonly varies between less than a mm to tens of cm. In the advance stage of pedogenesis adjacent glaebules may coalesce with each other. At this stage the porosity of the B-horizon is so low that the vertical movement of the water is hindered. Percolating meteoric water then tends to spread laterally after reaching the top of the B-horizon. The lateral spread of the solute-laden water produces discoid platy glaebules at the junction between the A- and the B-horizon. Continuation of this process eventually leads to formation of a new horizon characterised by coalesced platy glaebules. Pedogenically derived material may constitute more than 90% of such a horizon and is referred to as hardpan. If the authigenic component is chiefly carbonate the hardpan horizon is called a K or petrocalcic horizon. The hardpan horizon grows in an upward direction at the expense of the A-horizon.

Parts of a soil positioned close to the groundwater table or a zone of temporary, local water logging may show coloured blotches of pale yellow, grey and drab green. This feature is known as gley. In the vadose zone usually an oxidizing environment prevails, but in waterlogged parts the chemical environment may be reducing leading to reduction of iron-rich minerals into the ferrous state. Ferrous oxides impart a green or greenish grey colour to the soil. Such gleyed horizons are common in the low-lying areas of the floodplain where water table is perennially high. The soils dominantly comprising of the gleyed horizons are usually referred to as hydromorphic soils (Duchaufour, 1982). Gleys are often rich in preserved organic matter and peats are considered to be a variety of organic-rich gleyed soils.

Apart from the role of percolating meteoric water, plant rooting is the most important cause for pedogenic modifications. The mechanical penetration of the roots and rootlets causes structural disruption of the host fabric. Also, the respiration of the living roots produces a chemical microenvironment around them, which makes them more suitable sites for precipitation. For example the increased CO_2 concentration produced by root respiration causes supersaturation of carbonates around the roots. Precipitated carbonates thus form a sheath around the living roots and these are called rhizocretions. Moreover, the tubular voids left behind by the decay of dead roots serve as the conduits for percolation of meteoric water deeper into the soil profile.

Swelling of soils in response to the expansion of clays on wetting, and subsequent shrinkage on drying produces surfaces of brittle failure. Repeated wetting and drying of soils developed on fine-grained host produces arcuate, inclined and mutually intersecting slickenside surfaces that is typical of the upper parts of many soil profiles. The development of ice wedges, desiccation and burrowing are the other processes of structural modifications of the host during pedogenesis.

Formation of soil requires that there should not be any erosion and sedimentation at the site during pedogenesis. However, soils may also form in aggradational systems like alluvial plains as long as the rate of sedimentation does not overwhelm the rate of pedogenesis. At favourable sites a number of soil profiles may be stacked one upon another. The thickness of different soil horizons and their stacking pattern is largely dependent on the relative rates of sedimentation and pedogenesis. Compound soil profiles form when sedimentation rate is higher and a layer of unmodified sediment separates two weakly developed soil profiles. If the rate of pedogenesis exceeds that of sedimentation successive soil profiles may overlap giving rise to compound profiles. If sedimentation and pedogenesis go hand in hand, a thick single soil horizon may develop and is commonly referred to as cumulative soil profile.

ANCIENT FLUVIAL DEPOSITS

Based on the discussion in the previous section, we would now attempt to enumerate different criteria for recognition of ancient fluvial systems. We shall first discuss some of the general features of alluvial deposits and then proceed to identify more specific records of different types of river systems (viz. meandering, braided, anastomosed etc.). It should be kept in mind that our observation of ancient deposits is seriously constrained by the preservational bias of the depositional system and availability of suitable outcrop and sub-surface data. For example, gradual migration of channels over the floodplain obliterates deposits of the floodplain. Similarly, in outcrop ridges oriented parallel to the paleo-flow direction reconstruction of the cross-flow profiles of channels becomes difficult.

A succession of sedimentary strata bearing evidences of subaerial deposition and characterised by fining-upward succession of sedimentary strata overlying concave-upwards erosional surfaces, unimodal paleocurrent pattern of meso-scale bedforms, poor sorting and compositional immaturity can be confidently inferred as fluvial deposits. Fossil content, geochemical signatures, stratigraphic architecture and geotectonic setting may provide important clue to the recognition of alluvial deposits. However, isolated occurrence of any one of these features (e.g. a fining-upward succession) is not diagnostic and absence of any one of the above criteria (e.g. lack of channel form erosion surfaces) do not negate the strength of other evidences for fluvial interpretation.

Once a sedimentary deposit is identified as of fluvial origin, it should be further examined for more detailed reconstruction of the ancient fluvial system. Interpretation of the ancient channel pattern involves three-dimensional reconstruction of different orders of fluvial bedforms or "architectural elements", their patterns of organization in the deposit and the paleocurrent pattern ("Architectural Element Analysis" of Miall, 1985, 1988).

Three principal architectural elements of alluvial deposits are microform, mesoform and macroform. Microforms include ripples and plane beds and are the products of shallowest and weakest of the flows in the context of the alluvial channel and forms mostly in the highest topographic levels (higher parts of the bars or in shallow channels dissecting exposed bars during falling river stage). Dunes are regarded as mesoforms, which are scaled to local flow depth and velocity. Macroforms are channel-scale features with their dimensions related to the size and depth of the main channel, and are essentially compound bedforms or bars upon which micro and/or mesoforms are superimposed (Collinson, 1970; Banks, 1973; Allen, 1983; Miall, 1985). Accretion of the macroform produces inclined strata termed macroform strata that internally consist of micro and/or mesoform strata. The azimuths of the macroform strata may be orthogonal, oblique or parallel to that of the associated micro and mesoform strata. The first type of compound stratification is commonly known as epsilon cross-stratification or lateral accretion bedding and the parent macroform is termed laterally accreting bars e.g. point bar. The third type of compound cross-stratification popularly goes by the name of downcurrent-dipping cross-stratification and the associated macroform is called foreset macroform or downstream accreting bars.

Another type of macroform of the fluvial system is channel that is filled dominantly by vertical accretion producing plano-concave sediment bodies. Channels may be of different orders and are in many cases filled with sandy sediments; however, conglomeratic as well as muddy channel fills are also common. The principal diagnostic feature of channel-fill deposit is the concave-up, erosional contact with the underlying beds. The channel-fill sediment bodies produced by vertical accretion usually acquire the channel morphology and in transverse section the channel-form can be observed. The channel-fill may be a) massive, b) parallely stratified either horizontally or conformal to the cross-sectional channel geometry, c) stacked sets of massive strata, cross-strata and horizontal strata.

The different architectural elements produce different orders of bounding surfaces within a channel-belt sediment body (Fig. 5; Miall, 1988, 1995). The first order surface represents the boundaries between sets of micro/mesoform strata, whereas the boundaries of the cosets are of second order. Third order surfaces bound each macroform stratum. The boundaries of macroform bodies are fourth order surfaces. The fifth order surface represents the boundary between channel-belt and overbank deposits or between two different channel-belts.

A channel-belt sediment body is produced by superimposition of bar and channel macroforms

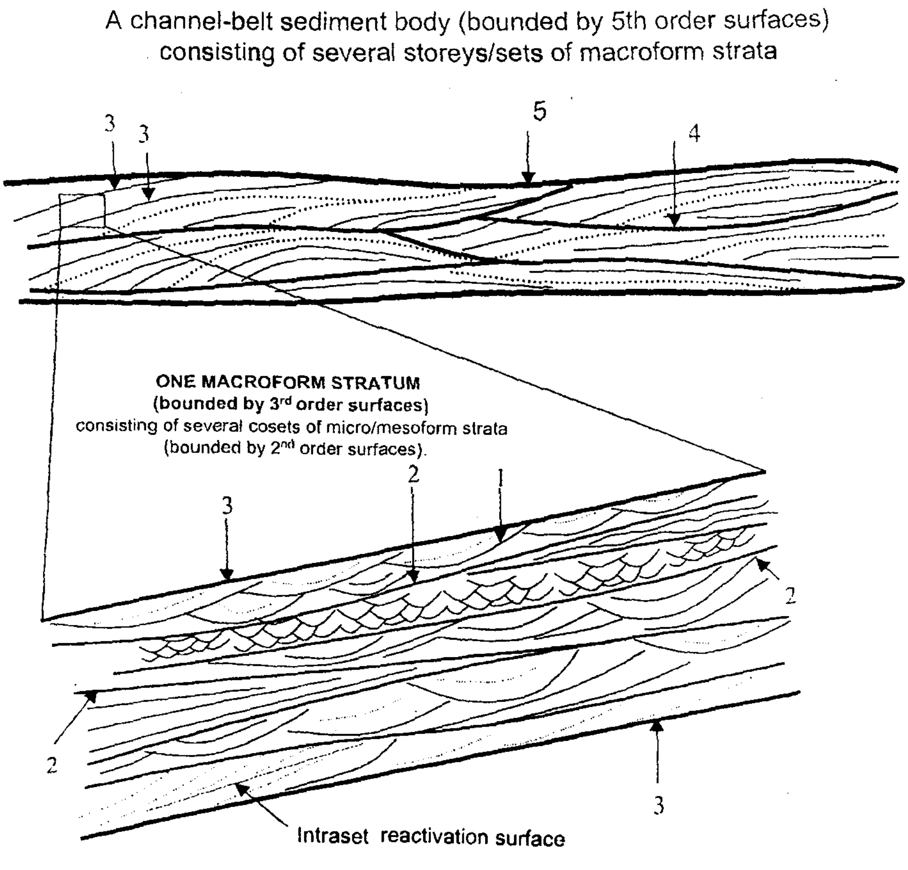

Fig. 5 Internal architecture of a channel-belt sediment body. Note different orders of bounding surfaces that may be present within the body.

(Fig. 5). The building blocks of ancient channel-belt sediment bodies are storeys; a storey is defined as the deposit of a single channel-bar and adjacent channel-fill and comprise of sets of macroform stratification (Bridge, 1993). A channel-belt deposit may consist of single or multiple storeys. Multistorey channel-belt deposit results due to migration or switching of individual channels within an aggrading channel-belt.

It is useful to describe the three-dimensional shapes of alluvial sediment bodies and their appearances in differently oriented vertical sections (Bridge, 1993). The convention is to describe the upper bounding surface first, followed by the lower bounding surface; for example plano-concave or plano-convex etc.

Channels within a river move downward (vertical incision) or upward (aggradation) in response to rise or fall of the base level respectively. They may also shift laterally (migrate) with or without a vertical component. The width of a channel-belt sediment body approximates the lateral extent of the alluvial plain within which the channels could change position by lateral shifting. The thickness, on the other hand, depends on the depth of channels and amount of aggradation and incision. The width:depth ratio of the braided rivers is usually higher than other types of channels and consequently channel-belt deposit of a braided river is sheet-like. Again, a meandering channel-belt would tend to produce sheet-like deposit if it is free to migrate laterally (Collinson, 1978), in contrast to ribbon-like channel-belt bodies produced by less-mobile meandering rivers or fixed channels of an anastomosed channel-belt (Bridge and Leeder, 1979).

In order to arrive at a comprehensive understanding of an ancient fluvial system, it is not only necessary to know the regional paleocurrent pattern but we also need to document the variation of paleocurrents between storeys and the relationships of the paleocurrent azimuths with that of the associated macroform strata.

In order to reveal the bounding surface geometry, paleocurrent pattern and shape of channel-belt sediment bodies, it is necessary to examine large, three-dimensional outcrops or closely spaced borehole logs. Combining the architectural elements in different ways Miall (1985) proposed twelve generalised models for fluvial deposits. It should be noted that this approach of architectural element analysis has thoroughly replaced the traditional approach of studying fluvial deposits based on analysis of vertical profiles alone. Most of the fluvial deposits develop a fining upward succession, but they vary remarkably with regard to nature and distribution of different types of architectural elements. Vertical profile analysis, being unable to recognize this three-dimensional distribution of different architectural elements and their lateral variability may lead to incorrect inferences (Allen, 1983; Miall, 1985; Collinson, 1996).

Meandering river deposits

The ancient record of meandering rivers is characterised by a fining-upward succession with

Fig. 6 Field photograph of a point bar deposit showing spectacular lateral accretion bedding defined by alternating layers of sandstone and mudstone. The sandstone layers comprise of ripple-lamination with paleocurrents normal to the plane of the section. Note several reactivation surfaces within the point bar deposit.

a flat lower bounding surface, usually marked by lag conglomerate, followed upward by lateral accretion deposits that grade-upward into fine-grained flood-plain sediments (Fig. 4; Allen, 1965; Puigdefabregas and Van Vliet, 1978; Smith, 1987). The lateral accretion strata are expressed either by surfaces, bounding sets or cosets of cross-strata or by alternate layers of sandstone and mudstone or siltstone (Fig. 6). It is, however, essential to establish a cross-flow direction of accretion of these inclined surfaces through measurements from interlayers/associated cross strata to assign a point bar origin. One gross measure of the higher sinuosity (> 1.5) of the paleochannels is the higher dispersion of the paleocurrent vectors measured preferably from meso-scale bedforms. Since smaller bedforms show higher dispersion (Miall 1974) compared to the mesoscale bedforms, clubbing of data from all scales of bedforms tend to overestimate the dispersion value. However, more precise estimate of paleosinuosity is now possible by applying methods developed by Ghosh (2000) (see below).

Another indication of the high sinuosity of paleochannels is development of scroll bars. Curvilinear, ridge and swale topography preserved on point-bar surfaces (Sundborg, 1956) represent scroll bars and mimic the edges of point bars at different instants. These ridges therefore represent the curvature of the channels flanking the point bars. Similar features have also been reported from ancient alluvial deposits (Puigdefabregas & Van Vliet 1978; Alexandar 1992; Collinson, 1996).

The geometry of point bars and the associated accretionary beds may be varied. Lateral accretionary beds may cover the entire point bar surface upto the bottom of the channel or may represent accretion only in the upper parts of the point bar. The latter is inferred to indicate perennially submerged condition in the lower part of the channel preventing finer grains from accumulating on the accretion surfaces. Accretion surfaces can be broken by benches or erosional scours reflecting multitier topography of the bar. Sets of lateral accretion strata may be separated by low-angle discordances (Fig. 6) and attributed to medium-scale stage fluctuation or slight change in accretion direction of the point bars (Maulik et al., 2000).

Isolated, laterally restricted sandbodies often show lateral accretion surfaces that terminate against clay-plugs (Nami & Leeder, 1978). The clay plugs denote eventual abandonment of the channel when fine-grained sediments settled from stagnant water to fill up the channel.

Knowledge gained through the study of modern meandering streams (Sundborg, 1956; Bridge 1977, 1992; Jackson, 1976; Bridge and Jarvis, 1976, 1982) has now allowed quantitative modelling of the point bar deposit. Several attempts have been made to quantitatively reconstruct the paleochannel characteristics and paleohydrology from the preserved point bars (Bridge & Diemer, 1983; Bridge & Gordon, 1985). Assuming that the preserved bed topography and bedforms to be in equilibrium with the flow, these paleohydraulic studies utilise the physical model of equilibrium flow pattern at modern channel bends to estimate the depth, width, discharge, mean velocity and sinuosity of paleochannels. Willis (1993) has run numerical experiments to reconstruct the three-dimensional geometry of the point bars and have developed techniques to reconstruct the three-dimensional geometry of the bars from two dimensional outcrop sections of the point bar deposits.

Several attempts have been made to understand the flow characteristics of meandering and braided rivers and to relate them to the orientation of the bedforms (Rust, 1972; Schwartz, 1978; Bluck, 1971, 1974; LeRoux, 1992). Systematic measurement of cross-strata from the surface of braid and point bars of Ganga River show a considerable variation in the paleocurrent pattern between braid and meander (point) bars (Shukla et al., 1999). Whereas the braid bars are characterised by unimodal paleocurrent pattern with high variance, point bars in the Ganga river show unimodal to polymodal distribution of paleocurrent azimuths. Moreover, this study reveals that there is considerable variation in the bedform orientation between upstream, central and downstream part of the bars.

Recently, sophisticated quantitative models have been developed to estimate channel sinuosity from the bedform orientations (Ghosh, 2000). This work utilizes a realistic model for point bar migration and fractal channel geometry to arrive at a quantitative relationship between flow vectors and channel sinuosity (Fig. 7). The uniqueness of the model is that it is found to estimate fairly accurately the sinuosity of the channels from both modern and ancient record and can be utilised as an independent tool for characterisation of the fluvial deposit in a braided or meandering model.

Lateral accretion deposits are not unique to meandering rivers. Migration of curved channel bends flanking mid-channel bars in low-sinuosity braided rivers can result in lateral accretion bedding (Bridge et al. 1986, Bridge, 1993). Isolated occurrences of lateral accretion deposits have been reported from successions that overall fit in a braided fluvial depositional model (Allen, 1983; Bristow, 1993). Similarly mid-channel bars do occur within sinuous meandering streams. Therefore, interpretation of a meandering channel pattern from ancient deposit should be done with circumspection.

Braided river deposits

The principal characteristics of the braided river deposits are coarse-sandy or gravelly downcurrent

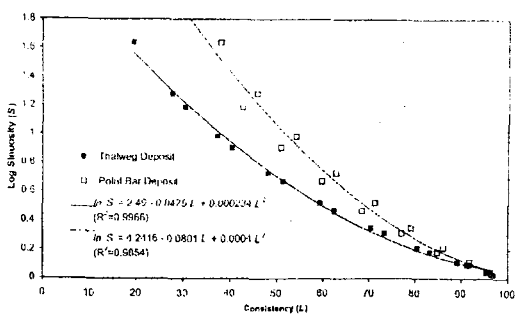

Fig. 7 Relationship between consistency and sinuosity of paleocurrent data measured from the different parts channel thalweg and bars. After Ghosh (2000).

accreting macroform strata, presence of numerous channel-fills of different dimensions, thin impersistent units of floodplain fines and a comparatively low spread of paleocurrent pattern. Study of modern braided rivers indicate that low-stage flow on emergent bars is highly variable (Bluck, 1976) and caution should be exercised in assessing dispersion of paleocurrent azimuths, and it is recommended that mesoforms, rather than microforms, be used during paleocurrent measurement. In contrast to meandering rivers, braid bars do not always produce a fining-upward grain size trend. On the contrary, winnowing action by shallow flows in the emergent bar tops may produce a coarsening upward trend (Bluck, 1976; Haszeldine, 1983; Bristow & Best, 1993). In spite of the fact that lateral accretion is common in many modern braid bars (Bridge et al, 1986; Bristow & Best, 1993), many of the braid-bars show down-current or oblique accretion with respect to local channel flow (DA architectural element). Large braid bars often referred to as sandflats, grow through many floods and have a complex erosional and depositional history. Thus the sandflat deposits may lack a simple fining- and thinning-upward organisation displayed by the point bar deposits.

In the Devonian Battery Point Formation, Canada, Cant and Walker (1976) observed occurrences of isolated, large cross-strata oriented oblique to that of the medium to small-scale cross-strata (Fig. 8). Their subsequent study in the South Sasketchewan braided river (Cant & Walker, 1978) showed that large 2-D transverse bedforms nucleate as cross-channel bars at zones of flow convergence. Nucleation of large cross-channel oriented bedforms inhibits flow further, attracting more bedforms to grow around the nucleus gradually leading to a large bar-complex or sandflats.

Through careful analysis of large outcrop sections of Carboniferous coal-measure sandstones, Haszeldine (1983 a,b) revealed the complex accretionary history of large (~10 m. thick) braid bars. A hierarchy of bedforms migrated downcurrent along

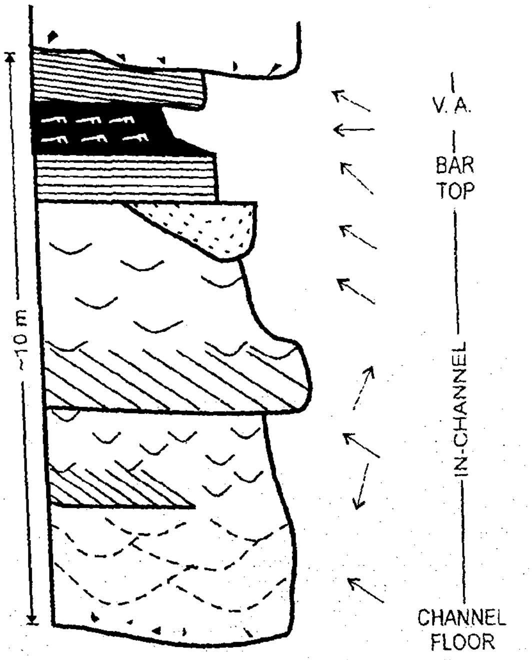

Fig. 8 Vertical profile of an ancient braid-bar deposit. Adopted from Cant and Walker (1978).

the gentle lee-slopes of the large in-channel braid bars resulting in cosets of downcurrent dipping cross-strata. Locally within this bar succession concentration of coarse granules, inferred to have been deposited through winnowing action of bar-top sediments at times of low-stage flow, gave rise to a coarsening-up trend (Haszeldine, 1983 a,b).

Thick and laterally extensive fluvial deposits have been recognized in Precambrian successions all over the globe (see Eriksson et al., 1998 for a review). Schumm (1972) postulated that most of the pre-Silurian fluvial systems were likely to be braided in nature due to absence of bank-stabilising and discharge-moderating land vegetation. Theoretical considerations by Fuller (1985) predicted that pre-Silurian streams might have very high width/depth ratio and may be confused with shallow marine deposits. Most of the Precambrian fluvial deposits are characterized by sandy or gravelly sediments, shallow-wide channels filled with sandy bedforms, lack or paucity of fine-grained floodplain deposits and close interaction with aeolian-processes (McCormick & Grotzinger, 1993; Fedo & Cooper 1990; Dott et al 1986; Roe & Hermansen 1993, Chakraborty & Chaudhuri, 1993; Chakraborty, 1996). Fine-grained sediments, marked by desiccation features, become more abundant as the gradient becomes gentler in the distal basinal setting (Winston, 1978; Sonderholm & Tirsgaard, 1998). Bars (down-current accreting macroforms), that characterise channel-belt sandbodies in most of the Phanerozoic braided river deposits (Hazeldine, 1983a; Bristow, 1993), are in many cases either absent (McCormick & Grotzinger, 1993; Fedo & Cooper, 1990) or when present are shallow ones (Roe & Hermansen, 1993; Chakraborty, 1999; cf., Bluck, 1976) within the Proterozoic alluvial deposits. However, there are reports of development of Proterozoic braidplains upto 15 km wide with channels about 10 m deep that were filled by large down-current accreting macroforms or sandflats (Rainbird, 1992). Floodplains though rare, when present may be sandy, interlayered with aeolian sediments and/or lateritic paleosols (Chakraborty, 1997; Eriksson et al., 1998). Due to absence of vegetation the Precambrian rivers experienced flashy discharge regime and sediment-gravity flow was likely to be more common in the Precambrian alluvial deposits (Eriksson et al., 1998).

Anastomosed river deposits

Anastomosed rivers are reported from humid (Smith & Smith, 1980) as well as arid climatic regimes (Rust 1981). However, the humid region model is largely applied to ancient deposits showing evidence of thick, narrow bodies of coarser-grained channel-fill deposits, flanked by crevasse splay lobes encased within extensive fine-grained floodplain, back swamp, and coal-accumulating peat-swamp deposits. Since anastomosed channels are laterally stable, they are likely to produce thick, narrow, shoestring sandbodies encased in mud (Smith & Putnam, 1980). However, in outcrops it is difficult to decipher the shoe-string geometry and most of the studied examples show isolated channel sandbodies seperated by floodplain fines in two-dimensional sections (Kirshbaum & McCabe, 1992; Nadon, 1994).

Confusion about the anastomsed river sedimentation model arises mainly due to our inability to demonstrate the contemporaneity of different ribbon-shaped channel sandstone bodies in the stratigraphic record. A meandering stream can also have well-developed floodplains and the channel body might be relocated to different parts of the floodplain producing a succession similar to that believed to be produced by anastomosed channels (cf. Nadon, 1994). However, a study of the Holocene fluvial record of Rhine-Meuse Delta (central Netherland) where geologic and geomorphic mapping has been combined with the data from 1,80,000 shallow bore holes and about 1000 ^{14}C-dates of the deposit (Tornqvist, 1993) appears to provide clues to the solution of this confusion. Both meandering and anastomosing channels are recorded from the succession and it has been found that anastomosing rivers produce much narrower channel sandstone bodies (width/thickness ratio 8 to 15) as compared to those of the meandering rivers (width/thickness ratio typically 50-65, Tornqvist, 1993). Also, the

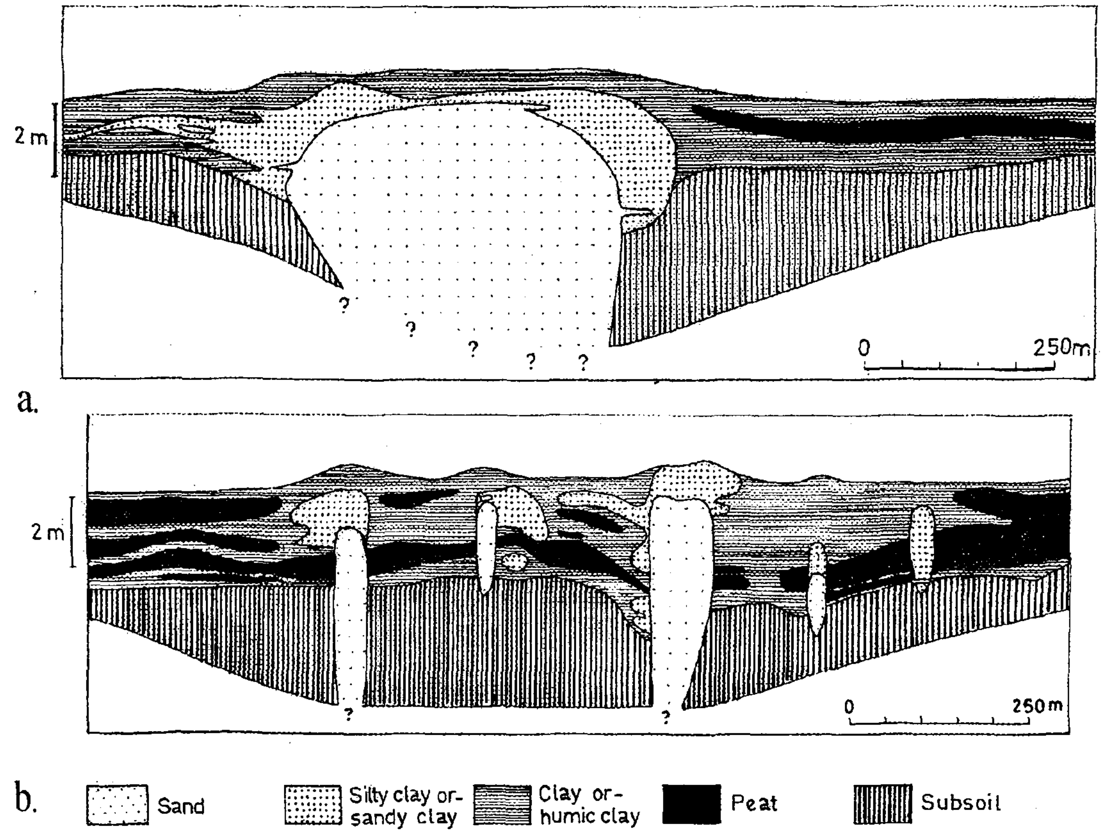

Fig. 9 Facies architecture in Holocene meander (a) and anastomosed (b) river deposits of the Rhine-Meuse Delta. Meander rivers are charactersied by wider channel deposits (sand) and extensive wedge-shaped levee deposits (sandy or silty clay). In contrast, anastomosed river deposits show narrow channel sands flanked by narrow levee complexes and numerous small crevasse channels. (Modified from Tornqvist, 2000).

›verbank levees are much broader for meandering ;tream that show a monotonic decrease in thickness ınd grain size away from the channels grading into lood-basin peat. The anastomosed channel systems, on the other hand, are characterised by much narrower levees and numerous crevasses (Fig. 9). The study reveals that the development of channel anastomosis is favoured by high rate of aggradation (1.5 mm/yr.) and presence of thick cohesive muddy subsoil (Tornqvist, 1993). Channel pattern in Rhine-Muese Delta seems to be less affected by changes in valley gradient. Nadon (1994) also argued that stratigraphic record of anastomosed channel system need not comprise vertical stacking of channel sandbodies (as visualised by Smith & Smith, 1980) because aggradation of channel-levee complex would invariably give rise to higher cross-valley slope eventually leading to avulsive relocation of channels in different locations within the floodplain. Nadon (1994) inferred the ribbon sandbodies with well-developed levee deposits of Late Cretaceous St. Mary River Formation represent deposition from anastomosed river.

Anastomosed rivers of Lake Eyre Basin, Australia and their sedimentary record bear broad similarity with the above mentioned model in terms of narrow ribbon-shaped channel body, cohesive muddy overbank sediments and somewhat subdued levee deposits flanking the channels (Gibling et al. 1999). However, the major differences include abundance of reworked mud-pellets, finer-grained sediment load, input from aeolian dunes and floodplain vertisols. Most of these differences are attributable to an arid climatic setting in the central Australian basin. It has however, been pointed out by Gibling et al. (1999) that certain features of anastomosed rivers have resemblance to single-thread, arid-region stream types and distinction between these two may be difficult in the rock record.

Overbank deposits

Sediments deposited in the vast area of the alluvial plain outside the domain of channels represent the overbank deposit. Bulk of the deposits is mudstone and siltstone. Red colouration of the mudstone usually reflect well-drained, oxidising condition, but greenish-grey colour or preservation of organic matter suggests high water table, water-logged condition and hence reducing nature of the pore water. The overbank strata include deposits of levee, crevasse splays and channels (described in an earlier section), small to medium sized lakes, peat-bog and marsh land, extensive muddy flats with features of subaerial exposure and intense pedogenesis or aeolian reworking.

Ephemeral to permanent lakes are abundant in alluvial plains, particularly associated with anastomosing and meandering channels. The typical lacustrine succession comprises thinly laminated mudstone and wave-reworked fine-grained sandstone (Collinson, 1996). Organic-rich mudstone, abundant root casts and hydromorphic soils are typical of the deposits associated with floodplain lakes. However, in drier climatic regime, the vegetal matter may be oxidised and evaporites and red mudstones may characterise floodplain ephemeral lakes. Small channels draining the floodplains may give rise to small lake-deltas producing coarsening upward succession (Tye & Coleman, 1989). Depending on the climatic conditions, these lake and marshland environments may support thick growth of vegetation facilitating formation of peat and coal deposits (Haszeldine, 1984).

Smith et al. (1989) have documented that development of large crevasse-splay bodies within floodplains are often related to relatively rapid diversion of major channels along newer paths - a process called avulsion. The processes and consequences of avulsion and the deposits left by them in the floodplain are discussed in the following section.

CHANNEL AVULSION, AVLUSION DEPOSITS AND AVULSION-CONTROLLED ALLUVIAL ARCHITECTURE

Avulsion and its deposits

Avulsion is a relatively rapid shift of a river in favour of a new course in the alluvial plain. This i entirely different from lateral switching of channels within a braided river tract and chute-, neck-cutoffs in meandering channels. Smith et al. (1989) have shown that channel-belt avulsion usually takes places through progressive enlargement and evolution of crevasse-splays (Slingestand & Smith, 1998) and the depositional record of the event preserved in the floodplain is known as avulsion deposits (Smith et al., 1989; Kraus and Wells, 1999). The process of avulsion is geologically instantaneous.

Mechanisms of avulsion : The fundamental cause of avulsion is the relative loss of sediment-fluid carrying capacity of a channel to such an extent that the existing path becomes unfavourlable for sustaining the flow, and the channel then finds a new course along a favourable gradient. The loss of capacity of a channel may take place in three ways:

1. An increase in the ratio of avulsion course slope and existing channel slope caused by the decreasing gradient of the existing channel (eg. Channel meandering, delta plain growth etc).

2. An increase in the ratio of avulsion course slope and existing channel slope caused by the increase in the gradient away from the existing channel (e.g. formation of raised alluvial ridges due exponential decrease in sedimentation rate across the flood plain, tectonic tilting of the valley floor etc.)

3. Non-slope related reduction in the capacity of the channel to carry water and sediment delivered to it (log or ice jam, change in the discharge, increase of sediment load etc).

The avulsions may be local where the new channel course converges with the old channel course in a down stream area, or regional where the avulsion channel occupies an entirely different path in the flood basin (Jones & Schumm 1999). Avulsion is called nodal, if successive avulsions take place from the same point in the alluvial plain and is termed random if divergence initiates at different points at different times (McKay & Bridge, 1995).

It should be understood that the instability of a channel (i.e. avulsion threshold) caused by the above factors does not necessarily induce avulsion unless a triggering event like flood or tectonic tremor occurs (Heller & Paola, 1996). It therefore, follows that avulsion frequency would depend on how the processes inducing channel instability interact with the triggering events (Jones & Schumm, 1999).

Avulsion deposits : Recognition of the importance of avulsion deposits in understanding the long-term behaviour of the fluvial system and its influence on the stratigraphic architecture, have recently motivated a number of investigations of the Holocene avulsion deposits (Morozova & Smith, 1999; Aslan & Blum, 1999). Most of these studies combine air-photo investigation of the abandoned channel-belt, data from shallow bore-holes and ^{14}C age data. Investigations suggest that average avulsion frequency varies from 600 to 1000 years and both instantaneous avulsion and gradual avulsion through development and enlargement of the crevasse-splay complex were common (Fig. 10). Increased rate of sea-level or lake-level rise appears to have acted as the primary cause of Holocene avulsion in these channel-belts. Periods of higher sea- or lake-level promoted avulsion through formation of the crevasse-splay complexes and development of multi-channel anastomosed river pattern. Stable lake- or sea-level promoted instantaneous avulsion through reoccupation of the older channel-belts; and the channel patterns were either single channel meandering or braided system (Morozova & Smith, 2000). Holocene avulsion deposits consist of fine sand or silt grade material and are laterally more extensive than channel-belt deposits and always occur below the coarser grained channel-levee deposits.

Kraus & Wells (1999) have identified avulsion deposits, which formed in a manner closely similar to that recorded from the Cumberland Marshes (Smith et al. 1989), from the Palaeogene alluvial deposits of the Bighorn Basin, Wyoming. Palaeogene avulsion deposits of the Big Horn Basin could be

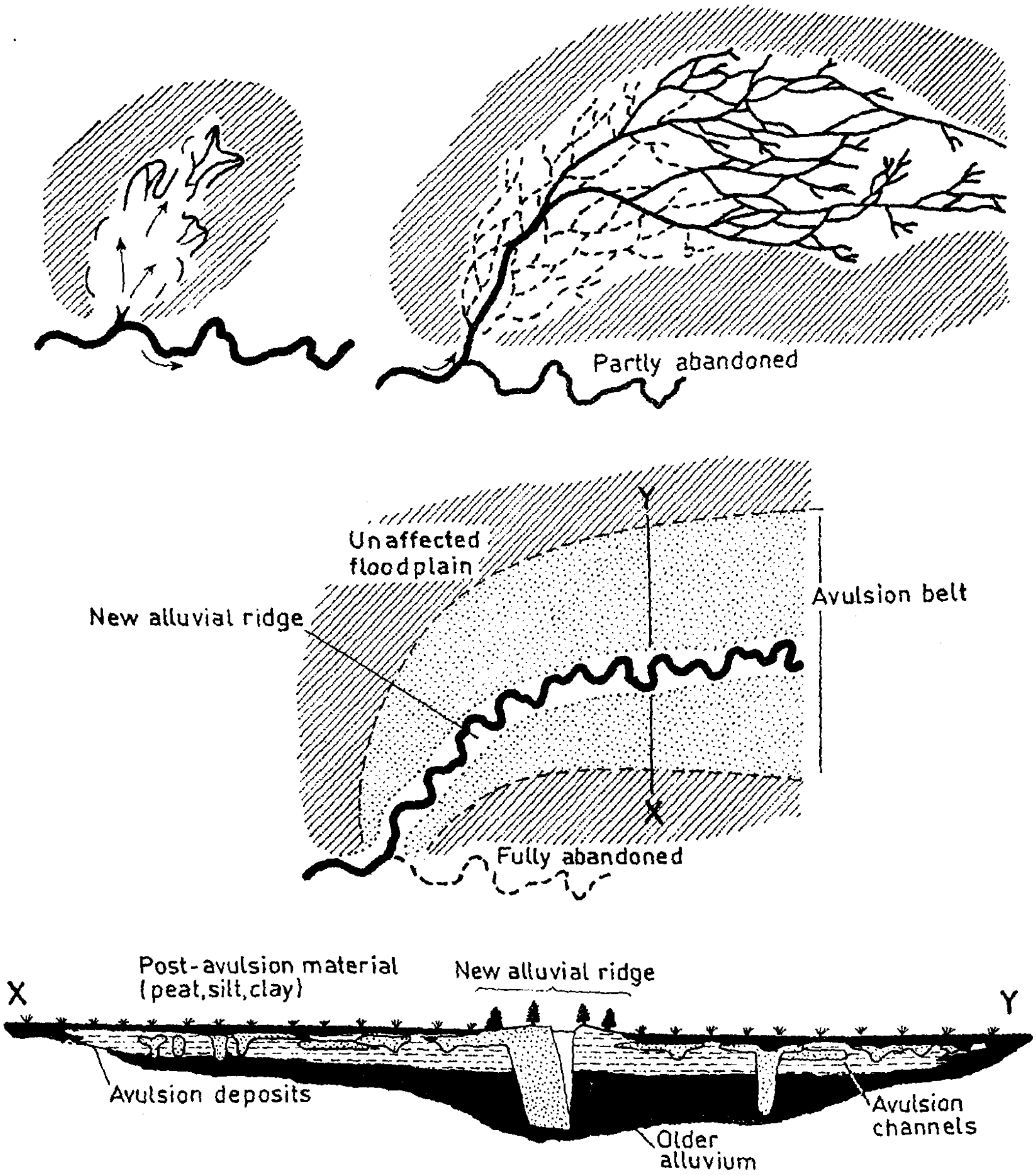

Fig. 10 Evolutionary pattern of a channel avulsion through the formation and enlargement of a crevasse lobe. Crevass initially form an extensive belt of avulsion deposit with a network of crevasse channels within the floodpla Smaller channels gradually evolve into a major channel and associated alluvial ridge till the next avulsior initiated. (Modified after Smith et al., 1989).

distinguished from the associated normal floodplain deposits using the following criteria: (1) large areal extent of these deposits, (2) the stratigraphic relationship of avulsion deposits with major channel-belt sandstones, which always overlie the avulsion related sediment packages rather than being laterally connected to them, (3) occurrence of peat or thick, mature paleosol deposits that over- and underlie the avulsion deposit but are not interlayered with them. Occurrence of cumulative or incipient paleosols in the avulsion deposits in contrast to thick paleosols or peat accumulations in the under and overlying floodplain deposits indicate high rate of floodplain sedimentation during avulsion.

Avulsion-controlled alluvial architecture

The term alluvial architecture refers to the geometry, proportion and latero-vertical distribution of different types of fluvial facies within the alluvial succession (Allen, 1978). A large number of computer models have been developed during the last two decades to simulate large-scale alluvial stratigraphy through known controls of sedimentation rate and frequency of avulsive relocation of the channel-belt (Allen, 1978; Leeder, 1978; Bridge and Leeder, 1978; Bridge & McKay, 1993a,b; McKay & Bridge, 1995; Heller & Paola, 1996). Most of the simulations develop two-dimensional models and are dominantly controlled by the differential sedimentation rate across the alluvial plain and a constant avulsion frequency. The recent simulation by McKay & Bridge (1995) develops a three dimensional model for the avulsive channel system and includes parameters for: (1) floodplain and channel-belt width, (2) bankful channel depth, (3) channel-belt and overbank sedimentation rate, (4) avulsion location and period, (5) compaction and (6) tectonism (tilting and faulting in alluvial plain).

One of the important findings of these simulations is that the proportion and connectedness of channel-belt sediment bodies relative to overbank deposits vary considerably with the rate of sedimentation at a constant avulsion frequency. If the spatially averaged sedimentation rate in the alluvial plain is assumed to proxy the rate of tectonically driven basin subsidence, it follows that the any observed vertical variation in the proportion and connectivity of channel-belt sediment bodies would reflect the varying rate of tectonic subsidence of the basin floor (Blakey & Gubitosa, 1984; Kraus & Middleton, 1987 among many others). A stratigraphic interval with isolated ribbon-shaped channel-belt sandstone bodies can then be co-related with periods of enhanced rate of basin subsidence and the intervals showing amalgamated channel-belt sediment bodies with periods of lower rate of basin subsidence. Whereas the above conclusion may be true in some cases, later studies show that a straight-forward correlation of alluvial architecture with basin subsidence rate is over simplified (Bryant et al., 1995; Heller and Paola, 1996). Internal and external controls of the alluvial architecture interact in a complex manner to produce contrasting alluvial architecture. Three-dimensional simulation model of McKay & Bridge (1995) shows that interconnectedness and proportion of channel-belt sandstone is a function of the distance from avulsion point. For the same sedimentation rate and valley slope, the sandstone bodies upstream from the avulsion point tend to have low width/thickness ratio and downstream of the avulsion point channel sandstone bodies tend to be interconnected resulting in the high width/thickness ratio. The interconnectedness of the sandstone bodies is critically dependent on the ratio of the channel-belt width (Wcb) and floodplain width (w). If the value of Wcb/w exceeds 0.75 all the channel sandbodies tend to be interconnected and thus have an amalgamed sheet-like geometry. Also, the nature of discharge (controlled by climate) as well as nature and amount of sediment input (controlled by tectonism and climate) can significantly alter the channel pattern and Wcb/w ratio resulting in significant change in the architecture of the resulting alluvial deposits (Bridge, 1996; Blum & Tornqvist, 2000).

ALLUVIAL SEQUENCE STRATIGRAPHY

The basic premise of sequence stratigraphy is to analyse a sedimentary succession in terms of he rates of accommodation space creation and filling. Accommodation space is the space between he basin floor and the base level surface and is generated due to relative rise of base level aided by subsidence (Fig. 11). In marine and lacustrine basins sea and lake level is the base level, whereas in alluvial basins the graded stream profile represents the base level. For ancient rock record, base level may be defined as '...the potential energy surface that describes the direction in which a stratigraphic system is likely to move, toward sedimentation and stratigraphic preservation or sediment bypass and erosion' (Barrel, 1917; Sloss, 1962; Wheeler, 1964). The accommodation space, available at any time in a sedimentary basin, is utilized by being progressively filled with sediment supplied to or produced in the basin. In marine and lacustrine basins the unfilled accommodation space, at any time, is represented by the water depth. Understandably, in alluvial basins the unfilled accommodation space is not apparent because the whole basin is subaerial (Fig. 11).

The sequence stratigraphers analyse development of a marine succession in relation to relative

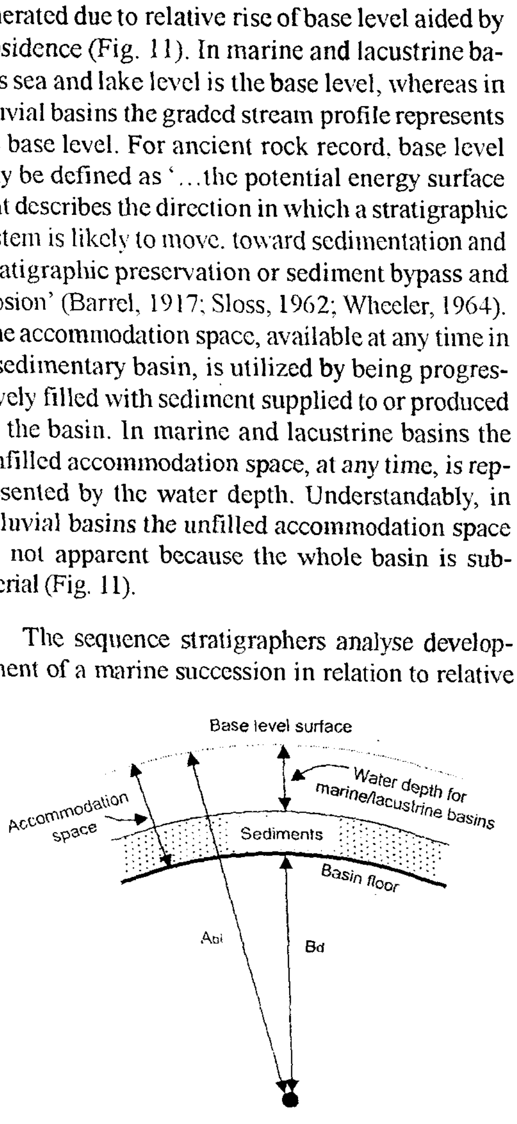

Fig. 11 Diagram illustrating definition of accommodation space in a sedimentary basin. Note that the space increases with increase in absolute base level (i.e increase in Abl) and subsidence (i.e. decrease in Bd).

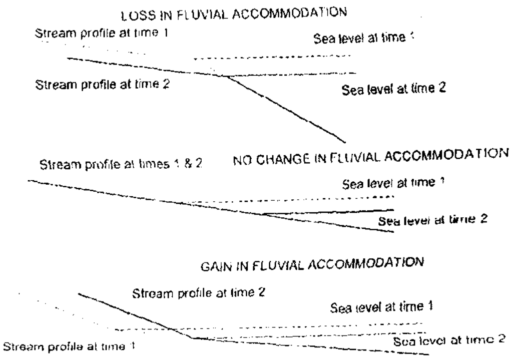

Fig. 12 Changes in fluvial accommodation in response to relative sea level fall under three different conditions of relative gradients between stream and shelf profiles.

sea level changes. Extrapolating the concept into the alluvial domain, several workers have studied the effects of relative sea level change on the non-marine base level revealing that displacement of non-marine base level in response to relative sea level changes varies depending on the relative steepness of the stream and shelf profiles (Fig. 12). Building up of an alluvial succession in consequence to relative sea level changes, however, can be analysed only if the succession contains coeval marine strata (Shanley et al., 1992). In alluvial basins located far away from the contemporary shoreline the non-marine accommodation space may change independently of relative sea level changes. In such basins accommodation space is affected by local tectonics and perhaps by climate. Therefore, the system of sequence stratigraphy for such basins is likely to be different from that of fluvio-marine basins.

As mentioned earlier, alluvial successions basically consist of two suites of strata: 1) channel-belt coarser sedimentaries and 2) overbank finer sedimentaries. Field studies of alluvial successions all over the world reveal that these two types of sediment bodies alternate vertically in the succession showing definite patterns. In vertical profiles, individual channel-belt sediment bodies may occur in isolation floating within the overbank sedimentaries. Again, a number of channel-belt

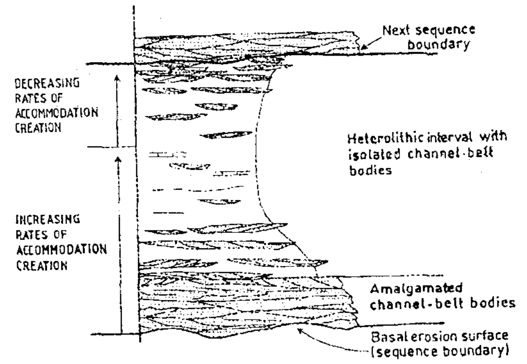

Fig. 13 A schematic vertical section through an alluvial succession showing different patterns of organization of channel-belt sediment bodies (shaded) and floodplain deposits (blank) in response to varying rate of accommodation space creation. Adopted from Olsen (1995).

sediment bodies may occur superposed upon one another defining a sediment body much larger in thickness and width than that of individual channel-belt deposit. It has also been mentioned that the degree of superposition of channel-belt sediment bodies changes vertically in a succession and this change has been qualitatively linked with change in the rate of accommodation space creation or (Fig. 13).

In alluvial basins the long-term rate of sediment supply (S) usually exceeds or equals the long-term rate of accommodation space creation (A) either by subsidence or upward movement of base level. Therefore, the long-term rate of sedimentation proxies the geological rate of subsidence and the A/S ratio will always have values between 0 and 1. If the available accommodation space is negative, the channel-belts incise valleys signifying disconformities. Unconformity without channel incision may result if the available accommodation space is zero and there is a lack of sediments. Assuming constant sediment flux, periods of higher subsidence will be characterized by higher A/S value and vice versa. It is likely that the lower and higher A/S values will have different expressions on the alluvial architecture and there will be surfaces within the alluvial succession marking abrupt changes in the A/S value (Martinsen et al., 1999; Olsen et al., 1995). Barring the unconformity surfaces that bound an alluvial sequence, the other surfaces commonly recognized in alluvial successions are :

1. Surfaces across which the stacking pattern of channel-belt and overbank sediment bodies changes but the fluvial style remains same. As mentioned earlier, periods of lower rate of accommodation space creation may be characterized by amalgamated channel-belt sediment bodies with little or no overbank fines whereas periods of higher rate of accommodation space creation could result in an alluvial succession showing isolated channel-belt bodies encased in overbank fines (Fig. 13; Burns et al., 1997; Martinsen et al., 1999).

2. Surfaces across which dimensions and flow directions of channel-belt change coupled with a change in fluvial style. Channel-belt dimensions increase during periods of lower subsidence rate with a dominant flow transverse to the basin axis and the fluvial style is usually braided. During periods of higher subsidence the alluvial plain may be characterized by meandering or anastomosed fluvial systems flowing roughly parallel to the basin axis. The factors that influence change in fluvial style are poorly understood. However, mean channel gradient and the proportion of fluid and sediment discharge is perhaps a major controlling factor. Braided channel-belt requires a larger gradient than meandering or anastomosed channel-belt. Higher discharge capacity of rivers and lower sediment load results into anastomosed network of channels, whereas lower discharge capacity and higher sediment load leads to a braided channel pattern (Burns et al., 1997)

3. Surfaces across which frequency, type and maturity of paleosol change (Fig. 14; Wright and Marriot, 1993; Kraus, 1999). Extremely low rate of sedimentation or non-deposition is a prerequisite for pedogenesis. That is why paleosols mark major

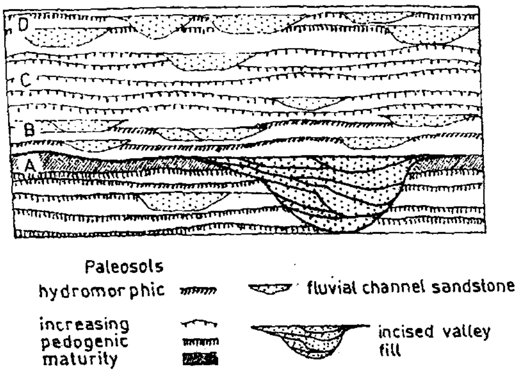

Fig. 14 Paleosol characteristics under different conditions of accommodation space creation in a sedimentary succession. The level A denotes little or no accommodation space generation signifying an unconformity and period of landscape stability. The rate of accommodation creation is highest in the level B leading to development of hydromorphic paleosols, which gradually decreases upwards resulting in increasing pedogenic maturity from level C to D. Adopted from Kraus (1999).

unconformities signifying long periods of non-deposition and non-erosion and landscape stability (Fig. 14). These paleosols are strongly developed, thick and solitary and are often laterally correlatable over long distances and are essentially allogenic i.e. their development is controlled by the prevailing tectonic and climatic conditions. Sedimentation in the alluvial floodplain is virtually episodic with intervening periods of non-deposition or erosion. Consequently, in the alluvial domain soils preferentially form on the floodplain deposits, which are essentially autogenic meaning that their development is controlled by the prevailing sedimentary processes. Periods of high subsidence and low sediment flux are characterized by development of hydromorphic paleosols. On the other hand, well-drained paleosols would form during periods of low subsidence and sediment flux (Fig. 14).

To summarise, three principal parameters, either independently or in combination, significantly control the proportion and connectedness of channel-belt sediment bodies in an alluvial succession: 1) avulsion frequency, 2) lateral mobility of the channel-belt and 3) rate of subsidence that approximates the rate of sedimentation in alluvial basins (Fig. 15; Bristow and Best, 1993)

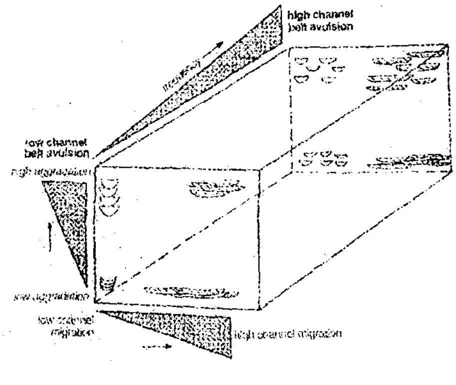

Fig. 15 Diagram illustrating the controls of different factors on the organization of channel-belt sediment bodies. Adopted from Bristow and Best (1993).

ALLUVIAL BASIN FILL: EFFECTS OF TECTONISM AND CLIMATE

The different fluvial styles discussed herein are associated with a typical combination of valley slope, discharge regime and sediment load. These intrinsic features may undergo dramatic changes in response to external (or allogenic) controls. Tectono-geomorphic studies indicate that the long-term evolution of drainage basins and consequential evolution of fluvial systems are largely controlled by allogenic factors such as tectonism, climate, eustatic sea level which leave record in the deposits.

Alluvial depositional systems are the principal agents of sediment distribution in intracontinental rift, strike-slip and foreland basins. The fills of these basins are typically asymmetric due to localization of tectonic activity along one basin margin causing maximum subsidence there. In rift basins the active margin is characterized by normal faults, in contrast to strike-slip and thrust faults that prevail along the margins of strike-slip and foreland basins respectively.

Geomorphologic studies in active extensional basin of SW Montana, USA (Basin and Range Province) has revealed that fault-induced tilting of the basin floor forces the meander-belts to preferentially migrate in the direction of tilt (producing a an asymmetric meander-belt). The resulting channel-belt sediment bodies are more extensive than would have been the case if there were no ground tilt (cf. Collinson, 1978). The migration of the meander-belt is achieved through the asymmetric growth of meander loops in the up-slope and the down-slope sides of the belt. The loops growing opposite to the ground tilt are preferentially preserved (Leeder & Alexander, 1987) and hence the lateral accretion surfaces dipping up-slope are more abundant in the resulting deposits. Recognition of such asymmetric distribution of lateral accretion surfaces in the ancient rock record may denote direction of tilt of the basin floor and location of active subsidence. Spatial changes in the channel pattern and development of terraces are also found to be related to the movements across fault planes (Alexander & Leeder, 1990).

In asymmetrically subsiding basins fluvial systems occur at two different orientations with respect to the basin axis (Fig. 16): 1) flowing at a high angle or normal to the basin axis originating from the basin boundaries (transverse system) and 2) flowing at a low angle or parallel to the basin axis (axial system). Studies of modern rift basins (Frostick & Reid, 1989; Leeder & Gawthrope, 1987) have revealed that footwall blocks are comparatively sediment starved due to rotation of the tilt block away from the basin and major transverse drainage network enters the basin through the hanging wall block forming large, low-gradient clastic cones. Tectonism exerts a significant influence on sedimentation, not only on the type and thickness of sediment, but also on the areal distribution of the transverse and axial fluvial systems within the basin. The role of tectonism in determining system distribution is especially important in asymmetrically subsiding basins, such as half grabens, foreland basins, and some strike slip basins

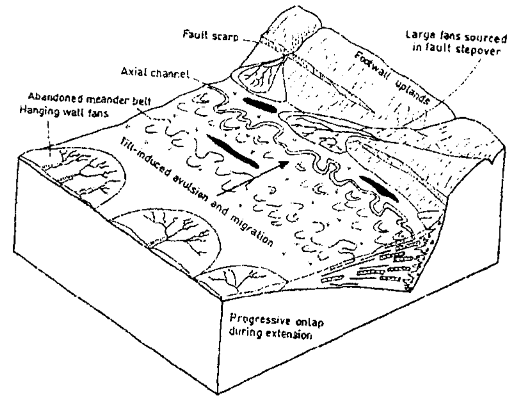

Fig. 16 Schematic representation of the disposition of different elements of the alluvial system in intracontinental, asymmetrically subsiding sedimentary basin. In extensional setting the active margin is marked by normal faults, where thrust faults prevail in compressional setting. Adopted from Leeder (1999).

(Alexander and Leeder, 1987; Cant and Stockmal, 1993; Nilsen and McLaughlin, 1985). Recent depositional models suggest that periods of rapid subsidence in asymmetrical basins result in deposition of fine-grained sediments from the longitudinal fluvial system, directly above the locus of maximum subsidence, close to the margin of the uplifted terrane (Blair and Bilodeau, 1988; Mack and Seager, 1990). Because denudation is significantly slower than the rates of uplift and subsidence, coarse-grained deposit of transverse fluvial system is restricted to a narrow zone directly adjacent to the uplifted terrane. However, as the rate of subsidence and uplift diminishes, erosion rate eventually surpasses subsidence rate and a sheet of coarse sediment spreads across the basin (Fig. 16; Burns et al., 1997). Thus, near the locus of maximum subsidence, finer-grained axial facies are syn-tectonic and coarser grained transverse facies are post-tectonic. In alluvial successions the lateral spread of the transverse axial deposits is thus a proxy measure of the relative rate of subsidence. The vertical variation of the spatial extent of successive packages of transverse fluvial systems in response to temporally varying subsidence rate has also been

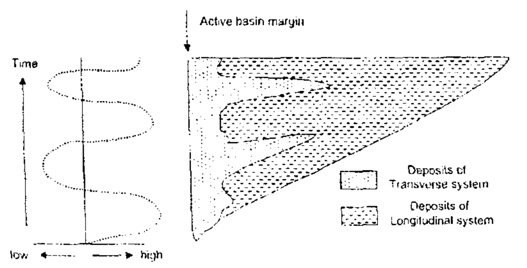

Fig. 17 Schematic representation of the results of numerical experiments illustrating the effects of subsidence rate on the lateral extent of the transverse fluvial system in the longitudinal profile of the basin. (After Paola, 2000)

numerically simulated using dynamic models of sediment diffusion (Fig. 17; see Paola, 2000).

Many of the responses of the fluvial system to tectonism as discussed above may also arise due to climatic influences. A number of well-constrained studies of the modern river basins and associated Quaternary deposits reveal major changes in planform pattern, discharge and sediment load characteristics of the river system over a period of 100 Ka. Based on available evidences these changes have been correlated with the changes in climate and sea-level rather than with the tectonism (for a detailed review see Blum & Tornqvist, 2000). Study of the major rivers of Europe (Straffin et al, 2000; Tornqvist, 1998) and USA (Blum & Valestro, 1994) reveal that during the last glacial period most of the rivers were coarse-grained braided type whereas during interglacial periods they were transformed into mixed-load meandering rivers. It is postulated that cyclic incision and aggradation documented from the Quaternary deposits of these river basins also reflect climatic and glacio-eustatic changes rather than the changes in tectonically induced subsidence.

ACKNOWLEDGEMENTS

We extend our sincere thanks to the Council of Geological Mining and Metallurgical Society of India for inviting us to prepare this overview. Financial assistance from DST, New Delhi is gratefully acknowledged. The Indian Statistical Institute provided the infrastructural facilities. We thank Mr. Ajay Das for drafting the figures. We are also indebted to a large number of students and colleagues of our Institute for providing enthusiastic support during preparation of the manuscript. We particularly thank Sanjoy Ghosh of ISI who was ever ready to help us compile this large manuscript against a tight schedule.

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