

# Tracing tectonic deformation using the sedimentary record: an overview

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**Abstract:** Tectonic activity, on a range of scales, is a fundamental control on sedimentary activity. The range of structural deformation within a region extends from the plate tectonic scale, governing, for example, rift initiation, to the basin scale, with the formation of basin-bounding faults. Internal basin configuration is also strongly influenced by tectonic activity. However, the relationship between tectonic activity and sedimentation is a complex one, given the many additional factors which can also influence sedimentary activity, including erosion, sediment transport, source area lithology, groundwater chemistry, range of depositional environments, climate, eustasy, and the relative location of an area and its distality to marine influences. In this paper we provide a selective overview of the issues associated with the interlinked themes of tectonics and sedimentation, examining the main basin types forming in both extensional and compressional plate settings. We then review the various models of sedimentation in the selected basins, both on a local and a basinal scale. Finally, we look to the future – providing a series of possible research areas, almost exclusively multidisciplinary, which would help to improve existing models of interlinked sediment–tectonics systems.

Sedimentary basins, and the depositional successions within them, provide the most tangible and accessible records of the lithospheric, geographical, oceanographic and ecological developments which occur in a specific area over a specific period of time. Tectonic activity, on a range of scales, is a major control on sedimentary activity. In recent years there has been an increase in the number of studies aiming to unravel the links between tectonic events and sedimentary response, both on a basin and intrabasinal scale (e.g. Blair & Bilodeau 1988; Heller *et al.* 1988; Cas & Busby-Spera 1991; Fisher & Smith 1991; MacDonald 1991; Williams & Dobb 1993; Schwans 1995; Cloetingh *et al.* 1997; Gupta 1997).

The range of structural deformation within a region extends from a plate-tectonic scale (e.g. rift initiation to oceanic-ridge formation) – affecting the changing pattern of the oceans and continents, and controlling the size and nature of large source areas, sediment transport pathways and the locations of sediment depocentres – down to the basin scale, where tectonics control the formation of major basin-bounding faults which determine basin form and location. Additionally, tectonic activity also controls the internal basin configuration, for example through the development of smaller intra-basinal faults (both synthetic and antithetic as well as transfer faults) that

influence the internal structure of the basin, segmenting it into related but separate depocentres (e.g. Jeanne d'Arc Basin, Tankard *et al.* 1989). These intrabasinal structures also influence the development of topography within a basin by controlling the location of both highs and lows which respectively act as potential sediment sources and sinks, and help to determine channel pathways for sediment (e.g. Alexander & Leeder 1987; Leeder & Jackson 1993; Anders & Schlichte 1994; Burbank & Pinter 1999). The broad pattern of faulting within a basin is determined by both the overall geodynamic setting (i.e. divergent, convergent or strike-slip), and by pre-existing crustal weaknesses which can strongly influence fault initiation and location.

Sedimentation results from the interaction of the supply of sediment, its reworking and modification by physical, chemical and biological processes and the availability of accommodation space, i.e. the space available for potential sediment accumulation. Many of these factors have a tectonic component. For example, sediment supply may vary in volume, composition and grain size, as well as in the mechanism and rate of delivery. These variations are largely controlled by the processes noted above. Similarly, accommodation space is controlled by sea-level variation, although relative sea-level changes may have a significant tectonic component.

Tectonic activity, therefore, is a very fundamental control on sedimentation and sedimentary activity. Similarly, the origin of the sedimentary sequences which are deposited within a basin can be related back to the tectonic activity which controlled them. The relationship between tectonic activity and sedimentation, however, is a complex one, given the large range of factors which can influence sedimentation within a basin, including: the rate and magnitude of tectonic activity, the number of faults which are active at any specific time within a basin (including their deformation histories), the rate and magnitude of sediment production (including erosion and sediment transport), the lithological composition of the source area(s), the chemistry of basinal waters, the range of depositional environments, climate, eustasy, and the location of the depositional area and its distance from marine influences (i.e. continentality). Given the inherent variability of all of these factors (together with the fact that many of them are interlinked, e.g. climate and erosion), any basin system is, by default, a complex one. Therefore, modelling of the evolution of the basin infill is difficult, since each individual basin will have its own particular tectonosedimentary signature. Additionally, there is the problem of differentiating between the various factors which influence the composition and distribution of the sedimentary succession within a basin.

Changes in our understanding of the inter-relationship of tectonic activity and sedimentation have occurred in several disciplines which play a central role in basin analysis. These include plate-tectonic theory (e.g. Cox & Hart 1986), new geodynamic models, as well as a revolution in our understanding of modern depositional systems, and consequent major advances in the sophistication of actualistic depositional models (e.g. Walker & James 1992; Miall 1997; Reading 1998; Leeder 1999). Petrological models relating sediment composition, especially sand and sandstone, to plate tectonic settings have also been developed (e.g. Dickinson & Suczek 1979; Dickinson 1988; Bahlburg & Floyd 1999), and this work has been extended into the fields of sedimentary geochemistry (e.g. Bhatia 1983; Roser & Korsch 1986; Clift *et al.* 2001) and single grain analysis (e.g. M. Smith & Gehrels 1994; Götze & Zimmerle 2000). New exploration techniques, especially seismic and sequence stratigraphy (e.g. Vail *et al.* 1977; Brown & Fisher 1979; Wilgus *et al.* 1988; Van Wagoner *et al.* 1990; Thorne & Swift 1991; Emery & Myers 1996) have led to a greater understanding of the importance of viewing basins as broad units (in a chrono-

stratigraphic sense) rather than isolated regions. Analysis of the sedimentary succession within a basin, therefore, enables us to determine some of the possible controls on the sedimentary record, and at a range of scales ranging from provenance or the examination of sedimentary structures, up to the recognition and classification of architectural elements and sedimentary sequences, and the reconstruction of depositional environments. Thus, the sedimentary record provides us with a unique opportunity to investigate the tectonic controls which are of significant interest in basin analysis.

Our objective in this paper is to provide a selective review of the linkages between tectonics and sedimentation, and more specifically, studies that have used evidence from the sedimentary record to reconstruct the tectonic history of a region. This overview will initially focus on the main types of tectonic settings and the sediments that are found in conjunction with them. Subsequent sections will examine the various models in use, summarizing with an overview of the current gaps in our knowledge and suggestions for future research areas.

### Basin classification

Basin classification schemes vary according to the particular needs of the user. For example, schemes which originate in the field of hydrocarbon exploration (e.g. Kingston *et al.* 1983a, b) are designed to be used in a predictive manner and tend to be limited to the main basin types (particularly those of interest to the hydrocarbon exploration industry). In contrast, academic classification schemes (e.g. Ingersoll & Busby 1995) tend to be more complex, since they tend towards inclusivity and completeness (Table 1). In this latter scheme, basin types are broadly grouped into those which are formed in divergent plate geodynamic settings (including continental rift basins), those which occur in intraplate settings (including intracratonic basins, oceanic islands and dormant ocean basins), those which form in convergent plate geodynamic settings (including arc-related basins, foreland basins, and trenches), those which are found in transform settings (including transtensional and transpressional tectonics) and a final group which includes basins located in hybrid settings (Ingersoll & Busby 1995). In our overview of basin types, we have chosen to follow the scheme proposed by Ingersoll & Busby (1995) but have simplified it by subdividing the basins into broad geodynamic contexts. Using this approach, it is clear that there are a number of different processes occurring within basins and that these

Table 1. Basin classification (modified after Dickinson, 1974, 1976a; Ingersoll, 1988b; Ingersoll & Busby, 1995).

Tectonic setting	Basin type
Divergent	Terrestrial rift valleys
Intraplate	Intracratonic basins
	Continental platforms
	Active ocean basins
	Oceanic islands, aseismic ridges and plateau
	Dormant ocean basins
Convergent	Trenches
	Trench-slope basins
	Fore-arc basins
	Intra-arc basins
	Back-arc basins
	Retro-arc foreland basins
	Remnant ocean basins
	Peripheral foreland basins
Piggyback basins	
Transform	Transtensional basins
	Transpressional basins
	Transrotational basins
Hybrid	Intracontinental wrench basins
	Successor basins

processes are mainly determined by the geodynamic context, but are also influenced by the locations of pre-existing weaknesses and intra-basinal processes (e.g. generation of synthetic and antithetic faults). We have divided our basins into two main groups – those which are formed within broadly extensional tectonic settings (and which would include basins found in convergent plate zones but which exhibit an extensional

character, i.e. which involve some component of rifting) and those from compressional settings. Such a subdivision greatly simplifies the characterization of the particular tectonic and sedimentary features of each basin.

## Extensional settings

### Introduction

Basins that form within an extensional tectonic setting are characterized by the development of depressions, bounded by normal faults, within which there is a direct relationship between fault activity and sedimentation. In their landmark paper, Leeder & Gawthorpe (1987) provided a clear outline of the influence of movement along an individual fault on the resultant sedimentary unit. The surface length of individual historical normal-fault ruptures range from 10–15 km (Leeder & Gawthorpe 1987), although longer basin-bounding faults (up to 50 km) occur in parts of the East African Rift (Ebinger 1989). In active extensional areas, individual fault displacements are of the order of several metres, although displacement varies from a maximum at the centre of the fault surface to zero at an elliptical tip-line (Fig. 1). Fault displacements vary systematically and there is a clear relationship between the amount of displacement and the size of the individual fault (Walsh & Watterson 1988) (Fig. 2). An exception to this rule would be the so-called ‘superfaults’, which are characterized by very large displacements occurring during a single slip event (Spray 1997).

Fault activity leads to the superimposition of a tectonically-induced gradient, the magnitude of which is determined by fault displacement, on to a pre-existing topographic one. As noted by

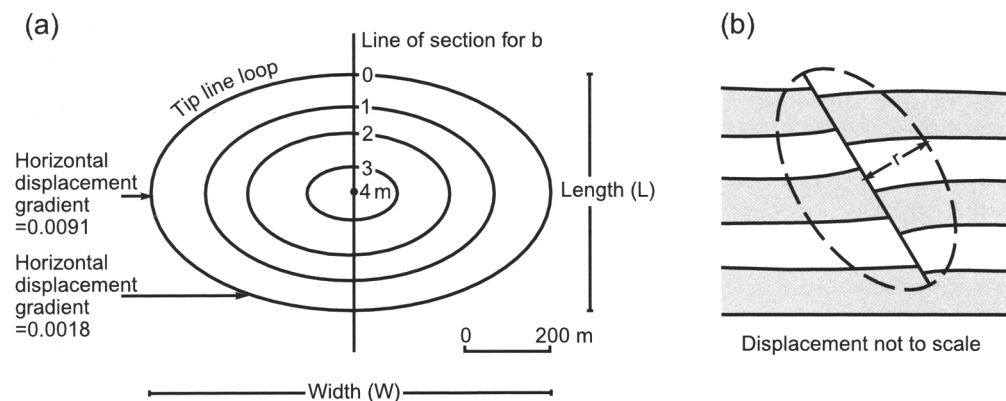
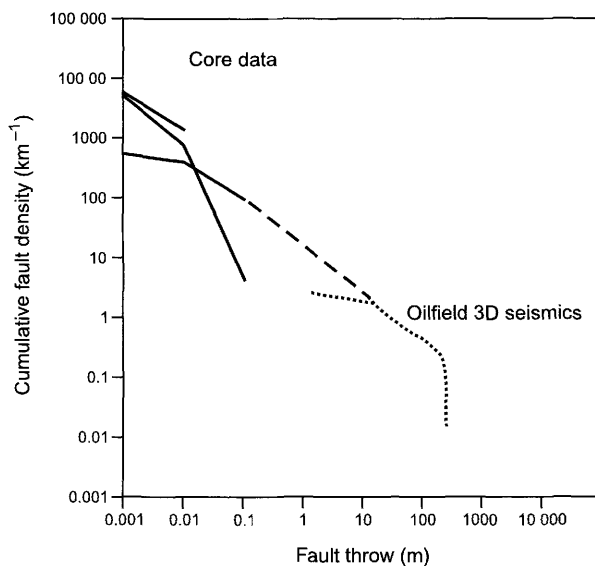


Fig. 1. (a) Schematic displacement contour diagram for a simple fault. View is normal to fault surface. (b) Cross-section showing perceptible reverse drag associated with simple fault (after Barnett *et al.* 1987).



**Fig. 2.** Comparison of fault displacement measurements on core data and oilfield three-dimensional seismics. Numbers of faults are normalized to cumulative fault density (number of faults per unit length of sample line), and displacements are displayed as fault throw. The core data were corrected to account for the fact that they are from vertical rather than horizontal sample lines. Despite the broad range of fault density, overall the measurements are consistent with a single power-law relationship (dashed line) spanning both core and seismic data, with a slope of  $c.-0.8$  (after Walsh *et al.* 1991).

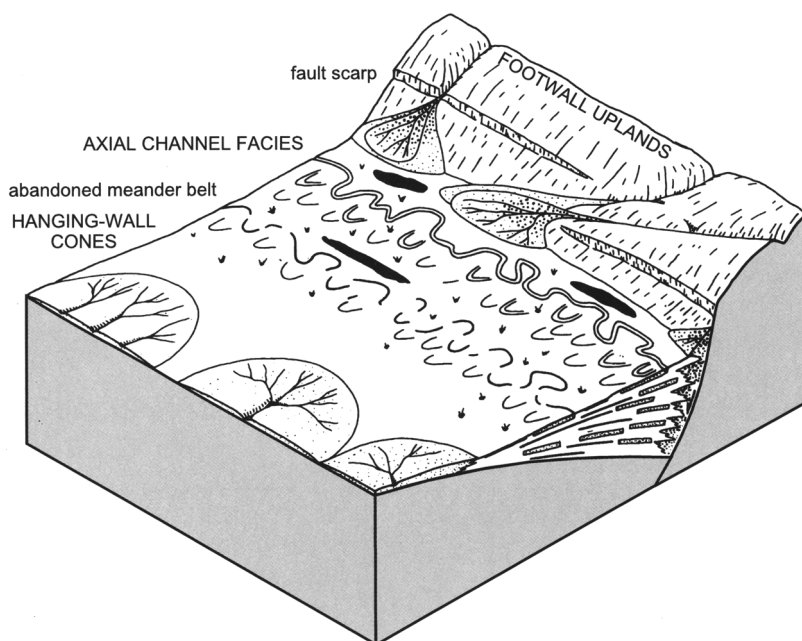
Leeder & Gawthorpe (1987) many geomorphological processes are influenced by gravity, and thus the increase in slope produced as a result of tectonic activity tends to directly influence a variety of sedimentation-related processes (e.g. Alexander & Leeder 1987; Collier *et al.* 1995; Burbank & Pinter 1999) (Fig. 3). The influence of fault deformation on surface processes has recently been confirmed by geodetic measurements which have characterized regional inter-seismic strain fields in many actively deforming areas (e.g. Norabuena *et al.* 1998). These measurements help to provide a more accurate picture of the tectonic forcing function at regional scales which drive long-term landscape development through the combination of tectonic and topographic gradients.

As noted above, faults increase their length with time, since fault displacement and length are positively related (Walsh & Watterson 1988). Fault segmentation, and the resultant interlinkage between various fault segments, however, complicate this relationship. Recent modelling has shown that fault interaction and linkage can lead to temporal variability within an evolving fault array (Cowie 1998). In addition, the segmented nature of normal fault zones suggests that two structural styles can occur contem-

poraneously along any one fault segment – the central parts of normal fault segments are characterized by surface fault breaks while growth folding dominates the ends of fault segments where the fault is blind (Gawthorpe *et al.* 1997).

Normal faults control the creation of accommodation space for syntectonic deposition in rift basins (Schlische 1991; Gawthorpe *et al.* 1994). Thus, the displacement history of a series of linked faults would be recorded within the synrift stratigraphy. However, because the spatial extent of the fault interaction is determined by the scale of the fault segments, synrift sequences will vary spatially along fault systems (Dawers & Underhill 2000). Thus, high displacement rates near segment centres may promote rift climax stratal patterns (cf. Prosser 1993) and facies associations, whereas shallow marine conditions may persist at fault tips and in overlap zones between unlinked faults (Dawers & Underhill 2000).

The overall effect of fault displacement on sedimentation and related processes (e.g. erosion, sediment transport) is a cumulative one, and one made more complex by the segmented nature of fault activity within fault zones. Thus, while there will be a close relationship between the history of



**Fig. 3.** Block diagram summarizing the major clastic environments present in a continental half graben system with through flowing axial drainage (after Leeder & Gawthorpe, 1987).

fault activity and the lithostratigraphic signal of the basin infill, the precise history is not always easy to determine.

### *Rift basins*

As noted by Ingersoll & Busby (1995), any model of continental rifting must consider the various ways in which the lithosphere behaves, including, for example, rheological differences within the lithosphere, contrasts in the composition and structure between the crust and the mantle, the differences between oceanic and continental crust, pre-existing heterogeneities, and the period of time over which strain operates. Two basic models have been proposed for the development of rift basins – the pure shear model of McKenzie (1978) which involves the development of a symmetrical rift structure flanked by major boundary faults (with associated antithetic and synthetic faults) and that of Wernicke (1981) which results in the development of an asymmetrical basin, associated with a deep (listric) fault along which associated antithetic and synthetic structures develop (Fig. 4). However, these two should only be seen as end members, and the variety of actual rift basin forms is much greater. In addition, modifications

of these models exist (e.g. Lister *et al.* 1986) (Fig. 4). Morphologically, rifts may be classified as:

- (1) solitary – e.g. Cambrian Tesoffi Rift, Africa;
- (2) rift stars – e.g. triple junctions, Nakuru junction, Africa, North Sea area;
- (3) rift chains, where several rifts are aligned end to end along linear/arcuate belts of rifting – e.g. East African Rift System, opening of Atlantic Ocean; and
- (4) rift clusters, where several subparallel rifts occur in roughly equant areas – e.g. Basin & Range, Aegean (Sengör 1995).

Active extension or stretching of continental lithosphere leads to surface deformation, volcanism and high heat flow due to the effects of normal faulting and the resultant changes in crustal and mantle thickness, structure and state (Ingersoll & Busby 1995). The tectonic environment of stretching is controlled by regional plate motions. Extension may occur in a variety of geodynamic settings, including continental crust adjacent to young oceans, back-arc basins, continental interiors and thickened crustal orogens (Ingersoll & Busby 1995). As defined by Sengör & Burke (1978) rifting may be passive (i.e.

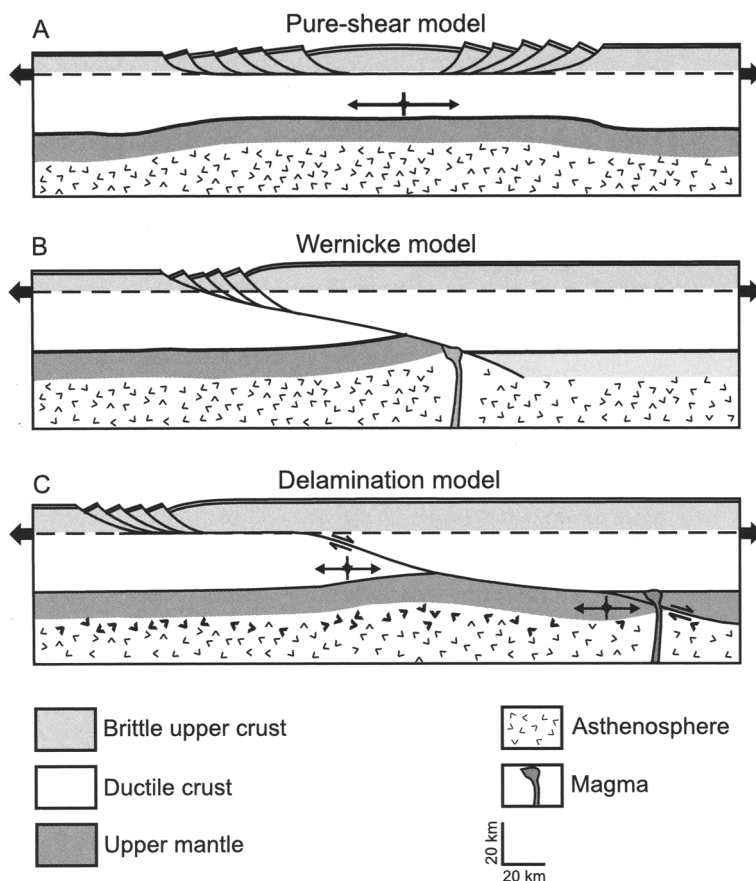


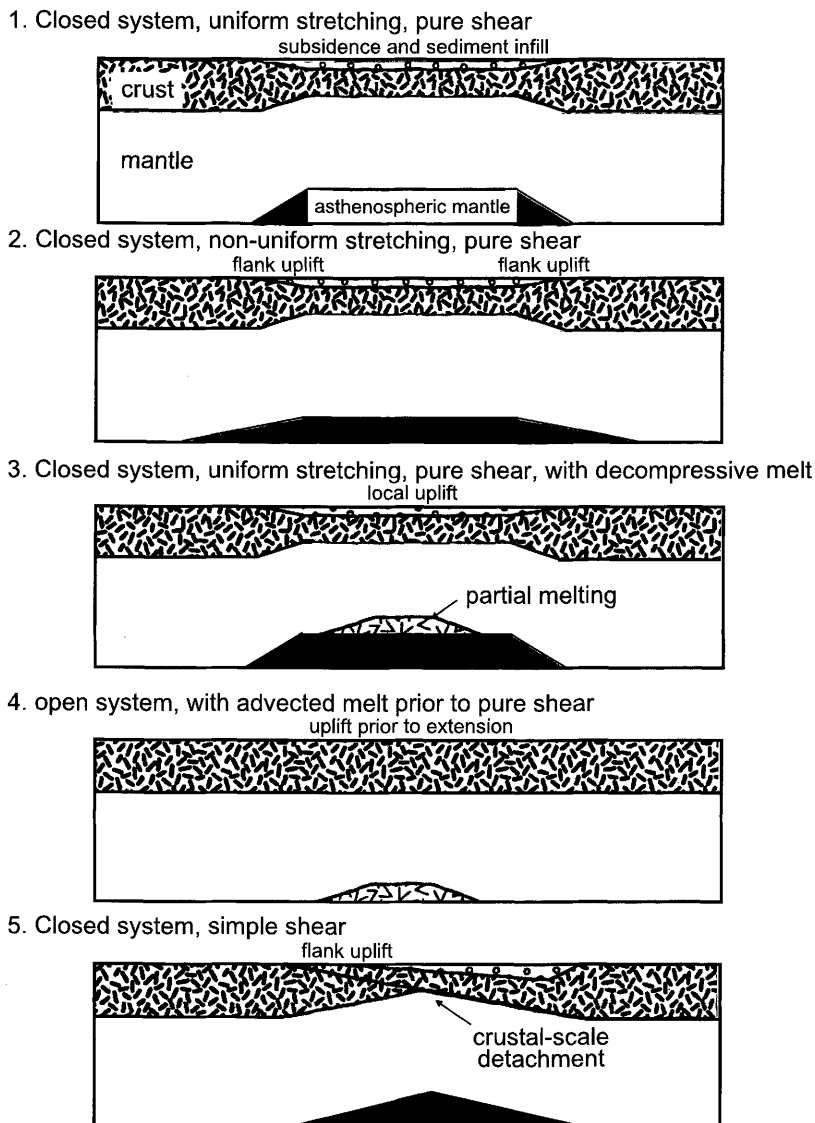
Fig. 4. Three end member models for continental extension (after Lister *et al.* 1986).

closed system, where the input of asthenospheric mass from outside the stretched lithosphere occurs passively as a response to lithospheric thinning) or active (i.e. open system, where rifting is accompanied by the eruption of voluminous volcanics, and the initial rising of the asthenosphere is independent of the magnitude of lithospheric extension) (Fig. 5). However, it is more probable that many rifts evolve under a combination or succession of these two processes (see discussion in Leeder 1995).

The basic structural element of a continental rift is now thought to be a half graben, comprising a single basin-bounding fault. Surface observations in the Basin and Range area have revealed that upper crustal extension is spatially very variable, resulting in local tectonic domains where the upper one-third to one-half of the crust has been removed (Wernicke, 1992). Several structural models have been proposed for half-graben development. These include:

- (1) domino faulting, where high-angle normal faults extend deep into the upper crust with nearly constant dip;
- (2) listric normal faults, which terminate downwards into subhorizontal detachment faults of regional extent and fault blocks are highly rotated; and
- (3) the flexural-rotation (rolling hinge) model, where an initially high-angle normal fault is progressively rotated to lower dips by isostatic uplifting resulting from tectonic denudation (Lucchitta & Suneson 1993) (Fig. 6).

Beneath these areas of extension, however, there is no upwarping of the Moho as would be anticipated if isostatic compensation of the extension occurred within the mantle. Thus, it is possible to find both heterogeneous upper crustal strain and uniform deep crustal structure across extensional domain boundaries resulting from

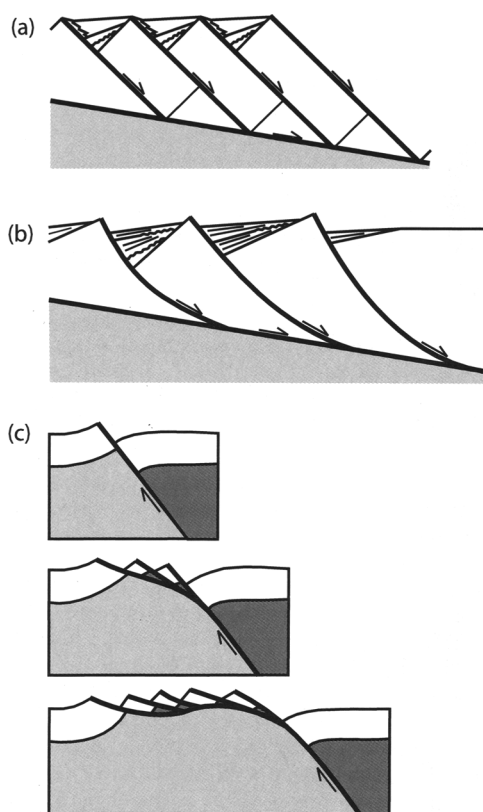


**Fig. 5.** Schematic diagrams to illustrate possible combinations of pure and simple shear, uniform or non-uniform stretching and magma generation. Local (Airy) isostatic compensation assumed throughout (i.e. lithosphere has small elastic thickness). Surface and upper crustal deformation by faulting not shown (after Leeder, 1995).

the effects of intracrustal isostasy (Ingersoll & Busby 1995). Models of rift basin evolution, incorporating a component of lower crustal flow, proceed from a core-complex mode (involving thick crust – *c.* 50 km, and high heat flow) to a wide-rift mode (weaker crust – *c.* 40 km, and high heat flow) to narrow-rift mode (thin crust – *c.* 30 km, and low heat flow) (Fig. 7).

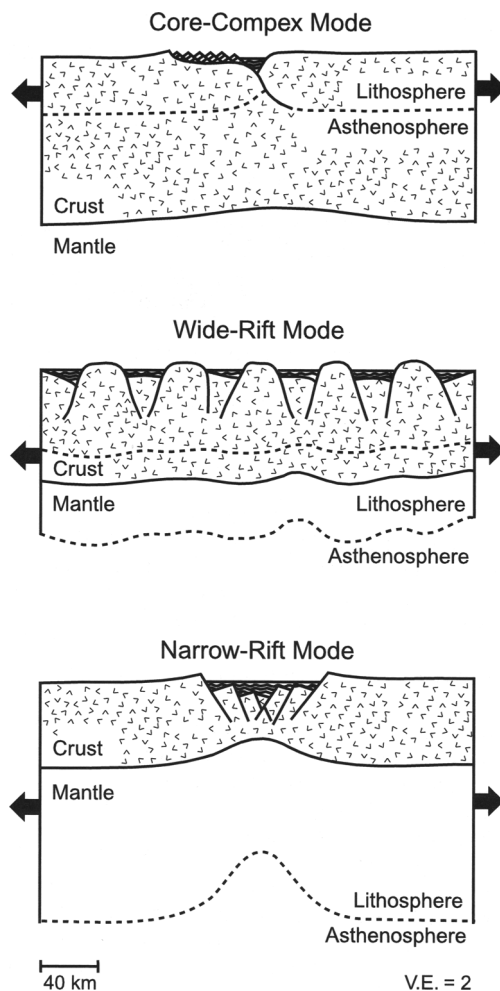
#### *Arc-related basins*

Volcanic arcs are generally arcuate or linear bodies, typically exceeding 1000 km in length and ranging from 50–250 km in width, which parallel subduction-zone trenches (see G. A. Smith & Landis 1995, and references therein for precise terminology of arc complexes). Arc-related basins are found in a convergent geodynamic



**Fig. 6.** Styles of upper-plate faulting. (a) Domino faulting where initial movement occurs on planar, high-angle, normal faults that subsequently rotate to lower angles with continued extension. The faults do not merge within the detachment zone, and the zone of intersection is brecciated and sheared. (b) Llistric faulting where movement occurs on curvilinear faults which flatten with depth and merge with the detachment fault. (c) Rolling-hinge model where an initially high-angle normal fault is progressively rotated to lower dips by isostatic uplift resulting from tectonic and erosional denudation. New high-angle faults are produced when the original faults are too rotated to accommodate extension (after Lucchitta & Suneson 1993).

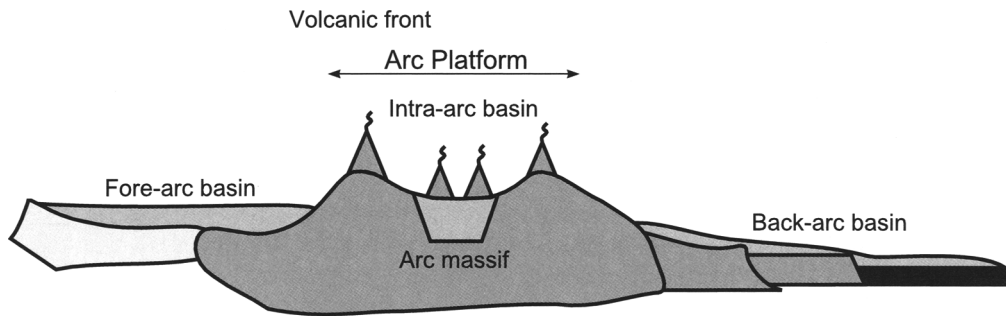
context but are all broadly extensional in terms of their tectonic activity. The three main basin types are related to the location of the volcanic arc, being located on the trench side of the arc (fore-arc), behind the arc (back-arc) or within this structure (intra-arc) (Fig. 8). Fore-arc basins are located between the trench axes, which mark the subduction zone, and the parallel magmatic arc where igneous activity is induced by the inclined descent of oceanic lithosphere (Dickinson 1995).



**Fig. 7.** Cartoon of the lithosphere in three modes of continental extension, emphasizing regions undergoing the greatest amount of continental strain. Lithosphere represents areas with effective viscosities greater than  $10^{21}$  Pa s. Crustal thicknesses vary from top to bottom, 50 km, 40 km and 30 km respectively. Modified after Buck (1991) and Ingersoll & Busby (1995).

Intra-arc basins are defined as basins located within or including the arc platform, which is the typically positive feature formed by the volcanogenic edifices which cap part or all of the arc massif. This latter feature is the region overlain by crust which has been generated by arc magmatic processes (G. A. Smith & Landis 1995). At the time of their formation intra-arc basins are spatially distinct from both fore-arc and back-arc basins (Fig. 8), but they may be just





**Fig. 8.** Diagrammatic cross-section through a convergent plate margin, showing location of arc platform relative to fore-arc and back-arc basins. Areas underlain by arc crust include basement to parts of forearc and backarc basins. Arc volcanoes are typically dispersed over a wide zone perpendicular to plate boundary, but most active volcanoes are aligned along a distinct volcanic front (after G. A. Smith & Landis 1995).

an evolutionary stage for the development of other basin types (e.g. back-arc basins). Back-arc basins occur behind volcanic island arcs and are common along continental margins as well as along convergent plate margins (Marsaglia 1995).

Because of arc migration, however, a single site may change between fore-arc, intra-arc, and back-arc settings, for example, the Luzon Central Valley, which, as a result of changes in subduction polarity and other processes, has successively occupied all three positions over the last 40 Ma (Bachman *et al.* 1983). It may, therefore, be difficult to recognize the precise basin type, i.e. fore-arc, intra-arc or back-arc, based only on their sedimentary and volcanogenic fill (G. A. Smith & Landis 1995). An additional complicating factor is the possibility of intrusion- or collision-accretion-related deformation. Such deformation, and any subsequent uplift and erosion, means that the spatial relationships of volcanogenic materials relative to the arc axis and the distinction – from a geodynamic point of view – of fore-arc, intra-arc and back-arc positions is not always clear (e.g. Lower Palaeozoic Welsh Basin and Lake District). Thus, basins containing volcanogenic material may, more generally, be referred to as arc-related basins.

Volcanic arcs produce large volumes of clastic material that may form much of the arc edifices, in addition to providing a variety of intrusive and extrusive igneous rocks. Intrusive igneous rocks are normally in the form of elongate composite granitoid batholiths. The extrusives include andesitic and dacitic rocks from stratovolcanoes, basalts from intraoceanic arcs and silicic ignimbrites from collapse calderas in continental-margin arcs. Arc settings, therefore, have a

significant component of volcanic debris in the resultant sediments.

*Fore-arc basins.* Forearc basins are extremely variable features, ranging in size from 25 to 125 km wide and between 50 and 500 km long. This variability is a result of the diversity of the controlling factors which govern their genesis (Dickinson 1995). Basins may be simple or compound (i.e. multiple fore-arc basins which lie parallel to one another). These latter features are comprised of strings of interlinked fore-arc depocentres which can extend for 2000–4000 km along modern arc-trench systems. The maximum thickness of sediment fill within a fore-arc basin ranges from 1 to 10 km.

Dickinson & Seely (1979) provided a classification of arc-trench systems, similar to that of Dewey (1980), and outlined plate-tectonic controls governing subduction initiation and forearc development. The factors which control forearc basin geometry include:

- (1) initial setting;
- (2) sediment thickness on the subducting plate;
- (3) the rate of sediment supply to trench;
- (4) the rate of sediment supply to the fore-arc area;
- (5) the rate and orientation of subduction; and
- (6) the time since initiation of subduction.

Subduction environments are extremely variable, although Jarrard (1986a) has recognized a number of distinct zones based on a series of factors, including arc curvature, the geometry of the Wadati-Benioff zone, the strain regime of the overriding plate, the convergence rate, 'absolute' motion (relative to hot spots), slab age, arc age and trench depth. His work demon-

strated that the strain regime within a subduction environment is probably determined by a combination of convergence rate, slab age and slab dip (Ingersoll & Busby 1995).

Fore-arc basins are bounded by volcano-plutonic assemblages with associated metamorphic rocks on the arc margin, and on the trench margin by uplifted subduction complexes composed of varying proportions of deformed and partly metamorphosed oceanic crust, sea-floor sediments, trench fill, and trench slope deposits (Fig. 4). Within the interior of fore-arc basins, either compressional or extensional deformation may occur during forearc sedimentation leading to the development of syndepositional folding, half-graben sub-basins, etc. Extension can be both normal to the subduction direction, or parallel to it (related to variable rates of lateral slippage along the arc–trench gap; McCaffrey 1992), although it is only likely within the fore-arc region within the first 10–20 Ma of initiation of an intraoceanic subduction zone (Stern & Bloomer 1992). Subduction obliquity can lead to strike slip movement (Jarrard 1986*b*). Deformational contrasts lead to corresponding contrasts in the subsidence history of the basin axis and in the uplift history of the trench–slope break, resulting in complex patterns of sediment distribution in both time and space. This complexity means that no single evolutionary model is applicable to all fore-arc basins (Beaudry & Moore 1985).

*Intra-arc basins.* Intra-arc basins are thick volcanic–volcaniclastic–sedimentary accumulations that are found along the arc platform, a region formed of overlapping or superposed volcanoes (Fig. 4). G. A. Smith & Landis (1995) suggest that there are two end member types for intra-arc basins, namely:

- (1) volcano bounded, which have poorly defined margins, thin sediment infill, and are not associated with arc rifting or the formation of oceanic crust (e.g. Larue *et al.* 1991); and
- (2) fault bounded, which are rapidly subsiding, arc parallel or arc-transverse basins caused by tectonically-induced subsidence of segments of the arc platform (e.g. Busby-Spera 1988).

Hybrid basins with characteristics of both types can also be found.

The structural histories of intra-arc basins can vary over time as the arc platforms undergo their complex histories of alternating uplift and subsidence related to angle, obliquity and rate of subduction, which in turn is partly related to the

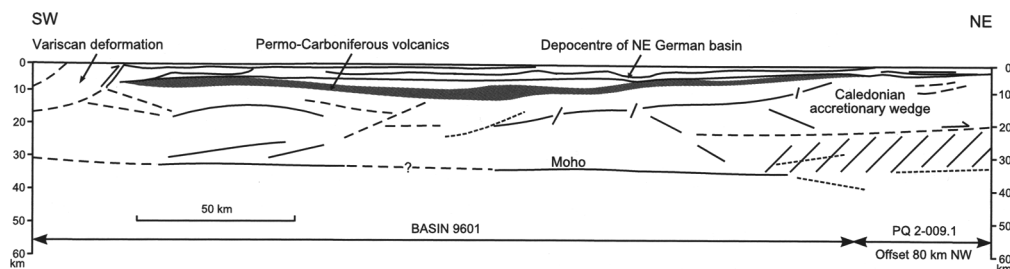
age, thickness and crustal type of the subducting lithosphere (G. A. Smith & Landis 1995). Uplift in the magmatic arc may be associated with crustal thickening and the thermal and physical effects of rising magma. Mechanisms for subsidence, however, are poorly understood, largely hypothetical, and more complex than can be explained by thermal-contraction and flexural-loading models typically applied to other basin types. Six possible mechanisms, acting singly or in combination, may be responsible, including plate boundary forces at the subduction zones, relative plate motions, variations in asthenospheric flow, regional isostasy, magmatic withdrawal and gravitational collapse (G. A. Smith & Landis 1995).

*Back-arc basins.* A back-arc basin is defined by Ingersoll & Busby (1995) as being either an oceanic basin located behind an intraoceanic magmatic arc, or a continental basin situated behind a continental-margin arc that lacks foreland fold–thrust belts. Back-arc basins initiate by crustal extension, firstly producing rifts and then new ocean crust by sea-floor spreading (Karig 1971; Packham & Falvey 1971). Various active and passive methods have been proposed, but no one theory adequately explains the formation of all back-arc basins, with different interpretations being proposed for different geographic regions (e.g. Carey & Sigurdsson 1984). A number of models for back-arc spreading have been proposed. These include extensive magma intrusion, mantle convection or mantle-wedge flow induced by the subducting slab, and thermal upwelling of a mantle diapir (see Marsaglia 1995 and therein for references). Three other types of back-arc basin are also recognized, including:

- (1) non-extensional, which include old ocean basins trapped during plate reorganization which causes a shift of the subduction zone;
- (2) back-arc basins which develop on continental crust and are transitional with retro-arc foreland basins; and
- (3) so-called ‘boundary’ basins, which can be produced by extension along plate boundaries with strike-slip components (Marsaglia, 1995).

#### *Intracratonic rift basins*

Intracratonic basins are saucer-shaped features which are found within continental interiors away from plate margins, and are floored with continental crust and often underlain by failed or fossil rifts (Klein 1995) (Fig. 9). The development



**Fig. 9.** Interpreted line drawing of part of the BASIN 9601 profile and its offshore extension PQ2.009.1, showing the main tectonic and stratigraphic features. Of particular interest is the saucer-shaped profile of the NE German Basin (after DEKORP-BASIN Research Group, 1999).

of an intracratonic basin involves a combination of basin-forming processes, including continental extension, thermal subsidence over a wide area, and later isostatic readjustments. From studies carried out on a number of intracratonic basins it is clear that their formation followed similar patterns. The processes, in order of occurrence, are:

- (1) lithospheric stretching;
- (2) mechanical, fault controlled subsidence;
- (3) thermal subsidence and contraction; and
- (4) merging of slower thermal subsidence with reactivated subsidence due to the isostatically uncompensated excess mass (see Klein 1995 for details).

The precise origin of intracratonic basins, however, is controversial and a variety of different hypotheses have been proposed, including factors which involve an increase in crustal density (due to eclogite phase transformation or thermal modification to the greenschist and amphibolite facies), or magmatic activity (related to igneous intrusions or partial melting and drainage of melt to mid-ocean ridge volcanism) (Klein 1991). Other factors, for example rifting-related hot-spot activity (e.g. Wilson & Lyashkevich 1996), the reactivation of pre-existing structures, far field effects, or changes in intraplate stress, may also occur.

Subsidence analysis studies from North America have shown that the initiation of subsidence of the Illinois, Michigan and Williston basins, and the initiation of subsidence of latest Precambrian and earliest Palaeozoic passive margins were coeval with late Precambrian-age supercontinent break-up (Bond & Kominz 1991). A similar relationship between supercontinental break-up and intracontinental basin formation is also noted from the late Proterozoic of Australia (Lindsay & Korsch 1989) and the

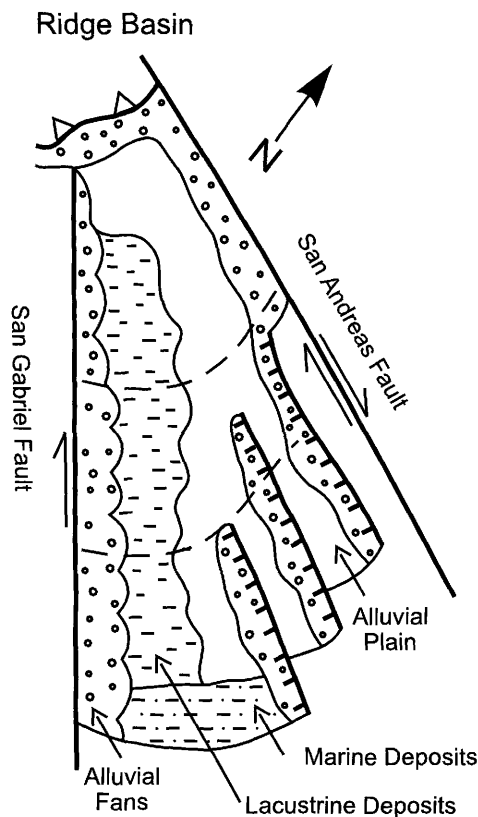
Mesozoic and Cenozoic of Europe and India (Klein & Hsui 1987). Additionally, the sedimentary sequences within intracratonic basins have coeval interregional unconformities and similar trends in thickness and volume (Sloss, 1963; Zalan *et al.* 1990; Klein 1995). The supercontinent break-up model, however, is not accepted by everyone. Some authors suggest that the subsidence histories of the basins are independent, and question the existence of anorogenic granites (which would result in crustal discontinuities) beneath the basins (e.g. Bally 1989). However, supercontinent break-up is not an instantaneous process. Instead, it occurs over a long period of geological time, and this can lead to variations within both basin formation times, and subsidence rates and magnitude across the cratonic area (Klein 1995).

### *Strike-slip basins*

The variability and complexity of sedimentary basins associated with strike-slip faults are almost as great as for all other types of basins (Nilsen & Sylvester 1995, 1999a, b). Christie-Blick & Biddle (1985) provided a comprehensive summary of the structural and stratigraphic development of strike-slip basins, based largely on the work of Crowell (1974a, b) (Fig. 10). The primary controls on structural patterns within strike-slip basins include:

- (1) the degree of convergence and divergence of adjacent blocks;
- (2) the magnitude of displacement;
- (3) the material properties of deformed rocks; and
- (4) the existence of pre-existing structures (Nilsen & Sylvester 1995, 1999a, b).

The formation of strike-slip basins depends largely on the orientation of the principal direc-



**Fig. 10.** Map of the Ridge Basin, California, a strike-slip controlled fault-bend basin, showing the asymmetrical basin morphology, the variable depositional facies, the combination of both axial (predominant) and longitudinal fill patterns, and distribution of sedimentary facies relative to the main basin-bounding faults (after Link 1982; Nilsen & McLaughlin 1985).

tion of extension relative to the direction of bulk shear strain, the overstepping arrangement of discontinuous and discrete fault segments, and on the bending geometry of the fault (Nilsen & Sylvester 1995). Transtensional (including pull-apart) basins form near releasing bends (Crowell 1974*b*), while basins associated with crustal rotations about vertical axes, within the rotating blocks (transrotational basins, after Ingersoll, 1988) may experience any combination of extension, compression and strike-slip.

### Compressional settings

In the Introduction to extensional settings the clear relationship between extensional fault activity and sedimentation/geomorphic processes

was noted. While there are many studies examining the nature of this relationship, there are few which do the same for reverse or thrust faults (see 'Understanding fault activity' below). Instead, displacement on compressional faults tends to be viewed more in terms of the overall geodynamic setting than in terms of its effect on a single fault. However, basins formed in compressional settings will have an abundance of folding and reverse fault activity.

### Foreland basin

Foreland basin systems are complex, large-scale features that develop in response to tectonic loading of a foreland plate by the emplacement of large fold-thrust sheets on their margins (Jordan 1981; Allen *et al.* 1986). The increase in thickness as a result of crustal loading leads to a corresponding isostatic adjustment in the crust, resulting in the formation of a down-flexed moat, which is the foreland basin *sensu stricto*. Subsequent erosion transfers mass from the thrust belt to the basin, resulting in uplift of the orogenic belt and increased subsidence in the basin area. Thus sediment-driven load subsidence amplifies and modifies the tectonic-driven subsidence (Jordan 1995). The stratigraphic record of a foreland basin, therefore, reflects the controlling mechanisms on basin formation, namely, regional subsidence related to flexure of the lithospheric plate on which the basin is located, and secondary controls such as local lithology, climate and eustatic sea level (Jordan *et al.* 1988).

Foreland basins may be broadly subdivided into two types:

- (1) peripheral or collisional foreland basins, which result from arc-arc, arc-continent, or continent-continent collision; and
- (2) retro-arc foreland basins, which form on the continental side of the magmatic arc formed during the subduction of oceanic plates (Dickinson 1976).

Distinguishing between the two types of foreland basin in the ancient record, however, may be difficult, since most orogens undergo several phases of accretion, changes in subduction polarity and changes in the angle of convergence, all of which lead to complications such as strike-slip displacement of the basin and source areas, or even the superposition of basins controlled by different tectonic mechanisms (Miall 1995). Changes in the tectonic style over the course of basin evolution may result in the formation of a hybrid basin that is difficult to classify in terms of its original plate tectonic origin.

### Basin-related magmatism

Magmatic activity within basins and the role and extent of mantle involvement in basin formation has been mentioned in the section on rifting (see above). Magmatic activity, both in terms of intrusion (dykes, sills and plutons), extrusion and withdrawal plays an important role in terms of broad basin dynamics and also in terms of the evolution of the basin infill. Evidence of magmatic activity provides important information on the relationship between heat, magma, pressure and the development of stresses (e.g. using volcanic alignments dyke/sill orientations as kinematics/palaeostress indicators) within basins (Sundvoll *et al.* 1992). Periods of active magmatism during basin formation are probably due to the combined effect of tectonic stress and heat flux. Subsequent magmatism can modify the stress distribution in a basin and lead to non-linear transient rheological heterogeneity in the lithosphere, affecting the lithosphere stress transmission on a regional scale (Ingersoll & Busby 1995).

As previously noted, rifting may be 'active' i.e. where the rifting process necessitates the presence of an upwelling convective plume at the base of the lithosphere prior to crustal extension, or 'passive' as a result of lithospheric extension, and without the need for any magmatic upwelling (cf. Sengör & Burke, 1978). Frostick & Steel (1993) have noted that 'active' and 'passive' rifts should be distinguishable on the basis of their sedimentary history. However, many rifts have features diagnostic of both types (Ingersoll & Busby 1995) since volcanism is present in many rifts. Thus, in order to fully understand the evolutionary history of a region it is necessary to understand the precise chronology of the magmatic, topographical, depositional and structural events.

### Models of sedimentation

The complexity and variability of tectonic settings gives rise to a corresponding complexity and variability within the basin infill of any given tectonic setting. Prediction of the types of sedimentary sequences which might be produced in each of the various basin types, therefore, is difficult. This predictive problem is further complicated by firstly the similarity of some of the basin types (e.g. intra-arc basin, fore-arc basin), and, by extension, the types of sedimentary sequences which will be produced within them; and, secondly the particular post-depositional history of an individual basin (including deformation, diagenesis, strike-slip movement altering original geographical relationships, etc.).

### *Models of sedimentation in an extensional setting (basin scale)*

Extensional basins are formed under tensional stress regimes and their tectonic evolution can be subdivided into the various extensional phases, namely, pre-rift, synrift, and post-rift. The sequential nature of the tectonic activity leads to the production of a correspondingly characteristic sediment sequence which can be related to the different phases of basin formation. The characteristic structural asymmetry of many rift basins exerts a fundamental control on the distribution of sedimentary environments and lithofacies (e.g. Gibbs 1984; Frostick & Reid 1987; Leeder & Alexander 1987; Leeder & Gawthorpe 1987; Alexander & Leeder 1990; Schlische & Olsen 1990; Lambiase 1991). This is particularly true along the basin margins, where transverse drainage systems evolve on the footwall and hanging-wall uplifts, transferring clastic sediments toward the basin centre.

Along the basin-bounding fault, the area of the newly created tectonic uplifts is controlled by the length of the tectonic slope produced during extension (Leeder *et al.* 1991). Coarse-grained cones, or aprons, of sediment are located along the length of these boundary faults. Within the resultant sediment sequences, however, there may be evidence of progradational-retrogradational cycles, the nature of which remains controversial. Some workers believe that clastic wedge progradation occurs during times of minimum tectonic activity along the basin margin, and that fine-grained intervals (lacustrine/shallow marine) correspond with times of high rates of basin subsidence (e.g. Leeder & Gawthorpe 1987; Blair & Bilodeau 1988; Heller & Paola 1992). These models assume constant sediment supply, where progradation results from reduction in accommodation during times of decreased subsidence. In contrast, Surlyk (1990) suggested that sedimentary architecture is controlled by episodicity in footwall-generated sediment discharge into depocentres subjected to continuous deepening.

### *Models of sedimentation in an extensional setting (local scale)*

The sediments which occur in fault-bounded half-graben basins have been widely studied in recent years (e.g. Coward *et al.* 1987). These basins develop progressively during extension, significantly controlling the local geomorphology and sediment transfer mechanisms (Leeder & Gawthorpe 1987; Alexander & Leeder 1990). During the development of an extensional basin,

distinct evolutionary sequences of basin fills may develop. The predominant symmetry of a half-graben basin is asymmetrical, with a steep footwall slope and a shallower hanging-wall slope (Fig. 3). The pattern of sediment distribution within the basin reflects this basic asymmetry, with the thickest sediment sequence being deposited adjacent to the region of maximum fault throw. Coarser sediments tend to be concentrated along the basin margin (e.g. McCann & Shannon 1993) where the decrease of gradient into the basin from the bounding fault causes rapid deposition and the construction of talus cones, alluvial fans, fan deltas and submarine fans (dependent on the prevailing water depths). In contrast, the hanging-wall source area has a broader, gentler slope and the sediments deposited in this region show a wider distribution. Basin centre environments are strongly controlled by climatic influences, with lake or playa systems forming according to the level and availability of local fresh water relative to evaporation. In arid closed basins, aeolian sand complexes may form. Where extension occurs at, or close to, sea-level then basin flooding may occur, leading to the formation of a marine gulf setting (Leeder & Gawthorpe 1987).

In arc-related environments the sedimentary input is characterized by the presence of volcanoclastic detritus, which in some cases (particularly that of the intra-arc setting) may be dominant. Furthermore, the complexity of arc-related settings makes it difficult to provide a single sequence which can be produced as a response to tectonic variables. In forearc areas the sediment infill comprises mainly interbedded sandstone and shale, with rarer conglomeratic intervals being restricted to proximal sites near the basin margins and along the sediment transport paths (e.g. submarine or fluviodeltaic channels). While clastic sediments usually predominate, carbonate sedimentation (related to water depth and geographical location) may also occur. Within intra-arc basins, the majority of the sediment is volcanoclastic in origin. These sediments are produced independent of weathering processes, and thus the sediment volumes and dispersal distances are larger than those found in other clastic depositional systems. Non-volcanoclastic sediments may be locally significant. Facies associations within the intra-arc setting, however, are not unique to these basins, and thus, the presence of vent-proximal volcanic rocks and related intrusions within the central facies association is critical to the correct identification of intra-arc basin settings (G. A. Smith & Landis 1995). The main sediment types recognized from back-arc basins are those derived from pelagic

fallout, airborne ash and submarine gravity flows (Klein 1985). The characteristic lithofacies are variable, reflecting the controls on their distribution. Volcanic components, however, are also present and include lava flows, breccias, pyroclastic rocks and reworked volcanoclastic materials.

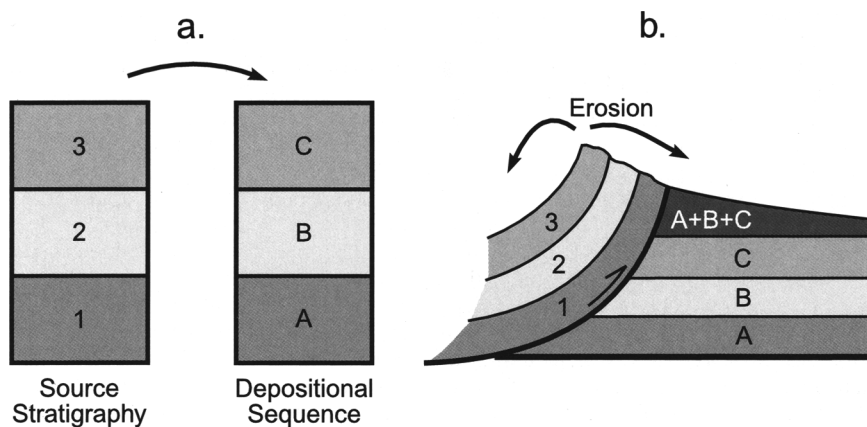
The sedimentary infill of a strike-slip basin may be very complex and variable, depending on whether the basins are submarine, lacustrine, subaerial or a combination, either spatially or temporally. Strike-slip basins tend to be asymmetrical, with diverse depositional environments (with characteristically abrupt facies changes), and an axial pattern of basin fill (Nilsen & McLaughlin 1985). Furthermore, the basin fill is derived from multiple basin margin sources that change over time, which may also mean that the basin sediments are petrologically diverse. In addition, basin fill is characterized by abundant synsedimentary slumping and deformation. Distinctive aspects of sedimentary basins associated with strike-slip faults include:

- (1) mismatches across basin margins;
- (2) longitudinal and lateral basin asymmetry;
- (3) episodic rapid subsidence;
- (4) abrupt lateral facies changes and local unconformities; and
- (5) marked contrasts in stratigraphy, facies geometry, and unconformities among different basins in the same region (Nilsen & Sylvester 1995).

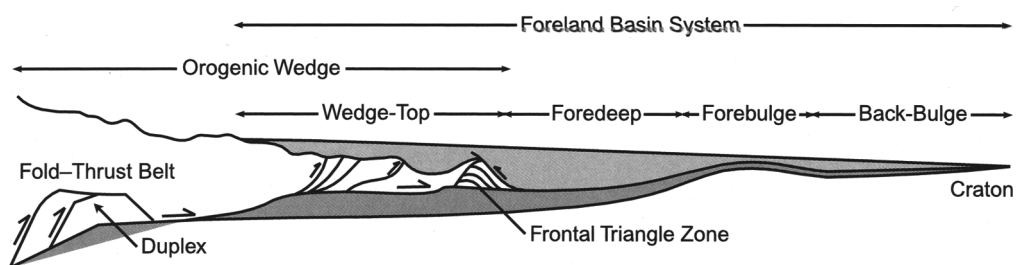
#### *Models of sedimentation in a compressional context (basin scale)*

While models for extensional areas are well developed, this is not the case for regions where compressional activity is predominant. The main models that exist for compressional settings are those that describe the evolution of foreland basin successions (e.g. Beaumont 1981; Jordan 1981). The first evidence of an arc-arc or arc-continent collision in the stratigraphic record may be the transfer of sediments, primarily derived from the fold-thrust belt, into a remnant ocean basin from a point of collision along strike. As the foreland basin develops and fills with sediment, the main trend is that of shallowing and coarsening of the sediment (Fig. 11).

DeCelles & Giles (1996) note that a foreland basin system is an elongate region of potential sediment accommodation (Fig. 12). Within a foreland basin system four discrete depozones, comprising wedge top, foredeep, forebulge and backbulge areas, may be recognized. As a result of the continuing evolution of the belt and the



**Fig. 11.** Composition of basin fill in terms of detrital source petrography: (a) classic inverted stratigraphy, (b) blended composition (after Steidtmann & Schmitt 1988).



**Fig. 12.** Schematic cross-section depicting a revised concept of a foreland basin system, with the wedge-top, foredeep, forebulge and back-bulge depozones shown at approximately true scale. The foreland basin system is shown in coarse stipple, and the diagonally-ruled area indicates pre-existing miogeoclinal strata, which are incorporated into the fold-thrust belt towards the left of the diagram. A schematic duplex is depicted in the hinterland part of the orogenic wedge, and a frontal triangle zone and progressive deformation (short fanning lines associated with thrust tips) in the wedge-top depozone are also shown. Note the substantial overlap between the front of the orogenic wedge and the foreland basin system (after DeCelles & Giles 1996).

basin itself, these zones are not fixed in either space or time and the interaction between them can result in an extremely complex sediment distribution pattern within any foreland basin system. Both subsidence and uplift can cause significant local variations in sediment erosion and deposition, while the relative sense of thrust movement can have significant influence on sediment transport pathways. Most suture zones form by the consumption of an ocean between irregular continental margins that do not match in shape when they collide. The suturing process, therefore, is a diachronous one, such that collision is progressive as the uplift and closure of the remnant ocean basin proceeds. Sediment transport is both axial to the fold-thrust belt and

normal to it (Jordan 1995). Variations in basin geometry and the composition of the stratigraphic fill may thus be interpreted in terms of the global geodynamic evolution.

#### *Models of sedimentation in a compressional context (local scale)*

The evolution of the basin fill in a foreland basin system in terms of sedimentary environment, succession thicknesses and vertical trends, is strongly dependent on the degree of compressional tectonic activity (Muñoz-Jimenez & Casas-Sainz 1997). Generally, foreland basins are initially marine, due to rapid downflexing (Jordan 1981; Flemings & Jordan 1989). At later stages,

sedimentation rates exceed subsidence rates, giving rise to continental sedimentation (Allen *et al.* 1986). Variations within the basin fill may be partially related to variations in the flexural response to loading, differences in the type of crust underlying the basin, the particular type of foreland basin that forms (peripheral or retroarc) or the age of the rifted margin underlying the foreland basin (Miall 1995). These variations will affect the different depozones (see previous section) in different ways, leading to a degree of basin segmentation where the subsidence pattern at any single point is distinctive. Depositional sequences, therefore, can show a high degree of lateral variation in their sedimentary architecture making regional correlation difficult. Such problems are only compounded by the structural complexity that can occur in these settings, for example, in structurally segmented foreland basins, such as thrust-top regions where numerous growth anticlines (related to underlying blind thrusts) are present (de Boer *et al.* 1991; Butler & Grasso 1993; Krystinik & Blakeney DeJarnett 1995). This lack of precise correlation can lead to problems in trying to establish the true relationship between the evolution of the fold-thrust belt and the foreland basin.

### *Sequence stratigraphic models*

Analysis of the sedimentary successions within basins tends to focus on the development of sequences separated by major interregional unconformities, and which record an almost complete transgressive-regressive cycle. In 1963 Sloss recognized a series of broad sequences from the cratonic succession of North America. These Sloss sequences, as they came to be known, were subdivided from each other by major tectonic events. According to Klein (1995) the sequences recognized from the North American Craton are comparable to the classic European geological systems and are unique to intracratonic settings in other regions of the world, including the Russian Platform, Brazil and Africa.

Subsequently, the development of the concept of sequence stratigraphy (e.g. Vail *et al.* 1977) concentrated on the subdivision of units into sequences which could be interpreted in terms of particular genetic parameters (i.e. lowstand, highstand etc.). Sequence stratigraphic concepts were initially developed in eustasy-driven passive margin settings. More recently, there have been attempts to extend this work both into continental settings (e.g. Emery & Myers 1996) and, of particular interest for this work, into tectonically active settings. For rift basins, a number of recent papers have explored the extent

to which it is possible to recognize rift episodes using characteristic sedimentary sequences (e.g. Prosser 1993; Gawthorpe *et al.* 1994; Howell *et al.* 1996; Ravnas & Steel 1998). The sequence stratigraphic models presented in these examples depart from the traditional passive margin sequence in a number of ways, most notably with regard to the spatial variability of sequence architecture (and, by extension, sedimentary architecture). Within rift basins there can be significant variations in subsidence, sediment supply, and physiography adjacent to extensional rift-basin margins, yet the variability in sequence development in three dimensions has only recently been investigated and is poorly understood (e.g. Dart *et al.* 1994; Gawthorpe *et al.* 1997). Typical two dimensional numerical models of tectonics and sedimentation do not account for along-strike variations in structural style and deposition. However, because many tectonic processes are inherently three dimensional, to be truly predictive and applicable, models are required that attempt to address this three dimensional nature (Hardy & Gawthorpe 1998). Such models allow quantification of the variability of stratigraphy and a better understanding of how different controls interact in three dimensions to generate spatially complex stratigraphy.

An additional point is the relative lack of sequence stratigraphic models for other tectonic settings. Some models have been created for foreland basin successions, especially those formed in broad ramp-like foredeep-forebulge type of settings (e.g. Weimer 1960; Lawton 1986; Miall 1991; Cant & Stockmal 1993; Deramond *et al.* 1993; Lopez-Blanco 1993; Plint *et al.* 1993; Posamentier & Allen 1993; Van Wagoner & Bertram 1995). For other tectonic settings, however, there is a lack of studies (see below).

### *Source area*

Much work has been carried out on the provenance of sedimentary rocks in order to differentiate between the various controlling factors, and to constrain the underlying tectonic controls on sediment production (e.g. Dickinson 1970, 1988; Zuffa 1985; Fontana 1991; Morton *et al.* 1991; Graham *et al.* 1993; Garzanti *et al.* 1996; Bahlburg & Floyd 1999). Sediment supply may be strongly asymmetrical (e.g. half-graben and foreland basin systems), and derived from either a few point sources or where these coalesce to approximate a line source. Models predicting sediment distribution within a particular tectonic setting are by necessity simplified versions of complex realities (e.g. Leeder & Gawthorpe 1987;



Steidtmann & Schmitt 1988) (Fig. 3). More precise evaluation of source-area geology can be determined by the analysis of specific minerals (e.g. Götze & Zimmerle 2000), or textures within lithic fragments (e.g. McPhie *et al.* 1993). In addition, specific heavy minerals (e.g. garnet, zircon), or heavy mineral associations (e.g. rutile–zircon, spinel–zircon) may be used to identify specific source rocks (e.g. Morton 1987; Morton & Hallsworth 1994). Geochemical and isotopic whole rock analysis (e.g. Henry *et al.* 1997) or specific chemical elements (e.g. Roser & Korsch 1986) can also be used to determine facts about the geology of the source region and to provide insight into the relative contributions of individual source rocks. More specifically, sediment infill can be analysed to investigate the geological evolution of the source area, for example, the degree of melting of volcanic source rocks (e.g. Najman & Garzanti 2000), pressure–temperature(–time) conditions of metamorphic source rocks (e.g. von Eynatten *et al.* 1996), or the proportion of juvenile to differentiated crustal materials in the source area using Sr and Nd isotopes (e.g. Najman *et al.* 2000).

#### *Influence of climate on sedimentation*

Climate can exercise a very significant control on sedimentation. For example, in foreland basins the climate in which the rising orogen develops is of great importance, both in terms of the tectonic style of the orogen and the architecture of the adjacent foreland basin. Areas of high precipitation (e.g. monsoonal areas) are characterized by rapid erosional unroofing, leading to a corresponding rapid uplift, deep erosion and the development of a foreland basin overfilled with non-marine sediments. In contrast, an arid environment would lead to less erosion, and so erosional unroofing would not compensate uplift, leading to the preservation of the fold–thrust belt and an underfilled foreland basin (Miall 1995). In addition, if a basin is located in a tropical/subtropical area where siliciclastic supply is reduced, then a carbonate template can be superimposed on the distribution of depositional environments within the basin (e.g. Leeder & Gawthorpe 1987; Burchette 1988).

#### *Influence of sea-level change*

Changes in relative sea-level influence the proportions of sediment deposited in a particular basin setting. For example in a foreland basin setting, where sea-level rise coincided with flexure-related subsidence, there would be a corresponding increase in the percentages of

marine sediments relative to non-marine. Indeed, eustatic sea-level is the prime control on whether a retro-arc foreland basin is marine or non-marine, since the thrust-load driven subsidence is not sufficient to submerge normal thickness continental lithosphere during low eustatic sea level. This is in marked contrast to peripheral foreland basins where subsidence commonly places the surface of the underlying plate beneath sea-level (Jordan 1995). Given the same amount of tectonic subsidence, retro-arc foreland basins may be marine (e.g. in the Cretaceous, a period of elevated sea levels), or non-marine (e.g. present day Andes region) (Jordan 1995). Similarly, the location of a rifted basin close to sea level would allow a very different sedimentary succession to evolve (e.g. Leeder & Gawthorpe 1987).

#### **From sediments to tectonics**

From the previous sections it is clear that the very complexity of basin models makes it difficult to fully ascertain the predominant controls on a particular setting. However, it is also very clear that the record contained within the sedimentary infill within a basin is of prime importance in being able to evaluate the tectono-sedimentary evolution of a region. Approaches to the analysis of the sedimentary infill are varied (see previous sections), but all share a common goal – to elucidate our understanding of the shared tectonic and sedimentary history of the basin under investigation. The following section will outline some problems associated with trying to trace tectonic evolution using the sedimentary record, as well as the varied techniques which can be applied, as well as introducing the various studies presented in this volume.

#### *Basin type and preservation potential*

The preservability of tectonostratigraphic assemblages is an important but seldom-discussed factor in basin analysis and palaeotectonic reconstruction. Some modern basin types are common and volumetrically important, whereas others are rare and volumetrically minor. In addition, even some common modern basin types are rarely found in the geological record because they are prone to uplift and erosion, and/or deformation and destruction (e.g. remnant ocean, back-arc basin). Their rarity in ancient orogenic belts is related to their susceptibility to erosion and deformation. Ingersoll & Busby (1995) have illustrated the typical life span of a selection of sedimentary basins versus their post-sedimentation preservation potential

(Fig. 13). It is clear that those basins which have a relatively high post-sedimentation preservation potential (e.g. intracratonic basins, terrestrial rift valleys) have a better chance for evaluation in terms of their tectonosedimentary characteristics than those basins where the sediment fill has a poor preservation potential (e.g. back-arc, transpressional and inverted basins).

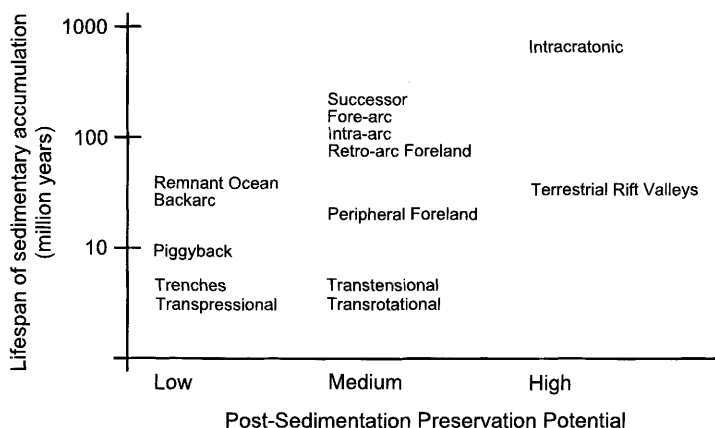
### Mapping

Detailed mapping of an area generally involves a combined approach using a variety of mapping techniques (structural, sedimentological, magmatic) in order to provide a broad picture of the geological evolution of a particular region. Such work also includes the mapping of specific features, for example, the architecture of infill *v.* time, or the spatial distribution of facies *v.* time, can be invaluable for the elucidation of the tectonic evolution of a region. Fernández-Fernández *et al.* (2003) have used a combination of structural and sedimentological mapping to investigate the Middle Jurassic to Cretaceous history of extension in the Betic Cordillera, Spain. The work of McCann *et al.* (2003) uses a similar approach, again using structural analysis and sedimentological investigation but also incorporating detailed mapping of the magmatic units in order to examine the Mid-Devonian-early Lower Carboniferous succession on the southern margin of the Donbas Basin, Ukraine. This work provides insight into the early phases of basin evolution in this complex region and shows the value of a multidisciplinary approach

to such studies. Christophoul *et al.* (2003) mapped a region in the foreland basin of the northern Pyrenees (France) in order to examine the tectono-sedimentary evolution of the thrust belt. Thrust wedge advance and the corresponding loading resulted in basin flexure and sediment infill. Similar work from the Variscan succession of Poland (Lamarche *et al.* 2003) has demonstrated the complexity of orogenic activity in this region. It has also been possible to subdivide the various tectonic episodes into pre-, syn- and post-orogenic phases, thus clarifying the tectonic evolution of this important region.

### Studying facies changing time and space

Tracing the changes in sedimentary facies evolution over time and space within an area can provide detailed information about the subtle ways in which tectonics and sedimentation interact in producing complex facies mosaics. Rieke *et al.* (2003) have used this approach to examine the upper Rotliegend succession from northeastern Germany in order to evaluate the importance of tectonic activity in terms of basin evolution. Previous models had suggested that basin evolution was controlled by a series of tectonic events. Rieke *et al.* (2003), however, clearly demonstrate that basin evolution was largely related to thermal subsidence within the region, although facies development was significantly influenced by climate. On a larger scale, Golonka *et al.* (2003) have examined the entire Polish Carpathian region, providing a



**Figure 13.** Typical life spans for sedimentary basins versus their post-sedimentation preservation potential. This latter term refers to the average amount of time during which basins will not be uplifted and eroded, or be tectonically destroyed during and subsequent to sedimentation. Sedimentary or volcanic fill may be preserved as accretionary complexes during and after basin destruction (modified after Ingersoll & Busby 1995).

series of maps which outline the changing palaeogeography of this region during latest Triassic–earliest Cretaceous times, a period of pronounced tectonic activity.

#### *Grain-dating – exhumation/erosion, source area*

Actualistic petrological models relating sediment composition especially sand and sandstone, to plate tectonic settings have been developed (e.g. Dickinson & Suczek, 1979). Cibin *et al.* (2003) have used petrography to characterize piggy-back basin fill successions and thereby to examine the evolution of thrusting within the northern Apennines, Italy. Augustsson & Bahlburg (2003) use the contrasting geochemical (including Nd and Sm), signatures from the sediment infill within an accretionary wedge sequence to differentiate the signature from the source area and that of the basin itself. Von Eynatten & Wijbrans (2003) have concentrated on a single mineral approach, in this case the Ar/Ar geochronology of detrital white mica, in the evaluation of the exhumation history of the Central Alps.

#### *Sequence analysis*

Sequence analysis is an important tool in exploring the broad evolution of a sedimentary basin. It enables different facies to be correlated and the underlying controls to be determined. Lazauskiene *et al.* (2003) have used this approach in the intracratonic Baltic Basin to investigate the Silurian succession, the period of maximum basin subsidence in the region, and relate basin development to tectonic activity along the Caledonian thrust front. On a smaller scale, Derer *et al.* (2003) have used sequence mapping across the Rhine Graben, Germany, to investigate the interrelationship of between fault activity and sequence formation within the region. Of particular interest is the fact that fault activity led to basin compartmentalization, leading to the evolution of different sedimentary successions on either side of the tectonic divide. Wartenberg *et al.* (2003) have used sequence analysis to investigate the evolution of a fore-arc basin succession within the developing collisional zone of western Australia.

#### *Basin modelling*

Increasingly, basin modelling is used in order to test certain ideas concerning the evolution of a basin. Carrapa *et al.* (2003) have integrated

structural and subsidence analysis data in order to investigate the Oligocene-Miocene history of a basin in northwestern Italy. On a larger scale, Artyushkov & Chekhovich (2003) employed subsidence analysis to investigate the evidence of tectonic subsidence in a region where major eustatic sea-level changes are not recognized. Similarly, Nalpas *et al.* (2003) have analysed the geometries of developing compressional structures, using both mathematical and analogue models, in terms of differing rates of sedimentation. Such work (see below) is of great importance in terms of broadening our knowledge base on compressional tectonic settings.

#### **Problems and future research directions**

It is clear from the previous sections that there are a large number of different tectonic settings and that the sediment sequences contained within them can be extremely variable. While there are particular sequences that are characteristic of particular tectonic settings (e.g. the broad marine-to-non-marine succession produced within peripheral foreland basins), it is not always easy to precisely determine from a particular sediment sequence what the dominant tectonic setting was. Some of these have been outlined above (e.g. influence of climate or sea-level), but there are other factors – broadly related to our lack of understanding of the relationship between sedimentation and tectonics – which are more problematic. In an overview of basin modelling problems, Cloetingh *et al.* (1994) noted that although the success of any individual basin model is often gauged by its ability to reproduce the observed sedimentary record, few models deal realistically with sediment transport and preservation. A lack of understanding of these factors can lead to false or oversimplified interpretations. It is, therefore, clear that there is a great need for additional research, preferably multidisciplinary, in these areas in order to improve interlinked sediment–tectonics models.

#### *Understanding fault activity*

There is now much better understanding of normal faulting (e.g. Roberts *et al.* 1991) and the scaling relationships that operate (e.g. Walsh *et al.* 1991; Walsh & Watterson 1991, 1992; Dawers *et al.* 1993; Dawers & Anders 1995), which provides some basis for the understanding of how faults nucleate, propagate and link together over time. These faults and their displacements are fundamental building blocks for uplift. However, similarly detailed data on the scaling and linkage of reverse and thrust faults do not

exist at present. Studies that fill this gap or define strain partitioning in transtensional and transpressional settings will help with strain quantification in regions of tectonic activity. Such research would greatly aid quantification models of sediment production, for example in basins evolving in compressional settings.

Fault segmentation and the resultant effect on sedimentation patterns is another area which requires investigation. Fault segmentation has been recognized from a variety of settings, including extensional (e.g. Larsen 1988; Peacock & Sanderson 1994; Walsh *et al.* 1999), compressional (e.g. Aydin 1988) and strike-slip settings (e.g. Peacock 1991). However, the precise interaction of the variations in stress generated by either the loss of displacement on individual faults or the transfer of displacement between fault segments, and the effects of these changes in displacement both on sediment basin location and sediment transfer patterns, remain to be studied.

#### *Understanding specific basin types*

Some basins are better understood and researched than others. This is particularly true of rift basins of the graben or half-graben type. However, other basin types require much additional research if we are to be able to really understand even the fundamental aspects of basin evolution in such systems. For example, the processes leading to crustal extension and subsidence in strike-slip settings are generally not as well understood as they are in other tectonic settings (Nilsen & Sylvester 1995). Furthermore, the complexity of strike-slip basins can vary according to their scale. Existing thermo-mechanical models for their formation as well as their structural and stratigraphic evolution are generally poorly developed.

Similarly, existing models for the development of intracratonic basins are largely related to ideas about supercontinent break-up and the resultant changes in heat flow. However, many intracratonic basins do not conform to the predicted subsidence histories. This, coupled with the fact that these basins have not been drilled to basement, leads to much speculation but little clarity.

Within arc-related settings, the situation is even more difficult. G. A. Smith & Landis (1995) note, with some degree of truth, that of all of the basin types considered by most workers involved in basin analysis, intra-arc basins remain the most poorly known. Dickinson (1995), for example, notes that in fore-arc basins little is known about the precise relationship of intra-basinal structures to relevant subduction para-

eters, such as plate convergence rate, the dip of the subducted slab and the motion of the arc massif relative to the roll-back of the subducted slab into the asthenosphere. All of these factors can influence tectonism within fore-arc basins. In addition, syndepositional deformation within fore-arc basins is varied and not well understood. The deformation may be partly related to the basin fill being underthrust by the subduction complex, or associated with backthrusting, both of which processes result in differential subsidence within the area (Dickinson 1995). In intra-arc settings there is little work done by sedimentologists, since the active processes within these basins are predominantly volcanic. In active arcs, young volcanic rocks may obscure older stratigraphic units and structures. Where more information exists (based on seismic evidence), there is a corresponding lack of information of the nature of the sedimentary and volcanic fills. In ancient sequences, the rocks are highly deformed and/or metamorphosed by later tectonic dismemberment or plutonism (G. A. Smith & Landis 1995).

Marsaglia (1995) notes that more detailed studies of the sedimentary facies architecture of backarc basins is lacking, partly because the depositional environments lack two or three-dimensional exposure upon which models could be constructed. There is also a lack of studies on particular sub-environments, particularly that of the volcanic apron, which, according to Carey & Sigurdsson (1984) could be the most diagnostic feature of back-arc basin sedimentation. In summary, the origin of basins within volcano-plutonic (magmatic) arcs is, in general, poorly understood, largely due to the paucity of studies that integrate volcanology, sedimentology and basin analysis (Ingersoll 1988).

#### *Differential tectonic response*

This occurs when parts of the basin are in compression while other parts are in extension. Thus the basin infill provides different tectonic signatures, which need to be compared and contrasted in order to be able to fully ascertain the overall basin history. A corollary of this is the increasingly recognized complexity of normal faults and their movement histories (e.g. Gawthorpe *et al.* 1997). It is extremely probable that such complexity also exists in compressional settings.

#### *Basin compartmentalization*

Basin compartmentalization is where a sedimentary basin is sub-divided by structural or

other barriers and where the various subbasins may produce different tectonosedimentary signatures. Within trench regions, for example, the subduction of interlinked fore-arc basins can lead to buckling of the basin chain, resulting in segmentation and differential subsidence. This relative isolation of the sub-basins has marked consequences for sedimentation patterns (including facies distribution). Similarly, in back-arc or intra-arc basins sediment transport and deposition patterns may be influenced by the locations of volcanic ridges and variable subsidence of rift blocks.

Problems associated with basin compartmentalization can be even more marked when the pattern is overlain by such secondary factors as sea-level variations. A study from northern Spain revealed the presence of a series of unconformities which had a very segmented nature (resulting from the boundary between zones of uplift and zones of subsidence). This pattern of segmentation was related to structural activity that alternated periods of synrotational forced regression (carving of surface below the prograding shoreface) and post-rotational transgression (accumulation of shale wedges prior to the next increment of tilting) (Dreyer *et al.* 1999). In effect, the segmentation of these unconformities demonstrated that there was insufficient time available for the formation of laterally extensive bounding surfaces in the region.

#### *Phase of basin development*

Basin evolution follows a general pattern of tectonic and sedimentary evolution. For example, in rift events we have the production of three clear sequences – the pre-rift, synrift and post-rift successions. Thus, in basin evolutionary models, for each phase of basin evolution (where basins are well understood) a characteristic succession will be produced, and a sediment sequence produced within a syn-rift regime will be very different to one produced in the post-rift thermal subsidence phase.

#### *Sediment budget within a basin*

Hovius & Leeder (1998) and Leeder (1999) note that, more than any other issue in basin research, there is a need to explore the consequences of temporal and spatial changes in water and sediment supply and to intersect time series of these variables with other basin-defining variables such as basin subsidence rate, sea- and lake-level change, catchment uplift rate and climate. Sediment budget or mass-balance methods aim at calculating the volumes of eroded sediment

(Leeder 1991) and can be used, in conjunction with other information (for example, catchment area size), to calculate the average erosion (Einsele *et al.* 1996) or discharge rates (Kuhlemann *et al.* 2001). While there have been a number of studies done in this area (see Burbank & Pinter 1999 for details), there is still a lack of understanding of the controls on sediment budget within a basin. Burbank & Pinter (1999) also noted that there was a need for better numerical models for erosion, and in particular, models which are supported by real data. Additionally, Schlunegger *et al.* (2001) have noted that when dealing with ancient settings, errors on budget methods can be very high, and that the results may be contrasting.

Sediment transport and post-depositional alteration within the basin also have a significant influence on the evolution of large-scale basin architecture through time, because the basin load modifies basin subsidence, and because post-depositional compaction and diagenesis of sediment affects accommodation space available for additional sediment (Schlager 1993).

#### *Differentiating between tectonic and other controls*

This is a very fundamental problem in terms of basin analysis. Sediments are, for the most part, preserved in basins, and the resultant succession records information related both to the depositional mechanisms operating within the basin, and tectonic mechanisms which control basin dynamics and determine the larger scale depositional setting within the basin. The sedimentary record preserved in a basin is thus a product of the interplay of these complex variables. Such factors would include sediment supply, continentality, sea-level variations and climate (e.g. Lindsay & Korsch 1989; Leeder *et al.* 1998; Mack & Leeder 1999). Interpretation of any particular basinal succession, therefore, involves understanding the many different controls on sedimentation. This can be problematic, however, in settings where different controls produce similar effects. In arc-related environments, for example, it can be difficult to distinguish between the interrelationship between tectonic activity and eustatic sea-level change, since tectonic deformation may result in significant changes in relative sea level. In such situations, it is necessary to use as varied an approach to basin analysis as possible in order to rigorously examine the various controls.

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