2 Modelling Channel Evolution and Floodplain Morphology

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2.1 INTRODUCTION

Floodplains result from the long-term cumulative action of the flow, erosion and depositional processes considered in this book. It is sometimes convenient in engineering or geomorphic practice to consider the floodplain (including the river channels) as a fixed geographic boundary over which flow properties and sediment transport are calculated to predict flow depths, scour potential, forces on structures, and rates and size distribution of sediment deposition. However, in many high energy fluvial regimes the rates of boundary modification from individual floods may be so great that the influence of recent and contemporaneous flooding in sculpting the floodplain cannot be ignored (e.g. Schumm and Lichty, 1963; Anderson and Calver, 1977, 1980; Graf, 1983; Kieffer, 1985; Nanson, 1986; Osterkamp and Costa, 1987; Baker, 1988; Miller and Parkinson, 1993). Even in lower energy fluvial systems, cumulative changes in channel pattern and floodplain morphology over time-scales of decades are often great enough to require their incorporation into engineering design and long-term sediment budgets. In addition, there are a range of issues such as groundwater flow in floodplains and petroleum reservoir characterization that require understanding of the internal stratigraphy of floodplains and hence of the history of floodplain development. The focus of this contribution is on recent developments and continuing research needs in the modelling of floodplain evolution over time-scales from years to centuries.

The long-term modelling approaches discussed here are based, directly or indirectly, upon the contemporary floodplain processes that are the topic of this book. But, of necessity, the long time-scales considered in this contribution require simplification of process laws, often to the point of using heuristic summaries of empirical observations. The major reason for the process simplification is the constraint that the long time intervals and large spatial scales place upon use of numerical models. Although this constraint is a receding horizon as computer capabilities improve, models based upon fundamental treatment of flow, erosion, transport, and deposition in channels and on floodplains extending over decades to thousands of years and over tens of square

kilometres remain for a future generation. Even were such models practical, more modest model structure with simplified assumptions still might be preferred. The evolution of channel planform commonly depends sensitively on local boundary conditions and small differences in initial conditions. This is particularly true of flood chute development, meander cutoffs, creation and abandonment of braided-channel anastomoses, and channel avulsions, as well as crevasse-splay sedimentation. This chaotic behaviour suggests that even the most detailed model has limited predictive power when extrapolated over timespans of decades or longer. Our understanding of a number of constraints and processes occurring on floodplains is sufficiently uncertain that the predictive accuracy of any model is thereby limited; this includes transport of mixed bed sizes, weathering of sediment temporarily stored in floodplains, the interaction of vegetation with erosional and depositional processes, processes controlling bank erosion, and future history of discharge and sediment load. Finally, the use of simplified models allows wide exploration of the effects of variations in process assumptions as well as boundary and initial conditions, while promoting a humility about the accuracy and generality of conclusions.

Models of floodplain development can be constructed for a variety of purposes, such as prediction of future channel pattern changes and associated erosion and deposition for engineering purposes, or simulation of floodplain stratigraphy for groundwater characterization and/or for understanding of the stratigraphic record. The spatial and temporal scales, as well as the level of detail to which processes are considered, will depend upon the model purpose. The models considered here are primarily oriented towards understanding of the long-term, broad-scale evolution of floodplains and channels and hence many process details are very simplified. However, some discussion is devoted towards prospects for incorporating more explicit process representation.

2.1.1   A spectrum of channel planforms and floodplain morphology

Floodplains and their associated channel systems are remarkably varied. Channel patterns are commonly classified into meandering, braided, straight, and anastomosing. Each of these has a wide range of variation among natural channels, and all intergrade. Although there have been a wealth of studies of the influence upon planform of channel and valley gradient, flow regime, sediment supply (both amount and size range), erosional history (e.g. tectonic deformation and long-term aggradation or entrenchment), and local physiography, there are no universal criteria for predicting planform type or even a universal agreement on leading causative factors (see, for example, recent reviews by Ferguson, 1987; Gregory and Schumm, 1987; Bridge, 1993; Germanoski and Schumm, 1993; Knighton and Nanson, 1993; Smith and Perez-Arlucea, 1994). Floodplains are commonly classified by the associated channel planform type (e.g. Lewin, 1978), but physiographic constraints, particularly valley width, are also important. Nanson and Croke (1992) classify floodplains by a combination of sedimentary environment, formative processes, and stream power per unit channel width. Their main three divisions are: (a) high energy floodplains with non-cohesive sediments; (b) medium energy floodplains with non-cohesive sediments, and subdivided into those formed by (i) braided and (ii) meandering channels; and (c)

low energy cohesive sediment floodplains of single or anastomosing channels. By further subdivision Nanson and Croke (1992) recognize a total of 15 floodplain types.

The most appropriate approaches to modelling floodplain temporal evolution vary with floodplain type. The selection of classes of floodplain models discussed here is based primarily on differences in model structure and technique, but these in turn are conditioned by channel planform, physiographic setting and the flow and sediment regime. Not all of the types of floodplains distinguished by Lewin (1978) and Nanson and Croke (1992) have been sufficiently well characterized to have been quantitatively modelled (e.g. the sandy floodplains subject to catastrophic channel widening during major floods discussed by Schumm and Lichty (1963) and Burkham (1972)). The broad categories of floodplains discussed here are: (1) strongly confined high energy; (2) braided channel; (3) meandering; and (4) avulsive and anastomosing.

The overall valley width is an important factor in all but the broadest of floodplains, such as the lower Mississippi River. A combination of five or more intergrading mechanisms may determine available valley width. In rapidly incising river systems the stream and the valley are essentially coincident, with valley width often no wider than can be scoured by the river in flood. Some channels are underfit, occupying valleys scoured by palaeofloods such as glacial meltwater (Dury, 1954, 1964a,b). Aggradation tends to widen valleys due to the layback of valley walls. Lateral channel migration, particularly in meandering channels, can undercut valley walls (Rich, 1914; Moore, 1926; Palmquist, 1975). Finally, if alluvial channels are restricted in their ability to downcut by a stable baselevel, valley walls will continue to erode, forming a widening pediment surface (as simulated by Howard, 1994, his Figure 19). This latter case is distinct from valley widening by lateral erosion, because the main channel complex does not necessarily occupy the entire valley width and fan-like pediments may extend from the valley edges to the centre of the valley. All of these processes may be influenced by tectonics (either active deformation or through rock structure) and lithology.

2.2   HIGH ENERGY CONFINED FLOODPLAINS

A wide spectrum of high energy confined floodplains exists, distinguished by floodplain width, flood frequency distribution, bed type, sediment characteristics and vegetation. The present discussion highlights three types selected from among this wide range, characterizing them slightly differently from Nanson and Croke (1992).

2.2.1   Chute canyon channels

Chute canyon channels are confined between narrow bedrock walls, usually within deep canyons, so that flood flows frequently extend from wall to wall. In some cases, such as portions of the granite narrows of the Colorado River in the Grand Canyon, even median or average flows fill the valley bottom, so that a floodplain is generally restricted to small fillets of deposited suspended sediment in slightly wider alcoves and an occasional gravel bar or small tributary fan exposed during low flow conditions. Slackwater sediments may be deposited along the lower course of tributary channels

(e.g. O'Connor *et al.*, 1994). Channel beds may be bedrock, sand or gravel/boulder. Channel gradients range from very steep with many rapids in boulder or bedrock channels, to quite gentle in sand-bed sections such as portions of the Marble and Echo Canyon sections of the Colorado River (Howard and Dolan, 1981). Also within this category are somewhat wider canyons that expose some gravel, bedrock or sandy valley bottom during most flow conditions. Baker and Kochel (1988) call these bedrock fluvial systems, although similar features are found in narrow canyons with alluvial beds. Characteristic of such channels are discontinuous gravel-bar floodplains and finer slackwater deposits in alcoves and along tributaries (Baker, 1977; Baker and Kochel, 1988; Kochel and Baker, 1988). The alluvial deposits in such canyons are transient, being remoulded by floods large enough to occupy the valley width.

2.2.2   Tributary fan canyons

In deeply incised canyons, steep tributaries occasionally disgorge large quantities of mixed sediment sizes as debris flows into the main river channel. In the very narrow canyons discussed above, this debris flow material may be rapidly eroded by main-stem flood flows, or it may form largely submerged bars and associated rapids (Dolan *et al.*, 1978; Howard and Dolan, 1981). In somewhat wider canyon sections, however, this debris flow material is deposited as a fan that laterally constricts the main-stem flow, generally creating rapids. The coarsest boulders in such debris flows are generally much coarser than the gravels transported through the main stem, so that main-stem flood flows winnow the finer sediment and create rapids with gradients at the threshold of motion of the boulders (Graf, 1979; Howard and Dolan, 1981; Kieffer, 1985). The steepness and length of the rapids thus depend upon the size and quantity of the supplied boulders, the highest post-deposition main-stem floods, and subsequent weathering, sorting, and abrasion of the fan deposits. Such systems of rapids are therefore temporally varying as older deposits are gradually eroded and local flooding creates new fans or replenishes existing ones (Howard and Dolan, 1981; Kieffer, 1985). The fans and associated rapids in turn provide a framework for suspended load deposition. Thin, discontinuous fillets of sandy to clayey sediments may accumulate on the valley walls along the pools between rapids, but most fine sediments accumulate in association with low-velocity zones associated with the tributary fans. Recirculation zones develop against the fans both upstream and downstream from the rapids (Figure 2.1), encouraging deposition of a veneer of suspended load deposits (Howard and Dolan, 1981; Rubin *et al.*, 1990; Schmidt, 1990; Bauer and Schmidt, 1993; Schmidt and Rubin, 1995). The fine sediments are generally deposited during flood flows, resulting in low benches exposed during low flows. Very high floods capable of moving the fan boulders will cause complete reworking and possibly partial removal of the suspended load benches. The extent of suspended load deposits depends upon the balance between addition of suspended sediments into the canyon (generally by summer and autumn flooding of desert tributaries in the case of the Colorado River) and redistribution and removal by large floods (snowmelt spring floods on the Colorado River). In the case of the Colorado River, disruption of sediment input by Glen Canyon Dam has resulted in partial erosion of the benches (Howard and Dolan, 1981; Schmidt and Graf, 1990).
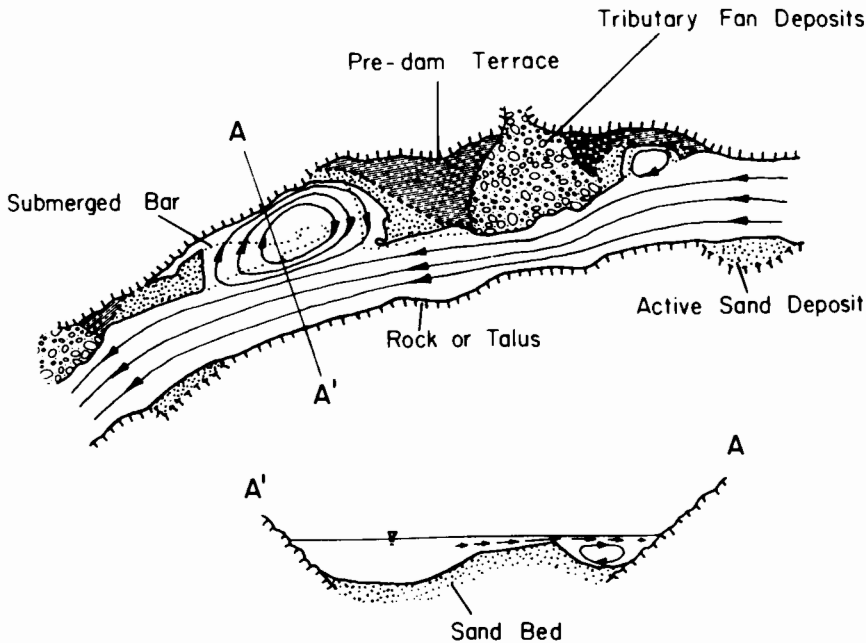
Figure 2.1 A tributary fan complex of the Colorado River in the Grand Canyon. Tributary fan deposits originate from debris flows from tributary side canyons. Both pre-dam terraces and active sand deposits are suspended load deposits. Sand deposits on the upstream side of the eddy are called separation bars, and the partially submerged bar on the downstream end is a reattachment bar (Schmidt, 1990). Illustration from Howard and Dolan (1981)

2.2.3 Vertical accretion floodplains

In somewhat wider and less steep confined valleys a pattern of vertical floodplain accretion during modest floods may be interrupted by widespread floodplain stripping during very large floods (Kochel *et al.*, 1982; Nanson, 1986; Kochel, 1988). Because of the time dependent floodplain height, Nanson (1986) terms these disequilibrium floodplains. Floodplain sediments coarsen upwards from gravels to sandy loam, with pronounced natural levees. Vertical accretion rates diminish as floodplain height increases, in accordance with observations in many floodplain environments (Wolman and Leopold, 1957; Everitt, 1968; Nanson, 1980). Stripping during large floods occurs most extensively near the main channel, and concomitant deposition may occur on more remote portions of the floodplain (Nanson, 1986).

Large floods also play a major role in somewhat wider valleys in hilly or mountainous terrain occupied by gravel or bedrock channels. The role of floods in such valleys has been described by Hack and Goodlett (1960), Williams and Guy (1973), Ritter and Blakely (1986), Ritter (1988), Miller (1990b), Harper (1991), and, most comprehensively, by Miller and Parkinson (1993). Channels in such valleys are commonly meandering where valley width permits, and often weakly braided. The meandering apparently occurs slowly enough that vertical accretion and stripping are

the dominant floodplain-forming processes. Tributary fans occur locally in these valleys (Miller and Parkinson, 1993), but do not control depositional and erosional processes as strongly as in tributary fan canyons. As with the narrower vertical accretion floodplains described above, floods with recurrence intervals less than about 50 years are primarily depositional, and floodplains are composed of a gravel base with generally less than 2 m of suspended load accretionary deposits. Major floods strip vegetation from and erode channel banks. Overbank flow locally scours the floodplains, often along pre-existing flood chutes and floodplain depressions, exposing the underlying gravel/boulder framework. These chutes appear to be semi-permanent features in which modest floods deposit sediment, but they are reoccupied and rescoured by large floods (Harper, 1991; Miller and Parkinson, 1993). Localized scour is also triggered by vortices generated by floodplain topography, trees and man-made structures. The most intense scour and most common location for chutes occurs on floodplain enclosed within meander bends (Figure 2.2). Scour also becomes concentrated at locations where flow leaves or returns to the main channel. In narrowly confined sections or across sharp bends vegetation destruction and floodplain stripping may be general enough to approach the catastrophic erosion described by Nanson (1986). Gravel scoured from the main channel or from stripped areas of the floodplain may be spread as a thin layer on essentially uneroded portions of the floodplain (Ritter, 1975, 1988; Miller and Parkinson, 1993). Modest amounts of suspended load deposition may occur along the margins of the floodplain where flow is slow. This pattern is somewhat modified in locations where extensive, concomitant debris
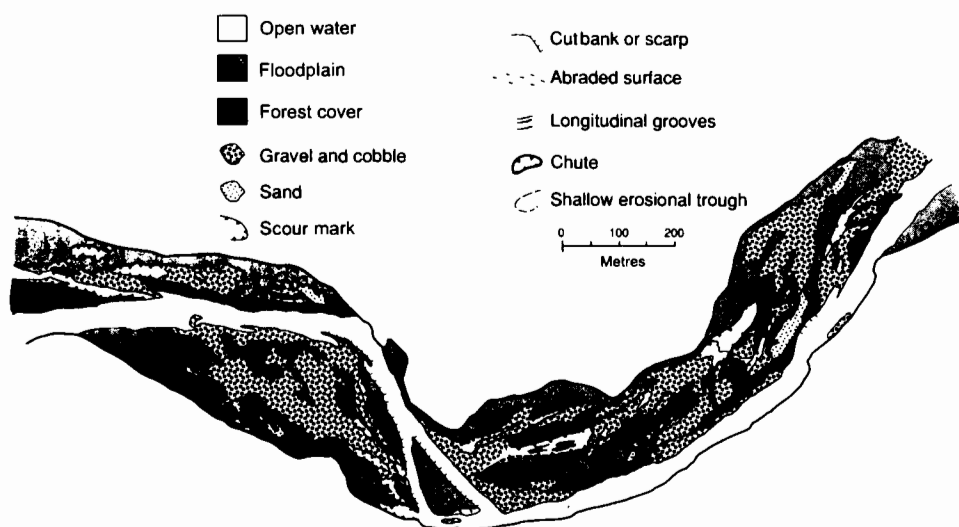

Figure 2.2   Patterns of flood scour and deposition on the North Fork of the South Branch of the Potomac River, West Virginia, resulting from 1985 flooding. Flow is from left to right. Gravel and cobble areas are partially exposed basal floodplain deposits and partially gravels transported onto the floodplain. Most of the forest cover originally bordering the channel has been stripped, and flood chutes have been excavated locally. Illustration from Miller and Parkinson (1993)

avalanches occur on steep mountain slopes due to high rainfall intensities (Hack and Goodlett, 1960; Williams and Guy, 1973; Miller, 1990b; Jacobson *et al.*, 1993), in that more extensive valley bottom sedimentation occurs, often accompanied by construction of levees and dams of trees stripped from debris avalanche sites and the floodplain. Miller (1990a) distinguishes a continuum of flood events ranging from very localized flooding due to high-intensity thunderstorms that have catastrophic effects only on headwater channels (e.g. Hack and Goodlett, 1960), through hurricane-induced rainfall extending over somewhat larger areas resulting in rainfall exceeding 300 mm in a few hours (e.g. Williams and Guy, 1973), to more regional rainfalls exceeding 100 mm over a period of a few days that primarily affect larger rivers (Wolman and Eiler, 1958; Costa, 1974; Gupta and Fox, 1974; Miller and Parkinson, 1993). Costa and O'Connor (1995) suggest that flood duration may be as important as flood magnitude in determining the degree of floodplain modification.

2.2.4 Prospects for modelling

Most of the model components necessary to quantitatively simulate geomorphic evolution of high energy floodplains are in a reasonable state of development, as manifested by the chapters of the present book. Because most of the geomorphic modification of high energy floodplains occurs during overbank floods, flow models capable of treating both within-channel and overbank flow are necessary (e.g. Knight and Demetriou, 1983; Yen and Yen, 1984; Ervine and Ellis, 1987; Knight, 1989; Gee *et al.*, 1990; Bates *et al.*, 1992; Miller, 1995; Knight and Shiono, Chapter 5, this volume; Sellin and Willetts, Chapter 8, this volume; Younis, Chapter 9, this volume). Routing of suspended sediment and its deposition on the floodplain can be treated by a variety of approaches ranging from the heuristic rules (Howard, 1992, and discussion below) to mechanistic treatment of advection and deposition from suspension (e.g. Carey, 1969; Pizzuto, 1987; Marriott, 1992, Chapter 3, this volume; Nicholas and Walling, 1995). However, the most challenging component of geomorphic models will probably be treatment of floodplain scour. The resistance offered by cohesive sediments and floodplain vegetation induce a threshold to scour that is difficult to quantify. Furthermore, scour is strongly affected by subtle variations in floodplain topography and vortices generated by flow obstacles (Ritter, 1988; Miller and Parkinson, 1993; Miller, 1995).

The greatest efforts in modelling of high energy floodplains have been directed towards understanding the balance of suspended load deposition and erosion in tributary fan canyons, particularly the Colorado River in the Grand Canyon. The mechanics of flow and suspended load transport and deposition has been investigated in field studies (McDonald *et al.*, 1994) flume experiments (Schmidt *et al.*, 1993) and theoretical studies (Nelson *et al.*, 1994; Smith, 1994). Miller (1994, 1995) has used depth-averaged flow modelling to investigate shear-stress distribution in variable-width valleys and in canyon reaches with tributary fans, showing that maximum stresses are much greater than in uniform reaches. However, modelling is still far short of full simulation of suspended load benches and recirculation bar development, or of the emplacement and subsequent modification and erosion of tributary fan deposits.

2.3 MODELLING OF BRAIDED STREAM DEVELOPMENT

Straight or irregular single-thread channel patterns are the exception on wide floodplains, where braiding, meandering and/or anastomosing patterns are most common. These more complicated channel patterns require both process(es) leading to the instability of straight, single-thread channels as well as other constraints or processes which prevent infinite elaboration of the pattern. Processes which may promote instability include (1) the non-linearity of sediment transport in response to applied stress, including the threshold of motion; (2) Interaction of grain sizes during transport; (3) secondary and recirculating fluid flows; (4) topographic effects upon fluid motion, including direction and magnitude of flow; (5) topographic effects upon sediment transport, including the threshold of motion, transport rate, and transport direction; (6) lags between change of bed shear and sediment transport rate; (7) effects of surface waves on flow and sediment transport, including out-of-phase relationships between the water surface and bed topography (Froude number effects); (8) discharge fluctuation; and (9) valley aggradation. Many of these instability mechanisms are outlined in Nelson (1990) and McLean (1990). Processes and constraints which limit the degree of planform elaboration include: (1) a finite valley width; (2) cutoffs; (3) finite quantity and depth of flow; and (4) topographic effects upon fluid motion and sediment transport.

Stream braiding results from two processes: (1) channel avulsion, that is, the partial or full diversion of flow across the channel bank (analogous to chute or neck cutoffs in meanders), and (2) growth of and emergence of within-channel bars and their subsequent dissection into two or more subchannels (Brice, 1964; Ashmore, 1991; Bridge, 1993; Bristow and Best, 1993; Ferguson, 1993; Leddy *et al.*, 1993). Many types of avulsion and bar dissection have been distinguished based upon channel geometry, origin by blocking of flow by sediment deposition versus headward erosion along the diversion path, degree of channel curvature and channel asymmetry, etc. So long as the emergent bars remain unvegetated, braided channel patterns are very changeable, with frequent addition and abandonment of anabranches (see Figure 2.4). The role of bars in creation of braiding has been documented in many field studies (e.g. Williams and Rust 1969; Cant and Walker, 1978; Ashmore, 1982; Church and Jones, 1982) and flume experiments (e.g. Ashmore, 1982, 1993; Fujita, 1989; Hoey and Sutherland, 1991; Germanoski and Schumm, 1993). Leopold and Wolman (1957) cite the development of braiding in gravel streams as resulting from growth of medial bars due to local accumulations of coarse grains that the stream is incompetent to transport. However, braiding also occurs in uniform sand channels at transport levels well above threshold, so that incompetence cannot be a universal explanation.

Most theoretical studies view braids as developing from bar-form instabilities resulting from bed topography effects on flow and sediment transport. Early theoretical approaches utilized linear perturbation analysis (e.g. Engelund and Skovgaard, 1973; Parker, 1976; Callander, 1978; Fredsoe, 1978; Blondeaux and Seminara, 1985; Struiksma and Crosato, 1989), and more recent approaches have broadened the analysis to finite amplitude bars, either mathematically or through mathematical simulation (Fukuoka, 1989; Nelson and Smith, 1989a,b; Seminara and Tubino, 1989; Nelson, 1990). The initial instability is due to topographic steering of flow and sediment

transport (Nelson, 1990), which is greater the shorter the bar wavelength. However, due to the lag between downstream changes in bed elevation and resulting boundary shear stress, an optimal bar growth rate occurs at a finite wavelength (Nelson, 1990). Finite-amplitude models include non-linear effects of flow and sediment transport as well as secondary flows (Nelson, 1990). These studies suggest that the degree of bar development is a strong function of the channel width W to depth H ratio (W/H) and a weaker function of channel gradient and flow stage (τ/τ_c where τ is bed shear stress and τ_c is the critical shear stress for sediment movement). For very narrow channels ($W/H < 10$) systematic bars cannot form, although curvature-induced (point) bars can occur. For larger W/H values alternate bars form, and for very wide channels, multiple bars occur, including both alternate bars at channel banks and linguoid bars in the central portions of the channel (Figure 2.3). The intensity of braiding has been described by a number of indices, including Brice's (1964) braiding index (twice the total length of bars divided by reach length), a total sinuosity defined as total channel length divided by reach length (Richards, 1982; Friend and Sinha, 1993; Robertson-Rintoul and Richards, 1993), and the average number of anabranches minus one (Howard *et al.*, 1970). The stability analyses suggest the bar mode, m, as a more fundamental quantity, which is related to the average number of scour holes and
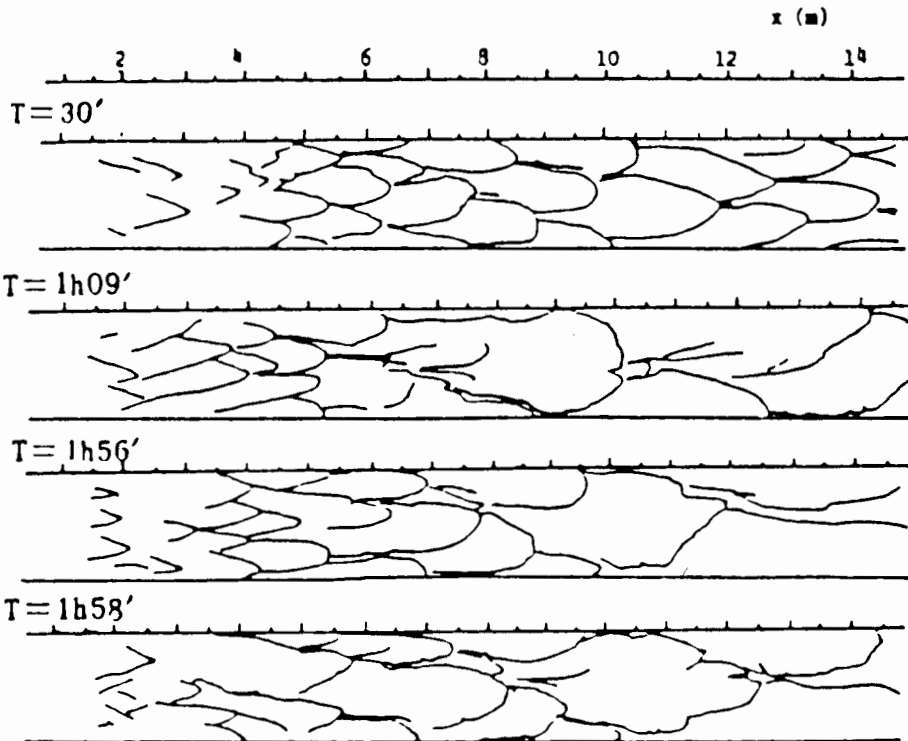

Figure 2.3 Temporal evolution of bars and braiding in a laboratory sand-bed channel (from Fujita, 1989). The linguoid bars in the initial stages are submerged, but portions of the bars in the final channel have emerged, forming a braided channel

channel thalwegs per cross-section, n (Fujita, 1989):

$$m = 2n - 1 \tag{2.1}$$

This index offers the advantage that it takes into account both distinct channels as well as submerged bars and it should not be strongly dependent upon flow stage, but it has the disadvantage that in-channel measurement is required (Bridge, 1993; Ferguson, 1993).

Field observations and flume experiments indicate that channels with alternate bars do not necessarily develop into braided channels. In such cases the finite channel width and non-linear effects of flow depth and bed topography stabilize the height of bars before they become emergent. However, for higher bar modes flume experiments suggest that, starting from a plane bed, braiding develops from a fairly regular pattern of generally linguoid bars. These bars tend to lengthen and increase in amplitude with decrease in m with time, as well as becoming more irregular through the development of smaller superimposed bars, until the largest bars may become exposed and relatively fixed in location, creating a braided pattern (Figure 2.3). Fujita (1989) relates the submerged bar mode in flume studies to channel width, depth and sediment grain size, d:

$$2.2 \ m^{2/3} < (W/d)^{2/3}(d/H) < 6.7 \ m^{2/3} \tag{2.2}$$

where the inequalities indicate that there is an overlapping range of widths and depths for each bar mode. Braided channels tend to have slightly higher values of width to depth ratio for a given bar mode than the relationship above (Fujita, 1989).

The above analysis suggests that braiding is related to a greater freedom of adjustment in wide channels in which the negative feedbacks limiting bar growth are insufficient to prevent bar emergence and the chaos of braiding. A fundamental unit in development of braids is the combination of a scour hole (often located at braid confluences; Ashmore, 1993) and a downstream depositional bar (Ashmore, 1982, 1991, 1993; Bridge, 1993; Ferguson, 1993), often called a pool-bar complex or a chute and lobe. The following enumeration of braiding behaviour is largely summarized from Ashmore (1991, 1993), Bridge (1993), Bristow and Best (1993), Ferguson (1993), Germanoski and Schumm (1993), and Leddy et al. (1993). Flow is usually convergent in the pools and divergent on the bars. The bars can become emergent even at constant discharge due to channel widening and flowstage lowering at the bar or migration and stalling of dunes or bedload sheets on the bar crest. Multiple channels can develop directly due to mid-channel bar emergence or by channelling and headward erosion by water flowing over the top of the bar. Incision may be aided by lowering river stage due to decrease in discharge. Individual braid channels can be curved, leading to asymmetric bar development (including point bars) and lateral channel migration. Pulses of bedload and temporal variation of water and sediment influx at junctions can cause aggradation and avulsions, blocking of channels, and migration and reorientation of confluences. Overall aggradation leads to a greater degree of braiding and more frequent channel shifting.

Braided channels are traditionally viewed as being distinguished from meandering channels by a threshold value of channel gradient and/or discharge (Leopold and Wolman, 1957; and numerous subsequent studies reviewed by Ferguson, 1987, 1993,

and Bridge, 1993). Additional criteria are bank strength, sediment size and quantity, stream competence, and width–depth ratio (e.g. Equation 2.2). Ferguson (1987) points out that most of these criteria are strongly interrelated by channel hydraulics. Although most studies have emphasized a threshold between meandering and braiding, the indices of braiding discussed above suggest a continuum from single-channel to strongly braided, and a few studies (e.g. Howard et al., 1970; Moseley, 1981; Richards, 1982) have related degree of braiding to hydraulic parameters.

Where valleys occupied by braided rivers are wider than the active channel complex, much of the floodplain is often essentially unmodified channels left behind after the active braid-train migrates to a new location (Figure 2.4). Portions of the floodplain which have been inactive for longer periods may become vegetated with thin caps of accreted fines from overbank deposition (Reinfelds and Nanson, 1993). Creation, abandonment and shifting of anabranches is a continuing process in braided rivers with negligible braid-bar vegetation. Islands are generally about the same dimensions as the anabranches, although the extent of bars/islands diminishes as they become submerged at higher stages (e.g. Mosley, 1982). In such braided systems small islands generally can be considered to have originated as submerged bars that have become emergent either by falling stage or by erosion and water surface lowering by the bordering anastomoses. Larger islands are generally just inactive portions of the braid-train. However, when climate, flood regime and sedimentary characteristics permit vegetational colonization of braid bars, these bars acquire a resistance to scour, and often increase in size by both lateral and vertical accretion. As a result, active channel width becomes restricted, with fewer, deeper and larger anastomoses. Furthermore, the frequency of channel shifting diminishes, often occurring only during large floods. Creation of new channels is primarily by creation of chutes and avulsions across existing vegetated islands.

The development of a simulation model of braiding and associated sedimentation and floodplain construction that incorporates a detailed description of flow and sediment transport would be a formidable undertaking because of the multiple channels, constantly changing boundary conditions (including creation and abandonment of channels and some meandering in the larger anabranches), and the sensitivity of the braiding pattern to discharge variation. As a result, extant modelling approaches have been very heuristic, with most having no explicit treatment of flow and sediment transport, and one (the Murray and Paola, 1994 model discussed below) having a very simplified hydraulic formulation. All of these models are targeted towards braided channels lacking appreciable vegetated islands.

Early attempts at modelling braided channel complexes utilized a downstream random walk with stochastic channel branching and recombination (Howard et al., 1970; Krumbein and Orme, 1972; Rachocki, 1981). The channel pattern is generated by marching downstream with existing channels extending downstream with an assigned probability of a lateral component of motion and of branching. The lateral component of motion may cause neighbouring channels to intersect and combine. The combination of branching and recombining generates a braided pattern. In order to create a braided planform with a statistically constant frequency of channels, there must either be a lateral constraint on floodplain width or a central tendency for lateral movements. Such stochastically generated braided planforms have statistical characteristics of braid

Figure 2.4 A braided tributary of the Toklat River, Denali National Park, Alaska. Flow from top to bottom

geometry that are similar to natural braided channel complexes (Howard *et al.*, 1970; Krumbein and Orme, 1972). A similar approach can be used to simulate alluvial fans (Rachocki, 1981) and deltaic distributaries (Smart and Moruzzi, 1972). This modelling, although static, indicates that branching and recombination of braided channels can be viewed as a stochastic, or possibly chaotic, process.

Webb (1994, 1995) has extended the random-walk model by accounting for the discharge and cross-sectional shape within each channel. Furthermore, the stratigraphy of aggradational braided channel floodplains is simulated by accounting for the deposition from each channel. Successive deposits are simulated by via generations of random-walk channel systems, with pattern changes between each generation being constrained by the previous pattern. Braided patterns simulated by Webb (1994, 1995) have morphometric characteristics that are closer to those of natural braided channels than the earlier models of Howard *et al.* (1970) and Krumbein and Orme (1972).

The major limitation of the random-walk models is that they do not account directly for the mechanics of flow and sediment transport that cause braiding, whereas the major advantage is their computational efficiency. Such models may prove to be useful for simulation of floodplain stratigraphy and topography if the rules of generation are empirically adjusted to replicate natural channel kinematics. The Webb (1994, 1995) approach seems promising in this respect, although the validation and calibration are limited by the sparsity of observations. The random-walk models offer little insight into the flow and transport mechanics that create braiding.

A novel approach to simulating braiding has been developed by Murray and Paola (1994). The model simulates flow and sediment transport over a mesh of nodes. Flow and sediment from each node is routed downstream, being divided among the three nodes lying directly downstream and to the left and right. The flow Q_i from a given node is apportioned between the downstream nodes in proportion to the gradient S_i to the downstream node:

$$Q_i = Q_u S_i^n / \Sigma S_i^n \qquad (2.3)$$

where Q_u is the total flow reaching the node from upstream and the exponent n is less than or equal to unity (flow is routed only to those nodes with positive S_i and the summation occurs only for positive S_i). An important additional feature is that if all S_i are negative (the local node is a depression), flow out of the node is apportioned among the downstream nodes, presumably accounting for flow momentum. Sediment discharge Q_{si} to downstream nodes is assumed to be a power function of gradient and discharge:

$$Q_{si} = K[Q_i(S_i + C)]^m \qquad (2.4)$$

where K is a constant, C is an additive constant on the order of the average gradient, and the exponent m is typically 2 to 3. The C term allows for a certain amount of upslope sediment transport. Certain other transport relationships also result in braided patterns. The realism of the braiding can be improved if a small amount of lateral transport is also permitted. Figure 2.5 (Plate I) shows bed topography and flow discharge produced by a simulation utilizing the Murray and Paola (1994) algorithm.

Braiding in the Murray and Paola (1994) model results from scour where flow converges and downstream deposition as a broad bar around which the flow ultimately

becomes diverted when the bar become sufficiently high. These high bars can become emergent when flow becomes diverted through adjacent scour holes. In this respect the model captures the evolutionary sequence observed in flume studies, including the reduction in bar mode through time and a continuing creation and abandonment of braid channels (cf. Figures 2.5a and 2.5b). The model does not account for curvature-related effects on flow and transport, including lateral migration and point-bar development. The channels simulated by the Murray and Paola (1994) model have a visual similarity to natural and laboratory braiding, but it remains to be seen how successfully the model can replicate the morphometry of braided streams, channel changes through time, and braid deposit sedimentology. None the less, the richness of pattern of the simulated braiding suggests that much of the diversity of braid geometry and of kinematic patterns of initiation, growth and decay of braid channels results from a single basic instability in flow and transport.

Despite the mathematical simplicity of the Murray and Paola (1994) model, restrictions on the allowable iteration time-step due to requirements for numerical stability severely limit the spatial size and temporal duration of simulations. An even simpler approach might be possible using the techniques recently developed for simulating aeolian dune fields by Werner (1995). In this approach sediment movement occurs as small unit sheets and the interaction of transport with topography and the fluid flow is highly idealized using heuristic rules.

Bridge (1993) shows how models of braid kinematics can be combined with sediment transport relationships and associated bedforms to predict braided river sedimentology.

2.4   MODELLING OF MEANDERING STREAM FLOODPLAIN DEVELOPMENT

Mackin (1937), Fisk (1944, 1947) and Wolman and Leopold (1957) specified the archetype for a meandering channel floodplain (type B3 of Nanson and Croke, 1992) formed by lateral channel migration with floodplain sediments dominated by sand-to-gravel point-bar deposits and finer sand, silts and clays deposited by overbank flooding. Subsidiary facies include sandy splays from levee crevasses, fine-grained fills in abandoned oxbow lakes, and organic-rich backswamp deposits.

The leading processes in the development of meandering channel floodplains are the lateral migration of meander loops and attendant cutoffs (Figure 2.6). The primary causative factor for meander development is the secondary circulation created by channel planform curvature, which leads to asymmetry of flow and bed topography in bends. Flow momentum causes greater shear stresses on the outer bank in a long bend, leading to enhanced erosion and concomitant point-bar deposition on the inside bank. This curvature-induced flow asymmetry is a sufficient explanation for development of

Figure 2.6 (*opposite*)   An unknown tributary of the Yukon River, Alaska. Note recent cutoff and oxbow lake at lower left and impending neck cutoff in lower centre of picture. Flow is from top to bottom

meanders (Ikeda *et al.*, 1981; Howard, 1984), although flow patterns associated with alternate bars may aid in initial stages of bank erosion (Lewin, 1976, 1978; Nelson, 1990). Bank erosion may also be affected by channel depth, which is also affected by secondary circulation. As will be discussed more fully below, cutoffs (and sometimes limited valley width) prevent infinite growth of meander bends.

The earliest models of meandering planforms were static simulations utilizing constrained random walks (Ferguson 1976, 1977). The Ferguson disturbed periodic model demonstrated that meanders can be considered to be generated by a periodic migration with superimposed irregularity. However, such static simulations cannot be used to simulate planform evolution or the topography and stratigraphy of the floodplain.

Prediction of channel migration requires models of the processes controlling bank erosion as well as flow patterns and bed topography in the meandering channel. Important observations of migration rates in natural channels were obtained by Hickin and Nanson (1975, 1984), Nanson and Hickin (1983, 1986), Hooke (1987), and Biedenharn *et al.* (1989), who showed that migration rates increase with bend curvature, reaching a maximum value at R/W (R is mean bend radius and W is channel width) of about 2–3, and decreasing rapidly for tighter curves (even becoming negative in the sharpest curves). However, Howard (1984), Howard and Knutson (1984), Parker (1984) and Furbish (1988) show that lateral migration rates must be an integral function of both local and upstream curvatures. Howard and Knutson (1984) and Ferguson (1984) introduced heuristic models relating local migration rates to weighted integrals of local and upstream curvatures. In addition, Odgaard (1987), Pizzuto and Meckelnberg (1989) and Hasegawa (1989a,b) found that spatial variations in migration rates in natural channels are linearly related to the magnitude of near-bank velocity.

These sets of observations relating relating migration rates to channel curvature and to near-bank velocities were shown to be consistent as models of flow in meandering channels were developed. In a seminal paper Ikeda *et al.* (1981) provided an analysis that showed that the cross-sectional velocity distribution and cross-sectional channel slope can be expressed as a function of local and upstream planform curvatures. Howard (1984), Howard and Knutson (1984), Parker (1984) and Furbish (1988) demonstrated that the Ikeda *et al.* (1981) model coupled with the assumption that bank erosion is proportional to near-bank velocity produces realistic simulated meanders with a relationship between rate of meander loop migration and loop curvature that is similar to that observed by Hickin and Nanson (1975).

2.4.1 Modelling meander migration and floodplain deposition

A simulation model of flow and bed topography in meander bends was coupled by Howard (1992) with simple assumptions about the rate of bank erosion and sediment deposition rates in the stream channel and on the floodplain to simulate the topographic evolution of floodplains. This model is briefly summarized below, together with extensions of the original model to permit the possibility of chute cutoffs and channel aggradation, as well as spatially variable bank erosional resistance due to valley walls or oxbow lake clay plugs.

Within-channel flow and bed topography

Several models have been developed to predict bed topography and flow structures in channels with arbitrary planform. The models of Ikeda *et al.* (1981), Johannesson and Parker (1989), Odgaard (1989a,b) and Parker and Johannesson (1989) are linearized one-dimensional (along-stream) models that treat downstream variations in velocity and depth structure explicitly but make simplifying assumptions about cross-sectional variations. A model of this type (the Johannesson and Parker, 1989, abbreviated J&P) was utilized by Howard (1992) for reasons of computational efficiency, although more general models exist that explicitly treat cross-sectional variations in depth and velocity (e.g. Nelson and Smith, 1989a,b). The J&P model treats only time-averaged, curvature-forced bed variations and thus does not account for migrating dunes and alternate bars. Local depth (h) and downstream vertically averaged velocity (u) are resolved into section means (H and U) and a dimensionless perturbation (h_1 and u_1):

$$u = U(1 + u_1) \qquad (2.5)$$

$$h = H(1 + h_1) \qquad (2.6)$$

Near-bank values of the perturbations are indicated by h_{1b} and u_{1b}. At the channel centreline u_1 and h_1 are assumed to be zero. The model also makes the simplifying assumptions that H, U and channel width, W, are constant downstream, that depth and depth-averaged velocity vary linearly across the channel, and that there are negligible sidewall effects on near-bank flows.

Bank erosion rate law

Howard (1992) notes that four processes may limit the rate at which channel banks erode to produce channel migration: (1) the rate of deposition of the point bar; (2) the ability of the stream to remove the bedload component of the sediment eroded from banks; (3) the rate of detachment of *in situ* or mass-wasted bank deposits; and (4) the rate of weathering of bank sediment. A similar assessment was made by Ikeda (1989). Because banks in meandering channels are generally cohesive, bank erosion is slow enough that (1) is seldom limiting. Constraint (2) would be appropriate for readily disaggregated banks composed dominantly of bedload-sized sediment. Even though stream banks are commonly initially undermined by erosion of bed sediment exposed at the foot of cut banks (Laury, 1971; Thorne and Lewin, 1979; Thorne and Tovey, 1981; Thorne, 1982; Pizzuto, 1984; Ullrich *et al.*, 1986; Osman and Thorne, 1988; Thorne and Osman, 1988), the cohesive upper bank sediment must be subsequently detached before further bank erosion can occur, governed by constraint (3). If banks are indurated then erosion may be limited by subaerial weathering, so that erosion rates may not be directly related to flow characteristics (constraint 4). All four constraints may vary in importance among streams, from place to place along a given stream, and through time at a given location.

Howard (1992) suggests a general relationship relating lateral bank erosion rate, $\partial n/\partial t$, to the near-bank velocity and depth perturbations:

$$\partial n/\partial t = E(au_{1b} + bh_{1b}) \qquad (2.7)$$

where a and b are coefficients and E is intrinsic bank erodibility. Most studies have suggested that the velocity perturbation is the important factor, so that $a = 1$ and $b = 0$ (Ikeda *et al.*, 1981; Beck, 1984; Beck *et al.*, 1984; Howard and Knutson, 1984; Parker, 1984; Parker and Andrews, 1986; Odgaard, 1987; Hasegawa, 1989a,b; Pizzuto and Mechelnberg, 1989). Hickin and Nanson (1984) and Nanson and Hickin (1986) relate bank erodibility to median grain size at the base of cut banks in the deeper scour holes. This suggests that bank erosion should be a positive function of depth (positive b in Equation 2.7); a similar assumption was made by Odgaard (1989a,b). On the other hand, Hasegawa (1989a,b) shows that if constraint (2) is governing, then b in Equation 2.7 should be negative; this occurs because the quantity of eroded bank sediment is proportional to channel depth.

If erosion rates are related to h_{1b} rather than u_{1b}, then meanders would tend to grow in amplitude with little downstream translation, because the depth perturbation is nearly in phase with, or may even lead the curvature for flow and sediment characteristics that are characteristic of natural channels (Howard, 1992). By contrast, u_{1b} lags changes in curvature significantly, so that maximum outer-bank velocities are downstream of the meander axis of symmetry, resulting in downstream migration. The simulations conducted below assume that $a = 1$ and $b = 0$.

Floodplain deposition

The third major component of the Howard (1992) model is an assumed rate law for floodplain sedimentation. Several studies have shown that the rate of overbank deposition on floodplains diminishes with floodplain age and floodplain elevation, because the higher locations are less frequently and less deeply flooded (Wolman and Leopold, 1957; Everitt, 1968; Nanson, 1980; Hupp and Bazemore, 1993). Further-more, floodplain deposition rates diminish with distance from the channel supplying water and sediment to the floodplain (Kesel *et al.*, 1974; Pizzuto, 1987; Walling *et al.*, 1992, Chapter 12, this volume; Mertes, 1994). Similar patterns have been observed in tidal marshes (Allen, 1990, 1992; French, 1993; French and Spencer, 1993; Falconer and Chen, Chapter 11, this volume). These observations were incorporated into a heuristic relationship for long-term overbank sedimentation rate, ϕ, by Howard (1992):

$$\phi = (E_{max} - E_{act})[\nu + \mu \exp(-D/\lambda)] \tag{2.8}$$

where E_{max} is a maximum floodplain height, E_{act} is the local floodplain elevation, ν is a position-independent deposition rate of fine sediment, μ is the deposition rate of coarser sediment by overbank diffusion, λ is a characteristic diffusion/advection length scale, and D is the distance to the nearest channel (both measured in channel-width equivalent units). The upper limit to deposition (E_{max}) in the Howard (1992) model is a rather arbitrary limit to maximum floodplain height. For the present contribution the deposition rate was given a more general formulation:

$$\phi = [\nu + \mu \exp(-D/\lambda)]\exp[-\gamma(E_{act} - E_{bed})^w], \tag{2.9}$$

where ν, μ, λ, D and E_{act} are as above, E_{bed} is the mean bed elevation, γ is the elevation decay rate, and w is an exponent, set in these simulations to 2.0. If $E_{act} < E_{bed}$ then the

second exponential term is set to unity. Thus there is no absolute upper elevation for sediment deposition.

Sedimentation occurs generally over the floodplain using Equation 2.9, but locations occupied by the channel have elevations determined by the flow and bed elevation model. When a channel migrates through and beyond a given floodplain cell, the initial elevation on which floodplain sedimentation occurs is the near-bank bed elevation of the channel predicted by the flow model on the side opposite to the direction of migration. This initial elevation is set to

$$E_{act} = E_{bed} + \eta_b \qquad (2.10)$$

where η_b is the relative bed elevation predicted by the flow model at the bank opposite to the direction of migration.

In the simulations reported below the deposition rate ϕ is assumed to be a long-term average resulting from the natural spectrum of flood flows; in other words, the effects of individual floods are not explicitly modelled, and all floods are assumed to result in net deposition.

In summary, the depositional model incorporates both a crude model of point-bar sedimentation expressed as a variable advancing bank initial elevation (Equation 2.10) and an overbank (vertical accretion) component (Equation 2.9).

2.4.2   Simulation results

This section summarizes the results of the Howard (1992) model, and subsequent sections explore modifications to that model. When the initial stream is straight except for small, random normal perturbations (Figure 2.7a), the early meandering pattern is very regular and nearly sinusoidal, but rapidly develops a classic very sinuous, upstream-skewed shape prior to cutoff. This shape is characteristic of the solutions to the governing equations (Parker, 1984; Parker and Andrews, 1986) and common in natural streams (Carson and Lapointe, 1983). Local differences in loop growth rate are due to the random perturbations of the initial input stream. After neck cutoffs begin (when the centreline of two channel segments approach within 1.5 channel widths), however, the stream pattern becomes much more varied in the form of the meanders, due to the disturbances that propagate throughout the meander pattern as a result of cutoffs. At these advanced stages, the pattern becomes much more similar to natural meandering streams with their commonly complex loop shapes (Figures 2.7b and 2.7c). As a result of chance occurrence of two or more cutoffs on the same side of the valley, the overall meander belt can develop a wandering path, as noted by Howard and Knutson (1984). The sharp bends that result from cutoffs are very rapidly converted to more gentle bends, commonly by reverse migration caused by maximum flow velocities occurring on the inside of very sharp bends. The development of varied meander forms from an initially regular pattern indicates that the combination of meander growth, the occurrence of cutoffs, and the influence of complicated initial and boundary conditions (including variations in bank resistance which are here held constant) implies a "sensitivity to initial conditions". That is, small differences in initial geometry or boundary conditions between two otherwise identical streams will cause different meander patterns. Also, predictability of future meander patterns decreases with time.
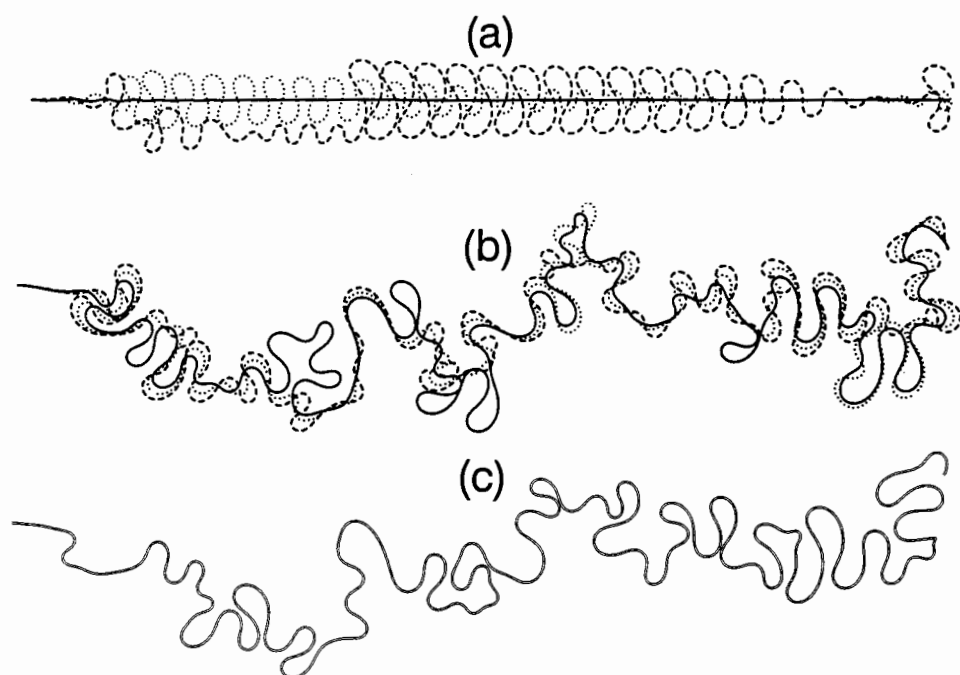
Figure 2.7 Simulated meander planform development. (a) Channel centrelines after 250 (solid), 500 (dotted) and 750 (dashed) iterations; (b) Channel centrelines at 4400, 4600 and 4800 iterations; (c) final channel at 5000 iterations

Floodplain age and elevation for the deposition model using Equations 2.9 and 2.10 are illustrated in Figure 2.8 (Plate II), which portrays the evolution of the central region of the stream shown in Figure 2.7. White areas have not been occupied by the meandering channel during the simulation. Similarly, Figure 2.9 (Plate III) shows meander and floodplain evolution for confined meanders developed between non-erodible valley walls. Confined meanders develop a characteristic asymmetric pattern with gentle bends terminating abruptly in sharp bends at valley walls, similar to natural cases (Lewin, 1976; Lewin and Brindle, 1977; Allen, 1982; Howard and Knutson, 1984).

These simulations exhibit many of the features of natural meandering streams, including overbank deposits gradually increasing in elevation away from the channel, rapid isolation of abandoned channels by filling near the main channel (modelled here as resulting from sediment diffusion from the main channel, but in natural channels advectional transport through the abandoned channel would also occur), and slower infilling of oxbow lakes primarily by deposition from suspension. Channel migration rates vary considerably from bend to bend, and the slope of the floodplain surface in the interior of bends is generally steeper the slower the migration. For high rates of overbank deposition (Figures 2.8b and 2.9b) most of the floodplain rapidly reaches a nearly uniform elevation and cutoffs are rapidly isolated into oxbows. The simulations with a low rate of overbank sedimentation (Figure 2.8c) exhibit a depositional feature that is a consequence of out-of-phase relationships between near-bank velocity and bed
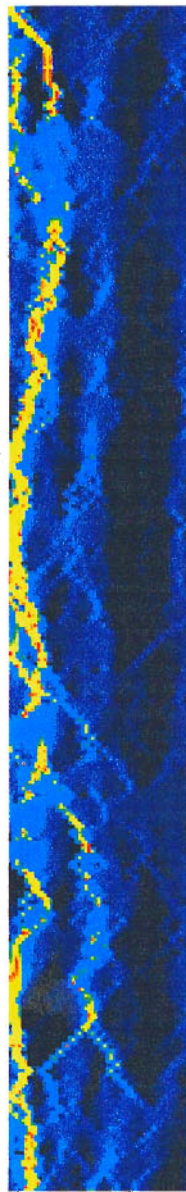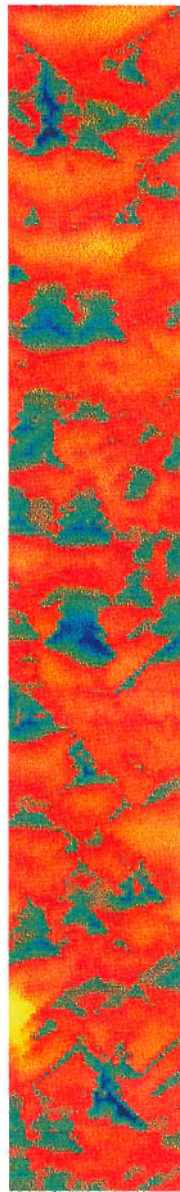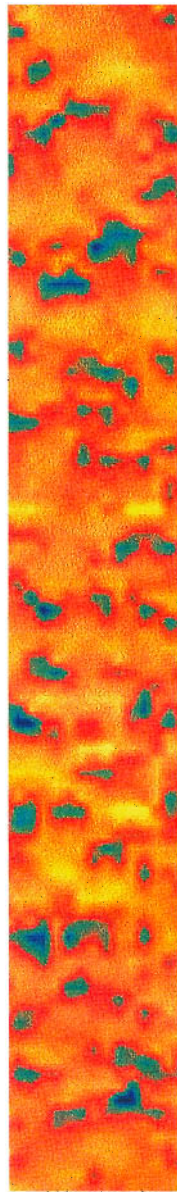
Plate 1 Figure 2.5 Simulation model of braided channel development. (a) Discharge over bed topography shown in (b); (b) final bed topography; (c) bed topography at an early stage of development. Areas of high topography and high discharge are yellow, low values are blue and green. Simulation is based upon the model of Murray and Paola (1994)
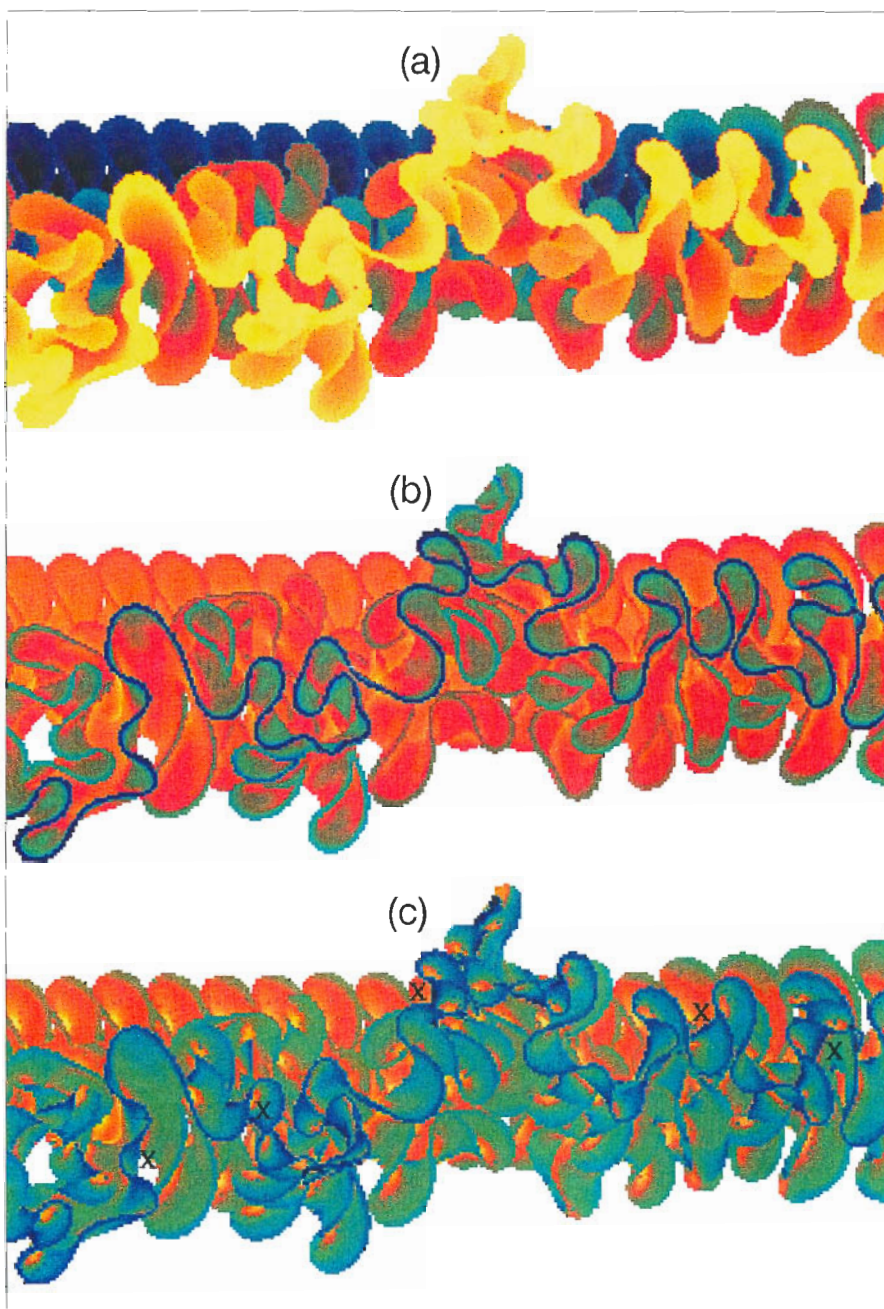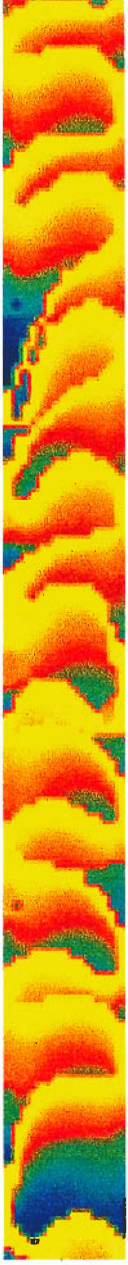
Plate II Figure 2.8 Map view of floodplain age (a) and floodplain elevation under conditions of high (b) and low (c) rates of overbank deposition. There are no lateral constraints on channel migration and no chute cutoffs in this simulation. Yellow areas are young in (a) and high areas are red and yellow in (b) and (c). All figures have been given a non-linear contast stretch to provide greater detail in young or high areas. This figure covers the central portions of the simulation shown in Figure 2.7

(a)

(b)
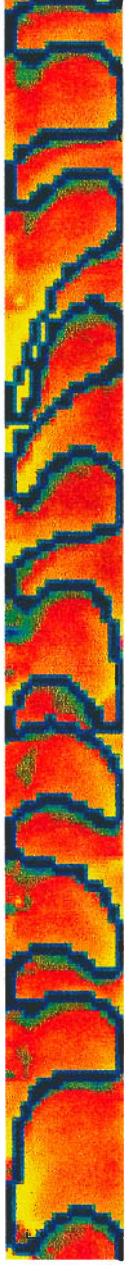
(c)

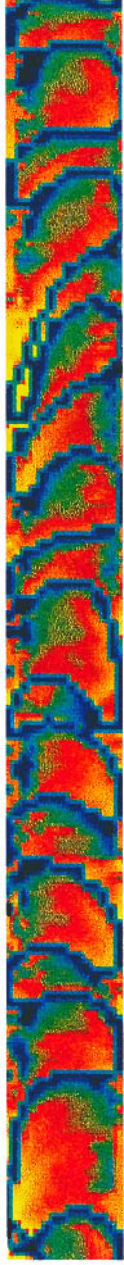**Plate III Figure 2.9** Map view of floodplain age (a) and floodplain elevations (b) with high rate of overbank deposition. Fixed valley walls restrict meander enlargement

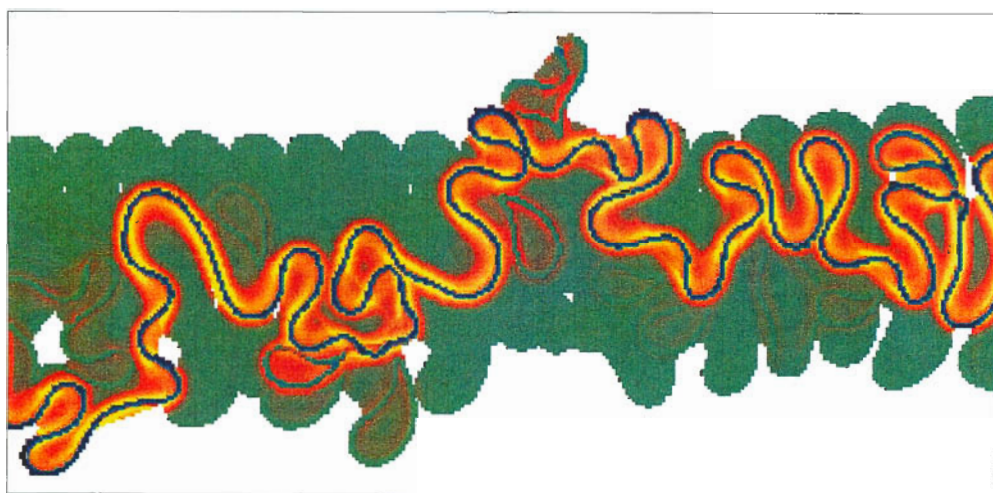**Plate IV Figure 2.14**   Floodplain elevation (high areas yellow) in a simulation similar to that shown in Figure 2.8b but with superimposed channel aggradation. Picture is contrast stretched to enhance portrayal of high areas

elevation perturbations. Where curvature changes abruptly downstream, the depth adjusts quite rapidly on the new outer bank, and generally overshoots its value for constant curvature, but velocity responds more slowly and migration is directed towards the inside of the bend. This means that, for short sections at the beginning of a sharp bend, deposition on the newly created floodplain on the outside bank must start from very low relative elevations (a scour hole). This zone of very low initial elevations is only about 2.5 width equivalents in length, followed by the more normal pattern of migration towards the outside bank. These short zones with lower-than-average initial floodplain elevations are located in consistent positions relative to bends as the channel migrates, in places leaving behind depressions, or sloughs, in the floodplain deposits. These sloughs are most commonly located in the axial position of sharp meander bends, and are best developed on the downstream end of the point bar near the curvature inflection leading to the next bend. Several of these depressions are labelled with an "X" in Figure 2.8c. These depressions are best developed in short abrupt bends near the site of recent neck cutoffs (Figure 2.8c) or where meander migration is confined. Howard (1992) points out that natural and experimental channels abound in similar features which may result from a mechanism similar to that incorporated in the model (e.g. the experimental runs of Friedkin (1945) and Wolman and Brush (1961)). Fine sediments accumulating in such sloughs and floodplain lows have been called concave bank benches or counterpoint bars (Cary, 1969; Woodyer, 1975; Hickin, 1979; Nanson and Page, 1983). Lewin (1978) attributes floodplain sloughs extending upstream from the outside bank of confined and unconfined sharp bends to formation as residual depressions from migrating deep scour pools – essentially the same mechanism as occurs in the simulations. These sloughs are best developed when vertical accretion deposition rates are modest (Nanson and Page, 1983). If this is not the case (large values of ν and/or μ in Equation 2.9), then bank and floodplain deposition is more uniform (Figures 2.8b and 2.9b).

The simulations suggest that bar and low floodplain features of wide meandering streams cannot be solely understood in terms of adjustments of bar form to contemporary flow and sediment transport, but are in addition intimately related to the kinematics of bank migration. The topography of these features should affect location of chute cutoffs, as is explored below.

Cutoffs

Loop cutoffs in the Howard (1992) model were limited to *neck cutoffs*, occurring when the growth of two loops cause mutual intersection (the criterion in these simulations was centreline approach to within 1.5 channel widths). The resulting channel pattern is very sinuous ($\mu > 3.5$, as shown in Figures 2.7 and 2.8); such highly sinuous meandering is rare in natural channels. The model has been revised to incorporate probabilistic *chute cutoffs*, which are longer flow diversions which generally result from formation and enlargement of a depression across a point-bar complex. Because the chutes have steeper overall gradient than the existing meander bend, the chute can often gradually increase in size during successive floods until it eventually accommodates almost all of the flow and the former channel becomes choked with deposited sediment. The cutoff model does not model the flood flow across the point

bar which creates the chute, but rather predicts the influence of planimetric and relief properties that determine where and when chute cutoffs might occur. Five such properties have been incorporated into the model: (1) the ratio R_c of gradients across the potential chute path to existing channel gradient; (2) the distance D_c across the potential cutoff; (3) the elevation E of the floodplain across which the chute develops; (4) the planimetric angle ψ between the existing channel direction and the path of the chute; and (5) the relative magnitude of the near-bank velocity at the site of the potential chute. The probability P of chute development at a given time and across a given chute path is expressed as:

$$P = K_c R_c e^{(-K_d D_c - K_e E + K_a \cos \psi + K_v u_{1b})} \tag{2.11}$$

where

$$R_c = (D_p - D_c)/D_p = (\mu_c - 1)/\mu_c \tag{2.12}$$

The gradient ratio R_c depends upon the relative distances (and hence also overall gradients) across the existing flowpath D_p and across the potential chute D_c, and it can also be expressed in terms of the cutoff sinuosity $\mu_c = D_p/D_c$. In evaluating Equation 2.12 the downstream channel gradient is assumed to be uniform. The terms in the exponential reflect the presumed smaller probability of cutoff across longer chute lengths and floodplains with higher elevation (E can be specified either as the maximum or average elevation across the chute, measured relative to the average bed elevation). Additionally, the probability of cutoff is assumed to decrease as the angle between the existing flowpath and the cutoff path increases. Finally, the probability of cutoff is assumed to increase if the near-bank velocity perturbation u_{1b} at the site of potential capture is large and positive. The coefficients K_c, K_d, K_e, K_a and K_v are assumed to be temporally and areally invariant. The sum of the exponent terms in Equation 2.11 is restricted to values ≤ 0.

Although there have been a number of studies of the relative frequencies of neck and chute cutoff in natural meandering channels and of their temporal development, no systematic observations of cutoffs in natural meandering channels have been made which would permit evaluation of the relative importance of the coefficients in Equation 2.11 or of the adequacy of the assumed functional form. Lewis and Lewin (1983) provide most comprehensive observations. Over half of the observed cutoffs in the Welsh rivers observed occurred as chutes. Cutoff locations vary from chord sites (across the point bar), *axial* sites (middle of loop), and *tangential* sites (loop-to-loop, connecting outer bends). Most cutoffs occurred across tight bends at positions from axial to tangential.

For the present contribution a number of simulations have been conducted to explore the characteristics of cutoffs and channel form resulting from the application of Equation 2.11. At regular intervals during the simulation (every 50 iterations in the present case) each point along the stream is examined in turn for the possibility of chute cutoff, progressing downstream. At each point all downstream locations up to a practical limit of 50 width equivalents are evaluated with respect to the probability P of chute development. For that point the potential cutoff path with the greatest probability P of chute development is compared with a randomly generated number from a uniform distribution. If that random number is $\leq P$, then the chute cutoff occurs. Locations

downstream from the cutoff are also examined in the same manner for additional cutoff. The probability P is assumed to be unity if D_c is less than 1.5 channel widths (a neck cutoff). In the model the chute cutoff is assumed to be instantaneous relative to the simulation time-scale (although natural chutes require an appreciable time to fully divert the main-stem flow).

A number of simulations were made with a range of values for the coefficients in Equation 2.11, generally by varying one coefficient while holding the others constant at zero, except that K_c was held at unity for runs with varying K_d, K_e, K_a or K_v. In addition, for runs with varying K_a or K_v the value of K_d was set at a value producing an intermediate degree of sinuosity of $\mu \approx 2$. Some of the simulated planforms are illustrated in Figure 2.10. The planforms are plotted at times just before the occurrence of cutoffs is allowed (every 50 iterations); this was done so that the very sharp bends that occur immediately after cutoffs would be smoothed by some channel migration. As a result, the plotted figures slightly exaggerate the average sinuosity during the simulation. By varying the coefficients any degree of overall sinuosity from near unity

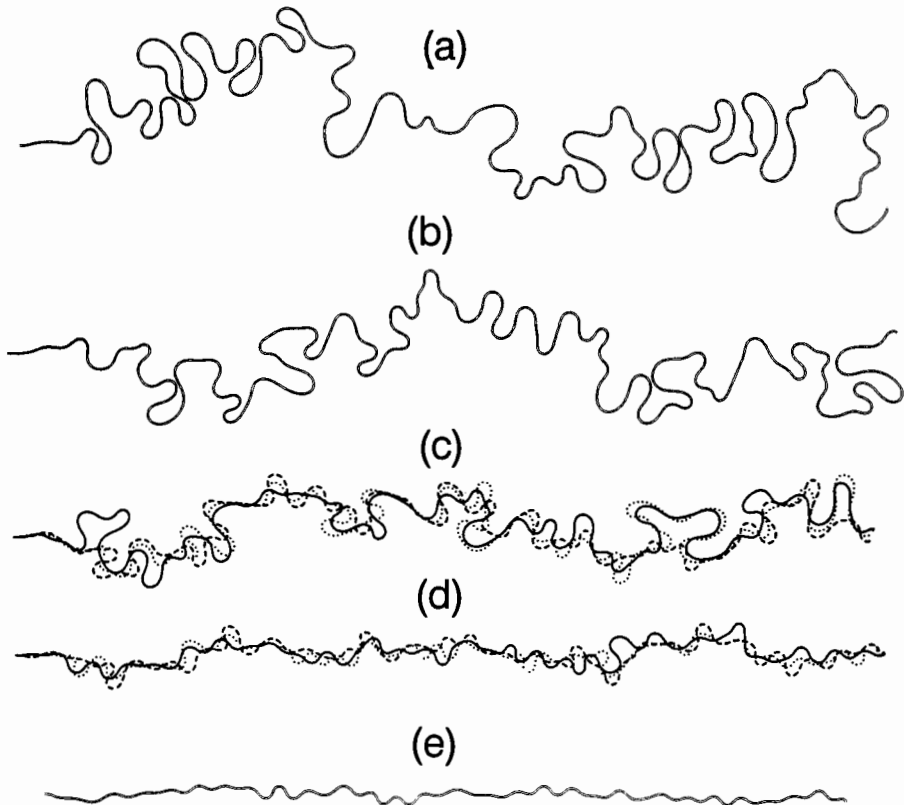**Figure 2.10**   Final channel planforms after 5000 iterations created by model with chute cutoffs controlled by Equation (2.11). For these simulations $K_c = 1$, $K_e = 0$, $K_a = 0$, $K_v = 0$ and K_d takes the following values: (a) ∞ (no chute cutoffs); (b) 0.3; (c) 0.2; (d) 0.1; (e) 0.03. Planforms at successive 200 iteration intervals are shown in (c) and (d). Statistical characteristics of cutoffs are shown in Table 2.1

to greater than three can be obtained, and in general the planforms for simulations with restricted sinuosity resemble natural examples with the same degree of sinuosity. As the probability of chute cutoffs is increased, not only does the sinuosity diminish, but the width of the meander belt diminishes.

Cutoff characteristics As K_c increases and K_d or K_e decreases, chute development becomes more probable and the planform becomes less sinuous (Table 2.1). However, variations in K_a or K_v have little effect on total sinuosity. Decreases in sinuosity are accompanied by decreases in the averages of cutoff path length D_p, cutoff sinuosity μ_c, and cutoff angle ψ, whereas the frequency of cutoffs and the average chute length D_c increase. Variations in K_c, K_d, K_e and K_a have little effect on the average value of the velocity perturbation u_{1b} at the cutoff sites; also, as noted above, variations in K_v have

Table 2.1 Summary statistics of chute cutoffs

Input parameters[*]					Summary statistics[†]							
K_c	K_d	K_e	K_a	K_v	μ	μ_c	f	D_c	D_p	E	ψ	u_{1b}
1.0	∞	–	–	–	3.56[‡]	–	–	–	–	–	–	–
1.0	0.43	–	–	–	3.17	18.89	0.024	3.69	57.5	13.4	84.9	0.02
1.0	0.30	–	–	–	2.88	8.64	0.050	5.31	36.2	13.9	77.3	−0.03
1.0	0.20	–	–	–	1.98	3.73	0.089	7.28	24.6	13.6	67.0	−0.01
1.0	0.10	–	–	–	1.39	1.50	0.171	9.17	13.9	13.7	48.3	0.00
1.0	0.09	–	–	–	1.22	1.43	0.181	9.55	13.6	13.8	43.0	−0.01
1.0	0.03	–	–	–	1.10	1.16	0.179	13.53	15.6	14.2	23.8	−0.02
1.0	0.06	0.15	–	–	2.14	5.27	0.069	7.8	28.3	6.9	66.1	−0.02
1.0	0.06	0.10	–	–	1.73	2.33	0.107	8.78	19.0	10.6	62.6	0.00
1.0	0.06	0.07	–	–	1.58	1.79	0.137	9.44	16.7	11.5	55.8	0.01
1.0	0.06	0.05	–	–	1.42	1.61	0.152	9.40	15.1	12.4	51.4	−0.01
1.0	0.06	0.035	–	–	1.29	1.49	0.157	10.04	14.8	13.6	45.7	0.03
1.0	0.20	–	0.5	–	1.94	3.11	0.104	7.16	20.7	13.5	53.2	0.00
1.0	0.20	–	1.0	–	2.19	2.51	0.126	7.26	17.7	13.0	46.4	0.00
1.0	0.20	–	2.0	–	1.69	1.71	0.193	7.42	13.2	12.4	38.2	0.00
1.0	0.20	–	–	4.0	2.41	2.86	0.113	8.03	22.1	13.7	60.4	0.30
1.0	0.20	–	–	2.0	2.40	3.09	0.102	7.42	21.1	14.1	65.4	0.17
1.0	0.20	–	–	1.0	2.05	3.20	0.091	7.34	22.1	13.5	68.5	0.11
1.0	0.20	–	–	−1.0	2.04	3.34	0.097	7.25	22.0	13.6	67.8	−0.09
1.0	0.20	–	–	−2.0	2.08	3.32	0.086	7.41	22.5	13.5	65.5	−0.18
1.0	0.20	–	–	−4.0	2.30	2.44	0.117	7.89	17.2	14.0	58.6	−0.31
0.8	0.03	–	–	–	1.11	1.20	0.185	12.67	15.1	14.5	23.4	0.00
0.5	0.03	–	–	–	1.14	1.27	0.163	13.07	16.4	15.2	29.3	0.01
0.25	0.03	–	–	–	1.19	1.42	0.133	12.76	17.9	15.4	37.2	0.00
0.1	0.03	–	–	–	1.49	1.81	0.102	12.31	21.4	16.4	48.3	0.03
0.05	0.03	–	–	–	1.41	2.31	0.081	12.41	26.3	16.6	55.8	−0.01
0.025	0.03	–	–	–	1.69	2.92	0.061	12.47	32.0	16.9	58.3	−0.01
0.01	0.03	–	–	–	2.41	4.16	0.038	12.12	39.9	17.0	67.8	0.03

[*] Input parameters defined in Equations 2.11 and 2.12).
[†] Average values of cutoff properties. Properties defined in Equations 2.11 and 2.12 except for f, which is the frequency of cutoffs (number per iteration).
[‡] Run with only neck cutoffs.

little effect on sinuosity, although it would be expected that high K_v would favour axial to tangential cutoffs, because u_{1b} is likely to be positive on the outer bank. Similarly, a high weighting on K_a would tend to favour tangential cutoffs, because ψ is small. Simulations with cutoffs due to high K_c resulted in smaller cutoff frequency, larger D_c, larger D_p and greater E than simulations of similar sinuosity resulting from high values of K_c, K_d, K_e or K_a. This is because the lack of restriction on distance or elevation across the cutoff for the K_c runs leads to longer cutoffs and K_c has to be set to a low value (low frequency of cutoffs) in order to maintain sinuosity appreciably greater than unity. Increases in sinuosity with variations in K_c and K_d are accompanied by increases in the average and maximum elevations across the cutoffs, whereas, as would be expected, average and maximum elevations decrease with increasing sinuosity as K_e is increased.

Planimetric and elevation characteristics of floodplains with chute cutoffs An obvious characteristic of simulations with frequent chute cutoffs is the restricted width of the meander belt (Figure 2.10). Even though there are no lateral restrictions imposed upon the simulated meandering (other than the fixed point at the upstream end of the simulated channel), when chute cutoffs are frequent the meander belt increases in width very slowly, and the average elevation of the floodplain is higher for a given average distance from the active channel. This is because each site within the meander belt stays closer to the active channel on the average.

Both the amplitude and wavelength (measured *following* the channel centreline) are smaller for meander planforms dominated by chute cutoffs as compared to those solely shortened by neck cutoffs for equivalent flow, bed topography and bank erosion parameterization. This results because meanders grow primarily in amplitude rather than along-valley wavelength (Howard and Knutson, 1984; Parker, 1984; Howard, 1992), and chute cutoffs primarily limit amplitude growth. Furthermore, the disturbances in the planform caused by cutoffs encourages growth of new meander loops. In addition, the simulations assume that the overall valley gradient is fixed over the time-scales necessary for full development of the meander pattern. As a result, the channel gradient is inversely related to sinuosity. Thus high sinuosity planforms (no chute cutoffs) have low velocity and relatively large channel depth; the preferred wavelength of meandering increases with increase in channel depth (Howard and Knutson, 1984).

Spatially variable bank erodibility

The simulations reported above assume that the erodibility of bank materials is spatially uniform. However, the pattern of meandering may be affected by at least two types of non-uniformity. The first of these occurs where meandering locally impinges on resistant materials such as bedrock or terraces bordering the active floodplain. This is termed confined meandering (or restricted meandering by Ikeda, 1989). Lateral migration may also be locally inhibited by exposure of resistant floodplain sediments, most commonly the cohesive sediments (clay plugs) deposited in oxbow lakes by overbank deposition (Fisk, 1947; Ikeda, 1989; Thorne, 1992), but sometimes older bankswamp deposits (Ikeda, 1989; Thorne, 1992). Ikeda (1989) calls this case confined

free meandering. Simulations incorporating spatially variable bank resistance are reported below.

Confined meanders   Narrow meander belts confined within valley walls or high terraces occur commonly. Meanders in such situations commonly develop an asymmetric pattern, with gentle bends deflecting abruptly at valley walls (Lewin, 1976; Lewin and Brindle, 1977; Allen, 1982). The effects upon meander pattern of a narrow floodplain was simulated by Howard and Knutson (1984) and Howard (1992) by the simple expedient of disallowing erosion beyond a fixed belt surrounding the centre of the floodplain (Figure 2.9), which is a reasonable model for some underfit valleys. The effect of resistant sidewalls is investigated using a more general approach in this contribution (a similar approach is used by (Sun *et al.*, 1996). The spatial variation in bank erodibility is explicitly modelled by overlaying a matrix of square cells on the floodplain and surrounding valley walls (or terraces), with each cell having dimensions equivalent to one channel width. Thus a floodplain of arbitrary width and pattern could be set as initial conditions. In the simulations reported here, meandering and floodplain development starts from a nearly straight initial channel with a floodplain width just equalling the channel width. Valley wall erodibility is assigned an areally uniform value of 0.2 times the erodibility of floodplain deposits. If the node representing a given location along the stream is located on a cell marked as a valley wall and if the bank erosion is directed towards the valley wall, then the rate of erosion is governed by the valley wall erodibility. However, if the channel node is located within a floodplain cell or is on a valley wall cell but erosion is directed away from the valley wall, the rate of erosion is proportional to the floodplain bank erodibility. The direction of erosion relative to the valley wall is determined in the following manner. Assume the direction of bank erosion is directed towards the east and north (so that the stream is flowing southeast or northwest). East and north are in the positive I and J directions respectively. If the local cell (I, J) is a valley wall cell and if any of the bordering cells at $(I, J + 1)$, $(I + 1, J + 1)$ or $(I + 1, J)$ is a valley wall cell, then the rate of erosion is governed by the valley wall erodibility, otherwise it is proportional to the floodplain sediment erodibility. Two variations of these rules were utilized; in the first, "strict" rules the bordering valley wall cells must not be occupied by the stream, whereas in the "liberal" rules the bordering cells may be occupied by the stream. Only slight differences in pattern resulted from the two sets of rules. Only the simulations for the strict rules are shown.

Figure 2.11 shows an early and later stage in the enlargement of the meander belt and retreat of the valley walls. In the earliest stages of valley wall enlargement (right side of Figure 2.11a) the channel pattern shows the asymmetric pattern discussed above. As valley walls retreat further, meander loops tend to become locked into reentrants eroded into the valley wall. This pattern is often seen in entrenched meandering. Upon further meander bend enlargement the stream impinges only locally against the valley walls (Figure 2.11b). Where this occurs the meander enlargement commonly becomes constrained against projections in the valley wall, resulting in highly distorted meander loops whose axes often point upstream. In some cases these distortions of the meander pattern induce neck cutoffs earlier than if the meandering were unconstrained, but in some locations highly sinuous loops form due to intervening valley walls. In the above
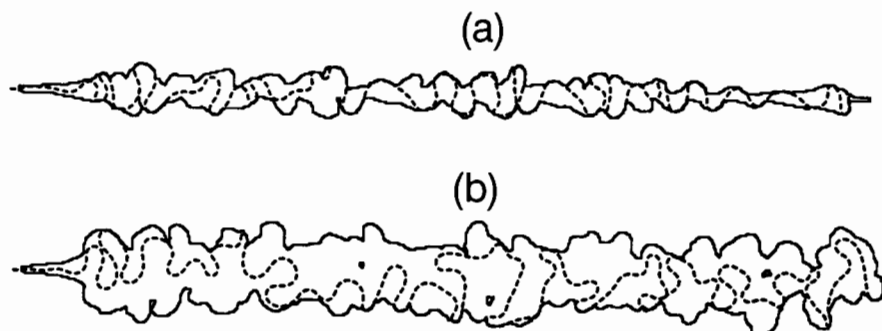
(a)

(b)

Figure 2.11 Successive stages (a and b) of valley widening and planform development with valley walls five times less erodible than active floodplain. Chute cutoffs are not permitted

simulations chute cutoffs are prohibited; when such cutoffs are permitted, the rate of floodplain enlargement is slower, the highly sinuous distorted loops are rare, and the valley width is more uniform with few deep reentrants (Figure 2.12).

The simulations do not show any tendency for development of "underfit" valleys, that is, a long wavelength valley meandering (generally with valley walls of fairly uniform width) with shorter wavelength channel meandering. Such a pattern is fairly common in nature and is attributed to valley enlargement during a past epoch of very high discharges (e.g. from glacial meltwater) followed by more moderate discharges related to the present meandering (Dury, 1954, 1964a,b). Thus the simulations provide circumstantial support for this interpretation.

Cohesive oxbow lake plugs Other simulations were conducted in which it was assumed that the cohesive sediments deposited in oxbow lakes are less erodible by a factor of five than normal point-bar and overbank sediments. The rules governing determination of local bank erosion rates are the same as for the valley wall simulations. Immediately after a meander loop is cutoff all of the cells along the cutoff path are assigned an erodibility of 0.2 except for portions of the loop closer to the existing channel of three channel-width equivalents. The portions of the cutoff loop close to the existing channel are assumed to be infilled with relatively coarse suspended sediment
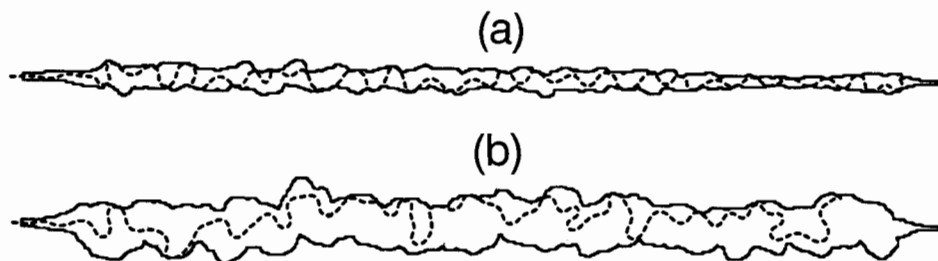
(a)

(b)

Figure 2.12 As in Figure 2.11 but with chute cutoffs permitted

diffusing and advecting from the main channel. Sun *et al.* (1996) utilize a similar approach to investigate the effects of clay plugs.

Figure 2.13a shows the channel pattern and location of resistant clay plug deposits for an advanced stage of meander belt development, under the conditions where only neck cutoffs are permitted. As with the valley wall simulations, impingement on clay plugs creates locally tortuous and distorted meander loops. The pattern of clay plugs over the floodplain becomes very complicated, with some locations evidencing a very heavy areal density of plugs. In some cases, a growing loop penetrates into the interior of a previously abandoned loop, whereupon the loop grows rapidly, more or less duplicating the earlier pattern of loop enlargement. The rate of enlargement then diminishes significantly when the channel reaches the plug on the far side of the filled oxbow.

Although there is some tendency for the channel to become constrained by the development of the resistant clay plugs (Ikeda, 1989), the simulations suggest that the channel can break through the plug belt on occasions, and in some locations the meander belt may migrate laterally, if plugs have formed preferentially on one side of the valley.

The pattern of lateral groundwater flow on the floodplain would be strongly affected by the areal distribution of plugs. As can be seen from the map, many floodplain regions are surrounded by one or more plugs; these would be expected to have relatively high water tables. On the other hand, in some locations a path extends from the active channel well into the floodplain without encountering plugs; such areas should be well drained. In addition, there is a zone surrounding the active channel with relatively few or only short, disconnected plugs that would be well drained.

In the above simulations chute cutoffs were not permitted. When chute cutoffs are allowed (Figure 2.13b), the overall width of the meander belt diminishes and the average size of oxbow plugs is smaller, as discussed above. In addition, the lower sinuosity of the channel system means that the axes of meanders loops and the resultant oxbow plugs are, on the average, nearly perpendicular to the valley axis. This contrasts with the greater variability of oxbow axes when only neck cutoffs occur (Figure 2.13a). Because of the narrower floodplain and more frequent cutoffs, oxbow plugs have a
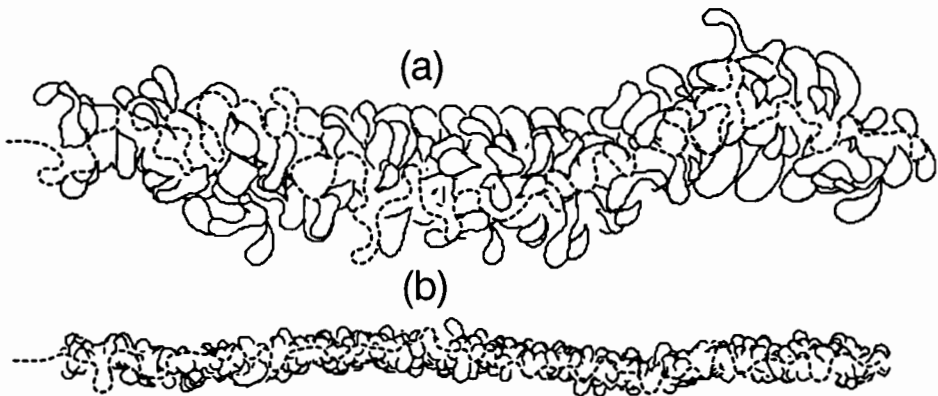

Figure 2.13 Final channel planform (dashed) and oxbow plug deposits (solid) in a simulation in which oxbow clay plugs are assumed to be five times less erodible than point-bar and overbank deposits: (a) after 5000 iterations with only neck cutoffs; (b) after 5000 iterations with chute cutoffs permitted

greater areal density when chute cutoffs are frequent. This does not necessarily mean that the prevalence of plugs would inhibit lateral migration strongly when chutes cutoffs are frequent, because in natural channels chute cutoffs are more common in stream systems with a relatively low supply of cohesive overbank sediments. Thus plugs are probably more erodible than where neck cutoffs predominate.

Natural levees and the effects of aggradation

Simulations of floodplain development by stream meandering under conditions in which the channel bed elevation remains constant fail to produce pronounced natural levees using the deposition model of Equation 2.9, even though the deposition rate is much higher near the active channel than at more distant locations. Slightly elevated banks do occur at crossings (inflection points), where the channel migration rate is very low (Figure 2.8b), but along bends levee development is discouraged because sediment deposited on the outside of bends is eroded by lateral migration soon after deposition, and deposition on the inside, or point-bar locations, occurs at low elevations because of the floodplain youth.

In a number of simulations the mean channel bed elevation was assumed to rise at a constant rate relative to the floodplain elevation. In these simulations pronounced levees were created (Figure 2.14 (Plate IV)). This suggests that channels with high, extensive natural levees (such as the lower Mississippi River) primarily occur where the channel bed is increasing relative to the floodplain level due to aggradation or floodplain lowering (due, perhaps, to sediment consolidation). The other main circumstance producing high natural levees occurs during disequilibrium floodplain sedimentation following catastrophic stripping of floodplain deposits (Nanson, 1986, and previous discussion).

Temporal evolution of the meander belt

The meander simulations typically start from a nearly straight initial stream. As the simulation progresses the stream migrates through an increasingly wide section of floodplain. Initially the growth of the meander belt width (defined as the maximum width of floodplain, measured normal to the valley axis, occupied by the meandering stream during the elapsed time) is slow because of the low sinuosity. As average bend size increases, the average rate of migration and the meander belt width increase rapidly. Cutoffs, however, begin to limit the rate of migration and the rate of meander belt growth slows. Figure 2.15 depicts the overall average width of the meander belt as a function of simulation time (arbitrary units). Meander belt width, W_b, during the nominal run, with no chute cutoffs, is well fit by a logarithmic growth function:

$$W_b = W + K_b \ln(t/t_0), \qquad \text{for } t \geq t_0$$
$$W_b = W, \qquad \text{for } t < t_0$$

(2.13)

where W is the channel width, t is simulation time, t_0 is a "lag time" for initiation of well-developed meandering from a nearly straight initial channel, and K_b is the growth rate constant. The inclusion of chute cutoffs decreases the rate constant, but not the overall logarithmic pattern of growth. Resistant valley walls delay the onset (t_0) and diminish the growth rate (K_b). Resistant oxbow plugs do not form until appreciable
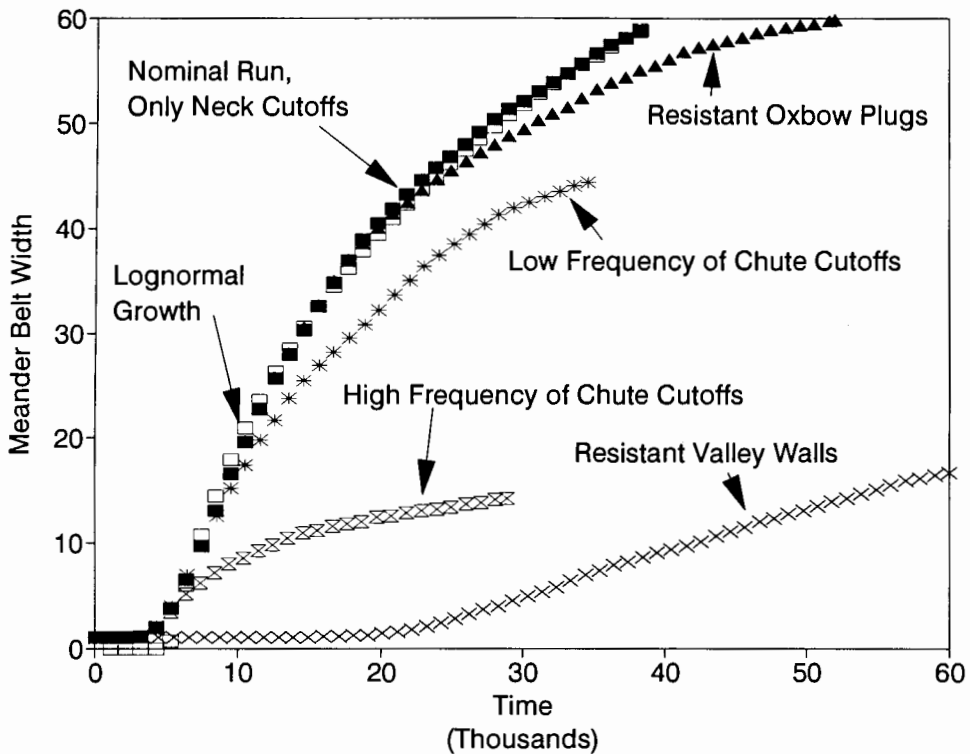
Figure 2.15 Cumulative meander belt width versus simulation time (arbitrary units). Meander belt width is total floodplain width (measured perpendicular to valley axis) occupied by the channel since the start of the simulation. Filled boxes for simulated meandering with uniform bank erodibility and no lateral constraints. Open boxes are a logarithmic growth curve (Equation 2.13) fit to the simulation results. Other curves show effects on growth rate of chute cutoffs, resistant clay plugs associated with oxbow lakes, and resistant valley walls

neck cutoffs occur (after some time $t_n > t_0$), and the growth constant (K_b) decreases for $t_n > t_0$. Ikeda (1989) suggested that meander belts in some rivers may be confined by cohesive fine-grained overbank sediments. The simulations suggest that, even in the absence of lateral constraints or of other controls such as tectonic tilting (Leeder and Alexander, 1987: Alexander *et al.*, 1994), the meander belt width is slow to increase beyond about two to three times the size of the largest meander loops.

The cumulative growth rate of the meander belt is of interest for several reasons: (1) the use of meander belt width to date the valley age; (2) inferring the geometry of sandstone aquifers and petroleum reservoirs; and (3) inferring the storage time of pollutants in fluvial deposits.

Comparison of simulated and natural meander morphometry

Because of the numerous simplifying assumptions made during the development of the meander simulation model, it is important to validate the model. In particular, the

prediction of the pattern of meander migration and attendant cutoffs depends both on assumptions regarding in-channel flow and bed morphology as well as assumptions regarding bank erosion processes. The most direct validation compares predicted patterns of channel migration with those observed; a number of studies have indicated that the combination of the Johannesson and Parker (1989) flow model with the assumption that bank erosion is proportional to the velocity perturbation (Equation 2.7) gives reasonable predictions of short-term migration patterns (Howard and Knutson, 1984; Hasegawa, 1989a,b; Pizzuto and Meckelnberg, 1989; Shane Cherry, personal communication, 1995). However, these techniques are limited by the duration of the historical record of channel migration.

Howard and Hemberger (1991) compared the planform of simulated and natural channel patterns using a suite of morphometric variables. Two linear discriminant functions composed of weighted combinations of these variables are quite successful in separating natural channel planforms (indicated by open circles in Figure 2.16) from those simulated by the Howard (1992) and Howard and Knutson (1984) model (open inverted triangles in Figure 2.16) and Ferguson's (1976, 1977) disturbed periodic model (which simulates meander planforms, but not migration, using an autoregressive approach; open diamonds in Figure 2.16). Discriminant function 1 primarily measures sinuosity and degree of upstream bend asymmetry, whereas function 2 is an index of the irregularity of the channel pattern (more irregular for positive values). The addition of chute cutoffs and spatially variable bank resistance to the model creates meandering patterns that approach closer to those of natural channels. Simulation runs with varying degrees of chute cutoff frequency (filled triangles in Figure 2.16) approach natural channels in having smaller sinuosity and less marked upstream bend asymmetry, but the patterns are still more regular than those of natural channels. Spatially variable bank resistance, due either to valley walls or oxbow lakes (filled boxes in Figure 2.16), both decreases sinuosity and increases pattern irregularity, particularly when combined with chute cutoffs. In some cases simulation runs with spatially variable bank resistance and chute cutoffs are classified as natural channels (Figure 2.16).

Not all meandering channels migrate

The meandering model presented above assumes that lateral migration is continuously active and that point-bar accretion is an important component of floodplain construction. Some highly sinuous channels, however, appear to have little to no active bank erosion. Examples include streams on the intermontane "parks" of the Rocky Mountains (Figure 2.17), channels on the lower coastal plain of the southeastern United States, low-gradient channels in England (Ferguson, 1981, 1987), and possibly meandering tidal channels (Ikeda, 1989). Channel pattern changes over scales of decades are slight to unnoticeable, the floodplains exhibit few scars from past cutoffs, and subaerial point bars are poorly developed or absent (inside banks generally are steep and well vegetated). These floodplains and channels share the following characteristics: (1) low supply rate of bedload; (2) a floodplain that is low and wide compared to channel width; (3) cohesive, strongly vegetated banks and floodplain surface; (4) low valley gradient; and, possibly, (5) a low frequency of large floods. These characteristics suggest that such channels develop from an initially nearly
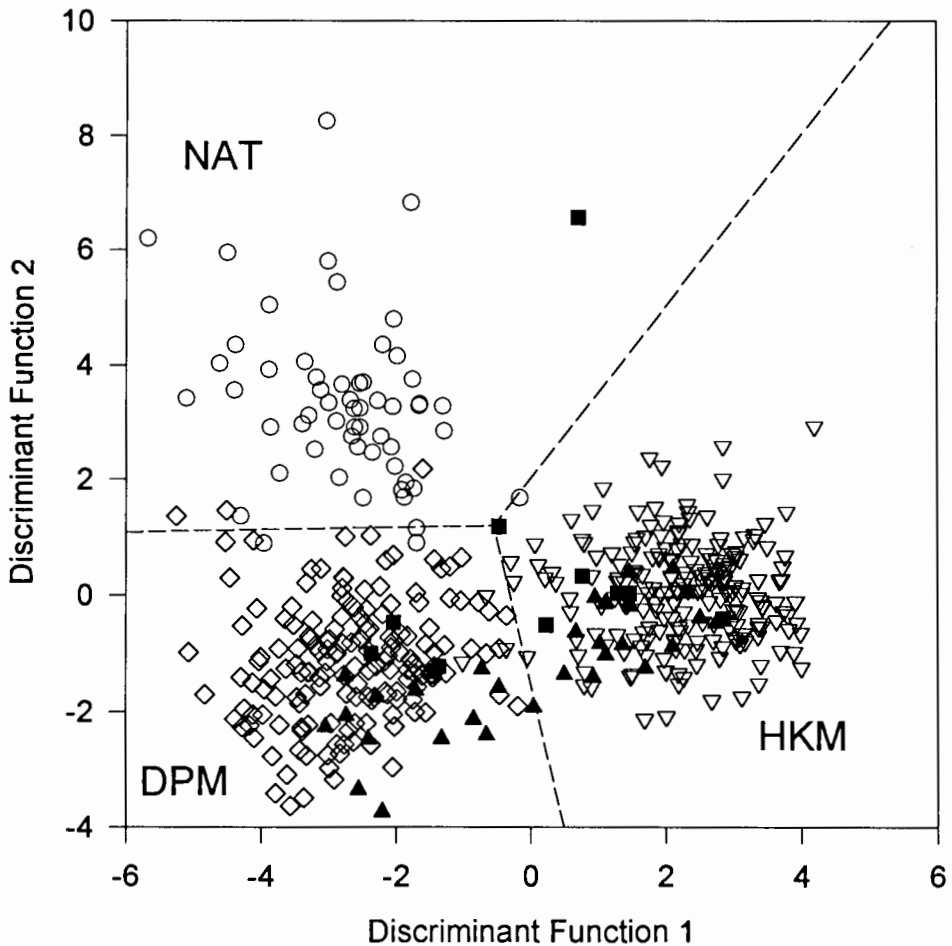
Figure 2.16 Classification of natural and simulated streams using two multivariate discriminant functions. Open circles are natural stream channels (domain classified as natural channels labelled NAT); inverted open triangles are simulated by the model of Howard and Knutson (1984) and Howard (1992) (domain labelled HKM); open diamonds are disturbed periodic simulations (Ferguson, 1976, 1977) (domain labelled DPM). Filled triangles are simulated channels with chute cutoffs, and filled boxes are simulations with resistant valley walls or resistant oxbow lakes

straight channel in which bankfull flows are barely competent to induce systematic bank erosion, but as sinuosity increases the decrease in flow velocity reduces the bank erosion rate to nearly zero before sinuosity reaches a value sufficient to induce cutoffs. During flood flows the low banks and wide floodplain limit within-channel flows. In addition, the downstream flow over the floodplain may interfere with flow in the channel, particularly since the overbank flow is largely perpendicular to the meander arms. Jarrett and Costa (1986) note that a dam break flood passing over the floodplain shown in Figure 2.17 had little effect upon channel pattern and bank erosion. The

Figure 2.17   Fall Creek, Estes Park, Colorado. Note the highly sinuous channel pattern with few oxbow lakes. Flow from bottom left to upper centre

slowness of lateral shifting may either be due to high bank resistance (constraint 3, above) or to a low supply of sediment to construct point bars (constraint 1), or both. An examination of the channel in Figure 2.17 reveals that banks occasionally fail by undermining and toppling, but the collapsed bank remains in the channel bed, anchored by the still-living grass, and apparently surviving with minimal break-up and erosion for several years (Deborah Anthony, personal communication). Low point bars form in these channels during high flows, but are eroded as flows diminish (Anthony, 1991). Lateral shifting in the channel system may have increased in recent years due to increased bed sediment supply originating from a debris fan constructed by the 1982 dam Break flood (Deborah Anthony, personal communication), suggesting that the ability to construct point bars may play a role in limiting bank erosion rates. Floodplains constructed by such relatively fixed meandering channels presumably occur primarily by vertical accretion, at least during their final stages.

Proposed model enhancements

Howard (1992) discussed a number of shortcomings of the meandering floodplain evolutionary model and strategies for overcoming these, as well as possible model enhancements. Only a synopsis is considered here.

Howard (1992) and the present contribution utilize the linearized J&P (Johannesson and Parker, 1989) flow and bed topography model. This model is utilized primarily

because of its computational efficiency, which is of primary concern when flow and bed topography must be recalculated thousands of times per simulation. However, the J&P model considers only time-averaged, curvature-forced bedforms. In wider meandering channels alternate bars may migrate through the channel (Kinoshlta, 1961; Fukuoka, 1989; Ikeda, 1989; Tubino and Seminara, 1990; Whiting and Dietrich, 1993a,b,c) and these bars may systematically affect bank erosion (Whiting and Dietrich, 1993c). For certain combinations of width/depth ratio and flow parameters, alternate bars become stationary, and, if their natural wavelength is the same as the meander wavelength, a "resonance" occurs under which conditions the linearized models, such as J&P, predict very large amplitude bars (Blondeaux and Seminara, 1985; Columbini *et al.*, 1987; Parker and Johannesson, 1989; Seminara and Tubino, 1989; Tubino and Seminara, 1990; Columbini *et al.*, 1992). The J&P model fails to converge to a stable solution under such conditions, although this might be correctable In a future version (G. Parker, personal communication). Furthermore, multiple bars may occur in large-amplitude meander bends (Whiting and Dietrich, 1993a,b), and the migration of alternate bars can be suppressed in sinuous meanders (Kinoshita, 1961; Fukuoka, 1989; Tubino and Seminara, 1990, Whiting and Dietrich, 1993c), both of which might affect near-bank flow and depth (and hence bank erosion) in systematic ways. The effects of alternate bars could be addressed by use of a flow and bed topography model that explicitly treats cross-stream topography and which has time-steps sufficiently short to resolve alternate bar migration (e.g. Nelson and Smith, 1989a,b; Shimizu and Itakura, 1989), but computational costs may prohibit incorporation into long-term evolutionary models. An alternative approach is to use the more explicit models to develop heuristic "corrections" to linearized models (such as J&P, 1989, and Odgaard, 1989a,b).

Bank erosion rates are assumed to respond to a dominant discharge. Discharge variation affects not only the magnitude of the velocity and depth perturbations, but also their distribution around the bend (lags in response to curvature changes become greater at higher discharges), as well as influencing the development and migration of bars. A frequency distribution of flow discharges could be incorporated into the model.

Equations 2.9 and 2.10 are based upon a simple diffusional conceptualization of floodplain deposition as embodied in the approach of Carey (1969) and Pizzuto (1987). However, as Pizzuto (1987) points out, advectional flow transport can lead to patterns of deposition rates and grain sizes not describable by the above parameters. This is particularly important for flows in chutes and sloughs, where both suspended load advection and bedload transport may occur. Techniques for modelling of overbank flows have been developed (Knight and Demetriou, 1983; Yen and Yen, 1984; Ervine and Ellis, 1987; Knight, 1989; Gee *et al.*, 1990; Miller, 1994; Bates *et al.*, 1992; Knight and Shiono, Chapter 5, this volume; Sellin and Willetts, Chapter 8, this volume; Younis, Chapter 9, this volume), and these can be coupled with models for suspended load deposition on floodplains (Carey, 1969; Pizzuto, 1987; Marriott, 1992, Chapter 3, this volume) to predict the spatial pattern and grain size distribution of deposited sediment (Nicholas and Walling, 1995). A frequency distribution of flood flows could be assumed, with deposition rates being based on a steady-state solution of floodplain flow for the flood peak.

An obvious extension would be to include grain size and stratigraphic information by modelling deposit thickness, bedform and grain size. Grain size and bedforms of point-bar deposits can be related to within-channel sorting and flow parameters much in the manner adopted by Allen (1970), Bridge (1975, 1978, 1984a) and Bridge and Jarvis (1977, 1982).

2.5   ANASTOMOSING AND AVULSIVE FLOODPLAINS

Very-low-gradient channel systems are commonly straight to weakly meandering with well-separated multiple channels, or anastomoses (Knighton and Nanson, 1993). Low gradients are associated with low bedload conveyance, tectonic backtilting (Burnett and Schumm, 1983; Gregory and Schumm, 1987), blocked valleys (Smith and Smith, 1980), and, usually, an aggradational regime (Smith *et al.*, 1989; Knighton and Nanson, 1993). Channel meandering is weak due to low stream power coupled with resistant banks (cohesive and/or strongly vegetated) (Harwood and Brown, 1993; Knighton and Nanson, 1993; Stanistreet *et al.*, 1993). Due to the aggradational regime, natural levees are usually well developed. Anastomoses generally develop due to avulsion at breaches in the natural levees. Also due to the aggradational regime (or, possibly, floodplain sinking), non-active portions of the floodplain are commonly swampy or occupied by shallow lakes. Splay deposits associated with multiple small distributaries are a common occurrence (Smith and Putnam, 1980; Smith and Perez-Arlucea, 1994).

Avulsions are not limited to the classic low-gradient anastomosed floodplains. Richards *et al.* (1993) note that avulsions may even occur in braided channel systems.

Geomorphic modelling of anastomosed or avulsive floodplains apparently has not been attempted, except for large-scale stratigraphic models of sedimentary facies (Allen, 1978; Bridge and Leeder, 1979). Some of the modelling techniques utilized for meandering channels, described above, could be adapted to anastomosed floodplains. Although a variety of processes and constraints may be involved in avulsions (see, for example, Richards *et al.*, 1993), most are strongly influenced by floodplain topography. Thus the development of avulsions could be modelled stochastically, similar to the approach used for chute cutoffs, with the probability of a diversion being a function of levee height and relative channel and floodplain elevations (e.g. Mohrig *et al.*, 1994).

2.6   LARGE SPATIAL DOMAIN MODELS

In the previous discussion the focus has been on models that attempt to simulate the history of individual channels and details of floodplain topography, and in some cases, floodplain stratigraphy. If the concern is for larger spatial and temporal domains, such as valley aggradation and entrenchment, or the overall development of deltas, alluvial fans or pediments, then faithful reproduction of channel morphology, channel pattern changes and detailed stratigraphy may be abandoned in order to efficiently simulate regional morphological and stratigraphical relationships. Models of sedimentary basin

development (e.g. Flemings and Jordan, 1989; Paola, 1989; Jordan and Flemings, 1990; Paola *et al.*, 1992; Slingerland *et al.*, 1994) have long used simplified approaches generally ignoring effects of individual channels and their deposits, or, in some cases, representing deposits of individual channels only in their bulk as channel complexes (e.g. Leeder, 1978; Bridge and Leeder, 1979). The present discussion will concern
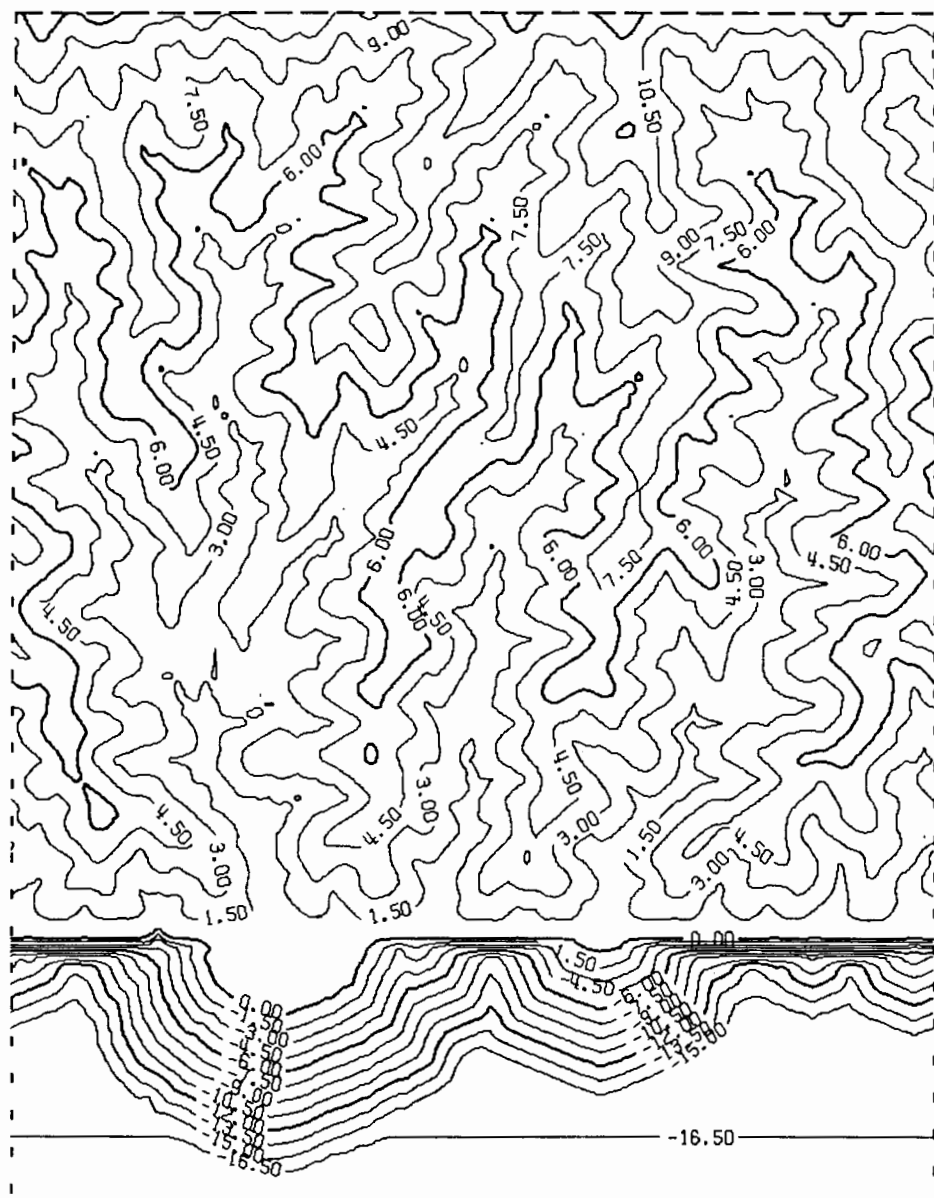

Figure 2.18 Simulation of drainage basin erosion and concomitant deltaic deposition using the simulation model of Howard (1994). Sea level is at an elevation of 0.0

spatially explicit modelling of fluvial depositional environments in which the topography of the deposit as well as, possibly, the stratigraphy are of interest.

In depositional environments, particularly deltas and alluvial fans, channel aggradation, possibly coupled with land surface lowering by sediment consolidation, results in frequent avulsions. In addition, the channel system commonly has multiple active channels as anabranches or distributaries. Rachocki (1981) simulated braided channels on alluvial fans with a random-walk model. However, this model was not coupled with deposition. Price (1974) simulated alluvial fan deposition with a probabilistic model for uplift, precipitation and fan deposition. In this model flow is routed down the fan, following a single course during each event, with the relative probabilities among possible directions at each node representing the fan surface being proportional to the slope gradients to adjacent nodes. Howard (1994) abstracts the depositional process further in a coupled model of basin erosion and sedimentation that simulates development of alluvial fans and alluvial pediments. Figure 2.18 shows an example of such a coupled model with delta development at a presumed shoreline. This model does not account for suspended load deposition as bottomset deposits, so that it is most applicable to the bedload-dominated Gilbert type of delta (Gilbert, 1890). The models of Howard (1994) utilize an approach to sediment routing and deposition that is computationally three orders of magnitude more efficient than finite difference modelling. In these models of alluvial fans, pediments and deltas, channels during each iteration follow the steepest path downstream across the alluvial cover. The deposition during each iteration is not intended to represent an actual flow and sedimentation event, but rather, the cumulative effect of multiple events. In aggradational settings the sedimentation occurring during each iteration tends to raise the current channel relative to older sections of the fan or delta, so that during the next iteration the steepest path follows a different course. In the Howard (1994) model the channels shift so frequently that the resulting landform is very symmetrical and often strongly affected by the restrictions of flow from any cell to one of eight directions (Figure 2.16). A more realistic simulation and presumably also a more realistic morphology (including, perhaps, birdfoot deltas) would occur if channel shifting was made more difficult, corresponding to the development of natural levees that constrain avulsions until the channel bed is well aggraded. Rules for such avulsions were discussed above. As with the floodplain meandering model described above, only the net quantity of sediment deposition is accounted for, and no attempt has been made to model sediment size or stratigraphy, although such features could be added.

2.7 DISCUSSION

Modelling of the geomorphic and stratigraphic development of floodplains is presently more of a promise than a reality, but most of the quantitative tools are at hand and computers are sufficiently capable that model development should mature rapidly. These concluding remarks address issues of model development and validation.

In the early days of the quantitative revolution in geomorphology, characterization of landform morphology and geomorphic processes revealed numerous intriguing relationships and patterns that exceeded our ability to explain. The rapid development of

theory and computational resources during the last two decades has often reversed this situation, such that validation of models is handicapped by the lack of appropriate empirical data. Sometimes this is unavoidable because of large spatial or temporal scales, as with models of drainage basin evolution. However, in the case of modelling of floodplain evolution, changes occur rapidly enough that short-term observation coupled with longer-term evidence from surveys, aerial photographs and floodplain stratigraphy can be utilized for model calibration and validation. Even so, there are few systematic databases on floodplain evolution that can be directly utilized in model validation.

Howard (1992) discussed a variety of validation approaches within the context of meandering floodplain evolution. Although many studies of stream meandering and braiding have collected information from individual short reaches, validation of large-scale models requires "reach-length" studies, extending downstream through multiple braids or meander bends. One such approach utilizes static comparison of simulated floodplain features with natural analogues, generally using multivariate statistical methods. For example, Howard and Hemberger (1991) showed that meander planforms simulated by the Howard (1992) model and the Ferguson (1976, 1977) disturbed periodic model are statistically distinguishable from natural meanders. Similar multivariate morphometric measurements have been applied to braided channels by Howard et al. (1970) and Webb (1994, 1995), and these could be used for validation of the braided channel model of Murray and Paola (1994).

A more powerful validation technique compares the temporal changes, or kinematics, predicted by a model with observed natural changes, thereby offering important clues as to specific sources of model deficiencies. Unfortunately, although a rich database of historical meander change exists, analyses of meander kinematics to date have been primarily qualitative or have yielded only summary statistics, such as average bank erosion rates (e.g. Brice, 1974a,b; Hooke, 1977, 1979, 1980, 1984; Dort, 1978). Studies of systematic variation in erosion rates with channel curvature and sediment properties utilized isolated bends (Hickin and Nanson, 1975, 1984; Nanson and Hickin, 1983, 1986) and have not fully accounted for upstream control of local flow and bed characteristics implied in theoretical models (such as the J&P model used here) (Parker, 1984; Furbish, 1988). Howard and Knutson (1984) have used an earlier version of the flow model to simulate several decades of channel shifting on the White River of Indiana, with generally encouraging results. Short-term predictions of channel shifting have also been made by Parker (1982), Beck (1984), and Beck et al. (1984). The flow and transport model has been compared to flume studies of meandering channels (Johannesson and Parker, 1989). Furbish (1988), Hasegawa (1989a,b), Pizzuto and Meckelnberg (1989), and Shane Cherry (personal communication, 1995) have compared observed bank erosion for individual bends or short reaches with model predictions. Although these comparisons are generally encouraging, they are too few and too rudimentary to comprise a thorough testing of the flow and erosion model. What is needed is systematic analysis of meander kinematics on a number of long reaches of natural and simulated channels and a comparison with model predictions. Howard (1992) outlines several direct and indirect techniques of reach-length comparison between historical records of change in natural meandering channels with simulated changes. Similar techniques could be utilized for braided channels, although

highly braided channels change so rapidly that year-to-year aerial photographic comparisons may not be useful.

By extending the types of observation made by Lewis and Lewin (1983), spatio-temporal information on neck and chute cutoffs and avulsions, together with information on floodplain topography and channel geometry could be used to quantify models of channel pattern changes in meandering and braided streams.

The validation of floodplain deposition models will prove troublesome because relevant data are difficult to obtain. Most easily assembled are statistics relating floodplain age, elevation and distance from stream channels (e.g. Everitt, 1968; Kesel et al., 1974; Hickin and Nanson, 1975; Nanson, 1980). Floodplain sedimentology and stratigraphy is more difficult to characterize. Cores and trenches are the obvious but tedious method, supplemented with archaeological and age-dating techniques to provide rate information (e.g. Brakenridge, 1984, 1985; Brown 1987, 1990; Walling and Bradley, 1989). Judicious use of sections exposed in cutbanks can also be useful. Ancient deposits in the sedimentary record serve as comparisons. Relatively short duration studies of rates and spatial patterns of deposition and erosion are quite practical for reach-length studies, using surveying, coring, and use of markers such as sand or gypsum (e.g. Hupp and Bazemore, 1993) or isotopic markers (e.g. Walling et al., 1992).

Geomorphologists developing theoretical models often find that appropriate empirical data are lacking. Sometimes this occurs because relevant data are difficult to obtain, but often geomorphologists have not appreciated the relevance of particular types of data or that data that have been collected are in the wrong form for model validation purposes. Hopefully, as it has in other branches of geomorphology, development of theoretical models will spur new field or experimental studies to acquire appropriate data sets on floodplain evolution.

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