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Tectono-sedimentary evolution of active extensional basins

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ABSTRACT

We present conceptual models for the tectono-sedimentary evolution of rift basins. Basin architecture depends upon a complex interaction between the three-dimensional evolution of basin linkage through fault propagation, the evolution of drainage and drainage catchments and the effects of changes in climate and sea/lake level. In particular, the processes of fault propagation, growth, linkage and death are major tectonic controls on basin architecture. Current theoretical and experimental models of fault linkage and the direction of fault growth can be tested using observational evidence from the earliest stages of rift development. Basin linkage by burial or breaching of crossover basement ridges is the dominant process whereby hydrologically closed rifts evolve into open ones. Nontectonic effects arising from climate, sea or lake level change are responsible for major changes in basin-scale sedimentation patterns. Major gaps in our understanding of rift basins remain because of current inadequacies in sediment, fault and landscape dating.

INTRODUCTION AND RATIONALE

Active extensional basins are important because their sedimentary fills and bounding tectonic structures provide: sinks with high preservation potential for sedimentary and fossil records of past changes in climate, sea/lake level and sediment/water supply, information on the growth, activity, decay and death of normal faults, vast economic reserves of hydrocarbons, water and minerals, a record of the early rifting of continents during cycles of plate break-up, a record of the back-arc extension of continents during cycles of plate convergence.

It is the purpose of this contribution to present updated (cf. Leeder & Gawthorpe, 1987) conceptual models for the tectono-sedimentary evolution of rift basins, with particular emphasis on: the three-dimensional (3D) evolution of basin linkage through fault propagation, the evolution of drainage and drainage catchments, the effects of changing climate and sea/lake level.

Numerous studies in recent years have shed light on some of the above topics, but major gaps remain because of the lack of fully integrated stratigraphic and structural studies, and due to current inadequacies in sediment, fault and landscape dating.

Colour and black-and-white versions of the conceptual models can be obtained from www.man.ac.uk/Geology/research/BSG/basin-res.html

TECTONIC AND STRUCTURAL FRAMEWORK

Current status

Recent developments in the understanding of normal fault arrays have come from a wide range of approaches, including investigations of active faults, description of ancient faults at outcrop and in the subsurface, and modelling of fault array evolution.

Active seismogenic normal faults

Studies of major active normal faults in areas such as the Basin and Range, Aegean and East Africa indicate that they are all segmented along strike (e.g. Jackson, 1987; dePolo *et al.*, 1991; Machette *et al.*, 1991). For major fault zones that break the seismogenic crust, segments are commonly 20–25 km long (Jackson & White, 1989), but may be longer where the seismogenic crust is inferred to be thicker (e.g. East African rift, Jackson & Blenkinsop, 1997). Fault segment boundaries are marked by local highs and lows in hangingwall and footwall elevations, respectively, and are commonly regarded as persistent (though not entirely efficient) barriers to fault rupture (e.g. Schwartz & Coppersmith, 1984; Crone & Haller, 1991; dePolo *et al.*, 1991; Machette *et al.*, 1991; Zhang *et al.*, 1991; Gawthorpe & Hurst, 1993; Roberts, 1996).

Segment boundaries are often marked by an increased density of small-displacement faults (Jackson & Leeder, 1994; Mack & Seager, 1995), as seen in the pattern of faulting during the 1915 earthquakes (Wallace, 1984) that cut the crossover basement ridge of the Sou Hills located between Pleasant and Dixie Valleys, Nevada (Fig. 1B). The reason why active fault segment boundaries are associated with marked relief variations compared to the centre of fault segments is fundamental to the study of basin evolution and to the apportionment of strain within a rift zone. For a given amount of regional extension, topographic elevation depends upon fault spacing, so areas of small-scale distributed faulting at fault segment offsets and crossovers stand at higher elevations than the hangingwalls adjacent to large-displacement border faults (Jackson & Leeder, 1994).

The Wasatch fault zone displays many of the key characteristics of large active normal fault zones (e.g. Schwartz & Coppersmith, 1984; Machette *et al.*, 1991). It is discontinuous along its length, comprising 10 segments with boundaries characterized by footwall lows and hangingwall highs. The longest segments, highest slip rates and highest footwall topography (proxy for displacement) are located in the centre of the fault zone, the magnitude of these parameters decreasing towards its ends. Furthermore, there is distinct evidence of episodic

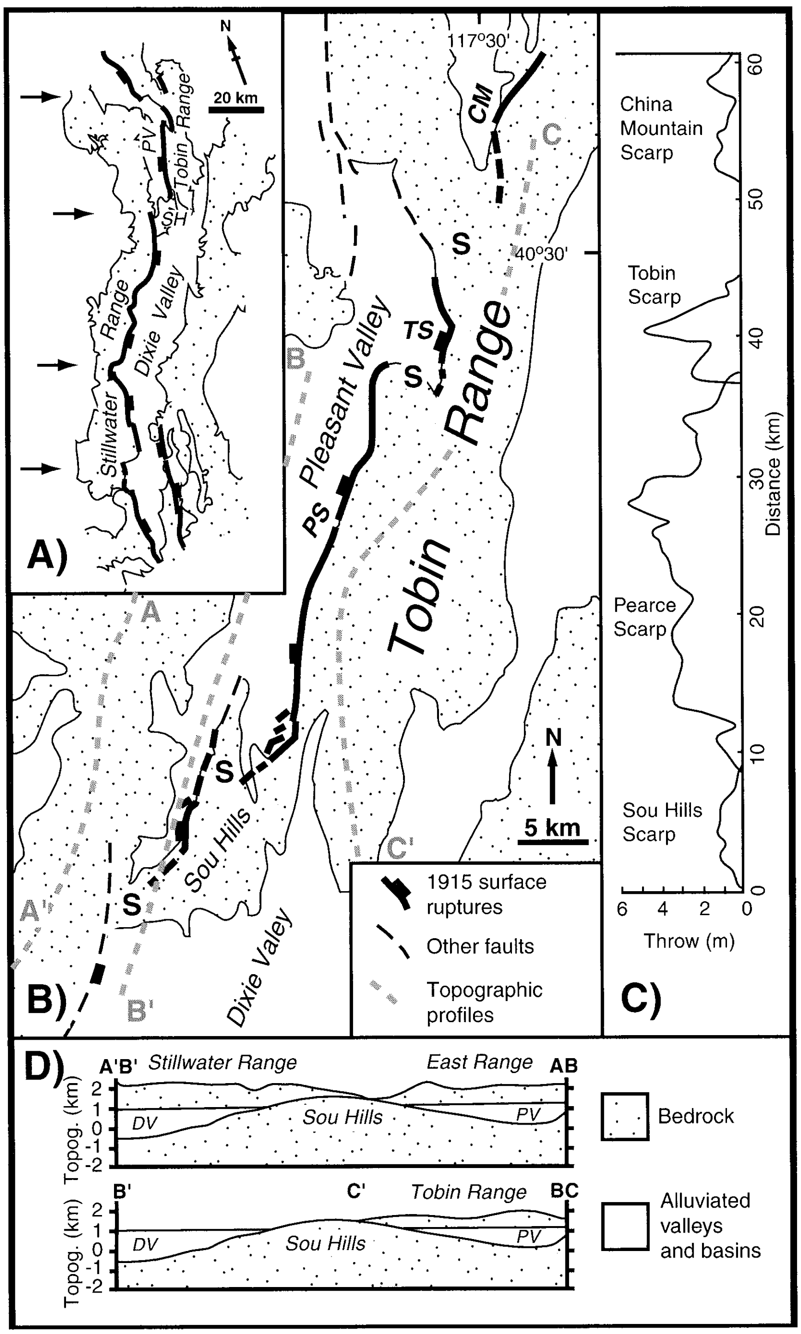

Fig. 1. Fault segments and slip and displacement profiles for the Dixie and Pleasant Valleys, Nevada, USA (modified from Wallace, 1984; dePolo *et al.*, 1991; Zhang *et al.*, 1991; Jackson & Leeder, 1994). (A) Major segment boundaries (arrows) and fault segments. (B) Map of the 1915 surface ruptures, showing rupture segment boundaries (S). Only some of the minor faults in the Sou Hills area are shown for clarity. (C) Slip profile for the 1915 ruptures. (D) Range crests and valley floor topography and estimated depth to basement profiles illustrating nature of long-term displacement profiles for the Dixie Valley and Pleasant Valley border faults. Note how the major segment boundary around Sou Hills is associated with high topography relative to the basins to the N and S. PS, Pearce scarp; TS, Tobin scarp; CM, China Mountain scarp; DV, Dixie Valley; PV, Pleasant Valley; SH, Sou Hills.

growth through clustering of earthquakes on particular parts of the fault zone (Machette *et al.*, 1991).

The advantages of analysing active structures is that many of the controls on their development can be constrained independently and slip rates and recurrence intervals can be determined. However, the rates and evolution of active structures have so far been quantified only over a few thousand to a few tens of thousands of years, which is insufficient to resolve medium- to long-term (10^4–10^6 years) evolution (Cowie, 1998). There is a clear need for studies that address longer-term fault development from the evidence of landscape evolution and basin stratigraphy. In both cases, accurate chronology is an essential prerequisite.

Ancient normal faults (outcrop and subsurface)

In the absence of quantitative data on medium- to long-term fault evolution, the easily derived relationship between fault length and displacement has stimulated numerous ideas as to how faults grow, interact and scale (Fig. 2). Studies at outcrop and in the subsurface suggest a scaling relationship between displacement and length such that $D = cL^n$ where D is maximum displacement, L is maximum trace length and c is a constant related to rock properties. The value of exponent n ranges between 1 and 2 (Watterson, 1986; Walsh & Watterson, 1988; Marrett & Allmendinger, 1991; Cowie & Scholz, 1992; Dawers *et al.*, 1993; Cartwright *et al.*, 1995; Dawers & Anders, 1995).

It has been suggested that some of the scatter in D–L relationships reflects fault growth by segment linkage (e.g. Peacock & Sanderson, 1994; Cartwright *et al.*, 1995; Dawers & Anders, 1995). Furthermore, for a discontinuous fault zone consisting of a number of segments, the D–L scaling for the whole zone is consistent with that for a single segment, and the D–L profiles for adjacent segments are often asymmetric and have steep gradients (e.g. Peacock & Sanderson, 1994; Cartwright *et al.*, 1995; Dawers & Anders, 1995; Willemse *et al.*, 1996; our Fig. 2). These observations suggest that there is mechanical interaction between segments before they link and that the segmented fault zone essentially acts as a single fault of the same length as the

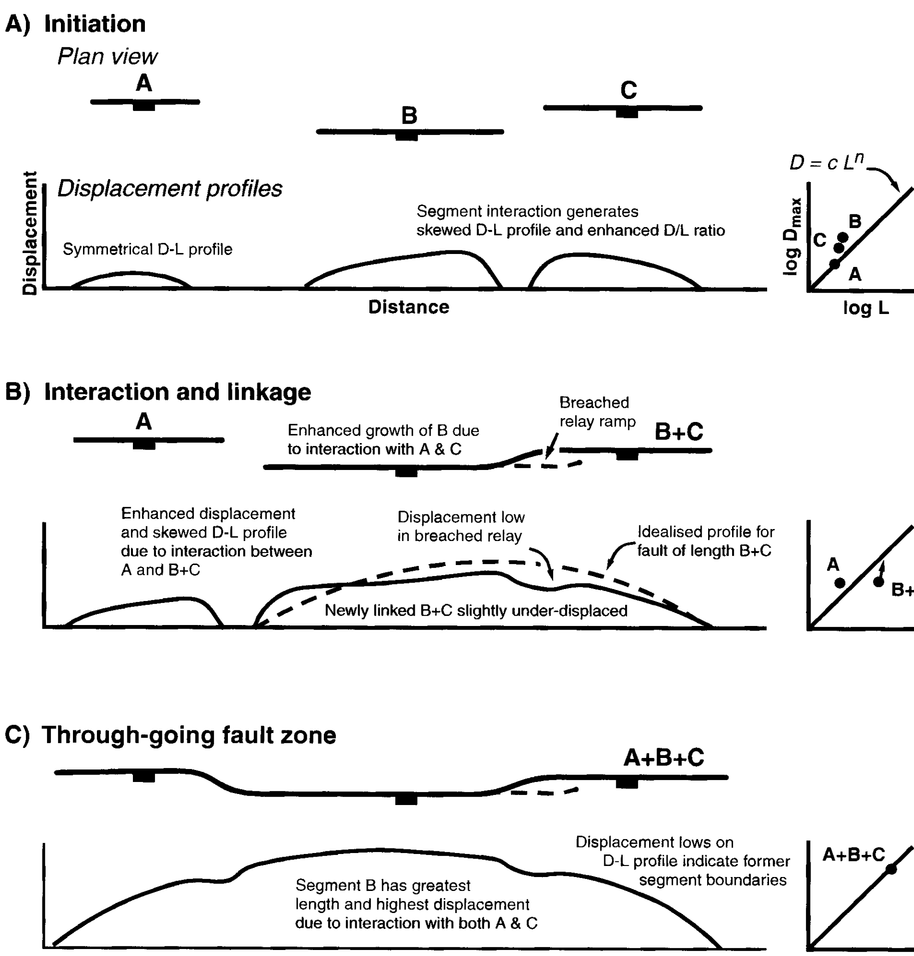

Fig. 2. Schematic evolution of three segments to produce a major border fault zone, such as illustrated by the evolution of segments A, B and C in Fig. 3. (A) Fault initiation stage, (B) interaction and linkage stage, (C) through-going fault zone stage. Note how interaction between segments produces skewed displacement profiles and that the displacement (D) and length (L) characteristics evolve so that the through-going fault zone has similar characteristics to those of an isolated fault segment (e.g. Fault A in A).

array of segments (Fig. 2). It also follows that, during segment interaction and linkage, the linked fault must accumulate displacement without increasing length in order to develop an appropriate D/L ratio.

Observations from a variety of rift basins indicate that folding is an important element in the development of normal fault zones. In particular, transverse folds (at high angles to fault strike) are associated with along-strike displacement gradients (e.g. Schlische, 1995). In the hangingwall of normal faults, transverse synclines define displacement maxima, whereas transverse anticlines are associated with displacement minima at segment boundaries (Gawthorpe & Hurst, 1993; Anders & Schlische, 1994; Schlische, 1995). Other folds associated with normal faults lie parallel to the fault zone and form due to ductile deformation ahead of the propagating fault tip (Fig. 4). These fault-propagation folds are commonly preserved as monoclines in the footwall of normal faults, and as hangingwall synclines (Withjack *et al.*, 1990; Schlische, 1995; Gawthorpe *et al.*, 1997; our Figs 3A and 4). Only when the fault tip can be reconstructed from landscape evolution or basin stratigraphy (e.g. Anders & Schlische, 1994; Gawthorpe *et al.*, 1997) can the results show the growth history of the fault, and even then the sequence of events may not yield the *rate* of tip growth.

Modelling studies

Analogue and numerical modelling studies have provided information on the progressive evolution of normal fault zones in two and three dimensions. Analogue models serve as useful templates for the analysis of natural fault arrays, displaying many of the segmentation and propagation characteristics of normal fault zones in natural rift systems. In particular, analogue models have provided insights into the deformation ahead of propagating faults, the breaching of relay ramps and the structure of accommodation zones (e.g. Ellis & McClay, 1988; Withjack *et al.*, 1990; McClay & White, 1995).

Cowie & Scholz (1992) and Cowie & Shipton (1998) have applied post-yield fracture mechanics to explain the observed D–L scaling relationships and the form of D–L profiles for normal fault segments and fault zones. Cowie & Shipton (1998) suggest complex segment growth histories, during which are phases of displacement accumulation and little propagation, alternating with intervals of marked propagation and limited displacement. These workers and colleagues (e.g. Willemse *et al.*, 1996; Crider & Pollard, 1998) have also investigated the mechanical interaction between fault segments where rupture on one segment enhances or inhibits rupture on a nearby segment due to static stress changes (King *et al.*, 1994). Results of such investigations provide a physical basis for the observed variations in D/L ratios and D–L profiles for segmented fault zones. In contrast, a kinematic approach, based on trishear fault propagation folding, has been applied by Allmendinger (1998) and Hardy & McClay (1999) to investigate folding and strain associated with normal fault propagation (Fig. 4).

On a larger scale, Sornette *et al.* (1990, 1994) and Cowie *et al.* (1993, 1995) have examined the temporal and spatial evolution of normal fault arrays, suggesting that they develop a fractal pattern and a power-law distribution of fault sizes. A consistent feature of these models is progressive localization of faulting. Cowie (1998) further developed the modelling of fault arrays by integrating the role of stress feedbacks on segment interaction and growth. Her results provide insights into the temporal development of fault arrays and suggest evolution from an initial nucleation phase with many isolated small segments, through a phase of enhanced growth where interactions between segments become increasingly important, to a phase where deformation is localized along a few major fault zones and other faults become inactive (Fig. 3). Episodic displacement accumulation involving alternate phases of rapid growth and relative quiescence is a general feature of segment growth, the increase in segment length and displacement rate being strongly influenced by position with respect to surrounding segments (see our Fig. 2). Segments located at fault zone centres (sites of rupture symmetry) have the highest displacement rates and lengths because they are most frequently loaded by laterally adjacent segments. In contrast, segments located in stress shadow zones have low displacement rates and may become inactive as deformation localizes (Figs 2 and 3).

Current issues

Sedimentary consequences of models for fault growth, propagation and evolution

The patterns of surface uplift and subsidence produced by the activity of growing normal faults enable some progress to be made in stratigraphic tests of theoretical and conceptual fault growth models. Testable features include the occurrence of presurface faulting monoclinal folds (Gawthorpe *et al.*, 1997), modifications to drainage (Jackson & Leeder, 1994) and sedimentation patterns (Gawthorpe *et al.*, 1997) and changes of stratal thickness in hangingwall depocentres (Gupta *et al.*, 1998; Sharp *et al.*, 2000a).

Consider surface fault growth with initially many small faults with small displacements, evolving to an arrangement whereby fewer and fewer but longer and longer faults with larger displacements and higher displacement rates are active (Fig. 3). This evolutionary scheme is testable in the basin record, since earliest basins should be numerous, isolated, bounded by small faults and show evidence for progressive abandonment. Furthermore, individual basins should show low subsidence rates and be widely preserved at depth across the hangingwall basement of later developed larger basins. Such scenarios are envisaged for the evolution of both the Gulf of Suez and northern North Sea rifts (see Gupta *et al.*, 1998;

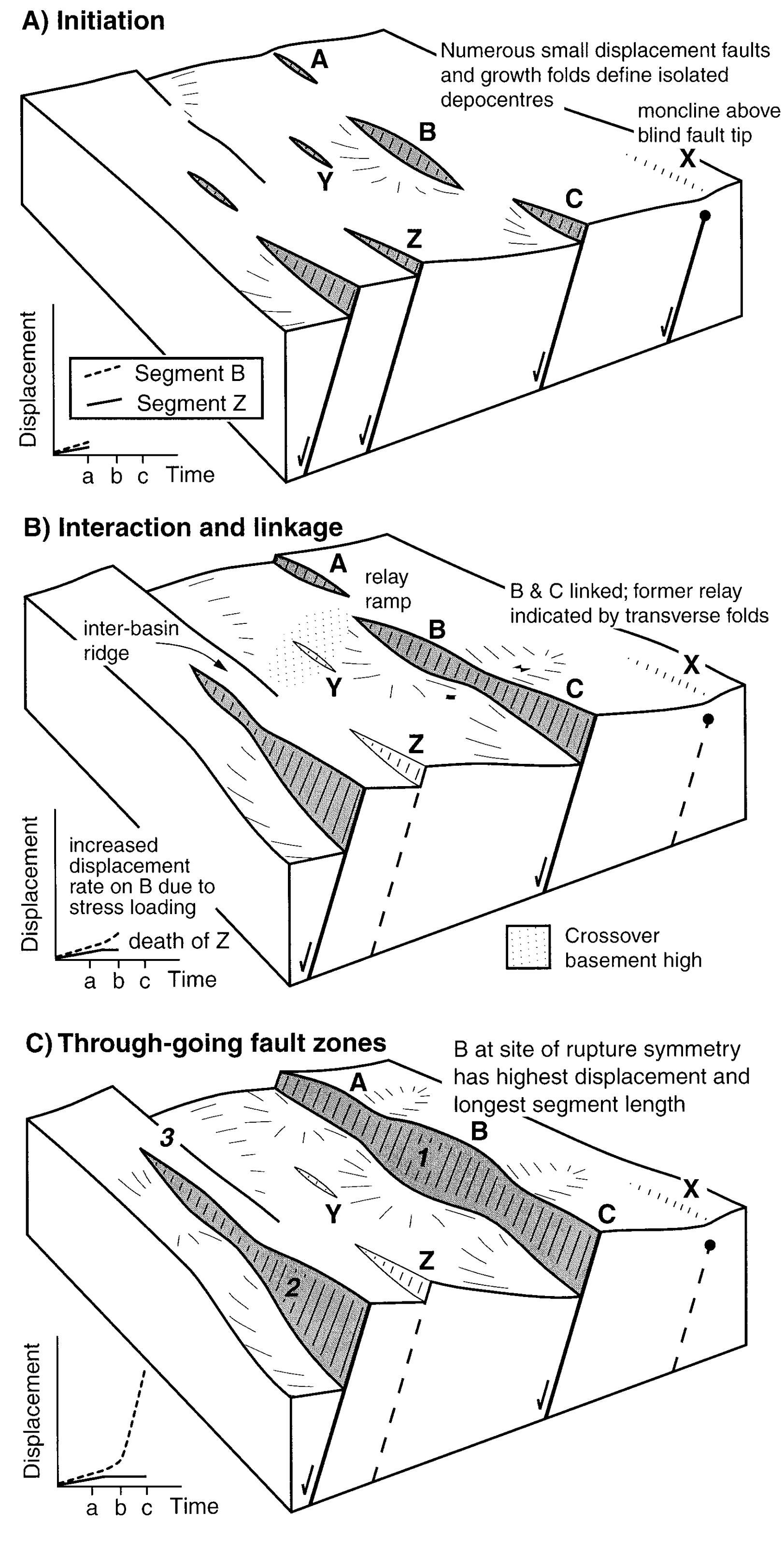

Fig. 3. Schematic 3D evolution of a normal fault array, with graphs illustrating displacement history of fault segments B and Z. (A) Fault initiation stage, characterized by a large number of small-displacement normal fault segments. Note the low surface topography influenced by fault propagation folds and surface-breaking normal fault scarps. (B) Fault interaction and linkage stage, where stress feedback between segments influences growth, and deformation in the fault array begins to become localized along major fault zones (A, B, C). Faults located in stress shadows begin to become inactive (X, Y, Z). (C) Through-going fault zone stage, where deformation is localized along major border fault zones (e.g. 1, 2 & 3) giving rise to major half graben and graben depocentres. Note the death of segment Z and increase in the displacement rate on segment B as deformation becomes localized on major fault zones.

Sharp *et al.*, 2000b), but the time-scales involved in the development of fault arrays are poorly constrained. For the Statfjord East fault in the Northern North Sea, Dawers & Underhill (2000) suggest progressive growth of the fault zone by linkage of individual segments over an interval of 20 Myr, whereas the transition from numerous isolated fault segments to major linked, crustal-scale fault zones occurred some 5 Myr after the onset of rifting in the Gulf of Suez (e.g. Patton *et al.*, 1994; Sharp *et al.*, 2000a,b).

An alternative model of fault growth is where most fault growth occurs at depth, from the base of the brittle upper crust upwards. In this scenario all 'natural selection' between competing small faults takes place subsurface. Eventually a fully formed major seismogenic fault breaks surface and very rapidly propagates laterally to reach its equilibrium length, approximately that of the thickness of the seismogenic layer (Jackson & White, 1989). No other large fault reaches surface to transfer crustal extensional strain. During the blind-fault phase a deformation front with anomalous seismicity at the subsurface fault tip must rise through the brittle layer. Any

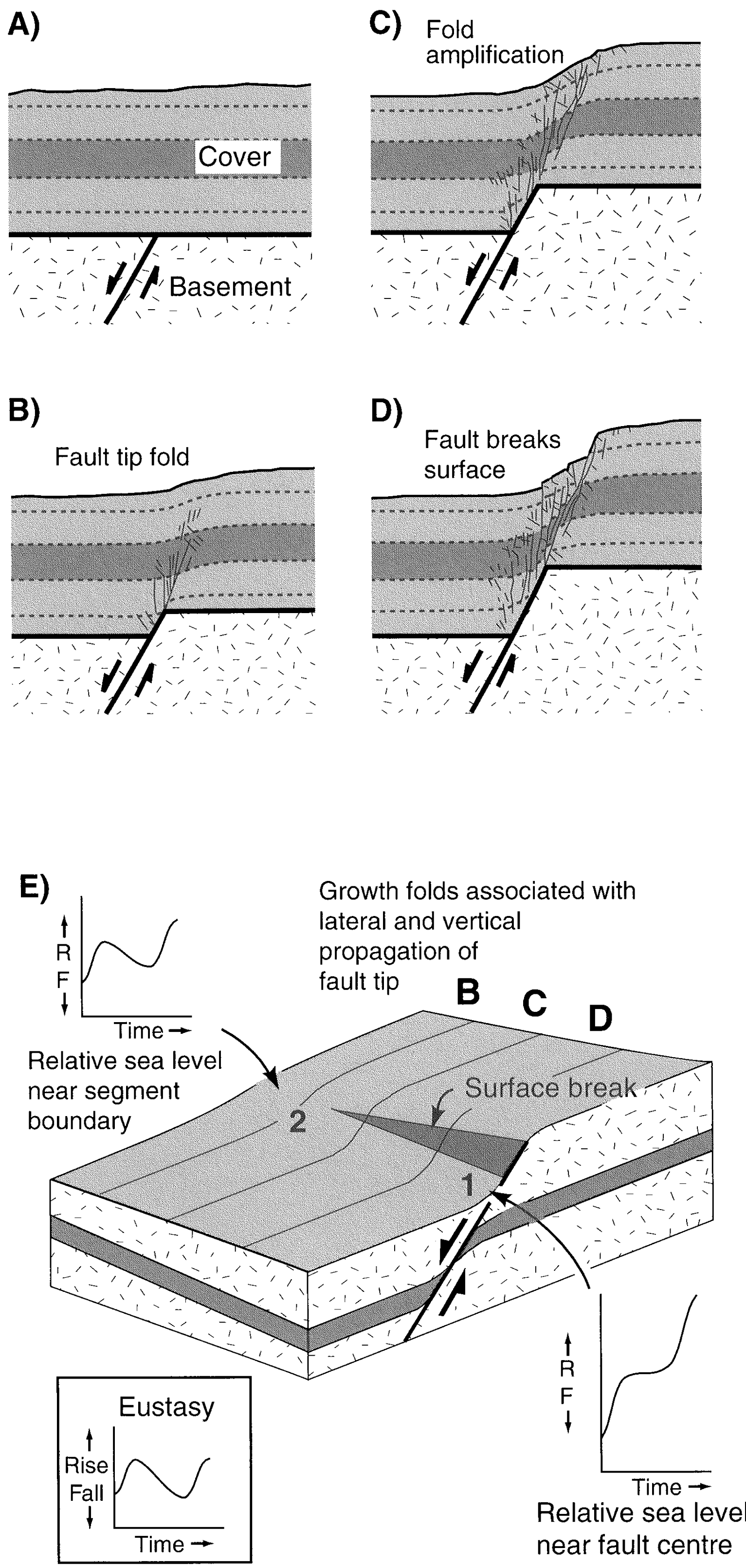

Fig. 4. Folding around propagating normal fault zones. (A–D) Experimental clay model of extensional fault-propagation fold associated with an upward-propagating, normal fault, dipping at 60° involving a rigid basement and a ductile sedimentary cover. Surface deformation above the blind normal fault is characterized by a growth monocline and the presence of high-angle to locally reverse secondary faults and fractures in the immediate hanging wall. Continued fault propagation beyond the stage illustrated in (D) would lead to a more typical half graben geometry. Redrawn from Withjack *et al.* (1990). (E) Strike variability in surface deformation for an idealized half fault segment (no scale implied). The fault-tip is associated with the development of a monoclinal fold above the blind, propagating normal fault. B, C and D represent the spatial position of cross-sections that would be equivalent to the structure illustrated at stages (B) (C) and (D) in the temporal evolution. Graphs illustrate the differences in relative sea level change due to along-strike variations in displacement rate. High displacement rates at fault segment centres (e.g. location 1) can outpace all but the fastest rates of sea level fall, leading to continual relative sea level rise, whereas slow rates of displacement near segment boundaries (e.g. location 2) give relative sea level falls. High-order glacio-eustatic sea level curve used in construction of the relative sea level curves is shown in the inset.

small surface faults which may exist are shallow features, initiating in the upper brittle layer or even in the sedimentary fill itself as surficial strain relievers.

Fault death

In many areas of active extensional tectonics, faults exist that are not currently active and which might thus be considered dead (Jackson, 1999). Leaving aside potential difficulties due to 'Lazarus faults', palaeoseismic indicators, geomorphology and basin stratigraphy reveal clear evidence for spatial and temporal migration of active faulting (e.g. Wallace, 1978; Wallace & Whitney, 1984). Such migrations and, more importantly, their rates and timing, have major implications for the dynamics of crustal extension. Unfortunately, the majority of published studies have poor time constraints. A particularly interesting example (Fig. 5A) comes from the Afar–Ethiopian rift (Hayward & Ebinger, 1996) where a decrease in crustal and lithospheric thickness from south to north is attributable to a southward propagation with time of 'oceanic'-style volcanism and extension. This is displayed in a south to north change in the nature of rift margin faulting, from long border faults bounding half graben, to full graben with felsic volcanism and finally to narrow, short graben with voluminous basic volcanics. The most interesting aspect is the preservation of both a time and space sequence, since the youngest stage in the northern Afar rift is preserved within an abandoned wide graben bounded by the earliest long faults. Within this old rift, antecedent transverse drainage now cut across younger central fault arrays and incise former subsiding basin fills along old major bounding faults. Other examples of fault death, abandonment, uplift and basin-fill cannibalization by rivers occur in Central Greece (our Fig. 5B; Leeder *et al.*, 1991; Leeder & Jackson, 1993; Jackson, 1999) and the southern Rio Grande rift (Mack & Seager, 1995).

SEDIMENTARY FRAMEWORK

Current status

Basins and their catchments

The fundamental tectonic control on the distribution of sedimentary environments and lithofacies is the structural asymmetry of rift basin margins. This causes contrasting transverse drainage catchments to evolve on newly created footwall and hangingwall uplands (Leeder & Jackson, 1993; our Fig. 6). Catchment drainage area is controlled by the length of the tectonic slope produced during extension, via Hack's power law expression $L = 1.4A^{-0.6}$ that relates drainage basin area, A, to principal stream length, L (Leeder *et al.*, 1991). This is important because, for any given climate and catchment bedrock lithology, drainage basin area controls the annual outlet discharge of water and sediment, and thus the magnitude of alluvial fans, fan deltas and submarine fans along graben margins. Several recent authors (e.g. Whipple & Traylor, 1996; Allen & Hovius, 1998; Parker *et al.*, 1998; Allen & Densmore, 2000) also see a major role for the rate of hangingwall subsidence in controlling fan size. A noticeable feature of both footwall and hangingwall catchments and their alluvial fans is the regularity of their spacing (Leeder *et al.*, 1991; Hovius, 1996; our Fig. 6B). This can be considered from the point of view of Hack's scaling law. Since faults are linear features and since each catchment must evolve with respect to the position of the fault line, the linear dimension of the catchment (i.e. catchment length), is more-or-less initially determined by the tectonics. However, the rate of development of the length is determined by the type and weatherability of catchment bedrock (Fig. 6B). Once fixed in this way by structure and bedrock, all similarly lengthened faults will have the same basin area. In such a way, regular lateral along-strike spacing is produced. Important exceptions result from antecedent drainage (Fig. 6B) and when drainage growth due to fault propagation causes previous depositional bajada to be incorporated in a new footwall catchment (Leeder & Jackson, 1993; our Fig. 6D).

Transverse sediment flux

The relief of footwall uplands adjacent to main basin-bounding normal faults increases as slip accumulates, eventually reaching an equilibrium with respect to the rate of local denudation. Short narrow and steep drainage basins develop as major range-parallel watersheds migrate away from the active fault trace. Generally, larger drainage basins evolve at the locus of maximum throw along a fault segment, although local variations in lithology also strongly influence drainage area (Fig. 6). Fault offsets and transfer zones in the footwall may feature larger-than-average drainage basins. Landslides occur on steep footwall slopes, particularly during faulting episodes (e.g. Madison Lake slide during the 1959 Hebgen Lake earthquake (Hadley, 1960); our Fig. 7D).

The abrupt decrease of gradient from the footwall uplands to the hangingwall depositional basin causes rapid deposition and the construction of talus cones, alluvial fans, fan deltas and submarine fans (Figs 6 and 7). More-or-less continuous bajada may form by the coalescence of individual fans. Recent modelling studies (Whipple & Traylor, 1996; Allen & Hovius, 1998; Parker *et al.*, 1998) have emphasized the role of tectonic subsidence in determining fan and bajada width. In many areas of the world severe Quaternary climatic and base level changes have also caused fan segmentation, progradation and retrogradation. Fan segmentation may also result where fault propagation causes higher fan segments to be cannibalized during footwall uplift and incision (Leeder & Jackson, 1993; Cohen *et al.*, 1995; our Fig. 6D).

Hangingwall catchments are initially much longer than those that gradually propagate into the footwall; they are also larger (the Hack relationship), and have gentler slopes. Hangingwall-sourced alluvial fans are thus larger than typical footwall-derived fans and coalescence of

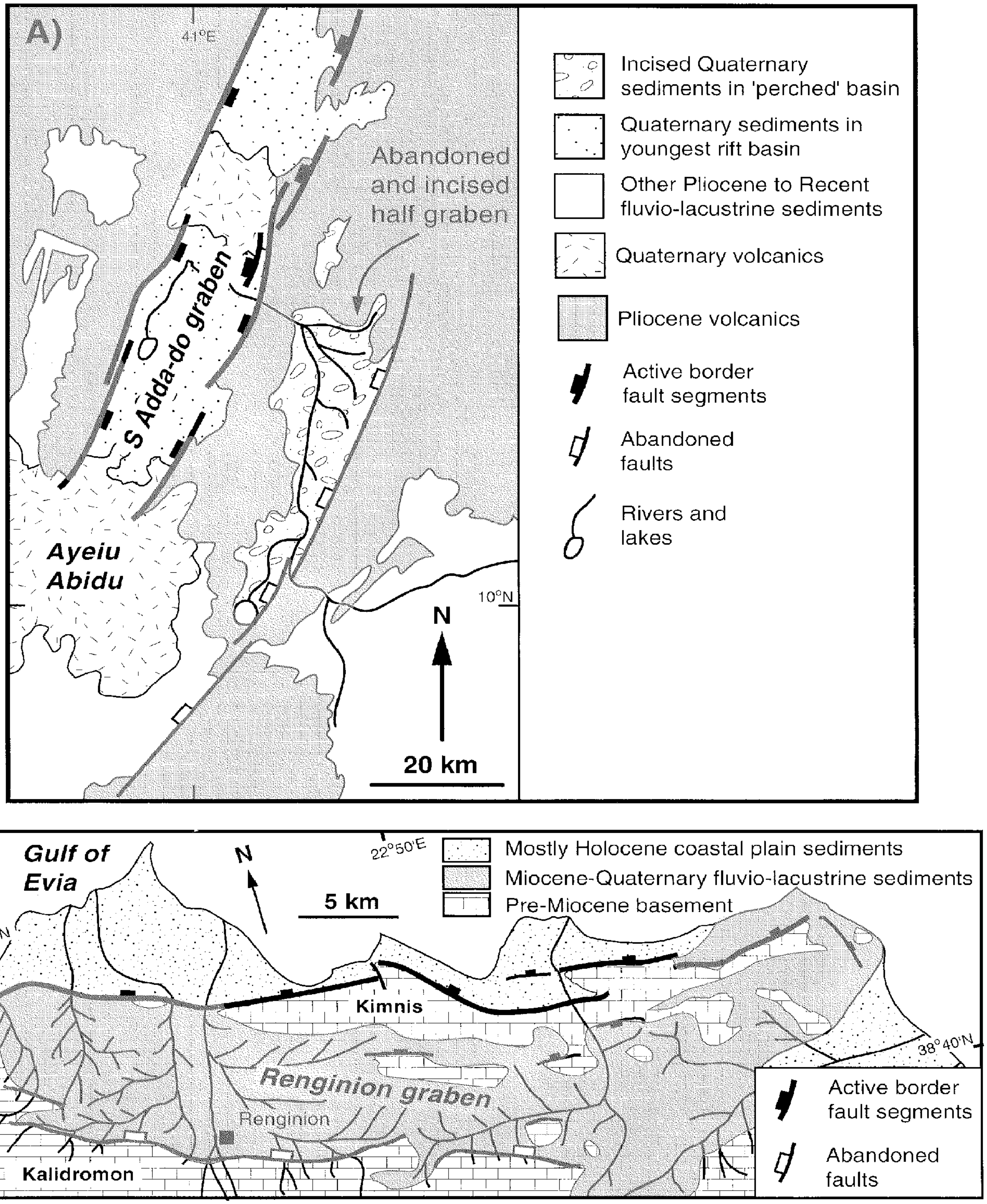

Fig. 5. Examples of the migration of the locus of active faulting resulting in 'fault death' and uplift and erosion of earlier extensional basin fills. (A) Southern Adda-do rift of Southern Afar (after Hayward & Ebinger, 1996). Note the older uplifted and incised fault segments and basin fill to the SE of the South Adda-do graben. (B) Abandoned and incised Renginion graben, central Greece (after Leeder & Jackson, 1993). Migration of the locus of active faulting to the north has uplifted Miocene–Quaternary sediments and the northerly dipping normal faults that bound the Kalidromon massif leading to incision and reworking of the basin fill.

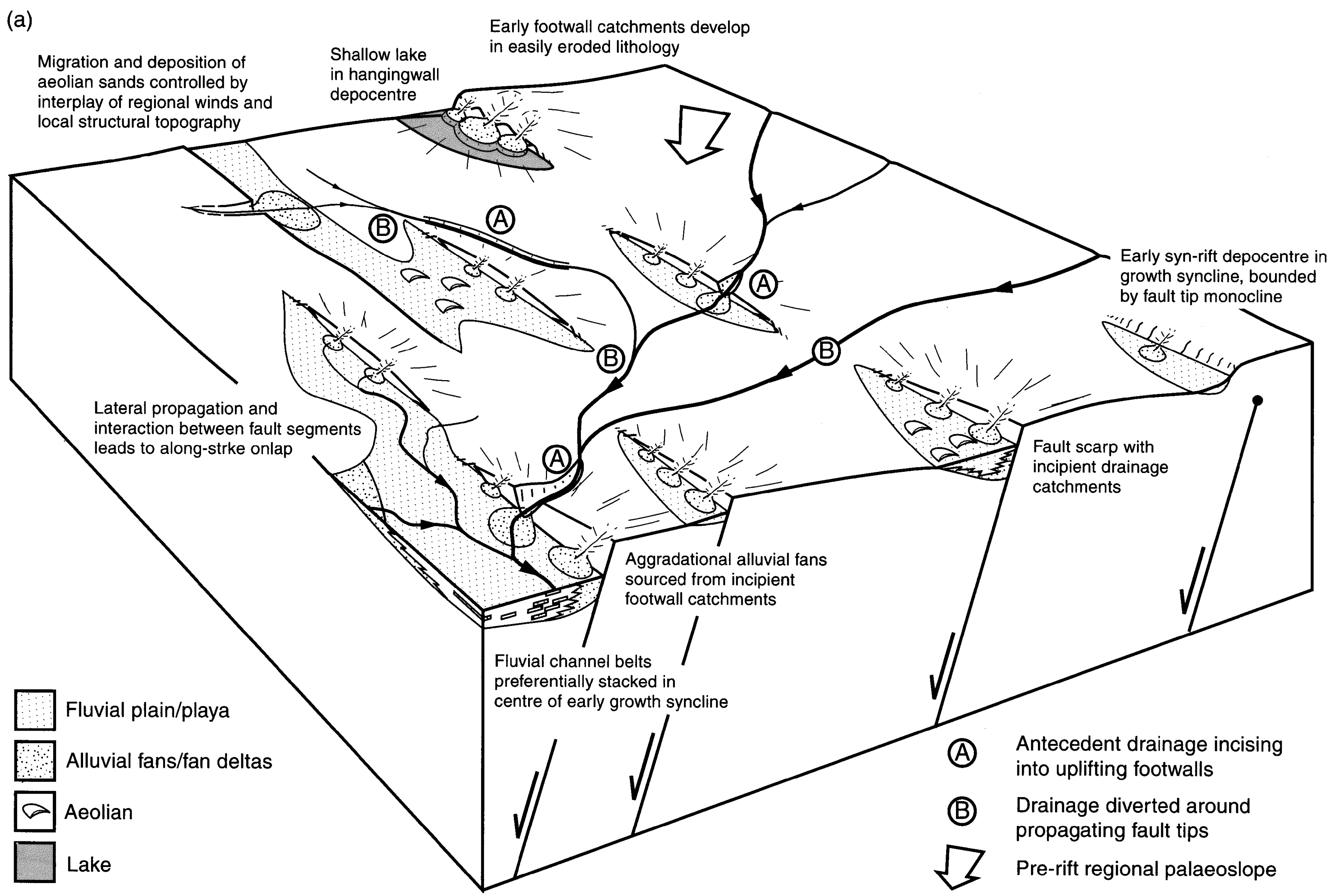

Fig. 6A. Tectono-sedimentary evolution of a normal fault array (continental environments); initiation stage. Numerous isolated fluvio-lacustrine subbasins in the hangingwalls of propagating normal fault segments. Major sediment transport pathways are dominated by antecedent drainage networks that are locally modified by surface topography associated with fault breaks and growth folds. Stratigraphic variability between individual basins is high, due to differences in sediment supply and whether surface deformation is associated with growth folds or faults.

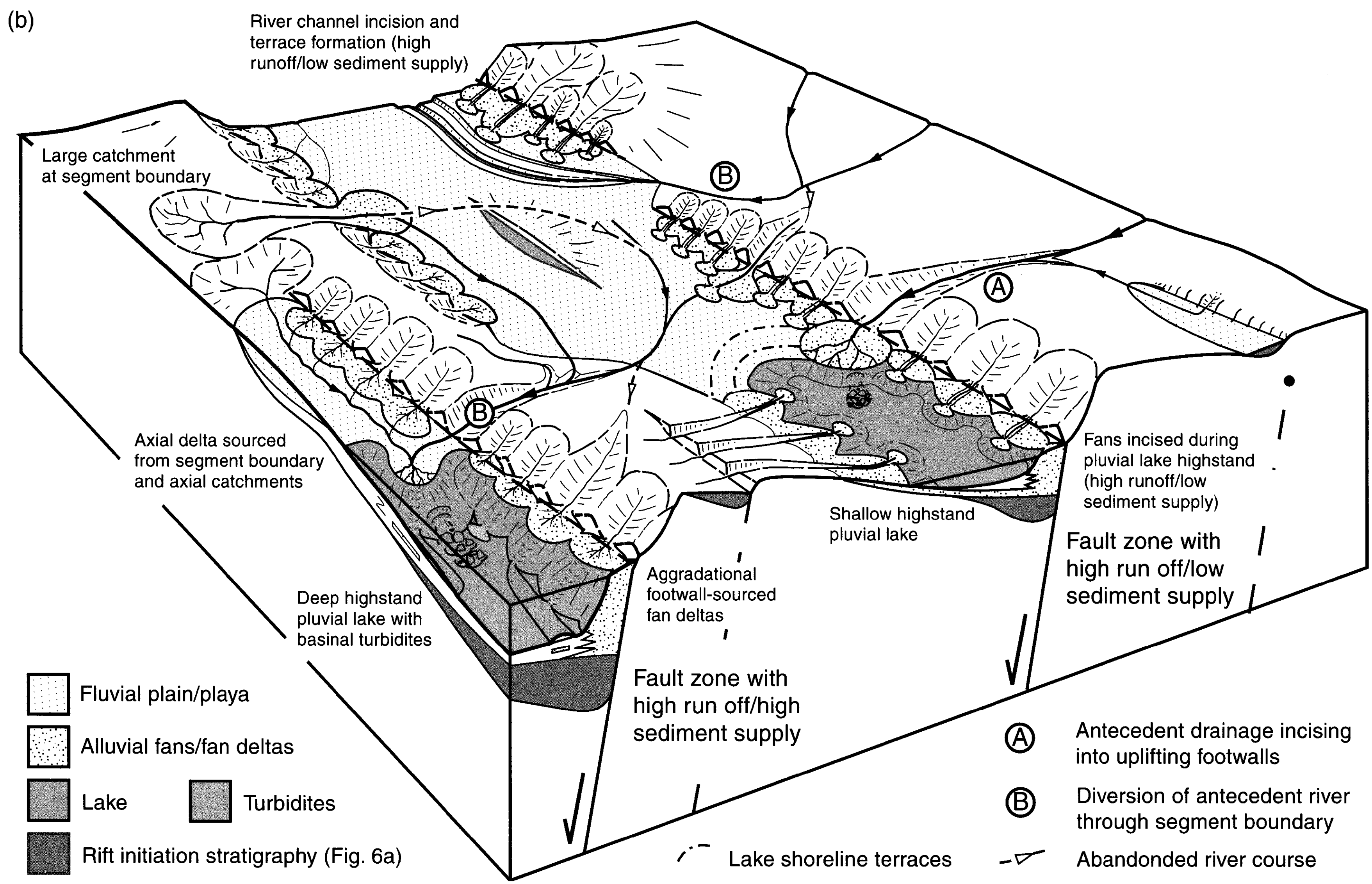

Fig. 6B. Tectono-sedimentary evolution of a normal fault array (continental environments); interaction and linkage stage. Lateral propagation and interaction between fault segments leads to enlargement and coalescence of early fault depocentres, whilst other fault segments become inactive (dashed on front face). Basin fills adjacent to inactive faults are buried and preserved if located close to the hangingwall of a major fault, or are uplifted, incised and reworked if near footwall crest. Consequent drainage catchments continue to develop along faceted footwall scarps and hangingwall dip-slopes and act as transverse sediment sources to developing half graben depocentres. Note decrease in size of footwall catchments and associated fans towards fault tips. Location of isolated lakes is largely controlled by fault segmentation. Right-hand fault zones are shown for Basin and Range style interglacial with high run off and low sediment yields. Left-hand fault zones are shown for interglacial with high run off and high sediment yield (e.g. East African lakes).

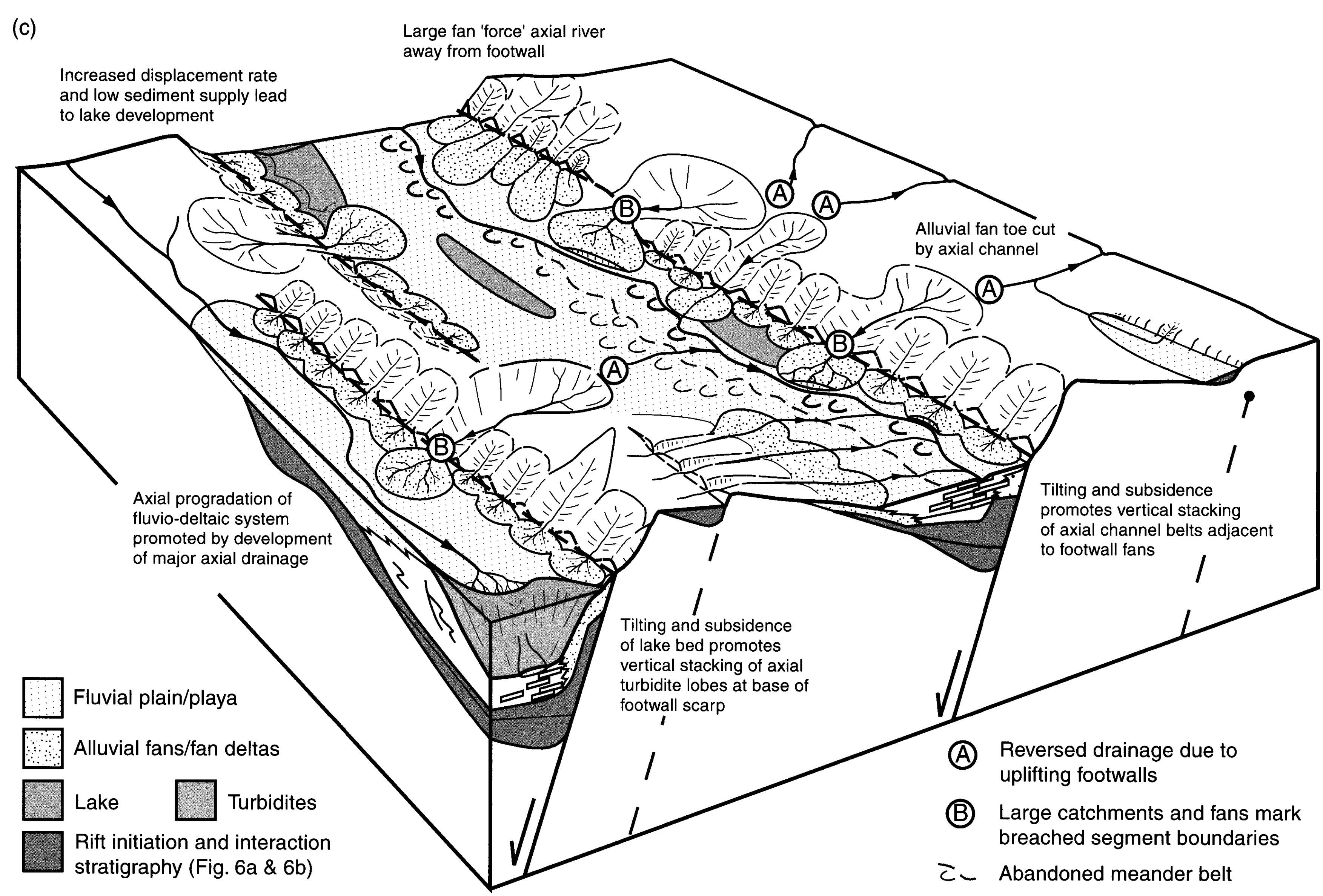

Fig. 6C. Tectono-sedimentary evolution of a normal fault array (continental environments); through-going fault stage. Linkage of adjacent fault segments creates major linked fault zones defining half graben basins. Displacement on linked faults reduces topography of former intrabasin highs, allowing axial river to flow between former isolated basin segments. Note asymmetric development of axial meander belt and interaction between meander belt and footwall fans. Localization of displacement causes increased displacement rates on active faults leading to the development of pronounced footwall topography and reversed antecedent drainage.

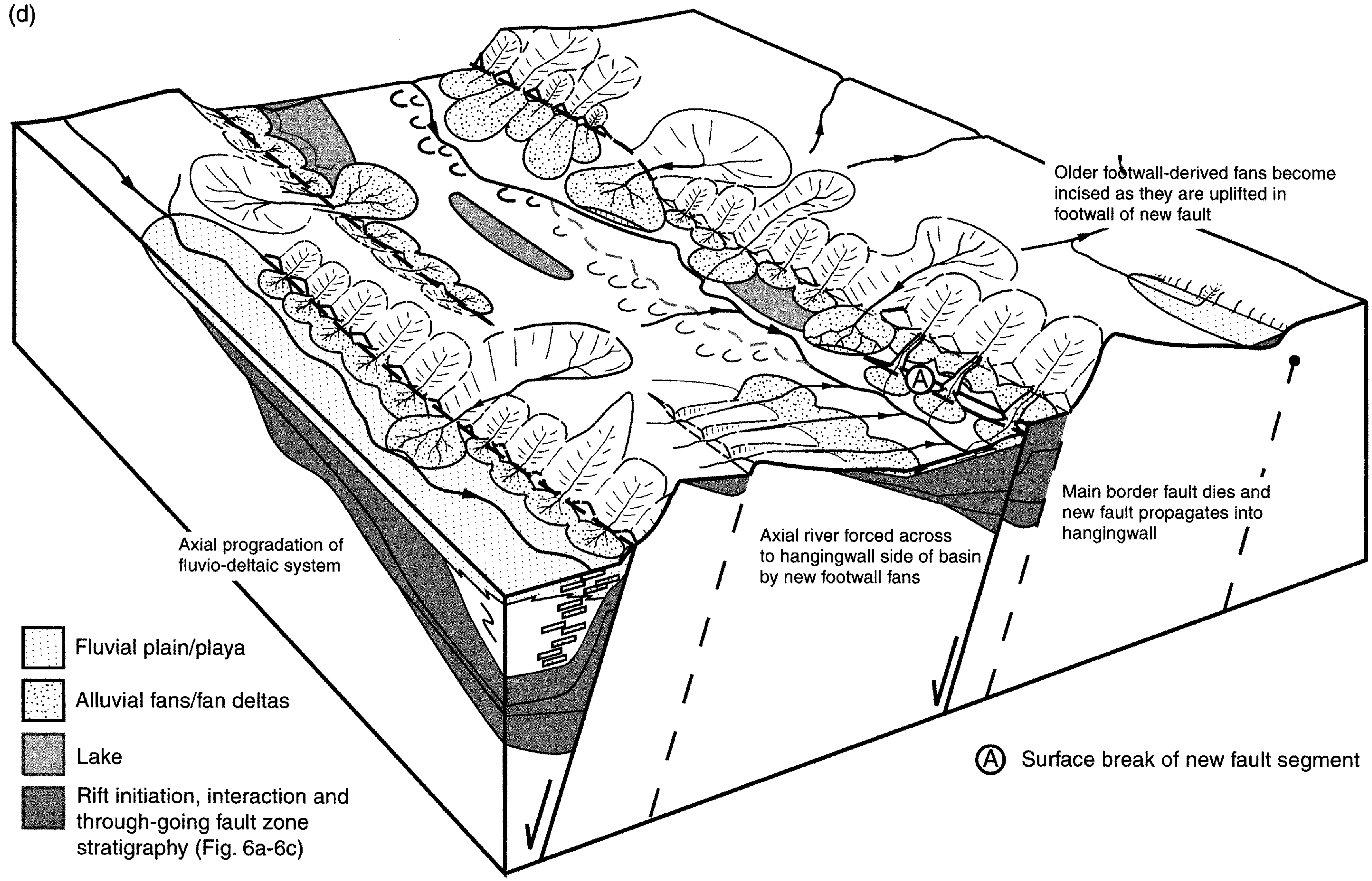

Fig. 6D. Tectono-sedimentary evolution of a normal fault array (continental environments); 'fault death' stage. Locus of active faulting migrates into hangingwall of right-hand fault zone causing uplift and incision of former footwall-derived fans and a shift of the axial river away from the rift shoulder.

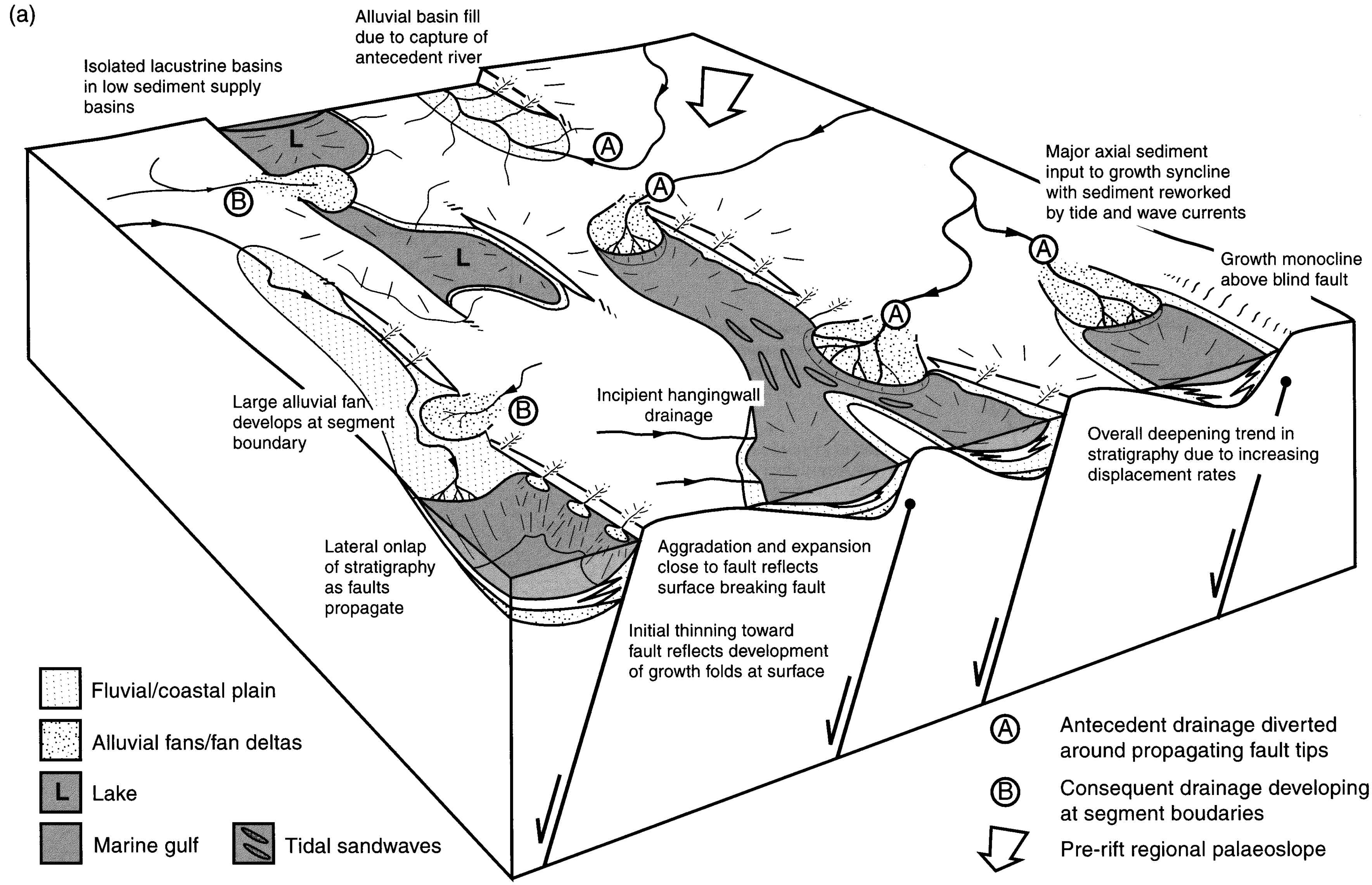

Fig. 7A. Tectono-sedimentary evolution of a normal fault array (coastal/marine environments); initiation stage. Early formed fault segments and growth folds form low-lying topography and define numerous isolated depocentres partially linked at highstands of sea level to form shallow, elongate marine gulfs and lakes. Antecedent drainage, locally modified by the evolving fault and fold topography, forms major sediment transport pathways. These early synrift depocentres display a marked variation in sedimentary fill, depending on their position with respect to sea level, sediment sources and the relationship between topography and sea level.

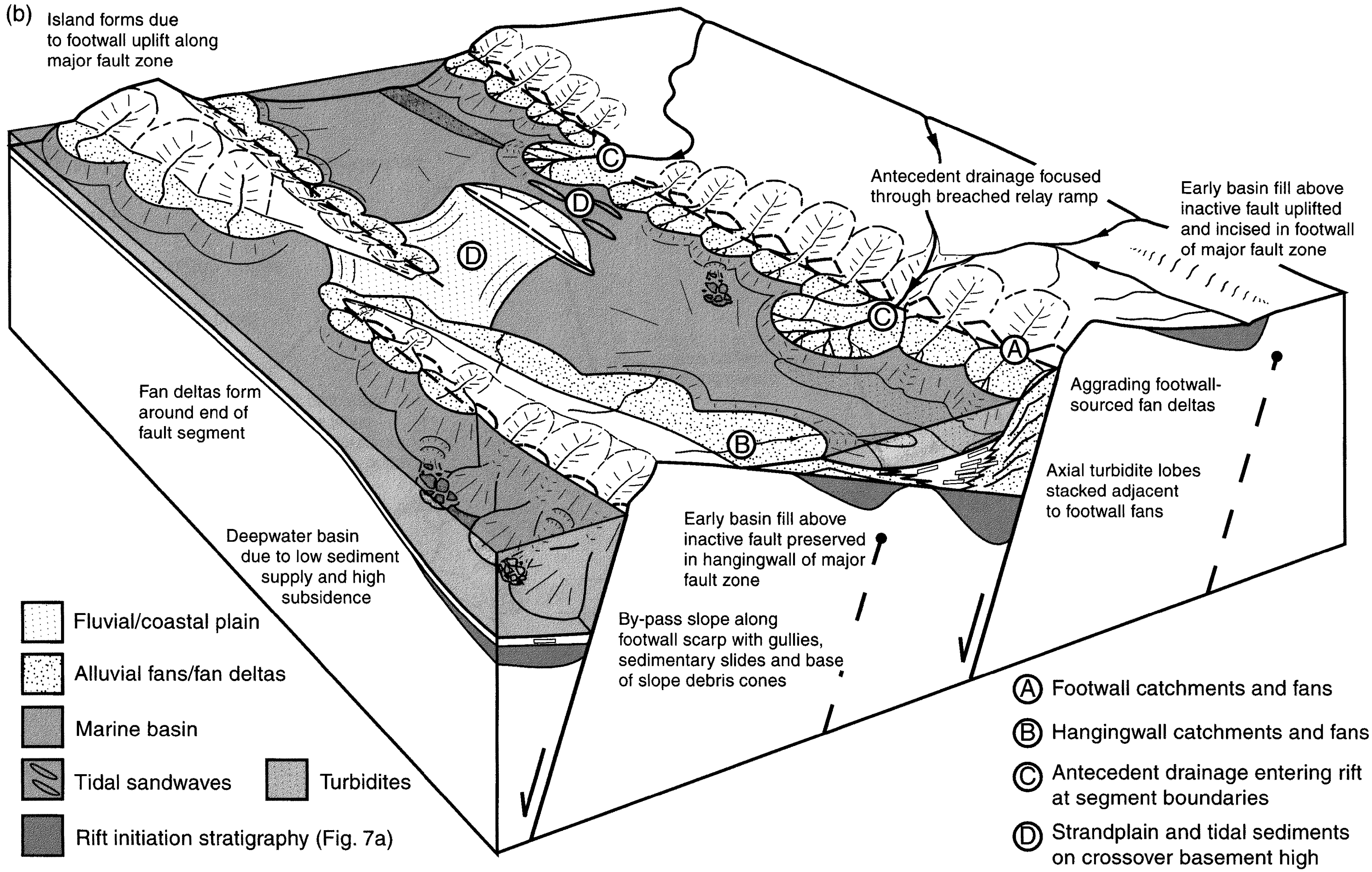

Fig. 7B. Tectono-sedimentary evolution of a normal fault array (coastal/marine environments); interaction and linkage stage during sea level highstand. Lateral propagation and interaction between fault segments leads to enlargement and coalescence of early fault depocentres, whilst other fault segments become inactive (dashed on front face). Development of drainage catchments along uplifting footwalls leads to the development of transverse sediment supply to footwall- and hangingwall-derived deltas. Right-hand fault zones along the rift shoulder are supplied by antecedent drainage that enters the rift through topographically low segment boundaries. Left-hand fault zones form isolated footwall islands. Limited transverse sediment supply from these islands leads to the development of starved, deepwater basins. Crossover basement highs at accommodation zones form shallow platforms along the axis of the rift that may become sites of shallow marine and tidal sedimentation. Tilting of the basin floor promotes axial transport of turbidites that stack and interfinger with the toes of footwall-derived deposits. Localization of deformation along major fault zones leads to increased subsidence rates that may outpace sediment supply and result in an overall deepening upward trend in the basin fill.

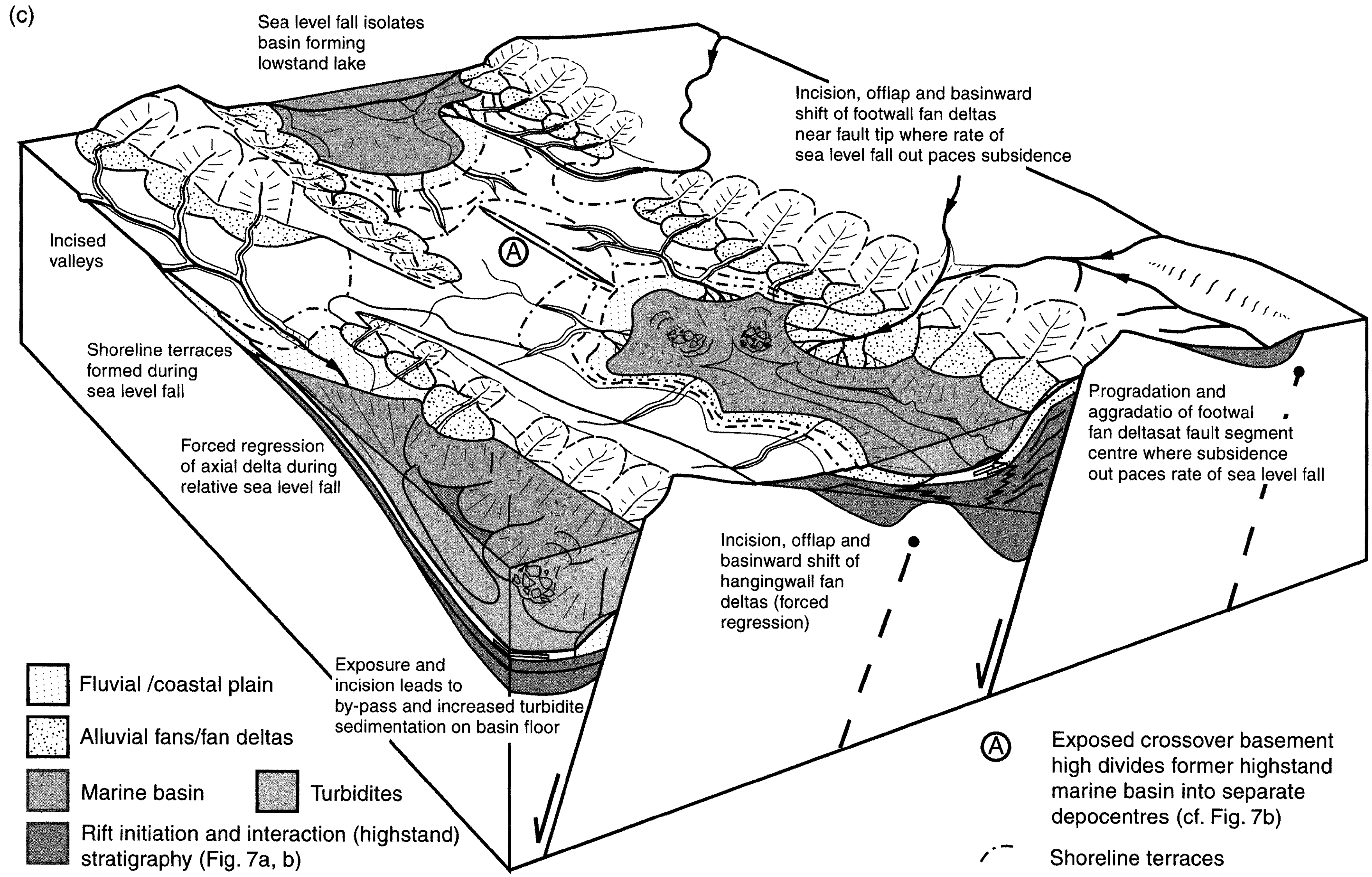

Fig. 7C. Tectono-sedimentary evolution of a normal fault array (coastal/marine environments); interaction and linkage stage during sea level lowstand. Eustatic fall in sea level results in subaerial exposure and incision in low-subsidence-rate settings (hangingwall dipslope, around fault tips/segment boundaries and crossover basement highs) resulting in marked basinward facies shifts. Exposure of crossover basement highs may lead to isolation of depocentres. In the immediate hangingwall of fault segment centres, high subsidence rates may outpace eustatic sea level falls, generating relative sea level rise and resulting in depositional systems prograding and aggrading.

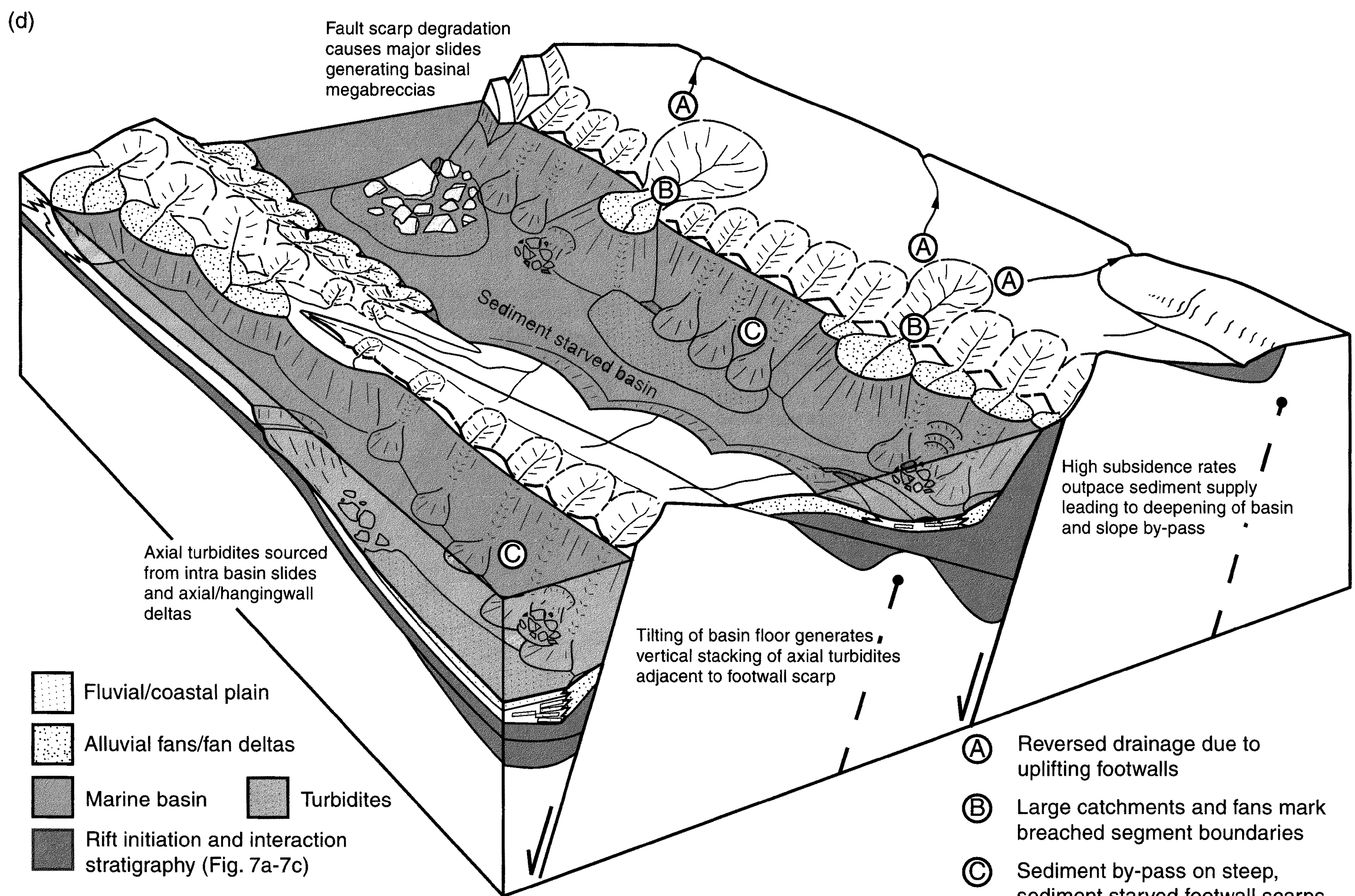

Fig. 7D. Tectono-sedimentary evolution of a normal fault array (coastal/marine environments); through-going fault zone stage. Continued fault growth and linkage leads to localization of deformation onto a limited number of major fault zones. Localization of deformation leads to high displacement rates on the active fault zones and to pronounced structural relief across the fault zones. Topographic highs associated with now breached segment boundaries along the linked fault zones become subdued and subbasins become linked. In general, high displacement rates outpace sediment supply along the fault zones, leading to retrogradation of footwall-derived depositional systems and sediment starvation in basinal areas. Fault scarps become areas of sediment bypass characterized by chutes and slump scars with base of slope talus and debris flows. Major zones of footwall instability can lead to the generation of degradation complexes.

hangingwall fans creates prominent low-angle bajadas (Figs 6B and 7B). Progressive tilting of the hangingwall dipslope towards the fault zone causes channel and fan-surface gradients to increase, resulting in fan-surface incision and lobe offlap (e.g. Hooke, 1972; Leeder & Gawthorpe, 1987).

Interaction of transverse and axial drainage

Axial rivers in open basins (Fig. 6C) are free to interact with transverse fans and drainages already established along the length of the half graben and to respond to any fault-induced tilting of the half graben floor. There must be an equilibrium balance between the magnitude of the transverse and axial water and sediment flux. Note that the magnitude of the axial sediment flux at any position is the sum of all upstream transverse components minus any deposition. High local relative transverse sediment fluxes, e.g. from large tributary streams draining easily eroded substrates, may lead to 'choking' of the axial reach, with frequent channel avulsions causing local onlap of down valley fans.

By lateral migration and avulsion, an axial channel must construct a floodplain in the space between the toes of transverse alluvial fans sourced from opposite sides of the half graben (Fig. 6C). This may be quite straightforward in basin segments that formerly held shallow lakes or playas, but in narrow segments this must proceed by active fan toe-cutting or onlap (Fig. 6C). Repeated periods of toe-cutting of transverse footwall-derived fans by the axial river are recorded in the southern Rio Grande rift and seem to be due to local tectonic mechanisms such as fault growth and hangingwall tilting (Leeder *et al.*, 1998; Mack & Leeder, 1999).

Current issues

Sea and lake level change and relative base level change

Base level (sea or lake level) change, together with tectonic subsidence or uplift, control the accommodation space in which marine or lacustrine sediment may be deposited. Furthermore, the rate of change of accommodation development (relative base level change) is one of the major controls influencing stratal geometry and facies stacking patterns, as discussed in sequence stratigraphic literature (e.g. Van Wagoner *et al.*, 1990). Of importance to the study of rift basin stratigraphy are the relative rates of base level change compared to rates of vertical deformation associated with basin development, and the nature of cyclical changes in relative base level that may arise from eustatic, lacustrine (climatic) and tectonic (fault array evolution) mechanisms (Gawthorpe *et al.*, 1994).

High-frequency global sea level changes are well documented for the late Quaternary glacial cycles (e.g. Imbrie *et al.*, 1984; Shackleton, 1987). Cycles have a dominant 4th order frequency of *c.* 100 kyr, a magnitude of $\approx$130 m, and are markedly asymmetric, with post-glacial sea level rise occupying only 20% of the time of a cycle. Rates of glacio-eustatic sea level change are rapid with an average 4th order fall of $\approx 1.5\ \mathrm{m\ kyr^{-1}}$, although higher rates of fall are associated with 5th order glacio-eustatic falls. In contrast, average rates of post-glacial sea level rise are $\approx 10\ \mathrm{m\ kyr^{-1}}$ and during the last deglaciation, around 12 ka, sea level rose at $\approx 20\ \mathrm{m\ kyr^{-1}}$ with a peak of 4 m per 100 years (Fairbanks, 1989).

Lake level changes are equally rapid and act as a major control on stratigraphic architecture in contintental half graben. For example, in the tropical rift lakes of East Africa (e.g. Lakes Malawi and Tanganyika) filling rates following the last glacial maximum are of the order of tens of metres over 10^2–10^3 years, with interannual lake level changes over the last century of the order of 6 m (e.g. Scholz & Finney, 1994; Cohen *et al.*, 1997). Over the last 100 kyr, lake level fluctuations in Lake Malawi have had amplitudes up to 350 m, with rapid rates of fall, as much as 140 m in 340 years (Scholz *et al.*, 1993).

Compilation of rates of coseismic deformation along active normal faults suggests they are comparable to all but the fastest rates of base level change (Fig. 4). As the rates of displacement vary spatially around rift basins, different structural settings will experience different relative base level histories, and thus potentially develop different sequence geometries (Gawthorpe *et al.*, 1994). Displacement rates on active normal fault zones can reach up to $5\ \mathrm{m\ kyr^{-1}}$ with an average for the centre of fault zones of 2–$2.5\ \mathrm{m\ kyr^{-1}}$. Rates of displacement on active structures show a clear enhancement for segments in the centre of fault arrays compared to distal segments. For the Wasatch fault zone there is an increase in Holocene displacement rates by a factor of five from distal segments to central segments (e.g. Machette *et al.*, 1991), and for average Quaternary rates on the Teton fault there is a similar pattern of enhancement, by a factor of 2.5 (Byrd *et al.*, 1994).

The fault displacement rates and their spatial variation are particularly important for relative sea level change and accommodation. In the centre of fault zones (location 1, Fig. 4E), where subsidence is greatest, relative sea level may continue to rise during a 4th order glacio-eustatic fall, albeit at a varying rate, because the rate of subsidence is always greater than the rate of eustatic fall. In contrast, towards the tips of fault zones (location 2, Fig. 4E), a relative sea level fall is likely during a glacio-eustatic sea level cycle. In these locations, subsidence rates are lower than at the centre of a fault segment, and the rates of subsidence are likely to be less than the rate of eustatic sea level fall.

In addition, displacement accumulation on fault zones is episodic, leading to pseudo-cyclical displacement histories with phases of earthquake clustering and rapid growth alternating with more quiescent intervals. This sort of episodic behaviour can occur on a wide range of time-scales, from millions of years (e.g. Fraser & Gawthorpe, 1990; Underhill, 1991) down to several thousand years (e.g. Machette *et al.*, 1991; Dorsey *et al.*, 1997). The results of numerical modelling of fault array evolution (e.g. Cowie, 1998; Cowie *et al.*, 2000) suggest that high-frequency clustering of activity may result from positive or negative stress feedbacks between adjacent fault segments, and thus the high frequency clustering is also spatially variable along and between fault zones. Longer-term variations in displacement rates on fault zones are also likely to be associated with the development of fault arrays (Fig. 3). Lower displacement rates characterize the early phase of evolution when displacement is distributed on numerous small faults, whereas higher displacement rates are characteristic of later in the evolution of the array once deformation has localized on a limited number of major bounding faults (e.g. Cowie, 1998; our Fig. 3). These marked spatial and temporal variations in displacement have a significant impact on relative base level and accommodation around normal fault zones. Thus, they have a profound impact on sequence evolution, with potential to develop a wide range of stratal and facies stacking patterns within contemporaneous sequences both along the strike of and between individual fault zones (Gawthorpe *et al.*, 1994; Hardy & Gawthorpe, 1998).

Sediment and water supply and climate change

The rates of water and sediment supply to a basin are fundamental in determining environments of deposition and stratigraphy. Climate, through its control on weathering rates, erosion rates, vegetation biomass and water balance, thus emerges as a major variable to consider when modelling stratigraphic architecture. A major research gap concerns the paucity of data on these variables and exactly how they control sediment and water fluxes over geological time. That supply is unsteady is in no doubt, but the magnitude and characteristic time of the unsteadiness are largely unknown.

A particularly interesting aspect of this problem in rifts concerns the delicate balance between axial and transverse sediment and water discharges. This balance determines whether the axial river will aggrade or degrade within an overall setting of basinal subsidence. The example of the ancestral Rio Grande is one of the most striking. Here the transverse and axial fluxes were in apparent overall equilibrium (disregarding local tectonic complications) for a period of >3 Myr in the Plio-Pleistocene and an overall aggradational sequence of fluviatile and alluvial fan sediments >150 m thick slowly accumulated in the rift (see Mack *et al.*, 1993). At approximately 800 ka this long-term aggradation was reversed. Since that time the axial river has periodically incised its own alluvium and that of adjacent transverse alluvial fans to a depth of up to 150 m along >1000 km of river length. Such major, basin-wide, changes in axial river behaviour seem most likely to have been due to climatically induced change in the magnitude and/or time distribution of sediment and water discharges from transverse catchments. Plentiful supply of sediment from tributaries causes lateral fan growth and gives a 'choked' axial system that responds by frequent avulsions and aggradation. Such scenarios accompanied deglaciation in temperate and high latitudes during Quaternary times and also characterized periods of changeover from tree-bearing to grass/scrub-bearing climates in Mediterranean and semi-arid climates (Leeder *et al.*, 1998). By way of contrast, if increased water discharge from tributaries to an axial river is accompanied by reduced sediment input, the result is likely to be incision, terrace formation and soil development (Reheis *et al.*, 1996).

As discussed in more detail by Leeder *et al.* (1998), Pleistocene catchments under Great Basin-type climates produced low sediment discharges during full-glacial climate and sea level lowstand. Under these conditions, incised valleys at the shelf/coastal break allowed sediment bypass and erosion to occur, but the lack of sediment availability limited development of forced regressive and lowstand shallow marine, deltaic and turbidite deposits. For lake environments the situation was different, with cool, wetter glacial maxima and lowered sediment yields coinciding with high lake levels due to a combination of increased run off and decreased evaporation. Here the development of upstream fluvial incision coincided with the production of highstand deltas whose low sediment supply causes low rates of highstand lake-margin delta progradation.

By way of contrast, during northern Mediterranean glacial intervals, catchments fed more sediment and water to marine and lake environments (Leeder *et al.*, 1998; Collier *et al.*, 2000). Marine lowstands thus coincided with aggradational conditions in river basins away from the influence of the shelf-coastal break. The high sediment supply allowed the development of volumetrically significant forced regressive and lowstand wedges and prominent submarine fans. During interglacials sediment supply was much reduced and rivers incised. In lakes, increased winter run off during glacial times led to high lake levels at the same time as enhanced sediment supply. Such variations in the stratigraphic response between lacustrine and marine settings, and between different climatic belts, highlight the importance of local variations in physiography, climate and sediment supply in controlling depositional sequences.

Basin linkage – evolution from closed to open rifts

The distinction between closed and open rift segments is a fundamental one for rift architecture. Rift segments are initially topographically closed systems, since drainage between adjacent rift segments is prevented by the development of hangingwall basement highs at the segment boundaries. Such closed basins are hydrologically distinct, water and sediment fluxes are largely conserved and the architecture of the rift sedimentary infill is distinctive, with lakes prominent (Fig. 6B). Open basins exhibit regional hydrological gradients, a significant part of the water and sediment flux is exported and the central basin axis and marginal fans are dominated by the erosive and depositional activities of axial drainage channels (Fig. 6C).

The key question here concerns how crossover basement ridges might be breached, enabling along-strike linkage of depositional systems. The question is analogous to discussions of structural linkage between developing faults in relay zones, which have to breach the relief features of relay ramps before they can link laterally with a near neighbour. In all cases linkage will be facilitated by any potential relief difference between the floor of one rift segment and another. On a large scale, this may be due to a regional gradient, such as that between the northern and southern ends of the Rio Grande rift. Thus the altitude of the basin floor in the Taos rift segment to the north is 2150 m, whereas the Mesilla basin at Las Cruces, 500 km to the south, is 1175 m. On a smaller scale, steep gradients may be due to the juxtaposition of segments whose faults differ significantly in spacing, such that a narrow, shallow basin can feed into an adjacent deep wide basin as between Pleasant Valley and Dixie Valley, in the Central Nevada Seismic Belt (Jackson & Leeder, 1994). There are several possible reasons why sedimentary linkage between fault segment depocentres might occur.

- Purely structural re-organization due to lateral fault growth, interaction and linkage (Figs 2 and 3).
- Sediment deposition may eventually outpace the growth of differential relief between basin floor and crossover ridge. Basement ridges are thus onlapped and eventually overwhelmed by sediment, enabling drainage from the higher segment to be transferred into the lower segment via overspill channels.
- In magmatically active rifts, volcanics may infill an isolated rift segment to brinkpoint at the crossover. This appears to have happened in the Northern Rio Grande rift segment infilled by the Taos Plateau volcanics at about 4 Ma.
- Steeper drainages incised on emergent basement ridges between basins of unequal depth may capture up valley by active headcutting, thus eliminating the drainage divide.
- Lake expansion in a narrow rift segment may be sufficient to cause spillover and cutting of an overspill channel through the basement ridge. This is an attractive general scenario for Pleistocene linkage of pluvial lakes.

TECTONO-SEDIMENTARY ARCHITECTURAL MODELS

Continental rift basins (Fig. 6)

Initial stages of extension (Fig. 6A) are represented by small fault- and fold-bounded basins. These early depocentres interact with pre-existing drainage until the tectonic slopes associated with faulting can dominate the landscape. The new tectonic slopes cause river deflection and topographic funnelling to occur; elsewhere antecedent drainages cut gorges through rising footwall blocks (Fig. 6A).

Lakes are common features of structurally isolated half graben (Fig. 6A, B). They are sinks for clastic sediment introduced by transverse drainages as well as for biogenic and chemical sediments. Lake environments include playa, semipermanent and permanent systems forming according to the level of local freshwater influx relative to evaporation. The rapid subsidence of rift floors enables very deep water conditions (cf. Lake Baikal, – 1640 m) to exist should the water balance be favourable.

Arid, closed continental basins commonly contain playa lakes whose chemical deposits reflect the ionic composition of groundwaters and run off from surrounding uplands. Shallow permanent lakes in less evaporative climates, such as those in the Baringo–Bogoria part of the Gregory rift, Kenya, and Lake Malawi, are fringed by fan and axial deltas, sublacustrine fans and basinal biogenic oozes and fine-grained clastic sediment (e.g. Tiercelin *et al.*, 1992; Le Turdu *et al.*, 1995; Soreghan *et al.*, 1999). Deeper lakes develop permanent stratification allowing good preservation of organic matter and the development of seasonal varves. For example, Lake Malawi is thermally stratified, with mixing only occurring in the upper 250 m of water, below which the lake is permanently anoxic (Halfman, 1993).

Lakes are highly sensitive to climatically induced changes in water balance; the resulting rise and fall of lake level exerts a fundamental control on basin architecture. Sedimentary cycles resulting from climate change are prominent in the records of all Pleistocene to Holocene lake basins (e.g. the sublacustrine fans in Lake Malawi documented by Soreghan *et al.*, 1999). Seismic reflection studies have identified both highstand and lowstand coarse-grained facies, including prominent subaqueous channels formed by fluvial down-cutting during lowstands. These channels and their well-developed levees are commonly parallel to the major bounding faults and their position may be controlled by intrarift synthetic faults (e.g. Soreghan *et al.*, 1999). During highstands, coarse clastic deposition occurs in the deep lake basins by turbidity currents issuing from subaqueous channels and by downslope turbulent dilution of sediment gravity flows and slumps. Small highstand deltas develop, but tend to be prone to erosion during lowstands. Lowstand deltas prograding into the much-reduced lakes have the highest preservation potential.

Wind reworking of alluvial and lake shoreline sands is prominent in many arid and semi-arid basins. The exact location of small ergs reflects the orientation of the basin relative to dominant prevailing winds. The aeolian sands interfinger with and are partially reworked by river channels (Fig. 6A). A fine example occurs in the San Luis basin, Colorado, where the Great Sand Dunes National Monument occupies the eastern end of a graben, close by a mountainous footwall which blocks the westerly sand-supplying winds.

As discussed previously, erosion and incision of cross-over highs causes linkage of basins, enabling discharge to become axial (Fig. 6C). Axial rivers are prone to the effects of tilting caused by individual or cumulative motions of the active bounding fault. The theoretical effects of such tilting were explored (Bridge & Leeder, 1979) long before adequate documentation of the effect was obtained from the Holocene sedimentary record (Leeder & Alexander, 1987; Alexander *et al.*, 1994; Blair & MacPherspon, 1994; Peakall, 1998). It seems from these studies that the axial river may either undergo steady towards-fault migration, leaving behind distinctive asymmetrical meander belts, or migrate towards the fault in a series of more sudden avulsive movements that may be reactions to individual faulting episodes (Peakall, 1998). Towards-fault movements of an axial river may be so severe that pre-existing alluvial fans suffer extreme toe-cutting as a result (Leeder *et al.*, 1998; Mack & Leeder, 1999). Such episodes have a high preservation potential since diminution of tilting by tectonic quiescence, erosion and deposition will cause subsequent rebuilding outwards of the fans.

Coastal/marine rift basin gulfs (Fig. 7)

There is little information on the structural style and stratigraphic response of marine-connected basins during the initial stages of extension. However, data from the Oligo-Miocene of the Gulf of Suez rift suggest a scenario of isolated, shallow-water, blind-headed gulfs with micro- to meso-tidal regimes fed by small-scale drainages or larger antecedent rivers (our Fig. 7A; Richardson & Arthur, 1998; Gawthorpe *et al.*, 1997; Sharp *et al.*, 2000a,b). Subsidence rates and structural relief between depocentres and footwalls are generally low at this initiation stage (Fig. 3A). Within individual structurally controlled depocentres two distinct tectono-stratigraphic styles can be identified: (i) an initial growth fold stage when the fault is a buried structure, and (ii) a subsequent surface faulting stage. During the growth fold stage, strata thin and become truncated towards monoclines above blind normal faults, and are rotated and diverge away from them, into hangingwall growth synclines. Sea level changes superimposed on such structural rotations lead to sharp-based, forced regressive shallow marine sand bodies that fine rapidly offshore into mudstones within growth syncline axes (Gawthorpe *et al.*, 1997). In contrast, once the fault breaks surface, strata form a divergent wedge, which is rotated and thickens into the fault, producing a stratal geometry more typical of half graben.

Fault linkage and interaction leads to concentration of deformation along a limited number of major faults and many early fault segments become inactive; their depocentres may become passively infilled, or if situated near footwall highs, incised and reworked (Sharp *et al.*, 2000b; Fig. 7B, C). During this stage the major faults experience higher displacement rates than during the initiation stage, as illustrated by backstripping analysis of well data from the Gulf of Suez (e.g. Steckler *et al.*, 1988). The development of pronounced footwall topography and tilting leads to characteristic graben and half graben structural styles, with transverse sourced fan deltas similar to those developed in lacustrine basins (cf. Figs 6B and 7B). Elongate marine half graben, typified by the Gulf of Corinth, are characteristic of the basin physiography during this stage of evolution (Fig. 7B, C), with basin depth depending upon rates of fault-induced subsidence relative to sediment infill. Important controls upon sedimentary regime are the magnitudes of tidal and wave-induced currents. The former depend upon basin morphology in relation to local tide-generating forces, the latter on the magnitude and direction of mean wave fetch. On Fig. 7(B) we indicate both wave-dominated accreting strandplains (e.g. Collier, 1990) and tidal current ridges developing along a narrow seaway between two subbasins at highstand, rather in the spirit of the former Isthmian tidal stream discovered by Collier & Thompson (1991). Along-strike variations in the balance between accommodation space creation and sediment supply may generate contrasting stacking patterns. Progradation is favoured by high sediment supply, low accommodation settings such as along hangingwall dipslopes or at segment boundaries, whereas retrogradation may occur in high-accommodation, low-sediment supply locations in the centre of fault segments (Fig. 7B, D). During lowstands of sea level the degree of fluvial incision and basinward shifts in facies may also be spatially variable due to the intrinsic variations in subsidence rates and slope gradients (our Figs 4E and 7C; Gawthorpe *et al.*, 1994; Hardy & Gawthorpe, 1998). Furthermore, during sea level lowstands, marine rifts may become isolated from their mother oceans and depending upon the local water balance develop into either deep freshwater or shallow saline lakes.

Important structural controls upon sediment sequences in marine rifts include the episodicity (Dorsey *et al.*, 1997) and migration (Leeder & Jackson, 1993) of the locus of active faulting. Thus in the Loretto rift, Baja California, Dorsey *et al.* (1997) recognize the occurrence of high-frequency (*c.* 10^4 years) fan delta retrogradation related to repeated clustering of earthquake-subsidence events along a basin margin fault. Along the southern margin of the Gulf of Corinth, successive generations of footwall-sourced fan deltas have formed in response to northward fault-migration (Ori, 1989; Dart *et al.*, 1994; Gawthorpe *et al.*, 1994). Modern fan deltas in the area are linked via submarine channels to submarine fans (our Fig. 7B–D; Ferentinos *et al.*, 1988). Steep subaqueous slopes, mantled by fine clastic and biogenic hemipelagic sediment, are cut by subaqueous channels that issue from both axial and transverse alluvial channels and fan deltas and feed basinal turbidites. The bypassed slopes and delta fronts are susceptible to mass failure during earthquakes, and sidescan and high-resolution seismic profiling reveal a plethora of slump scars, slumps and debris flow lobes (e.g. Papatheodorou & Ferentinos, 1993; our

Fig. 7B–D). Some of these are transformed to turbidity currents as they move downslope (Fig. 7C, D). These features are characteristic of many sediment starved marine rift basins at the through-going fault stage (Fig. 7D), such as the Late Jurassic of the North Sea (e.g. Rattey & Haywood, 1993) and the middle Miocene of the Gulf of Suez (Patton *et al.*, 1994).

The arid Miocene Suez rift exhibits only relatively small-scale alluvial fans and fan deltas whose largely inactive fringes are extensively colonized by patch reefs. Detailed studies in areas close to major transfer zones along the faulted margins reveal the initial presence of successive and aggradational fan delta lobe deposits sourced from larger drainage basins located within transfer zone uplands (e.g. Gawthorpe *et al.*, 1990; Gupta *et al.*, 1999; Sharp *et al.*, 2000a). Some of these are organized into spectacular growth folds and display marked along-strike variations in stratal geometry and stacking patterns between fault segment centres and propagating fault tips (see Fig. 4E). Away from areas of significant clastic input, spectacular reef and reef-talus development occurs.

CONCLUSIONS

It has been the purpose of this contribution to present updated conceptual models for the tectono-sedimentary evolution of rift basins. Recent developments have shown in particular that basin-fill architecture depends upon a rather complex interaction between 3D evolution of basin linkage through fault propagation, the evolution of drainage and drainage catchments, the effects of changing climate, and variations in sea/lake level. In particular, we identify the processes of fault propagation, growth and death as major tectonic controls on basin architecture. Current theoretical and experimental models of fault linkage and the direction of fault growth are testable using observational evidence from the earliest stages of rift development. Basin linkage by burial or breaching of relays and crossover basement ridges is the dominant process whereby hydrologically closed rifts evolve into open ones. Nontectonic effects arising from time-varying climate and sea/lake level are responsible for major changes in basin-scale sedimentation patterns. Major gaps in our understanding of rift basins remain because of current inadequacies in sediment, fault and landscape dating.

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