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Rift basins and supradetachment basins: intracontinental extensional end-members

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ABSTRACT

Two end-members characterize a continuum of continental extensional tectonism: rift settings and highly extended terrains. These different styles result in and are recorded by different extensional basins. Intracontinental rifts (e.g. East Africa, Lake Baikal) usually occur in thermally equilibrated crust of normal thickness. Rift settings commonly display alkali to tholeiitic magmatism, steeply dipping (45–60°) bounding faults, slip rates <1 mm yr^{-1} and low-magnitude extension (10–25%). Total extension typically requires $>$ 25 Myr. The fault and sub-basin geometry which dominates depositional style is a half-graben bounded by a steeply dipping normal fault. Associated basins are deep (6–10 km), and sedimentation is predominantly axial- or hangingwall-derived. Asymmetric subsidence localizes depocentres along the active basin-bounding scarp.

Highly extended continental terrains (e.g. Colorado River extensional corridor, the Cyclade Islands) represent a different tectonic end-member. They form in back-arc regions where the crust has undergone dramatic thickening before extension, and usually reactivate recently deformed crust. Volcanism is typically calc-alkalic, and 80–90% of total extension requires much less time (<10 Myr). Bounding faults are commonly active at shallow dips (15–35°); slip rates (commonly >2 mm yr^{-1}) and bulk extension (often $>$ 100%) are high.

The differences in extension magnitude and rate, volcanism, heat flow, and structural style suggest basin evolution will differ with tectonic setting. Supradetachment basins, or basins formed in highly extended terrains, have predominantly long, transverse drainage networks derived from the breakaway footwall. Depocentres are distal (10–20 km) to the main bounding fault. Basin fill is relatively thin (typically 1–3 km), probably due to rapid uplift of the tectonically and erosionally denuded footwall. Sedimentation rates are high ($\sim$1 m kyr^{-1}) and interrupted by substantial unconformities. In arid and semi-arid regions, fluvial systems are poorly developed and alluvial fans dominated by mass-wasting (debris-flow, rock-avalanche breccias, glide blocks) represent a significant proportion (30–50%) of basin fill. The key parameters for comparing supradetachment to rift systems are extension rate and amount, which are functions of other factors like crustal thickness, thermal state of the lithosphere and tectonic environment. Changes in these parameters over time appear to result in changes to basin systematics.

INTRODUCTION

The last 10 years have produced a revolution in the understanding of extensional tectonics. This is due chiefly to the recognition of highly extended terrains and development of conceptual frameworks which explain increasingly abundant observations (e.g. Wernicke & Burchfiel, 1982; Buck *et al.*, 1988; Davis & Lister, 1988; Spencer & Reynolds, 1989; Holm *et al.*, 1994). Simultaneously, the past decade has produced a revolution in approaches to basin analysis. This results from the development and refinement of new analytical techniques (e.g. backstripping, multi-channel seismic data, numerical modelling, isotope stratigraphy, palaeobathymetry), greatly improved geochronology (single-crystal Ar/Ar, magnetic polarity stratigraphy, U–Pb concordia), and new conceptual frameworks for basin evolution (e.g. Royden & Keen, 1980; Leeder & Gawthorpe, 1987; Wilgus *et al.*, 1988; Flemings & Jordan, 1989).

Despite these advances, the controls on extensional basin fill history and evolution are still poorly understood. There is debate concerning which processes exert strong influence and under what circumstances, and the processes themselves are poorly understood in many cases. There is also a paucity of calibrated models for sedimentation in extensional basins. Several proposed models (e.g. Leeder & Gawthorpe, 1987; Schlische, 1991) explain the first-order, two-dimensional stratigraphy of simple half-grabens for intracontinental rift basins rather well. However, these models have been widely and sometimes inappropriately applied to extensional settings where they may not reflect the wide range of tectonic and depositional styles.

This paper compares rift basins and basins in highly extended terrains, or supradetachment basins (Friedmann & Burbank, 1992; Friedmann *et al.*, 1993). It also presents a new geometric model (like Leeder & Gawthorpe, 1987) based chiefly on field observations and, as such, serves to explain a variety of geological, geomorphical and geodynamic data. Perhaps more importantly, the supradetachment model represents an end-member in a spectrum of extensional tectonics which reflects the complex nature of competing processes and feedbacks. We suggest the model can be most useful and best tested as a template for observation and comparison with other intracontinental basins.

DEFINITIONS

A *supradetachment* basin is a basin which forms above a low-angle normal fault system. As such, it is a type of 'piggy-back' basin forming above its primary bounding fault. The term also represents the end-member model presented here. Although rift basins, or traditional rift basins, have been interpreted by many authors to form above normal faults which root into detachments at depth, these faults are usually of moderate- to high-angle in the upper crust (Jackson *et al.*, 1988). The term *backshed*, or extensional hinterland, means that portion of the footwall which shed water and detritus into the basin. It is synonymous with footwall catchment.

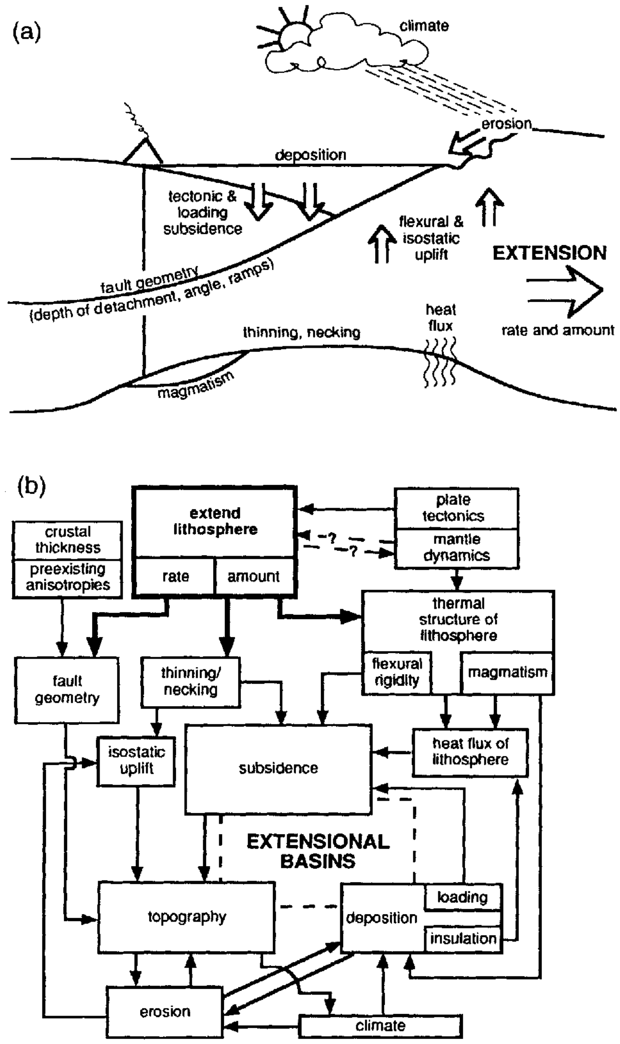

Fig. 1. (a) Schematic extensional basin, illustrating many key features and processes developed during extension and deposition. (b) Flow diagram illustrating the non-negligible factors which determine basin history. Each box represents a complicated set of processes which act in concert towards the evolution of the basin. Thick borders and arrows represent the key driving function and its direct outputs. Note the complex feedbacks between boxes.

MAJOR PROCESSES

Extensional tectonism results from the complex interplay and feedback of numerous tectonic processes and factors (Table 1, Fig. 1). Many features which are recognized as significant factors in highly extended terrains, e.g. magmatism, lower crustal flow or flexural uplift, are

Table 1. Processes of extensional settings.

First-order inputs	First-order effects	Second-order effects
Thermal state of lithosphere	Geometry of basin	Topographic evolution
Crustal thickness	Uplift/denudation	Type of fill
Mantle dynamics	Deformation history	Facies type
Crustal anisotropy	Subsidence rate	Facies distribution/stacking
Stretching/thinning	Subsidence magnitude	Magma chemistry
Amount of extension	Crustal geotherm	Absolute altitude
Rate of extension	Flexural rigidity	Micro-climate effects
Plate tectonic setting	Crustal rheology	
Magmatism(?)	Heat flow	
	Fault geometry	

poorly understood in terms of scale of occurrence, magnitude and overall significance (compare Keen, 1987; Buck *et al.*, 1988; Armstrong & Ward, 1991; Xiao *et al.*, 1991; Lister & Baldwin, 1993). In a schematic representation of a basin developing during extension (Fig. 1a), the driving force is lithospheric extension, which results in all the other features of the basin (save those related to climate). This extension may be induced by other factors, such as plate boundary forces or mantle thermodynamics.

Few factors in extensional environments are independent, and most factors or processes have a number of complex feedbacks (Fig. 1b). For example, crustal rheology, flexural ridigity and fault geometry depend strongly on anisotropies of earlier tectonism, heat flow through the crust and crustal thickness (Buck, 1988; Braun & Beaumont, 1989; Kusznir & Egan, 1989). These factors change over time as a function of amount and rate of extension. Similarly, the rate of erosional denudation of the footwall plate and sediment supply into the basin in part controls the isostatic uplift of the footwall and the flexural loading of the basin (Leeder, 1991). These two parameters can influence the rate of footwall exposure and type of exposed substrate, which feeds back into sediment supply and erosion (Leeder & Jackson, 1993). Similarly, lithospheric thinning can generate small-scale mantle convection (Buck, 1986) which can further thin the lithosphere. This interdependence of many significant factors greatly complicates the interpretation of most extensional settings (see Kusznir & Egan, 1989; Boutilier & Keen, 1994; Leeder, 1995). As such, the simple distinction between 'active' and 'passive' rift settings (Sengör & Burke, 1978) is not particularly helpful in delineating the importance of competing and evolving processes in a basin.

RIFT SYSTEM

Tectonic overview

The African rift system (e.g. Lakes Tanganyika, Rukwa, Turkana, Malawi), the Gulf of Suez and northern Red Sea, the Mid-continent Rift system, the Newark rift system and Lake Baikal all serve as the examples of traditional rifts (Table 2). Of these rifts, only the Red Sea is a proto-oceanic basin (cf. Leeder, 1995).

Rifts characteristically have slow extension rates and small amounts of net extension (Table 2). Extension is relatively long-lived, usually persisting for $\gg$10 Myr and typically for 30–40 Myr. During this time, only limited extension occurs (total extension <20%, β values ~1.2). Geometrically, these rifts can be described as chains of linked half-grabens, typically with relatively steep (>45°) bounding faults and occasionally significant antithetic faults (Fig. 2). These structures can be concave-upward (listric) faults (e.g. Rosendahl, 1987; Ebinger, 1989) or faults which remain planar to great depths (Jackson, 1987; Jackson *et al.*, 1988). Displacement along these faults is transferred or relayed from one to the next along accommodation zones (e.g. Faulds *et al.*, 1990; Morley *et al.*, 1990). These accommodation zones can either result from lateral propagation of fault strands towards each other or initiate as zones of complex fault overlap. They rarely form as simple strike-slip transfer faults. Either way, accommodation zones along major bounding structures are sites of intrabasinal highs, and as a result have thinner sedimentary covers (Anders & Schlische, 1994). Individual rift segments are commonly long (50–100 km), narrow (20–40 km) and structurally deep (6–16 km undecompacted) (Table 2). Magmatism associated with rifting is typically tholeiitic to alkaline, which suggests melting of mantle sources.

An important characteristic of most rifts is that extension occurred in crust which had no significant tectonism during the preceding 150 Myr (Table 2). This means that lithospheric thickness and geothermal gradient were probably 'normal' before rifting (e.g. 30–40 km crust, 120 km to base of lithosphere, 10–15 K km^{-1}). Although many rifts exploit anisotropies inherited from earlier tectonism (e.g. Swanson, 1986; Grieling *et al.*, 1988; Ring, 1994), orogeny ceased in the region of the rifts long before extension. Thus, rifts are typically found within old plates above cold, normal crust.

Sedimentation

The Newark and Hartford basins (Olsen, 1991), Lakes Tanganyika and Malawi (Ebinger, 1989; Crossley, 1984), the Gulf of Corinth (Leeder *et al.*, 1991) and the Gebel Zeit region of Suez (Moretti & Colletta, 1988; Gawthorpe *et al.*, 1990) provide excellent modern and ancient analogues for basins and depositional systems associated with rifting (Table 2). Many of the depositional and stratigraphic characteristics of these basins were summarized by Leeder & Gawthorpe (1987) and added to by other authors (e.g. Schlische & Olsen, 1990; Cohen, 1991).

The basic structural and depositional unit of a rift basin is a half graben, with asymmetric subsidence and with the depocentre along or close to the main bounding fault. Clastic influx is largely derived either axially or transverse from the hangingwall (Fig. 2). The chief depositional environments are fluvial–deltaic, deposited on the hangingwall basinward of a pivot zone, or lacustrine, chiefly turbidites and mudstones. Arid systems are dominated by traction- and sheetflood-dominated alluvial fans and playa environments. Minor fans and fan deltas are shed from the footwall scarp, with limited runout (less than 4 km). Individual half grabens can accumulate very thick fill, often over 6 or 7 km and as much as 16 km thick (Table 2). Sediment accumulation rates are generally less than 0.5 m kyr^{-1}, although

Table 2. Basins and known basin characteristics.

	'Traditional' rift basins					Supradetachment basins				
	Newark Basin	Lake Tanganyika	Lake Baikal	Mid-Continent Rift	Suez (lat. Geibel Zeit)	Shadow Valley	Artillery Mts	Strymon	Chemehuevi	Pickhandle (pre 18 Ma)
References	a, b	c, d, e	f, g, h	i, j	k, l, m	n, o, p	q, r	s, t	u, v	w, x, y
Bounding fault geometry	Moderate (60–40°)	Steep (~67°)	Steep (~70–50°)	Steep (60–65°)	Moderate (60–35°)	*31 ± 3° initial*	*16° (ave)*	*Shallow*	12–26° initial	<20–27° (alt. 30–41 / 15–19°)
	multiple ?planar	Listric, multiple cuspate	Listric, planar	multiple planar	multiple planar	curviplanar, corrugated	curviplanar, corrugated	curviplanar, corrugated	curviplanar, corrugated	unknown
Total extension (km)	7	<16 (beta = 1.25)	10–25	??	11–18 (beta = 1.2–1.4)	11–26	53–79	>25	40–75	20–40
Extension rate (km Myr^{-1})	~0.2	~1.0–0.6	0.3–0.84	??	0.4–0.7	12–1	5.3–8.5	5.1–2.2	10.1–4.5	8–3.5
Duration of sediment record (Myr)	30	15–25 (in progress)	35–30 (in progress)	15–30	24 (in progress)	<7	7–10	7–12	7–9	5–6
Fill thickness (km)	>7	4–6.8	4–6	3.3–7.5	5	3 km	2–3	2.5–4.0	2–3	0.5–1
Accumulation rates (m kyr^{-1})	0.83 (Stockton)	0.45 (max)	0.2 (max)	0.5 (max)	0.5 (max)	~1.2–0.3 (middle)	0.42	0.6 (max)	0.5 (upper)	0.3 (max)
	0.5–0.1 (Lockatong-Passaic)	0.16 (min)	0.14 (min)	0.11 (min)	0.1 (min)	>2.0 (lower)	0.2	0.18 (min)	0.2 (min)	0.1 (min)
Predominant provenance	Hangingwall	Hangingwall	Hangingwall	??	multiple	footwall	footwall	footwall	hangingwall and footwall	hangingwall and footwall
Predominant transport paths	Axial, hangingwall	Axial, hangingwall	Hangingwall	??	multiple (axial)	transverse, ext. parallel	transverse, ext. par.; min. ax.	transverse, ext. parallel	transverse, ext. parallel	transverse, par. & anti-par.
Sedimentary style	fluvial (meandering)	fluvial (meandering), delta	fluvial (meandering), delta	fans, fluvial (braided, meandering)	fluvial (meandering, braided), delta	mass wasting, fans	mass wasting, fans	mass wasting, fans	mass wasting, fans	mass masting, fans
	deep lake (muds, turbidites)	deep lake (muds, turbidites)	deep lake (muds, turbiditic)	shallow lake, mudflat	reef, deep marine, evaporites	major lake (playa, perennial)	fluvial; playa / perennial lacustrine	minor playa, shallow marine	fluvial (?) sandstone	playa, peren, shallow lake
Associated magmatism	Tholeiitic	Alkalic, tholeiitic	Alkalic, tholeiitic	Alkalic, tholeiitic	*Alkalic, tholeiitic*	calc-alkaline *(minor)*	calc-alkaline	calc-alkaline *(minor)*	calc-alkaline	calc-alkaline, minor alkaline
Tectonic lead time (Myr)	~150	~160	~150	~400	~600	~65	~40	~30	~40	~60

(a) Schlische & Olsen (1990), (b) Olsen (1991), (c) Ebinger (1989), (d) Morley *et al.* (1990), (e) Cohen (1991), (f) Hutchinson *et al.* (1992), (g) Logatchev & Zorin (1987), (h) Kiselev (1987), (i) Chandler *et al.* (1989), (j) Sims *et al.* (1993), (k) Jackson *et al.* (1988), (l) Patton *et al.* (1994), (m) Evans (1988), (n) Friedmann *et al.* (in press), (o) Fowler (1992), (p) Davis *et al.* (1993), (q) Spencer & Reynolds (1991), (r) Yamold (1994), (s) Dinter & Royden (1993), (t) Zagorchev (1992), (u) Miller & John (1988), (v) John & Howard (1994), (w) Dokka (1993), (x) Dokka *et al.* (1991), (y) Glazner *et al.* (1994). ext., extension; par., parallel; min. ax., minor axial.

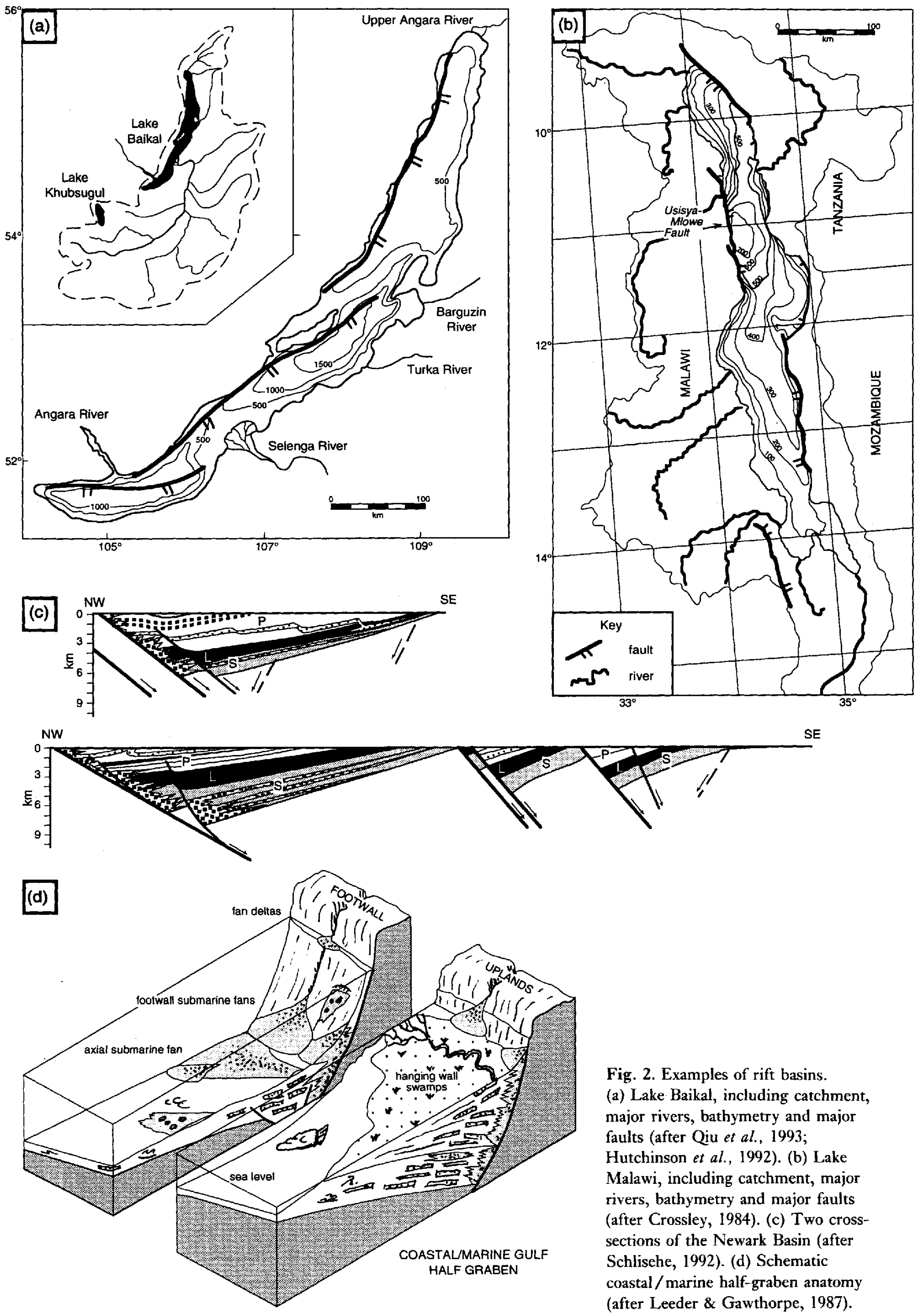

Fig. 2. Examples of rift basins. (a) Lake Baikal, including catchment, major rivers, bathymetry and major faults (after Qiu *et al.*, 1993; Hutchinson *et al.*, 1992). (b) Lake Malawi, including catchment, major rivers, bathymetry and major faults (after Crossley, 1984). (c) Two cross-sections of the Newark Basin (after Schlisehe, 1992). (d) Schematic coastal / marine half-graben anatomy (after Leeder & Gawthorpe, 1987).

they vary as a function of supply, which is in part controlled by both climate and eroded lithotype (Leeder, 1991).

Within an episode of rifting, unconformities tend to be relatively rare and poorly developed (e.g. Morley, 1989; Hutchinson *et al.*, 1992). Although hiatuses occur between members, they are generally short lived and paraconformable (e.g. Olsen, 1991). Minor local unconformities tend to develop in association with block rotation and a shift in the locus of extension (e.g. Moretti & Colletta, 1988), but seldom reflect a more fundamental hiatus. A commonly observed sequence in rift basins is that fluvial environments near the base of the section are replaced by lacustrine fines upsection. Schlische & Olsen (1990) argue that this transition from fluvial to lacustrine deposition does not necessarily reflect a change in rates of tectonism, but can be explained as a consequence of basin widening given a relatively constant clastic influx.

SUPRADETACHMENT BASIN SYSTEMS

Tectonic overview

Supradetachment basins occur in highly extended terrains, such as the United States Cordillera, the Cyclade Islands, the Miocene of the Northern Aegean Sea and the D'Entrecasteaux Islands (Fig. 3). Although some authors suggest that these systems can lead to oceanic rifting (Ethridge *et al.*, 1989), the systems discussed below occur within continental crust and serve as another intracontinental end-member (Table 2).

A detachment's history may be relatively short lived (5–10 Myr) and characterized by high amounts (50–200%) and rapid rates (up to 15 mm yr^{-1}) of extension (see Fig. 5). Although there is considerable controversy about the surface cut-off angles for these fault systems, many workers agree that they have a very shallow average angle and in some cases were demonstrably active very near the surface at shallow (<30°) angles (Table 2). Single detachment strands are commonly ~50 km in length, with displacement relaying onto another detachment or into a transfer structure. In three dimensions, the primary shape of the bounding faults is commonly corrugated (e.g. John, 1987; Davis & Lister, 1988; Spencer & Reynolds, 1991; Dinter & Royden, 1993). Major detachment corrugations have wavelengths of 10–30 km and amplitudes of 1–2 km (Fig. 3). Magmatism associated with detachment faulting is typically calc-alkaline (Table 2; Gans *et al.*, 1989) and contains evidence of crustal contamination and magma mixing.

Many detachment systems are characterized by extremely high slip rates (>5 mm yr^{-1}) along master faults (Foster *et al.*, 1990; Spencer & Reynolds, 1991; Wernicke *et al.*, 1993) and high footwall topography (Hill *et al.*, 1992; Lee & Lister, 1992). Rift systems typically show much slower slip rates (>1 mm yr^{-1}) compared to detachment systems (Table 2).

Detachment systems often occur within crust which is relatively young, or has been tectonically active within only a few tens of millions of years. For example, some Cordilleran core complexes developed within the Sevier thrust belt within only one or two million years of thrusting (Armstrong, 1982), and all developed within 50–70 Myr (Coney & Harms, 1984). Similarly, Aegean detachments often reactivated Eocene structures in the Miocene–Pliocene, typically within 40 Myr (e.g. Schemer *et al.*, 1990). Some authors have argued that recently thickened crust, even overthickened crust, is requisite for the development of highly extended terrains (e.g. Coney, 1987), which may require high heat, strong anisotropy or high gravitational potential (Glazner & Bartley, 1985; Dewey, 1988; Braun & Beaumont, 1989). Thus, the tectonic environment of detachment basins is typically in thick or overthickened, recently active, warm crust in a back-arc setting.

Sedimentation

Supradetachment basins (Fig. 4) are not well represented in the literature, even in terms of stratigraphic descriptions. We use the Shadow Valley Basin, the Chemehuevi Basin and the Artillery Basin of the south-west Cordillera, with the Strymon Basin of Northern Greece, as primary examples which serve to illustrate the supradetachment end-member. Other basins (Pickhandle, Sacramento Mts, GABS and Trobriand basins) with many similar characteristics (Table 1) are included in the discussion.

Although the structural relief on detachment faults can be quite large (10–20 km), depth-to-basement within the associated basins is rarely greater than 3 km, usually between 1 and 2 km (Table 2). This means that most of the crustal isostatic response is manifested in uplift of the denuded footwall (Buck, 1988; Block & Royden, 1990) rather than in subsidence. Davis & Lister (1988) argue that the footwall of metamorphic core complexes must rise far more than the hangingwall subsides. This can often result in impressive footwall topography, which then sheds detritus into the basin (Davies & Warren, 1988). Aerially restricted rock types in the footwall can sometimes be identified as sources for clasts in the basin fill (Miller & John, 1988; Zagorchev, 1990; Dinter & Royden, 1993; Yarnold, 1994; Friedmann *et al.*, in press) and much basinal sediment can be footwall derived (Table 2). This geometry is different from half graben settings, where basin subsidence greatly exceeds footwall uplift (Stein & Barrientos, 1985; Leeder, 1991).

The basin fill in supradetachment basins in semi-arid and arid settings is commonly dominated by coarse conglomerates, usually deposited in large alluvial fans. Debris flow conglomerates and megabreccias are com-

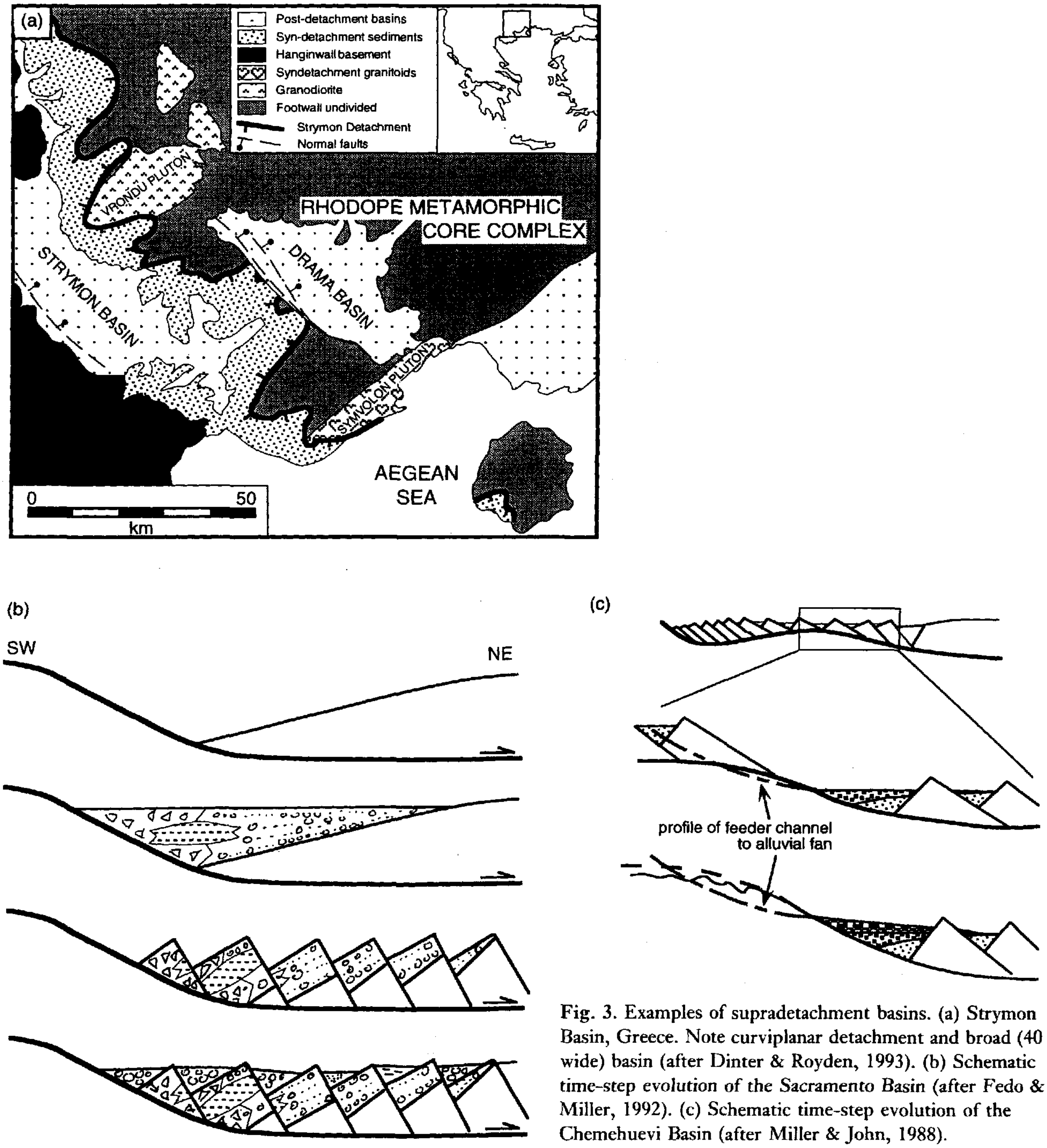

Fig. 3. Examples of supradetachment basins. (a) Strymon Basin, Greece. Note curviplanar detachment and broad (40 km wide) basin (after Dinter & Royden, 1993). (b) Schematic time-step evolution of the Sacramento Basin (after Fedo & Miller, 1992). (c) Schematic time-step evolution of the Chemehuevi Basin (after Miller & John, 1988).

mon features of the basin fill, and gravity-driven glide blocks are present irregularly. Lacustrine systems are usually short lived, and typically evolve as playa environments with evaporites and carbonates in arid conditions. Occasionally, braided stream networks and distal fan environments are preserved. The type of depositional settings preserved are at least in part climatically controlled, as most detachment basins described to date have occurred in arid or semi-arid settings. In many supradetachment basins, the primary drainages flow for great distances in the direction of extension (Figs 3 and 7), i.e. transversely, from the footwall across the hangingwall. Put another way, depocentres occur distal to the bounding fault. Part of the explanation for long run-out and a distal depocentre is that footwall uplift near the range-front faults excludes accommodation space, and sediment is quickly bypassed to a more distal position. In addition to the major transverse footwall drainages, minor, axial drainages would prograde towards the depocentre from other basin margins. Detachment basins contain little detritus shed from the hangingwall towards the bounding scarp (Miller & John, 1988; Zagorchev, 1990; Ingersol *et al.*, 1993; Friedmann *et al.*, in press) although there are notable

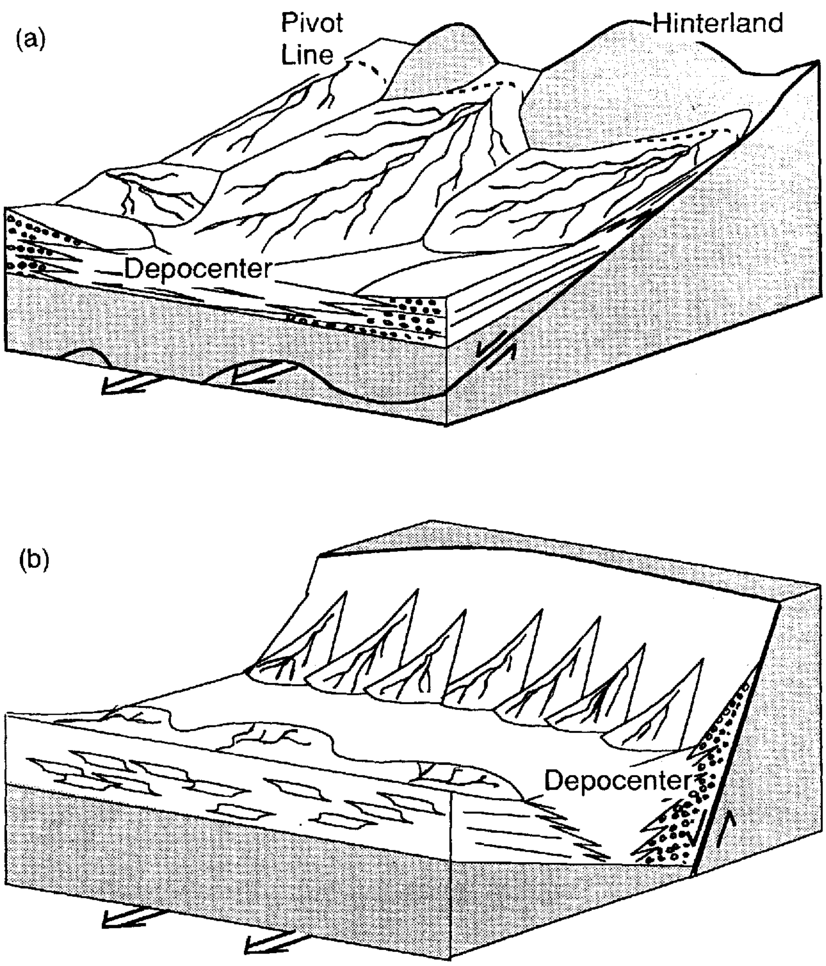

Fig. 4. (a) Closed arid terrestrial supradetachment basin (after Friedmann *et al.*, in press). (b) Closed arid terrestrial rift basin (after Leeder & Gawthorpe, 1987). Note differences in fault geometry, drainage direction and sediment character.

exceptions (Topping, 1993a,b; Fillmore, 1995). It is worth noting that as the detachment system evolves, basin drainage is likely to change (see below; also, Yarnold, 1994).

Substantial angular unconformities are typical in supradetachment basins. These result from tilt-block faulting within the upper plate (e.g. Miller & John, 1988; Fedo & Miller, 1992; Fowler *et al.*, 1995) or other deformation during extension (Friedmann *et al.*, in press). Although relatively short lived (<2 Myr), the change in tilt angles and depositional systems across unconformities can be profound. Tilt-block faulting typically begins a few million years after basin initiation, cutting the pre-existing supradetachment strata with high-angle faults (Fedo & Miller, 1992; Topping, 1993; John & Howard, 1994). Fowler *et al.* (1995) argue that upper-plate block faulting may initiate from stress caused by the isostatic uplift of the lower plate.

The Shadow Valley Basin: an example of the supradetachment end-member

The Shadow Valley Basin is located in the easternmost Mojave Desert (Fig. 6). The major, basin-bounding fault of the Shadow Valley Basin is the Kingston Range/Halloran Hills detachment fault (Burchfiel *et al.*, 1983; Davis *et al.*, 1993), which extends from the northern Kingston Range over 40 km south to the Mescal Range, and possibly further south. Detachment faulting began at approximately 13.4 Ma and probably ceased before 7 Ma (Friedmann *et al.*, in press). A 12.5-Ma pluton, the Kingston Peak pluton, intruded the active detachment fault in the Kingston Range. This event immobilized the detachment in the northern Kingston Range, though extensional faulting and deposition continued south of the pluton in the Shadow Mountains and Halloran Hills (Davis *et al.*, 1993).

The primary map-view geometry of the detachment is curviplanar, with corrugation amplitudes up to 1.5 km and wavelengths of 10–15 km (Fig. 6). Corrugations of the detachment cut Mesozoic faults and fabrics of the Clark Mountain thrust belt, locally at a high angle (Friedmann *et al.*, 1994). The corrugations are primary features and cannot be attributed to later transverse folding, as evidenced by the unfolded, planar geometry of thrusts in the detachment footwall (Davis *et al.*, 1993). Structural studies have constrained the initial dip of the fault system and the amount of extension associated with detachment faulting (Fowler, 1992; Bishop, 1994; Fowler *et al.*, 1995). Translation of the upper plate above the detachment is limited to 5–9 km by regional geological

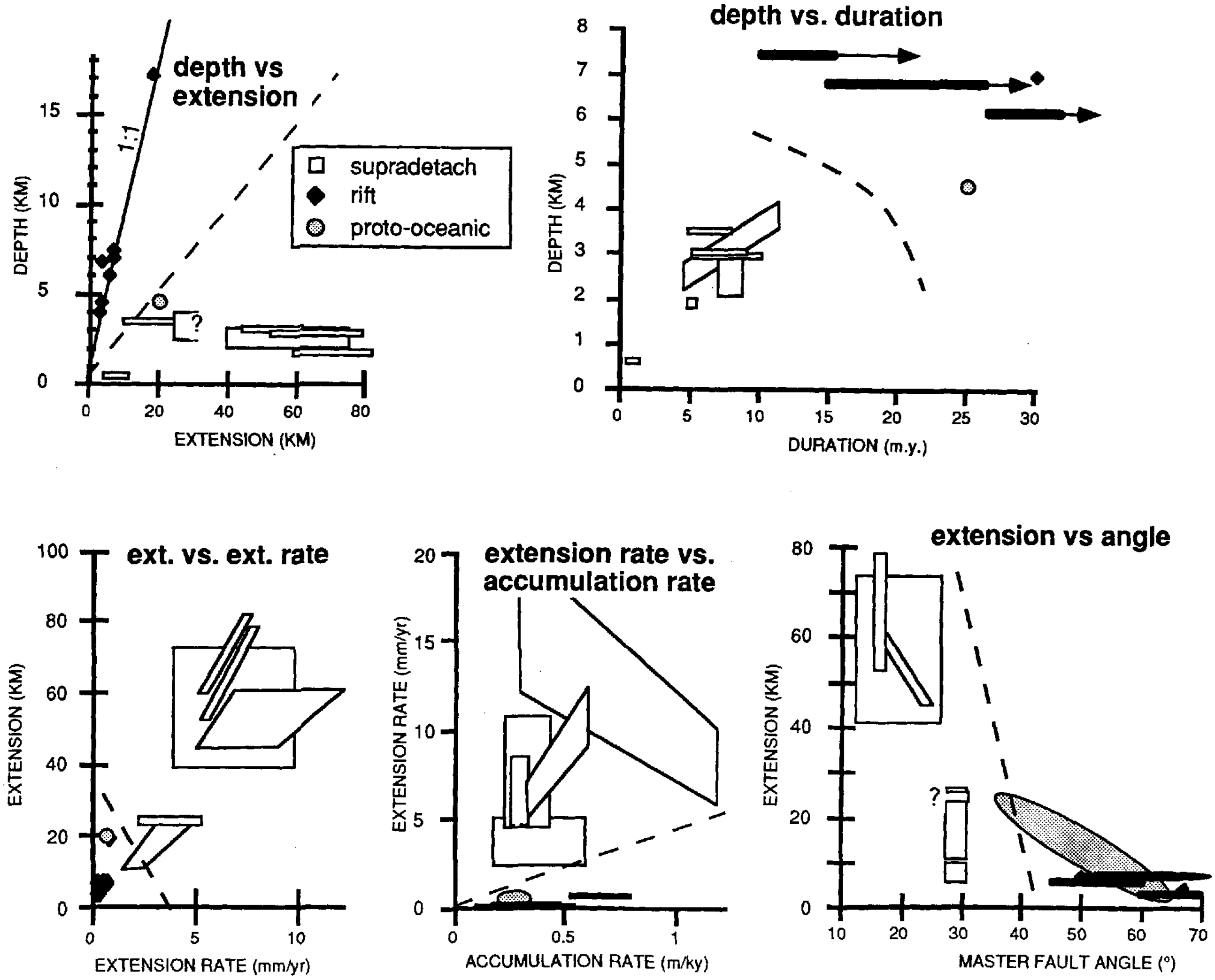

Fig. 5. Five graphs contrasting the end-member characteristics presented in Table 2.

relationships, while extension within the upper plate is probably limited to an additional 3–8 km (10–25%), based on bed dips, and preserved cut-off angles. The initial dip of the detachment at the latitude of the Kingston Range is 30–35°, as constrained by structural relationships there and palaeomagnetic data from the Kingston Peak pluton (Fowler, 1992).

The Shadow Valley basin contains approximately 3 km of middle–late Miocene strata which were deposited during, and as a consequence of, regional extension (Friedmann *et al.*, 1994). At least four unconformity-bounded packages of strata constitute the basin fill. These four members consist predominantly of fanglomerate deposits, mega-breccias, glide blocks, thick perennial and playa lake deposits and volcanic rocks. Most of the strata (members 1–3) were deposited before the initiation of upper-plate tilt-block faulting and thus represent accumulation during the translational phase of extension (Fowler *et al.*, 1995).

New radiometric ages and magnetostratigraphic control allow for the determination of local sediment accumulation rates (Friedmann *et al.*, in press). The rates (Table 2) are maximum and minimum values which consider both error and competing interpretations of data. Undecompacted sediment accumulation rates are of the order of 1.0 mm yr^{-1} (m kyr^{-1}) and range from >5.0 to <0.4 mm yr^{-1}. High accumulation rates are consistent with the predominance of mass-wasting deposits, and are comparable to the few other published rates from supradetachment basins (Woodburne *et al.*, 1990; Holm *et al.*, 1994).

Data from palaeocurrent, provenance and facies-belt analyses in the Shadow Valley Basin strongly suggest that E–W transverse drainage dominated throughout deposition of members 1–3 (Fig. 7). During deposition of members 2–4, a subsidiary drainage centred around Kingston Peak shed detritus across a northern accommodation zone (Friedmann *et al.*, 1994, in press). Most of the sediment which entered the basin, especially detritus within the E–W transverse system, was shed from the footwall of the detachment system. No evidence for a significant hangingwall source exists until member

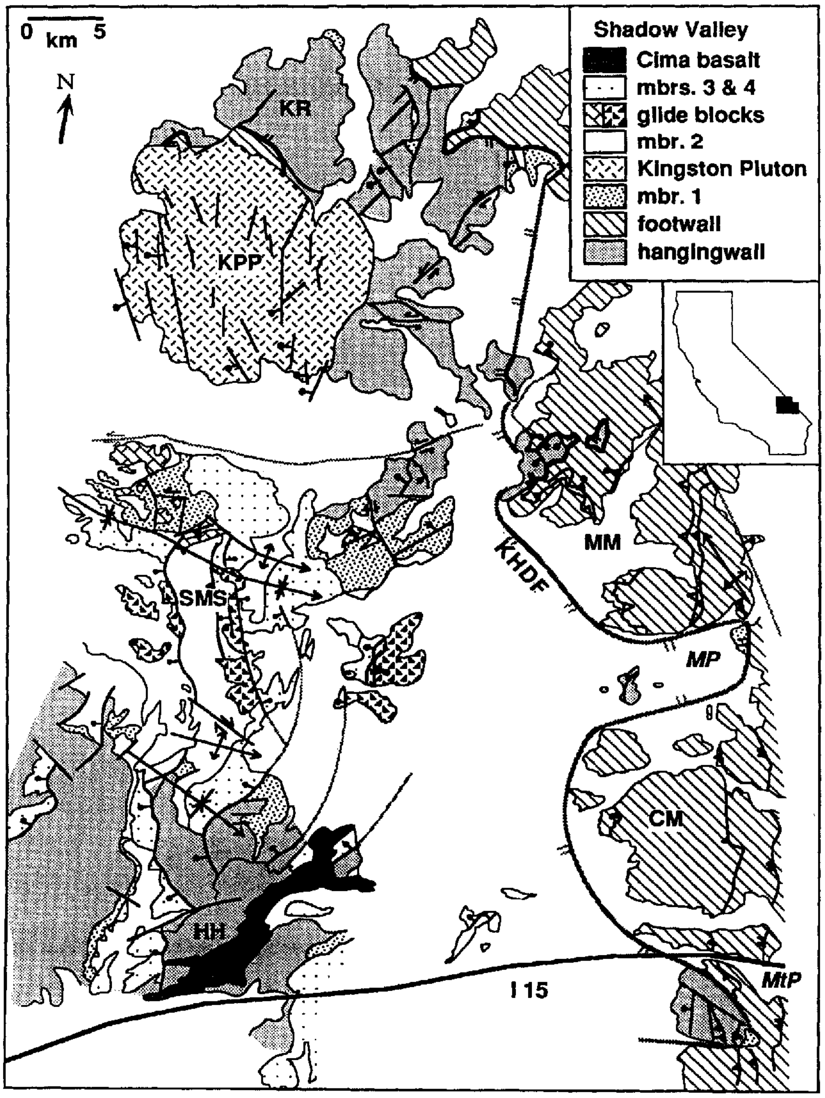

Fig. 6. Map of the Shadow Valley Basin. Kingston/Halloran detachment fault indicated as thick solid or grey line. Note curviplanar detachment and upper-plate normal faults. CM, Clark Mountain; HH, Halloran Hills; I15, Interstate highway 15; KHDF, Kingston Halloran detachment fault; KPP, Kingston Peak pluton; KR, Kingston Range; MM, Mesquite Mountains, *MP*, Mesquite Pass; *MtP*, Mountain Pass; SMS, Shadow Mountains.

4 deposition, during upper-plate faulting. Moreover, facies data suggest that the depocentre lay far from the active range-bounding fault. Upon restoring the extension associated with upper-plate faulting and rigid translation, the depocentre always lay at least 15 km from the active range-front (Friedmann *et al.*, 1994, in press).

The distal position of the Shadow Valley basin depocentre with respect to its eastern breakaway, and long E–W transverse drainage (Figs 7 and 8), are not compatible with models for half-graben depositional systems (Fig. 4). In simple half-grabens, the depocentre forms essentially adjacent to the range front, while long drainages with high sediment yields tend to form within the hangingwall. This is because, in extensional settings that are characterized by steep bounding faults, hangingwall subsidence produced by faulting greatly exceeds associated footwall uplift, usually by a factor of five (Stein & Barrientos, 1985). The accommodation space created by this hangingwall subsidence is rarely completely filled. Moreover, the large volume of sediment shed from the hangingwall tends to displace the depocentre towards the footwall block, which is the lesser sediment source. In contrast, in supradetachment basins and in Shadow Valley in particular, extensive footwall uplift creates a major source area in the backshed and reduces the accommodation space at the range front (Fig. 8). The predominance of footwall-derived detritus, long, transverse drainages, and distal depocentres in supradetachment basins suggest that footwall uplift during extension is a major process in basin evolution.

THE SUPRADETACHMENT BASIN END-MEMBER MODEL

We present here a geometric model for basin evolution in highly extended terrains based on the data discussed above. In this model, a shallow basin (<3 km) develops above and adjacent to an active detachment system (Figs 4 and 8). The detachment may have a corrugated shape which results in a ridge-and-swale topography in the

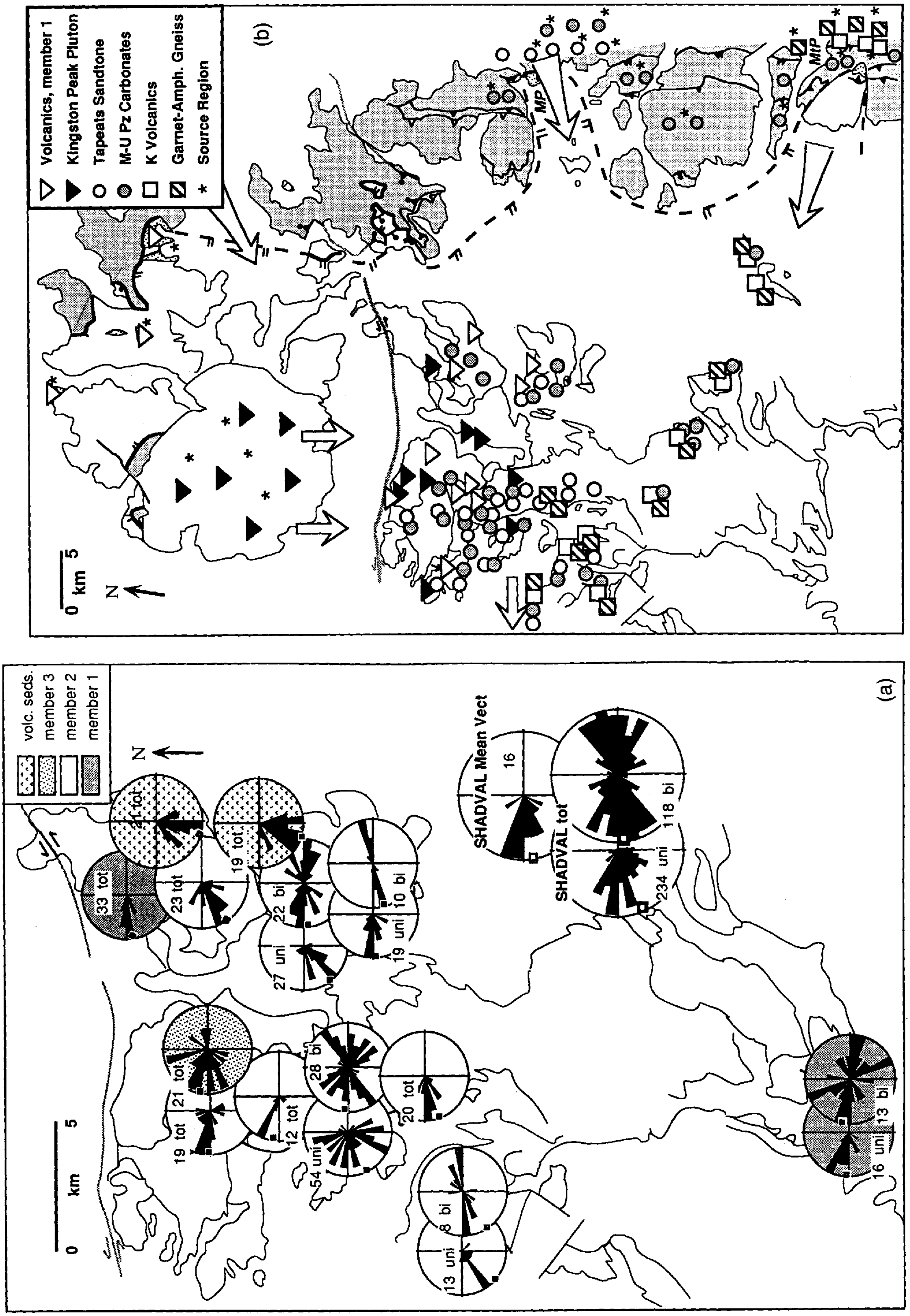

Fig. 7. (a) Palaeocurrent map of the Shadow Valley Basin. Note predominant E–W transport. (b) Provenance map of the Shadow Valley Basin. Note hinterland sources, basin-wide clast occurrence and mixing zones between the three separate main drainages.

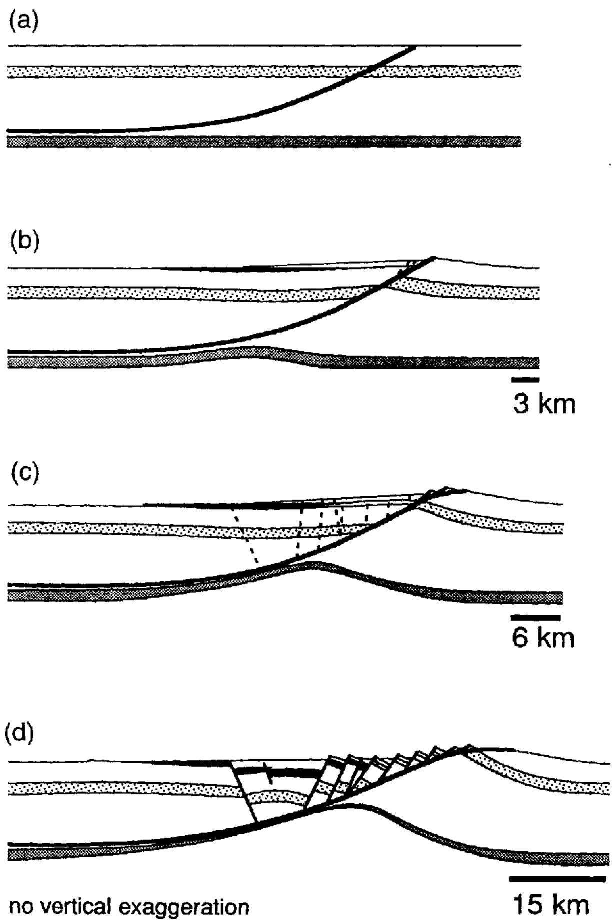

Fig. 8. Schematic time-step evolution of the Shadow Valley Basin. Note translational phase before initiation of upper plate faulting.

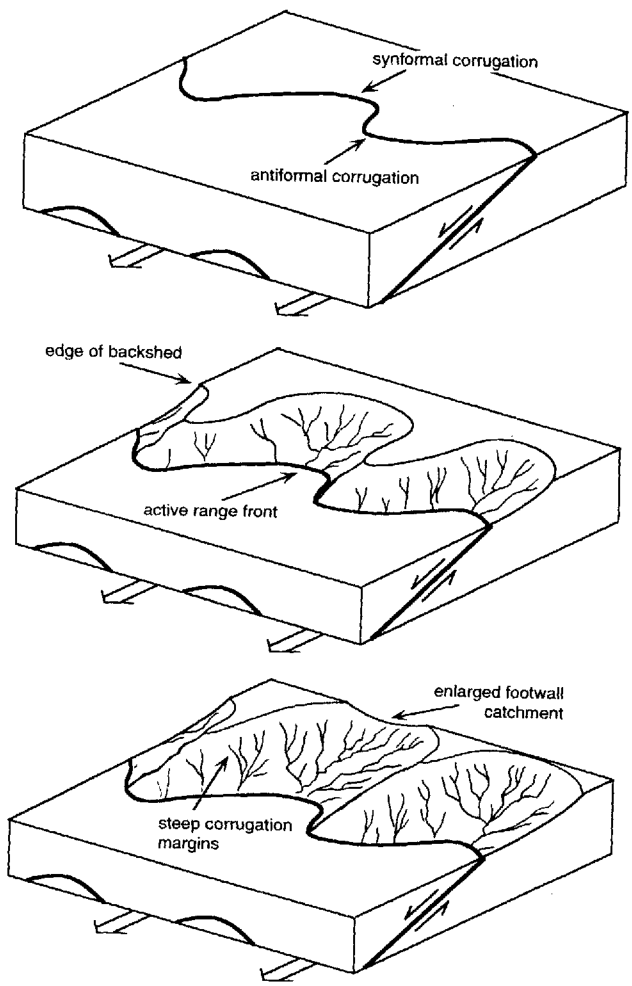

Fig. 9. Schematic block diagram of possible topographic evolution of the detachment hinterland.

backshed. This fundamental basin geometry will remain constant until tilt-block faulting dismembers the upper plate and diverts major drainages.

Footwall uplift, driven chiefly by tectonic denudation, produces several important features. Near the active fault, uplift will maintain a shallow basin by precluding regional subsidence. Little accommodation space and high uplift rates produce long transverse fans sourced from the footwall. These fans prograde towards a distal depocentre, which may in part be localized by upper-plate structures. If the detachment is corrugated, the heads of these fans should lie in corrugation troughs.

Backshed topography will evolve through weathering, knickpoint incision, erosion and mass wasting. The corrugations could act as long-lived sediment transport zones for major drainages (Fig. 9), as is seen in Shadow Valley. This pattern can be seen in bedrock fault scarps, albeit on a much smaller scale. There, corrugation troughs undergo breaching, erosion and preferential surface flow while, in contrast, corrugation crests undergo relatively little modification (Stewart, 1993). Also, brecciation along the fault surface and metre-scale transfer of hangingwall rock into the footwall produces a thick, pre-fractured source terrain (Stewart & Hancock, 1988). Pre-fracturing can produce the detritus needed for debris-flow deposition and is a necessary condition to rock-avalanche deposition (Keefer, 1984; Keefer & Wilson, 1989). Because the steepest slopes would probably shed the greatest detritus, the steep margins of large-scale corrugations could be the chief mass wasting source (Miller & John, 1990; Friedmann *et al.*, 1993, 1994; Yarnold, 1994). Concurrent with these processes, roll-over of the footwall (Wernicke & Axen, 1988) should change the length and shape of drainage catchments.

DISCUSSION

Competing processes

The differences between traditional rifts and basins in highly extended terrains (Fig. 4, Table 2) reflect the predominance of different processes for each end-member. Several factors can influence basin configuration. One is thermal state and structure of the lithosphere before stretching. Highly extended terrains are uniformly associated with subduction magmatism, greatly thickened crust, and thermally and mechanically weakened lithosphere (Glazner & Bartley, 1985; Coney, 1987; Doser, 1987; Dewey, 1988; Davies & Warren, 1988). The combination of high geothermal gradients from magmatism and radiogenic heat flow, gravitational instability and antecedent weaknesses in the crust seem to weaken the crust sufficiently for the development of low-angle decollements that accommodate extension (Sonder *et al.*, 1987; Braun & Beaumont, 1989). Overthickened crust, a hot lithosphere and possibly magmatism would provide sufficient driving buoyancy for the high rates of uplift which are coupled to high rates of extension. In contrast, old, cold crust is strong, and without gravitationally induced deviatoric stress, such crust will tend to fracture at higher angles, producing scarps between 30 and 70° dip (Jackson, 1987).

Rates and amount of extension must also play a determining role in basin configuration and crustal response (Kusznir & Park, 1987; Braun & Beaumont. 1989). These parameters are governed in part by plate tectonic forces, such as slab pull or subduction rollback (Royden, 1993), mantle dynamics and intraplate tension. The space–time distribution of extension, which affects rates within individual basins, is subject as well to local effects (e.g. anisotropies, local heat flow). Whatever the cause, it seems that rates of extension of the order of 4 mm yr^{-1} or more typify supradetachment basins and may be a prerequisite for their formation.

The two-dimensional geometry of the basins (distribution of subsidence) is most strongly controlled by a combination of fault geometry and footwall uplift. The high rates of footwall uplift in detachment settings result in a broad, elevated backshed which sheds detritus across the bounding fault into the basin, little accommodation space near the breakaway, bypassing at the basin margin and deposition in a distally disposed depocentre. In contrast, uplift in rift settings is localized at the rift shoulders due to the geometry of the bounding faults. The depositional environments and facies distribution follow this general geometry (Fig. 4).

High rates of sediment influx also appear to be characteristic of detachment basins. Though speculative, erosional denudation of the footwall may be an important tectonic process in highly extended terrains. Catchment areas are large, sedimentation rates are high (Table 2) and mass-wasting is a significant component of basin fill. Keefer & Wilson (1989) document in modern examples that large earthquakes ($M > 5.5$) typically cause from 10 to 200 mm of mean denudation for areas of thousands of square kilometres per earthquake as a result of landsliding. Transfer of detritus from footwall to the basin may result in isostatic rebound of the backshed area (Molnar & England, 1991; Burbank, 1992). Put another way, if supradetachment basins are filled mostly by footwall detritus, a 2-km-thick basin fill probably resulted in 1–2 km of erosional footwall denudation in only a few million years, and may produce comparable isostatic uplift of the footwall. In contrast, footwall erosion in rift basins is small, especially where resistant rock is exposed (Leeder & Jackson, 1993). Therefore, the component of erosion-induced uplift in rift settings is small over short time-scales.

Death Valley basins

Implicit in its name, the Basin and Range province contains a great number of Cenozoic basins formed during extensional tectonism. Most of these basins are preserved in subsurface strata yet, based on teleseismic, seismic reflection and geomorphic data, the basins have the character of traditional rift basins. Most have moderate to steep (45–70°) bounding faults, large hangingwall sources and small amounts of extension across the major bounding faults (e.g. Leeder & Jackson, 1993). Although as a whole the province is extending rapidly (Minster & Jordan, 1987), individual fault-bounded basins are undergoing relatively slow rates of extension. Many of the modern, rift-type basins occur in areas which had previously been highly extended terrains (Wernicke *et al.*, 1987), and young high-angle faults cut older low-angle detachments. This sequence can be clearly seen in Death Valley (Fig. 10), where a rift-like basin was superposed across a supradetachment-like basin, providing an illustration of how a single extensional system can evolve through time.

Central Death Valley (Fig. 10a) is an excellent natural laboratory to study the relationship between detachment basins and rift basins (Table 3). The modern valley is an asymmetric half-graben, with maximum subsidence located along the bounding scarp. Large, hangingwall-derived fans shed the majority of detritus into a playa lake, with individual fan bodies having run-out lengths of 8–10 km (Hunt & Mabey, 1966). In contrast, the footwall-derived fans are small, even where they are derived from easily eroded Tertiary strata, with run-out lengths up to 2 km and averaging less than 1 km. Although the valley contains up to 3 km of sedimentary fill, probably only 1 km is related to the modern system.

However, Death Valley is a rift basin which developed above an older detachment basin. The late Miocene Greater Amaragosa Chaos-Buckwheat-Sperry Hills (GABS) Basin (Topping, 1993a; Holm *et al.*, 1994) formed above an active detachment whose breakaway was near the Kingston Range during tectonic denudation of the Black Mountains (Stewart, 1983; Holm *et al.*, 1994). The GABS Basin was shallow (1.5–2.0 km) and filled in two phases. Phase one (10.5–7.8 Ma) contains chiefly footwall- and hangingwall-derived fans and minor playa sediments. Footwall-derived sediments formed a substantial quantity of fill, predominantly megabreccias and debris-flow-dominated fans. Hangingwall-derived fans had a run-out length of 7–10 km from the active detachment (Topping, 1993a). During phase two (7.8–4 Ma) the basin was cut by the Amaragosa detachment

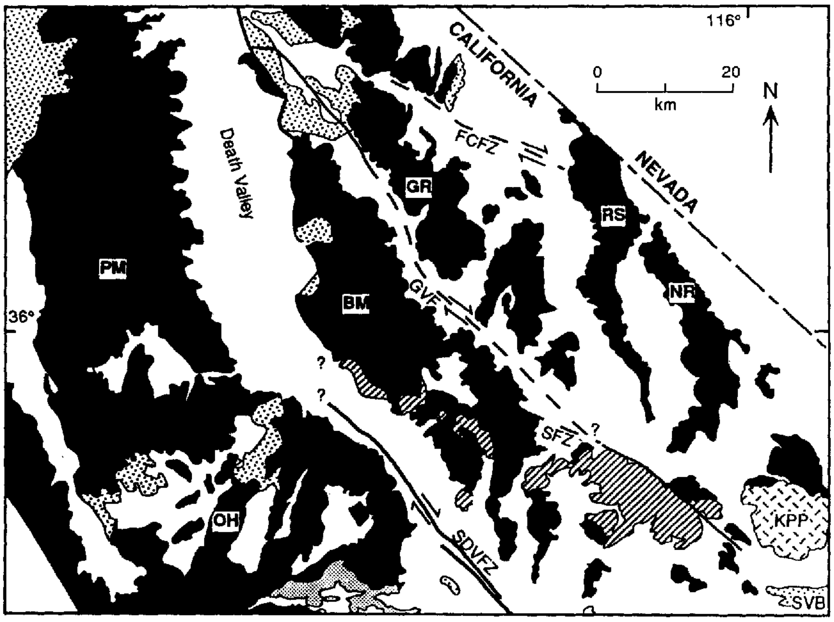

Fig. 10. Map of the Death Valley region, California. Stippled pattern indicates Tertiary sedimentary rock, and the diagonal stripes represent the GABS Basin. BM, Black Mountains; FCFZ, Furnace Creek fault zone; GR, Greenwater Range; KPP, Kingston Peak Pluton; NR, Nopah Range; OH, Owlshead Mountains; PM, Panamint Mountains; RS, Resting Spring Range; SDVFZ, Southern Death Valley fault zone; SFZ, Sheephead Fault Zone; SVB, Shadow Valley Basin. After Serpa (1990) and Topping (1993a).

fault. Most of the 1.5 km of fill, dominated by footwall-derived conglomerates, was deposited during this phase. Extension from 10.5 to 4.0 Ma equals ~55 km (Holm *et al.*, 1994), was most rapid during the earliest phases and migrated westward away from the breakaway over time. After 4 Ma, the detachment and basin were cut by high-angle faults. The past 4 Ma saw 15 km of extension along a system of high-angle faults, one of which is the Modern Death Valley bounding fault.

The GABS Basin was shallow, had long run-out footwall-derived fans, and formed above an active detachment. As such, it has attributes of the supra-detachment end-member. It should be noted, however, that palaeoslopes calculated from traction-dominated fans shed from the hangingwall block argue that the depocentre may have been towards the eastern side of the basin (Topping, 1993b). This suggests that the basin also had rift characteristics, and was perhaps a hybrid basin. As the region evolved, the modern Death Valley system was superposed across the older basin. The modern basin is bounded by high-angle faults, has very large hangingwall-derived fans with almost no footwall contribution, is dominated by playa environments and is currently characterized by slow rates of extension ($<1\ mm\ yr^{-1}$). As such, it is more like a traditional half-graben. We suggest that as the crust returned to

Table 3. Basin characteristics of some superposed extensional systems.

Basin name	Modern Death Valley	Early GABS Basin	Modern Rio Grande (Camp Rice / Palomas)	Early Rio Grande (Lower Santa Fe)
References	a, b	c, d, e	f, g, h	h, i
Bounding fault geometry	moderate (35–60°) curviplanar(??)	unknown unknown	Steep (65–75°) planar (?)	low? (40–0° present) unknown
Total extension (km)	15	25–30 km	10–20 (beta = 1.1–1.2)	>100 (beta = 1.6–3.0)
Extension rate ($km\ Myr^{-1}$)	3.8–0.3	9.2–10.1	<0.2	unknown
Duration of sedimentary record (Myr)	3–4	2.7	9.3–3.0 (4.0)	unknown
Fill thickness (km)	~1	1.5–2	~1 (structural relief = 10)	<2 km
Accumulation rate ($m\ kyr^{-1}$)	1.1	0.5–0.7	0.1–0.3	unknown
Predominant provenance	hangingwall	footwall / hangingwall	axial, hangingwall	footwall
Predominant transport paths	transverse	transverse	axial	transverse
Sedimentary style	traction-dominated fans playa, dunes	mass-wasting, playa debris / traction fans	braided river traction-dominated fans	debris / traction fans playas
Associated magmatism	tholeiitic, alkalic	calc-alkaline	tholeiitic, alkalic	calc-alkaline
Tectonic lead time (Myr)	<5	<5	~10	~15

(a) Hunt & Mabey (1966), (b) Butler *et al.* (1988), (c) Holm *et al.* (1994), (d) Wright *et al.* (1991), (e) Topping (1993a, b), (f) Olsen *et al.* (1987), (g) Mack & Seager (1990), (h) Morgan *et al.* (1986), (i) Seager *et al.* (1984).

more average thickness (30–40 km), rates of local extension slowed due to strengthening of the lithosphere and the supradetachment system was replaced by a rift system. It is worth noting that neither the Amaragosa Chaos Basin nor the modern Death Valley system fully represent either basin end-member (Table 3), probably due to the complex interaction of competing stratigraphic and tectonic processes. The basin dynamics are probably also complicated significantly by the influence of strike-slip tectonism (e.g. Burchfiel & Stewart, 1966; Cemin *et al.*, 1985) which would be likely to influence sediment dispersal and tectonic style.

The Rio Grande rift is another portion of the Cordillera in which a classic rift developed across a highly extended terrain. The modern setting is one of the classic rift settings of the world, displaying steep bounding faults, high heat flow, small (5–15%) extension, alkali–tholeiitic magmatism and small footwall fans supplementing a predominantly axial drainage (Table 1; also Leeder & Gawthorpe, 1987). The modern setting developed between 10 and 3 Ma, with the ancestral Rio Grande developing at approximately 4 Ma (Seager *et al.*, 1984). However, between 30 and 20 Ma, portions of the Rio Grande region underwent 50–200% extension along what are currently low-angle faults, producing broad, shallow basins filled predominantly by footwall detritus (Seager *et al.*, 1984; Morgan *et al.*, 1986). This time interval was characterized by calc-alkaline magmatism, later transitional to alkaline magmatism. Pre-rift palaeogeography is not well known, yet it is clear that the early episode of extension followed within 20 Myr of Laramide thickening. Sediment-accumulation rates and early structural history within the rift are not well constrained, but the characteristics of the tectonism and basin characteristics suggest that deep half-grabens developed across a network of supradetachment basins.

In addition to the Rio Grande rift, other regions in the Cordillera show superposition of rift basins on supradetachment basins (e.g. Safford Basin, Kruger *et al.*, 1995; Barstow Basin, Glazner *et al.*, 1994; Woodburne *et al.*, 1990). In these regions, the period of most rapid extension and tectonic denudation occurred during the earliest phase of basin evolution. This tectonic environment was replaced by slow, less dramatic extension and more 'rift-like' basins. This is also the case in the Northern Aegean, where the Strymon Basin was cut by later high-angle faults of the active Aegean extensional province (Dinter & Royden, 1993).

CONCLUSIONS

The complex, multi-variate and time-dependent nature of continental extensional tectonics makes it difficult to characterize and rank the processes which control basin evolution. Nonetheless, the recognition of tectonic end-members, rifts and highly extended terrains suggests stratigraphic end-members, rifts basins and supradetachment basins. Detailed analysis of the sedimentological and structural evolution of extensional basins can help to characterize basins in the context of the competing processes which produce them. It must be stressed that these two models for extensional basins serve only as end-members for guiding inquiry and discussion.

Two broad associations can be seen. Rifts are generally associated with normal-thickness, cold, old lithosphere, probably because it promotes high-angle faulting and inhibits high strain rates. In contrast, supradetachment basins are associated with thick, hot, young crust, which permits rapid extension and promotes low-angle faulting. Within rift basins, fault-controlled subsidence dominates basin geometry and facies distribution. Within supradetachment systems, footwall uplift, isostatic and flexural, strongly affects basin evolution. Footwall uplift produces long run-out drainages transverse to the main bounding fault, major sediment influx from the footwall, abundant mass-wasting, diminished accommodation space close to the range front and distal depocentres. The geomorphological evolution of detachment footwalls is strongly controlled by shape of the detachment and characteristics of footwall lithology.

Finally, the models presented here represent end-members within a continuum of extensional basin types. From previous investigations, it appears that basins may display characteristics of both end-members as well as evolve from one sort of basin into another over relatively short time-scales. The controls on basin geometry and long-term evolution need to be better understood and quantified, hopefully through increasingly detailed study of and comparison between basins.

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