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CONTENTS

CONTENTS (continued)

CONTENTS (continued)

The nature and origin of structures formed during extensional orogeny

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Introduction

It is remarkable that the nature and origin of structures formed during extensional orogeny should remain controversial for more than one decade, and that one topic should generate so much heated debate.

What is the true nature of extensional orogeny? What drives the extension process? Can extension be associated with widespread regional metamorphism? Do shear zones formed in an extending orogen pass through the entire crust and into the underlying mantle, or do they end in zones of 'bulk pure shear', broadening, or anastomosing until their individual identities are lost? How do metamorphic core complexes form? What is the relation between granite magmatism and the formation of metamorphic core complexes? Do granites rise through the crust, ductilely ballooning the rock above, forming the characteristic domes associated with the metamorphic core complexes? Or do they rise to a point where they form a cushion at the base of a shear zone, acting as a lubricant while the crust above is dragged off sideways? Or is the notion of molten granites in shear zones acting as lubricants nonsensical, requiring metaphysical rates of cooling and magmas many orders of magnitude stiffer than ever measured? Are major normal faults listric? How can some faults be listric and others remain planar to great depths? Do shallow-dipping normal faults of large areal extent really form at low-angles, or are all such 'detachment faults' really rotated high-angle normal faults? What is the nature of the brittle-ductile transition at the termination of great normal faults? What is the role of fluids during the extension process? Do large volumes of fluid flux along detachment faults, and is the role of fluid essential to the formation of these faults? Do detachment fault-related orebodies form because young epithermal systems interact with pre-existing structure, or is the role of the fluid emanating from granites at depth and rising along shear zones essential to the formation of young sedimentary basins? Can we image 'detachment faults' in these basins on seismic records? How does this change our ideas on the evolution of major sedimentary basins? What are the effects of asymmetric extension on the distribution of heat production during the formation of sedimentary basins? How does this influence the maturation history of hydrocarbons? And so the list goes on.
In this lecture, we will step back, and examine some of these issues.

How does the continental crust stretch today, in modern extending orogens?

The upper part of the continental crust appears to extend as the result of the operation of normal faults (Fig. 1). The lower crust extends by ductile stretching. Early models (Fig. 2) envisaged the transition between the region of brittlely-stretching upper crust and the ductilely-stretching lower crust as a fault. Such models sought to explain the 'detachment faults' of the Basin and Range Province, as equivalent to this brittle-ductile transition. In fact, the situation must be considerably more complex.

The strength of rock increases gradually downwards, and may attain a maximum in the region where gradually increasing temperature allows sufficient ductile flow (e.g., as the result of crystalline plasticity) so as to prevent the large deviatoric stresses necessary for seismic failure from being achieved. As the continental crust extends, stresses are transferred to this layer, so that it acts as a stress guide, controlling the behaviour of rock above and below this layer.

Large earthquakes occur when this continental-stress-guide catastrophically fails. The focus of the earthquake is generally at depths between 8-12 km. The seismic rupture propagates upwards, sometimes reaching the surface, as well as propagating downwards, sometimes reaching depths of 15-16 km. Contrary to popular belief, the fault-plane in such major ruptures does not appear to be listric. The dip on the fault-plane may vary less than 10° in this entire depth range.

In the period preceding the earthquake two things must happen in the continental-stress-guide. First, there must be a gradual build-up in the level of deviatoric stress, as the external constraints posed by the tectonic system force the crust to try to continue stretching. Above this diffusely-bounded stress-guide the crust fails by faulting, but the faults are relatively small, and they do not penetrate the entire crust. Below the stress-guide, ductile flow relaxes the accumulating elastic stresses, so that only relatively low stress levels are maintained. Second, as the stress reaches critical levels, dilatancy increases, and fluids are sucked into the region surrounding the fault plane.

At the time of rupture, fluids in cavities associated with the rupture will drop pressure and temperature abruptly. A portion of their dissolved content will be deposited. Ore concentrations may form under the appropriate circumstances. After the seismic rupture has propagated, these fluids are expelled upwards along the fault plane, where they may become involved in shallower-level circulation systems, and other types of ore deposits may form. These fluids are also driven downwards, below the normal level of the brittle-ductile transition, into regions where fluid pressures are generally lithostatic. This is because the seismic rupture propagates downwards from the stress-guide, into regions which under normal circumstances would flow in a ductile fashion. During a seismic event however, strain-rates are high enough to cause the normally ductile rock to behave brittlely. The fault will be associated with dilation, for example around zones of non-planarity, or where jogs in the fault-plane occur. This dilation will suck fluid from higher levels in the crust downwards, into the zone below the stress-guide, where pore-fluids are normally lithostatically pressured.

At these depths, stress levels before the earthquake are relatively low. The propagation of the rupture meant that a stress rise occurs in this region. In the period of post-seismic relaxation, these stresses must be relaxed, in large part by ductile creep. As this occurs, the fluids which have been driven downwards into the lower parts of the fault are sucked into the shear zone, as the result of small but penetrative dilatancy associated with ductile deformation. This fluid will serve to allow retrograde shear zones to develop (such as at Broken Hill). The ductile shear zones formed in this region must be shallow-dipping, to accommodate strain incompatibilities between the brittle upper crust and the ductile lower crust.

The domino mechanism for extension of the continental crust

The above concepts suggest that continental extension should take place involving planar normal faults separating crustal-scale tilt blocks, which rotate as extension proceeds. This gives rise to the domino model for deformation of the upper continental crust. Figure 3a shows large (crustal-scale) tilt blocks. Extension occurs as the result of these tilt-blocks rotating. If extension exceeds about 70 per cent, the tilt blocks must have rotated about 30° and normal faults which were originally dipping at 60° are now to be found at about 30° dip (Fig. 3b). At this stage, a new set of normal faults may initiate, as shown in Figure 3c, and by the time extension has reached 200 per cent (i.e., an additional 70 per cent extension), these faults will have also rotated 30°, so the situation as shown in Figure 3d arises. A new generation of normal faults will eventually be generated, as shown in Figure 3e. By the time extension reaches 300 per cent, an additional 30° of rotation takes place, and once-horizontal strata are vertically dipping, with the second generation high-angle normal faults now horizontal. Such a picture allows representation of many of the aspects of the geology of economically-interesting areas such as at Yerrington, or in the Eldorado Mountains, in Nevada.

Evidence for detachment faults

However, this is not an entirely accurate account of the exact situation. In many of the core complexes, the youngest faults are shallow-dipping normal faults of large areal extent, as illustrated in Figure 4. This additional element to the geology, namely a basal detachment fault, is often of large areal extent. In the Whipple Mountains, California, the fault presently mapped as the Whipple detachment fault appears to be only the latest surface (or, perhaps, a composite of late-generation surfaces) that sliced through the fractured upper plate, accomplishing considerable excisement of the lower portions of the upper plate in the process (Fig. 5). This fault is a primary feature formed during the extensional process. It does not represent a reactivated thrust. It does not represent a former high-angle fault which has been rotated into shallow attitudes by ongoing extension. It propagated to near-surface levels with a dip of between 10^{o}-25^{o}. It cannot represent an ancient brittle-ductile transition in the Earth's crust. Moreover, the detachment fault formed after the mylonites in the lower plate had become kinematically inactive, at a time when the lower plate was deforming brittlely. This complex system of detachments evolved with considerable rapidity, and appears to have accomplished a horizontal translation in excess of 40 km within a probable time-frame of less than two Ma duration. The extreme rapidity of extension is a common feature of the terranes in which detachment faults have been recognised.

The role of volcanism

Opinions are divided concerning the role of volcanism in the extension process. The regions in which the most extension took place (where metamorphic core complexes outcrop) are also areas in which extensive and voluminous volcanism also took place, synchronously with the extension process. Few places on earth today have such intense volcanic activity, although examples do exist (e.g., the Taupo Zone, New Zealand). Some data suggest that voluminous volcanic activity took place prior to the rapid extension process which led to the formation of metamorphic core complexes. However, these data are disputed, and it may be that significant extension had taken place in some cases before extension volcanism took place. There is no resolution of the question as to whether volcanism triggers extension, or vice versa. In any case, large horizontal motions are involved, and earlier estimates of the amount of Basin and Range Province extension, which suggest that extension was limited, are now seen to be

The role of fluids in detachment faulting

In California, and Arizona, large volumes of fluids have moved along these detachment faults. The detachment fault is marked by a microbreccia, which is a zone of ultracataclasite. Detailed studies of such rocks show they have been repeatedly crushed, and then healed again into indurated rock, forming generation after generation of "microbreccias". In one hand specimen from the Whipple Fault, generations of fault movement can be recognised. Clasts of cataclasites exist within clasts of cataclasite which themselves exist within clasts of cataclasite, and so on. These ultracataclasites, when crushed, were extremely mobile, and acted apparently in a superplastic fashion. The healing processes has involved growth of quartz crystallites in a high-permeability matrix, until the growing crystallites impinged upon one another. These crystallites were then overgrown by quartz with a minor amount of clay which imparted their characteristic green and red colours. The "microbreccias" are, in fact, dirty cherts. Presumably silica was dropped-out of fluids as they migrated up the detachment fault.

Below the microbreccia, the lower plate is extensively brecciated, metasomatised, and chloritised. The rock underwent the transition from ductile to brittle at the same time as pore-fluids became involved in the microstructure, suggesting that the ductile-brittle transition was induced by the arrival of pressured pore-fluids.

In the Basin and Range Province, these fluids may have emanated from crystallising granites, intruded during the extension process. Extension is, in any case, accompanied by violent and prolific volcanic activity. I believe these fluids play a fundamental mechanical role in allowing this type of extension process to take place.

Economic significance

Do structures such as those recognised in the Basin and Range Province exist outside of that Province, and what possible significance can this have for the mineral and hydrocarbon industries?

In terms of mineral exploration in Australia and the southwest Pacific, there is no doubt that detachment terranes will be of some interest. I believe already we can target three, if not four, possible detachment terranes on land in Australia and the southwest Pacific. Some of these are already interesting prospects for gold mineralisation. We will investigate further the real significance of detachment fault mineralisation in the Basin and Range Province, with possible implications for these other terranes.

In terms of the hydrocarbon industry, the interest in these structures is possibly less direct. In subsided sedimentary basins on the Australian passive margin, there are many structures typical of asymmetric extension models, and application of detachment models has already provided a fruitful and productive area of research. For example, the detachment concept can be simply and elegantly applied to the evolution of the Atlantic-type passive margin (for example the Baltimore Canyon segment). A model has been envisaged in which initially there is an offset between extension at the surface, and extension in the deep crust (Fig. 6a), but as extension continues, the area of intensely disrupted upper crust expands until it encounters intensely stretched middle to lower crust (Fig. 6b). Mantle melting is earliest induced in this area, so this is where final continental break-up takes place. Subsequent subsidence leads to the development of thick sag sequences in the seawards side of the margin, because here the greatest amount of lithospheric extension has taken place (Fig. 6c). On land, adjacent to the margin, little subsidence has taken place, although significant extension took place in the upper crust. The detachment concept has led, and I think it will continue to lead, to the re-interpretation of the structural evolution of hydrocarbon-producing basins in Australia, and it is therefore of considerable interest.

Acknowledgements

Work reported in this paper has been carried out in collaboration with Professor Greg Davies, of the University of Southern California, Los Angeles; Dr Steve Reynolds, of the Arizona Geological Survey, Tucson; and Dr Mike Etheridge and Phil Symonds, BMR, Canberra. An extended bibliography of relevant articles is available on request.

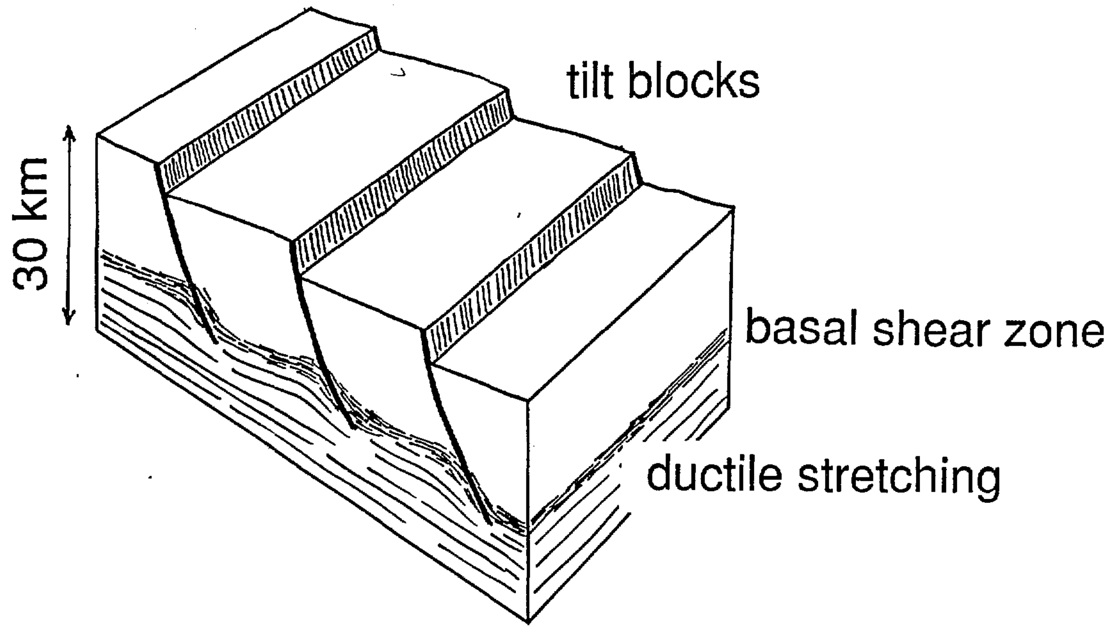

Figure 1: A model for the extension of the continental crust based on seismological observations in contemporary extending terranes. An array of steeply-dipping, deeply-biting planar normal faults is shown. These faults are seismically active, and occasionally on-going extension results in great earthquakes. The ruptures propagate both upwards, towards the surface, and downwards, beneath the depth of the brittle-ductile transition, as it pertains to average strain rates, between seismic events. A basal shear zone forms to maintain strain compatibility between these blocks, and the ductilely-stretching lower crust.

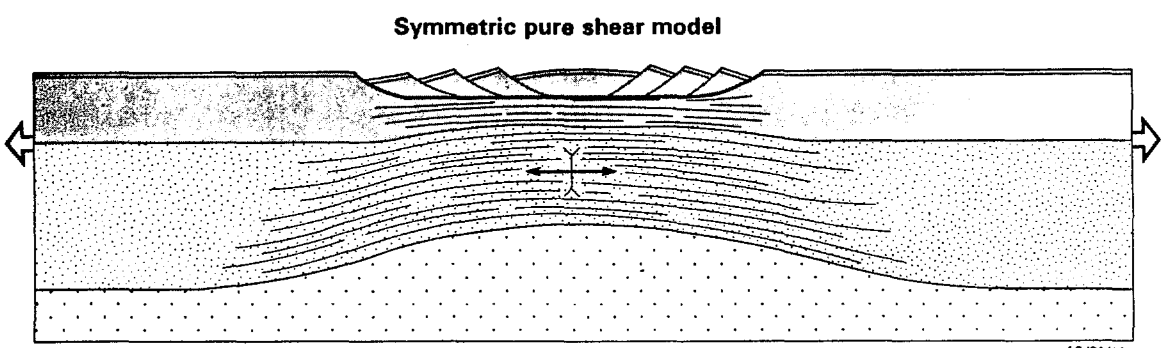

Figure 2: An earlier model for continental extension by pure shear which envisaged the transition between the brittle-upper crust, and the ductile lower crust as a flat-lying detachment fault.

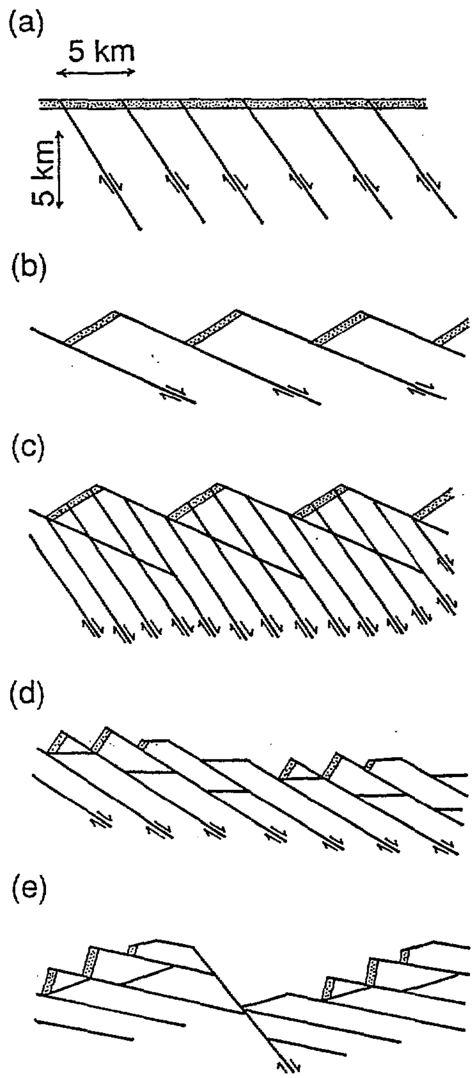

Figure 3: Crustal-scale tilt blocks (a) form dominoes which rotate as extension proceeds, (b) new normal faults form, defining new dominoes, (c), which then rotate so that earlier-formed faults are highly segmented, but now horizontal. If yet another generation of steeply-dipping normal faults forms (the present-day Basin and Range Province faults, for example), additional rotation will ensue, so as to produce flat-lying tilt blocks with vertical strata, (e), for example, as reported at Yerrington, or in the El Dorado Mountains.

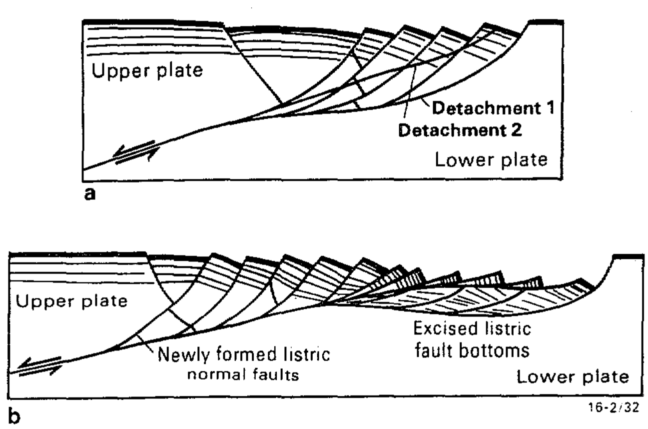

Figure 4: The youngest structures in many of the core complexes are shallow-dipping normal faults of large areal extent, termed detachment faults. These form with an episodic history (a) and complex structures result as the consequence of multiple generations of such detachment faults. Diagram (b) shows the effect of excisement of listric fault bottoms.

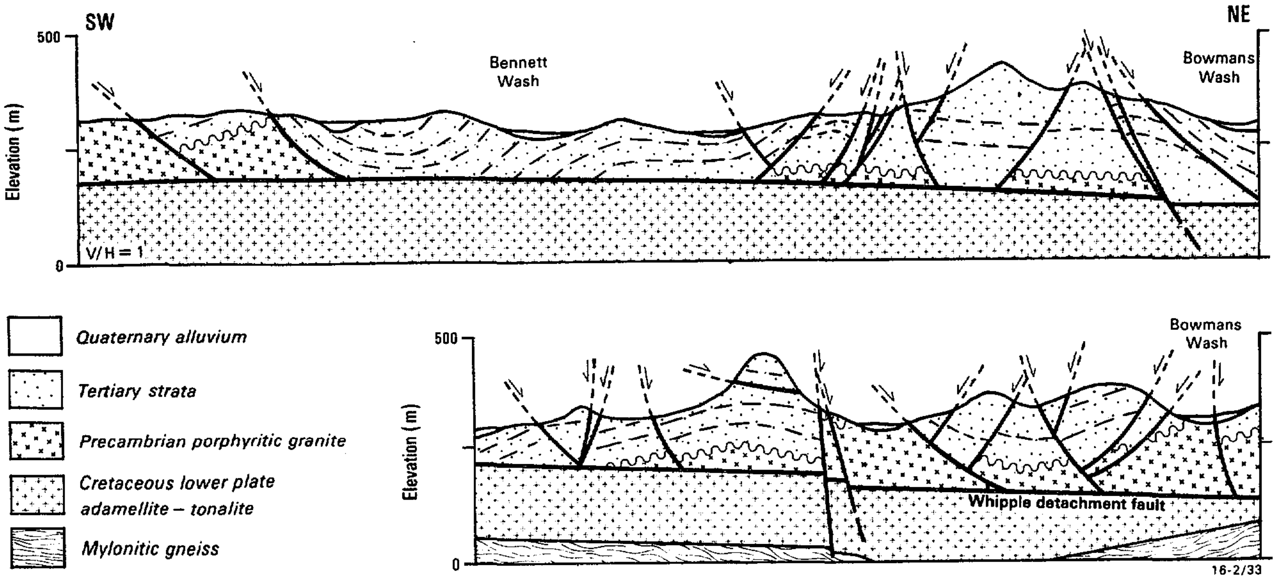

Figure 5: An example of excisement is presented from the Whipple Mountains (after Thurn). The detachment fault is clearly the youngest structure evident in the section.

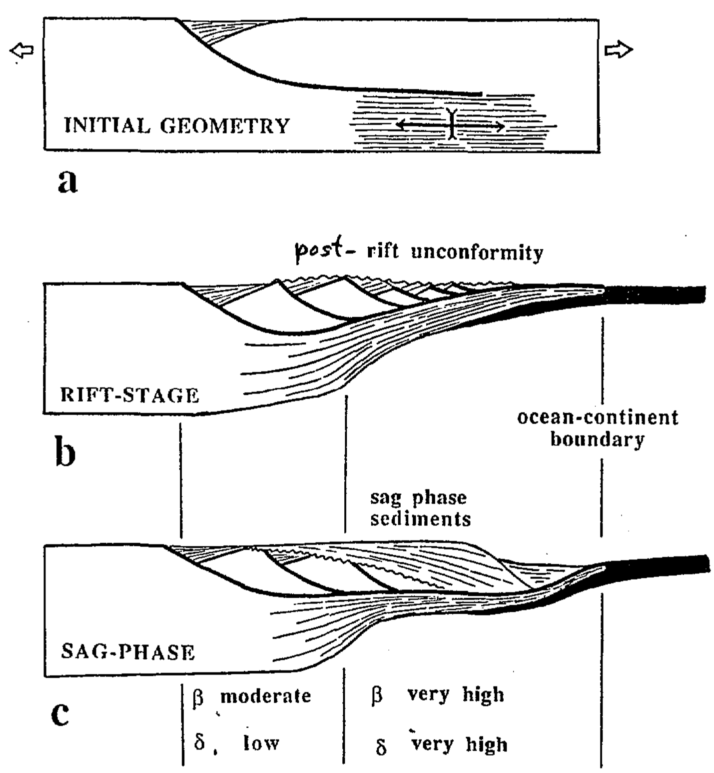

Figure 6: The evolution of the Atlantic passive margin can be explained in terms of the concept of detachment. Initially, the detachment fault links the area of surficial extension with the ductilely-stretching middle to lower crust (a). As extension proceeds, the area of surficial extension expands (b), and encounters rocks stretched previously at depth. Uplift persists after rifting ceases, and the post-rift unconformity develops (b). Subsequently thermal anomalies relax, and subsidence takes place. Break-up took place where lithospheric extension was greatest, because here the amount of partial melting in the mantle was most pronounced. This region is subject to the most subsidence, as here the thermal anomalies were the greatest.

Continental extension and related igneous history in the Basin and Range Province, USA

S J Reynolds, Arizona Geological Survey, USA

The Basin and Range Province (BRP) of the southwestern USA is one of the world's classic examples of regional crustal extension. Surface elevation varies from near sea level in the southern BRP to over 2 km in the higher basin floors in the northern BRP. The province is underlain by thin continental crust that is 25 to 35 km thick. Seismic reflection data reveal that the reflection moho is generally not offset by major faults and is overlain by a lower crust characterised by numerous discontinuous, subhorizontal reflectors.

Crustal extension affected the region during middle to late Cainozoic time and is generally divisible into two distinct episodes: one during the middle Tertiary (Oligocene to mid-Miocene), and another from the mid-Miocene to present. These two episodes are generally distinct in (1) style of deformation; (2) direction of extension; and (3) type of associated volcanism. The middle Tertiary (MT) episode resulted in the larger amount of crustal extension, but the more recent episode is responsible for most of the present block-fault physiography.

Middle Tertiary extension

The MT episode of crustal extension largely occurred between 30 and 15 Ma and resulted in large amounts of east-northeast to west-southwest extension, commonly exceeding 100 per cent. The upper-crustal expressions of MT extension include: (1) tilted fault blocks bounded by low- to high-angle normal faults; (2) regional low-angle shear zones with tens of km of normal displacement; and (3) aligned dyke swarms and elongated plutons emplaced in tension fractures oriented perpendicular to the extension direction.

Tilted fault blocks that were formed during MT extension vary in width from small-outcrop-scale examples to 20-km-wide mountain ranges. Stratal dips are generally less than 60^{o} in the largest fault-blocks, but are steeper in some smaller fault-blocks. Normal faults bounding the blocks have as much as 5 km of normal displacement and mostly formed at moderate to steep dips, but have been rotated to gentle dips in areas of extreme extension. Half-grabens formed during fault-block rotation contain syntectonic sedimentary and volcanic rocks, commonly with fanning dips. Most half-graben fill includes kilometre-long blocks or sheets of megabreccia, representing catastrophic debris avalanches; such megabreccias are shattered, but commonly preserve the original stratigraphy of the rocks from which the debris avalanche was derived.

Regional low-angle normal faults, termed 'detachment faults', accommodated large amounts of upper-crustal extension. These faults are interpreted as having formed with dips of less than 30°, except at their breakaway where they were probably listric. The larger detachment faults have cumulative displacements of tens of kilometres and have brought mid-crustal footwall rocks to the surface because of tectonic denudation of upper-plate rocks and associated isostatic uplift. The term 'metamorphic core complex' is applied to areas where these deep-level rocks are exposed at the surface.

Structures in the lower plate of detachment faults (in metamorphic core complexes) record the ductile-to-brittle evolution of the rocks as they were uplifted to high structural levels by gently-dipping mylonitic zones formed by ductile shearing along the deeper segments of the detachment zone. These mylonitic zones are cut by superimposed brittle zones formed after the rocks were uplifted through the brittle-ductile transition during detachment faulting. The observed structural sequence is consistent with models that interpret the detachment faults and core complexes in terms of an evolving shear zone that accommodated crustal extension.

Upper-plate rocks above detachment faults are commonly brittlely-distended into tilted fault blocks, like those described above. The faults bounding upper-plate fault blocks are truncated downward by, or merge with the underlying detachment fault. Most upper-plate faults are planar, but some are listric. Strata in upper-plate fault blocks commonly have a uniform direction of stratal tilt (and consistent opposing dip of normal faults) over large areas or tilt-block domains. The dip of strata in such domains is generally antithetic to the direction of upper-plate transport on the underlying detachment fault. Some boundaries between opposite-dipping tilt domains are parallel to extension and are analogous to transfer faults. Boundaries perpendicular to the extension direction (parallel to the strike of tilted fault blocks) are commonly offset into segments by transfer faults.

Late Tertiary crustal extension

The late Cainozoic Basin and Range (BR) episode of crustal extension began at 15 to 13 Ma in the southern BRP (Arizona, New Mexico, and southern California). Basin-and-Range-style faulting gradually encroached northward into the northern BRP (Nevada, Utah, and southern Idaho) with the northward migration of the Mendocino triple junction and associated lengthening of the San Andreas transform system. The locus of active normal faulting is concentrated near the margins of the province, and much of the southern BRP has been tectonically inactive since about 5 Ma.

BR extension resulted in an average of 10 to 30 per cent crustal extension, or much less than MT extension. BR extension was east-west directed and formed large, moderately-dipping normal faults that outline many of the present-day basins and ranges. Compared with MT extension, BR extension formed wider, less-rotated fault blocks. Also, except for the Sevier Desert detachment in central Utah, BR extension evidently did not form major gently-dipping detachment faults, at least at the surface.

Tectonic setting and magmatism

The two episodes of crustal extension occurred in different tectonic regimes and were associated with different types of magmatism. MT extension was broadly synchronous with generation of large volumes of compositionally-expanded (mafic, intermediate, and felsic) magmas. MT silicic ash-flow tuffs and silicic to mafic lava flows cover broad areas. Consanguineous granitic plutons and dyke swarms are likewise widespread, especially in the tectonically denuded rocks beneath detachment faults. Detachment faulting, therefore, has provided exposures of different crustal levels of the MT magmatic system.

MT magmatism occurred in a continental-arc setting as oceanic lithosphere of the Pacific Ocean was being subducted eastward beneath the western margin of North America. Before MT time, high convergence rates caused the subducted oceanic slab to dip at too shallow an angle for generation of a typical continental-margin arc. As convergence rates slowed at about 40 Ma, the gently-dipping subducted slab began to steepen in dip and lose coherence. Because of this slab steepening and distintegration, MT magmatism is time-transgressive. In the southern BRP, magmatism started in the east at 35 Ma and had reached the western edge of the province by about 20 Ma. At 20 Ma, magmatism became somewhat bimodal in character, with roughly equal volumes of mafic rocks (basalt and basaltic andesite) and silicic flows (rhyolite to dacite). In the northern BRP, the magmatic arc swept southwestward out of the northern Rocky Mountains and across the province between 40 and 15 Ma. This magmatic sweep was punctuated by eruption of extensive ash-flow sheets and caldera collapse.

In at least the southern BR, there was a marked change in the character of magmatism at 15 to 12 Ma, corresponding with the end of MT extension and a change from subduction to transform tectonics. Unlike the earlier compositionally-expanded magmatism, late Tertiary and Quaternary BR volcanism was fundamentally basaltic or bimodal (basalt-dacite) in character. Nearly all volcanic rocks erupted after 12 to 15 Ma, including silicic rocks in bimodal fields, have initial $^{87}Sr/^{86}Sr$ less than 0.7055, whereas those erupted before 12 to 15 Ma (i.e. in MT time) have initial ratios greater than 0.7055. This change in Sr isotopic composition reflects a change from MT magmas with large crustal components to BR magmas with a dominant mantle signature. This change, in turn, probably reflects a strengthening of the lower crust in late Tertiary time, in that BR mantle-derived magmas could traverse the crust via through-going fractures, whereas earlier MT magmas were contaminated upon encountering a ductile, partially-molten lower crust. A strengthening of the crust with time would account for the change in style of extension from (1) extreme extension accommodated by gently-dipping detachment faults to (2) more moderate amounts of extension along spaced moderate- to high-angle normal faults. Finally, an inflow of mantle-derived melts during both episodes may account for the presently-flat character of the moho.

References

Crittenden, M.D., Jr., Coney, P.J., & Davis, G.H., (Editors), 1980 - Cordilleran metamorphic core complexes. Geological Society of America Memoir 153.

Davis, G.A., Lister, G.S., & Reynolds, S.J., 1986 - Structural evolution of the Whipple and South Mountains shear zones, southwestern United States. Geology, 14, 7-10.

Davis, G.H., 1983 - Shear-zone model for the origin of metamorphic core complexes. Geology, 11, 342-347.

Klemperer, S.L., Hauge, T.A., Hauser, E.C., Oliver, J.C., & Potter, C.J., 1986 - The Moho in the northern Basin and Range Province, Nevada, along the COCORP 40^{O}N seismic–reflection transect. Geological Society of America Bulletin 97, 603–618.

Miller, E.L., Gans, P.B., & Garing, J.D., 1983 - The Snake Range decollement: An exhumed ductile–brittle transition. Tectonics, 2, 239–263.

Smith, R.B., & Eaton, G.P., (Editors), 1978 - Cenozoic tectonics and regional geophysics of the western Cordillera. Geological Society of America Memoir 152.

Wernicke, B., 1981 - Low–angle normal faults in the Basin and Range province: nappe tectonics in an extending orogen. Nature, 291, 645–648.

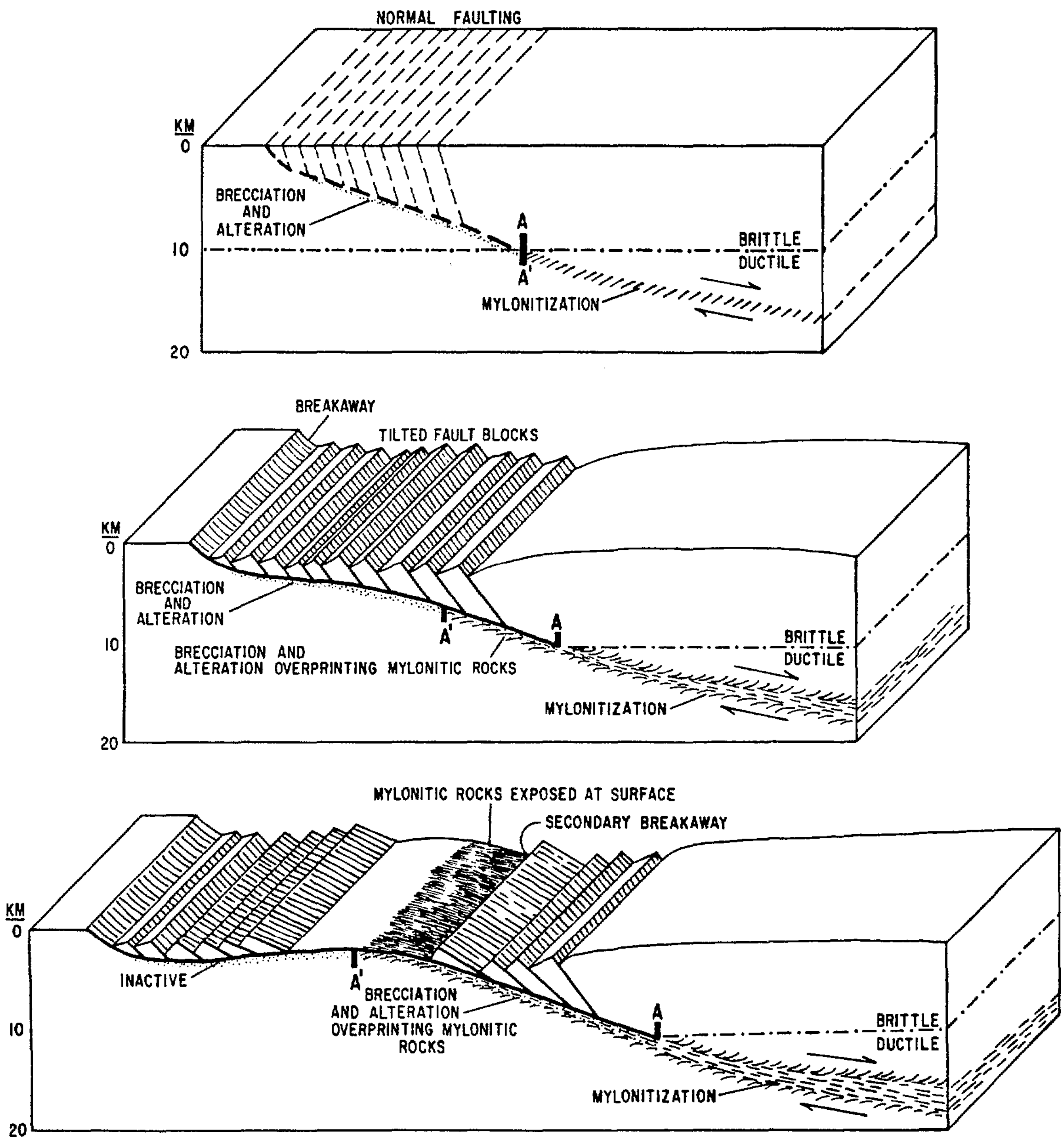

Figure 1. Low-angle shear-zone model for the evolution of detachment faults.

Finite deformation models of lithospheric extension during continental rifting

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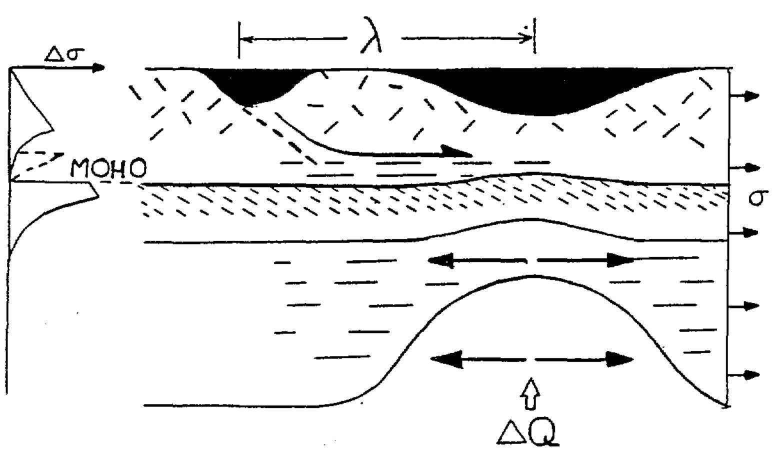
We use a finite element model to investigate the dynamics of finite extension of the continental lithosphere during rifting. The model includes a zone of crustal weakness, from previous faulting or suturing, that is laterally offset by λ from a zone of lower lithosphere weakening from heating above a mantle plume or from a magma injection. Under what circumstances is pure-shear extension of the lower lithosphere transferred along a lower crustal shear channel to reactive simple-shear extension at the offset weakness?

Factors including: initial lithospheric properties (composition and ambient thermal state, which determine rheology, in particular lower crustal viscosity); the far field tension σ; and rate of lithospheric heating, determine the maximum value of λ for reactivation, as well as the timing of rifting and final shape of the necked lithosphere.

A range of rifting styles, that includes pure shear of the whole lithosphere and simple shear of the crust as end members, is predicted. That these styles look like the styles of rifted continental margins reinforces our view that mantle rheology controls the overall form of lithospheric necking. Crustal extension differs from mantle extension reflecting the strength, and inherited weakness, of the upper brittle layer and the ability of low-viscosity crustal-channels to transfer strain to the weak zones.



Geophysical character of continental extended terranes

P Wellman, BMR

Gravity and magnetic anomalies generally do not reflect the primary structure of extensional basins, rather they provide supplementary information of the geometry of the basin fill. In young sedimentary basins without carbonates or volcanics, the gravity mainly reflects the difference between the attraction of the sediments and their isostatic compensation, while the magnetic anomalies are most useful in estimating the depth to magnetic basement. The following examples are interpretations of the gravity and magnetic anomalies over a variety of extensional basins in Australia and New Zealand.

The Bass Basin is an intra-cratonic basin north of Tasmania, formed by Late Jurassic to Early Cretaceous extension along a northwest-trending axis (Etheridge & others, 1984; Williamson & others, 1985) (Fig. 1, A-C). The basin contains 10 km of sediments. Free-air gravity anomalies show a broad low over the basin of about 200 $\mu m.s^{-2}$, the low having a shape similar to that of the basin. It is to be expected that the rift sediments would have overlain hot, and therefore weak, lithosphere, with essentially local isostatic compensation. Magnetic anomalies show no obvious correlation with the distribution of sediments in the basin. The anomalies are attributed to pre-rifting basement intrusions, and Cainozoic mafic volcanics. The magnetic gradients have two dominant directions; these directions parallel the transfer and the normal faults of the rifting.

The McArthur Basin (Northern Territory) is a Middle Proterozoic crustal rift, which extended in stages over the period 1700-1400 Ma ago (Fig. 2, A-C). Using mapped geology, together with gravity and magnetic anomalies, Plumb & Wellman (1987) have mapped the rift margin, the transfer faults, the inferred sediment thickness, and sediment type. Over much of the basin, magnetic anomalies due to dolerite dykes and sills interfere with estimates of depth to basement. The basin is unusual in that three major sediment groups give different gravity and magnetic anomalies. The rift margin is inferred from a combination of sediment thickness pattern from geology, the greater amplitude of gravity and magnetic anomalies within the rift, and the truncation of inter-rift anomalies at the rift margin.

The Adelaide Orogen is a late Proterozoic to early Phanerozoic rift in South Australia (Von der Borch, 1980) (Fig. 3, A–B). The rift is bounded by the Gawler Craton in the west, and by the Curnamona Cratonic Nucleus in the northeast; other boundaries are poorly defined. The area of the rift is a magnetic low, with major elongate magnetic anomalies along parts of the western margin. The magnetic low is consistent with the general lack of volcanics in the rift-fill. The main part of the rift, south of Leigh Creek, can be divided into two strips, a western strip that is a relative gravity low, and an eastern strip that is a relative gravity high and which underlies the main axis of the Flinders/Mount Lofty Ranges. Seismic results are interpreted as indicating that there is little or no crustal root under the ranges. The distribution of rock types and folds appears unrelated to the gravity strips. The cause of the two-fold gravity division of the rift is not understood.

The Central Volcanic Region in New Zealand is an active back-arc basin within continental crust (Fig. 4, A–B). Seismic refraction work is interpreted as indicating that under the rift there is a thin crust (15 km) of mainly 6.1 $km.s^{-1}$ seismic velocity, with a low upper mantle seismic velocity of 7.4 kms^{-1}. There is high heat flow. The Region has a low residual gravity anomaly of about 500 $\mu m.s^{-2}$, with steep sides, and a broad magnetic high of amplitude about 50 nT. Relative to the rest of the Region, the volcanically active part (called the Taupo Volcanic Zone) has a gravity anomaly about 100 $\mu m.s^{-2}$ more negative, and magnetic anomalies about 100 nT higher with a very irregular pattern. Similar volcanic rocks of Carboniferous age occur in eastern Queensland near Georgetown (Newcastle Range Volcanics, Fig. 4, C–D), and near Mount Coolon (Bulgonunna Volcanics, Fig. 4, E–F). These volcanic rocks occur in elongate depressions that are thought to be formed in either a back-arc or a continental extension environment. The gravity and magnetic anomalies over these elongate depressions are similar to those over the Taupo Volcanic Zone. The similarity in geophysical anomaly may be due solely to similar rock types in a depression with similar geometry, or may be due in part to similar lower crustal structure.

References

Etheridge, M.A., Branson, J.C., Falvey, D.A., Lockwood, K.L., Stuart-Smith, P.G., & Scherl, A.S., 1984 - Basin-forming structures and their relevance to hydrocarbon exploration in Bass Basin, southeastern Australia. BMR Journal of Australian Geology and Geophysics, 9, 197–206.

Malahoff, A., 1968 - Origin of magnetic anomalies over the Central Volcanic Region of New Zealand. In, Knopoff, L., Drake, C.L., & Hart, P.J., (Editors)

- The crust and upper mantle in the Pacific area. American Geophysical Union, Geophysical Monograph, 12, 218–240.

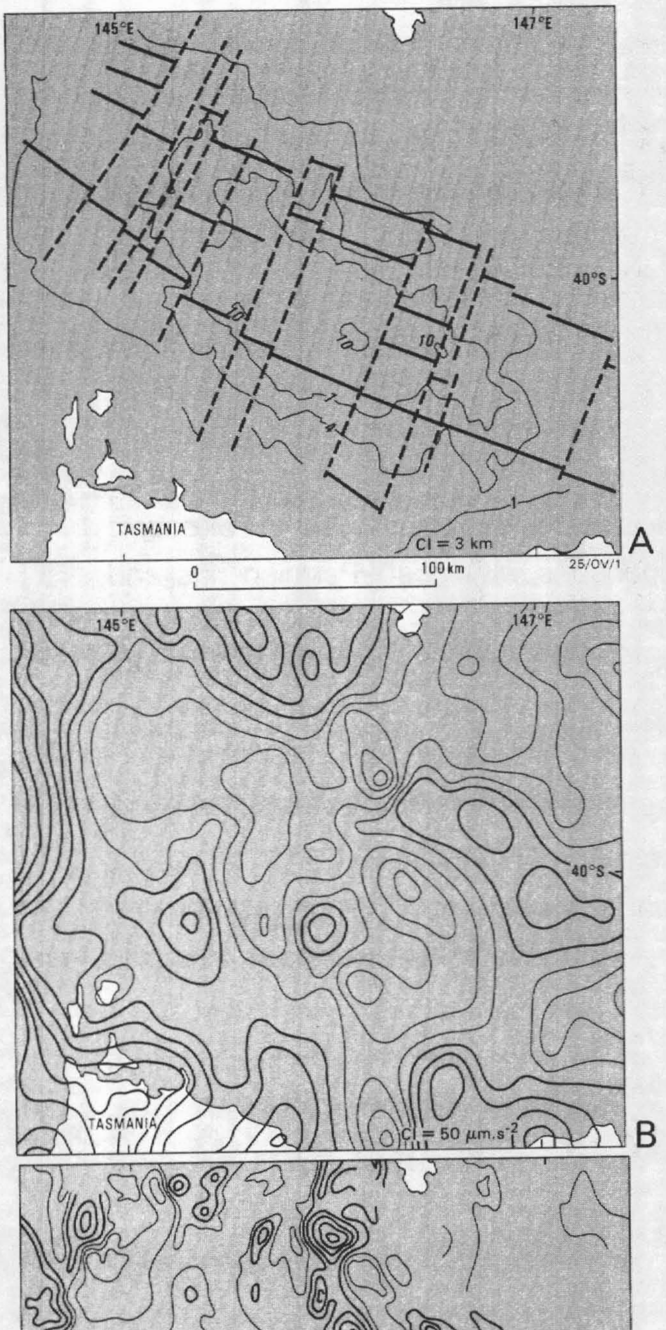
Plumb, K.A., & Wellman, P., 1987 - McArthur Basin, Northern Territory: mapping of deep troughs using gravity and magnetic anomalies. BMR Journal of Australian Geology and Geophysics, 10, 243–251.

Stern, T.A., 1986 - Geophysical studies of the upper crust within the Central Volcanic Region, New Zealand. In, Smith, I.E.M., (Editor) - Late Cenozoic volcanism in New Zealand. Royal Society of New Zealand Bulletin, 23, 92–111.

Von der Borch, C.C., 1980 - Evolution of Late Proterozoic to Early Palæozoic Adelaide Fold Belt, Australia: comparisons with post–Permian rifts and passive margins. Tectonophysics, 70, 115–134.

Watts, A.B., Karner, G.D., & Steckler, M.S., 1982 - Lithospheric flexure and the evolution of sedimentary basins. Philosophical Transactions of the Royal Society of London, A305, 249–281.

Williamson, P.E., Pigram, C.J., Colwell, J.B., Scherl, A.S., Lockwood, K.L., & Branson, J.C., 1987 - Pre–Eocene stratigraphy, structure, and petroleum potential of Bass Basin; a national achievement. APEA Journal, 25, 362–381.



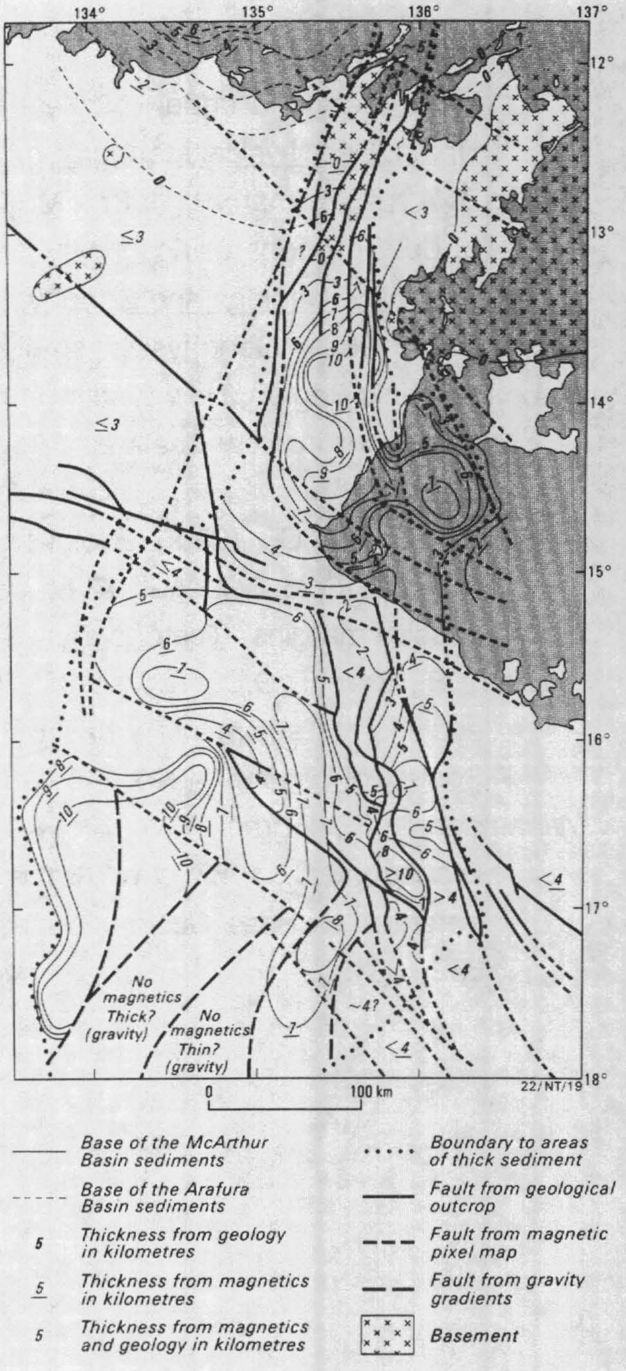
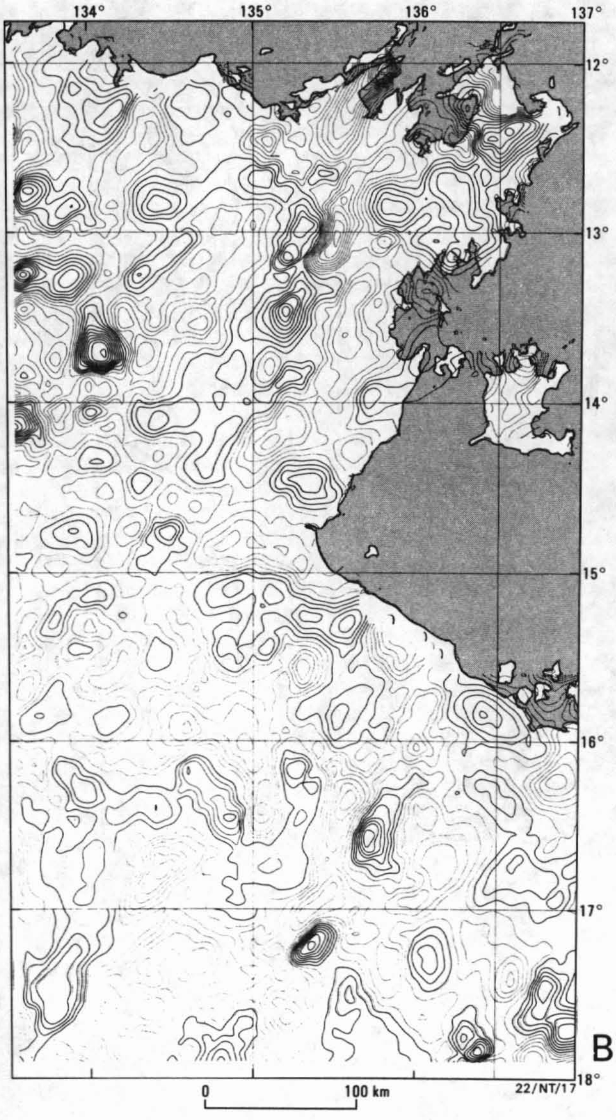


Fig. 2A McArthur Basin. Inferred depth of sediments and faults.

Fig. 1 Bass Basin. A. Contours give depth to ?Jurassic pre-rift unconformity; solid lines give rift-forming normal faults, and dashed lines give transfer faults. B. Free-air gravity anomalies (thick lines positive). C. Magnetic anomalies (thick lines positive).



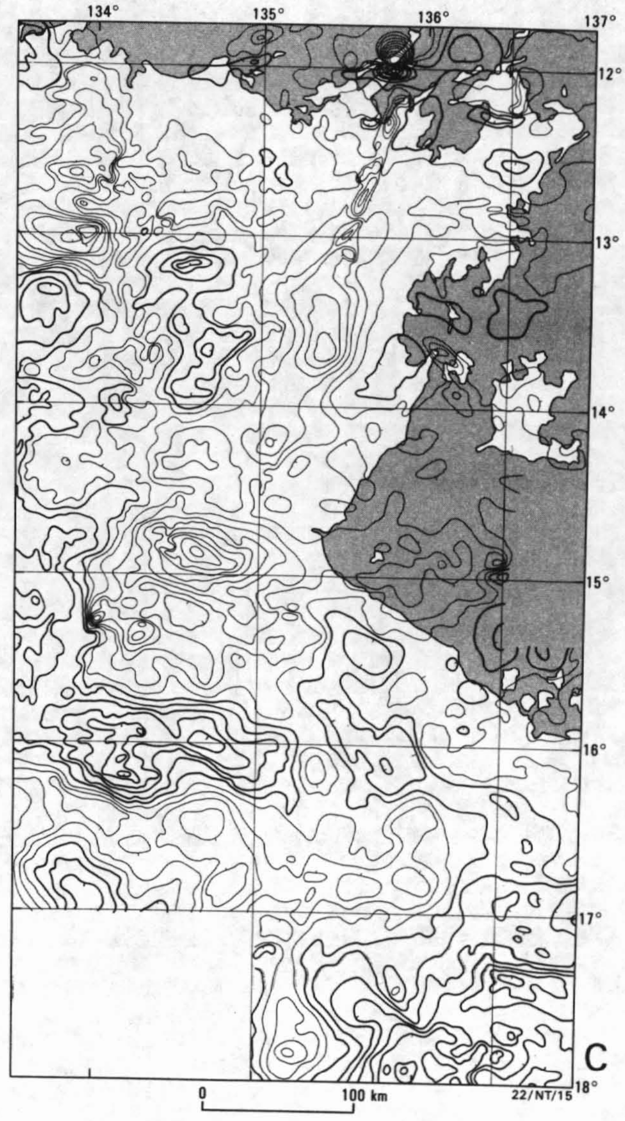


Fig. 2B McArthur Basin. Gravity anomalies, second vertical derivative. Contour interval 50 $\mu m.s^{-2}/(11\ km)^2$, negative contours thick.

Fig. 2C McArthur Basin. Magnetic anomalies. Contour interval 25 nT, negative contours thick.

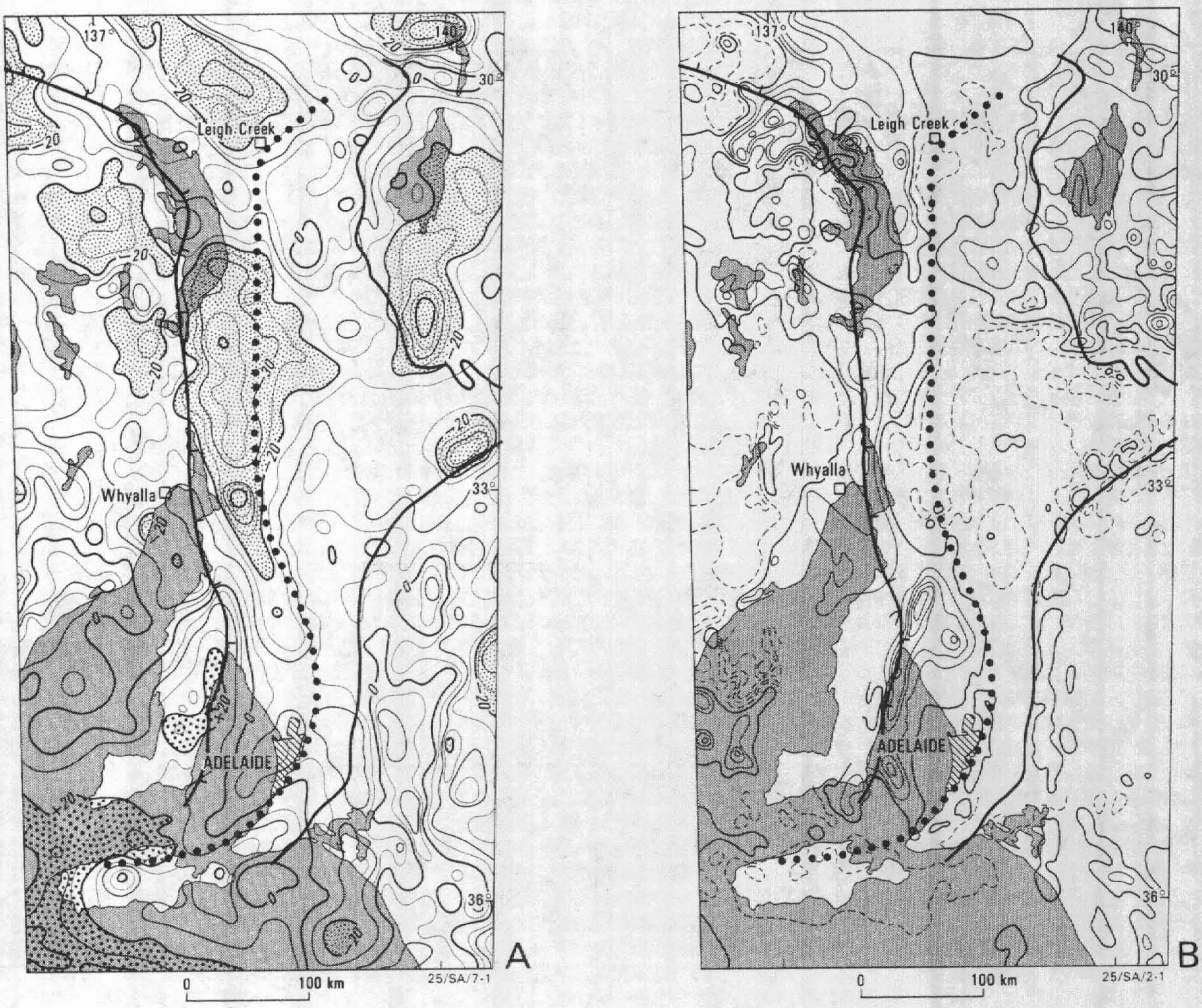


Fig. 3 Adelaide Orogen. A. Bouguer gravity anomalies, contour interval 50 $\mu m.s^{-2}$, density 2.67 $t.m.^{-3}$. B. Magnetic anomalies, contour interval 250 nT, negative contours dashed. Solid line shows rift margin; dotted line shows gradient in A.

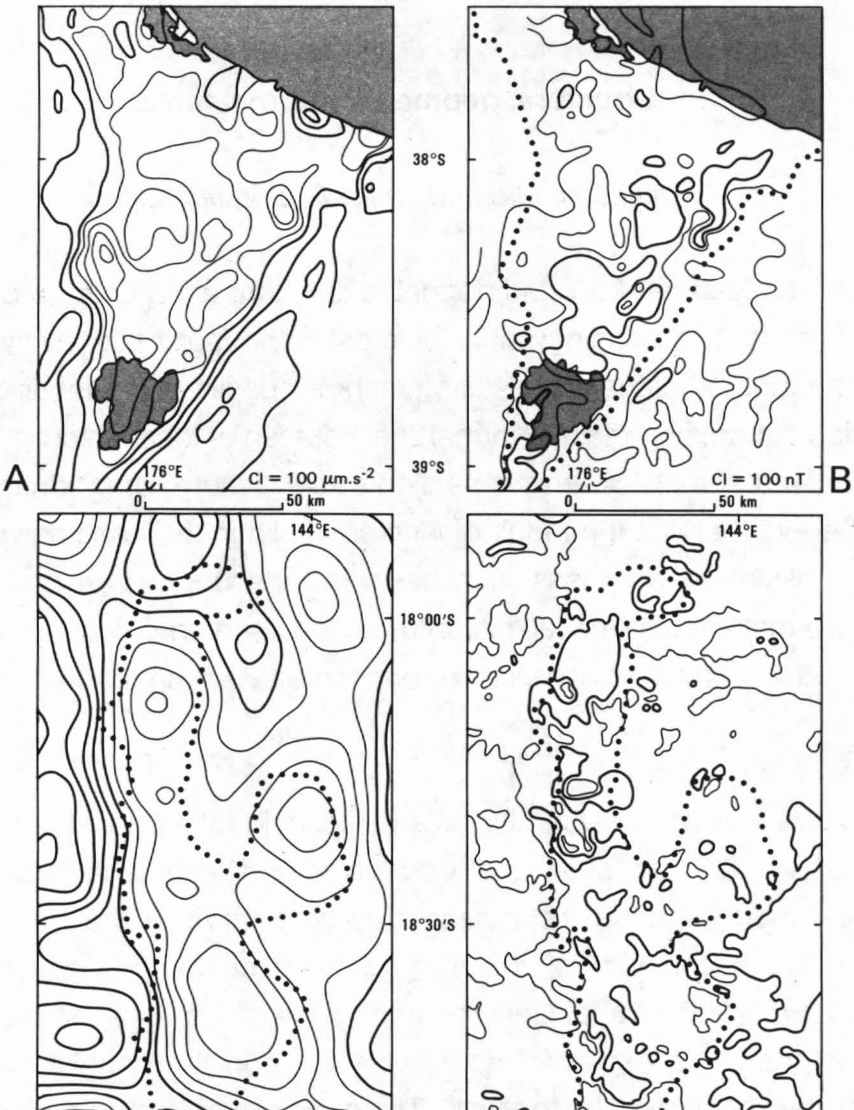


Fig. 4 A-B. Central Volcanic Region of New Zealand; C-D. Newcastle Range Volcanics, Queensland. E-F. Bulgonunna Volcanics, Queensland. A, C & F. Residual gravity anomaly. B, D & F. Residual magnetic anomaly. Dotted line gives rift margin; positive contours are thicker.

Structural geometry: normal faults

A Gibbs, Midland Valley Exploration Ltd

The diagnostic feature of a listric normal fault is that a dip change occurs across the fault from foot to hangingwall. For most listric faults this involves a rotation towards the footwall of the pre-faulting marker. Basinward depositional dips and compaction dips may remain. Normal drag is seen at local outcrop scale but it is rarely observed at map or seismic section scale. Where "drag" does occur some authors believe that it is the result of increased slip in the hangingwall away from the fault, possibly due to loss of cohesion in the upper part of the roll-over. Variable slip models of listric faults can rarely be used analytically or quantitatively and are usually invoked to explain phenomena which cannot be otherwise explained.

All listric faults can be analysed by balance techniques to construct the shape of the fault from the roll-over and vice versa. These techniques assume a geometrical model for the movement of each particle in the hangingwall. The simplest such technique is the so called "Chevron" construction (Fig. 1) introduced by Peter Verral of Chevron in a series of lectures. This assumes that the geometry can be explained as if every point in the hangingwall moved by vertical simple shear over the footwall. The assumption implies nothing about real deformation mechanisms and is frequently criticised as being geologically unrealistic and "modified" Chevron constructions have been introduced which are claimed to be geologically more realistic.

Figures 2 through 6 show the principal modifications that have been proposed. They are included here as some of the construction techniques are not published in a readily available form. The present figures are taken from JAPEC Course Notes 49 by Coward and Gibbs published by the Geological Society of London following the work of Williams, Vann and others. The final Figure in this series, 7, shows the effect of varying the amount of shear by, for example, changing cohesion on the fault plane.

The modified constructions make assumptions such as layer parallel shear, or uniform inclined shear with a general model proposed by Kempler (1987, in press). Inclined shear models work well on single roll-over problems but present insuperable problems with multiple fault systems where the order of faulting must be known. Where the hangingwall contains an upper detachment perhaps with antithetic faults it may not be possible to use a single shear sense for the whole hangingwall.

Layer parallel shear is a useful approach in thrust belt folding but cannot be applied to cases where there is non-uniform non-parallel layering. This is its major failing, as growth sequences are the rule rather than the exception in basin formation. Inclined simple shear problems are resolved iteratively and lack the geometric simplicity of the basic Chevron construction which allows intermediate bed positions to be calculated from well data once a single roll-over and fault have been constructed. The flexibility and extreme simplicity of graphical use of Chevron is its major feature. Experience with industry seismic data has shown that the faults constructed using this construction are invariably within the resolution of the technique and depth conversion assumptions. Moreover Chevron is independent of vertical scale exaggeration, so can frequently be used directly on seismic data where velocity variations within the section are not too great. Figure 6 shows the differing fault detachments which some of the constructions give.

Listric faults may either be normally concave upwards or concave downwards, as, for example, where a fault cuts down as a staircase trajectory through alternating strong and weak lithologies. Chevron construction techniques apply equally well to such cases as to a single simple roll-over. Some further examples of this are discussed in a later contribution in this symposium.

In some cases no rotation between foot and hangingwall markers is observed and such faults must be essentially planar. Jackson (e.g., in Jackson & others, 1982) argues that most major basin-forming faults are of this nature and that listricity seen on many seismic profiles is always the effect of velocity. However the only criterion which is geologically valid to distinguish and identify listric or domino faults is relative rotation of a marker on the foot to hangingwall.

All domino faults require some geological process to accommodate deformation at the "keels" of the blocks and at the ends of the fault arrays. This has led some to suggest that only the sole fault of the array need be listric and the intervening blocks can be essentially planar. In my view this is simply a special case of a listric array.

With domino arrays a further criterion is that the stratigraphic fill between the rotated blocks must all be of the same age. In other words, the array must function by continuous rotation on all of the blocks, unlike a listric array which may operate by either foot or hangingwall collapse with the consequent diachronous fault-block fills. This feature may be used to differentiate between a highly extended originally listric array on a flat and a true domino array.

In listric faults the depth to detachment is given by cross-sectional area in the roll-over divided by the extension. With a domino array it is not possible to use this criterion as the same excess area and extension is seen regardless of whether the dominoes are short and stubby or long pencil-like. Depth to detachment calculations will therefore give the minimum depth to detachment. In the field, however, where domino arrays can be seen, they usually have an aspect ratio of about 1:1 or 1:2 so that the minimum depth to detachment from an excess area calculation is likely to be a realistic estimate in many cases.

Very few fault problems can be solved by reference to a single roll-over and fault. Synthetic fault arrays are very common, with significant antithetic faults present in many cases. Both syn and antithetic faults are likely to be sequential in development and, as with thrust faults, only one active fault is necessary with the other faults in the array behaving passively in the hangingwall. Formation of a new fault is likely to bypass extension on older parts of the array. Once again balanced section construction and modelling the stratigraphic response is the key to understanding how a particular array has developed.

Gibbs (1984) showed how complex linked-fault systems can account for both single and paired graben (Fig. 8). Graphical or computer modelling can be used to investigate more complex systems with breached extensional arrays such as may be found in the Basin and Range Province (Wernicke, 1984; Gibbs, 1984, 1987).

The important factor is that to allow the basin to develop, the fault systems must be linked together, and may contain both domino and listric faults providing that the deformation is balanced. For an extensional dip-slip basin, the criterion of a correct solution to the geometry is that it should be restorable by sequentially removing displacement on the linked faults and compensating for compaction, subsidence and erosion as necessary. For more complex oblique and strike-slip problems, balance cannot be achieved on a single section and must be done in 3D. In these cases particular attention must be paid to transfer elements (see Etheridge, this symposium).

The problems with restoring a complex–fault array as an aid to validating the interpretation or understanding and developmental history of the structure can be solved only by iterative modifications to the section and sequential restoration. Two related issues are central to this process; firstly, is that all the faults must be linked together in cross–section, and ultimately on the map in 3D, and the "regional" must be identified. Different fault linkages may produce valid but non–unique solutions and the sensitivity of any particular problem to this must be known.

The "regional" is defined as the former position of a marker–horizon before deformation. In some cases the regional is simply the former position of a bed in the hangingwall. In the more general case the regional represents the pre–basin surface before faulting, erosion, sedimentation, isostatic uplift, and thermal subsidence, etc. Restoration to the regional is the ultimate objective of balanced section construction and incidentally produces serial palinspastic sections as part of the process. Changes that the interpreter makes to the selection of the regional will radically alter the solution and geological history of the area. Choice of the appropriate regional for each stage in the analysis is non–trivial and is at the heart of the process.

In the course of a single analysis the interpreter is likely to select the simplest possible set of regional assumptions and geometry for the regional, for example ignoring thermal and isostatic effects. Once a restoration can be effected using the fault geometries alone it is then appropriate to ask if thermal or isostatic effects would significantly alter the model. Different interpreters will approach the problem in a variety of ways and experience will allow short–cuts to be made without loss of accuracy.

Although many basins extend along strike for considerable distances with the same tectonic elements present relatively unchanged, transfer components are necessary at all scales. These transfers allow deformation to be balanced between structural components or even have different polarities as pointed out by Bally (1981) and more recently by Rosendahl and his co–workers in the African Rift System.

Transfer faults may therefore share the same detachment as the hangingwall or footwall or create major lateral ramps along strike in the active detachment fault. In the latter case these transfers may detach at base crust or base lithosphere in the case of major tectonic components, or at any intermediate level. Where a transfer fault terminates on the map its displacement is completely transferred to the detachment on which it sits and its three-dimensional shape will be controlled by the detachment on which it steps-down to the fundamental regional detachment.

Transfer faults in basin development differ from strike-slip faults which they resemble in many of their features in the same way as oceanic transform faults. Gibbs (1984) figured transfer faults which off-set a basin trend in a direction which is opposite to the sense of slip on the fault. For this reason reconstruction of palinspastic maps by matching isopachs is invalid. Off-set depocentres may have been off-set since their inception and are pinned to footwall and sidewall ramps.

Regionally, whole basin systems may be bounded by transfers and at this scale the distinction between major strike-slip-zones may become blurred. However, as at plate-scale deformation, slip in a structural compartment must be completely transferred in a geometrically compatible way with deformation or non-deformation in adjacent compartments. The whole system of extension, compression and strike-slip has to be linked to the plated boundaries for all large deformations.

The simplest and probably most common transfer geometry is where the faults are vertical and run perpendicular to the basin boundary faults. In these cases transport of the hangingwall is parallel to the transfers and space problems do not develop. Kinematically this is the simplest relationship and many basins have this geometry. A common but still simple system is where the transfers are still in the transport direction but are oblique to the basin-margin fault, which in this case must be extensional oblique slip. Gibbs (1987) has pointed out that on such faults a small but significant number of oblique slip structural indicators such as reverse faults and en echelon folds are to be expected, these are diagnostic of the true slip system.

Oblique transfers which have both a wrench component and a down-dip slip are apparently less common. In such systems very complex deformation is possible. Figure 10 shows hangingwall displacement oblique to both the basin margin and the transfer. Here the whole hangingwall is listric in three dimensions and the bounding faults both have associated roll-overs. Commonly such systems continue to step in the same sense along the entire length of the basin margin; where the margin steps onboard towards the footwall the transfer component must be either inactive and carried on a lower detachment, or will be transpressional. Mechanically transpressional transfers in basins may be unstable and their existence strongly suggests that the whole system should be regarded as a mixed mode or oblique-slip basin rather than an extensional basin.

The recognition that transfer in both extensional and contractional environments may not be parallel to transport is important in indicating the possibility of oblique-trending structural and stratigraphic traps and of the probable presence of a number of compressional elements.

In basin deformation transfer faults commonly occupy the hangingwall but in some cases the footwall may also be involved and this should be taken as geometric evidence of a lower detachment. Where basin systems are linked on base crustal detachments so as to involve a whole series in intraplate deformations, identification of a fixed footwall reference may be difficult and ultimately the structural balance must be taken to the plate boundaries.

At an intermediate scale, it is commonly seen in strike-section or on map-section that the extensional system is comprised of different tectonic leaves. Each leaf will be subdivided internally by transfers and extensional detachments and bounded along strike by major transfer systems (Fig. 11). Within a single leaf an individual detachment fault will ride on a floor fault and have a corresponding roof fault which floors the structurally overlying sequence. Where such systems do not consist of slip parallel orthogonal transfers the resulting internal arrangement of roll-over extension folds and ramp contraction folds will be very complex and again is probably indicative of major oblique or strike slip components of deformation in the system as a whole.

For the simpler cases of stacked leaves and transfers within each leaf offset from each other, the important consequence is that fault-bounded reservoir compartments may be offset. Transfer controlled traps may then be stacked in an offset manner both down dip and along strike. This situation is likely to be common in the upper levels of basin-fill where there is significant stratigraphic development of over-pressured shale and salt. Stacked fault systems in the carapace of a basin may provide prolific trapping situations.

Transfer-fault-trends of all types are significant in that they provide access for sediments into the basin and between basin compartments. Especially where footwall uplift occurs, only a minor amount of sediment may enter the basin across the basin boundary and associate synthetic fault arrays. Erosion and transport in these cases may be primarily down the roll-over towards the proximal fault. Subsequently, sediments ponding in these areas may be spilled into the next compartment across the transfer faults. In these areas significant stratigraphic anomalies are likely to occur, and it is no accident that oil fields such as the North Sea Brae Field sits in such a stratigraphic build-up located on a major transfer step in the basin.

In basin development it is now possible to identify structural suites of detachment faults, buried and emergent ramps, and transfer components which are diagnostic of its development and behaviour. Basins which are mapped without these components linked together in a way which allows the basin to be reassembled step by step (palinspastically-balanced) have been inadequately described and economic opportunities and potential undoubtedly missed. Modern techniques of iterative section balancing to test the evolving interpretation and forward modelling are the vital elements to any basin analysis.

References

Bally, A.W., 1981 - Atlantic type margins. In Geology of passive continental margins. Am. Assoc. Petrol. Geol. Course note series 19, 1-48.

Etheridge, M.A., Branson, J.C., & Stuart-Smith, P.G., 1987 - The Bass, Gippsland and Otway Basins, southeast Australia: a branched rift system formed by continental extension. In Beaumont, C., & Tankard, A., (Editors) - Sedimentary basins and basin-forming mechanisms. Can. Soc. Petrol. Geol. Mem. 12.

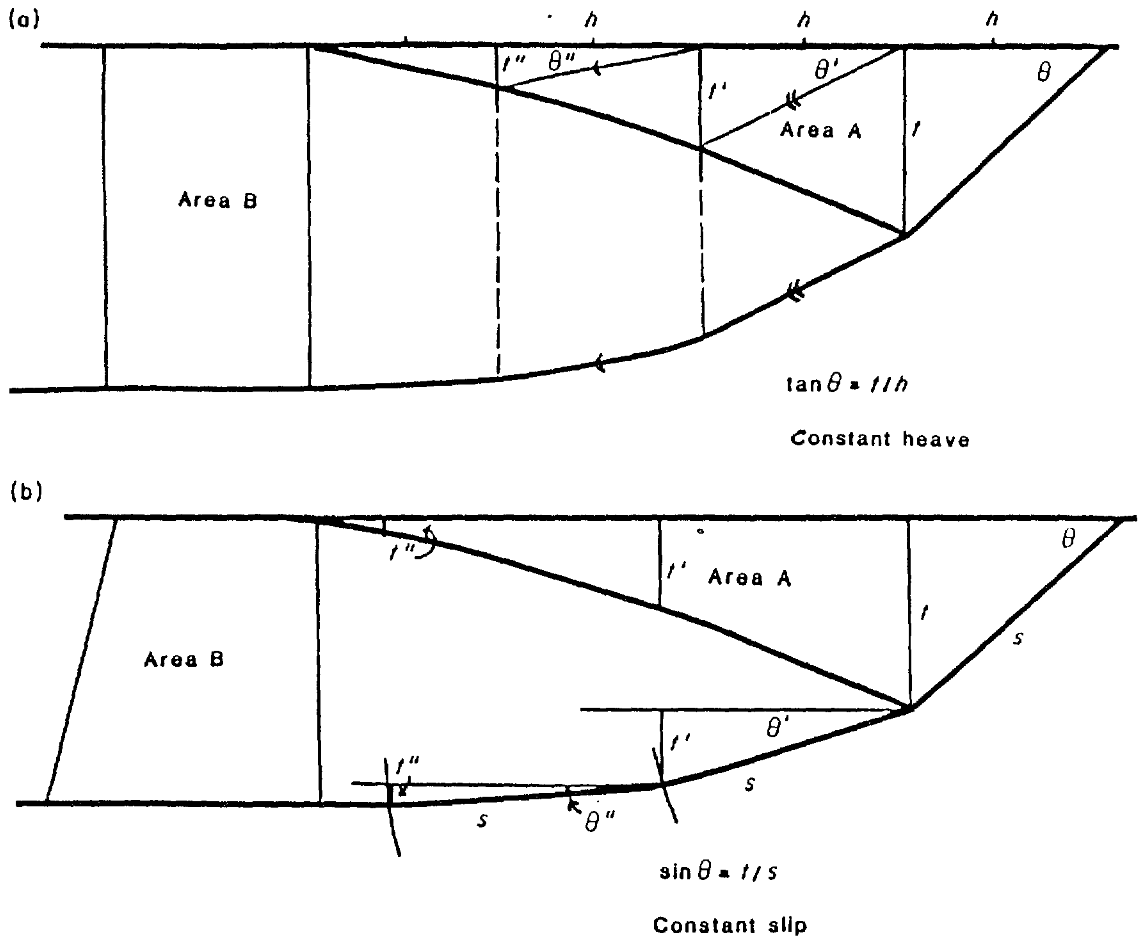
Gibbs, A.D., 1984 - Structural evolution of extensional basin margins. J. Geol. Soc. Lond.,141, 609-620.

Gibbs, A.D., 1987 - Development of extension and mixed-mode sedimentary basins. In Coward, M.P., Davey, J.F., & Hancock, P.L., (Editors) - Continental extensional tectonics. Geol. Soc. Lond. Special Pub, 28, 19-33.

Rosendahl, B.R., 1987 - Architecture of continental rifts with special reference to east Africa. Earth Plan. Sci. Let. (in press).

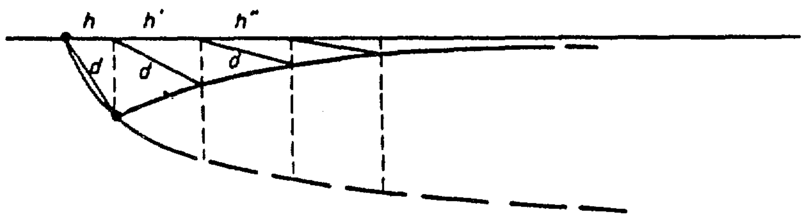
Wernicke, B., 1985 - Uniform-sense normal simple shear of the continental lithosphere. Can. J. Earth Sci., 22, 108-125.

Fig 1



Construction of fault trajectory and depth of detachment from rollover. All points in the hanging wall are assumed to have moved an equal distance, h, to the left, the heave of the fault. The throw, t, decreases as the fault plane dip, θ, changes. b. Construction of fault trajectory and depth of detachment from the same rollover as in Fig. 3a. In this case the slip, s, along the fault plane is constant and heave changes. Areas A and B correspond but, because heave is not constant, the marker rotates as it moves to the left and the slip along the fault plane is greater than the heave in the hanging wall.

MODIFIED CHEVRON CONSTRUCTION I.



to find shape of fault

1 draw displacement vector, length d, from regional to roll-over

2 draw fault parallel to displacement vector for column width h'

assumes constant displacement d along fault

displacement constant for any one vertical line

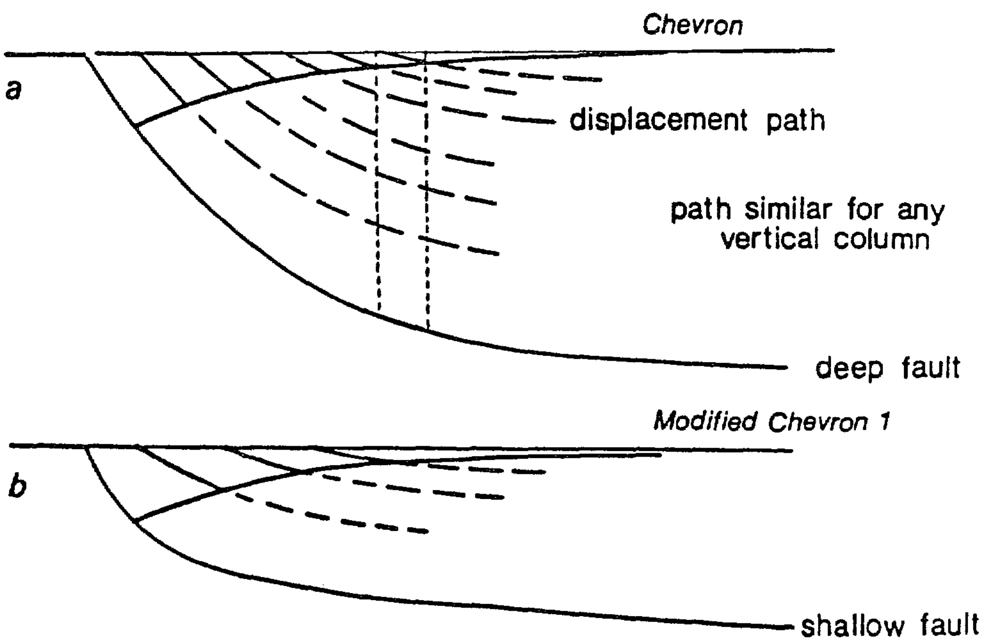


Fig 2

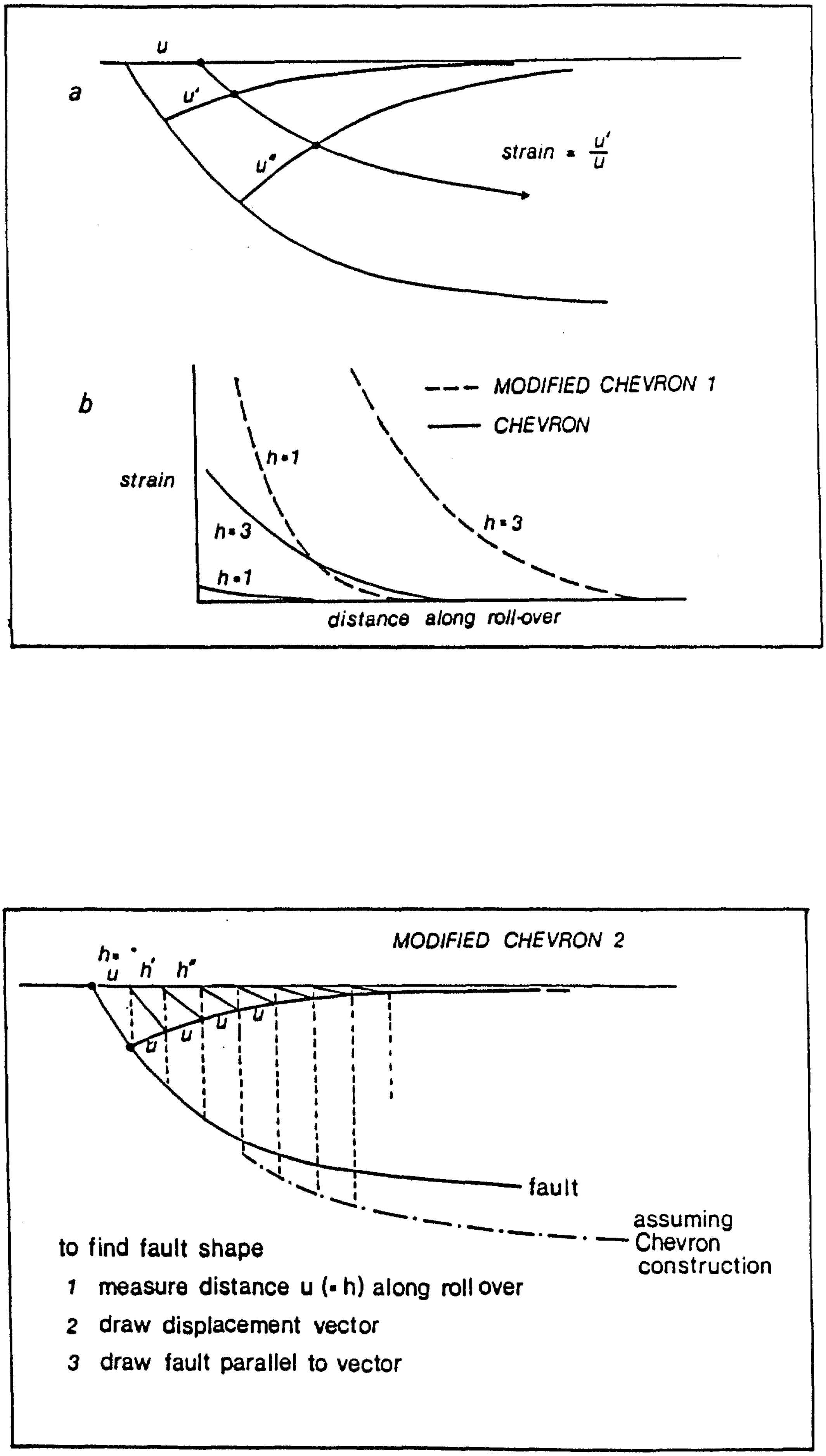
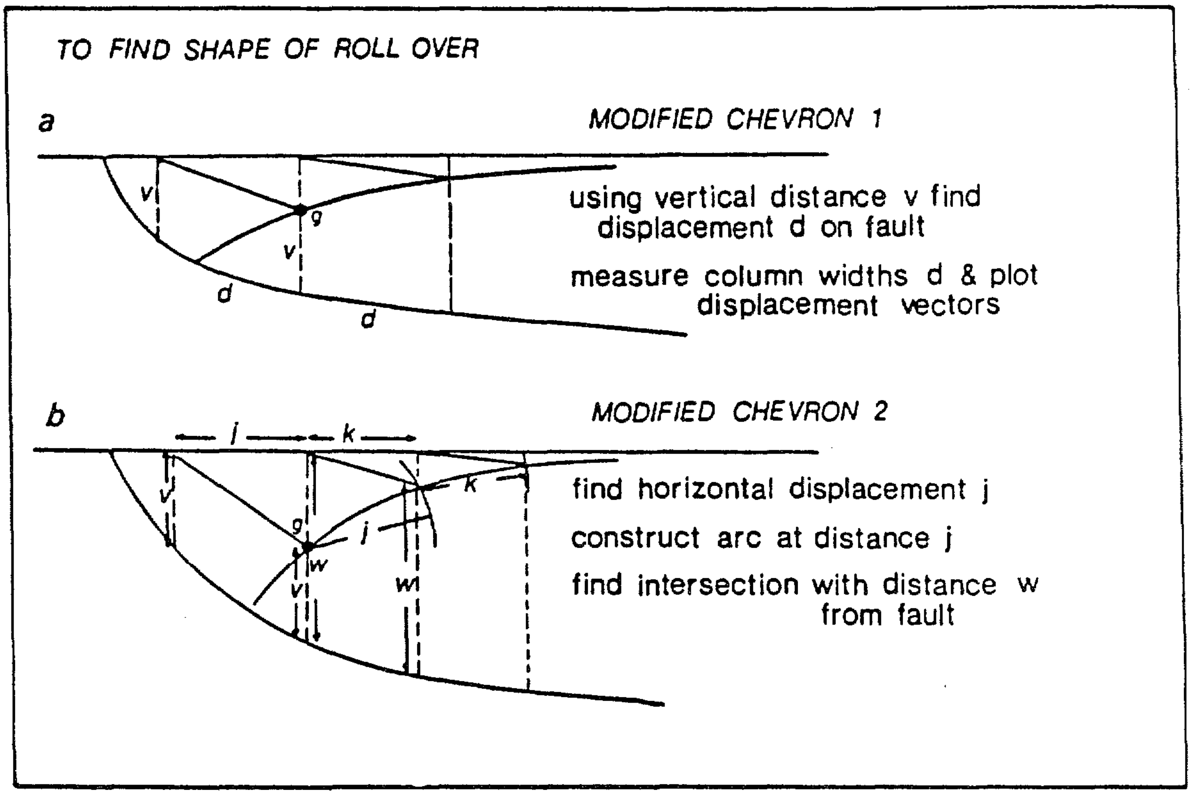

a
u
u'
u"
strain = u'/u
b
MODIFIED CHEVRON 1
CHEVRON
strain
h=1
h=3
h=3
h=1
distance along roll-over
MODIFIED CHEVRON 2
h=
u
h'
h"
u
fault
assuming
Chevron
construction
to find fault shape
1 measure distance u (= h) along roll over
2 draw displacement vector
3 draw fault parallel to vector

Fig 3


TO FIND SHAPE OF ROLL OVER
a
MODIFIED CHEVRON 1
v
g
v
d
d
using vertical distance v find displacement d on fault
measure column widths d & plot displacement vectors
b
MODIFIED CHEVRON 2
j
k
v
g
j
w
v
w
k
find horizontal displacement j
construct arc at distance j
find intersection with distance w from fault

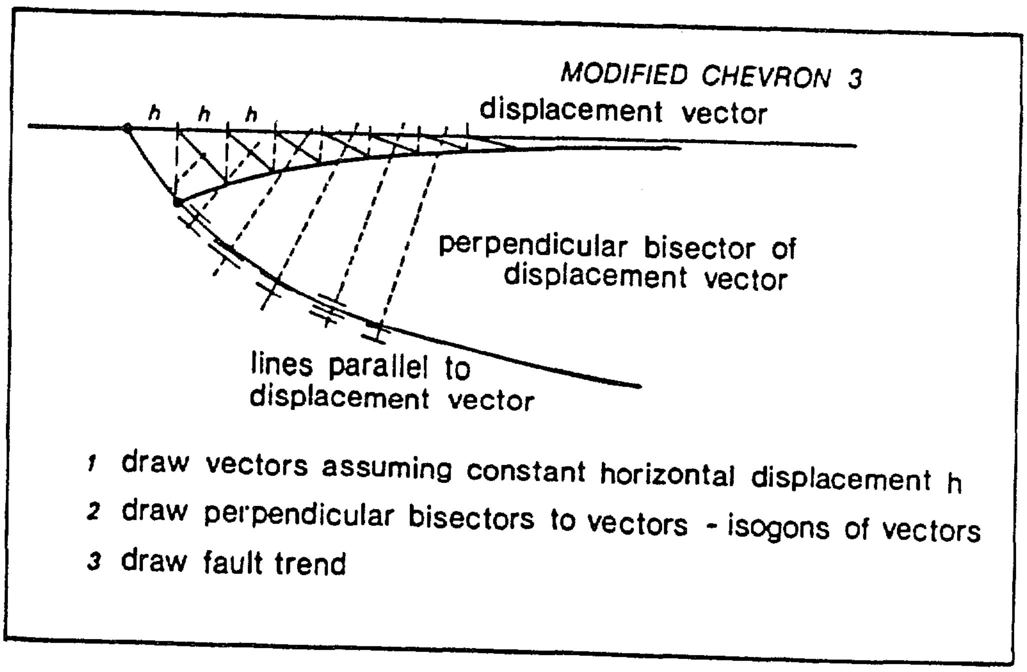
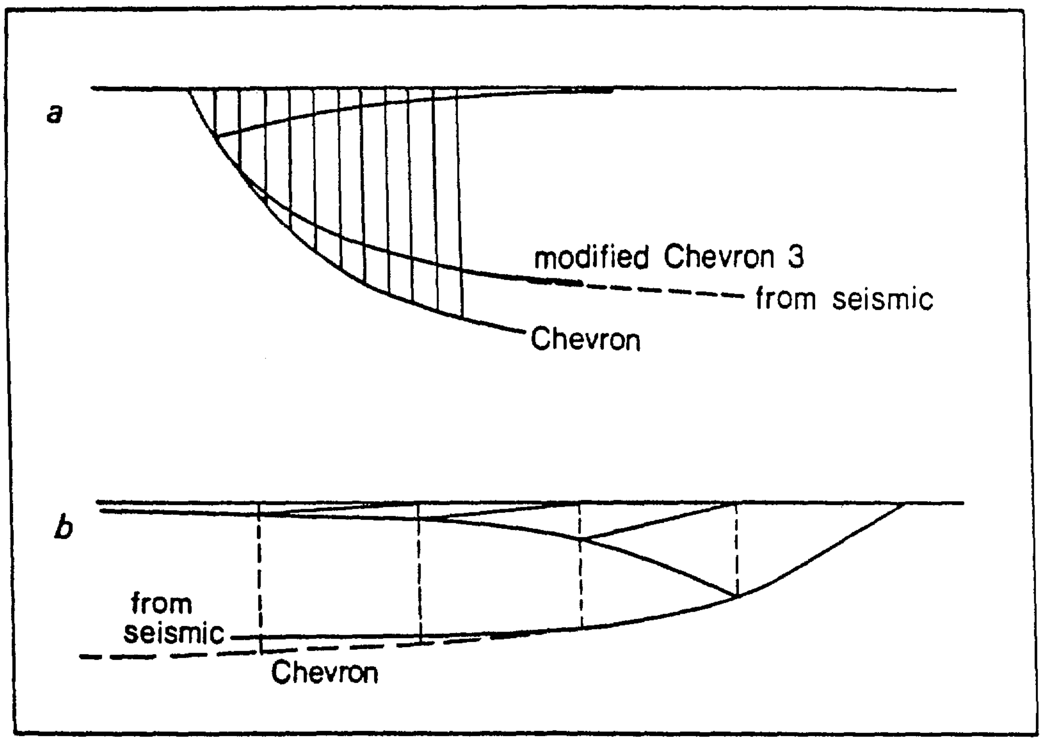

MODIFIED CHEVRON 3
h h h
displacement vector
perpendicular bisector of displacement vector
lines parallel to displacement vector
1 draw vectors assuming constant horizontal displacement h
2 draw perpendicular bisectors to vectors - isogons of vectors
3 draw fault trend

Fig 4


a
modified Chevron 3
from seismic
Chevron
b
from
seismic
Chevron

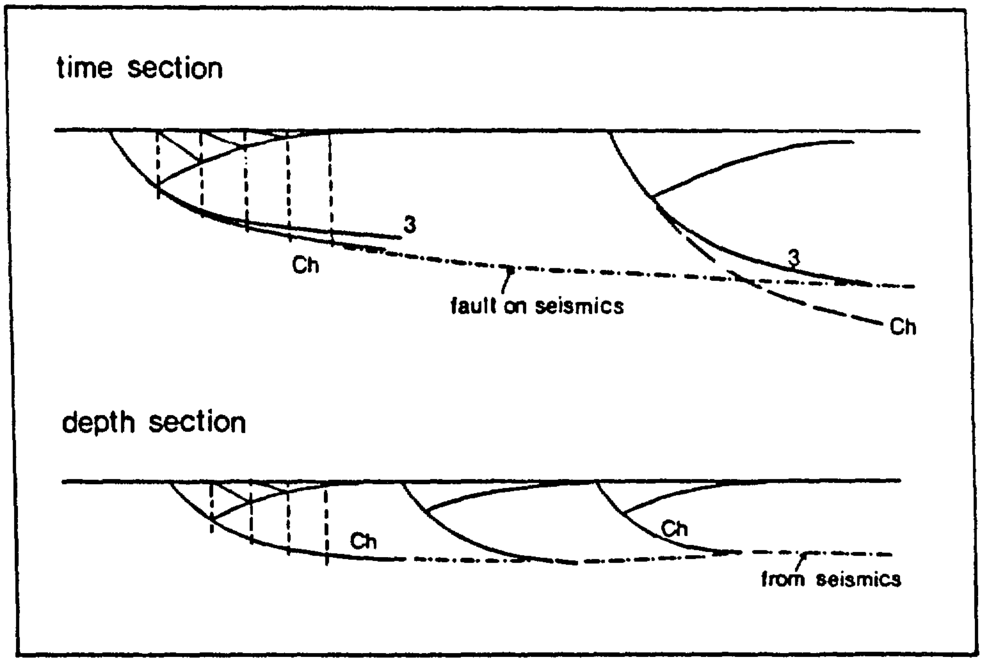
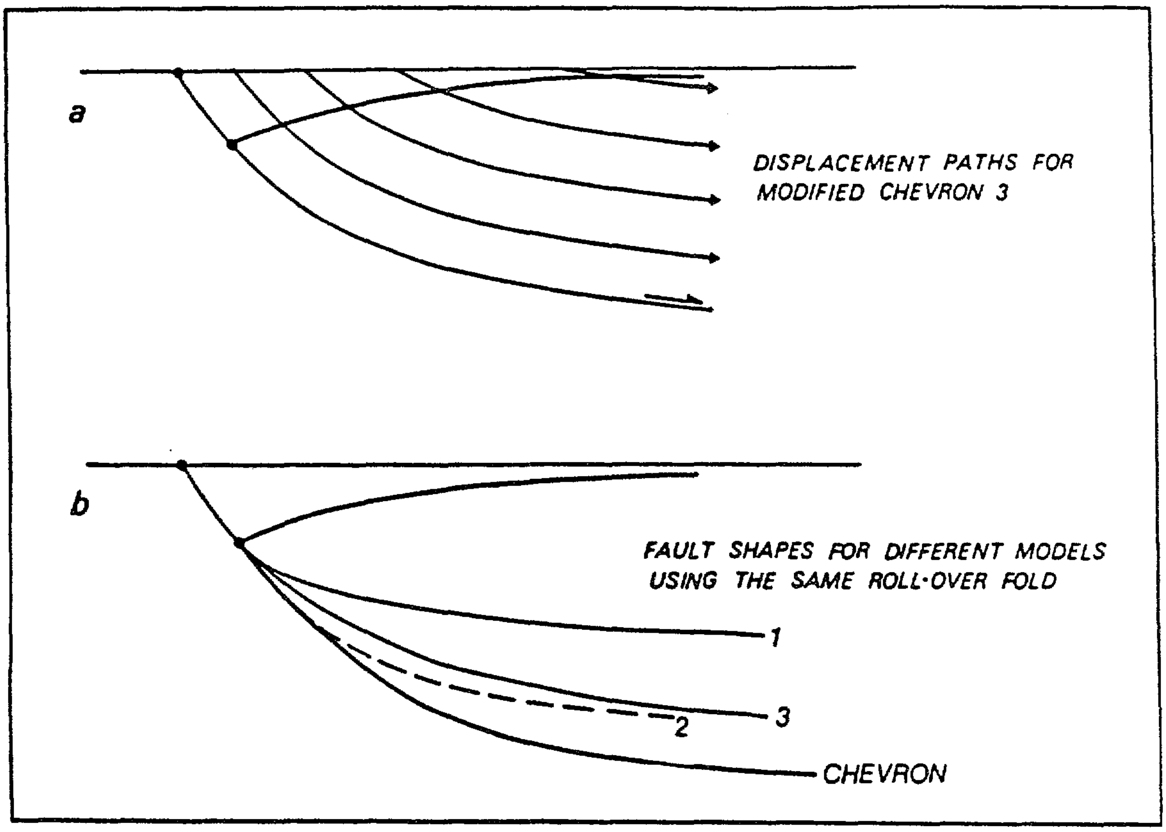

time section
3
Ch
fault on seismics
3
Ch
depth section
Ch
Ch
from seismics

Fig 5


a
DISPLACEMENT PATHS FOR MODIFIED CHEVRON 3
b
FAULT SHAPES FOR DIFFERENT MODELS USING THE SAME ROLL-OVER FOLD
1
3
2
CHEVRON

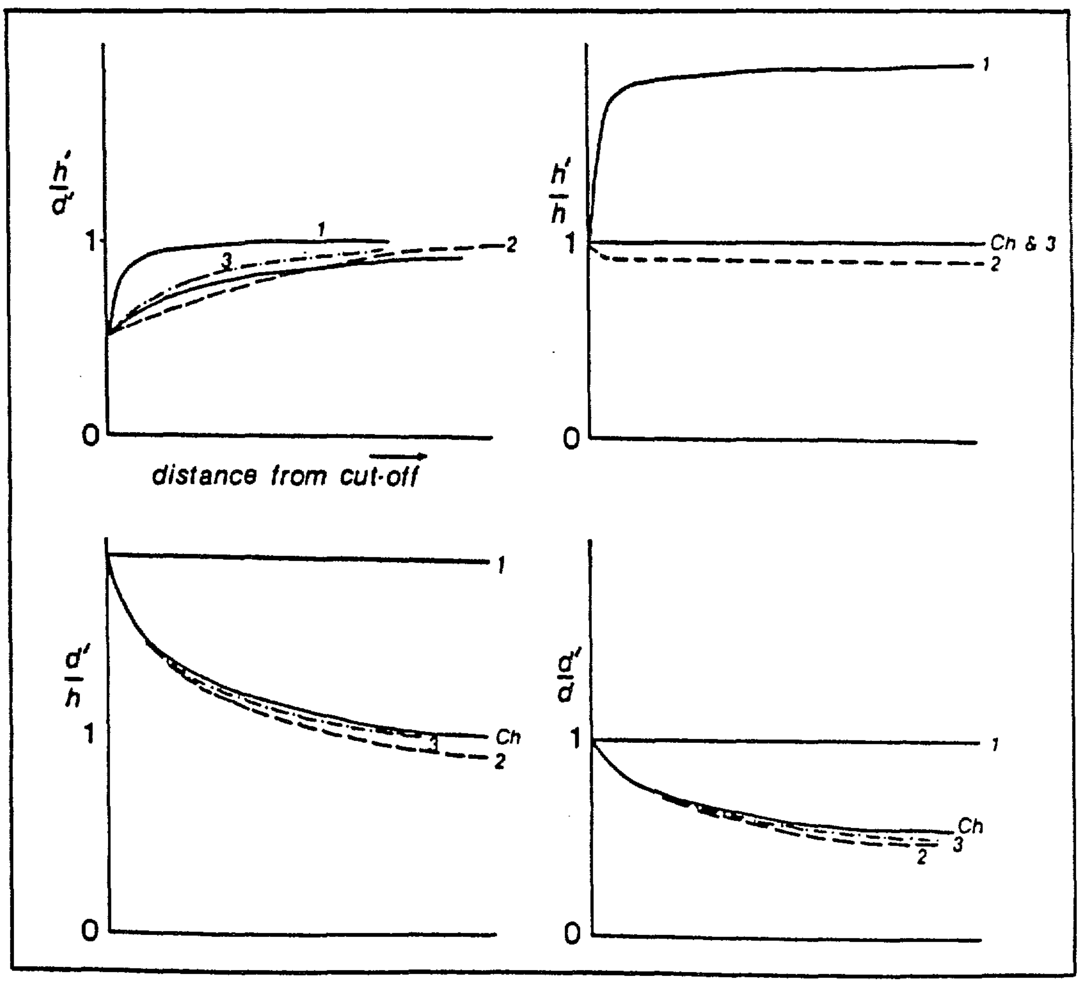
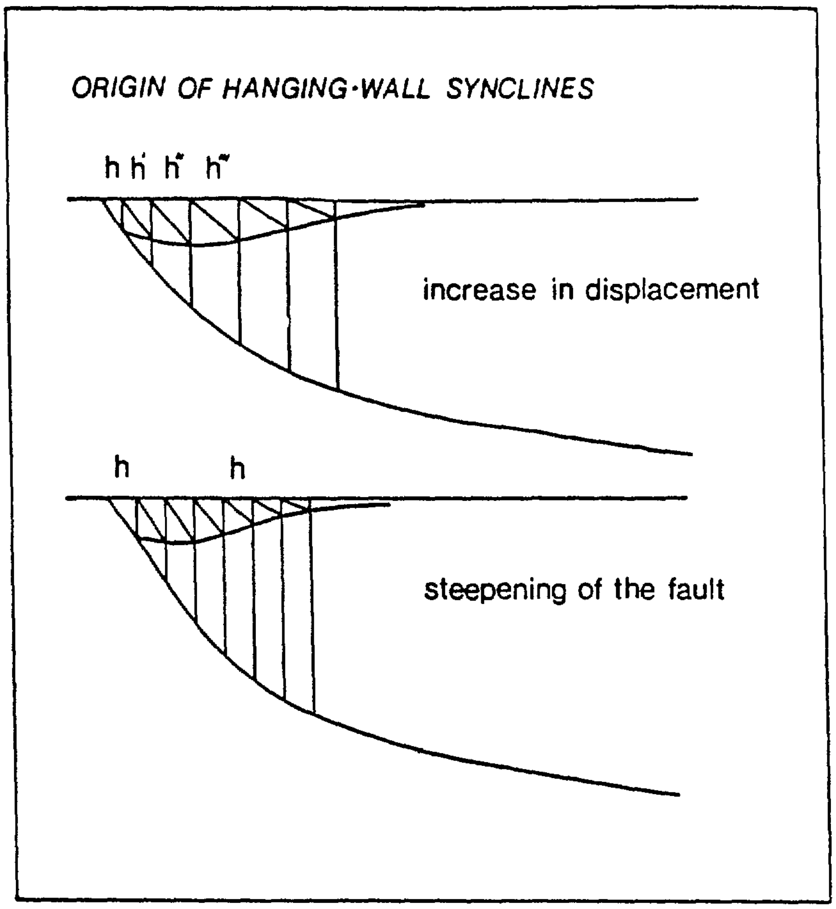

h'/d'
1
0
distance from cut-off
h'/h
Ch & 3
2
d'/h
Ch
d'/d
1
2
3

Fig 6


ORIGIN OF HANGING-WALL SYNCLINES
h h' h'' h'''
increase in displacement
h h
steepening of the fault

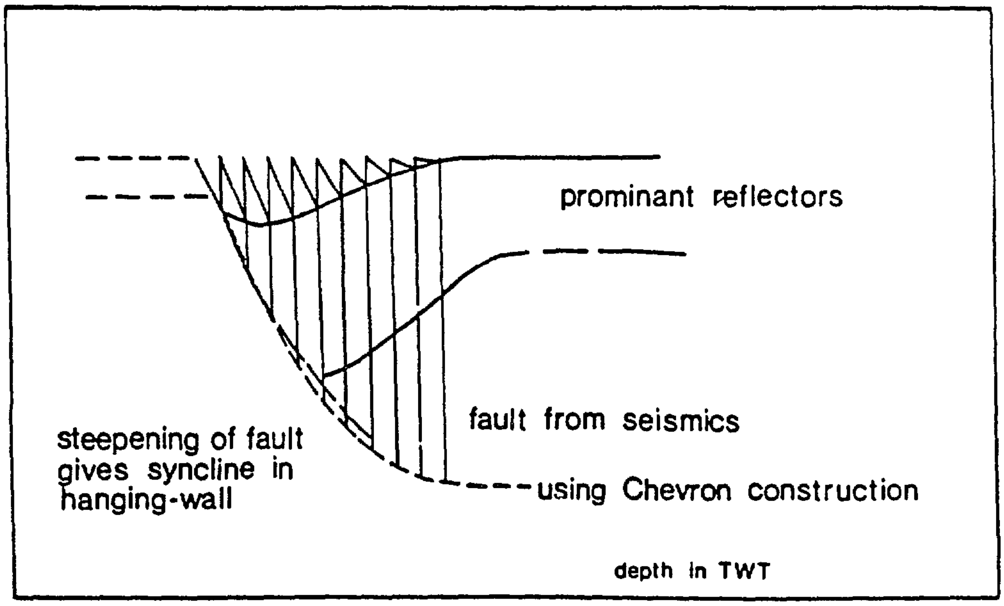
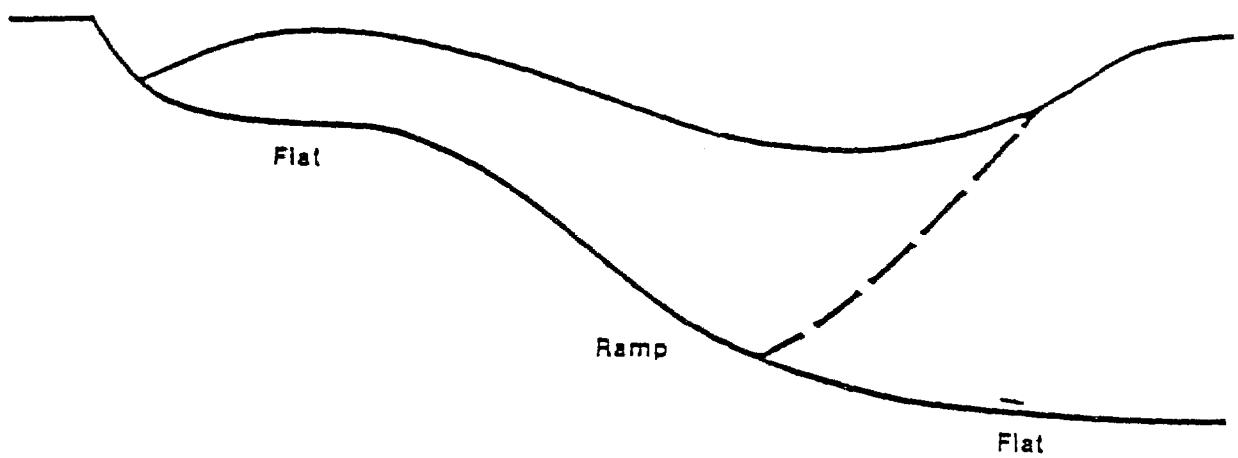
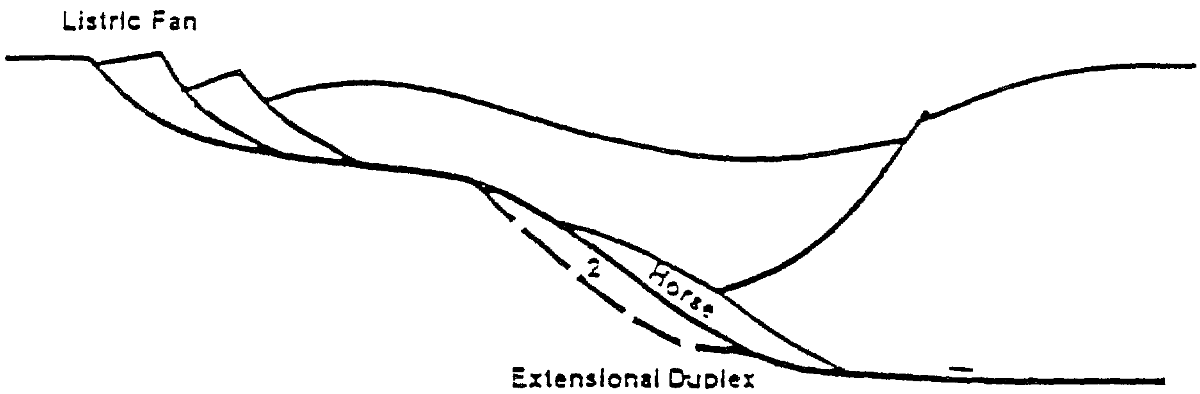

prominant reflectors
fault from seismics
steepening of fault gives syncline in hanging-wall
using Chevron construction
depth in TWT

Fig 7

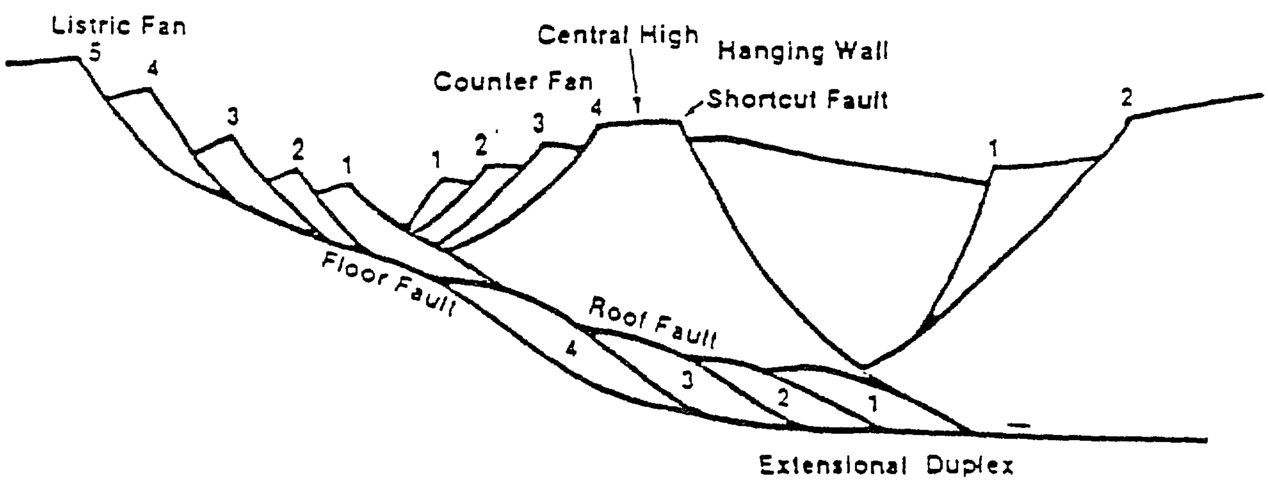
Fig 8



(a) Listric fault with flat and ramp on footwall with necessary hanging wall anticline and syncline. Antithetic fault developing on hanging wall in response to ramp.



(b) Listric fault with listric fan developed on traling edge of hanging wall and migration of footwall fault on the ramp to form a duplex zone with extensional horse.



(c) Fully developed system with listric fan and counter fan, extensional duplex with a central high developed above ramp by formation of a short cut fualt on hanging wall

Breached extensional fault array.

Note overturning turning of early extension faults.

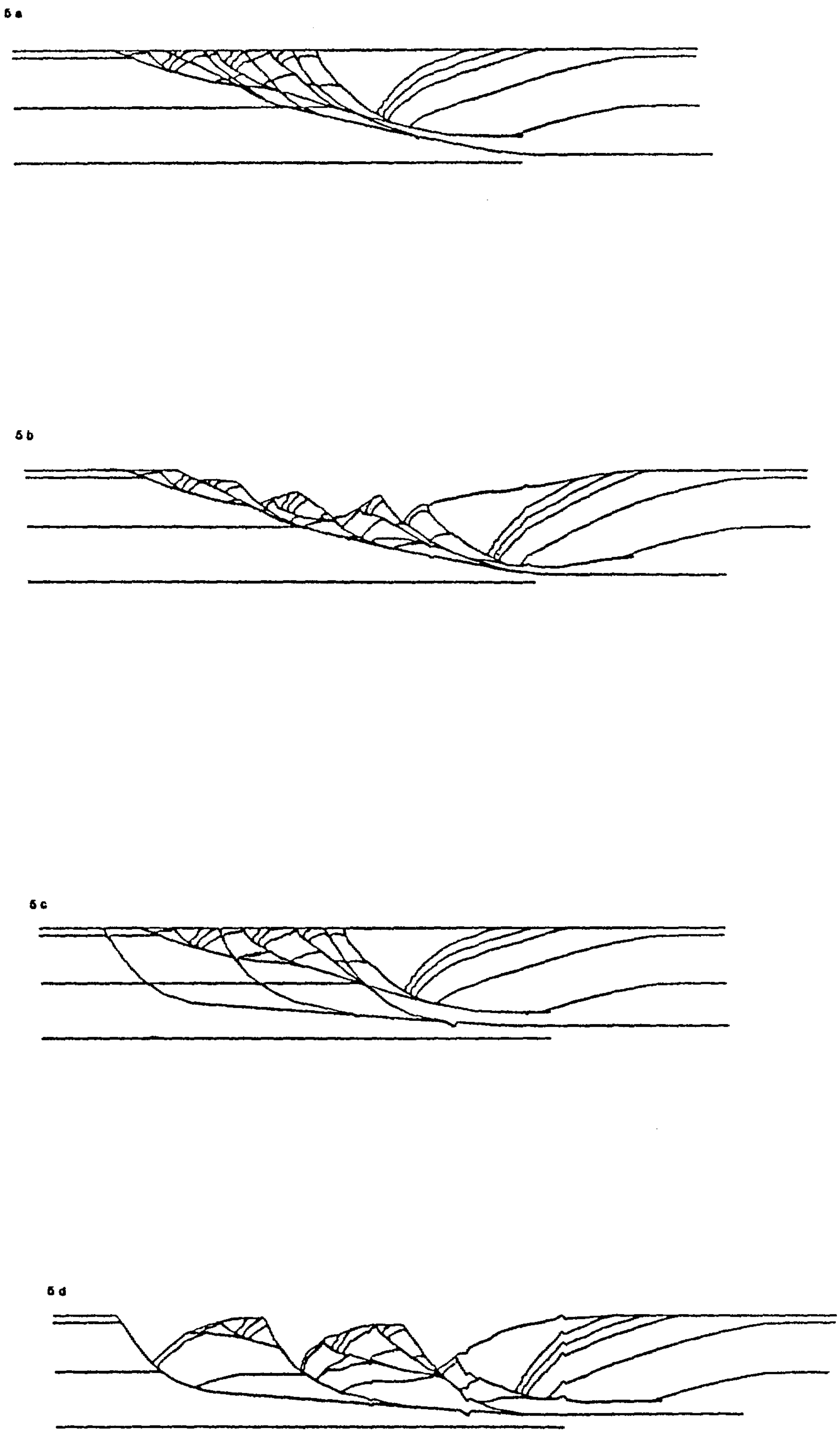


Fig 9

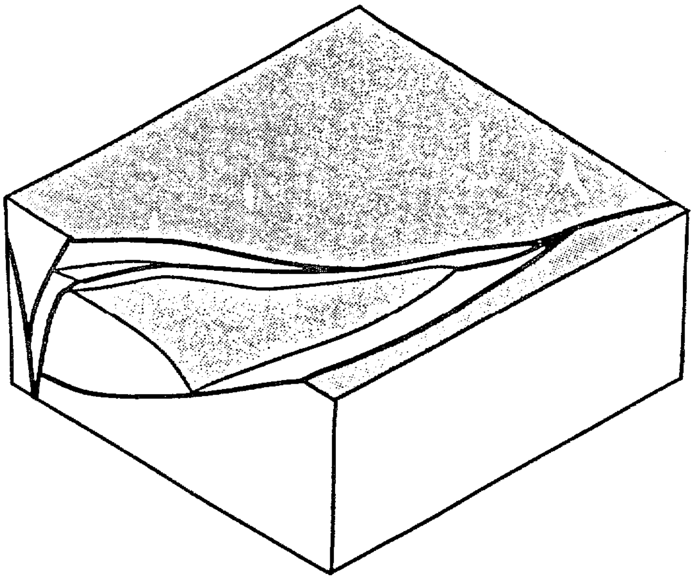
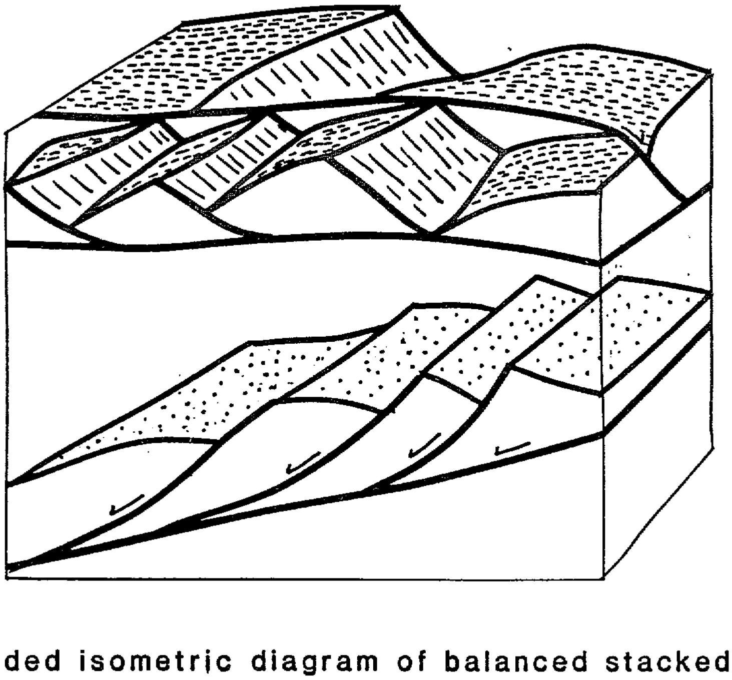


Fig 10

Linked detachment of oblique slip and strike slip bounded compartment. The strike slip fault forms a sidewall to the extensional compartment



Exploded isometric diagram of balanced stacked detachments. The upper leaf has a single transfer and overlies the lower leaf with opposite polarity. A shale or salt horizon decouples the two leaves.

Fig 11

Structural geometry in extensional terranes : transfer faults

M A Etheridge, BMR

Extension of the upper part of the continental lithosphere takes place largely on two sets of normal faults; large area, non-rotational faults with low initial dip (detachment faults), and smaller rotational normal faults that commonly sole into the detachment (listric or domino faults). Variations in the along-strike geometry of both fault sets result in a third set of accommodation structures known as transfer faults (Bally, 1981; Gibbs, 1984, 1987; Etheridge & others, 1985; 1987). Transfer faults are kinematically analogous to oceanic transform faults, but their geometry and relationship to the normal faults may be substantially more complex, especially where extension is significantly oblique. The foregoing paper by Alan Gibbs highlights the geometry of extensional normal faults, and this paper concentrates on the role of transfer faults both during and after extension.

Transfer faults are essentially accommodation structures, which allow along-strike variations in the position, spacing and/or dip of the normal and detachment faults. Their geometry is therefore largely determined by the geometry and displacements of the normal faults that terminate against them. They do not extend horizontally or vertically beyond the outer terminating normal or detachment faults. Strain compatibility requires that normal and transfer faults intersect in a common movement vector. Where the normal faults are close to pure dip-slip, the transfer faults will be sub-vertical and perpendicular to them (Fig. 1; e.g., the Bass Strait basins, Etheridge & others, 1985, 1987). However, where there is significant oblique slip on the normal faults, the transfer faults will also be oblique slip and the fault traces will not be orthogonal (Fig. 2). The latter geometry is typical of regions of strongly oblique extension, where the distinction between normal and transfer faults may break down, resulting in two similar sets of oblique slip normal faults (e.g., parts of the African Rift System, Rosendahl, 1987; Rosendahl & others, 1986; the Moray Graben, Gibbs, 1984).

Transfer faults operate on a wide range of scales. They accommodate 180° switches in the dip of the master detachment fault, and therefore in the gross structural asymmetry of extensional terranes. Flips in asymmetry with a length scale of several hundred kilometres have been interpreted for the Atlantic margin (Lister & others, in press). Switches in normal fault dip and half-graben facing are common to many of the world's continental rifts, generally with a length scale of about 100 km (Fig. 3; Bass Strait basins, Etheridge & others, 1987; African Rift System, Rosendahl & others, 1986, this symposium; Canning Basin, Drummond & others, in press, this symposium; Mohave Desert region, California, Dokka, 1986). Individual rift basins may terminate at a transfer fault (e.g., southeastern end of the Bass Basin), although the extension will generally be transferred to an adjacent basin. On a smaller scale, one to ten kilometre offsets of the rift-bounding (and intra-rift) normal faults may be widespread (e.g., Fig. 4, southern margin of the Gippsland Basin, Etheridge & others, 1985), giving rise to apparently curved to sinuous fault traces where the transfer faults have not been recognised.

Because they are accommodation structures, transfer faults have movement patterns that are dictated by the geometries of the normal faults that abut them. In particular, they are not simple strike-slip faults. They will generally be imaged on reflection seismic profiles as moderately to very-steeply dipping normal faults, but their amount and even apparent sense of displacement may change rapidly along strike (Fig. 5). Where seismic profiles are oblique to, and therefore intersect, both normal and transfer fault sets, geometries may result that are very difficult to interpret unless the general three-dimensional fault geometry is known (Fig. 6).

Transfer faults generally form scarps, and can therefore influence syn-rift sedimentary facies development in a fashion similar to the classical rift-bounding normal faults. They can (1) be primary sources of clastic input to the half-graben, with alluvial fans and similar clastic aprons that could form petroleum reservoirs (e.g., Hibernia oil field, Scotian Shelf, Tankard & Welsink, in press; Fig. 7); (2) provide 'along-strike' barriers to half-graben, giving rise to restricted lacustrine environments ideal for deposition of hydrocarbon-source facies (Powell & Bradshaw, this symposium); and (3) exert a substantial influence on sediment transport geometry, especially in the terrestrial environment that typifies active rifts (e.g., African rift lakes, Rosendahl & others, 1986).

Even after extension ceases, steeply-dipping transfer faults are ideally oriented to take up differential thermal subsidence along the axis of the rift, and therefore to act as growth-faults influencing facies development (albeit in a more subtle fashion than during extension) throughout the whole life of the basin. This will be true particularly of transfer faults that accommodate changes in detachment geometry related to variations in the location and/or distribution of the thermal anomaly produced by mantle thinning. As long-lived zones of movement, transfer-fault zones are also likely to have high permeability, and therefore to be preferred fluid-flow paths. The focussing of fluid-flow will have important implications for both hydrocarbon migration and the localisation of many types of hydrothermal mineralisation (e.g., MVT Pb-Zn sulphide, Mount Isa/McArthur River - type Ag-Pb-Zn mineralisation, some types of stratabound Cu deposit, and hydrothermal precious metals in extensional terranes). Deep-seated, vertical transfer-fault zones are also potential sites of igneous intrusion/extrusion (Oversby, this symposium).

Because transfer faults are commonly vertical zones of weakness with depth ranges of several kilometres to in excess of 100 km (the base of the lithosphere), they are readily reactivated in a wide range of post-extensional stress fields (Etheridge, 1986). In particular, compression oblique to the original extension direction is likely to induce strike-slip reactivation of the transfer faults, giving rise to the classical wrench structures (en echelon anticlines, positive and negative flower structures; e.g., Gippsland Basin, Etheridge & others, 1985; Canning Basin, Begg, 1986; Drummond & others, this symposium; Miocene structuring in Carnarvon Basin, WA). Many orogenic terranes originate as extensional basins, and the extensional segmentation on major transfer faults can be preserved through subsequent compressional orogeny. For example, tear faults which segment thrust-belts may have originated on pre-existing transfer faults. Important lineaments that originated as transfer or strike-slip faults may be repeatedly reactivated during cycles of extension and compression. Transfer faults formed on passive continental margins may propagate on to the craton under suitable conditions of reactivation, to form subtle topographic lineaments that may extend for hundreds of kilometres. The Murray-Darling topographic lineament is interpreted to have formed by propagation of the major transfer fault (now an ocean - continent transform) at the head of the Tasman Sea, as a result of the east-west compression that dominates the present stress-field in eastern Australia.

Finally it is emphasised that transfer faults or similar accommodation structures are increasingly being recognised as characteristic features of extensional terranes. They have been recognised in most Mesozoic and Cainozoic continental rifts, and on many passive margin segments. They rank equally with the better known faults in terms of their contribution to rift geometry, depositional facies evolution, structural development and, most importantly, control of both petroleum and mineral resource potential.

References

Bally, A.W., 1981 - Atlantic-type margins. In Geology of passive continental margins. American Association of Petroleum Geologists, Education Course Note Series, 19, 1/1-48.

Begg, J., 1986 - Structuring and controls on Devonian reef development on the north-west Barbwire and adjacent terraces, Canning Basin. Australian Petroleum Exploration Association Journal, 27, 137-151.

Dokka, R.K., 1986 - Patterns and modes of Early Miocene crustal extension, central Mohave Desert, California. Geological Society of America Special Paper, 208, 75-95.

Drummond, B.J., Etheridge, M.A., Middleton, M.F., & Davies, P.J., - Half graben model for the structural evolution of the Fitzroy Trough, Canning Basin. Australian Petroleum Exploration Journal (in press).

Etheridge, M.A., 1986 - On the reactivation of extensional fault systems. Philosophical Transactions of Royal Society of London, A317, 179-194.

Etheridge, M.A., Branson, J.C., & Stuart-Smith, P.G., 1985 - Extensional basin-forming structures in Bass Strait and their importance for hydrocarbon exploration. Australian Petroleum Exploration Association Journal, 25, 344-361.

Etheridge, M.A., Branson, J.C., & Stuart-Smith, P.G., 1987 - The Bass, Gippsland and Otway Basins, southeast Australia: a branched rift system formed by continental extension. In Beaumont, C., & Tankard, A., (Editors) - Sedimentary basins and basin-forming mechanisms. Canadian Society of Petroleum Geologists Memoir, 12.

Gibbs, A.D., 1984 - Structural evolution of extensional basin margins. Journal of the Geological Society of London, 141, 609-620.

Gibbs, A.D., 1987 - Development of extension and mixed-mode sedimentary basins. In Coward, M.P., Dewey, J.F., & Hancock, P.L., (Editors) - Continental extension tectonics. Geological Society Special Publication, 28, 19-33.

Lister, G.S., Etheridge, M.A., & Symonds, P.A., 1986 - Application of the detachment fault model to the formation of passive continental margins. Geology, 14, 246-250.

Lister, G.S., Etheridge, M.A., & Symonds, P.A., - Detachment model for the formation of passive continental margins. Tectonics (in press).

Rosendahl, B., 1987 - Architecture of continental rifts with special reference to east Africa. Earth plan. Sci. Lett., 15.

Rosendahl, B., Reynolds, D., Lorber, P., Burgess, C., McGill, J., Scott, D., Lambiase, J., & Derksen, S., 1986. In Frostick, L., (Editor) - Sedimentation in the East African rifts. Geological Society of London Special Publication, 25, 29-43.

Tankard, A.J., & Welsink, H.J., - Extensional tectonics and stratigraphy of the Hibernia oil field, Grand Banks of Newfoundland. J. American Association of Petroleum Geologists, (in press).

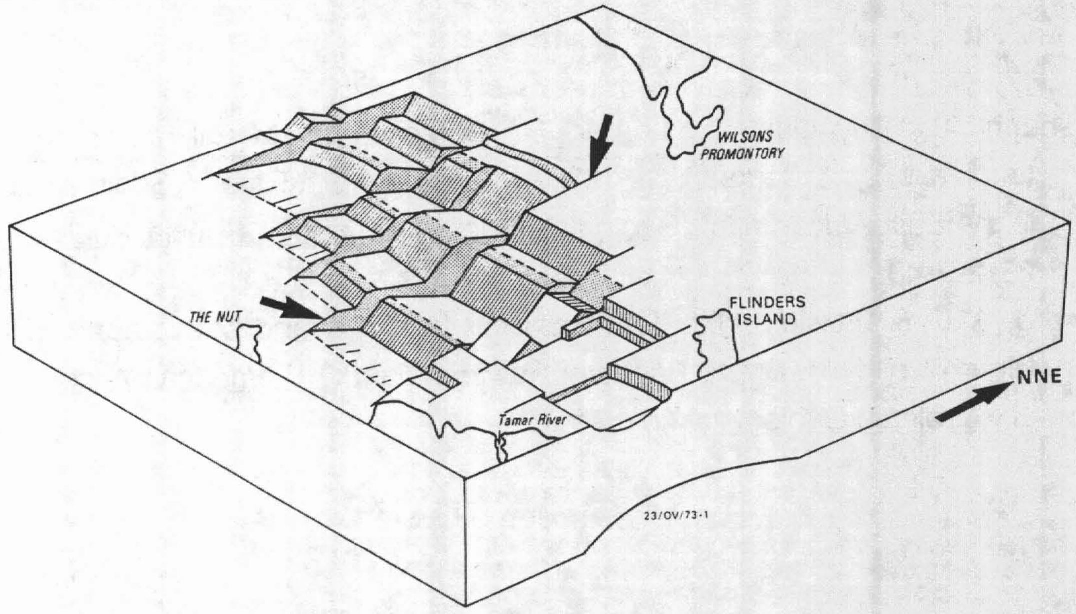


Figure 1: Schematic projection of orthogonal normal and transfer faults geometry in the Bass Basin, showing only the major transfer faults; the arrowed transfer fault accommodates a significant change in the spacing between the main normal faults (after Etheridge & others, 1984).

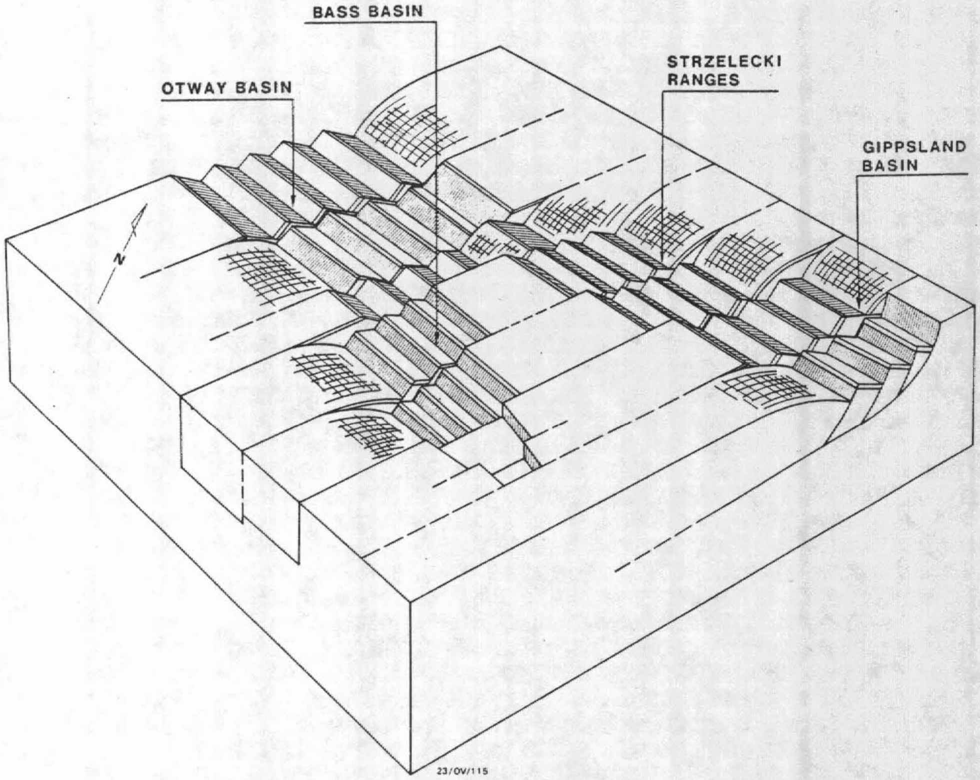


Figure 2: Schematic representation of the linkage of the major Bass Strait basins by a transfer fault array; note the transfer of extension from one basin to another (e.g., southeastern end of Bass Basin to Gippsland Basin), and the switches in polarity or asymmetry of tilt blocks and associated half-graben (e.g., within Gippsland Basin), (after Etheridge & others, in press).

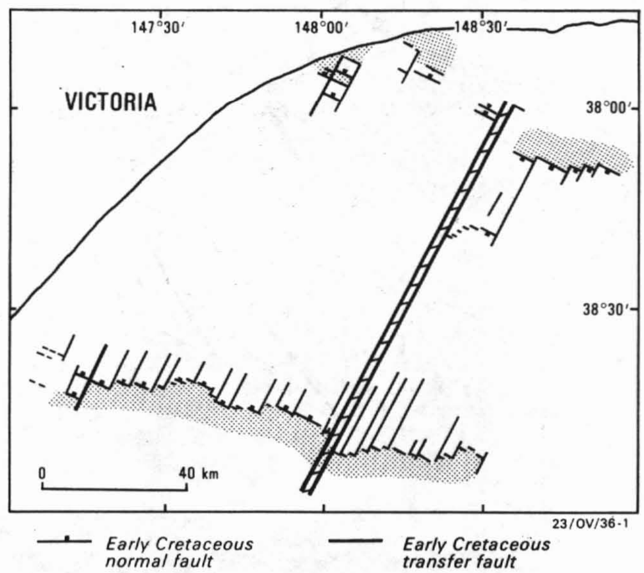


Figure 3: Mapping of offsets in the basin-bounding normal faults and/or hinge boundaries of the Gippsland Basin revealed the location of minor transfer faults; the hachured structure trending north-northeast across the basin is the major transfer fault that accommodates a switch in structural symmetry (after Etheridge & others, 1985).

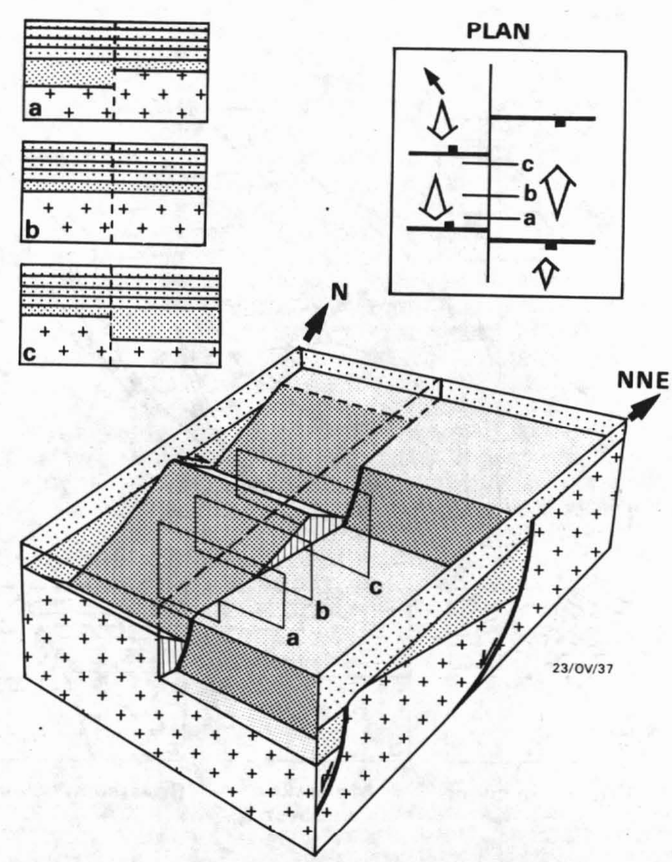


Figure 4: Because tilt blocks and half-graben are offset across transfer faults, displacement patterns of basement and/or syn-rift sediments may be complex, making mapping of transfer faults difficult. Transfer faults are hard to recognise in seismic section unless the pre-rift basement surface is clearly visible. They are best imaged on strike-lines (after Etheridge & others, 1985).

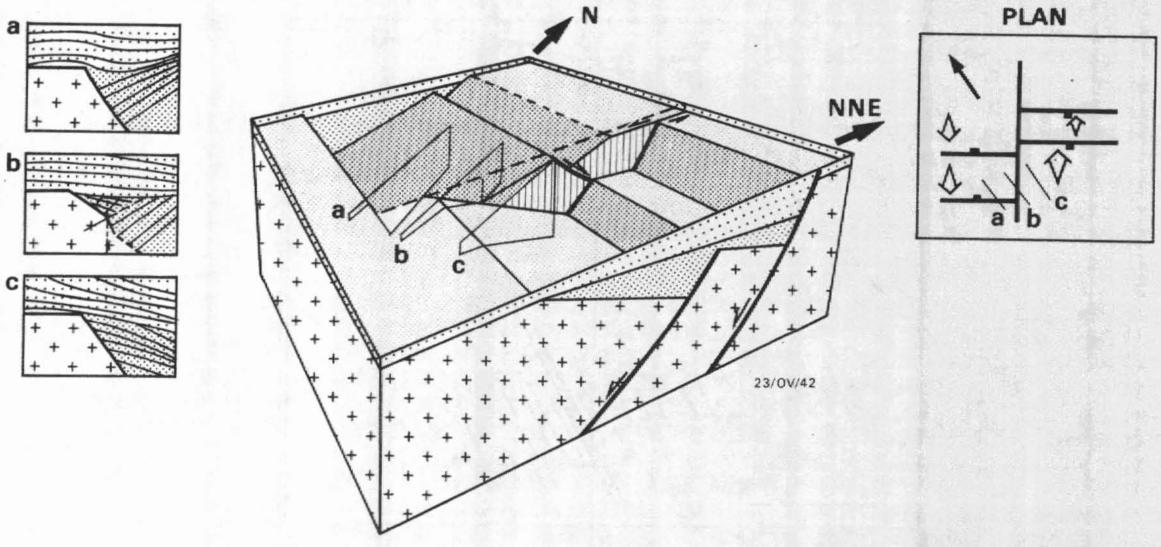


Figure 5: Seismic sections oblique to the normal- and transfer-fault trends will commonly cross both fault sets, potentially giving rise to complex geometries at the syn-rift and basement levels. This example is based on sections across the major polarity-switching transfer fault in the Gippsland Basin (after Etheridge & others, 1985).

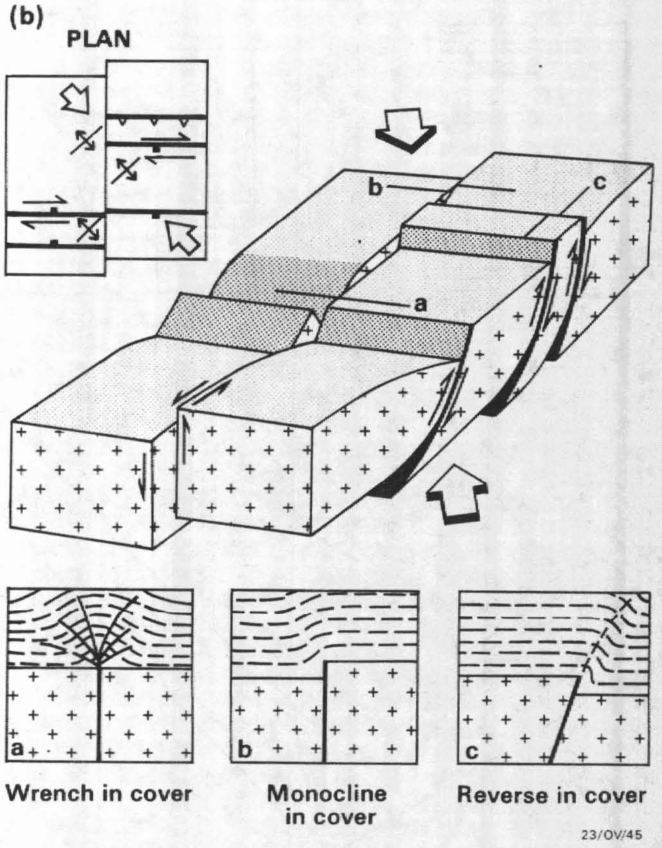


Figure 6: Transfer faults are readily reactivated by subsequent changes in the stress orientation; this sketch illustrates the styles of reactivation induced by compression oblique to the transfer fault direction. Example of this style of reactivation are seen in the Gippsland and Canning Basins, and may be present throughout the Carnarvon Basin (after Etheridge & others, 1985).

Architecture of African rifts with special reference to passive margins

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The East African Rift is composed of zones of linked half-graben. The key to interpreting seismic data from such areas is recognising that linking geometries have a direct bearing on rift-basin morphology. Dip profiles across overlapping, opposing half-graben show two half-grabens that face each other. This creates an apparent full-graben with some sort of complex axial structure, usually antiformal in character. Such structures have been termed "interference accommodation zones" (INAZ) because they accommodate the subsidence of the facing half-graben. In some cases the accommodation can involve transverse, strike-slip, and even compressional tectonics, although the overall stress regime is extensional. Depending upon the geometry, INAZ can be very large features and they often terminate in platforms, which serve as the access routes for fluvial clastics entering rift basins with this type of geometry. Dip profiles across non-overlapping, opposing half-grabens show two half-grabens facing away from each other. They are usually separated by backbones of relatively unsubsided country rock, termed "isolation accommodation zones" (ISAZ). These can be sites of considerable strike-slip faulting. Unlike INAZ which usually are not modern-day barriers to fluvial sediment dispersal, ISAZ often act as nearly-complete dams throughout the active rifting phase. Fluvial clastic input to half-grabens linked in this fashion is mainly from the shoaling or ramping sides. Although it is common for one sense of polarity to dominate in African rift zones, there is never an exclusive direction of asymmetry and usually no more than two adjacent half-graben display the same polarities. This contrasts with the Triassic rift basins of the eastern United States, where all but a few half-grabens have their border fault systems on the west. The difference probably relates to reoccupation of Palæozoic thrust planes in the Triassic case, and lack of any systematic low-angle thrusts in the old, cold, and brittle crust of pre-rifted east Africa. The areas between similar polarity half-grabens are often platforms, and fluvial clastics may enter these half-grabens across such platforms, as well as from their shoaling sides.

These concepts should prove very useful to the petroleum exploration industry. For example, the style of linking determines the loci of fluvial clastic input, which bear on the distribution of reservoir rocks. Also, the nature of interference accommodation zones is dynamic - these features evolve with the adjacent depocentres, usually remaining high relative to the depocentres. The stratigraphic and hydraulic consequences of this, combined with the scale of these features and the fact that they merge into platforms, make them a primary target for petroleum exploration in rifts. The Agua Grande Field in the Reconcavo Basin is a possible example. On the other side of the coin are the isolation accommodation zones, which usually are poor synrift exploration targets.

The PROBE rifting concepts also offer new insight into the progression from continental rift to juvenile ocean basin to passive continental margin. For example, arrangements of facing half-grabens, hence INAZ, seem to be a necessary step toward successful rifting, which probably begins with juvenile half-graben expressions and ends with the success of one spreading centre and abandonment of one or more others. A related application is the realisation that symmetric conjugate rift margins are an unlikely outcome of ripping the Tanganyika or Malawi rifts apart. Instead, asymmetry should rule. Probably most of the rifted terrain is left on only one of the margins, in a pattern that alternates crudely between the opposing margins along strike.

Active extension in the D'Entrecasteaux Islands, Papua New Guinea

E J Hill, Monash University

Introduction

Immense domes of high-grade metamorphic rocks are emerging from the sea to form mountainous islands, over 2 000 m high, in an area of active rifting on the Solomon Sea Plate (Fig. 1). The D'Entrecasteaux Islands, offshore eastern Papua New Guinea, are made-up of a number of broad, partially fault-bounded domes of gneissic and mylonitic rock (Figs 2 and 3); they show remarkable similarity in characteristics to the metamorphic core complexes of the United States Cordilleran region.

These domes were originally described by Davies & Ives (1965). Ollier & Pain (1980) suggested that doming and shearing was caused by forceful granite intrusion, while Masson (1984) concluded that the domes were fold interference patterns. Senior & Billington (1987) undertook a detailed photogeological survey of part of one of the domes. The domes began uplifting in the Late Pliocene (Davies, Symonds & Ripper, 1984) and uplift is believed to be continuing at present. Uplift has been associated with abundant igneous activity including intrusion of large granodiorite bodies in the cores of the domes, calc-alkaline and peralkaline volcanism associated with faulting, and current hydrothermal activity.

It is the intention of this presentation to describe the structure of the region, its regional tectonic setting, and to establish a relative geochronology. Detailed field work has been carried out only on the northern side of Fergusson Island and in a small area on Goodenough Island.

Tectonic setting

The D'Entrecasteaux Islands lie on an active seismic lineation which is believed to be the western extension of the actively-formed Woodlark Basin, and east-west - trending seafloor spreading system (Ripper, 1982). The basin lies in the complex collision zone between the northward-moving Indo-Australian plate and the northeastward moving Pacific plate. The westward propagation of the Woodlark spreading system (which began in the Pliocene) into a region of continental crust, has resulted in a complex seafloor structure to the east of the

D'Entrecasteaux Islands. There have been two main areas of seafloor spreading; a currently heavily-sedimented, failed southern arm, and an active northern arm with evidence of hydrothermal venting (Binns & others, 1987). Further propagation of the spreading system has caused formation of three fault-bounded depressions: Milne Bay, Mullins Harbour, and the seismically-active Goodenough Bay (to the west of the D'Entrecasteaux).

Peralkaline silicic volcanics occur on eastern Fergusson Island. They have been erupted during two phases of activity, the later from three north-south - aligned volcanic centres. This volcanic type is characteristically associated with extensional areas, typically in major continental rift structures, and appear to be related to the rifting of the Woodlark Basin (Smith, 1976).

All available evidence suggests that the D'Entrecasteaux Islands are located in an area of active continental extension.

Fergusson Island: Geology

Fergusson Island is centred by a large igneous body, the Omara Granodiorite, which is surrounded in part by irregularly-deformed metamorphics - mostly gneiss and amphibolite - and capped by a shell (200-300 m thick) of strongly sheared and mylonitised metamorphics, which define two large domes: The Mailolo dome in the northwest and the Oiatabu-Morima arc in the southeast. They are believed to be faulted-off parts of a single original dome, the Omara dome (Davies & Ives, 1965; Davies, 1973).

The irregular metamorphics of the core consist dominantly of quartzo-feldspathic gneisses, calcareous gneisses, migmatites, eclogites, marble and granite. They are believed to be high-grade equivalents of Mesozoic sialic rocks of the Owen Stanley Range. The shell consists of retrogressed, sheared and mylonitised equivalents of these rocks. The basement rocks contain minor intrusions of granite and gabbro as well as abundant granitic veins, pegmatites and dolerite dykes. Structural features are summarised in Figure 4.

Around the edge of the dome the mylonitic foliation dips between 15^{o} and 30^{o} and is parallel to the surface of the dome. Towards the centre of the dome the foliation has been broadly folded into a number of irregular antiforms and synforms which may also fold the dome-bounding fault in some places (e.g., around Wapolu Point). Isoclinal intrafolial folds are associated with the mylonitic foliation. Low angle ($10^o - 40^o$), undulating shear zones cut through the foliation but are broadly parallel to it; they are associated with tight asymmetric folding. High-angle normal shears (generally greater than 45^o) commonly occur in conjugate sets and in some areas are intruded by sheared granitic veins. All fold hinges are parallel to a strong mineral lineation which, over the northern half of Fergusson Island, trends consistently east-northeast, with the only exception in the vicinity of the Oredi Fault, northeast Fergusson Island, where it trends northwards.

Hanging-wall rocks of the dome-bounding faults are poorly outcropping and consists of largely serpentinised, altered ultramafic rocks. Ultramafics also occur as isolated patches on the hanging-wall blocks of the north to northwest-trending faults, and as xenoliths within the granodiorite core. The ultramafics are believed to be part of the Papuan Ultramafic Belt which was thrust over the Owen Stanley Metamorphics during the Lower Eocene.

Two main types of low-angle faults cross-cut the islands; they dip around 15^o to 30^o outwards from the centres of the domes. Northwest-trending faults with dominantly normal sense of movement bound the northern and eastern edges of the domes; and north to northeast-trending faults with considerable strike-slip motion, as well as normal sense of movement, have caused displacement of the dome-bounding faults and distortion of the domes. These faults generally coincide with an abrupt change in topography from highlands to narrow, flat coastal plains.

1. Dome Bounding faults (Mwadeia and Elologea Faults): Approaching the dome-bounding fault the basement becomes increasingly deformed by undulating biotite-rich shear zones which broadly follow the trend of the mylonitic foliation. Close to the fault these shear zones are overprinted by later shears containing biotite, chlorite and/or actinolite. In some areas there is a fault-bounded strip of schists and breccias around the dome-bounding fault known as the "transition zone". It forms an area of moderate relief between the high basement inland and the low, flat coastal strip. The schists commonly show one or more generations of crenulation development, and, where uncrenulated, the schists may show a distinct lineation which parallels the trend of the gneissic lineation. The transition zone consists of mixed slices of altered and brecciated ultramafics and basement rock. The main faults are marked by brecciated schists and amphibolite. Around the fault area the hanging-wall ultramafics are generally

brecciated and may be hydrothermally altered, causing strong silicification and pyritisation.

2. North and northwest-trending faults (Barrier Islands fault system including the Wabawe and Kwakwau Faults, the Lake Lavu Fault System, the Wadelei Fault, and the Oredi Fault): These faults commonly form as major fault-zones containing several faults. Where they intersect the dome-bounding faults they cause complicated faulting patterns, which appear to be associated with mineralisation. These faults are associated with major centres of calc-alkaline volcanism and thermal activity. Normal movement along the Wadelei Fault and Lake Lavu Fault System has caused the centre of the Omara dome to collapse exposing the Omara Granodiorite.

Goodenough Island: Geology

The Barrier Islands Fault System separates the Omara dome of Fergusson Island from the Goodenough dome. The geology of Goodenough Island is similar to that of Fergusson Island except that the mylonitic shear zone is thinner (less than 200 m), and the dome is smaller in diameter. That the larger Fergusson Island structure is not as high as the Goodenough dome is attributed to the later downward faulting of the centre of the Omara dome. The Luboda Granodiorite forms the core of Goodenough dome. The mylonitic foliation dips outward from the centre of the dome around $20^{o}-30^{o}$, and the ubiquitous mineral lineation has a trend consistent with that of Fergusson Island, i.e., towards the north-northeast. A distinct break in topography marks the dome-bounding fault (the Wakonai Fault) on the northeastern side of the island. The hanging-wall rocks are largely covered by alluvium except for a small outcrop of ultramafic rocks in the southeastern corner of the island.

Mineralisation and thermal activity

Mineralisation is structurally controlled and is related to both the dome-bounding faults and the north to northeast-trending fault systems. Gold mineralisation accompanied by strong silicification and variable pyritisation is generally found in shallowly-dipping zones in the hanging-wall rocks, i.e., in the transition zone rocks and in ultramafics. At the Wapolu prospect, northwestern Fergusson Island, it is considered that a hydrothermal system developed along the fault breccia and a "gravity sliding breccia" (McManus, 1986) within the transition zone;

strong mineralisation is confined to the hanging-wall rocks due to the poor permeability of the crystalline basement rocks.

Active hot-spring activity is confined also largely to areas associated with major faulting systems (e.g., around Wapolu Point) or volcanic activity (e.g., Deidei and Iamelele thermal areas). However small hot-springs commonly emanate from low-angle biotite shear zones in the basement gneisses.

Summary: igneous-tectonic history

1. Emplacement of ultramafics over sialic crust occurred during the Eocene to Oligocene. The gneisses of the D'Entrecasteaux are believed to be continuous with the sialic crust of the mainland, and as such would comprise a deeper section of the Papuan obduction system, resulting in the much higher grades of metamorphism exposed on the centre of the islands.

2. Uplift of the domes is believed to coincide with a sudden increase in grain size of sediments deposited in the Cape Vogel Basin (to the north) around Pliocene-Early Pleistocene times. Uplift is closely associated with the intrusion of large granodiorite bodies on Fergusson and Goodenough Islands, dated around 2 Ma (Goodenough Island, Webb in Davies, 1973). These granodiorite bodies may be subvolcanic plutons related to the peralkaline volcanism (Davies, 1973). Initial movement on dome-bounding faults probably commenced in the Pliocene, parts of the faults being obscured by both the calc-alkaline and peralkaline phases of volcanism (Pliocene to Pleistocene in age, Davies, Symonds & Ripper, 1984). Volcanism is strongly spatially-related to the north to northeast-trending fault systems, which both displace and are covered by the volcanics. Progressively younger structures are associated with lower grades of metamorphism, and the grade of metamorphism and coarseness of mylonitic fabrics decreases from west to east.

Conclusions

These structures are similar to metamorphic core complexes in that they consist of a mylonitic shear zone in metamorphic rocks, with a generally consistently-trending lineation, which is bowed up into a domal structure with a bounding fault on one side.

Granodiorite intrusion appears to be intrinsic in the formation of the domes and constitutes a possible mechanism for uplift of the shear zones. Collapse of the centre of Fergusson Island may be related to evacuation of part of the magma chamber caused by peralkaline volcanic activity. This collapse may also have caused the broad folding in the gneissic layering of the Mailolo dome.

Decreasing metamorphic grade and grain-size of mylonitic fabric from west to east suggests that shear zone movement in the east was occurring at temperatures lower than in the west.

Acknowledgements

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References

Binns, R.A., Scott, S.D., & PACLARK Participants, 1987 - Western Woodlark Basin: potential analogue setting for volcanogenic massive sulphide deposits. Australasian Institute of Mining & Metallurgy Pacific Rim Congress 87, Proceedings, 531-535.

Davies, H.L., 1973 - Fergusson Island, Papua New Guinea - 1:250 000 Geological Series. Bureau of Mineral Resources, Australia, Explanatory Notes, SC/56-5.

Davies, H.L., & Ives, D.J., 1965 - The geology of Fergusson and Goodenough Islands, Papua New Guinea. Bureau of Mineral Resources, Australia, Report 82.

Davies, H.L., & Smith, I.E., 1971 - Geology of Eastern Papua. Geol. Soc. Amer. Bull. 82, 3299-3312.

Davies, H.L., Symonds, P.A., & Ripper, I.D., 1984 - Structure and evolution of the southern Solomon Sea Region. BMR Journal of Australian Geology & Geophysics, 9, 49-68.

Davies, H.L., & Warren, R.G., 1983 - Kyanite eclogite in Cenozoic domed layered metamorphics in Southeastern Papua New Guinea. Abs. with Progs. Geol. Soc. Amer., 15, 554.

Masson, P.H., 1984 - Structure of Fergusson Island, D'Entrecasteaux Group, PNG. Exxon Minerals Company (unpub.).

McManus, J.B., 1986 - Annual Report for the Year ended 14/2/86 on Mineral Exploration of PA 539-Wapolu, Fergusson Island, D'Entrecasteaux Islands, Milne Bay Province, PNG (unpub.).

Ollier, C.D., & Pain, C.F., 1980 - Active rising surficial gneiss domes in Papua New Guinea. J. Geol. Soc. Aust., 27, 33-44.

Robertson Research (Australia) Pty Ltd & Flower Doery Buchan Pty Ltd, 1985 - Petroleum potential of the Cape Vogel, Bougainville and New Ireland Basins, Papua New Guinea. Volume 1.

Ripper, I.D., 1982 - Seismicity of the Indo-Australian/Solomon Sea plates boundary in the Southeast Papua region. Tectonophysics, 87, 355-369.

Senior, B.R., & Billington, W.G., 1987 - Structural interpretation of northwest Fergusson Island. Australasian Institute of Mining & Metallurgy Pacific Rim Congress 87, Proceedings.

Smith, I.E.M., 1976 - Peralkaline rhyolites from the D'Entrecasteaux Islands, Papua New Guinea. In Johnson, R.W., (Editor) - Volcanism in Australasia. Amsterdam, Elsevier, 275-286.

Smith, I.E.M., & Compston, W., 1982 - Strontium isotopes in Cenozoic volcanic rocks from southeastern Papua New Guinea. Lithos, 15, 199-206.

Smith, I.E.M., & Milsom, J.S., 1984 - Late Cenozoic volcanism and extension in Eastern Papua. In Kokelaar, B.P., & Howells, M.F., (Editors) - Marginal Basin Geology. Geol. Soc. Spec. Publ., 16, 163-171.

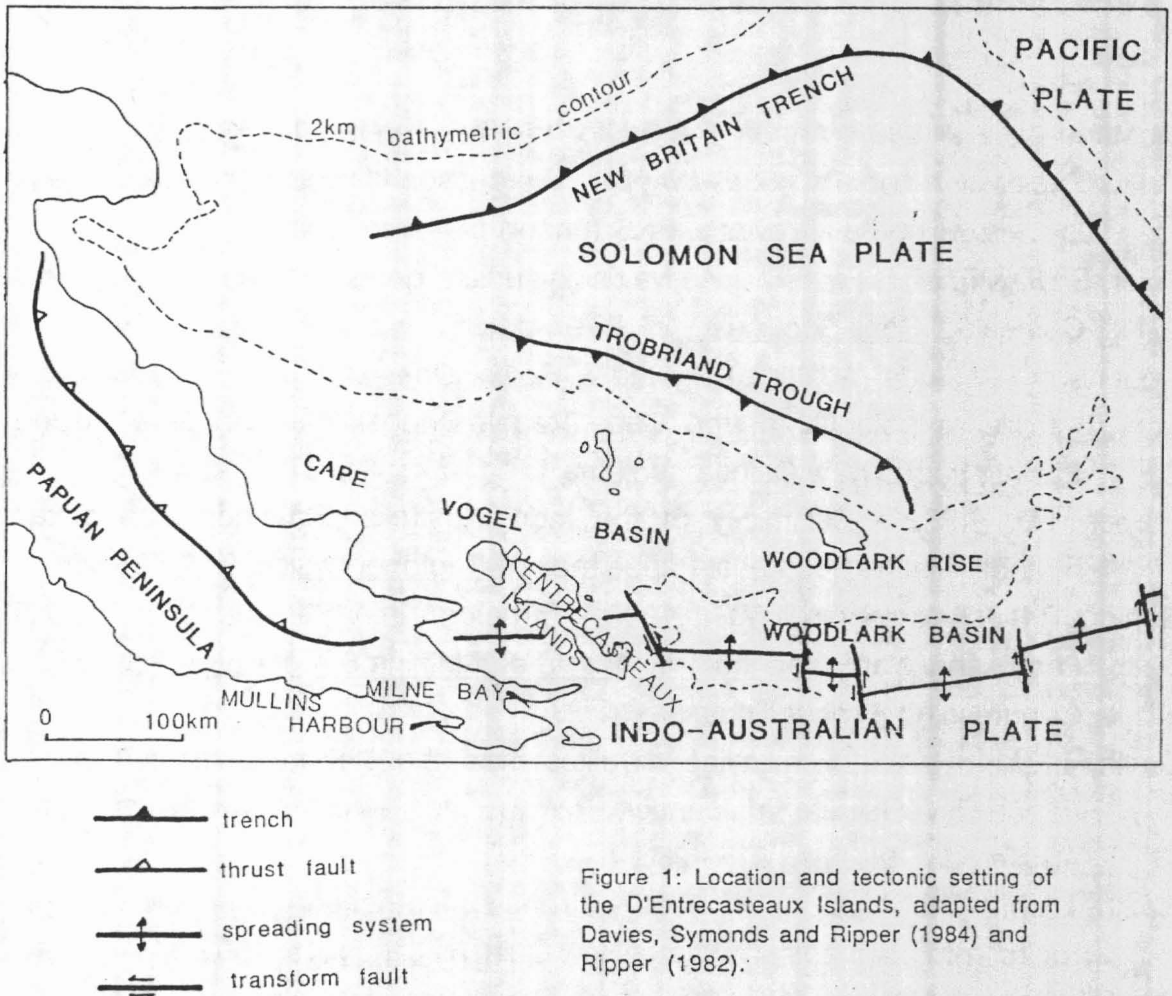


Figure 1: Location and tectonic setting of the D'Entrecasteaux Islands, adapted from Davies, Symonds and Ripper (1984) and Ripper (1982).

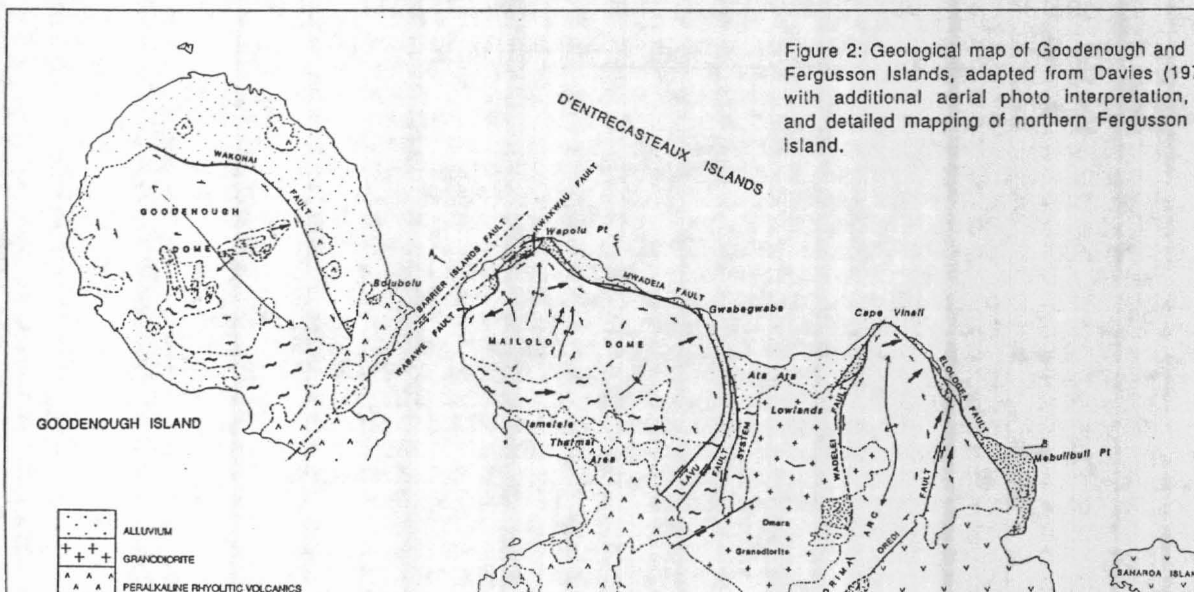


Figure 2: Geological map of Goodenough and Fergusson Islands, adapted from Davies (1973) with additional aerial photo interpretation, and detailed mapping of northern Fergusson island.

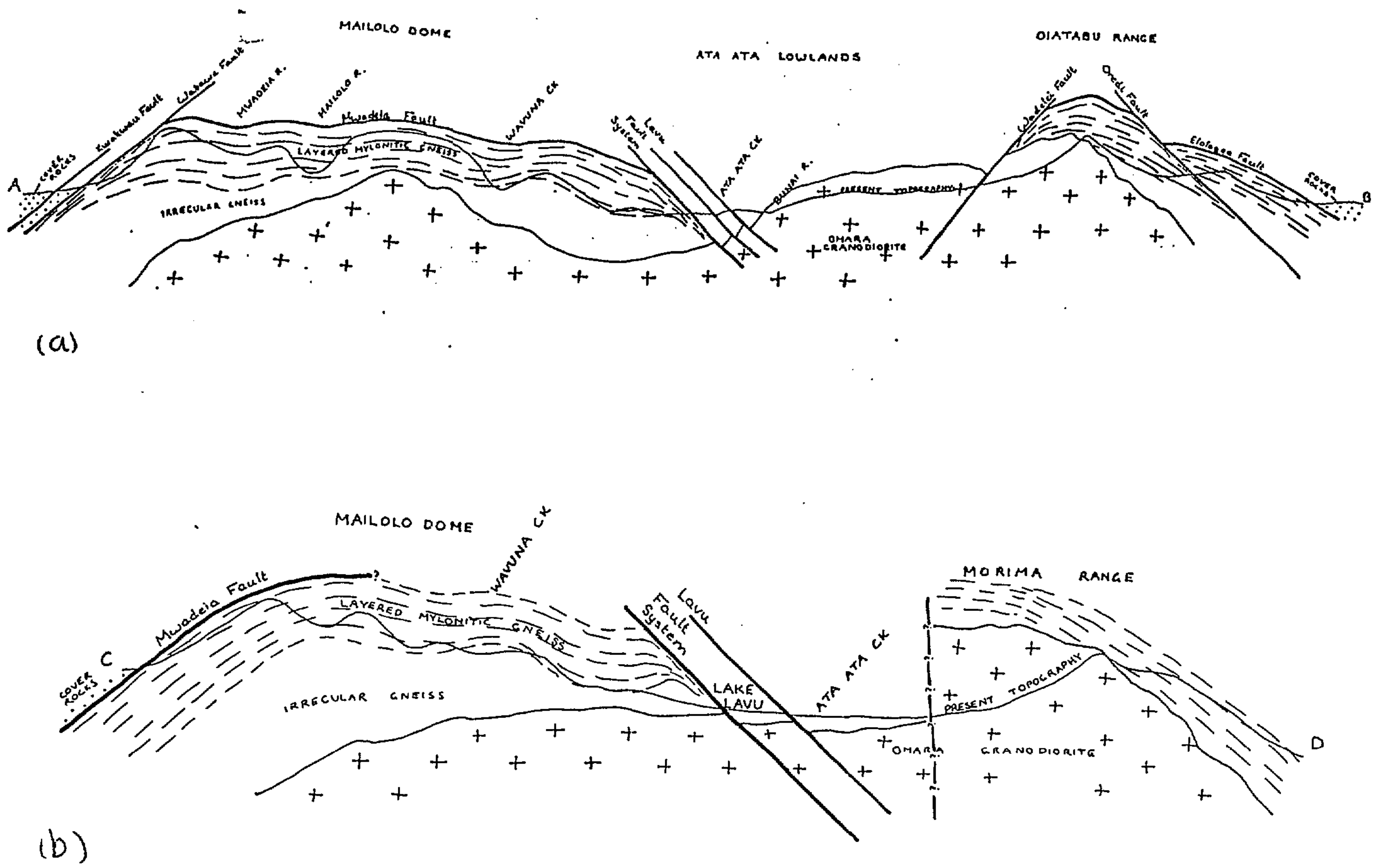


Figure 3:
(a) East-west section across Fergusson Island
(b) North-south section across western Fergusson Island

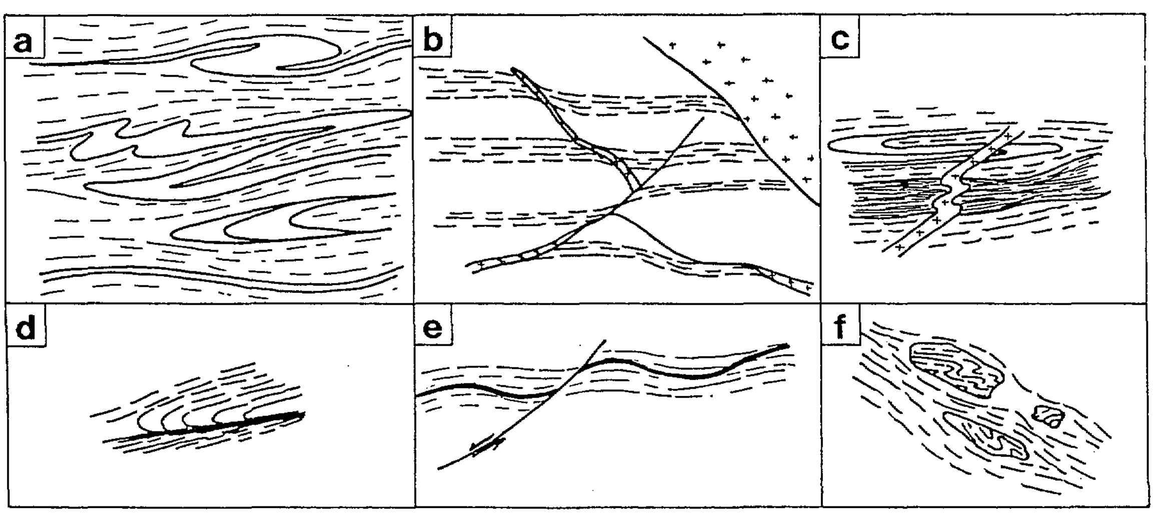


Figure 4: Summary of structural features. (a) First generation intrafolial folds with mylonitic axial plane foliation;(b) Conjugate high angle shear zones intruded by granitic veins; (c) low angle biotite shear zone cutting the mylonitic foliation at a low angle and folding a pegmatite vein; (d) sheared off lower limb of second generation fold; (e) normal fault displacing mylonitic foliation and biotite shear zone (dark line), faults commonly flatten with depth; (f) breccia along dome bounding fault containing fragments of gneiss with second generation folds in a matrix of biotite-actinolite-chlorite schist.

R8705107

Extensional tectonics in the Shuswap Terrane of the southern Canadian Cordillera

R L Brown, Carleton University, Canada

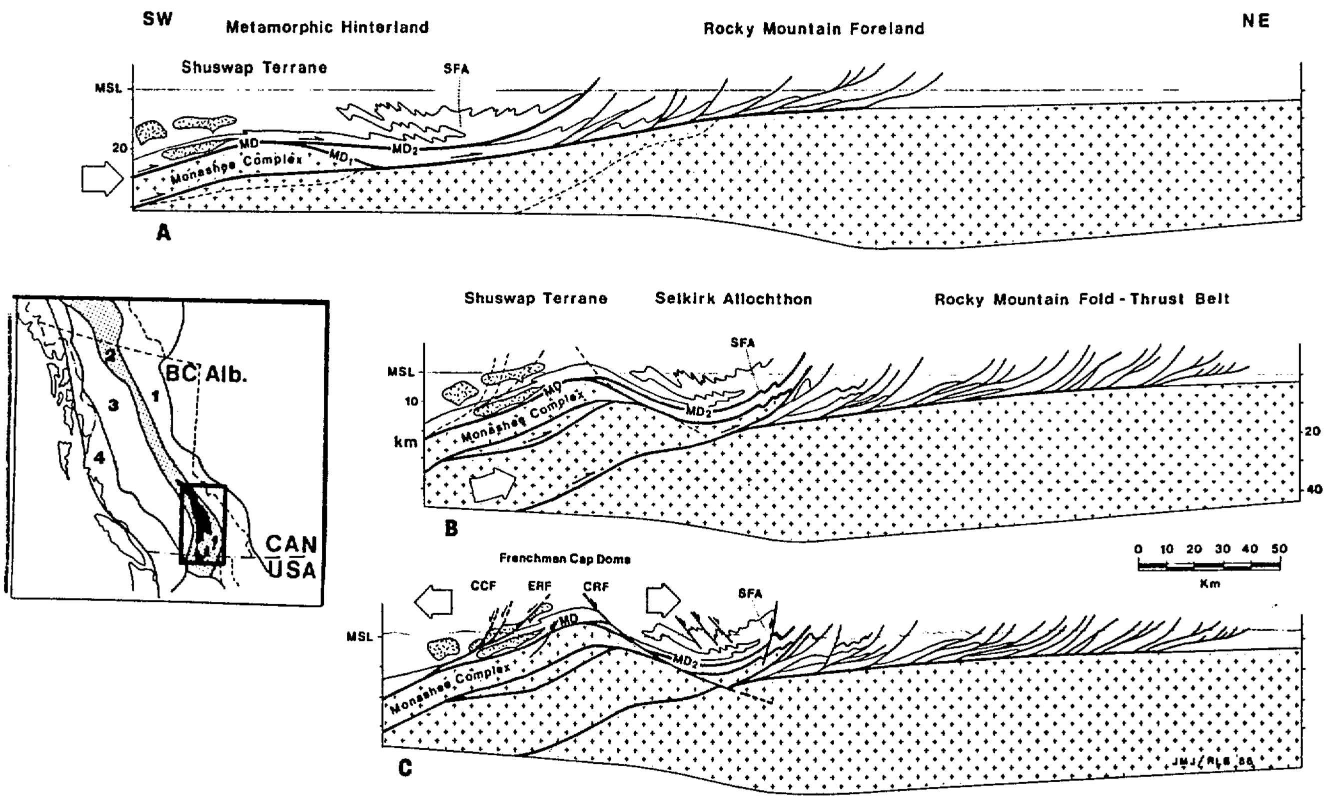
Precambrian (2.2 Ga) crust that is now exposed in the uplifted core of the metamorphic hinterland was initially extended in a Proterozoic episode of continental margin rifting. Stratabound carbonatite, mafic intrusions and lead–zinc mineralisation reflect this early rifting event. Prior to extension, shallow clastics and carbonates accumulated in a platformal shallow marine basin that may have evolved as the upper plate of an asymmetric rift margin.

Mesozoic marginal basin collapse, and accretion of exotic terranes to the North American continental margin prism, induced crustal thickening through duplexing of basement slices above deep crustal compressional shear zones. The crustal and possibly lithospheric welt, which achieved its maximum thickness in latest Cretaceous to early Tertiary time, was rapidly uplifted and denuded by the development of crustal extensional shear zones and associated detachment faults.

The extensional shear zones dip at moderate angles (30°C) away from the crustal culmination, and one has been traced on LITHOPROBE seismic reflection profiles down to near the base of the crust (27–30 km).

Above the brittle–ductile transition, hydraulic fracturing and veining has disrupted the mylonitic footwall rocks of the shear zones. These highly altered zones are being actively explored for economic mineral deposits.

Tertiary extension is interpreted to be directly related to the previous history of crustal thickening, and to have been facilitated by thermal weakening and associated magmatism. Both the compressional history and superimposed extension evolved in a transpressional plate boundary setting.



SW
Metamorphic Hinterland
Rocky Mountain Foreland
NE
Shuswap Terrane
SFA
MSL
20
MD
MD_1
MD_2
Monashee Complex
A
BC Alb.
1
2
3
4
CAN
USA
Shuswap Terrane
Selkirk Allochthon
Rocky Mountain Fold - Thrust Belt
SFA
MSL
10
km
MD
MD_2
Monashee Complex
20
40
B
Frenchman Cap Dome
CCF
ERF
CRF
SFA
MSL
MD
MD_2
Monashee Complex
C
0 10 20 30 40 50
Km
JMJ/RLB 86

Extensional structures in the Tumut Trough, southern New South Wales, Australia

P G Stuart-Smith, BMR/ANU

Recent work in the Brungle area of the Tumut Trough, a small Silurian basin about 100 km southwest of Canberra, has established features consistent with an early extensional history. In this area, rocks form two distinct domains: an ?Ordovician basement (Bullawyarra Schist) and a Silurian sedimentary and volcanic cover sequence (Brungle Creek Metabasalt, Wyangle Formation and Blowering Formation) (Fig. 1). These two domains are separated by a sharp discontinuity marking an abrupt change in rock type, structure, metamorphic grade and deformation style. The cover sequence has undergone only one major deformation during the Early Devonian, involving lower greenschist facies metamorphism and upright folding, as a result of east-west shortening. By comparison, the basement underwent at least two additional older deformations at upper greenschist facies (Table 1), and has distinct high-strain zones subconcordant with the basement/cover contact. The high-strain zones, characterised by a ubiquitous north-northwest - trending mineral-elongation lineation, record a progressive discontinuous history of ductile to brittle behaviour consistent with an extensional origin.

The structural and metamorphic discontinuity separating basement from the Silurian cover is characterised by extensive cataclasis and alteration, and is interpreted as a major detachment fault associated with extension and the development of the Tumut Trough in the Early to middle Silurian. Serpentinite was emplaced in the high-strain zones prior to intrusion of the Blacks Flat Diorite into the basement (Bullawyarra Schist) during the main period of movement on the detachment. This preceded deposition of Silurian trough-sediments and volcanics which unconformably overlie and onlap older units. The Brungle Creek Metabasalt represents an allochthonous sequence, possibly of Ordovician age, which probably has been considerably attenuated.

The development of the Tumut Trough, in the Brungle area, bears many similarities with Cordilleran metamorphic core complexes. Such a model is consistent with a back-arc marginal-sea environment suggested for the area by previous workers and may have a wider application to the development of similar Silurian "trough" sequences throughout the Lachlan Fold Belt.

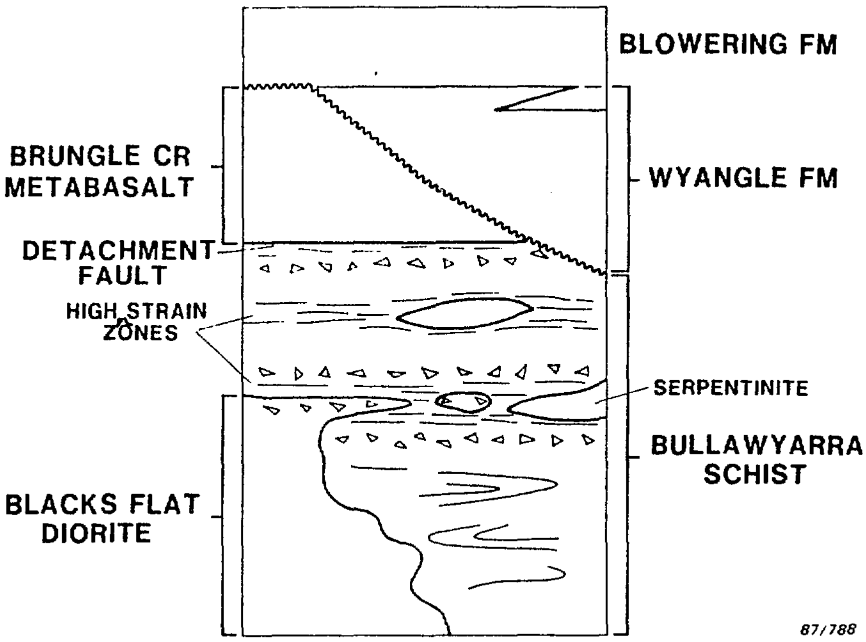


Figure 1. Schematic diagram showing components and relationships of basement and cover units in the Brungle area of the Tumut Trough.

		ORDOVICIAN		EARLY TO MIDDLE SILURIAN	EARLY DEVONIAN
		Upper greenschist facies metamorphism		Retrograde metamorphism	
	Rock type	S1	S2	Cataclasis	S3
MINERAL ASSEMBLAGES	Metabasalt/ metadolerite	Prismatic actinolite, epidote, sphene, magnetite, quartz, plagioclase	Fibrous actinolite epidote, sphene, magnetite, quartz, plagioclase	Carbonate, chlorite, epidote	Chlorite, carbonate, epidote
	Chert/ Quartzite		Biotite, muscovite		muscovite
	Meta-gabbro		Tremolite, zoisite		
	Serpentinite		Antigorite, carbonate, talc, magnetite	Chlorite, talc	
STRUCTURES		Isoclinal folding (F1) metamorphic foliation (S1) with mineral-elongation lineation (L1)	Recumbent folding (F2) with axial crenulation cleavage (S2), locally mylonitic with stretching lineation (L2)	Localised brecciation and formation of tectonic melange	Upright, open to tight folding (F3) axial crenulation cleavage (S3)

Table 1. Summary of structural and metamorphic history of the Bullawyarra Schist

Proterozoic continental extension in the Mount Isa Inlier, Queensland, Australia

P R Williams, A J Stewart, BMR, & P J Pearson, University of Queensland

The Mount Isa Inlier (Fig. 1) is a Proterozoic igneous and metasedimentary complex in northwest Queensland, which hosts important copper, lead-zinc, silver, gold, cobalt and uranium deposits. It is overlain by flat-lying sedimentary rocks of Late Proterozoic and Early Cambrian age. A recent research program in the Mount Isa area, coordinated by BMR, has included a detailed examination of structures in the central part of the inlier, where brittle extensional faulting and associated dolerite intrusion preceded thrust faulting and folding and upright compressional deformation. Narrow imbricate fault-slices, low- and high-angle normal faults, listric-normal-fault fans, fault breccias, brecciation of quartzite units, and large regional stratigraphic omissions with related detachment structures formed during the extensional event. Some anomalous geometric patterns that developed during later ductile compression are attributed to rotation on early extensional faults. The early brittle structures underwent, and therefore preceded, ductile compressional deformations and may be related to basin-forming crustal or lithospheric extension.

The geological history of the inlier indicates that it was an area of intracratonic rifting during the period 1800-1670 Ma (Etheridge & others, 1987; Blake, 1987), with at least two rift events and associated thermal sag phases accounting for the sedimentary and volcanic sequences. The oldest extensively exposed rocks are the felsic Leichhardt metavolcanics (1860 Ma) of the Kalkadoon-Leichhardt Belt and Blockade Block (cover sequence 1; Blake, 1987). These are overlain by metabasalt (Magna Lynn Metabasalt), felsic metavolcanics with subsidiary feldspathic sandstone (Argylla Formation), quartzite (Ballara Formation) and banded calcsilicate and metabasalt (Corella Formation). The calcsilicates are overlain by feldspathic sandstone, mica schist and banded granofels (Deighton and White Blow Formations). These belong to cover sequence 2 of Blake (1987), and were deposited between 1810 and 1750 Ma. Cover sequence 3, which includes the Mount Isa Group, host to the Mount Isa shale-hosted sulphide deposits, was deposited between 1700 and 1670 Ma in the western part of the inlier.

Little structural evidence for continental rifting has been presented to date. In the western part of the inlier, north of Mount Isa township, several authors (e.g. Smith 1969) have discussed syn-depositional faulting, but Bell (1983) argued that these faults are in fact thrust-faults separating horses of a south-directed thrust-complex. These compressional events affect cover sequence 3 and occurred long after rifting.

Mesoscopic structures related to ancient continental rifting have been best described in the Basin and Range province in Nevada, where low-angle extensional detachment faults separate lower plate rocks deformed in a ductile manner from upper plate rocks deformed in a brittle manner (Crittenden & others, 1980). Brittle deformation above the detachment fault commonly overprints mylonitic rocks of the lower plate in a progressive deformation sequence (Davis & others, 1986; Wust, 1986); the overprinting may indicate a separate phase of brittle faulting with reactivation of older ductile shear-zones (Gaudemer & Tapponier, 1987). The upper plate is characteristically intensively shattered, brecciated and faulted. High-angle normal faults are common, as are large fault-blocks cut by low-angle normal faults. Also, listric faults within larger blocks converge downward onto a decollement surface (Seager, 1970; Anderson, 1971). Individual blocks cut by scoop- and bowl-shaped faults have also been reported (Seager, 1970), as have normal imbricate fault-complexes (Davis & Hardy, 1981).

In the central part of the Mount Isa inlier, Passchier (1986) has identified a set of imbricate structures of apparently extensional nature, with associated brittle deformation features. In the nearby Jimmy Creek area normal faults now rotated by two generations of folds were originally extensional imbricate fans (Fig. 2) and were the locus of dolerite intrusion and brecciation. Mylonitic decollement surfaces form the base of these extensional imbricates. Pearson & others (1987) have attributed stretching lineations and mylonite in the Wonga Belt to early crustal extension and argue that these structures are synchronous with granite intrusion. In the Little Beauty area (Fig. 3) rotational-block-faulted Ballara Quartzite forms the limbs of a regional syncline. The fault planes are clearly rotated by the regional fold, and some have been reactivated. Quartz-tourmaline fault breccias and dolerite intrusions parallel to the faults show an early stretching lineation and are also rotated by the regional upright folding. The core of the regional syncline is marked by an intense shear zone along which hydrothermal alteration has taken place. There is a major stratigraphic omission across the shear zone, with Corella Formation faulted against Leichhardt metavolcanics. To the west the rotational-block-faults die out against a ductile shear zone in metabasalt. To the east strain increases towards the sheared granites of the Wonga Belt.

On the eastern margin of the Deighton klippe (Fig. 1), the Corella Formation is separated from the Leichhardt metavolcanics by a fault zone of brecciated, tourmaline-rich Corella-type lithologies. Conformable relationships elsewhere between Corella Formation and Deighton Quartzite militate against a thrust nappe emplacement of the quartzite (Loosveld & Schreurs, 1987). The extensive brecciation, faulting which brings younger over older sequences, and asymmetric high-angle normal faulting, all of which predate the regional compressional deformations affecting the central Mount Isa inlier, are strong evidence of an early extensional brittle tectonic environment.

Because of the widespread evidence for early extensional structures in the central part of the inlier we suggest that a phase of regional continental extension resulted in brittle disruption of cover sequence 2, synchronous with emplacement of granite, development of regional shear zones and intrusion of mafic dyke swarms. We infer a shallowly-dipping low-angle detachment surface separating brittle and ductile deformation regimes (Fig. 4). The direction of movement on the surface is still uncertain, with south-directed movement indicated in the Wonga belt and west-directed movement in the Deighton area. Southwest and southeast movement has been inferred in the Alligator and Jimmy Creek areas respectively. The regional extension was possibly related to formation or sedimentary basins in which cover sequence 3 accumulated to the west.

References

Anderson, R.E., 1971 - Thin skin distension in Tertiary rocks of southern Nevada. Geological Society of America Bulletin 82, 43-58.

Bell, T.H., 1983 - Thrusting and duplex formation at Mount Isa, Queensland, Australia. Nature, 304, 493-497.

Blake, D.H., 1987 - Geology of the Mount Isa Inlier and environs, Queensland and Northern Territory. Bureau of Mineral Resources, Australia, Bulletin 225.

Crittenden, M.D., Coney, P.J., & Davis, G.H., 1980 - Cordilleran metamorphic core complexes. Geological Society of America Memoir 153.

Davis, G.H. & Hardy, J.J., 1981 - The Eagle Pass detachment, southern Arizona: product of mid-Miocene listric (?) normal faulting in the southern Basin and Range. Geological Society of America Bulletin 92, 749-762.

Etheridge, M.A., Rutland, R.W.R., & Wyborn, L.A.I., 1987 - Orogenesis and tectonic process in the Early to Middle Proterozoic of Northern Australia. In Kroner, A., (Editor) - Proterozoic lithospheric evolution. American Geophysical Union, Geodynamics Series 17, 131-147.

Gaudemer, Y., Tapponier, P., 1987 - Ductile and brittle deformation in the northern Snake Range, Nevada. Journal of Structural Geology, 9, 159-180.

Loosveld, R. & Schreurs, G., 1987 - Discovery of thrust klippen northwest of Mary Kathleen, Mt Isa Inlier, Australia. Australian Journal of Earth Sciences, 34, 387-402.

Passchier, C.W., 1986 - Evidence for early extensional tectonics in the Proterozoic Mount Isa inlier, Australia. Geology, 14, 1008-1011.

Pearson, P.J., Holcombe, R.J., & Oliver, N.H.S., 1987 - The Mary Kathleen Fold Belt, northwest Queensland: D1 - a product of crustal extension. Geological Society of Australia, Abstracts, 19, 37-38.

Seager, W.R., 1970 - Low-angle gravity glide structures in the northern Virgin Mountains, Nevada and Arizona. Geological Society of America Bulletin 81, 1517-1538.

Smith, W.D., 1969 - Penecontemporaneous faulting and its likely significance in relation to Mount Isa ore deposition. Geological Society of Australia Special Publication, 2, 225-235.

Wust, S.L., 1986 - Extensional deformation with northwest vergence, Pioneer core complex, central Idaho. Geology, 14, 712-714.

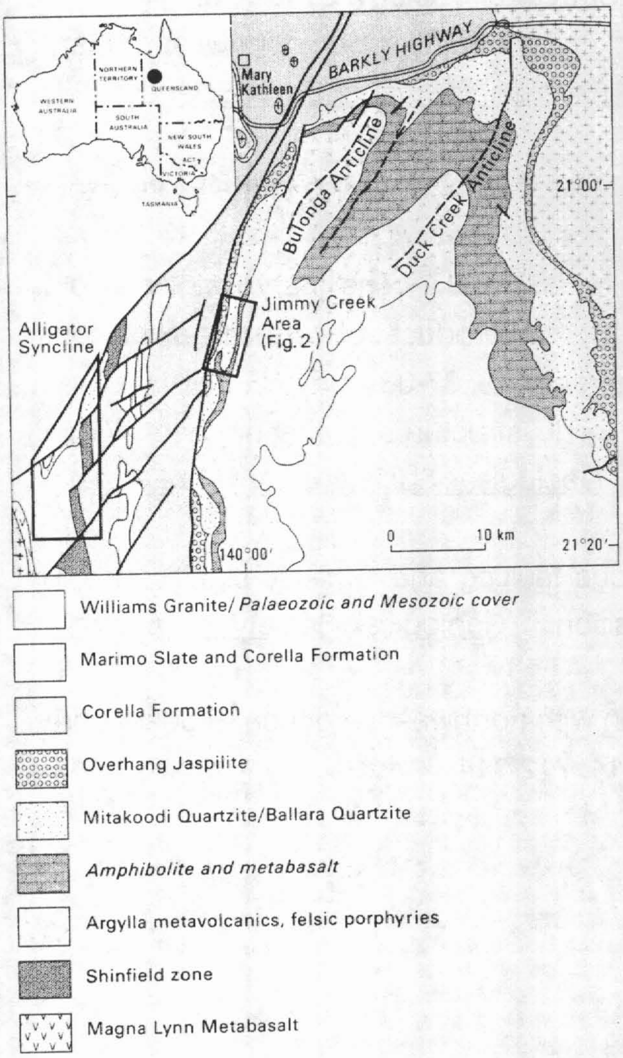
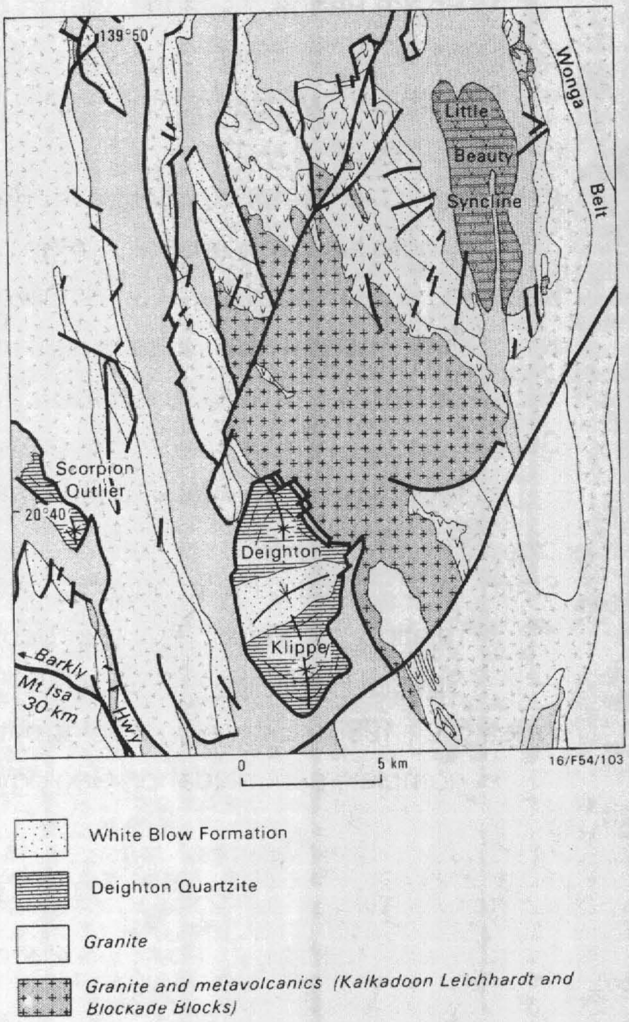


Figure 1A. Locality map of the Mount Isa Inlier and the Jimmy Creek area



1B. Locality and generalised geology of the Deighton Klippe and Little Beauty syncline

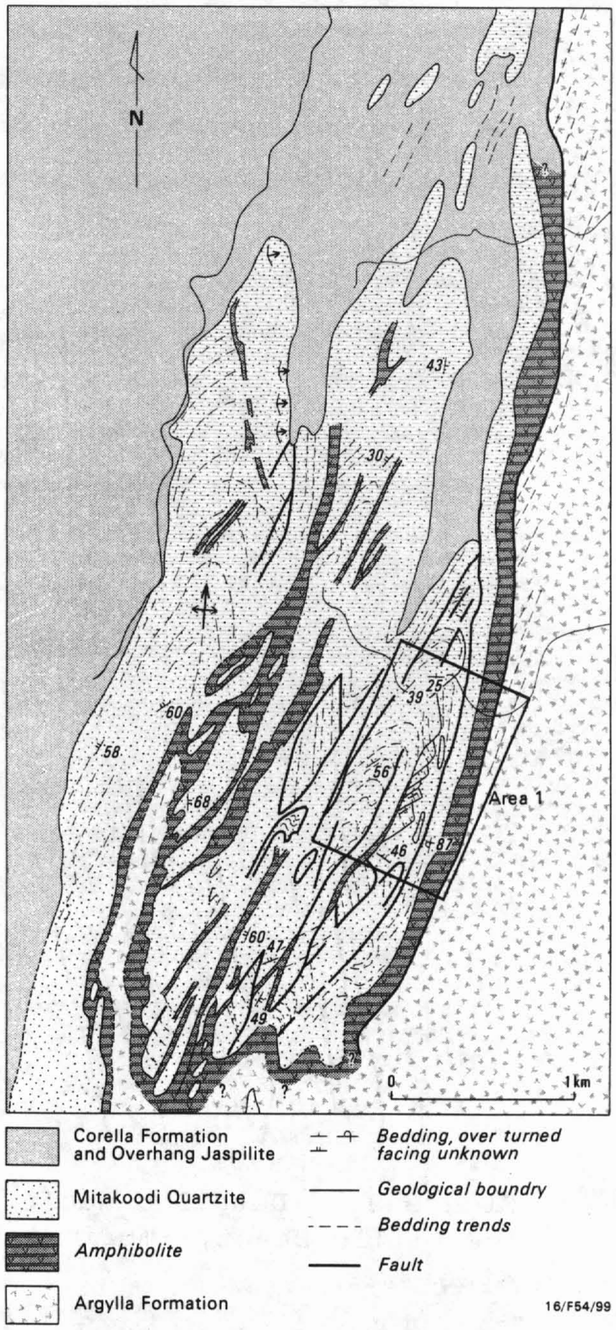


Figure 2. Geology of part of the west limb of the Bulonga Anticline, showing normal fault fan and pattern of amphibolite bodies

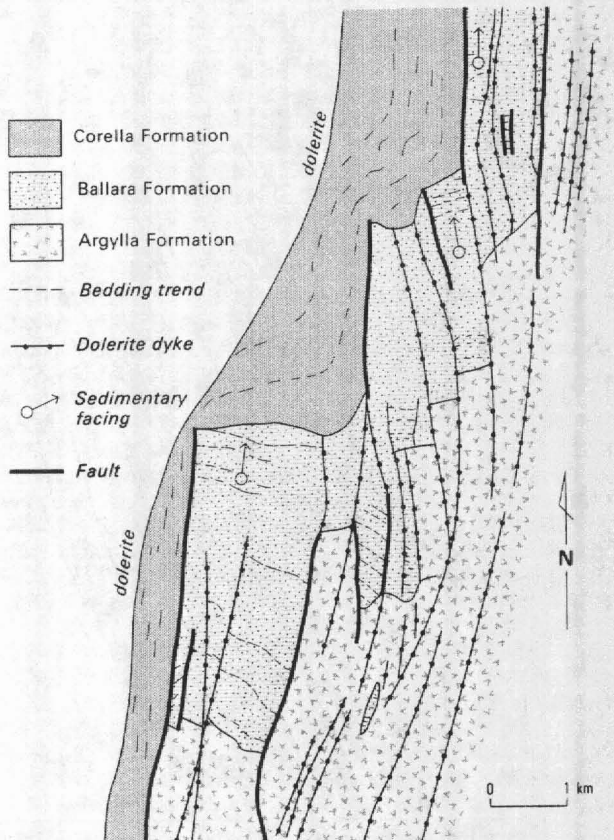


Figure 3. Detailed geology of the Little Beauty imbricates and dyke swarm

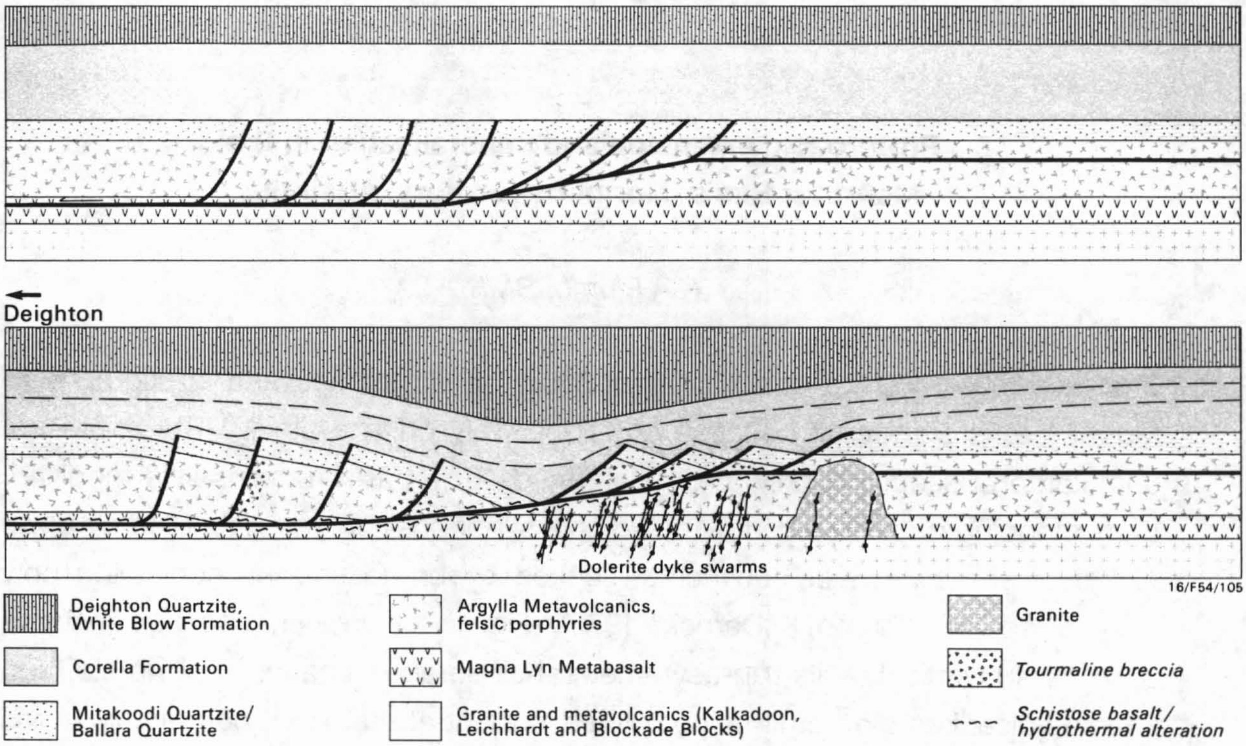


Figure 4A. Possible cross-section reconstruction of the Deighton Klippe after extension, with a restored section indicating location of early faults.

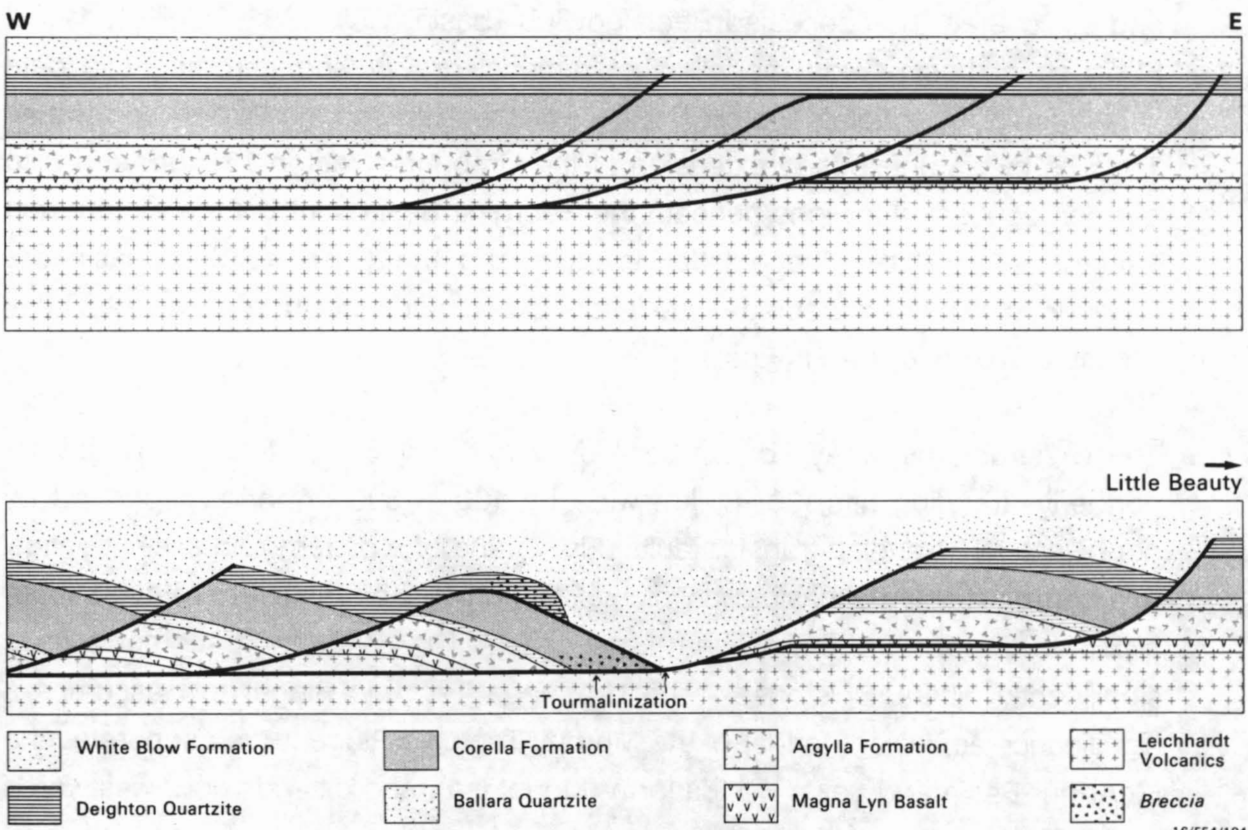


Figure 4B. Possible schematic cross-section through the Little Beauty area, prior to compressional deformation. Undeformed section indicates possible location of early faults.

Proterozoic extension and mineralisation in the McArthur Basin, Northern Territory, Australia

K A Plumb, BMR

The ca. 1700 – 1400 Ma – old McArthur Basin, of similar age and stratigraphy to Mount Isa, is significant because of its relatively pristine state. The rocks are unmetamorphosed and only mildly deformed. The basin remains little modified. Early mapping outlined a central meridional syndepositional graben filled by up to 12 km of shallow – water sediments, flanked by broad shelves containing only 4 km of section (Plumb & Derrick, 1975) (Fig. 1). However, because features such as subvertical faults, an asymmetric half – graben shape, axial horsts, and sparse volcanism did not fit rifting models current at that time, nongenetic terms such as trough or graben were used and Plumb & others (1980) developed a structural model based on strike – slip faulting.

Many critical zones were exposed poorly, but a multidisciplinary geophysical profile confirmed the general deep structure across the southern McArthur Basin, and discovered the concealed Beetaloo Sub – basin (Cull, 1982; Collins, 1983). Subsequent basin – wide modelling of regional gravity and aeromagnetic data showed that the "original Batten Trough" is separated into the Walker and Batten Troughs by an unexposed extension of the Urapunga Tectonic Ridge, and also outlined concealed troughs beneath the Georgina Basin south of the Kilgour and Tanumbirini Uplifts. These latter troughs and the Beetaloo Sub – basin are continuous with outcropping Tomkinson Creek beds north of Tennant Creek (Plumb & Wellman, 1987) (Figs 1, 2).

These structures may be modelled very simply in terms of several northerly – trending rifts, 30 – 80 km wide by 100 – >300 km long, separated by northwesterly – trending transfer faults (Fig. 3) and of a pattern and scale identical to modern extensional basins, in which extension may be related to rotation about a pole to the north of the Arnhem Land coast. The different rift segments have significantly different histories. The Urapunga Tectonic Ridge now assumes a significance equivalent to that of the Murphy Tectonic Ridge, which separates the McArthur Basin and Lawn Hill Platform (Mount Isa), and extends northwestwards as the Kilgour – Tanumbirini Uplift.

The McArthur Basin succession comprises four main sequences, separated by regional unconformities. Various shallow – marine and continental environments have been interpreted from the sequences in the southern part of the basin (Jackson & others, 1987), and may be inferred for less well-known areas in the north, but revised correlations modify significantly basin-wide palæogeographic models (Plumb, 1985; cf. Plumb & others, 1980, 1981). Data are inadequate for precise balanced or palinspastic cross-sections or estimates of extension. Steepening by later deformation hampers particularly the reconstruction of fault zones at depth. However, the basin's evolution still can be reconstructed schematically, and with regard to reasonable control of thickness and scale (Figs 4, 5). In such models the Emu Fault almost certainly represents the surface relic of the major detachment beneath the Batten Trough, but a more complex history and structure preclude identification of the equivalent detachment beneath the Walker Trough.

Where exposed, most Tawallah Group units display uniform thicknesses and shallow-marine, lacustrine, and alluvial sandstone and volcanic-lutite-carbonate facies throughout the southern McArthur Basin, except for a dramatically thinned block within the Emu Fault Zone (Fig. 4B) (Jackson & others, 1987). Outcrops in the uplifted Tawallah Range, although now modified considerably by post-depositional block-faulting (with apparent throws of up to 6 km), nevertheless are tilted and faulted regularly in a pattern consistent with extension (Fig. 4A). However, there is little change in stratigraphy between blocks, or evidence of growth-faulting, and so the tilting shown schematically in Figure 4 may simply reflect accommodation of later (McArthur Group) extension.

The hypersaline paralic to lacustrine carbonate units of the McArthur Group are confined to the Batten Trough or adjacent areas. Striking variations in the basal (Masterton) sandstone unit are consistent with tilting of the underlying Tawallah Group. Features such as lack of metamorphism at the exposed base of the apparently thick (10 km) basin sequence, proximal source areas, marked unconformities, missing sequences, and other post-Tawallah Group stratigraphic irregularities, all indicate that the Tawallah Range tilt blocks were elevated during McArthur Group and later time. Lower McArthur Group stratigraphy (above the basal sandstone) is elsewhere regionally uniform, so few, if any, faults reached surface at this time. Extension is demonstrated most strikingly by the Batten Subgroup and underlying Barney Creek Formation, which show hypersaline lacustrine sequences with abundant tuffaceous volcanism, deposited in a half-graben; striking stratigraphic changes between fault blocks; and demonstrable growth faulting and talus breccias along the Emu and related fault-zones.

Although the uplift blocks are still apparent in the distribution of basal conglomerates in the Nathan Group, the main lacustrine or paralic carbonates show no evidence of growth faults, etc.; the sequence may represent a sag phase. The much younger shallow-marine blanket-sands and intervening lutites of the Roper Group show a marked shift in depositional pattern. A gradual thickening towards the southwest again resembles a sag phase above the Batten Trough, but the geophysical signature of the concealed Beetaloo Sub-basin suggests faulted (rifted) margins.

The history of the northern McArthur Basin differs from that in the south. The sandstone-rich Parsons Range Group (Tawallah Group equivalent) has no equivalent outside the Walker Trough and, significantly, has no volcanics. Extension was accommodated by closely-spaced extensional faults in the basement; these cut out mostly at the base of the basin (Fig. 5). The McArthur Group in the Walker Trough displays structure and facies similar to those farther south in the Batten Trough, but the equivalent Katherine River Group on the Arnhem Shelf comprises mostly fluvial sands alternating with volcanics, carbonates, and lutites. Local rifts developed at the northwestern edge of the basin. The Mount Rigg Group is similar in facies and distribution to the equivalent Nathan Group, and only a very thin Roper Group was deposited on the Arnhem Shelf. There is no evidence as to whether the Mitchell Range was elevated during basin subsidence (cf. Tawallah Range), or whether all central uplift (up to 10 km) was post-depositional.

The controlling faults of the McArthur Basin are inherited from at least the Early Proterozoic. Pre-McArthur Basin movements have been demonstrated on the Calvert and Bulman Faults, and on faults within the Mitchell Range (Plumb & others, 1980), while the Urapunga Tectonic Ridge is on-trend with the Early Proterozoic South Alligator Hinge Zone (adjacent to Mt Callanan Basin, Fig. 1). Northwest strike-slip faults (the transfer fault system) are prominent in rocks of all ages throughout northern Australia (Plumb, 1979). The grabens comprise two, apparently conjugate, sets - north-northwest to north, and northeast - subparallel to similar systems in the Mount Isa area. Post-depositional uplift and thrusting was concentrated on the more northerly set. The northeast and northwest faults formed a later set of conjugate strike-slip faults, as at Mount Isa. The latest deformation involved unrelated strike-slip displacements on the north-northwest to north set.

Recent models describe extensional terranes as favourable environments for mineralisation. The terranes may develop a variety of sites for mineral deposition and circulate mineralising brines, because of both high heat flows and the development of favourable channel-ways, particularly along the major listric and transfer fault zones.

The lacustrine or lagoonal facies of the HYC and neighbouring shale-hosted lead-zinc deposits are clearly localised by growth-faulting on the major (listric?) fault of the Batten Trough, the Emu Fault. The restricted lake sediments acted as a hydrologic trap for diagenetic brines, which were derived demonstrably from the fault zone (Walker & others, 1978; Williams, 1978b; Logan & Williams, 1984). A newly-recognised class of widespread small lead-zinc and copper deposits in the McArthur Basin comprises discordant karstic and unconformity-related deposits (Jackson & others, 1987). These deposits are also clearly localised along major fault zones of the rift systems (Fig. 1), probably because unconformities are best developed on the uplifted tilt blocks, and the mineralising brines are derived demonstrably from a source similar to that of the HYC deposit (Williams, 1978a; Walker & others, 1983; Muir & others, 1985).

References

Collins, C.D.N., 1983 - Crustal structure of the southern McArthur Basin, northern Australia, from deep seismic sounding. BMR Journal of Australian Geology & Geophysics, 8, 19-34.

Cull, J.P., 1982 - Magnetotelluric profiles in the McArthur Basin of northern Australia. BMR Journal of Australian Geology & Geophysics, 7, 275-286.

Jackson, M.J., Muir, M.D., & Plumb, K.A., 1987 - Geology of the southern McArthur Basin, Northern Territory. Bureau of Mineral Resources, Australia, Bulletin 220.

Logan, R.G., & Williams, N., 1984 - Sedimentary controls on the hydrothermal system that formed the HYC deposit at McArthur River, Northern Territory. Geological Society of Australia, Abstracts, 12, 335-340.

Muir, M.D., Donnelly, T.H., Wilkins, R.W.T., & Armstrong, K.J., 1985 - Stable isotope, petrological, and fluid inclusion studies of minor deposits from the McArthur Basin: implications for the genesis of some sediment-hosted base metal mineralisation from the Northern Territory. Australian Journal of Earth Sciences, 32, 239-260.

Plumb, K.A., 1979 - Structure and tectonic style of the Precambrian shields of northern Australia. Tectonophysics, 58, 291-325.

Plumb, K.A., 1985 - Subdivision and correlation of Late Precambrian sequences in Australia. Precambrian Research, 29, 303-329.

Plumb, K.A., 1986a - Batten Subgroup, McArthur Basin - modern analogues for an evolving ancient lake. Geological Society of Australia, Abstracts, 15, 159-160.

Plumb, K.A., 1986b - Batten Subgroup, McArthur Basin - evolution of a 1650 Ma-old lake complex. 12th International Sedimentological Congress, Canberra, Abstracts, 244.

Plumb, K.A., & Derrick, G.M., 1975 - Geology of the Proterozoic rocks of the Kimberley to Mount Isa region. In Knight, C.L., (Editor) - Economic Geology of Australia and Papua New Guinea. 1. Metals. Australasian Institute of Mining and Metallurgy, Melbourne, Monograph Series, 5, 217-252.

Plumb, K.A., Derrick, G.M., Needham, R.S., & Shaw, R.D., 1981 - The Proterozoic of Northern Australia. In Hunter, D.R., (Editor) - Precambrian of the Southern Hemisphere. Developments of Precambrian Geology, 2, Elsevier, Amsterdam, 205-307.

Plumb, K.A., Derrick, G.M., & Wilson, I.H., 1980 - Precambrian geology of the McArthur River-Mount Isa region, Northern Territory. In Henderson, R.A., & Stephenson, P.J., (Editors) - The Geology and Geophysics of northeastern Australia. Geological Society of Australia, Inc., Queensland Division, Brisbane, 4, 71-88.

Plumb, K.A., & Wellman, P., 1987 - McArthur Basin, Northern Territory: mapping of deep troughs using gravity and magnetic anomalies. BMR Journal of Australia Geology & Geophysics, 10, 243-251.

Walker, R.N., Logan, R.G., & Binnekamp, J.G., 1978 - Recent geological advances concerning the HYC and associated deposits, McArthur River, NT. Journal of the Geological Society of Australia, 24, 365-380.

Walker, R.N., Gulson, B., & Smith, J., 1983 - The Coxco Deposit - a Proterozoic Mississippi Valley-type deposit in the McArthur River district, Northern Territory, Australia. Economic Geology, 78, 214-249.

Williams, N., 1978a - Studies of the base metal sulphide deposits at McArthur River, Northern Territory, Australia: I. The Cooley and Ridge deposits. Economic Geology, 73, 1005-1035.

Williams, N., 1978b - Studies of the base metal sulphide deposits at McArthur River, Northern Territory, Australia: II. The sulphide-S and organic-C relationships of the concordant deposits and their significance. Economic Geology, 73, 1036-1056.

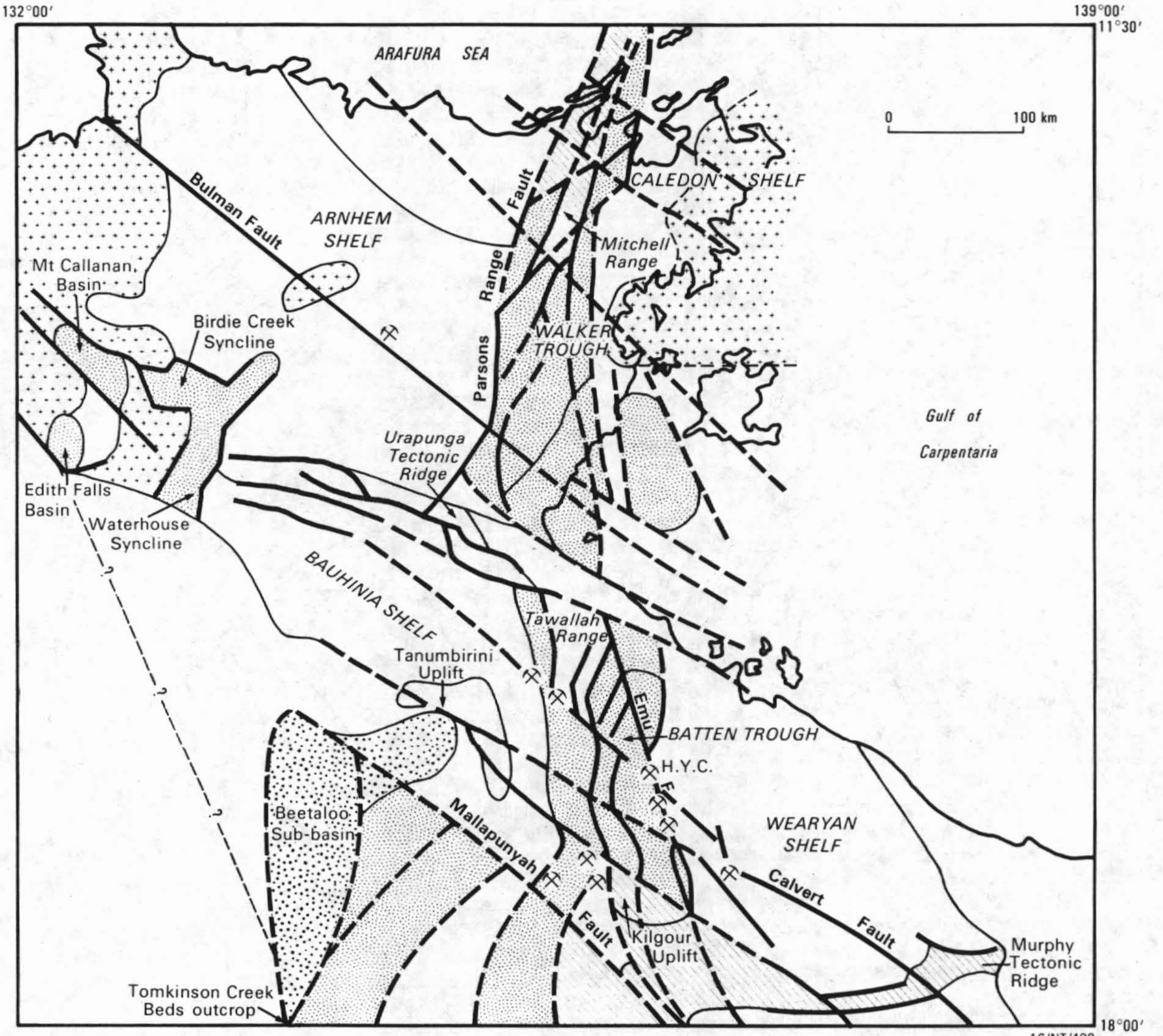


Figure 1. Principal tectonic elements of the McArthur Basin, adapted from Plumb & Derrick (1975) and Plumb & Wellman (1987).

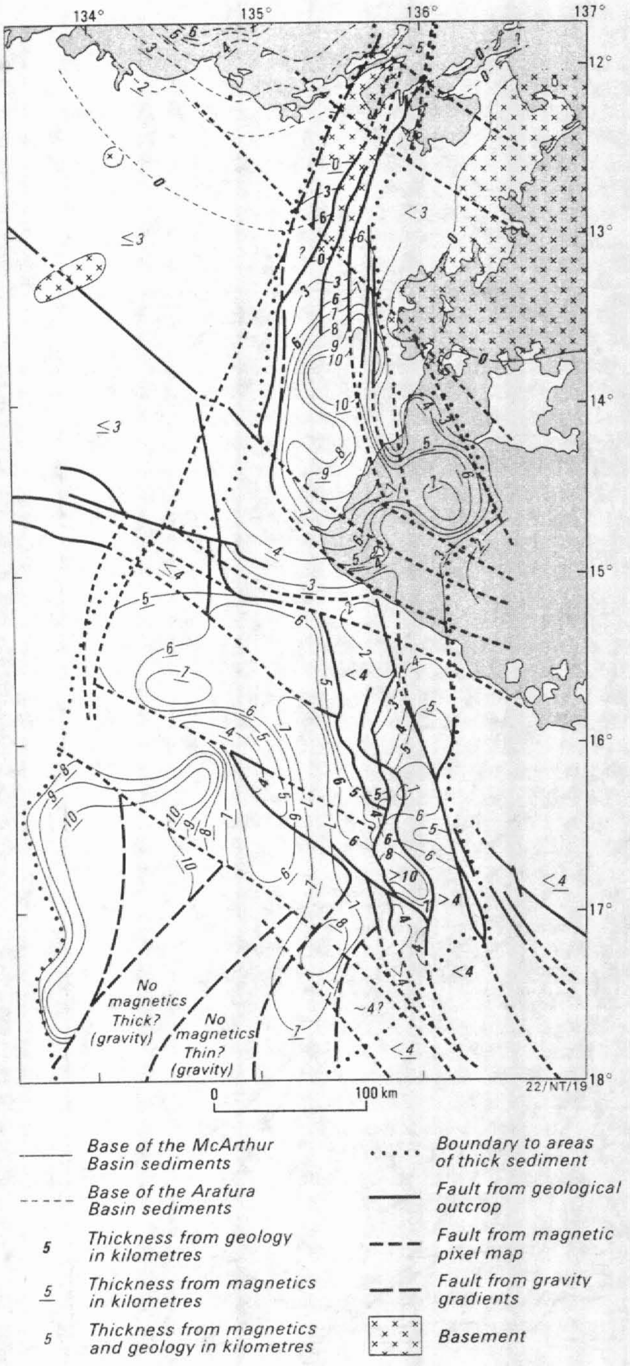


Figure 2. Inferred thickness of sediment in part of the McArthur Basin from magnetic anomalies and geology, after Plumb & Wellman (1987).

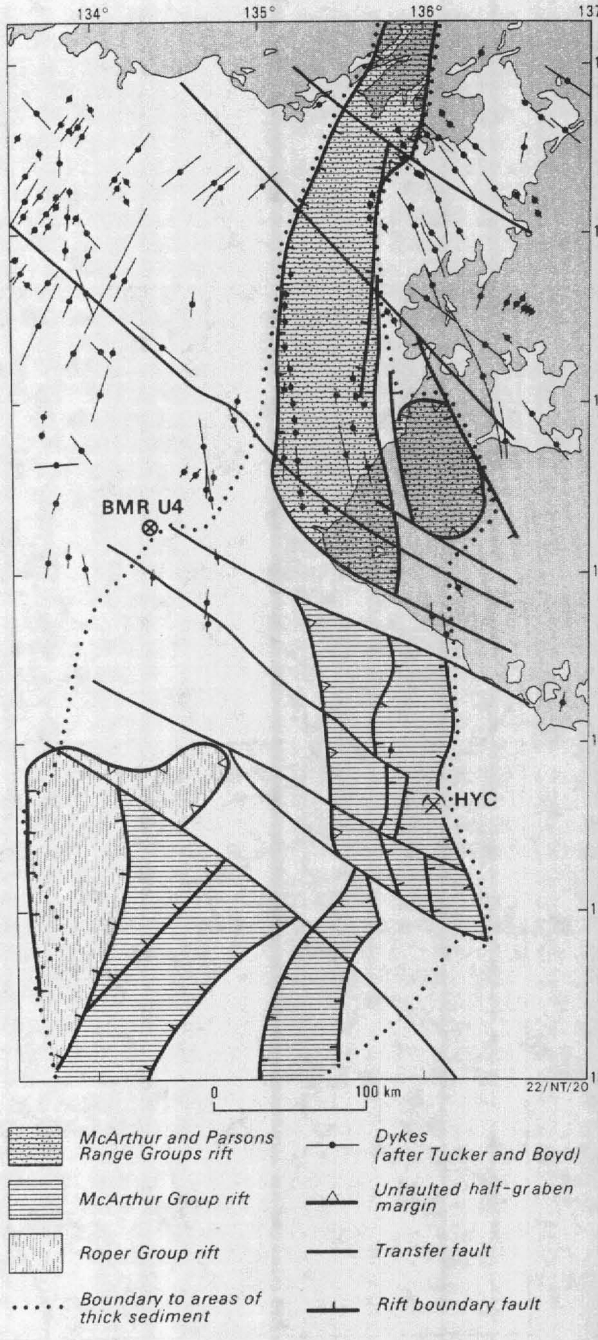


Figure 3. Tectonic model for part of the McArthur Basin, from Plumb & Wellman (1987).

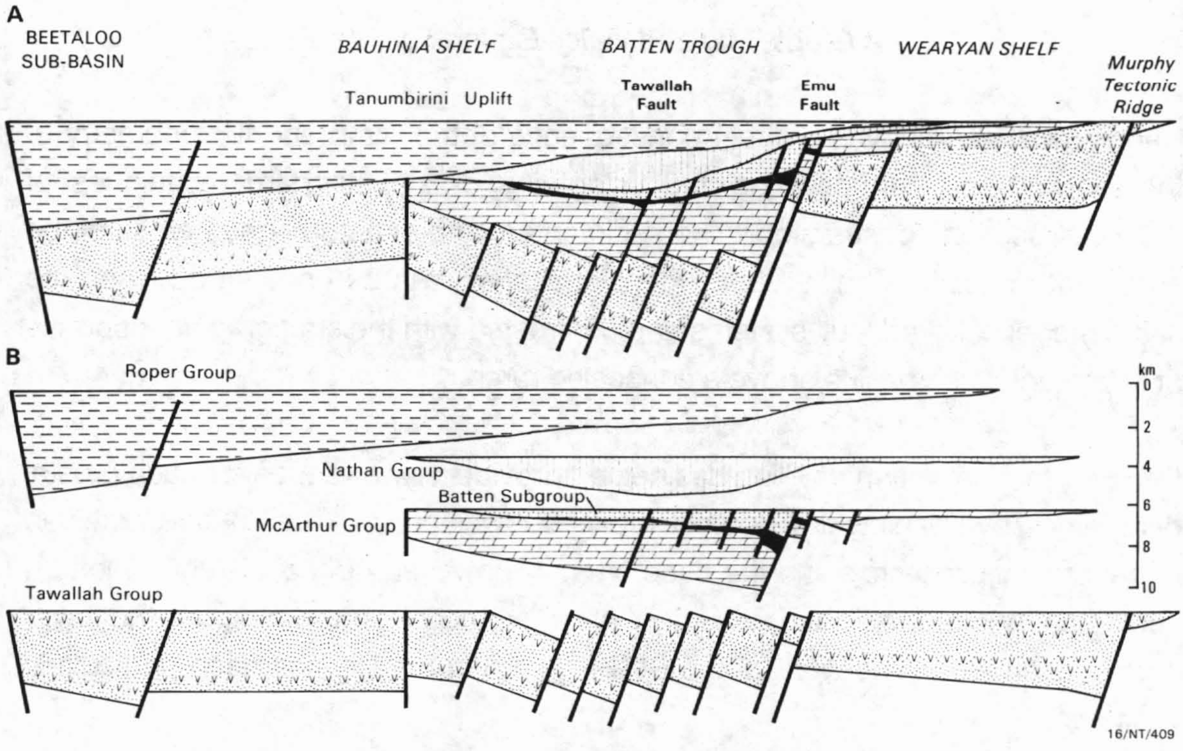


Figure 4. Schematic development of the southern McArthur Basin, from the Beetaloo Sub-basin to the Wearyan Shelf, through the Batten Trough.

A. Schematic cross-section at the close of Roper Group sedimentation.

B. Schematic development of the principal depositional units.

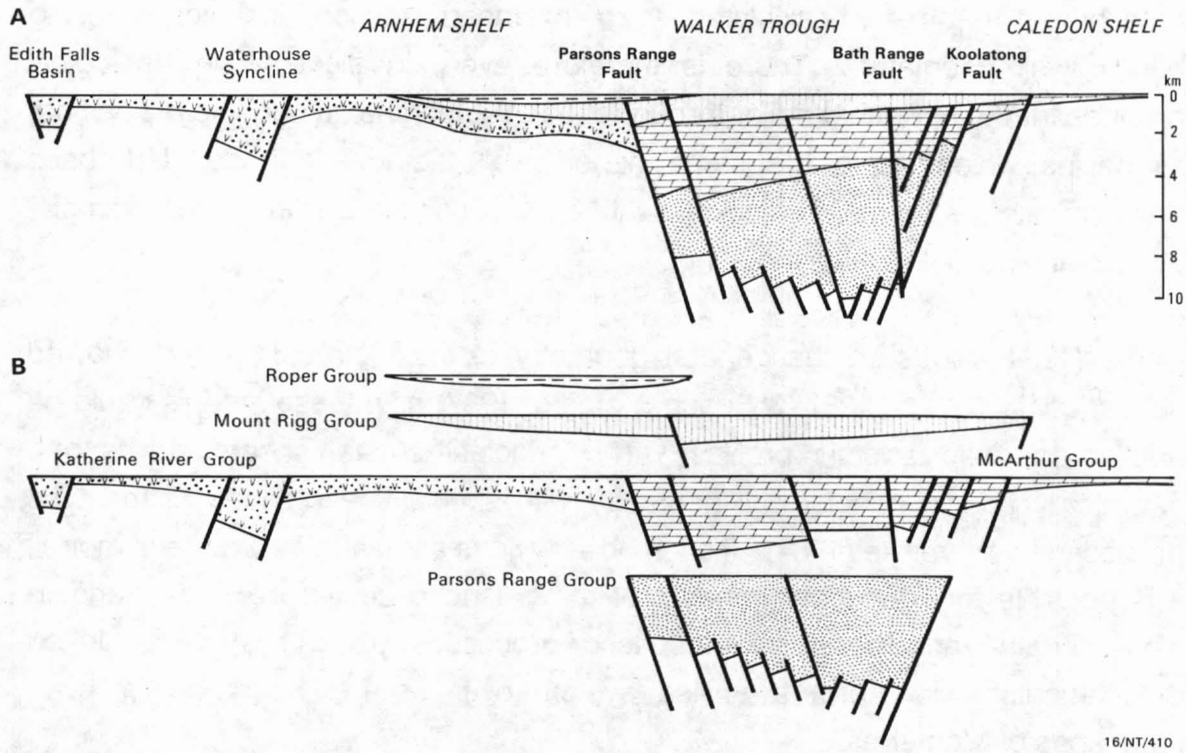


Figure 5. Schematic development of the northern McArthur Basin, from the Arnhem to Caledon Shelves, through the Walker Trough.

A. Schematic cross-section at the close of Roper Group sedimentation.

B. Schematic development of the principal depositional units.

Hangingwall geometry of normal faults

A Gibbs, Midland Valley Exploration Ltd

Hangingwall geometry is a direct consequence of the ramp and flat geometry of the footwall and can be predicted using any of the standard techniques of balanced section construction. The sedimentary and structural geometries of the emergent ramps in a half-graben extensional basin can be understood using the same model of linked fault systems and contrasted with the stratigraphic response of linked unfaulted basins above a concealed ramp.

Deep seismic reflection profiling has helped provide new evidence for such crustal geometries by imaging the geometry of the lower crust. The linkage between upper crustal processes and the deep crust is, however, not consistently seen on such data, and models have to be built which are consistent with upper and lower crustal geometry. The work of the BIRPS group in the UK has been particularly successful.

On these profiles, a continuous reflector linking the structure of the upper crust to that of the lower crust has not yet been observed as a single reflector. In some cases discontinuous groups of reflectors align, running in a dipping zone from upper to lower crust (Fig. 1). This geometric alignment is coincident with the detachment geometry calculated from balanced section techniques using hangingwall geometry. There is therefore every justification on geological grounds for the belief that major through-crustal detachments underlie many if not all basins which share the same hangingwall features. In the UK these detachments are normally coincident with older lineaments defining, for example, Hercynian and Caledonian thrusting.

Gibbs (1984) modelled the general geometry of such structures and showed subsequently (Gibbs, 1987, & Beach & others, 1987) how these models could be applied to deep seismic profiles. The principal crustal shears correspond seismically and in hangingwall geometry with simple shear zones of the type proposed by Wernicke (1984). If such observations are taken as a general model, it is possible to extend the concept of the balanced section to understanding whole crustal and possibly lithospheric processes on a system of linked detachments which form a unified system over several hundreds or even thousands of kilometres.

Clearly the isostatic and thermal effects of such linked systems will eventually be critical to understanding and modelling their behaviour. In the first instance, however, there can be derived a purely mechanical model which has major implications to our understanding of structural growth of the basin, and stratigraphic response and sedimentary history. In this contribution I discuss the development of such a system.

Figure 2 shows the simplest form of a linked crustal detachment comprising an emergent extensional fault linked to a flat, and one or more, mid-crustal ramps before detaching on the base of the brittle crust. The most obvious element of this system is the half-graben basin defined by the emergent ramp. Additional synthetic arrays and antithetic faults will be present in almost all cases. Each will exhibit a structural/stratigraphical facies similar to that illustrated in Figure 3. Individual facies belts may be suppressed or absent depending on the rate of faulting, subsidence and eustatic changes in depositional base.

The general geometry of the half-graben as it grows by continued extension will be of off-lap on the distal hangingwall, and stacking of the growth wedges sequentially towards the emergent fault. The position of off-lap is defined by the point at which the listric fault merges with the detachment parallel to the "regional". This position is fixed relative to the footwall and forms a "rolling hinge" in the hangingwall. In cases where there is no mid-crustal flat the hinge will be positioned over the link between the listric ramp fault and the flat detachment (Fig. 4). Even where isostatic effects such as footwall uplift, or thermal effects, are taken into account, the "rolling hinge" is still fixed relative to the footwall in this way, but its record in the stratigraphic record will migrate in response to changes in footwall rather than hangingwall geometry. The simple off-lap stratigraphic response may be masked by subsidence but the syn-sedimentary position of the hinge will inevitably be associated with facies boundaries etc., and its identification is of prime importance to the exploration geologist.

In petroleum exploration the identification of the rolling hinge is the key to understanding stratigraphic plays on the distal roll-over. Where major basin systems have an asymmetry controlled by a through-crustal shear, mapping of the evolution of this hinge and modelling its stratigraphic response is a key element in formulating successful exploration potential. Beach (1986) pointed out the apparent asymmetry by one order of magnitude of the petroleum discovery sizes and numbers in the North Sea on the UK sector (emergent ramp) and Norwegian (crustal hinge) sectors.

Figure 5 shows how hangingwall basins occur above buried ramps. These basins may be of virtually any size ranging from a few metres across in the case of a detachment fault in the sedimentary carapace within one of the extensional compartments of Fig. 2 (see also Gibbs, 1987) to tens or hundreds of kilometres across for a crustal-scale ramp basin. Throughout the size range there is a common family of geometric components. Figure 4 shows these. Again the basin is tied to a fixed position relative to the footwall and hence the proximal basin margin is dominated by overstep and the distal margin by off-lap in a position homologous with the rolling hinge of the emergent ramp basin.

Stratigraphic wedges for each extensional increment of the ramp basin are stacked laterally (Fig. 5). Each wedge will be asymmetric in overstepping on to its proximal margin and off-lapping on its distal margin. For each major sedimentary province these geometries will hold although the lithology present may be dominated by carbonates in one case or clastics in another.

An important feature of ramp basins is that they may have no significant faulting and may have been previously misidentified as thermal or flexural basins. Flexural and purely thermal basins should be distinguishable on the basis of this simple geometric model and will lack the fundamental growth asymmetry of ramp basins.

Where the detachment fault has a major mid-crustal ramp, the ramp may become the site of a short-cut or breakthrough fault (e.g., Gibbs, 1984). The short-cut fault will often be on the site of the proximal margin of the former ramp basin (Fig. 6). With footwall uplift a new ramp may form, bypassing the upper extensional system (see Wernicke (1986) and Gibbs (1984). On a linked system (as in Figure 6) a very deep basin may form on such a breakthrough fault and a chain of upper detachment half-grabens may end in an anomalously-deep basin. This situation may account for the change in geometry from the Shetland Terraces to the Viking Graben axial basin (e.g., Gibbs, 1987). With detachment geometries such as shown, equal increments of extension will create very different amounts of fault subsidence for differing detachment depths (following the relationship A = e.d).

Where the breakthrough basin forms on the site of an earlier-stage ramp basin the breakthrough fault will normally form on the overstep proximal margin of the basin. Hence the stacked basin of the ramp stage will lie in the hangingwall of the new fault and the up-dip early basin margin will now be structurally down-dip. This leads to apparent inversion of the proximal margin of the early-formed ramp basin (Fig. 6) and the superposition of emergent ramp stratigraphic geometries.

These simple structural relationships are the key to successful exploration in both extensional and inverted basins (Fig. 7). Modelling various ramp-flat combinations with their erosional and depositional response and the timing of breakthrough faults is a powerful tool in understanding the complex stratigraphy of basins. Where sufficient data exist enabling the construction of balanced cross-sections, forward modelling can be used to test the implications of alternative fault-linkages and to predict the likely sequence of stratigraphic units. Work on detailed mapping and modelling of stratigraphy is now clearly a major opportunity to calibrate and test structural models derived from both geophysical and theoretical studies. The necessary criteria to "finger-print" different combinations of structural linkage and thermal and isostatic effects will prove a challenge in basin modelling in the next few years.

References

Beach, A., 1986 - A deep seismic reflection profile across the northern North Sea. Nature, 323, 53-55.

Beach, A., Bird, T., & Gibbs, A., 1987 - Extensional tectonics and crustal structure: deep seismic reflection data from the northern North Sea Viking graben. In Cowards, M.P., Dewey, J.F., & Hancock, P.L., (Editors) - Continental extensional tectonics. Geol. Soc. London Special Publ. 28, p. 467-476.

Gibbs, A.D., 1983 - Balanced cross-section construction from seismic sections in areas of extensional tectonics. J. Struct. Geol. 5, 153-160.

Gibbs, A.D., 1984 - Structural evolution of extensional basin margins. J. Geol Soc. London, 141, 609-620.

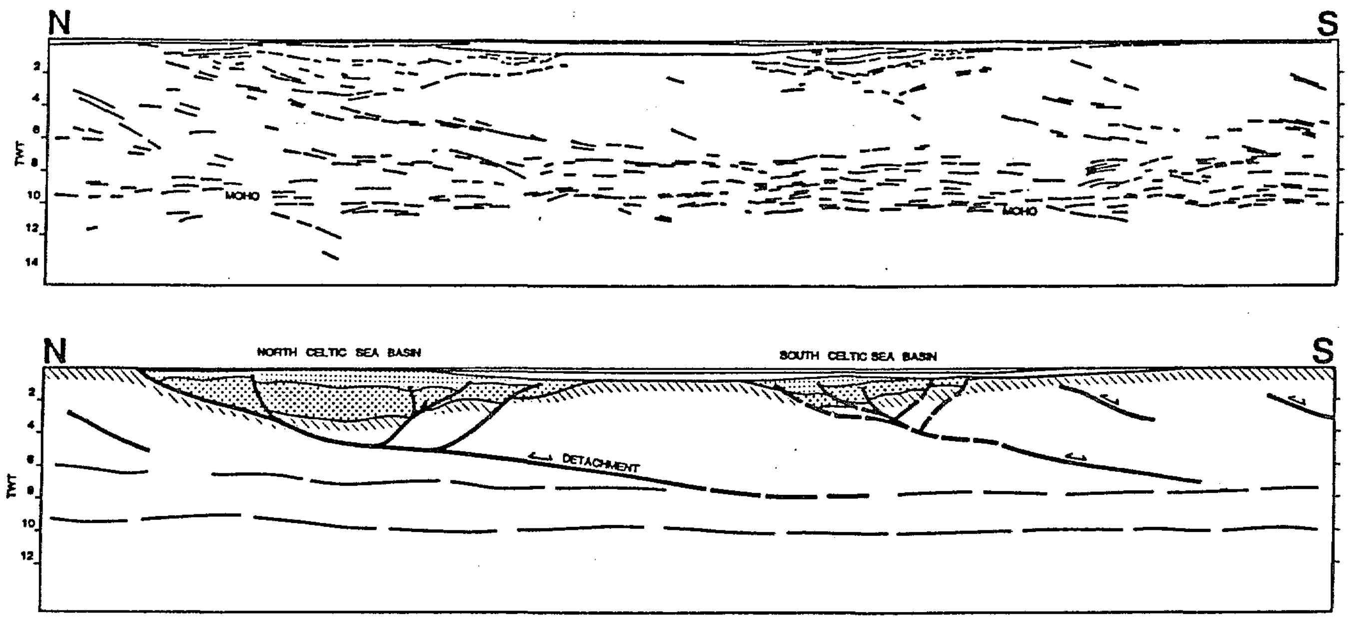
Gibbs A.D., 1987 - Development of extensional and mixed mode sedimentary basins. In Coward, M.P., Dewey, J.F., & Hancock, P.L., (Editors) - Continental extensional tectonics. Geol. Soc. London Special Pub. 28, 19-33.

Gibbs, A.D., 1987 - Basin development, examples from the UK and comments on hydrocarbon prospectivity. Tectonophysics, 133, 189-198.

Wernicke, B., 1985 - Uniform sense of normal simple shear of the continental lithosphere. Can. J. Earth Sci., 22, 108-125.

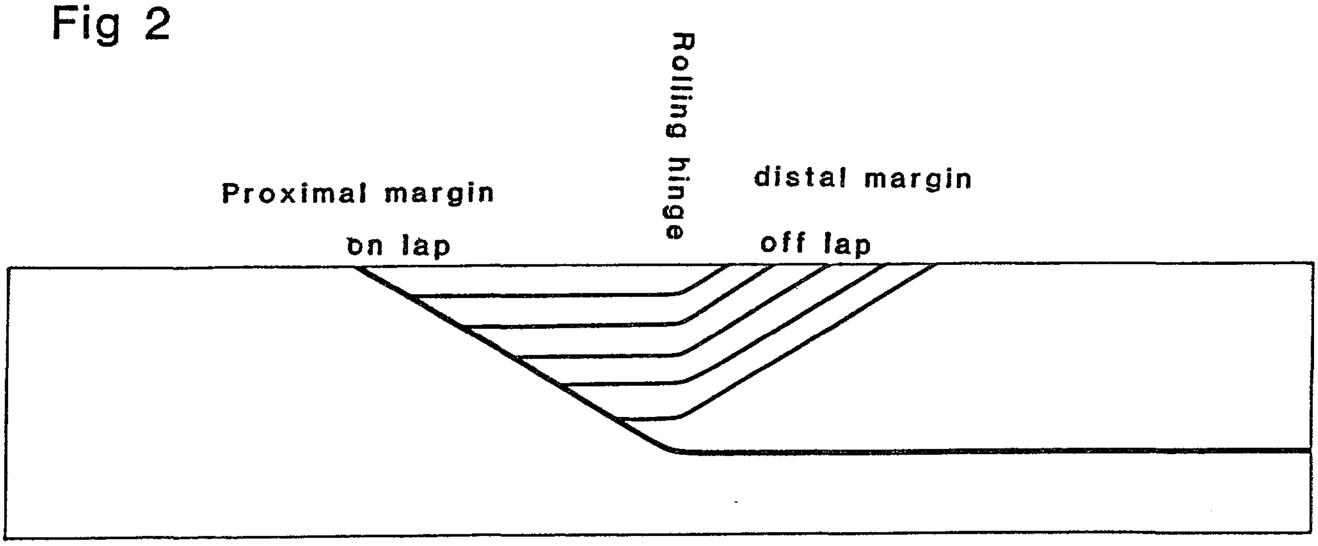
Wernicke, B. & Burchfiel, B.C., 1982 - Modes of extensional tectonics. J. Struct. Geol. 4, 105-115.

Fig 1



SWAT 4 PROFILE

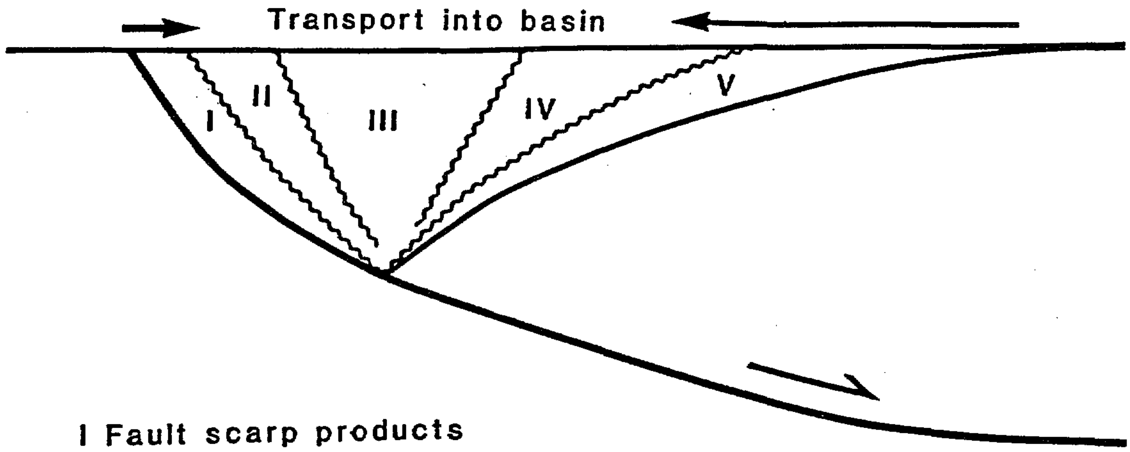
Fig 2



Asymmetry over simple ramp flat system

Fig 3

Structural / Stratigraphic Facies



I Fault scarp products

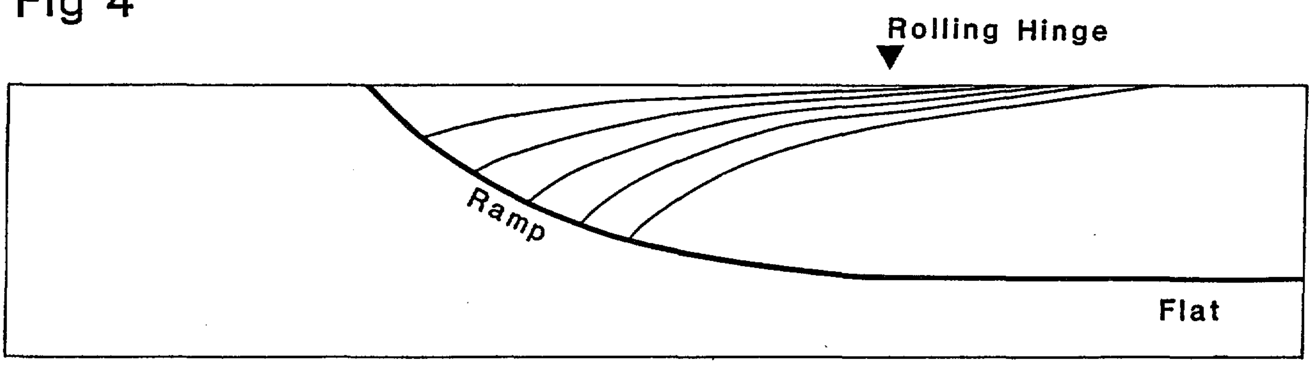
II Tectonic depocentre

III Compactional and axial depocentre system

IV Build ups prograding down slope

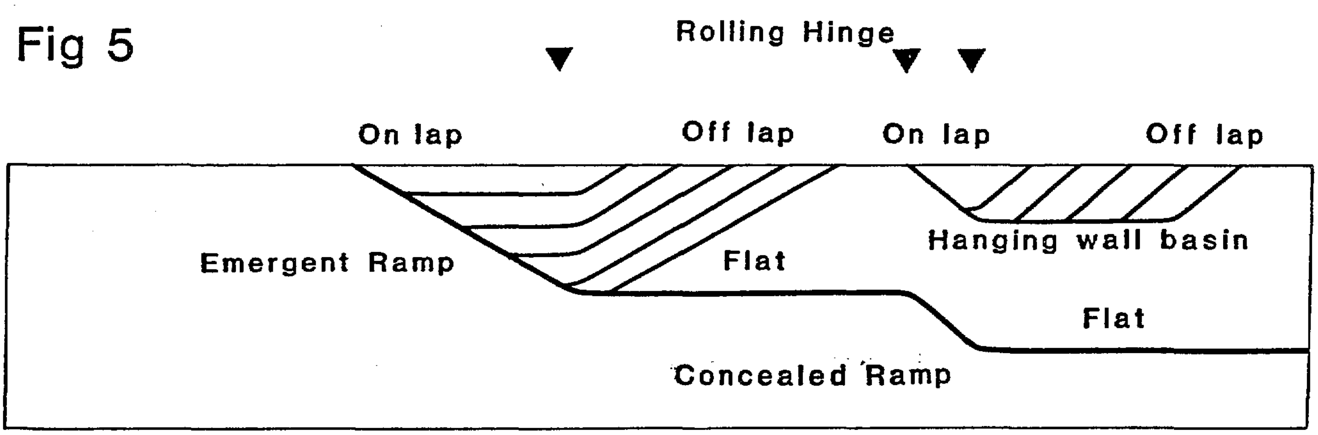
V Main transport slope mixed erosion and build-ups

Fig 4



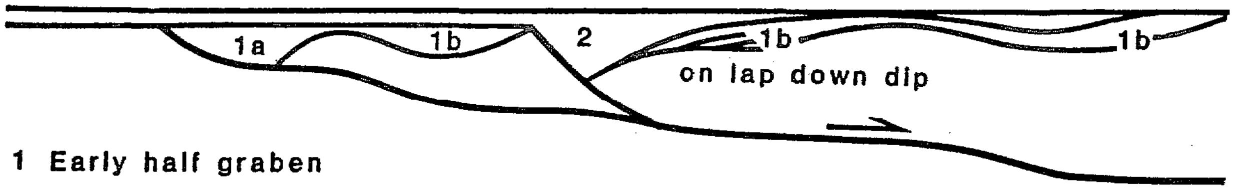
Listric ramp with asymmetric on and off lap basin margins

Fig 5



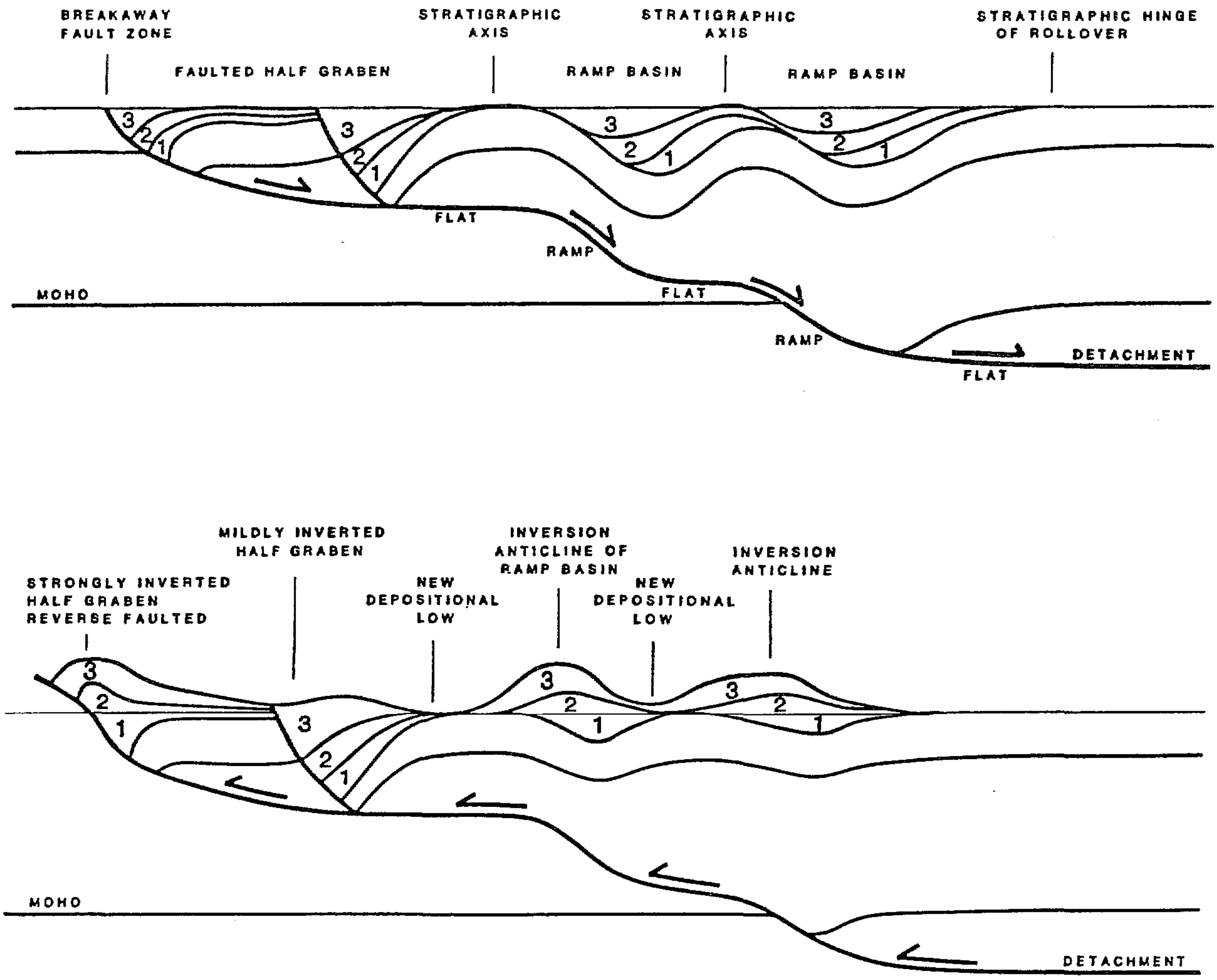
Asymmetric linked basins

Fig 6



1 Early half graben
1b Hanging wall basins
2 Basin formed on break through fault from ramp

Fig 7



Extensional models for passive margin evolution

M A Etheridge, BMR, G S Lister, Monash University, & P A Symonds, BMR.

The asymmetric detachment model of continental extension that has been developed largely from studies in the Basin and Range province (Wernicke, 1981, 1985; Wernicke & Burchfiel, 1982; Reynolds & Rehrig, 1981; Lister & Davis, in press; Davis & Lister, in press) is applied to the evolution of passive continental margins. The inherent asymmetry of extension by detachment leads to two broad classes of passive margins - upper plate margins comprising rocks above the detachment, and lower plate margins comprising the footwall of the detachment commonly overlain by faulted upper plate remnants (Fig. 1). Upper and lower plate margins have distinctive architectures, structural styles, uplift/subsidence paths and thermal histories. However, their specific character depends to a large extent on the detailed geometry of the detachment(s) and on the amount and distribution of ductile pure shear stretching beneath the detachment.

Five separate detachment models, incorporating ramp/flat geometries and sub-detachment stretching are presented, together with their consequences for passive margin evolution (Fig. 2). These models, together with allowance for multiple detachments, provide a rational explanation for many of the previously enigmatic structural and morphological features of passive margins, such as marginal plateaux, outer rises, unstructured syn-rift sag basins, perched rift basins and the wide range in observed structural styles. Transfer faults (Bally, 1981; Gibbs, 1984, 1987; Etheridge & others, 1985; Etheridge, this symposium) accommodate substantial variation in the along-strike architecture of passive margins, including switches in detachment dip-direction and therefore in basic asymmetry (Fig. 3; Lister & others, 1986).

Numerical modelling demonstrates that the wide variation in uplift/subsidence behaviour and syn-rift thermal effects observed around passive margins can be explained by varying the detachment geometry and sub-detachment pure shear in relatively simple fashion (Fig. 4). Specifically, upper plate margins undergo substantial uplift, which can partly survive post-extension thermal subsidence if crustal underplating by mantle-derived melts takes place, providing an explanation for the widespread passive margin mountains (e.g., Eastern Australian Highlands). Marginal plateaux can result from a flat, mid-crustal detachment below a weakly-extended upper plate, and are shown to be emergent or very shallowly submerged throughout much of the extension, with post-rift subsidence to intermediate water depths. The subsidence of lower plate margins depends largely on the distribution of stretching above and below the detachment, providing an explanation for the combination of rift basins perched on the continent, eroded but submerged tilt blocks and rapidly subsided segments on a single lower plate margin (e.g., Carolina Trough segment, eastern USA, Lister & others, in press). The modelling also predicts horizontal offset of upper crustal stretching (rift basins) from deeper stretching and consequent post-rift subsidence and thermal maturation (e.g., Jerboa-1, Australian southern margin). None of the models, however, accounts easily for the widespread evidence of the emergence of lower plate margins throughout and even following extension, and additional buoyancy derived from such processes as underplating, secondary convection or mantle plumes is required.

Examples of a classical lower plate margin (Australian southern margin, Willcox & others, this symposium) and a complex upper plate margin (Queensland Plateau, Symonds & others, this symposium) are interpreted in terms of the detachment model elsewhere in this symposium. The complementary asymmetry of conjugate margin pairs is illustrated from the North and South Atlantic and the Tasman Sea. The North Atlantic between eastern USA and northwest Africa also demonstrates the switching of asymmetry across major transfer faults.

The detachment model is presented in two-dimensional form; however, oblique extension is likely to be widespread, and three-dimensional analysis of the extensional structure should always be undertaken. The detachment models presented here have a number of other important implications; 1) magnetic quiet zones can be explained in large part as highly attenuated (and possibly underplated) lower plate pulled from beneath the conjugate margin; 2) matching of major transfer faults and careful overlapping of conjugate upper and lower plate segments provides a superior means of precise continental reassembly; 3) continental transfer faults provide important controls on the location of oceanic transform faults.

The detachment model has important implications for hydrocarbon exploration of passive margins and extensional basins, because of the close relationship between the specific detachment and pure shear geometry of a terrane and its structural, stratigraphic, palaeogeographic, subsidence and thermal evolution (Gibbs, 1984, 1987; Harding, 1984; Etheridge & others, 1985). In particular, (i) hydrocarbon maturation is related primarily to the mantle-derived thermal anomaly which will commonly be offset from the surface rift basin, leading to immature rifts with otherwise excellent petroleum prospects; (ii) the half-graben style and transfer fault compartmentalisation of rift basins provide fundamental controls on both source and reservoir facies distributions; (iii) The characteristic normal and transfer fault geometries developed during asymmetric continental extension impact on seismic survey design, seismic interpretation concepts, and the development of post-extensional structures which commonly provide the most important hydrocarbon traps.

References

Bally, A.W., 1981 - Atlantic-type margins. In Geology of passive continental margins. American Association of Petroleum Geologists, Education Course Note Series, 19, 1/1-48.

Bosworth, W., 1985 - Geometry of propagating continental rifts. Nature, 316, 625-627.

Coward, M.J., 1986 - Heterogeneous stretching, simple shear and basin development. Earth plan. Sci. Letters, 80, 325-336.

Crittenden, M.D., Coney, P.J., & Davis, G.H. (Editors) 1980 - Tectonic significance of metamorphic core complexes of the North American Cordillera. Mem. Geol. Soc. Amer. 153.

Davis, G.A., & Lister, G.S., - Detachment faulting in continental extension: perspectives from the southwestern U.S. Cordillera. In John Rodgers Symposium Volume, Mem. Geol. Soc. Amer., (in press).

Etheridge, M.A., Branson, J.C., & Stuart-Smith, P.G., 1985 - Extensional basin-forming structures in Bass Strait and their importance for hydrocarbon exploration. Australian Petroleum Exploration Association Journal, 25, 344-361.

Etheridge, M.A., Branson, J.C., & Stuart-Smith, P.G., 1987 - The Bass, Gippsland and Otway Basins, southeast Australia: a branched rift system formed by continental extension. In Beaumont, C., & Tankard, A., (Editors) - Sedimentary basins and basin-forming mechanisms. Canadian Society of Petroleum Geologists Memoir 12.

Frost, E.G., and Martin, D.L. (Editors) 1982 - Mesozoic-Cenozoic tectonic evolution of the Colorado river region, California, Arizona, and Nevada. In The Anderson-Hamilton Volume, Cordilleran Publishers, San Diego, California.

Gibbs, A.D., 1984 - Structural evolution of extensional basin margins. Journal of the Geological Society of London, 141, 609-620.

Gibbs, A.D., 1987 - Development of extension and mixed-mode sedimentary basins. In Coward, M.P., Dewey, J.F. & Hancock, P.L. (Editors) - Continental extension tectonics. Geological Society Special Publication, 28, 19-33.

Harding, T.P., 1984 - Graben hydrocarbon occurrences and structural style. J. Amer. Assoc. Petrol. Geol., 68, 333-362.

Keen, C.E., Beaumont, C., - Geodynamics of rifted continental margins. Geol. Soc. Amer. DNAG special publ. (in press).

Lister, G.S., & Davis, G.A., - A model for the formation of metamorphic core complexes and mylonitic detachment terranes. Tectonics (in press).

Lister, G.S., Etheridge, M.A., & Symonds, P.A., 1986 - Application of the detachment fault model to the formation of passive continental margins. Geology, 14, 246-250.

Lister, G.S., Etheridge, M.A., & Symonds, P.A. - Detachment model for the formation of passive continental margins. Tectonics (in press).

Rosendahl, B.R., - Architecture of continental rifts with special reference to east Africa. Earth plan. Sci. Lett., (in press).

Tankard, A.J., & Welsink, H.J., - Extensional tectonics and stratigraphy of the Hibernia oil field, Grand Banks of Newfoundland. J. Amer. Assoc. Petrol. Geol. (in review).

Wernicke, B., 1985 - Uniform-sense normal simple shear of the continental lithosphere. Can. J. Earth Sci., 22, 108-125.

Wernicke, B., & Burchfiel, B.C., 1982 - Modes of extensional tectonics. J. Struct. Geol., 4, 105-115.

Wernicke, B., 1981 - Low-angle normal faults in the Basin and Range Province: Nappe tectonics in an extending orogen. Nature, 291, 645-648

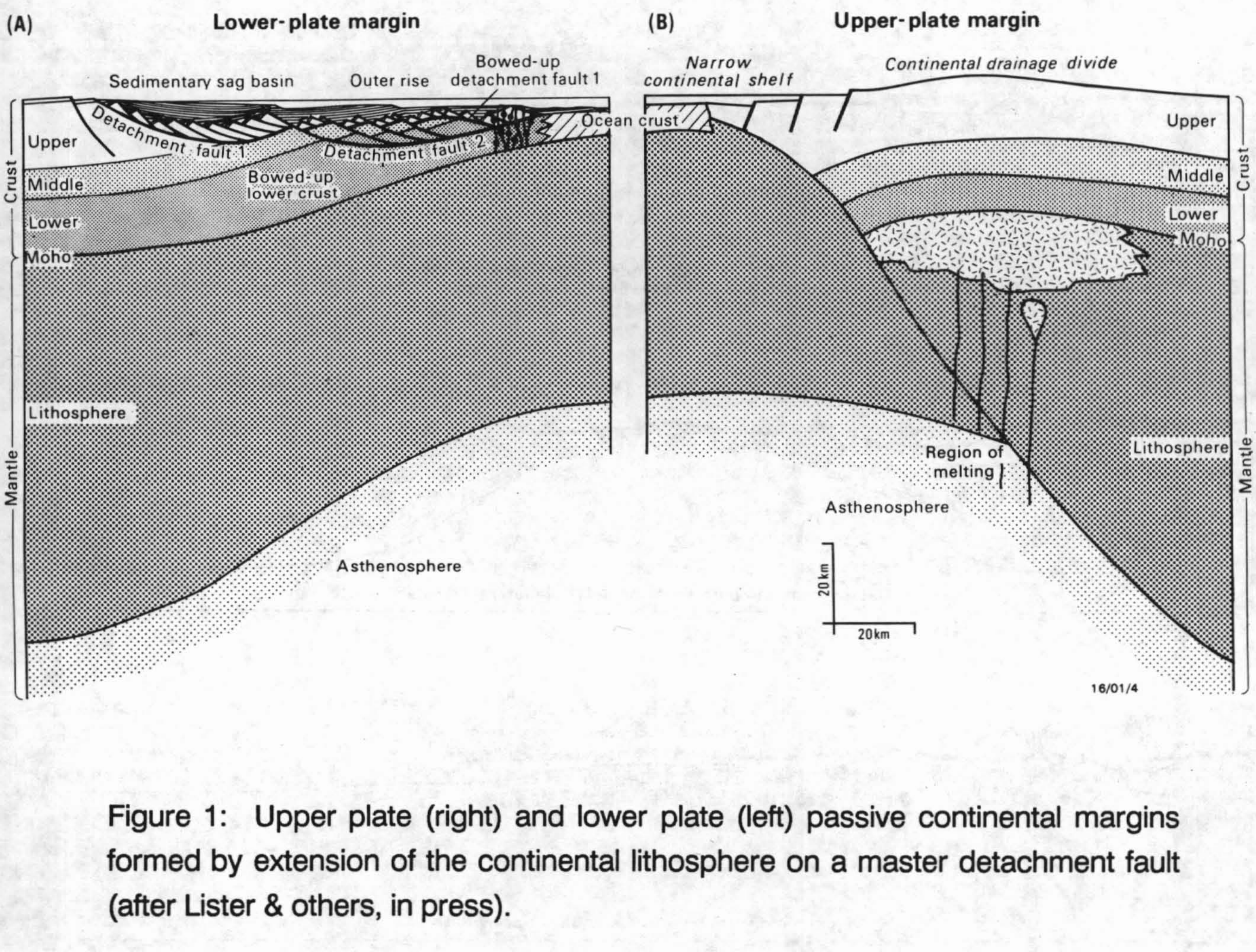


Figure 1: Upper plate (right) and lower plate (left) passive continental margins formed by extension of the continental lithosphere on a master detachment fault (after Lister & others, in press).

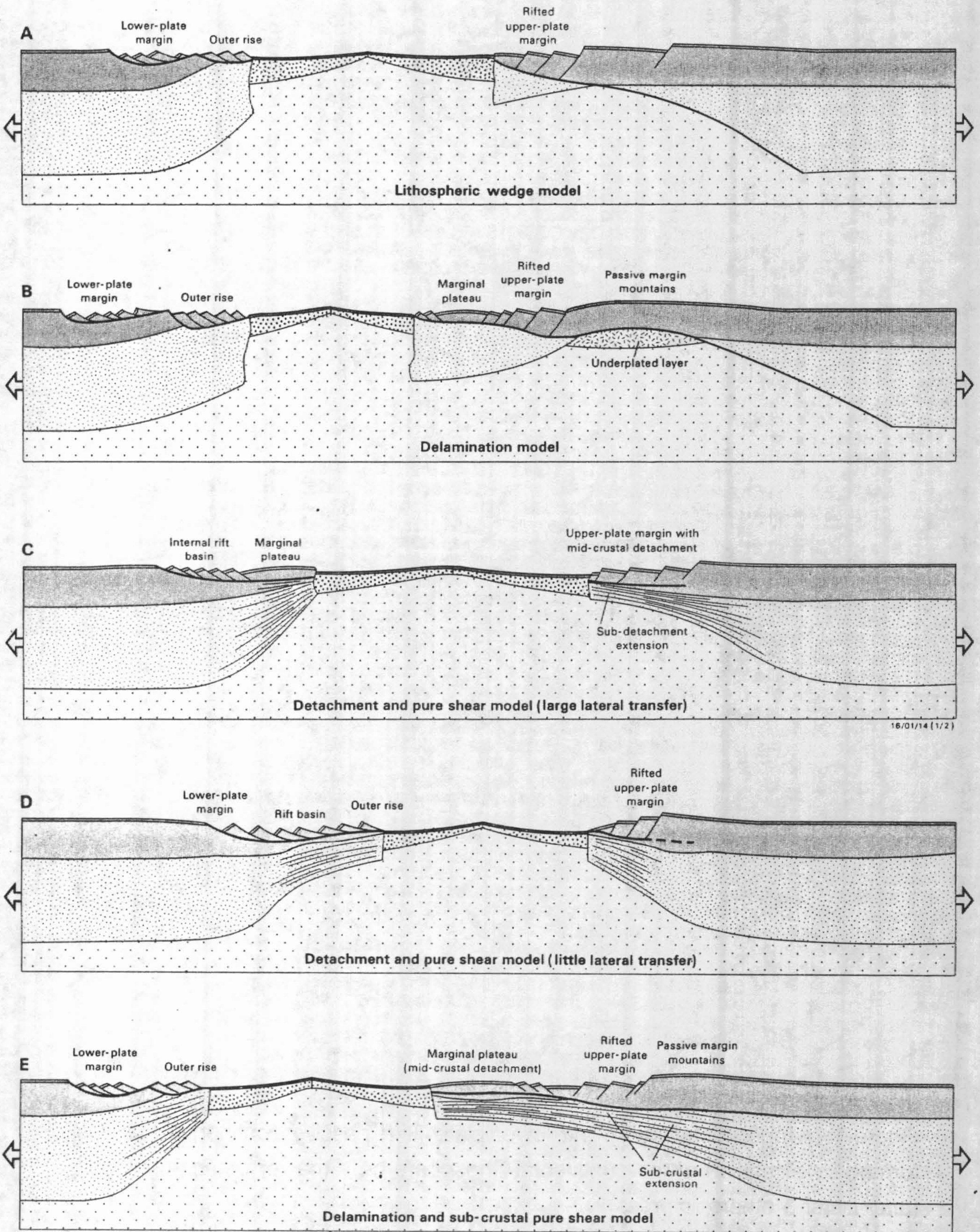
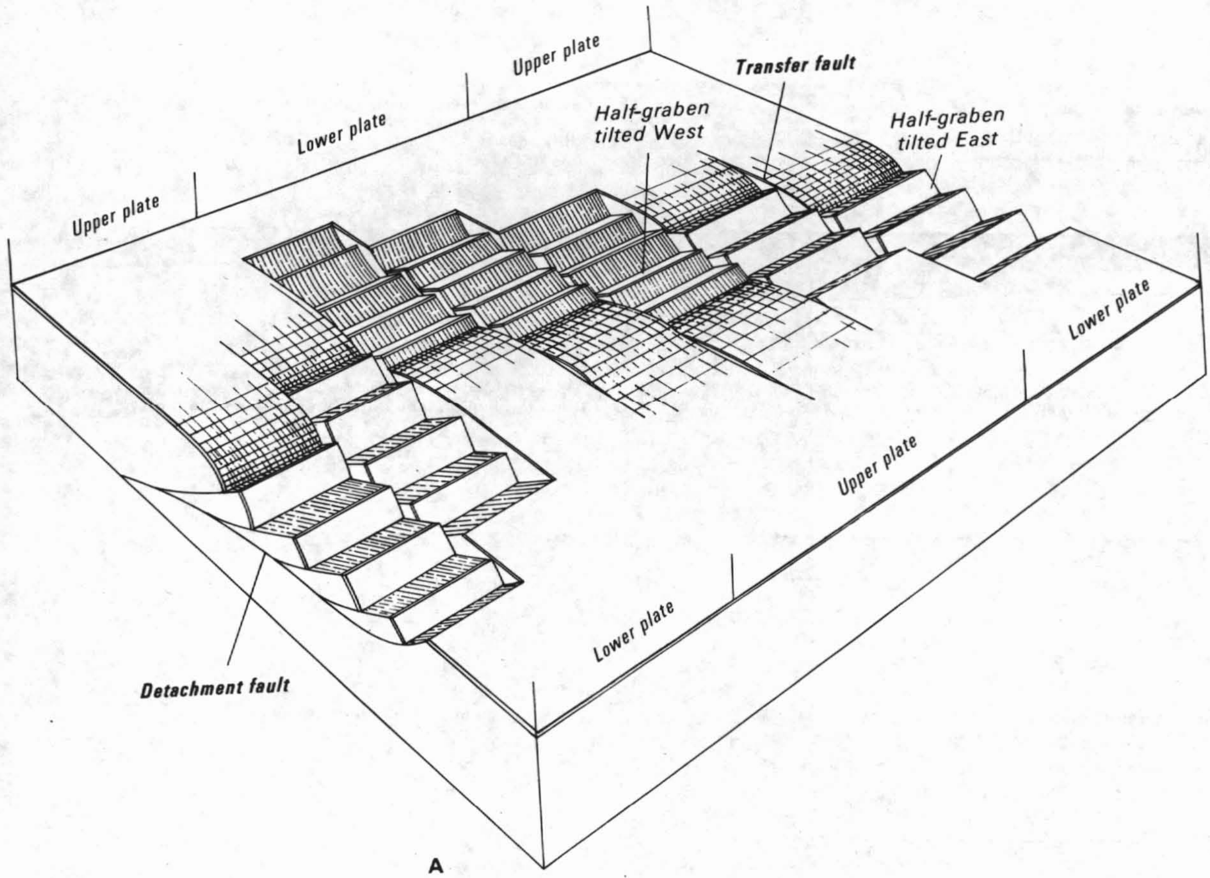


Figure 2: Examples of the potential variation in passive margin architecture that can result from different detachment geometries and from different distributions of ductile stretching below the detachment. Note the appearance of marginal plateaux above flats on the detachment (after Lister & others, in press).



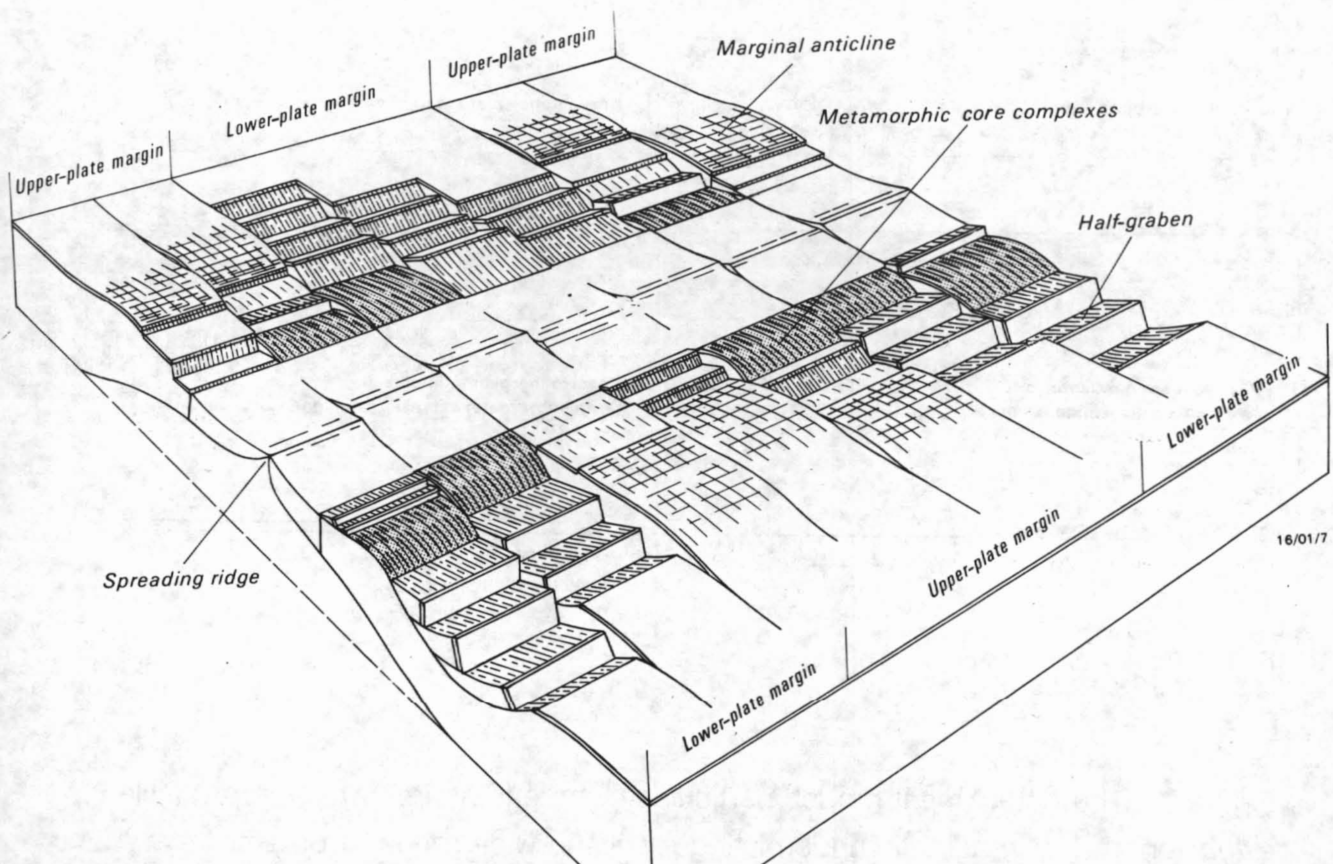


Figure 3: Transfer faults play a key role in accommodating variations in the along-strike geometry of passive margins and continental rifts, including substantial offsets and switches in asymmetry (after Lister & others, 1986).

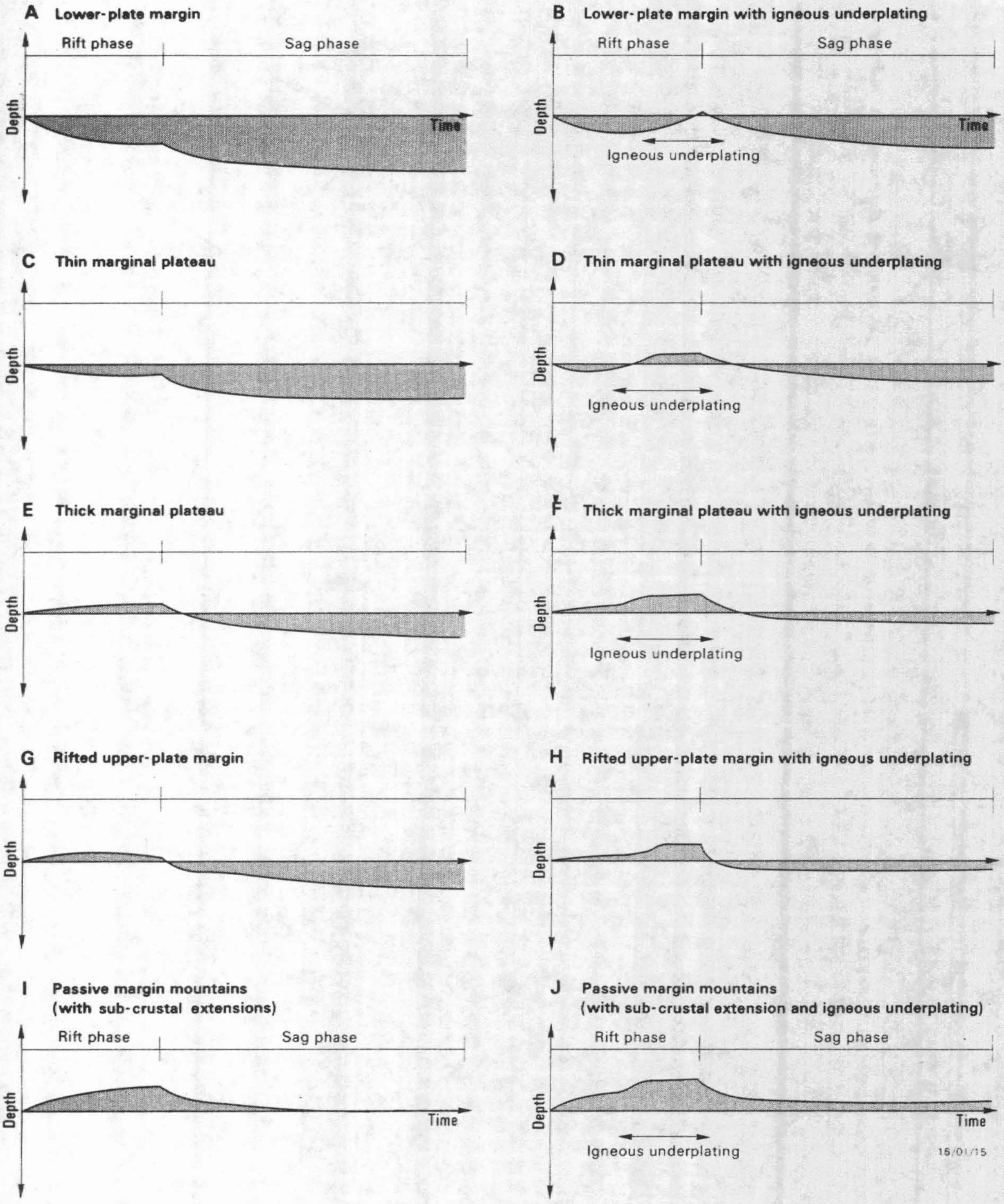


Figure 4: The detachment model, particularly where it incorporates deep ductile extension and/or magmatic underplating results in wide range of potential syn- and post-rift uplift/subsidence histories. These examples are based on one-dimensional modelling of the major structural environments illustrated in Figure 2 (after Lister & others, in press).

Petroleum source rock facies in extensional basins: relevance to Australian petroleum geology

T G Powell & M Bradshaw, BMR

Deep stratified water bodies can occur in both marine and continental rift systems. The anoxic bottom conditions lead to the preservation of organic matter and the formation of thick source beds. The East African Rift provides a modern analogue for lacustrine source rocks in ancient rift. Deep starved lakes occur at the 'Mature Half-Graben' stage of rift development (Rosendahl, 1987). Subsidence outstrips sediment supply since drainage is directed away from the rift by back-tilting blocks along the rift margin. In tropical lakes, oxygen contents are low because of high temperatures and there is no seasonal overturn. Salinity stratification is common and the hypolimnion can remain permanently anoxic resulting in the deposition of organic-rich shales and marls (15 per cent TOC in Lake Kivu). Carbon contents in ancient lacustrine source rocks vary widely depending on sediment supply. Organic matter types range from II/III for clastic lacustrine source rocks to Type I in source rocks deposited in deep, starved lakes (Powell, 1986). The Jizhong Depression, China, is an example of a lacustrine petroliferous system in a Tertiary rift regime (Wu & Liang, 1984). Up to 5 km of Eocene to Oligocene lake sediments were deposited in half-grabens. The source rock is up to 400 m thick and contains low amounts (circa 1 per cent) of Type II/III organic matter. Hydrocarbon reservoirs occur in adjacent basement horst blocks and overlying sands.

Marine petroliferous rifts occur in the North Sea and the offshore Jeanne D'Arc Basin, Canada. In the North Sea, the thick Kimmeridgian source rock is confined to the graben complex (Demaison & others, 1984). The source facies is juxtaposed to reservoirs in submarine fans along the rift margin (e.g. Brae Field). In the Jeanne D'Arc Basin, the rapid deepening of the rift in the Kimmeridgian (Tankard & Welsinck, 1987) coincided with a relatively high sea-level stand resulting in the formation of an anoxic basin (Powell, 1985). Transfer faults controlled the bottom topography and the distribution of the source rock. In both examples, the rifts developed within an epeiric sea which minimised clastic input during their initial phases. Organic carbon contents are high (up to 10 per cent) and the organic matter is Type II.

Australia offers examples of source rocks deposited in both marine and continental rifts. In the late Jurassic, Australia was still part of Gondwana, being attached to Antarctica via a continental rift valley along the southern margin. On the Northwest Shelf, continental breakup had commenced in the Callovian (Veenstra, 1985) and inboard of the new seafloor, deep marine troughs occupied failed rifts and sags (Kopsen & McGann, 1985). Although Australia occupied high southerly latitudes, climates were temperate and humid. Seismic sections from the Eyre Sub-basin (Bein & Taylor, 1981) on the southern margin show thick wedges of Jurassic sediments in fault-controlled half-grabens. Jurassic sediments have been found in only a few wells and consist predominantly of dark carbonaceous clays and lignites deposited in lacustrine and fluvio-lacustrine environments. The existence of proven fluvial-lacustrine source rocks in the Jurassic of the Eromanga Basin, Queensland, (Vincent & others, 1985) is evidence that the Jurassic climate and vegetation of Australia was suitable for the formation of continental oil-prone source rocks. However, it would be expected that lakes formed on the southern margin would be deeper and more persistent and would be capable of accumulating greater thicknesses of source rocks, than the ephemeral lakes of the Eromanga flood plain. Extensional conditions persisted into the Cretaceous along the southern margin, but sedimentation rates were high particularly in the east. Lacustrine environments favourable for preservation of organic matter probably persisted, but dilution from the large input of clastic sediments is a significant risk. The composition of offshore oil seeps in South Australia in the Otway Basin are consistent with a lacustrine source facies in that area (McKirdy & others, 1986).

On the northwestern margin of Australia, source-rock environments are related to deep, restricted, fault-bounded marine troughs such as the Lewis Trough and Vulcan Graben. The narrow deep troughs were embayments on the southwestern shore of Tethys, partly enclosed by land areas. This geometry in conjunction with active subsidence and a high eustatic sea level produced oxygen-depleted, deep-water environments during the Late Jurassic and especially the Kimmeridgian. Proven source rocks were deposited in these environments as indicated by the hydrocarbon accumulations located around the margins. Very few wells have intersected the Late Jurassic source facies, but the available data show organic carbon contents of 2 to 3 per cent and a Type III+ to Type II kerogen. The high content of terrestrial organic matter in the sediments of the Lewis Trough is a consequence of high clastic sedimentation rates in this land-locked embayment.

References

Bein, J., & Taylor, M.L., 1981 - The Eyre Sub-basin: recent exploration models. APEA Journal, 21(1), 91-98.

Demaison, G., Holck, A.J.J., Jones, R.W., & Moore, G.T., 1984 - Predictive source bed stratigraphy; A guide to regional petroleum occurrence. Proc. Eleventh World Petroleum Congress, 2, 17-29.

Kopsen, E., & McGann, G., 1985 - A review of the hydrocarbon habitat of the eastern and central Barrow-Dampier Sub-basin, Western Australia. APEA Journal, 25(1), 154-176.

McKirdy, D.M., Cox, R., Volkman, J., & Howell, V., 1986 - Botryococcane in a new class of Australian non-marine crude oils. Nature, 320, 57-59.

Powell, T.G., 1985 - Palaeogeographic implications for the distribution of Upper Jurassic source beds, offshore eastern Canada. Bull. Canadian Petroleum Geology, 33, 116-119.

Powell, T.G., 1986 - Petroleum geochemistry and depositional setting of lacustrine source beds. Marine and Petroleum Geology, 3, 200-219.

Rosendahl, B.R. 1987 - Architecture of continental rifts with special reference to east Africa. Ann. Rev. Earth & Planetary Sciences, 15, 445-503.

Tankard, A.J., & Welsinck, H.J., 1987 - Extensional tectonics and stratigraphy of the Hibernia oilfield, Grand Banks of Newfoundland. AAPG Bulletin (in press).

Veenstra, E., 1985 - Rift and drift in the Dampier Sub-basin, a seismic and structural interpretation. APEA Journal, 25(1), 154-176.

Vincent, P.W., Mortimore, I.R., & McKirdy, D.M., 1985 - Hydrocarbon generation, migration and entrapment in the Jackson-Naccowlah Area, ATP 259P, southwestern Queensland. APEA Journal, 25 (1), 62-84.

Wu, H., & Liang, D., 1984 - Analysis of oil formation of three oil fields in Jizhong Depression. Beijing Petroleum Symposium, September 20-26, Beijing, China. Circum-Pacific Council of Mineral & Energy Resources. Preprint.

Fluid dynamics in extensional basins

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Basins produce a variety of resources that are related by the fact that all formed as the result of fluid movements. Since all basin resources reflect the basic fluid-flow processes that operate in basins and the distribution of the more permeable geologic units within basins (sediments or structures), all can contribute information that is useful in exploring for the others. In this sense the traditional division between hard- and soft-rock geology has been unfortunate since it has probably impeded development of a clear view of the movement of fluids in basins.

The purpose of this paper is to stimulate discussion by suggesting: (1) that basin resources fit a context of basin evolution as shown in Table 1, (2) that convection and compactive expulsion are the fluid-drive mechanisms that are economically important, and (3) that basin resources reflect characteristics of the "high permeability" parts of a basin as indicated in Table 2.

A simple formula that describes temperature perturbations in a basin caused by fluid flow through a shallow-dipping network of linear fault and aquifer segments is used to reach several fluid dynamic conclusions and insights. The formula and methodology may be useful to the investigation of many other basin questions.

The proposed relation between basin resources, basin evolution, fluid-drive mechanism, and fluid type is summarised in Table 1. It should be particularly noted that Table 1 does <u>not</u> relate basin resources to topographically-driven flow. Following McKenzie (1978) a fundamental distinction is made between the initial rifting or "tectonic" stage of basin development and the later "thermal" subsidence that is related to lithosphere regrowth. Hydrothermal convection is the economically important fluid-flow process during the tectonic stage. This is discussed in my second talk. Compactive expulsion of pore fluids is the economically significant process in the thermal stage of basin development. This is the subject of the present discussion.

The characteristics of the basin resources produced during the late tectonic or thermal stage of basin development that are most important for understanding basin fluid dynamics are:

Shale hosted massive sulphide deposits (SHMSD): Deposition from ~200°C saline solutions near active faults at the margins of ~60 km-wide, ~5 km-deep grabens. Example: McArthur River deposit in Australia.

Kupferschiefer-type copper-silver-(lead-zinc) mineralisation: Deposition in shale "caps" overlying red-bed-filled grabens. Disseminated and antitaxial (hydrofracture) veinlet mineralisation. Additional example: White Pine, Michigan.

Hydrocarbons: Expulsion from low-permeability organic-rich units such as shales into permeable units such as sands or fractures when temperatures reach 80-120°C. Migration driven by buoyancy or the "sweeping" effect of other pore fluids in the permeable units to traps.

Mississippi Valley-type (MVT) lead-zinc deposits: Precipitation at the distal margins of mature basins from basin brines with temperatures similar to those in the deeper parts of the associated basins (80-120°C). Deposition in open space within 1 km of the surface. Deposition in pulses separated by sulphide dissolution. Deposits were thermal anomalies at the time of formation but could not have remained thermally anomalous for more than a fraction of a million years (thermal maturity of hydrocarbon constraints).

The fluid-drive mechanism responsible for these resources may be appreciated by analysis of one particularly instructive deposit type- MVT deposits. We ask the simple (and least restrictive) question: How rapidly must basin brines move along a basal aquifer to steadily transmit temperatures typical of the deeper parts of the basin to depths of ~1 km at margins with <1% slope? The question is answered (Fig. 1) by requiring a balance between (a) the heat introduced (or removed) by fluid moving in the aquifer and (b) the heat removed (or introduced) by an elevated (or depressed) geothermal gradient. The heat balance requirements shows the temperature along a straight aquifer segment with shallow dip (<~10°) is determined by the temperature at one location as shown in Figure 2.

The equation can be used to calculate the temperature in a basin underlain by an arbitrarily segmented aquifer with arbitrary fluid throughput by "daisy chaining" the formula. For example: the temperature in the first (inflow segment) is calculated by specifying ambient temperature near the surface (e.g., recharge by ambient temperature waters), temperature in the second segment is calculated using the temperature at the end of the first, etc.

Figures 3-8 show some sample calculations and their consequences. Figure 3 shows that a throughput of ~440 m^3 per million years is required to warm a basin margin as required by MVT deposits. Figure 4 shows this result is quite general (independent of basin depth) provided the unperturbed thermal gradient is $<60^o$/km. Figure 5 shows that a flow rate of 440 m^3 per million years is three orders of magnitude greater than could be provided by steady compactive basin dewatering even if all the compactively-expelled water enters the basal aquifer. Thus MVT deposits require episodic rather than steady compactive dewatering.

Figures 6 and 7 show that cross-basin hydrologic flow will cool as well as warm a basin. Figure 8 shows that cross basin flow rates of the magnitude required by MVT deposits will quickly flush any brine from the basal aquifer, in contradiction of observation. Figure 9 shows the steady-state temperature distribution beneath a basal aquifer subject to cross-basin hydrologic flow. This figure shows that the marginal temperature anomalies are caused by the elevated temperature at shallower depths. These anomalies cannot survive fluid throughflow.

References

Cathles, LM, & Smith, AT, 1983 - Thermal constraints on the formation of Mississippi Valley-type lead-zinc deposits and their implications for episodic basin dewatering and deposit genesis. Economic Geology, 78, 983-1002.

Garven, G, 1985 - The role of regional fluid flow in the genesis of the Pine Point deposit, Western Canada Sedimentary Basin. Economic Geology, 80, 307-324.

Knauth, LP, & Beeunas, MA, 1986 - Isotope geochemistry of fluid inclusions of Permian halite with implications for the isotopic history of ocean water and the origin of saline formation waters. Geochim. et Cosmochim. Acta., 50(3), 419-433.

McKenzie, DP, 1978 - Some remarks on the development of sedimentary basins. Earth and Planet. Sci. Lett., 40, 25-32.

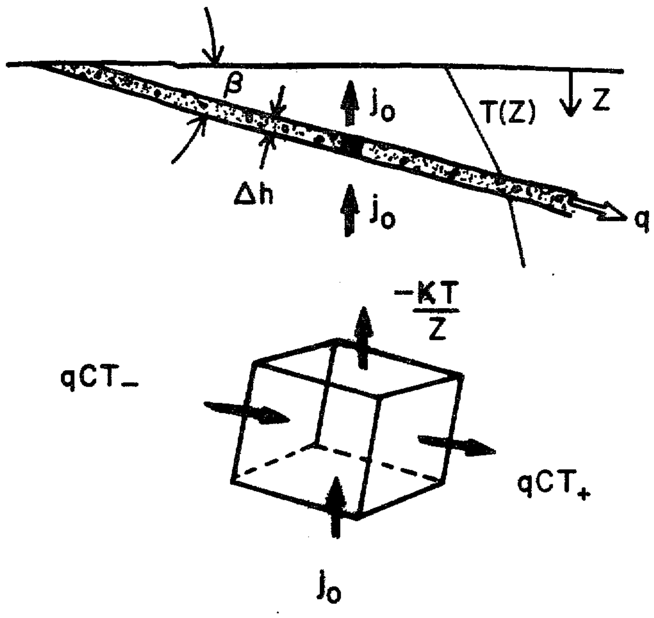
Table 1: Basin resources in relation to the stage of basin development, fluid drive and fluid type

Resource	Tectonic Stage	Flow Drive	Fluid
Au			Fresh
	Tectonic	Convection	
VMSD			Sea Water
SHMSD			
"Kupferschifer"			
MVT	Thermal	Expulsion	Brine
OIL			
Au			Fresh
		Hydrologic	

Table 2: Implications of basin resources for the permeability structure of basins.

Resource	Requirements	Permeability Implications
Au	large volume of hydrothermal discharge at one location	faults that retain permeability over long periods
VMSD	intrusive body with thermally cracked margins	permeability is temperature-dependent
SHMSD	expulsion out border faults from capped graben	permeable border faults with hydrologic connection to basin interior
Kupfersch.	distributed fluid expulsion from margin toward center	provides information on the permeability of a shale cap under geopressured conditions
MVT	rapid expulsion to surface	low ration of volume of high permeability units to volume of dewatering pulse
OIL	accumulation in structures over traps within basin	high ratio of volume of permeable units to volume of dewatering pulse
Au	metamorphism to amphibolite grade	mineralization records expulsion of fluids from geopressure to hydro-pressure

Figure 1: **HEAT BALANCE DETERMINES STEADY STATE TEMPERATURE AT BASIN MARGIN**



$$\boxed{\frac{\partial T}{\partial Z} = \frac{j_o}{K} - \frac{T}{Z} \quad \frac{K}{Qc \tan\beta}}$$

Figure 2

$$\Delta T = (j_o z / K)(\alpha - B\{z_c/z\}^{(\alpha+1)}) / (\alpha+1),$$

$$B = (\Delta T_c / \Delta T_{o,z_c})(\alpha+1) - \alpha$$

$$\alpha = K/Q\,c \tan\beta.$$

Figure 2. Formula giving the temperature distribution along an aquifer segment with fluid throughput Q.

Q = cm^3 of water per cm strike length of aquifer per sec

ΔT_c = integration constant = temperature in °C at z_c

$\Delta T_{o,z_c}$ = "normal" or unperturbed temperature at z_c

K = harmonic average thermal conductivity of the sediments above the aquifer in cal/cm-sec-°C. (= 3.5×10^{-3} in calculations)

j_o = unperturbed or original heat flow in cal/cm^2-sec

c = the heat capacity of water in cal/g-°C

β = dip of the aquifer segment measured positive clockwise from the horizontal projection of the flow direction. β can be negative as well as positive. The resulting α is positive for inflow and negative for outflow. A value of α = -1 must be avoided (by taking α = -1.01 or -.99 for example) but otherwise the equation is quite general.

Figure 3

TEMPERATURE AT TOP OF BASEMENT FOR VARIOUS FLOW RATES

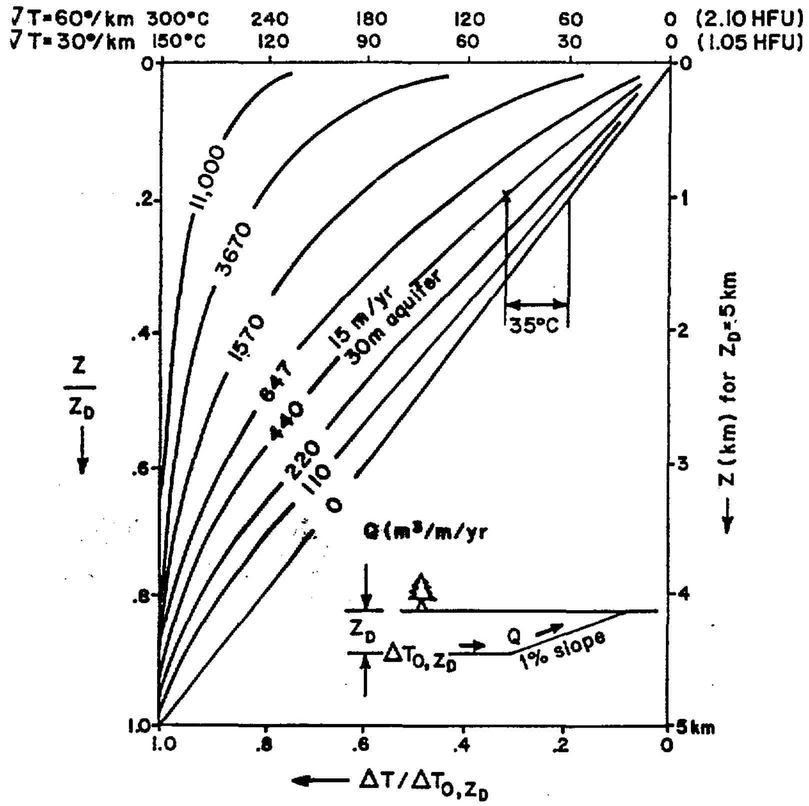
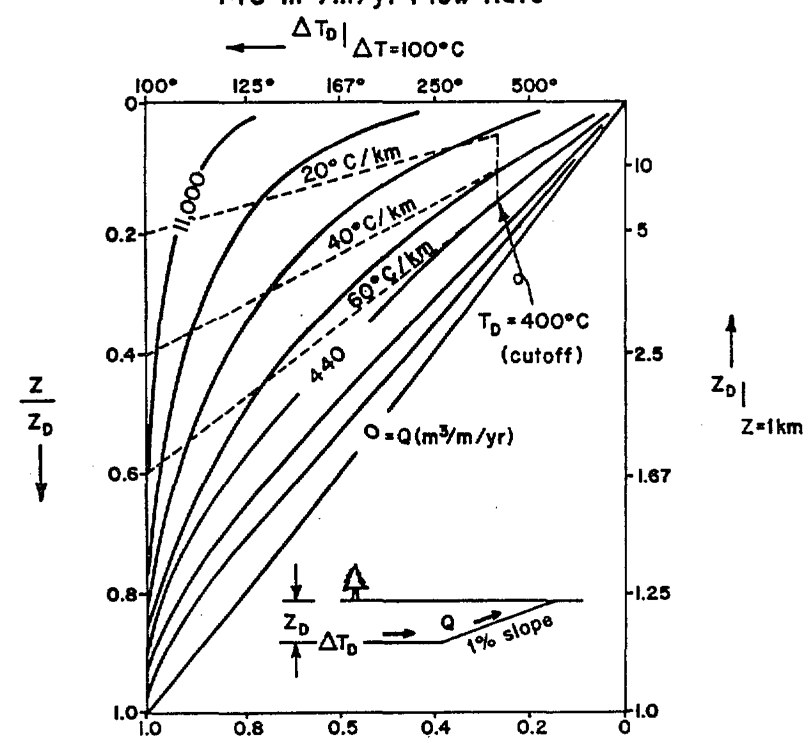


Figure 4

ΔT = 100°C at 1 km Depth Requires ~440 m³/m/yr Flow Rate



$$\frac{\Delta T}{\Delta T_{0,Z_D}} = \frac{\left(\frac{Z}{Z_D}\right)^{\alpha} + \alpha\left(\frac{Z}{Z_D}\right)}{\alpha + 1}, \quad \alpha = \frac{K}{cQ\tan\beta}$$

Figure 5

EXPULSION MUST BE EPISODIC TO PERTURB TEMPERATURE

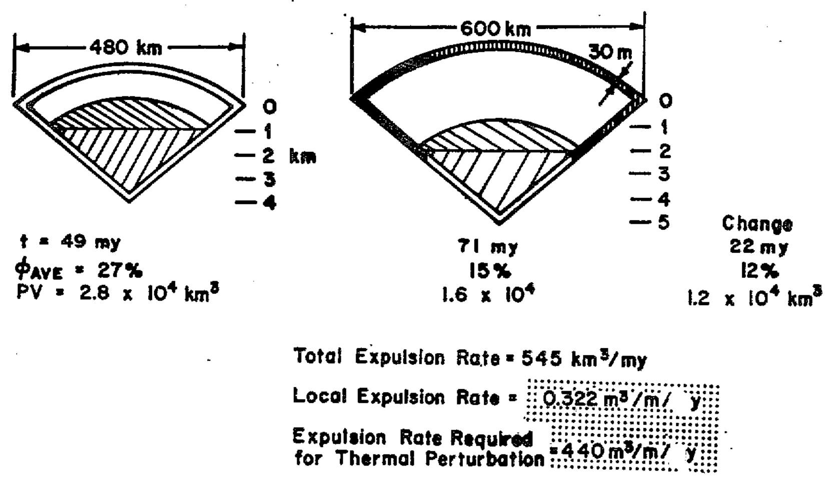


Figure 6

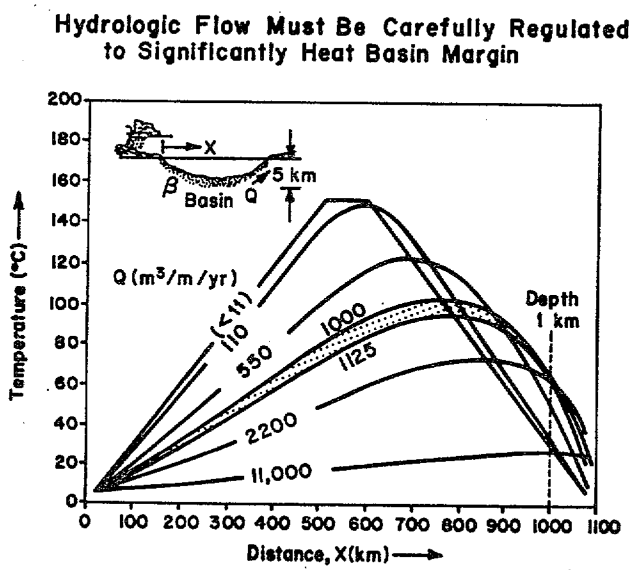


Figure 7

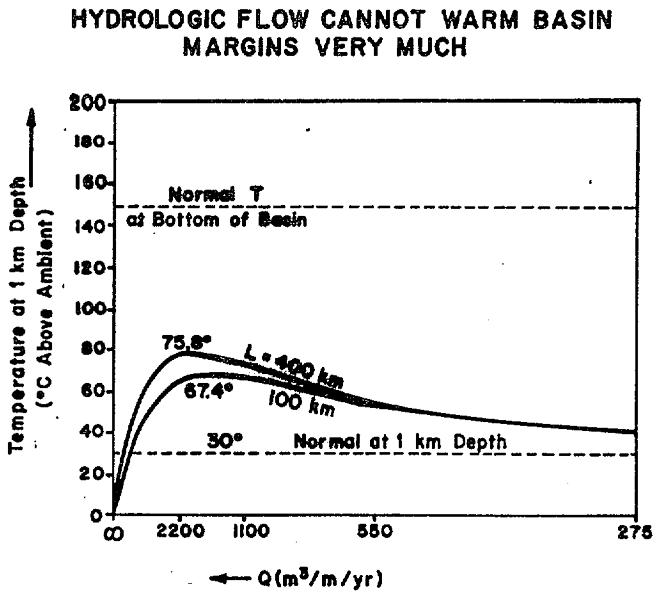
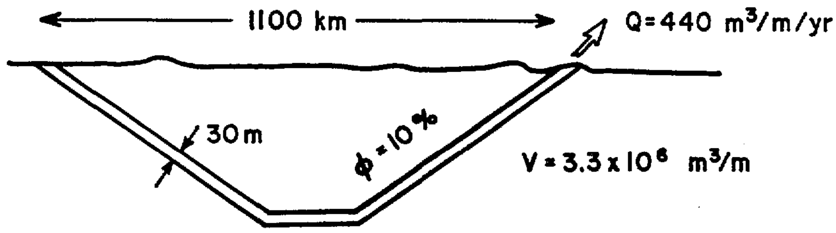


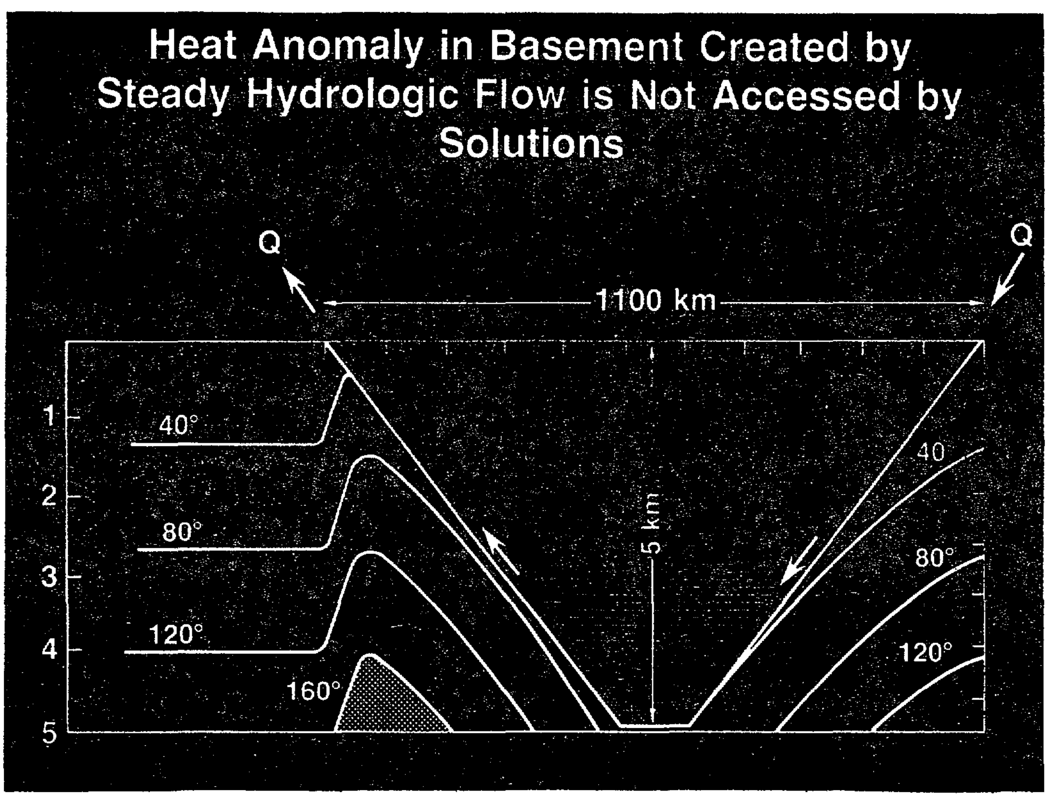
Figure 8

MAIN PROBLEM WITH CROSS BASIN FLOW: BRINE WOULD BE FLUSHED TOO QUICKLY FROM AQUIFIER



$$\text{FLUSH TIME} = \frac{V}{Q} = \frac{3.3 \times 10}{440} = 7500 \text{ yrs}$$

Figure 9



Simplified heat flow and subsidence histories for asymmetric extensional basins

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Horizontal extension of the lithosphere is now widely accepted as a mechanism of sedimentary basin formation. There are many possible geometrical distributions of lithospheric strain that result in thinning of the crust and separation of the basin margins, but there are two important end-members of the range of extension solutions, for both of which the thermal evolution of the lithosphere is simply analysed. These simplified models bound the behaviour of other more complex (and perhaps more realistic) models.

At one end of the range is the 'pure shear' class of models in which lithospheric extension takes place by means of a distributed ductile deformation, analogous to the stretching of a stick of toffee. A basin formed by this mechanism should be approximately symmetric in terms of the thermal anomaly introduced by this extension, although faulting of the near-surface layer may impart a superficial asymmetry to the basin. McKenzie (1978) analysed the thermal evolution of the simplest model of this class (a single layer, rapidly and uniformly stretched by a factor β). This analysis, and derivatives of it, have been widely used to model observed subsidence rates in intracratonic basins and continental margins (e.g., Sclater & Christie, 1980; Sclater & others, 1980; Steckler & Watts, 1978; Royden & Keen, 1980; Royden & others, 1980; Hellinger & Sclater, 1983; Barton & Wood, 1984).

At the other end of the range is the 'simple shear' class of models, the simplest of which concentrates all of the horizontal strain on a narrow, dipping detachment zone that cuts right through the lithosphere, dividing it into upper and lower plates (Fig. 1). The detachment surface is generally pictured as an aseismic normal fault near the surface, grading into a ductile shear zone at depth. This type of model was proposed by Wernicke (1981, 1985) with reference to the recent extension of the western USA. Its implications for the development of continental margins have been described by Lister & others (1986) and a number of extensional regions have been the subject of new interpretations in which at least some of the extension takes place on detachment surfaces (e.g., Beach, 1986; Fitzgerald & others, 1986; Mohr, 1987).

The thermal anomaly introduced by extension on a detachment zone is inherently asymmetric with respect to the axis of the basins; a simple formulation of the thermal evolution was recently given by Voorhoeve & Houseman (1987) and is summarised below.

In both simple-shear and pure-shear models there are 2 phases of subsidence. Initial subsidence takes place on the same time-scale as the extension and is caused by thinning of the crust and heating of the lithosphere. There follows a slower phase known as thermal subsidence caused by the subsequent cooling and thermal contraction of the lithosphere. With appropriate assumptions, the temperature as a function of time t and depth z within the lithosphere of thickness L is:

$$T = T_1 \left[\frac{z}{L} + \sum_{n=1}^{\infty} b_n \exp[-n^2\pi^2\kappa t/L^2] \sin[n\pi z/L] \right]$$

consisting of a linear steady-state geotherm plus a series of decaying transient terms (T_1 is the temperature at z = L and κ is the thermal diffusivity). The Fourier coefficients b_n are calculated from the initial post-extension geotherm. For rapid extension, with the geometry shown in Figure 1,

$$b_n = \frac{2}{n\pi} \left\{ (1-c) \cos(n\pi a) + \frac{1}{n\pi} \sin(n\pi c) \right\}$$

where a and c are geometrical factors that define the depth to the detachment surface and to the asthenosphere.

The surface heat flow follows from the temperature gradient at the surface, and the basement level (elevation or subsidence) is obtained by isostatic balance of the extended column with the original unstretched column. With suitable approximations, the following explicit expressions for the surface heat flow and subsidence as a function of time are simplified from those obtained by Voorhoeve & Houseman, 1987:

$$F = \frac{kT_1}{L} \left\{ 1 + \pi \sum_{n=1}^{\infty} n b_n \exp[-n^2 \pi^2 \kappa t / L^2] \right\}$$

$$S_T = h \left(1 - \frac{1}{\beta}\right) (\rho_m - \rho_c)/(\rho_a - \rho_s)$$

$$S_t = S_T - \alpha T_1 L \rho_m \left\{ \sum_{m=1}^{\infty} \frac{2}{(2m+1)\pi} b_m \exp[-(2m+1)^2 \pi^2 \kappa t / L^2] \right\}$$

where k is thermal conductivity, F is surface heat flow, ρ_m and ρ_c are zero-temperature densities of mantle and crust respectively, α is the coefficient of thermal expansion, h is the original crustal thickness, h/β is the crustal thickness after extension, ρ_a is asthenosphere density, ρ_s is sediment (or water) density, S_t is the subsidence at time t and S_t is the total subsidence ($t = \infty$). These expressions are equally valid for the pure shear model if the coefficients b_n are appropriately modified.

Figure 2 shows surface elevation and heat flow profiles across a basin produced by 100 km of extension on a detachment surface that dips at 12°. The crustal thickness distribution (and thus the final subsidence profile) is symmetric about the basin axis if the dip of the detachment surface is constant, but the underlying thermal anomaly is asymmetric because of the variable depth to the detachment surface. The asymmetry of the subsidence profiles is most marked at time zero, immediately after the extension, and gradually diminishes with time, whereas for surface heat flow, the apparent asymmetry increases with time. On the upper-plate side of the basin the thermal pulse has first to heat-up the overlying plate before it can be conducted away, so the heat flow first increases and then diminishes with time (Fig. 3). The magnitude of the peak heat flow decreases, and the time of the peak increases, with increasing depth to the detachment surface. The time-integrated heat flow at any point also diminishes (roughly linearly) as the detachment surface gets deeper. An important difference from the pure shear model is that, for a given crustal extension factor β, the time-integrated heat flow in the centre of the basin is only about half the amount predicted by the pure shear model.

The basin-wide asymmetry of heat-flow history will presumably control basin-wide variations in the degree and the timing of thermal maturation of overlying sediments. Studies of maturation indicators may, in fact, be the best way to demonstrate lithospheric-scale asymmetry of an extensional basin.

References

Barton, P., & Wood, R., 1984 - Tectonic evolution of the North Sea basin: crustal stretching and subsidence. Geophys. J.R. Astr. Soc., 79, 987–1022.

Beach, A., 1986 - A deep seismic reflection profile across the northern North Sea. Nature, 323, 53–55.

Fitzgerald, P.G., Sandiford, M., Barrett, P.J., & Gleadow, A.J.W., 1986 - Asymmetric extension in the Transantarctic Mountains and Ross Embayment, Antarctica. Earth planet. Sci. Lett., 81, 67–78.

Hellinger, S.J., & Sclater, J.G., 1983 - Some comments on two–layer extensional models for the evolution of sedimentary basins. J. Geophys. Res., 88, 8231–8270.

Lister, G.S., Etheridge, M.A., & Symonds, P.A., 1986 - Detachment faulting and the evolution of passive continental margins. Geology, 14, 246–250.

McKenzie, D.P., 1978 - Some remarks on the development of sedimentary basins. Earth planet. Sci. Lett., 40, 25–32.

Mohr, P., 1987 - Structural style of continental rifting in Ethiopia. EOS Trans. Amer. Geophys. Union, 68, 721–730.

Royden, L., & Keen, C.E., 1980 - Rifting process and thermal evolution of the continental margin of eastern Canada determined from subsidence curves. Earth planet. Sci. Lett., 51, 343–361.

Royden, L., Sclater, J.G., & Von Herzen, R.P., 1980 - Continental margin subsidence and heat flow. Important parameters in formation of petroleum hydrocarbons. Am. Assoc. Pet. Geol. Bull., 64, 173–187.

Sclater, J.G., & Christie, P.A.F., 1980 - Continental stretching: an explanation of the post mid–Cretaceous subsidence of the central North Sea basin. J. Geophys. Res., 85, 3711–3739.

Sclater, J.G., Royden, L., Horvath, F., Burchfiel, B.C., Semken, S. & Stegena, L., 1980 - Formation of the intra–Carpathian basins as determined from subsidence data. Earth planet. Sci. Lett., 51, 139–162.

Steckler, M.S., & Watts, A.B., 1978 - Subsidence of the Atlantic type continental margin off New York. Earth planet. Sci. Lett., 41, 1–13.

Voorhoeve, H., & Houseman, G., 1987 - The thermal evolution of lithosphere extending on a low–angle detachment zone. Basin Research (in press).

Wernicke, B., 1981 - Low angle normal faults in the Basin and Range province: nappe tectonics in an extending orogen. Nature, 291, 645–647.

Wernicke, B., 1985 - Uniform sense normal simple shear of the continental lithosphere. Can. J. Earth Sci., 22, 108–125.

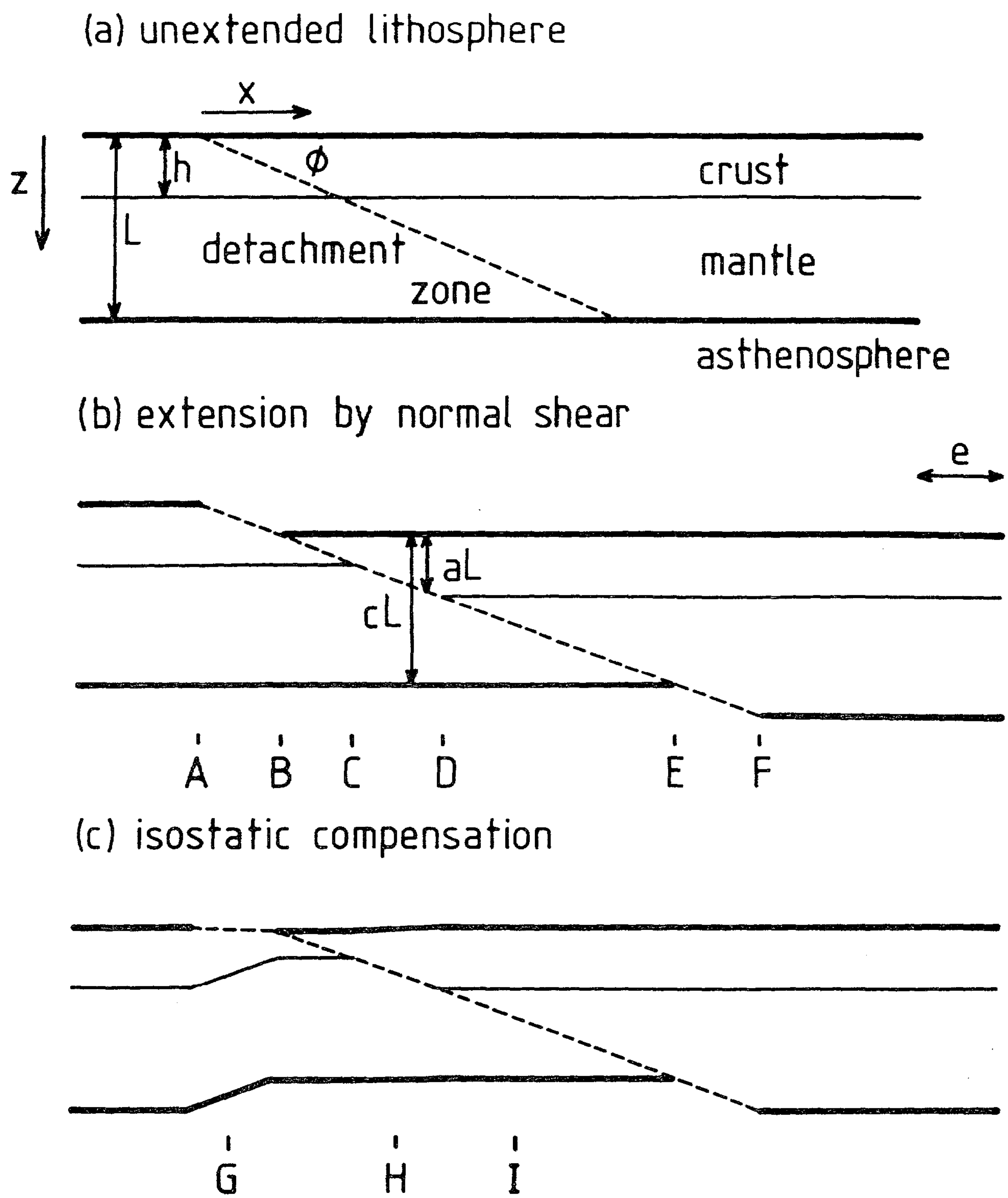

Figure 1 The geometry of extension on a detachment surface: (a) shows unextended lithosphere in vertical section; the dashed line is the detachment zone with dip φ. (b) extension is accomplished by normal shear on the detachment surface; the horizontal separation of the two plates is e, the depth to the detachment surface is aL and the depth to the base of the thinned lithosphere is cL. (c) local isostatic compensation requires vertical adjustment of every column to the right of the origin; the distances e, aL and cL remain as shown in (b).

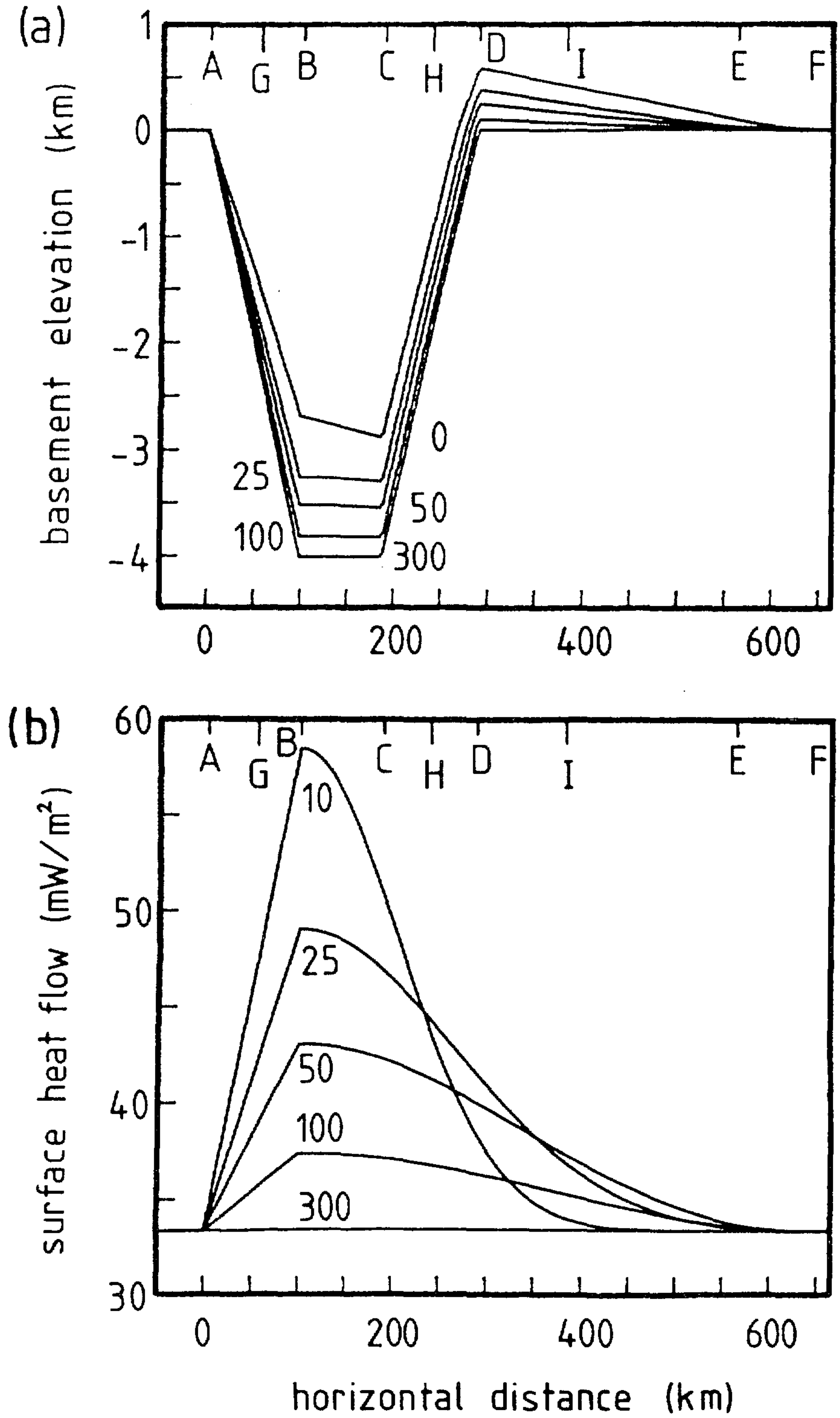

Figure 2 (a) Surface elevation versus x-coordinate at several different times, e=100 km, φ=12°, L=120 km, h=40 km, α=3.4×10^{-5} K^{-1}, T_1=1333 °C, ρ_m=3.33 Mg/m^3, ρ_c=2.919 Mg/m^3 and κ=8×10^{-7} m^2/s. Subsidence below zero level is assumed to be water loaded (ρ_s=1.03 Mg/m^3) while elevation above zero is in air. The number beside each curve is the time in m.y. after extension and the letters A-F correspond to the locations in Figure 1. (b) Surface heat flow versus x-coordinate at several different times for the same parameters and k=3.0W/m/K. The time in m.y. is shown beside each curve. At t=0 the heat flow between A and B is undefined (∞) and elsewhere it is equal to the pre-extension heat flow.

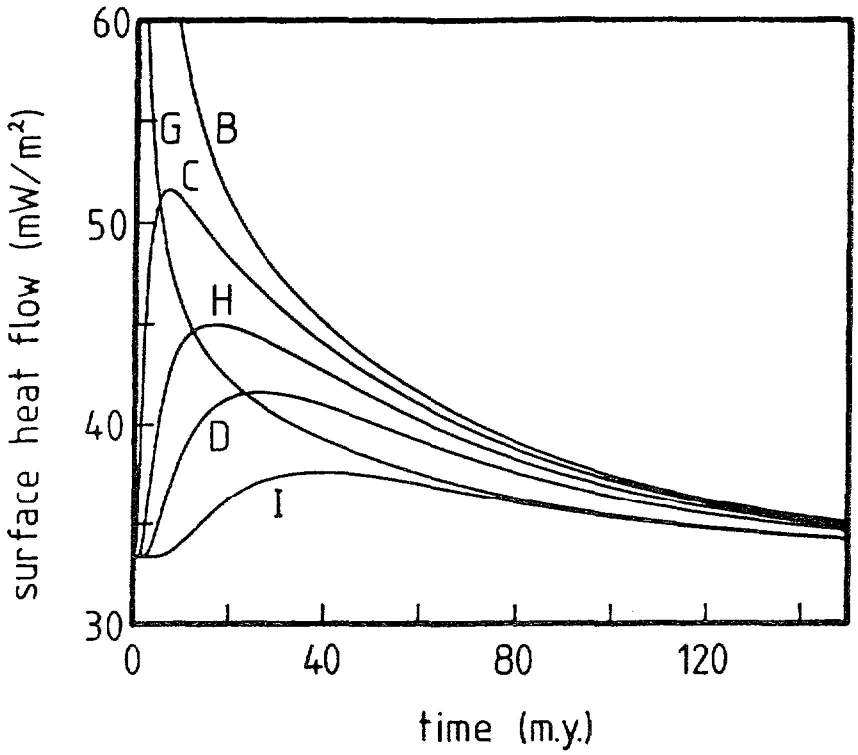

Figure 3 Surface heat flow versus time at 6 different x-coordinates. See Figure 1 for the x-location. Curves G and B decrease monotonically from infinity at t=0 while the other curves increase from the pre-extension value to a maximum. All curves decay back to the pre-extension heat flow.

Application of fission track dating to extensional terranes and their basins

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Fission track dating depends on the accumulation of radiation damage from the spontaneous fission of ^{238}U over geological time in uranium-bearing minerals. The radiation damage occurs in the form of fission tracks, linear defects in the host crystal which are produced by the passage of the heavily-ionizing nuclear fragments resulting from fission decay. The tracks can be made visible by chemically etching a polished surface on the crystal so that they can be observed and counted under an optical microscope at high magnification. The number of these fission tracks will steadily increase through time, at least for the simplest case where the sample remains at low temperatures. If the rate of decay and the uranium concentration are known, then a geological age can be calculated from the number of tracks that have accumulated over the lifetime of the crystal. The most frequently used mineral for fission track dating is apatite, because of its common occurrence, suitable uranium abundance and low-temperature track fading properties which enable its application to a variety of palæotemperature reconstruction and thermotectonic problems. Useful summaries of various aspects of the fission track dating method may be found in Naeser (1979), and Gleadow & others (1983).

Like other forms of radiation damage, fission tracks begin to fade as temperature is increased and are only stable for long time periods at relatively low temperatures. The fading of the etchable fission tracks during heating is known as thermal annealing and occurs by a gradual shortening of the etchable tracks as the radiation damage is progressively repaired and eventually disappears. The temperatures required for this to occur depend on the duration of heating and the particular mineral, each having its own characteristic annealing properties. The same process, which occurs during natural annealing in geological environments over long time periods, can be reproduced in the laboratory at high temperatures for much shorter times. Much of the emphasis in fission track studies over the past five years has concerned firstly defining the full extent of the information contained within the fission track record (e.g., Gleadow & others, 1983), Green & others, 1987), and secondly, developing techniques to extract this information as a quantitative thermal history (e.g., Laslett & others, 1986).

Information about past variations in temperature is preserved in the distribution of fission track lengths. During thermal annealing each track will shorten to a length which is essentially characteristic of the highest temperatures experienced. As new fission tracks are continuously being added to the crystal throughout its lifetime by radioactive decay, the final distribution of track lengths will be a function of the thermal history experienced. The annealing process is a function of both temperature and time and has been studied in detail in the laboratory by examining the track shortening produced by heating at various combinations of temperature and time (Green & others, 1986; Laslett & others, 1986). On the basis of these annealing studies the pattern of observed fission track parameters which would be expected from any given thermal history can now be modelled.

Fission track analysis of apatites from uplifted and eroded basement rocks and sedimentary basin sequences can thus provide a unique record of the low-temperature (<150^{o}C) thermal history experienced in such environments. This information can be used to constrain and test various models of the tectonic and structural evolution of extensional terranes. The low-temperature thermal evolution of sedimentary basins formed in such environments also has important implications for the timing of hydrocarbon maturation in suitable source rocks contained within them. A number of examples of the application of these techniques are described in the following reference list.

References

Fitzgerald, P.F., & Gleadow, A.J.W., 1987 - Fission track geochronology, tectonics and structure of the Transantarctic Mountains in northern Victoria Land, Antarctica. Isotope Geosci., (in press).

Fitzgerald, P.G., Sandiford, M., Barrett, P.J., & Gleadow, A.J.W., 1986 - Asymmetric extension in the Transantarctic Mountains and Ross Embayment. Ear. Planet. Sci. Lett., 86, 67-78.

Gleadow, A.J.W., & Brooks, C.K., 1979 - Fission track dating, thermal history and tectonics of igneous intrusions in East Greenland. Contr. Min. Pet., 71 45-60.

Gleadow, A.J.W., & Fitzgerald, P.F., 1986 - Uplift history and structure of the Transantarctic Mountains: new evidence from fission track dating of basement apatites in the Dry Valleys area, southern Victoria Land. Ear. Planet. Sci. Lett., 82, 1-14.

Gleadow, A.J.W., Duddy, I.R., & Lovering, J.F., 1983 - Fission track analysis: a new tool for the evaluation of thermal histories and hydrocarbon potential. APEA J., 23, 93-102.

Gleadow, A.J.W., Duddy, I.R., Green, P.F., & Lovering, J.F., 1986 - Confined fission track lengths in apatite - a diagnostic tool for thermal history analysis. Contrib. Mineral. Petrol., 94, 405-415.

Gleadow, A.J.W., Duddy, I.R., Green, P.F., & Hegarty, K.A., 1986 - Fission track lengths in the apatite annealing zone and the interpretation of mixed ages. Ear. Planet. Sci. Lett., 78, 245-254.

Green, P.F., 1986 - On the thermo-tectonic evolution of Northern England: evidence from fission track analysis. Geol. Mag., 123, 493-506.

Green, P.F., Duddy, I.R., Gleadow, A.J.W., Tingate, P.R., & Laslett, G.M., 1986 - Thermal annealing of fission tracks in apatite: I - A qualitative description. Isotope Geoscience, 59, 237-253.

Green, P.F., Duddy, I.R., Gleadow, A.J.W., & Lovering, J.F., 1987 - Apatite fission track analysis as a palæotemperature indicator for hydrocarbon exploration. In Naeser, N.D. (Editor) - Springer-Verlag, (in press).

Hegarty, K.A., Duddy, I.R., Green, P.F., Gleadow, A.J.W., Fraser, I., & Weissel, J.K., 1987 - Regional evaluation of the tectonic and thermal history of the Gippsland Basin. In Glenie, R.C., (Editor) - Second southeastern Australia oil exploration Symposium, Petroleum Exploration Society of Australia, 65-74.

Laslett, G.M., Green, P.F., Duddy, I.R., & Gleadow, A.J.W., 1986 - Thermal annealing of fission tracks in apatite: II - A quantitative analysis. Isotope Geoscience, 65, 1-13.

Moore, M.E., Gleadow, A.J.W., & Lovering, J.F., 1986 - Thermal evolution of rifted continental margins: new evidence from fission track dating of apatites from southeastern Australia. Ear. Planet. Sci. Lett., 78, 255-270.

Naeser, C.W., 1979 - Fission-track dating and geologic annealing of fission tracks. In Jäger, E., & Hunziker, J.C., (Editors) - Lectures in isotope geology. Springer-Verlag, 1979, 154-169.

Analogue modelling experiments of structuring during normal and oblique extension

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Regional-scale extensional structures are common features of passive margins throughout the world and dominate the structure of many major petroleum fields. A study of extensional deformation styles, kinematic history and associated stress regimes has been carried out using sandbox analogue models. Dry quartz sand is used as it has a similar rheological behaviour to the brittle crustal rocks of passive margins (Sanford, 1959), particularly the bulk modulus to cohesion ratio and Coulomb fracture response. No attempt is made to quantify results; this would be of little value since forces and strain-rates in brittle deformation have minimal effect on the resultant geometry of deformation features.

Basic structural styles interpreted from seismic profiles have been modelled with the observed master-fault simulated by imposing a detachment of similar slope and curvature (e.g., Figs. 2a & 2b). The major stress regimes associated with movement on the fault are inferred from associated structures and regional interpretation. To assist in understanding stress regimes associated with extensional systems, a graphical representation of stress trajectories and Mohr-Coulomb fracture trajectories was constructed (see Fig. 1), based on similar work by Hafner (1951). These diagrams illustrate the relationship between the fault geometry and footwall or hanging-wall migration. Also predicted is the convex upward geometry of faults conjugate to listric faults (concave upward) observed in some models (e.g., Fig. 2).

The following structural styles have been identified and modelled:

(i) Hangingwall deformation over listric normal faults. The style of deformation in a listric fault hangingwall is directly related to the dip and curvature of the fault-plane. Fault planes of even curvature and gradual flattening have roll-over 'blocks', where the material at the front of the hangingwall dips toward the fault plane at an increasing angle with displacement. There is little internal deformation within the block. A syn-deformational sedimentary package may accumulate above the block as it moves down the listric master-fault. Rotation of the roll-over block causes extension in the hangingwall material immediately behind it. At high angles of roll-over this creates an asymmetrical graben, termed the 'roll-over graben' (Fig. 2b). Bounding this graben from the roll-over block, and accommodating the rotational displacement associated with roll-over, is the 'main synthetic fault system' (Fig. 2b). These convex-upward faults are of type B, as predicted in Figure 1b, and run along the top side of the roll-over block. This may form a potentially large hydrocarbon trap. On the other side of the graben the bounding fault system consists of spaced, lower-angle, concave-upward faults that are antithetic to the master-fault. These correspond to type A faults as illustrated in Figure 1b. The lower sections of both bounding fault systems simply fade out where the roll-over block pivots and below which there is no extension. A subtle antiform, termed the 'basal antiform', may be observed in this lower area, suggesting a weak compressive régime. This may be a boundary effect, although if proven otherwise would certainly be of economic significance. The analogue model (Fig. 2b) exhibits all of the structures observed in the seismic profile (Fig. 2a), although the antithetic bounding faults are modified by adjacent structures and the main synthetic faults are interpreted as listric. It should be noted that real fault angles are significantly flatter and less concave upward due to the profile being a compressed, unmigrated section.

Low angle listric faults have less roll-over and may form 'nose structures' (Fig. 3). In nose structures the roll-over graben forms by pure shear without bounding fault systems (faults shown in the seismic section are due to later structuring). This slight roll-over without associated faulting may produce good anticlinal petroleum traps, depending upon the profile along strike.

If the fault curvature is sharp or has a change of slope, the area that would otherwise be an undeformed roll-over block becomes regularly faulted by antithetic faults (Fig. 4) of type A in Figure 1B. In the seismic profile (Fig. 4a) this style of fault is developed on the right-hand side over a low-angle décollement surface. Low-angle faults or décollements commonly have ramp-flat geometries, as modelled in Figure 5 using only a small ramp. Observation of this model development revealed the ramp basin to be pervasively faulted by synthetic faults (type B, Fig. 1b), which results in poor seismic data in such areas and may confuse interpretation of the basin. The fault set to the right end of Figure 5 is formed during the initial movement of the ramp.

(ii) Extension within a wrench régime. Pull-apart basins are produced by localised subsidence over areas of extension within basement wrench régimes. Such extension may be produced by dilational jogs or Riedel bends (Hempton & Neher, 1986). Fault geometry in the basin sediments is controlled at the surface by the bulk shear-stress and at depth these faults bend into the basement fault geometry (Fig. 6). Strike-slip faults have increased normal displacement as they approach the basin. The dual controls of faulting produce complicated geometries such as helicoidal and pseudo-reverse faults. Surface morphology of a pull-apart basin is characteristically sigmoidal, with increasing depth toward the centre. Interpretation of wrench-related extension structures is complicated by the strike-slip component on what appear to be normal faults but are actually rotational blocks, where fault blocks slide along strike into the basin.

(iii) Reactivation (inversion structures). With the high incidence of post-extensional compression events in extension dominated margins, it was useful to model structure from such events. Reactivation of a listric fault by a basement compression event may produce structures very similar to the 'positive flower' structure found in wrench régimes. The 'flower' is formed by successive faulting at a decreasing angle with the master fault, as illustrated by the fault sequence in Figure 7. Subsidence may occur between the faults even though they are within a compressive régime. In the seismic profile (Fig. 7a) the uppermost faults bear good resemblance to those formed in the model, illustrated in Figure 7b.

The analogue modelling program proved extremely successful in reproducing structural styles observed from seismic sections whilst providing valuable information on the mode and sequence of fault formation, and the associated stress régimes.

This project was made possible by the generous financial and technical support of Woodside Offshore Petroleum.

References

Hafner, W., 1951 - Stress distributions and faulting. Geol. Soc. Amer. Bull., 62, 373-398.

Hempton, M.R. & Neher, K., 1986 - Experimental fractures, strain and subsidence over en échelon strike-slip faults: implications for the structural evolution of pull-apart basins. J. Struct. Geol., 8, 597-605.

Sanford, A.R., 1959 - Analytical and experimental study of simple geological structures. Geol. Soc. Amer. Bull., 70, 19-52.

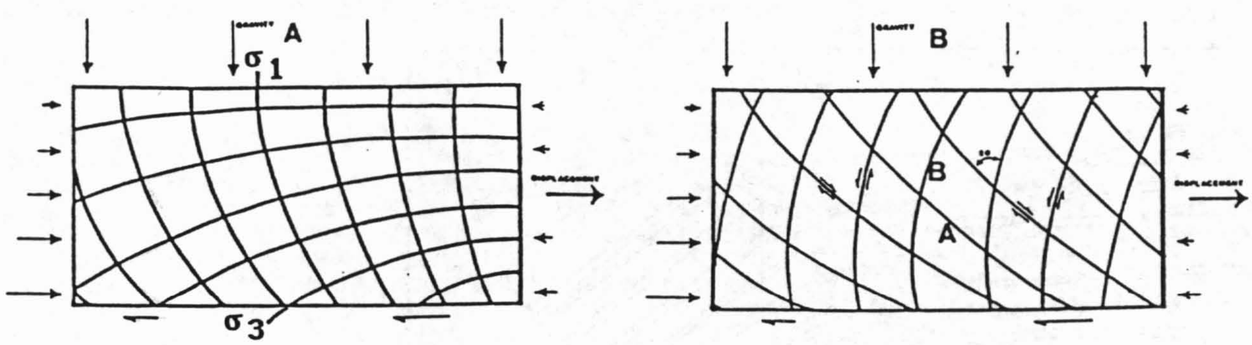


Figure 1. Geometry of principle stress trajectories (left) and corresponding Coulomb fractures (right), in hanging-wall block with relaxed confining stress on right side boundary.

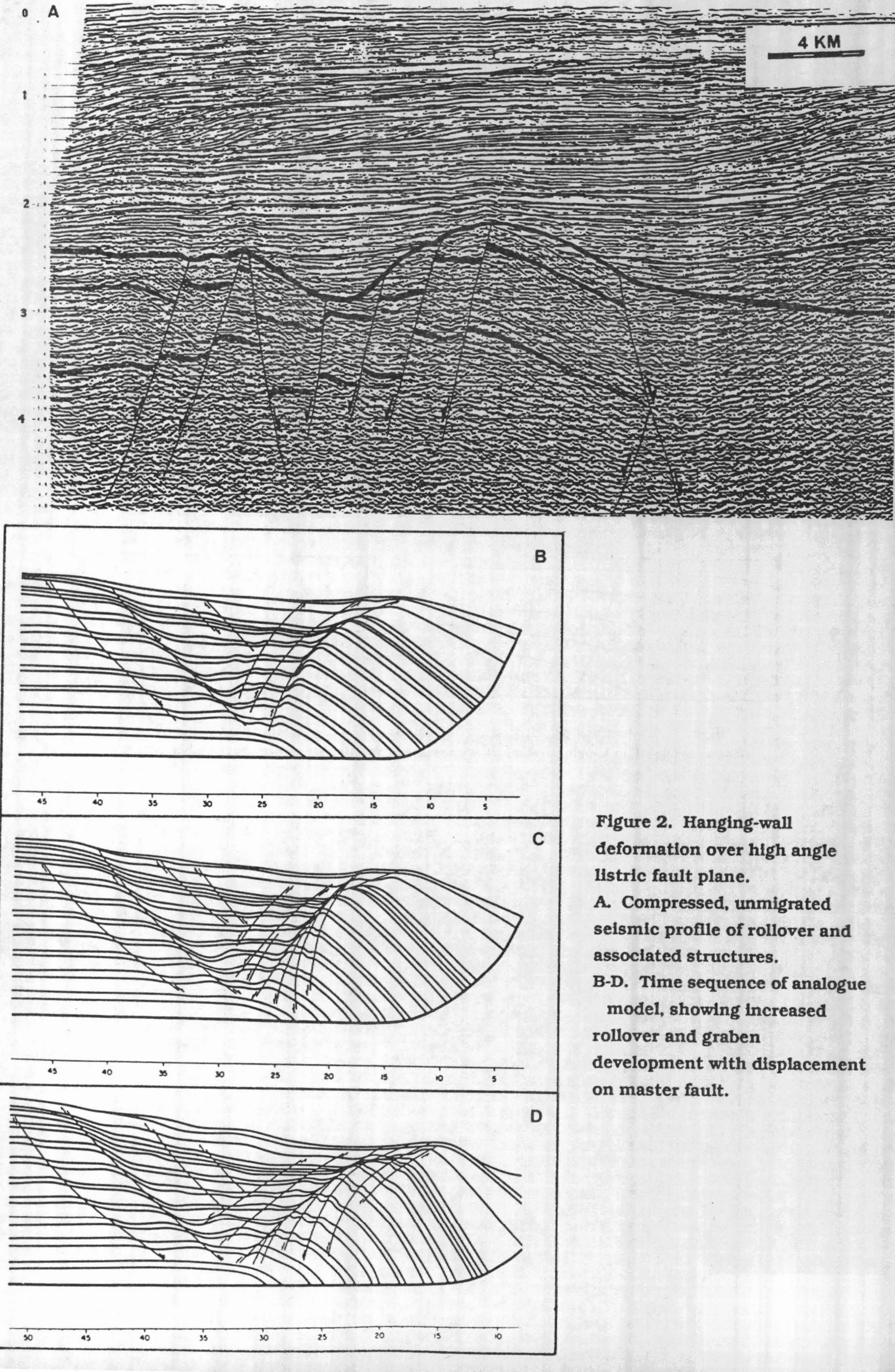
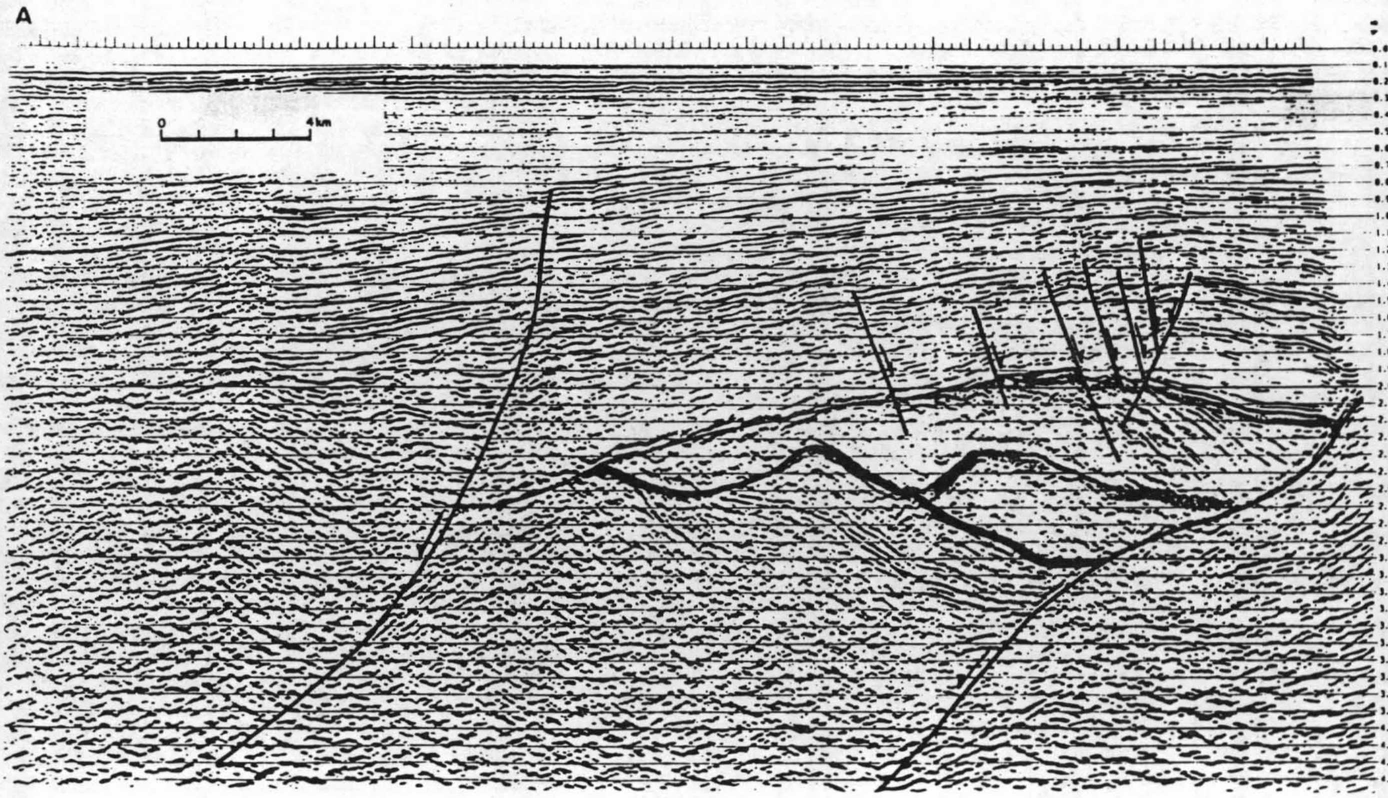


Figure 2. Hanging-wall deformation over high angle listric fault plane.

A. Compressed, unmigrated seismic profile of rollover and associated structures.

B-D. Time sequence of analogue model, showing increased rollover and graben development with displacement on master fault.



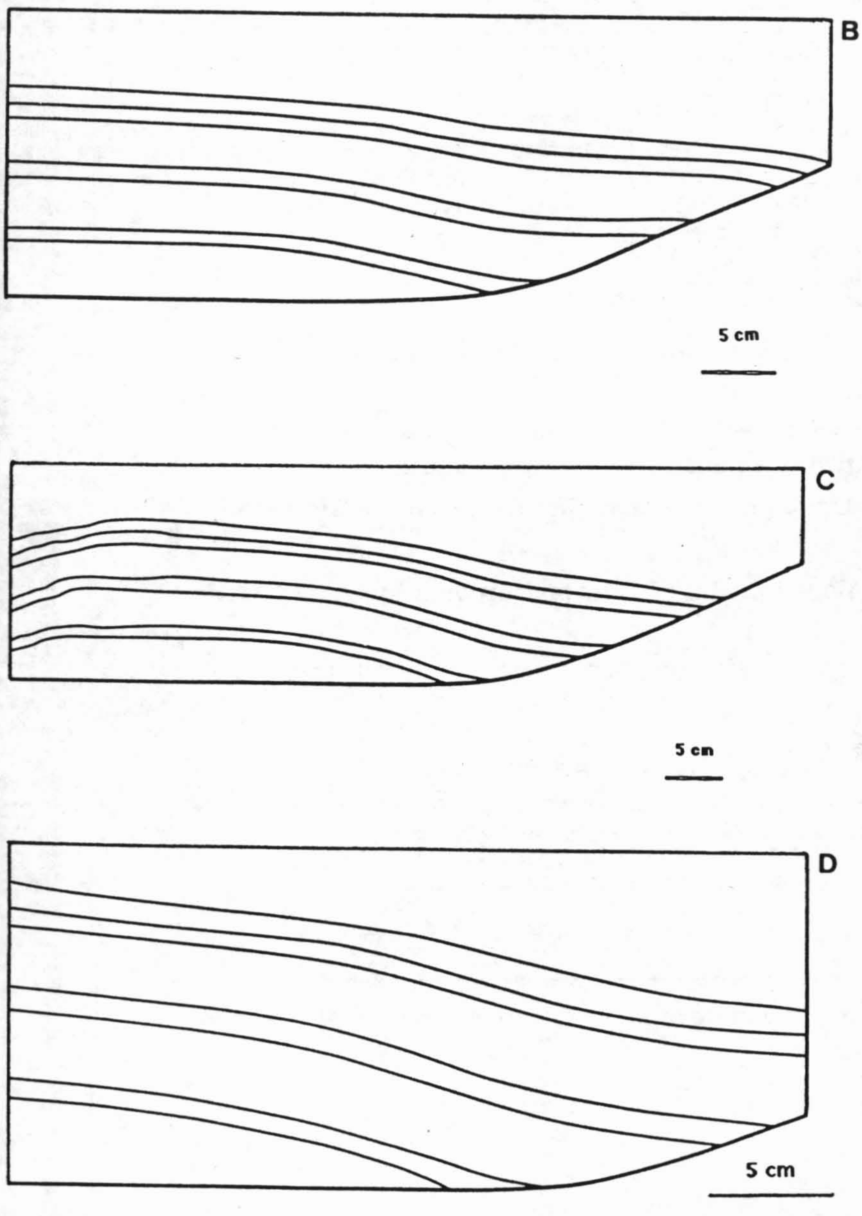


Figure 3. Hanging-wall deformation over low angle listric fault plane.

A. Compressed, unmigrated seismic profile of nose structure.

B-D. Time sequence of analogue model, showing development of nose structure.

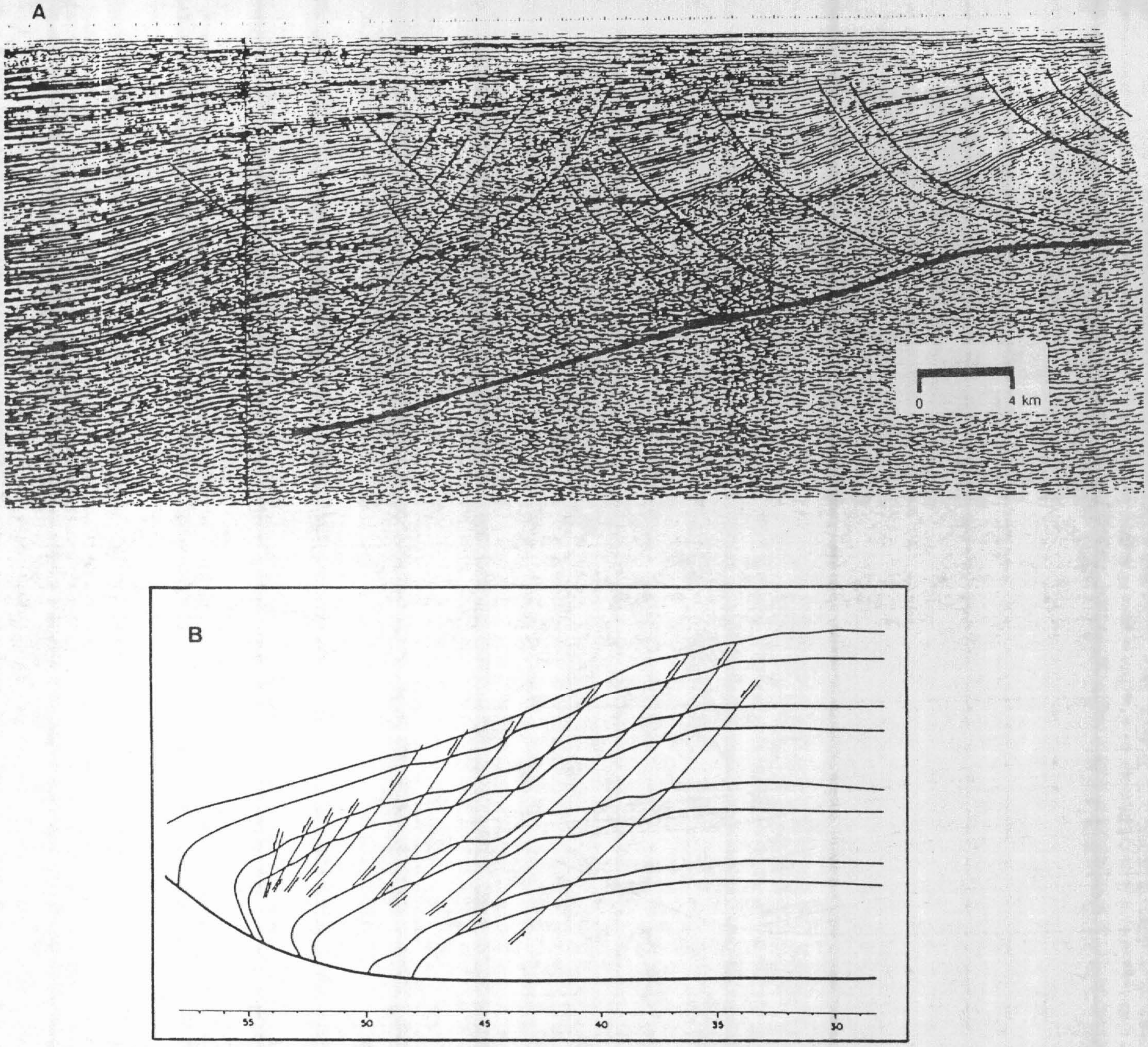


Figure 4. Antithetic faulting during extension on low angle faults.

A. Compressed, unmigrated seismic profile of antithetic faulting on large scale décollement.

B. Antithetic faulting in an analogue model of deformation over low-angle listric fault.

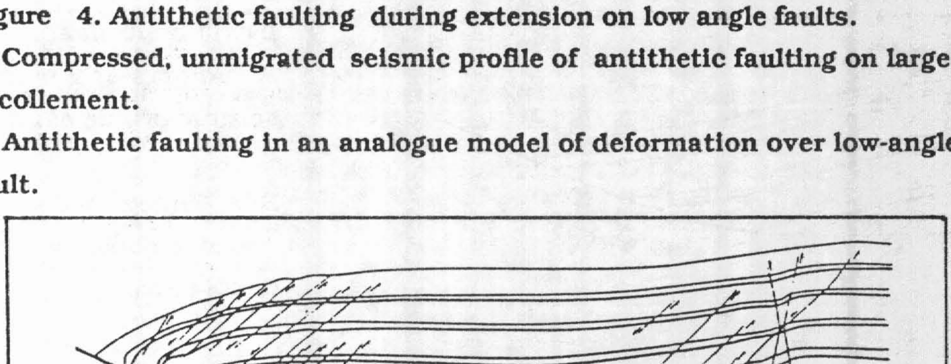


Figure 5. Analogue model of ramp basin formation during extension of ramp-flat fault.

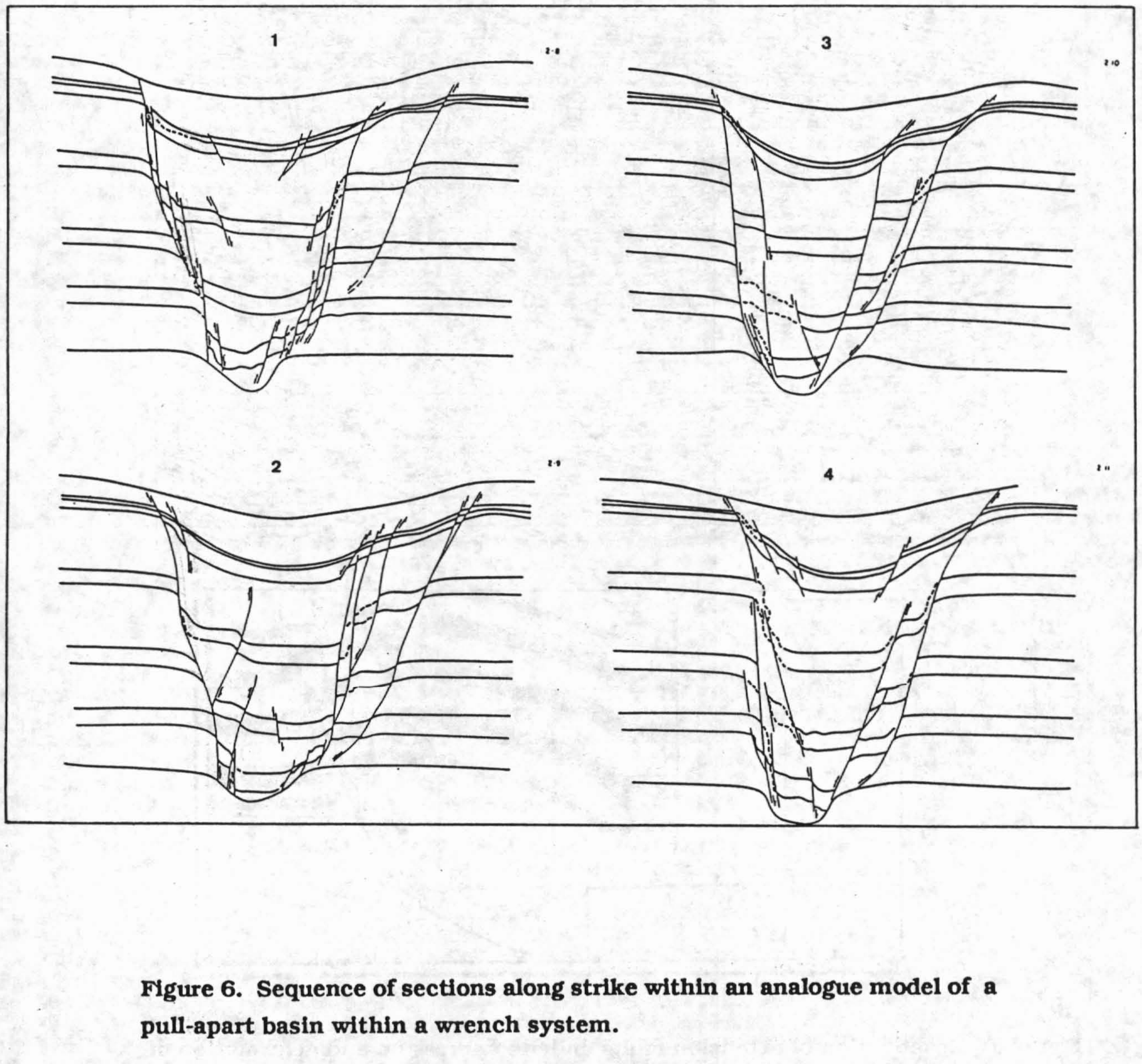


Figure 6. Sequence of sections along strike within an analogue model of a pull-apart basin within a wrench system.

A

B

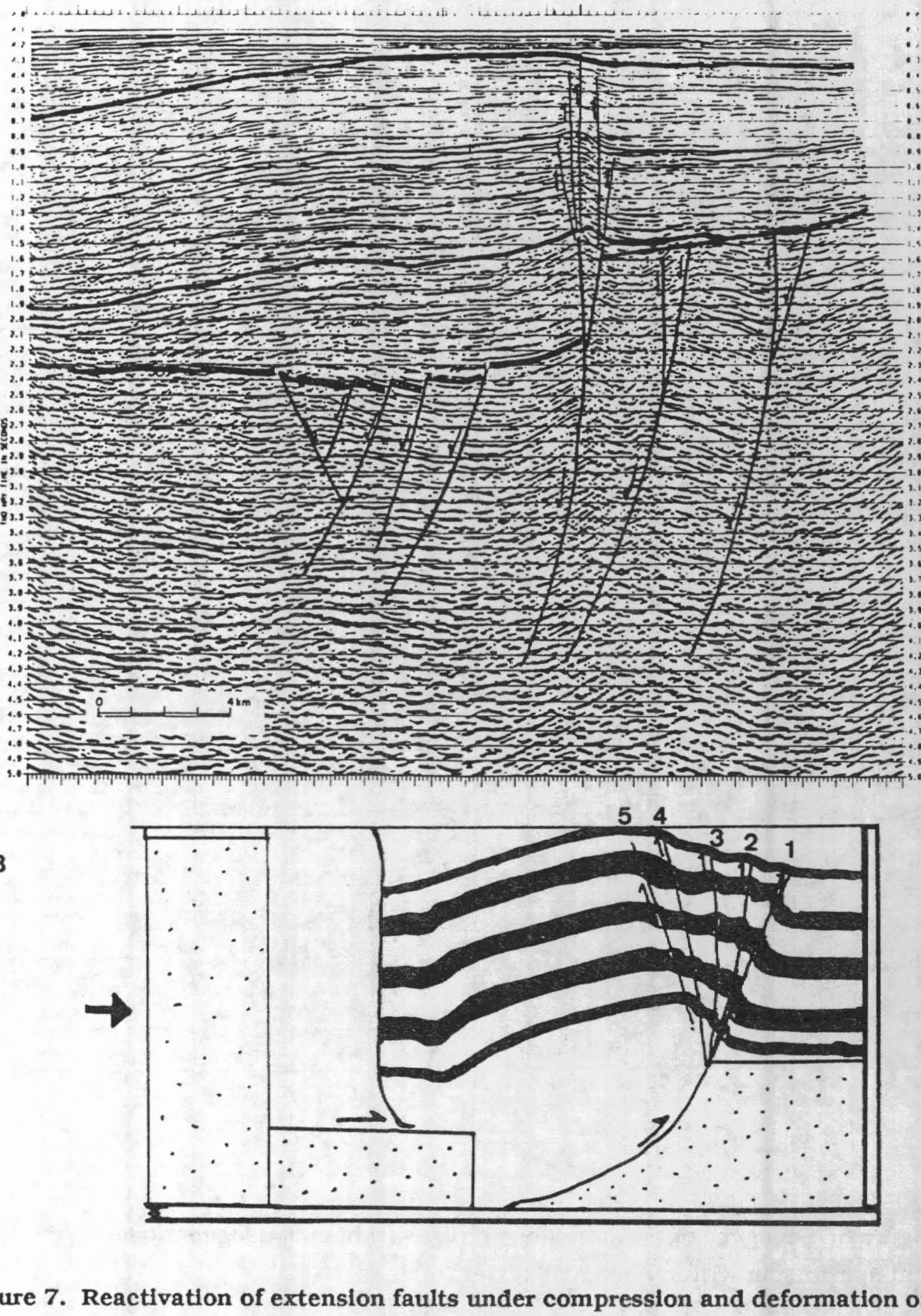


Figure 7. Reactivation of extension faults under compression and deformation of post-extensional sediment.

A. Compressed, unmigrated seismic profile of inversion structure, with splay set of reverse faults in post-rift sediments.

B. Analogue model illustrating the formation of reverse faults in undeformed material over a reactivated listric fault.

Basin evolution in the northern Browse Basin, offshore from northwest Australia

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The Browse Basin is essentially an intracratonic basin (Allen, Pearce & Gardner, 1978), although it is located on the continental shelf of northwest Australia, close to the edge of the continental crust (Fig. 1). While its tectonic setting suggests that it has been involved in the crustal extension leading to separation along the northwest margin, its structural history, deformation styles and sedimentation patterns indicate a markedly different timing of events from what might be predicted using a more simple rift–drift model (Falvey & Mutter, 1981; Veevers & Cotterill, 1978).

The northern Browse Basin is underlain by a thick, northwest–facing passive–margin sequence of Permian and Triassic sediments (MacDaniel, in press). Up to 700 m of Permian and 2 000 m of Triassic sediments have been recorded from exploration wells, but seismic data indicate that, over parts of the area, the Triassic alone is probably up to 4 500 m thick. This thick pile of relatively–unlithified sediments formed the tectonic basement on which a succession of Jurassic basins developed.

The passive–margin depositional pattern was terminated during the early Jurassic (Crostella, 1976; Barter, Maron & Willis, 1978) when a number of broad, intracratonic depocentres, or sag basins, developed along what is now the northwest continental margin. These depocentres, examples of which are the Cartier and Tamar Troughs, took the form of broad, shallow, oval–shaped depressions up to 40 km across, elongated in a northeast–southwest direction and flanked by broad, faulted arches also up to 40 km across (Fig. 2). The troughs probably formed a series of lakes or interconnected coastal embayments and were filled with sediments ranging in age from Hettangian–Sinemurian to Bathonian.

Following the cessation of deposition in the sag basins, a widespread unconformity surface developed across the entire area. Deposition on this surface commenced in the Oxfordian or slightly earlier during a rapid regional transgression. The Bathonian–Callovian erosional interval corresponds to the period immediately preceding continental separation along the present northwest margin, while the Oxfordian transgression corresponds to the beginning of seafloor spreading along the margin (Larson, 1975).

On the Ashmore Plateau, immediately west of the northern Browse Basin, the erosion surface extends in an almost planar manner across a terrace of tilted fault blocks. Triassic sediments within the fault blocks dip up to 15^{o} eastwards and are cut by west-dipping normal faults. Over part of the platform, the faults appear to join a major, sub-horizontal detachment surface near the base of the Triassic sequence. This deformation is probably related to the extension which eventually led to continental separation along the northwest margin. The Ashmore Platform was probably the easternmost area affected by the deformation.

In the northern Browse Basin the Oxfordian transgression coincides with a period of rapid faulting resulting in the development of a major fault system along the eastern margin of the basin and the formation of long, relatively narrow, northeast trending fault blocks which exerted a strong control on the Late Jurassic depositional pattern. The extent to which the Late Jurassic faults followed the trend of the faults on the arches flanking the sag basins is uncertain, but at least one sag basin, in the Skua-Eclipse area, has been strongly dislocated by later faulting.

The Late Jurassic depositional cycle was terminated during the Early Cretaceous Valanginian when another major, regional unconformity surface developed across the area. The timing of this Valanginian erosional interval is probably related to uplift along the flanks of the Browse Basin proper, immediately prior to the onset of thermal subsidence. The regional transgression which followed during the late Valanginian marks the beginning of thermal subsidence centred on the Browse Basin. Cretaceous and Paleocene sediments thicken consistently southwards across the northern portion of the Browse Basin and are at least 3 000 m thick in the Browse Basin depocentre to the south.

Eocene sediments, in contrast, prograde and thicken northwest across the northern Browse Basin, clearly in response to the flexure of the continental margin in that direction. This pattern continued until the Early Miocene when the area was again subjected to extensional deformation. Many of the Jurassic faults were reactivated and penetrate to the present seafloor. The common development of antithetic faults has led locally to complex, faulted mosaics of high-level grabens above deeper horst blocks.

The Cartier Trough also reactivated during the Early Miocene and is an excellent example of a present-day sag basin (Fig. 3). It is relatively symmetric in cross section and forms a prominent seafloor depression with water depths of up to 420 m compared with 200 m on the adjacent arches. The 1 000 m basal Miocene to seafloor isopach outlines a northeast-southwest-trending, oval-shaped area 40 km across and 120 km long. In the centre of the trough late Tertiary to Recent sediments, mainly limestone, are approximately 1700-1800 m thick.

When it reactivated, the trough initially formed a symmetric sag basin, with little or no faulting on its marginal arches. Faulting on the arches began later, with faults downthrowing both towards and away from the trough axis and with small half-grabens locally controlling sediment thicknesses. This pattern of trough subsidence and marginal faulting continued during the remainder of the Miocene and Pliocene and is still active today.

The recent structural studies described above have outlined a complex history on what was previously regarded as part of a typical passive margin. Unfortunately, no deep seismic lines have yet been acquired which might link the overall pattern with deeper crustal structures. However, the study has shown that different styles of depositional troughs formed at different times and that their nature may be partly influenced by their underlying basement and relative timing during the overall evolution of the area.

References

Allen, G.A., Pearce, L.G.G., & Gardner, W.E., 1978 - A regional interpretation of the Browse Basin. APEA Journal, 18(1), 23-33.

Barter, T.P., Maron, P., & Willis, I., 1982 - Results of exploration, Browse Basin, Northwest Shelf, Western Australia. Trans. 3rd Circum Pacific Energy and Mineral Resources Conference, 105-115.

Crostella, A., 1976 - Browse Basin. In Leslie, R.B., Evans, H.J., & Knight, C.L., (Editors), Economic Geology of Australia and Papua New Guinea. Australasian Institute of Mining & Metallurgy, Monograph 3, Petroleum, 194-9.

Falvey, D.A., & Mutter, J.C., 1981 - Regional plate tectonics and the evolution of Australia's passive continental margins. BMR Journal of Australian Geology & Geophysics, 6, 1-29.

Forrest, J.T., & Horstman, E.L., 1986 - The northwest shelf of Australia - geological review of a potential major petroleum province of the future. AAPG Memoir, 40, 457–486.

Larson, R.L., 1975 - Late Jurassic seafloor spreading in the eastern Indian Ocean. Geology, 3, 69–71.

Veevers, J.J., & Cotterill, D., 1978 - Western margin of Australia - Evolution of a rifted and arch system. Geological Society of America, Bulletin, 89, 337–355.

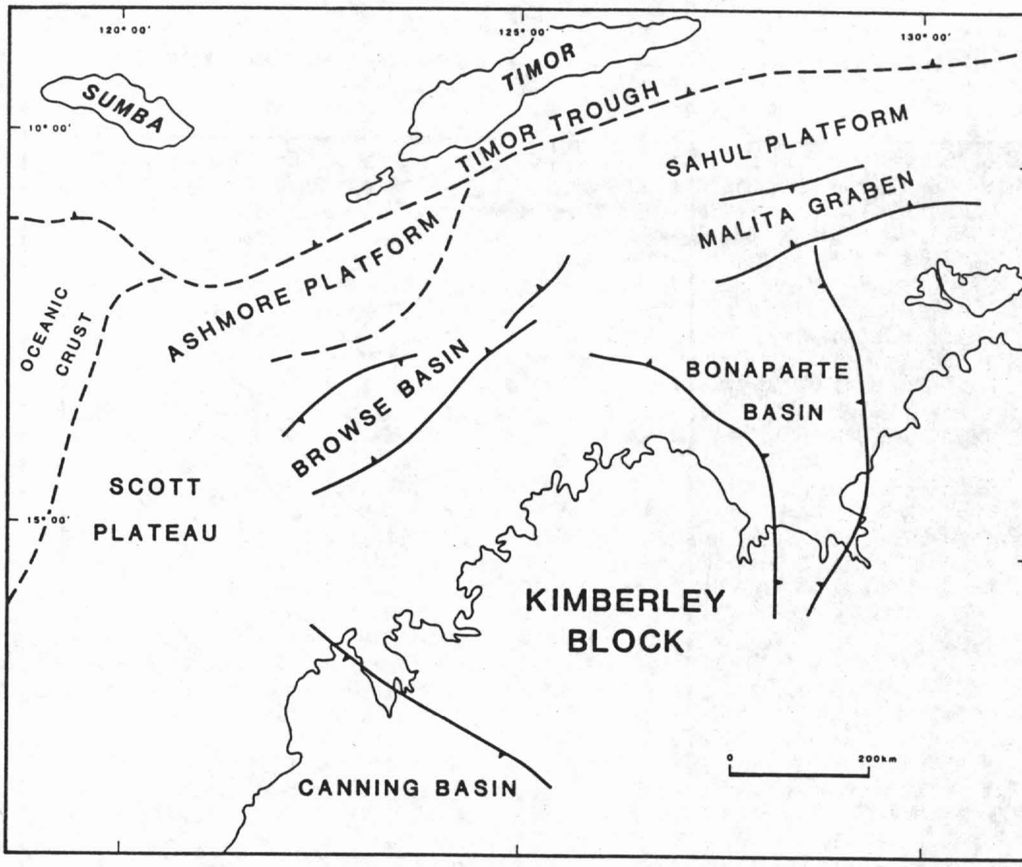

Figure 1. Location map, showing main tectonic features

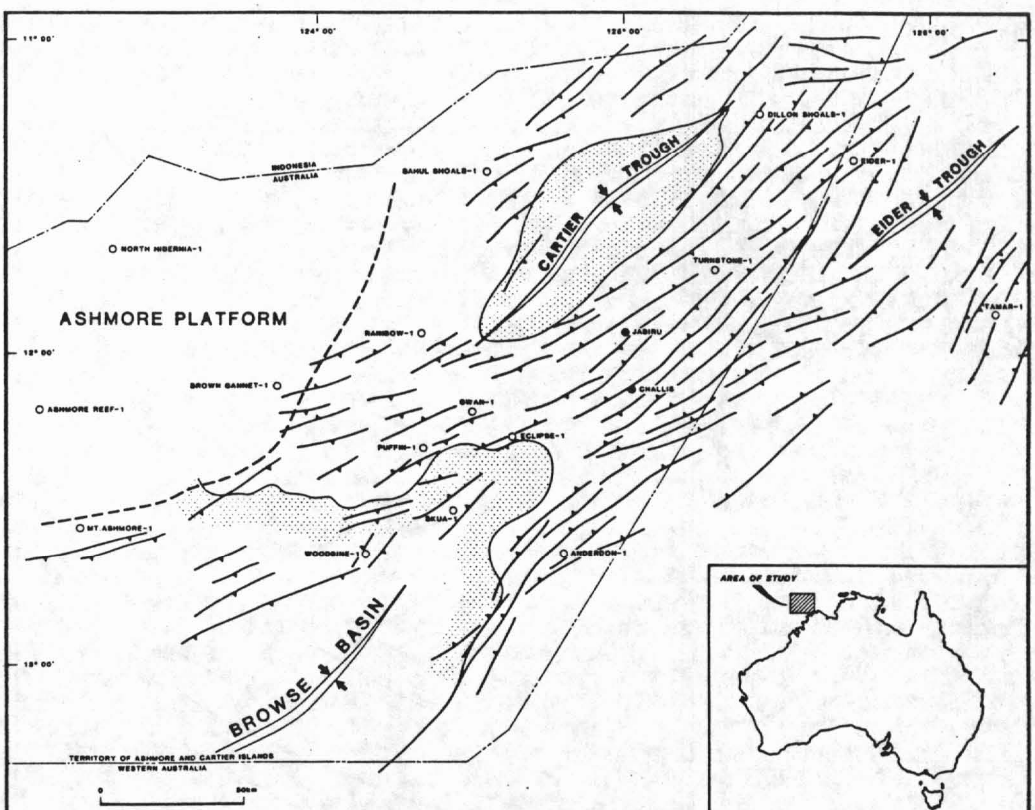

Figure 2. Northern Browse Basin, showing main structural features. Cartier Trough outlined by Miocene-Recent 1000m isopach and northern end of Browse Basin outlined by Cretaceous-Paleocene 1000m isopach.

(Not all exploration wells are shown)

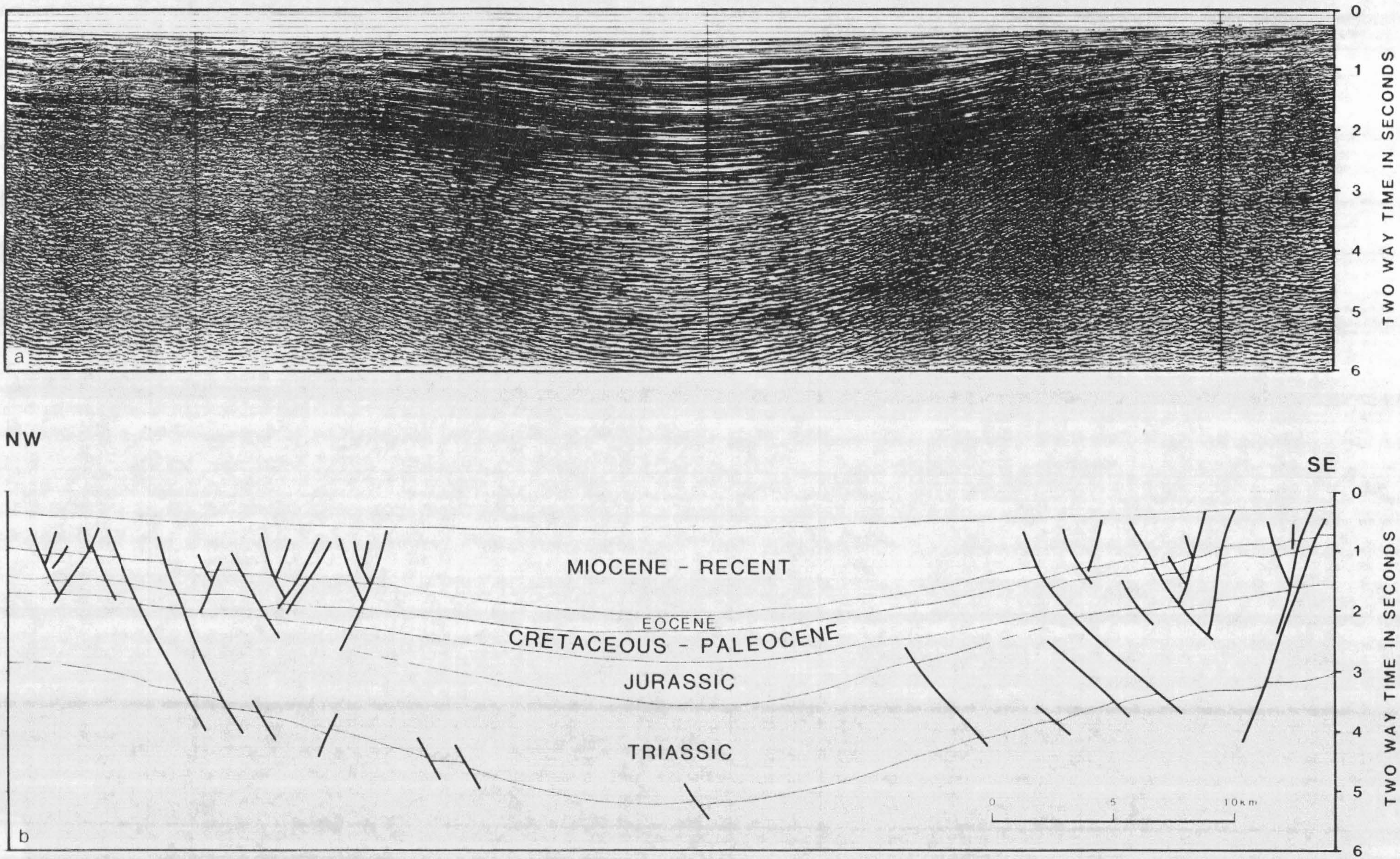

Figure 3. Central Cartier Trough, illustrated by a. northwest southeast oriented seismic section and b. interpretation of seismic section Note the vertical scale is two way time in seconds

The Bowen Basin, Queensland, Australia: an upper crustal extension model for its early history

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Introduction

The Bowen Basin, occurring in eastern central Queensland, contains a locally thick, Permian to mid-Triassic succession of fluvio-deltaic and shallow marine sedimentary rocks. It is currently the source of most of Australia's export black coal and contains significant gas reserves.

Aims

The intentions of this work are to briefly outline the origin of an extension model for the Bowen Basin, and to illustrate the extent to which extensional fault patterns have influenced subsequent regional deformation. A brief assessment of the importance of the models with respect to mining and exploration activity is also made.

Setting

The Bowen Basin is recognised as the northern continuation of the composite Bowen-Gunnedah-Sydney Basin (Murray, 1985) which is developed in the hinterland of the New England Orogen of eastern Queensland and northeastern New South Wales (Fig. 1). It is overlain, and obscured over large areas, by Jurassic rocks of the Surat Basin.

The Bowen Basin sequence was deposited on a basement of Silurian, Devonian and Carboniferous rocks, largely with magmatic or marine sedimentary origins, and belonging to the New England Orogen. Early Palaeozoic metamorphic rocks of the Anakie Metamorphics are exposed to the west and also locally form basement to the basin. The New England Orogen has been interpreted as a volcanic arc, fore-arc basin and accretionary prism association, although the details from model to model differ widely (e.g., Harrington & Korsch, 1985; Murray & others, 1987).

The conventional interpretation of the Bowen Basin is that it represents a retro-arc basin associated with the resumption of westward subduction in the Early Permian (Murray, 1985; Murray & others, 1987). Uplifted basement terranes along the basin's eastern margin (Connors & Auburn Arches) are partially mantled by arc-type volcanics ("Camboon Volcanic Arc" after Day & others, 1978) which correspond in age to the oldest sediments present in the basin.

Early Permian extension

Superficially it is difficult to envisage a possible role for crustal extension models (e.g. Lister & others, 1986) in the settings outlined above. However, evidence exists that indicates rifting and the development of oceanic basins (probably not floored by oceanic crust) in the Early Permian.

1. Marine basaltic volcanism and flysch sedimentation is preserved in the Gogango Overfolded Zone (Fig. 2).
2. Similar ocean basin associations occur in the Gympie terrane (central New England Orogen).
3. Bimodal rift-type volcanics in the Sydney and Gunnedah Basins.

Although not conclusive, and commonly interpreted as localised basins formed along major strike slip zones (Harrington & Korsch, 1985; Murray & others, 1987), these occurrences lend considerable weight to the following interpretation of the Bowen Basin.

Gross structure of the Bowen Basin

The basin can be readily subdivided into morpho-tectonic features as follows:

i) Springsure Shelf (basement high to southwest)
ii) Denison Trough (early marginal sub-basin to southwest)
iii) Collinsville Shelf (basement high to northwest)
iv) Comet Ridge (basement high between Denison and Taroom Troughs)
v) Taroom Trough (= Mimosa Syncline - synclinorium with very thick Late Permian and Early Triassic sequence - basin axis)
vi) Dawson Tectonic Zone (restricted belt of close- to tightly-folded Late Permian rocks)
vii) Nebo Synclinorium (= northern extension of Taroom Trough - discontinuous large-scale synclines).

In addition to these elements, basement rocks along the basin's eastern margin readily divide into 3 separate belts (Fig. 2):

i) Connors Arch - Carboniferous magmatic terrane
ii) Gogango Overfolded Zone - Thrust-fold style deformed zone entraining
  (a) Early Permian intermediate volcanics, volcanogenic sediments and ocean basin-type rocks (Grantleigh Trough = basal Bowen Basin sequence), and
  (b) Silurian to Carboniferous volcano-sedimentary basement rocks.
iii) Auburn Arch - Carboniferous magmatic terrane.

Thrusting

The structure of the basin is superficially dominated by thrusting. Hobbs (1985) identified a structural zonation across the central part of the basin with which he drew parallels to the Appalachian Orogen in North America, a well-documented thrust terrane (e.g., Cook & others, 1979). In such a thrust interpretation, the basement terranes along the basin's eastern margin would have been uplifted on buried ramps, of which one is apparently exposed in the north (Fig. 2). Similarly, the Gogango Overfolded Zone can be interpreted as shallower-level thrust imbrication of rocks equivalent to the Bowen Basin sequence with less basement involvement.

In addition, the morphology of the basin changes from north to south in a step-like manner across numerous transverse zones that can be interpreted as buried tear faults and lateral ramps. Some of these transverse zones, or "corridors" mark significant regional changes in the basin's character (Fig. 2).

Smaller-scale structures are consistent with this assessment, and in particular include spectacular meso- and macroscopic thrust structures exposed in some of the open-cut coal mines.

Extension models

Ziolkowski & Taylor (1985) assembled deposition and deformation models for the Denison Trough. They illustrated the development of deep half-grabens in the earliest Permian, at the time of the accumulation of thick sequences of largely intermediate volcanics along the present eastern margin of the basin (Fig. 3). Two compressional episodes partially inverted half-grabens in the Denison Trough during the deposition of most of the remaining Bowen Basin sequence. The younger episode culminated in a thrust event in the mid-Triassic.

Despite the retro-arc basin interpretation of the Bowen Basin, the unequivocal evidence for an early extensional episode has prompted the application of extensional models (e.g. Lister & others, 1986) in an endeavour to explain aspects of the basin's geology. In particular:

1. the rapid development of an oceanic trough to the east of the present day Bowen Basin in the earliest Permian,
2. the unusual transverse corridors that effectively divide the basin into domains with distinctly different characters,
3. the "out of place" occurrence of major reverse faults (e.g., the Shotover Thrust and Jellinbah Fault),
4. the abrupt and faulted western limit of early Permian volcanics immediately north of the basin (Fig. 3),
5. morpho-tectonic features in the basin like the Comet Ridge and basement arches in the east where the succession is relatively thin,
6. and the mid- to late Permian regional sag phase (Ziolkowski & Taylor, 1985) eventuating in coal-measure deposition.

An extension model involving localised belts of half-graben development explains the transverse corridors as major transfer faults (Gibbs, 1984). Their present character can be attributed to subsequent utilisation as lateral ramps and tear faults. The apparent localisation of intrusive activity on or near corridors is consistent with the deep-seated nature of transfer faults. Major, out of place, reverse faults can be interpreted as half-graben bounding faults reactivated during mid-Triassic thrusting (Fig. 4).

Such a model also explains the rapid development of oceanic sedimentation east of the present basin as an area of more extreme extension, and the apparently restricted extent of Early Permian volcanics in the north as a consequence of their confinement to half-grabens.

The Comet Ridge can be interpreted as a marginal plateau-like area where half-grabens did not develop, and which was potentially enhanced by the immediately post-extension partial-inversion episode (Ziolkowski & Taylor, 1985). Similar arguments can be applied to basement arches on the basin's eastern margin, though they were subsequently thrust-over the basin on reactivated extensional faults (Fig. 4).

A sag phase in the mid- to Late Permian would also be consistent with the widespread extension (and upper crustal thinning) that is proposed for the earliest Permian.

Implications

The Bowen Basin can be subdivided into domains that contain internally-consistent basinal structure, but which are commonly markedly different from adjacent domains. This provides those concerned with coal-mining operations with an enhanced ability to predict more accurately the type and style of structure likely to be encountered. This type of work is being actively pursued at CSIRO's Division of Geomechanics.

The nature of the extension models provides for regionally significant transfer fault systems that are necessarily deep-seated structures providing an ideal "plumbing" system for the rise and emplacement of magma and fluid circulation. Numerous intrusives, some large, mark the trace of postulated transfer faults. In addition, these features appear to have been active during Cretaceous and Tertiary tectonism. These aspects indicate the importance of this extension model as an exploration tool in the region, and potentially the importance of recognising extensional episodes and identifying their related fault patterns in any terrane. This observation is particularly appropriate for Australia, given the abundance of preserved passive margin and the inferred extension episodes.

As a closing comment it should be indicated that an Early Permian extensional episode has not yet been recognised or considered in tectonic reconstructions of the New England Orogen.

References

Cook, F.A., Albaugh, D.S., Brown, L.D., Kaufman, S., Oliver, J.E., & Hatcher, R.D., 1979 - Thin skinned tectonics in the crystalline southern Appalachians; COCORP seismic-reflection profiling of the Blue Ridge and Piedmont. Geology, 7, 563-567.

Day, R.W., Murray, C.G., & Whitaker, W.G., 1978 - The eastern part of the Tasman Orogenic Zone. Tectonophys., 48, 327-364.

Gibbs, A.D., 1984 - Structural evolution of extensional basin margins. J. Geol. Soc. Lond., 141, 609-620.

Harrington, H.J., & Korsch, R.J., 1985 - Tectonic model for the Devonian to middle Permian of the New England Orogen. Aust. J. Earth Sci., 32, 163-179.

Hobbs, B.E., 1985 - Interpretation and analysis of structure in the Bowen Basin. Geol. Soc. Aust., Abs., 17, 151.

Lister, G.S., Etheridge, M.A., & Symonds, P.A., 1986 - Detachment faulting and the evolution of passive continental margins. Geology, 14, 246-250.

Murray, C.G., 1985 - Tectonic setting of the Bowen Basin. Geol. Soc. Aust., Abs., 17, 5-16.

Murray, C.G., Fergusson, C.L., Flood, P.G., Whitaker, W.G., & Korsch, R.J., 1987 - Plate tectonic model for the Carboniferous evolution of the New England Fold Belt. Aust. J. Earth Sci, 34, 213-236.

Ziolkowski, V. & Taylor, R., 1985 - Regional structure of the north Denison Trough. Geol. Soc. Aust., Abs., 17, 129-137.

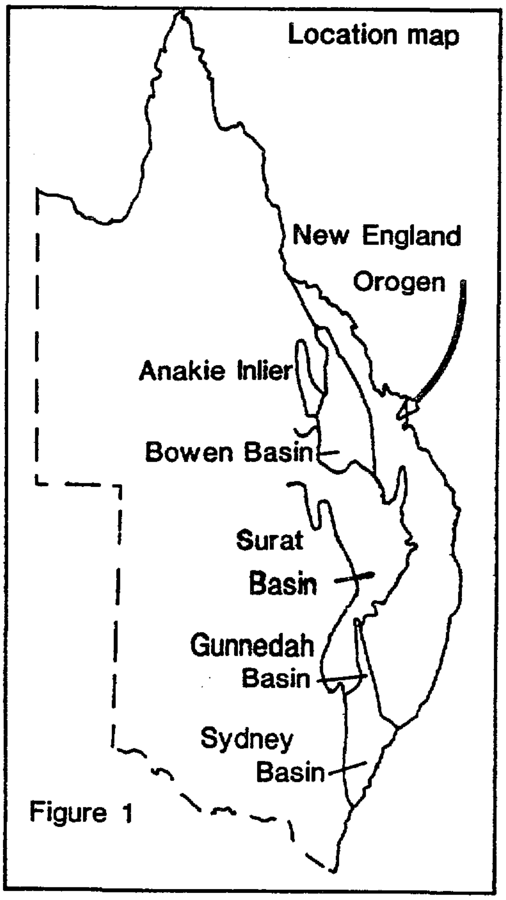

Figure 1. Illustration of the distribution of Permian and younger depositional basins in eastern Australia.

Bowen Basin - Coal measure sequences.

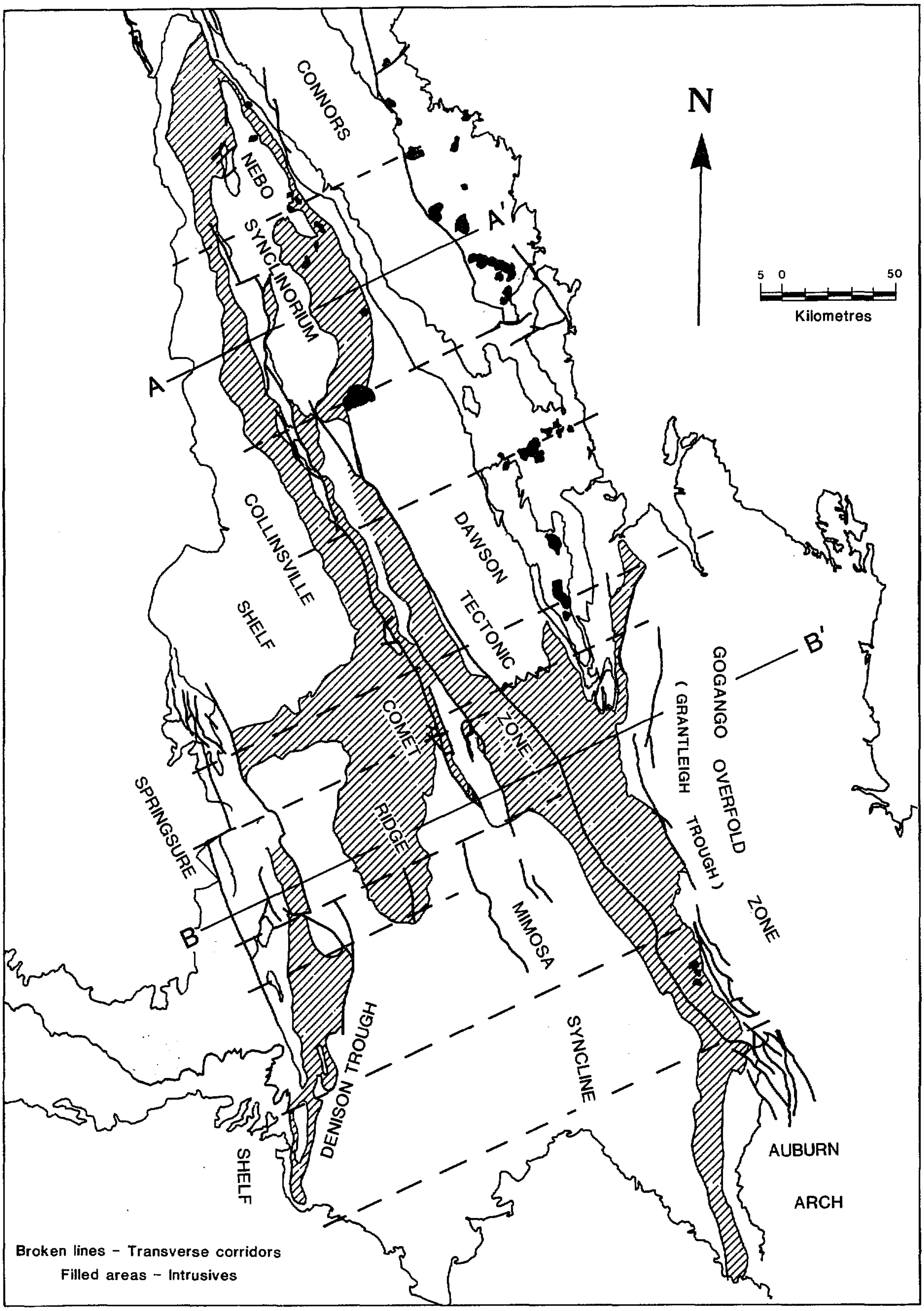

Figure 2. Regional sketch map of the Bowen Basin illustrating the distribution of mid- to late Permian coal-measure sequences (sag phase) and the gross morphology of the basin. Names refer to morpho-tectonic elements, filled areas are larger Cretaceous acid intrusives, broken lines represent known major corridors, and lines A-A' and B-B' locate cross-sections depicted in Figure 4.

Exposed Early Permian Units

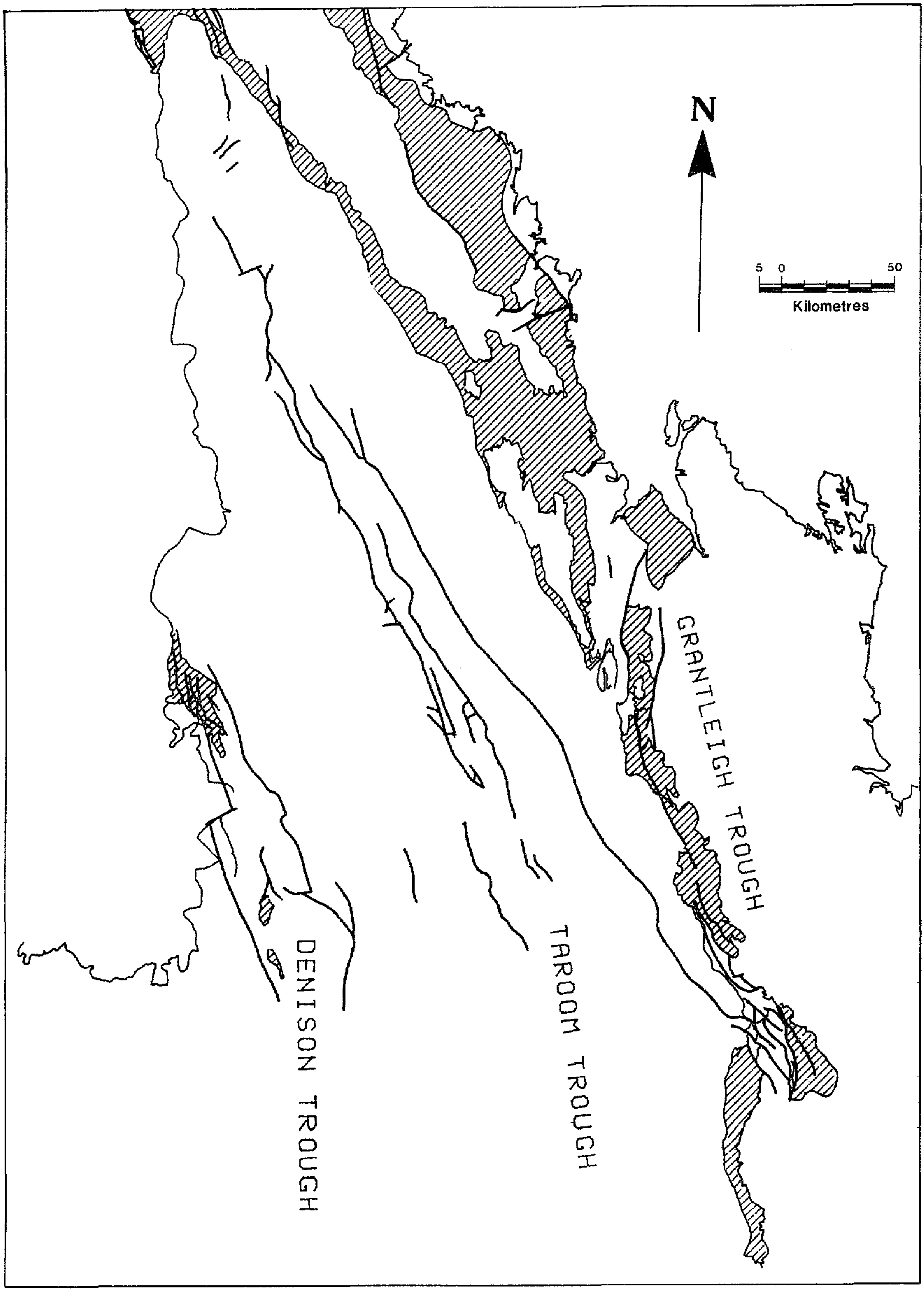

Figure 3. Regional sketch map of the Bowen Basin illustrating the distribution of exposed Early Permian units (extension phase). Lacustrine coal-measure rocks are abundant sub-surface in the belt marked 'Denison Trough'. Exposures along the east side of the basin are largely volcanogenic. Note the abrupt-faulted western limit of early Permian units in the far north, and the exposed core of basement (Connors Arch) on which the sequence is thinner.

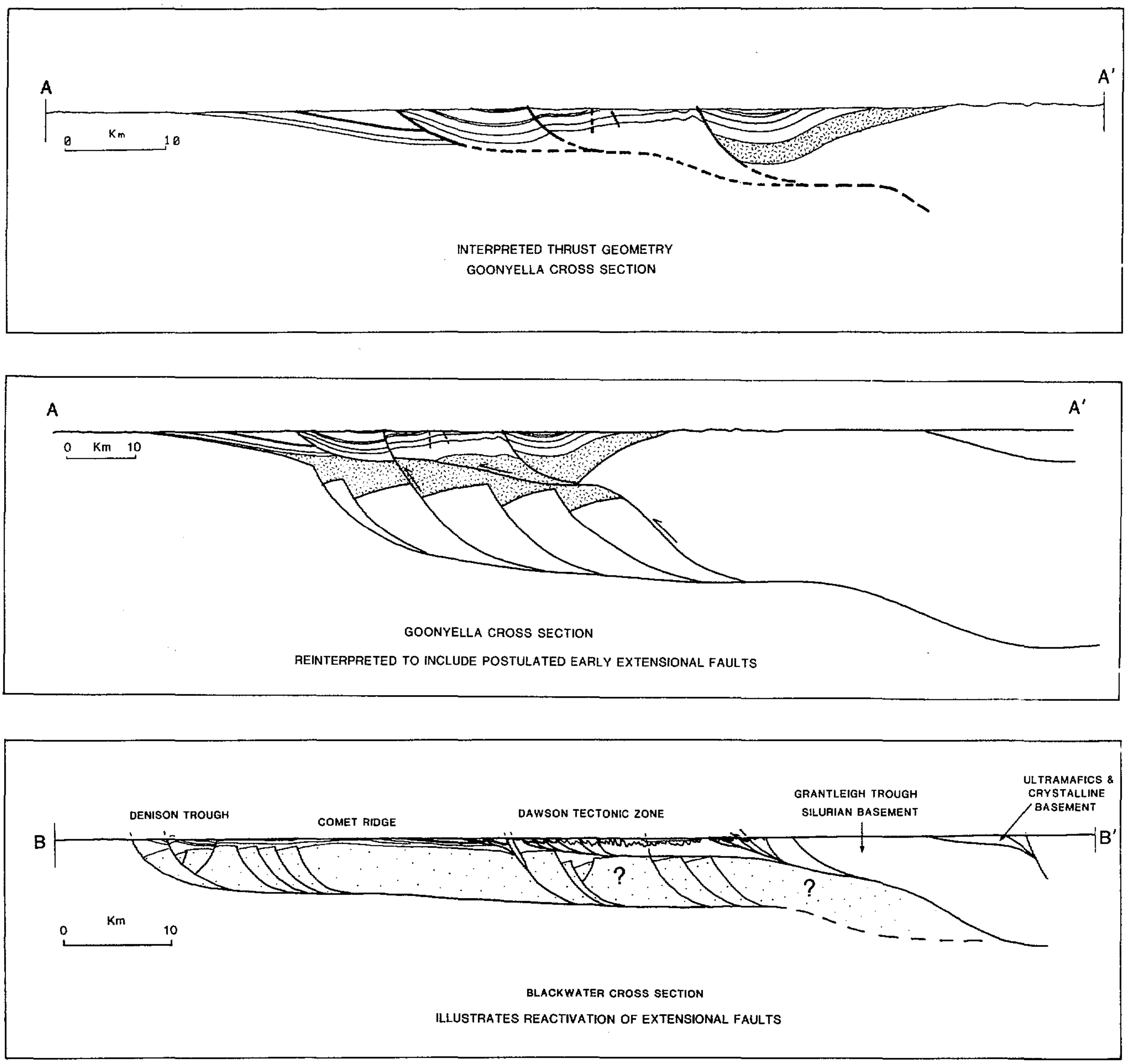

Figure 4. Two interpretations of the section A-A' are depicted (note vertical exaggeration). The upper section is a conventional thrust interpretation, whereas the lower attempts to account for the abrupt western limit of volcanics in the far north, the presence of 'out-of-place' reverse faults, and better account for the character of the Connors Arch (Fig. 2) rather than simply infer a buried basement ramp. The section B-B' (no vertical exaggeration) illustrates the greater complexity of the southern Bowen Basin, the origin of features like the Comet Ridge and Denison Trough (Fig. 2), and the much greater shortening (and hence possibly also extension initially) required to account for the Folded Zone (Dawson Tectonic Zone). (Note that scale bar on section B-B' should read 0 to 30 km.)

An extensional model for the formation of the Surat Basin, eastern Queensland, Australia, based on deep seismic profiling

K Wake-Dyster, B J Drummond, R J Korsch, D M Finlayson, M J Sexton, D W Johnstone & R Bracewell, BMR

The Surat Basin formed in the Jurassic and Cretaceous across a Palaeozoic basement in southern Queensland. Seismic profiling across the region (Fig. 1) by the Bureau of Mineral Resources (BMR) has produced an image of the crust that suggests an extensional model for the formation of the basin. A line drawing and interpretation of the seismic data are shown in Figure 2.

Several features are clear on the seismic section. The Surat Basin contains a predominantly flat-lying blanket of sediments marked 'Sag' in Figure 2b. They are draped over the Nebine Ridge in the west and over two asymmetric packages of sediments lying in fault-bounded sub-basins in the east ('Rift'). The sub-basins are bounded by normal faults along their eastern margins and by hinge-lines in the west, and are sub-surface correlatives of the Permo-Triassic Sydney-Gunnedah-Bowen Basins.

Below the sediments, the crust is characterised by numerous discontinuous reflections. The base of the crust is interpreted as the bottom of the reflective zone; farther west in the Eromanga Basin, the base of the reflective zone correlates with the Moho defined by seismic refraction profiling. The Moho in Figure 2 has considerable topography. It is deepest in the west but shallows abruptly near station 2000 along a band of reflections which penetrates the entire crust (AA'). Farther east it shallows again by about 2-3 s (6-9 km) over a distance of about 120 km between stations 2750 and 4250. In the region where the Moho flattens (below station 4500), numerous reflections project upwards to the east towards the base of the faults that bound the rifted basins.

The seismic data suggest a tectonic model for the region in which the older Permo-Triassic sub-basins formed as the rift phase of an extensional event associated with a major detachment penetrating through the crust and displacing the crust/mantle boundary. The crust was extended along the detachment zone by about 120 km, and in the vicinity of station 4250 was thinned by 6-9 km. The fault between the two sub-basins was acting as a transcurrent fault as late as the Cretaceous. The BMR seismic data do not indicate whether the rifting formed a single sub-basin that was later disrupted by transcurrent faulting to form the two sub-basins seen in Figure 2, or whether the eastern rift basin formed above a rotated fault block associated with the detachment. The Surat Basin then formed during the thermal subsidence phase after the cessation of rifting.

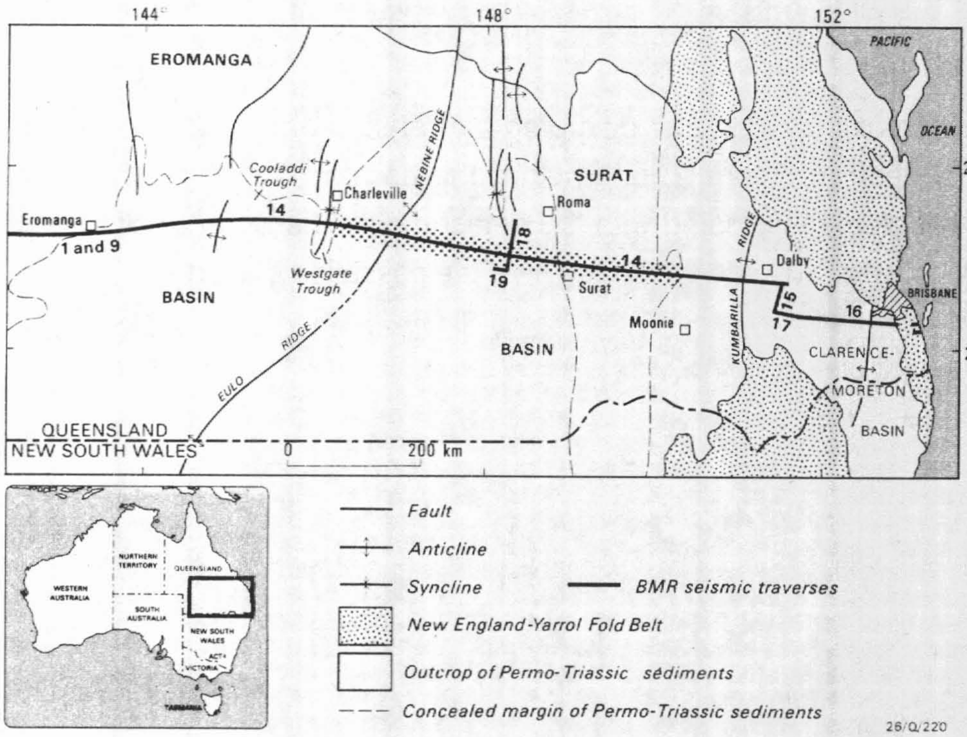

Figure 1: Location map showing BMR traverses and geological setting in southern Queensland. Line drawings of the data from the hachured part of the seismic traverses are shown in Figure 2.

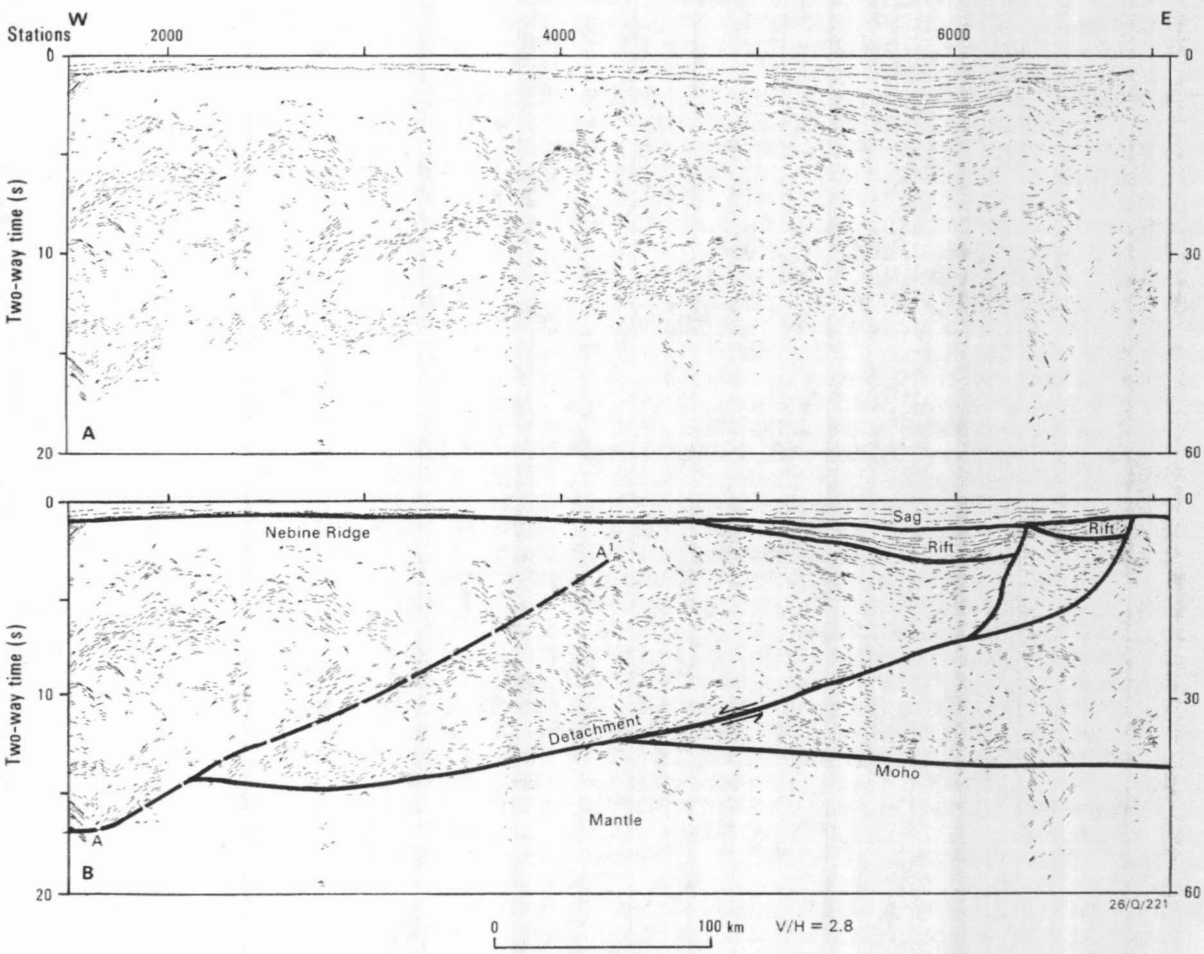

Figure 2: (a) Line drawing and (b) interpretion of the seismic data along a 450 km-long section of traverse in southern Queensland.

The geometry of extensional structures in the Fitzroy Trough, Canning Basin, Western Australia

B J Drummond, M A Etheridge, (BMR) & M F Middleton (Geological Survey Western Australia)

The Canning Basin lies in northwest Australia. It contains Palaeozoic sediments in a series of sub-basins with an overall northwest/southeast tectonic grain, and a Mesozoic sequence, restricted mainly to the offshore areas, deposited during and after the breakup of the northwest continental margin. The basin formed on a Precambrian basement between the Kimberley and Pilbara Cratons. Summaries of the geological evolution of the basin (eg., Brown & others, 1984) suggest that it formed in the Ordovician as a broad sag, followed by rifting in the Middle Devonian to Early Carboniferous to form the Fitzroy Graben along the northeastern margin of the basin (Fig. 1). Many of the normal faults associated with the rifting phase were reactivated as dextral wrench faults during the breakup of the continent in the Mesozoic, in some areas resulting in complicated structural patterns.

The onshore Fitzroy Graben has 2 major depocentres: the Fitzroy Trough (Fitzroy Sub-basin in Figure 1) in the northwest and the Gregory Sub-basin in the east. Traditionally, the graben has been regarded as symmetric (eg., Brown & others, 1984; see also Fig. 2), with networks of major normal faults bounding both the northeastern and southwestern margins. Begg (1986) reinterpreted much company seismic reflection data from the Fitzroy Trough, and proposed that it was segmented by numerous transfer faults that trend perpendicular to the rift margins. However, he retained the concept of an essentially symmetric graben.

As part of the planning for a deep seismic profiling program in the Canning Basin, we examined selected company seismic traverses from the northeastern and southwestern margins of the Fitzroy Trough and conclude that, at any place along strike, the trough is an asymmetric rift, with a normally-faulted margin on one side and a flexed margin, or hinge, on the opposite side. There are insufficient data in the centre of the trough to determine whether it is a single half-graben or an array of parallel half-graben. Furthermore, in some places the faulted margin is on the southwestern side, while in other places it is on the other side, so that the sense of asymmetry switches along the axis of the basin.

The switching of asymmetry occurs at transfer faults, most of which coincide with the major transfer zones recognised by Begg (1987). In addition, other transfer faults accommodate small to moderate offsets in the boundary fault of the half-graben. A cartoon sketch of our preferred faulting style is shown in Figure 3. The gross symmetry of the trough results from the repeated (at least three times) switching in asymmetry along its length. The character of the individual margin-segments is also obscured by the Mesozoic inversion, which commonly produced reverse movements on both faulted and hinged margins. On hinged margins, the reverse reactivation occurred on both the basement surface and on one or more early Palaeozoic horizons.

The asymmetric character of the half-graben and the widespread transfer faulting in the Fitzroy Trough have important implications for petroleum and mineral exploration. The environments of deposition were markedly different on opposed margins, with rapidly subsiding blocks on the rifted side and more gentle subsidence on the hinged side. The sections and stratigraphic column of Towner & Gibson (1983) show platform and marginal basin conditions on the northeastern margin where we have mapped a hinged boundary, and marine conditions where a faulted boundary occurs. The different conditions would have had a marked effect on the development of the Devonian reef complexes that formed on the margins of the trough. The reefs are a popular play for petroleum explorers. Those on a gradually subsiding hingeline would have been able to respond to changes in sea level by moving out of and into the trough as the seas transgressed and regressed, while pinnacle-reefs are more likely to have developed on the subsiding fault blocks.

The major normal faults also acted as conduits for the migration of petroleum, and also the fluids responsible for the deposition of Mississippi Valley-type lead/zinc deposits. The absence of major faults on the side with the hingeline may affect prospectivity. The transfer faults which mark the switching of asymmetry also serve to relieve stresses within the basin sediments. They are often marked by zones of crushing or secondary porosity that may serve as suitable repositories for minerals and petroleum.

References

Begg, J., 1987 - Structuring and controls on Devonian reef development, northwest Barbwire and adjacent terraces, Canning Basin. Australian Petroleum Exploration Association, Journal, 27(1), 137-151.

Brown, S.A., Boserio, I.M., Jackson, K.S. & Spence, K.W., 1984 - The geological evolution of the Canning Basin - implications for petroleum exploration. In Purcell P.G., (Editor) - The Canning Basin, WA. Proceedings of the Geological Society of Australia/ Petroleum Exploration Society of Australia Symposium, Perth, 1984, 85-96.

Etheridge, M.A., Branson, J.C. & Stuart-Smith, P.G., 1985 - Extension basin-forming structures in Bass Strait and their importance for hydrocarbon exploration. Australian Petroleum Exploration Association Journal, 26(1), 344-361.

Towner, R.R. & Gibson, D.L., 1983 - Geology of the onshore Canning Basin, Western Australia. Bureau of Mineral Resources, Australia, Bulletin, 215.

Yeates, A.N., Gibson, D.L., Towner, R.R., & Crowe, R.W.A., 1984 - Regional geology of the onshore Canning Basin, WA. In Purcell P.G., (Editor) - The Canning Basin, WA. Proceedings of the Geological Society of Australia/Petroleum Exploration Society of Australia Symposium, Perth, 1984, 23-55.

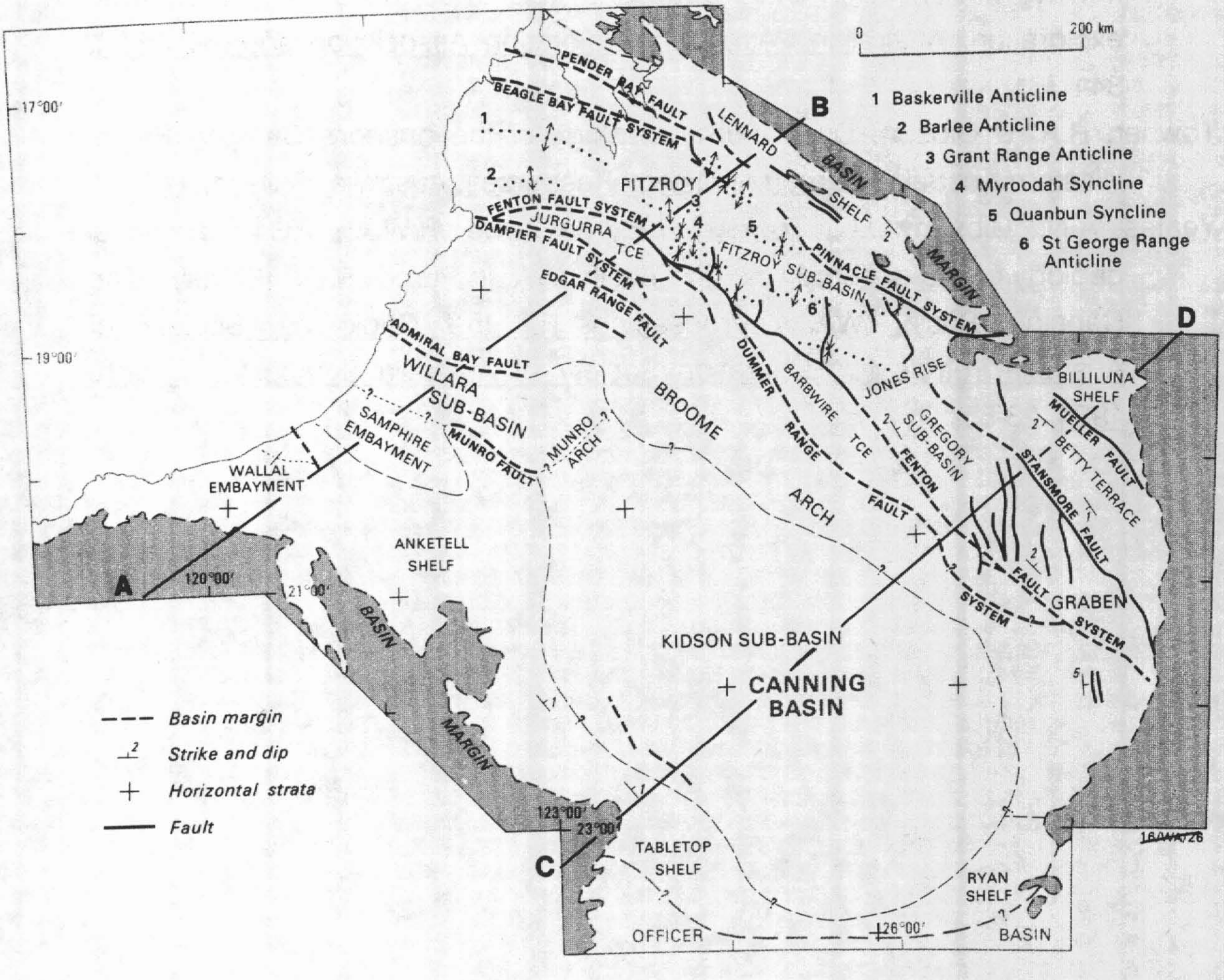

Figure 1: Structural features of the Canning Basin (from Yeates & others, 1984, Figure 4).

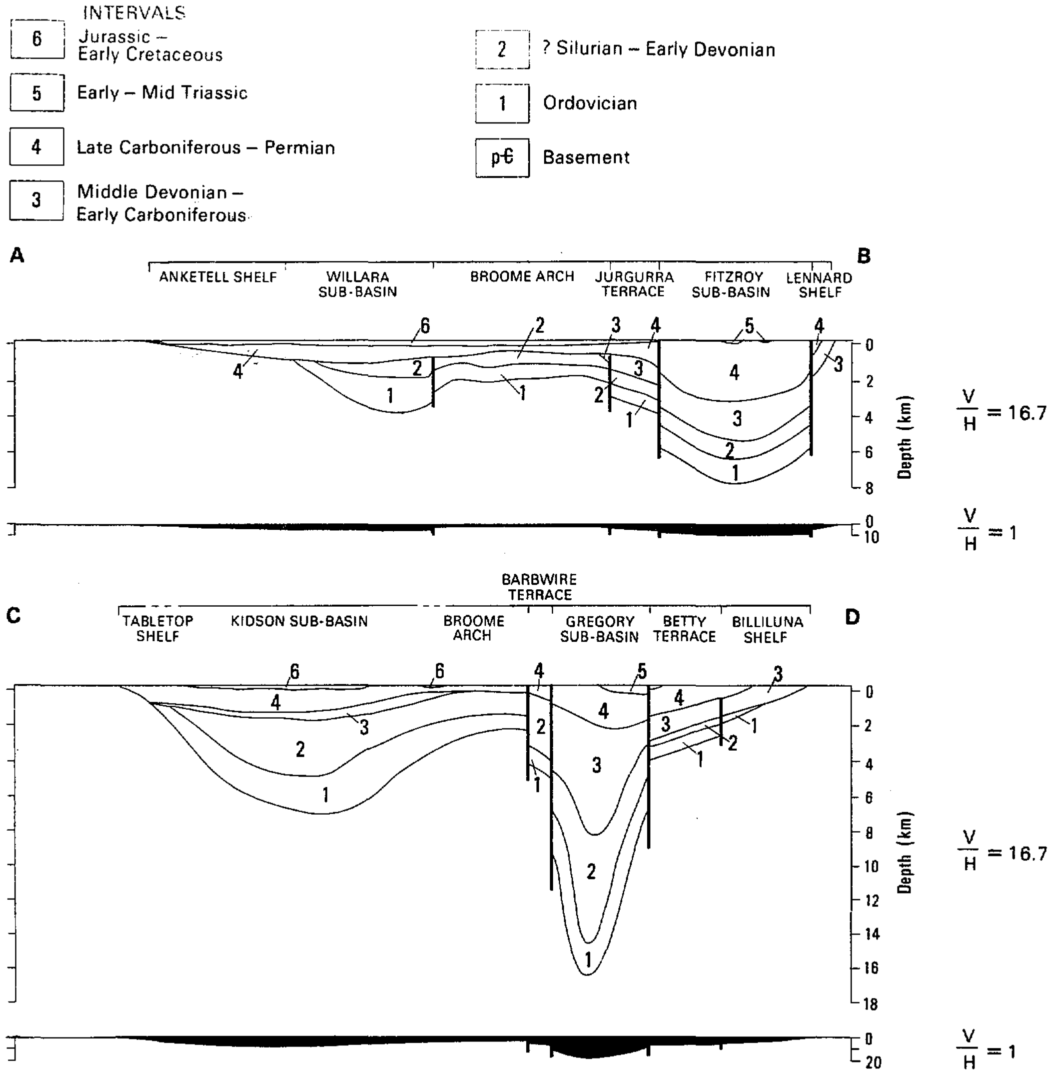

Figure 2: Cross-sections based on the traditional understanding of the Fitzroy Trough/ Gregory Sub-basin as symmetric grabens with major normal faults bounding both margins. The positions of the sections are shown in Figure 1 (based on Yeates & others, 1984, Figure 3).

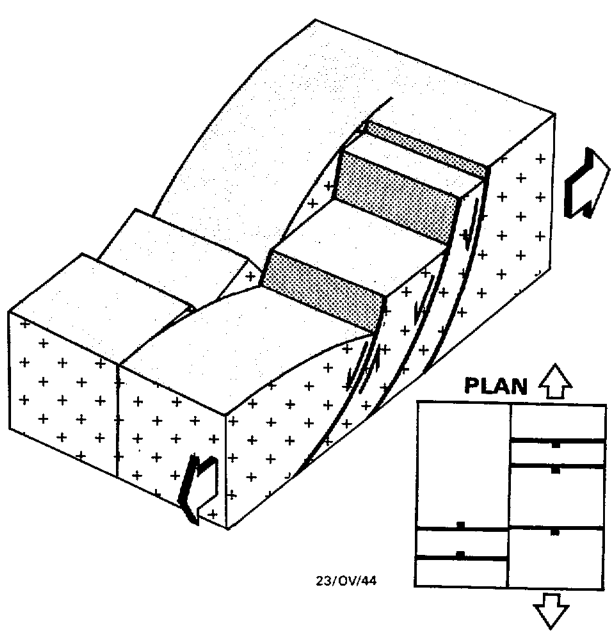

Figure 3: Cartoon sketch of the extensional and transfer fault array in the basement of an extensional basin such as the Fitzroy Trough (Etheridge & others, 1985, Figure 14a)

Modern processes in a continental rift lake: analogues for synrift facies distribution

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Examination of over 12 000 km of 28 kHz echosounder profiles from Lake Malawi, Africa, has produced a snapshot of structural and depositional processes in the synrift environment. Fracture frequency is sporadic, showing apparent concentrations of activity on the master border-fault-systems, on the shoaling sides of the northern two half-grabens, and on some of the accommodation zones that occur between adjacent half-grabens of opposite polarity. Areas of structural quiescence are associated with the first wedge blocks lakeward of the northern border-fault-complexes. Apparent quiescence is also observed lakeward of the intersection of the current rift zone and the Permo-Triassic Ruhuru Trough where large sediment loads are entering the lake. There is a general increase in fracture frequency in the northern half of the lake, resulting in a more dynamic sedimentary environment than in the south. Recognition of the dynamism of sedimentary processes not normally associated with anoxic, deep-water basins is perhaps the most significant result of this study.

Classification and correlation of echo character shows that the northern border-fault-complexes are buried in fanglomerates. Evidence of mass wasting is common, including slump scars and debris flows. Coalescing fans create slope aprons, particularly lakeward of the northernmost border-fault-complex. Structurally-guided channels with well-developed levees are common even in the deepest parts of the northern basins. A large, sublacustrine fan originates from the mouth of the Ruhuru River, which drains a Karoo rift basin. This fan follows the Karoo trend (northeast-southwest) cross-cutting infra-basinal Cainozoic structures (north-south), across the lake for over 75 km, where it is finally diverted by the master border fault system, continuing southward for some 20 km. The eastward boundary of the southern lobe is delimited by the western boundary of the accommodation zone created between the two facing, opposite polarity half-grabens which define this basin. The echo character of this facies is indicative of coarser-grained deposits and dynamic processes. Echo character of this type extends into the deepest part of the basin, where traditional models predict an undisturbed pelagic and/or biogenic facies. Accommodation zones often correlate precisely with acoustic facies distribution. However, the distinction is between two subclasses both of which have been inferred to represent open, pelagic deposition, with slightly coarser-grained material being deposited on top of the accommodation zones.

The apparent tectonic quiescence in the south is mimicked in the acoustic stratigraphy, which has been documented to result from blanket deposition of diatomaceous ooze. The uniformity of this blanket is disturbed locally by bottom currents, especially in the region of the southern bottleneck. Even the near-shore depositional environment appears relatively static, compared to similar environments in the north. The border-fault-complexes lack the fanglomeratic echo-facies that characterise their northern counterparts, suggesting that little subsidence has occurred along the southern border faults over the time-frame of these observations. There is a basic symmetry of echo-class patterns in the south that belies a modest structural asymmetry.

Tectonic evolution of the central Southern Margin of Australia

B Willcox, P Symonds & H Stagg, BMR

The southern margin of Australia was originally thought of as a relatively simple, classical passive margin (Sproll & Dietz, 1969; Smith & Hallam, 1970; Griffiths, 1971), which commenced its development as a thermogenically-derived rift-zone (or ? valley) in the Early Cretaceous (Falvey, 1974), and culminated with breakup and north-south seafloor spreading in the Paleocene (Anomaly 22, Weissel & Hayes, 1972; Fig. 1). The rift within Gondwanaland - possibly a propagating rift - was envisaged to extend for over 3000 km, from the Naturaliste Plateau region in the southwest to the South Tasman Rise region in the southeast. A branch of this rift was considered to have extended into Bass Strait, being a precursor to the Bass and Gippsland Basin.

Investigations of the Southern Margin since the early 1970s have included regional studies of margin evolution (Boeuf & Doust, 1975; Deighton & others, 1976; Willcox, 1978, 1981; Falvey & Mutter, 1981; Lister & others, 1986); crustal structural studies (Karid & Talwani, 1977; Mutter, 1978; Talwani & others, 1979; Konig, 1980); seismic structural/stratigraphic studies of specific features such as the Ceduna Terrace (Fraser & Tilbury, 1979), Eyre Sub-basin (Bein & Taylor, 1981), Otway Basin (Williamson & others, 1987) and southeast Otway/Tasmania margin (Hinz & others, 1986); studies of seafloor spreading (Cande & Mutter, 1982; Veevers, 1986) and studies of the conjugate Antarctic margin (Eittreim & others, 1985; Veevers & Eittreim, in press). Continuing investigations of the central Southern Margin are presently underway, following two cruises by BMR's R/V 'Rig Seismic' in October - December 1986 (Willcox & others, in press; Fig. 6).

Several different lines of evidence have shed light upon the extensional history of the Southern Margin:

- Gravity Map of Australia (BMR, 1976; see also Haxby & others, 1983). This shows that many gravity provinces can be enclosed by parallelograms with sides (presumably fault-traces) trending approximately east-northeast and southeast.

- Eyre Sub-basin. This contains 6000 m of post-Middle Jurassic synrift and post-breakup sediments, including a marine transgressive sequence of late

Albian or Cenomanian age. It is structured apparently by east-northeast-trending normal faults and southeast-trending transfer faults (Figs 1, 3, & 4).

. Polda Trough. The depth of magnetic basement map presented by Nelson & others (1986) shows basement depressions which trend east-northeast, but are progressively offset to the southeast (Fig. 5).

. Great Australian Bight (GAB) Basin. Structural maps of the basement surface (Gawler Block) along the northern boundary of this deep basin (Fraser & Tilbury, 1979) are again, we believe, consistent with an east-northeast and southeast-trending configuration of faults (Fig. 1). The southwest flank of the Ceduna Terrace has yielded the 'transpressional' structures of pre-Cenomanian age as predicted by Willcox & others (1986). These presumably reflect an underlying transfer fault of major proportions.

. Duntroon Basin. This seems to exhibit the same east-northeast and southeast trends as the GAB Basin. Wrenching may have been an important factor in creating a central basin uplift.

. Otway, Bass & Gippsland Basins. Within this family of basins the extensional normal faults trend east-southeast, nearly orthogonal to those in the GAB Basin.

. Seafloor spreading anomalies. The anomalies south of the magnetic quiet zone (MQZ), originally identified as 19-22, have been remodelled by Cande & Mutter (1982) as anomalies 20-34, created during a phase of slow-spreading (Fig. 1). This revision, and a refinement by Veevers (1986), places breakup in the Cenomanian-Turonian, that is, 95 $\pm$ 5 Ma. Our study of anomaly-trends using existing Lamont-Doherty Geological Observatory data, and the trend of Anomaly 34 using 'Rig Seismic' profiles (Willcox & others, in press), suggests that, in detail, the oldest anomalies trend 075° and hence the initial direction of spreading was probably 165°.

Lister, Etheridge & Symonds (1986) have used the gross architecture of the Eyre Terrace region (Figs 1, 3, section X-X^1) to create a detachment model for the central Southern Margin (Fig. 2). The major implication is that a lower plate Australian margin has been pulled out from beneath an upper plate Antarctic margin. Highly-extended (200 per cent) remnants of the upper plate occur only beneath the MQZ, including most of the Ceduna Terrace. Regions with less extension (approx. 20 per cent) which have consequently undergone less subsidence - most notably the Eyre Sub-basin - occur north of the MQZ. It is probably the boundary of these zones of different extension (itself a detachment branch) that creates the prominent magnetic trough (MT).

The data now available indicate that a major change in the direction of extension, from approximately southeasterly on the central Southern Margin to north-northeasterly in the Otway, Bass and Gippsland Basin, must occur somewhere between the Duntroon Basin and Beachport Terrace (Fig. 1). We consider that a temporal change in rift development has also taken place: rifting having been pre-Middle Jurassic on the central Southern Margin and probably Early Cretaceous further east.

The geometry of the extensional systems indicates that wrench tectonics would have been prevalent, probably from the Jurassic through to Cenomanian, on the southwest flank of the Ceduna Terrace, in the proto- Otway Basin, and probably in the Duntroon Basin. The 'Rig Seismic' results confirm this scenario for the southwest Ceduna Terrace (Figs 7 & 8). The extent to which Jurassic strike-slip movements have created deep structuring in the Otway Basin, and in basins further east, is unknown. However, the configuration of features such as the Robe and Penola Troughs suggests that they are the products of extensional and strike-slip movements, respectively.

The detailed application of detachment models to the Southern Margin enables us to predict areas where extensional and wrench-type structures are likely to be found. Furthermore, the degree of extension should have a direct bearing on palæo-heatflow and hence basin maturity: for example, we would predict that a perched basin such as the Eyre Sub-basin (extension 20 per cent) would have been cool in comparison to more extended terranes such as the Ceduna Terrace (200 per cent). Such conclusions could lead to a more-focussed petroleum exploration effort. Further study is needed in areas such as the Duntroon Basin and western Otway Basin where the overprinting of extensional regimes may be present.

References

Bein, J., & Taylor, M.L., 1981 - The Eyre sub-basin: recent exploration results, Australian Petroleum Exploration Association Journal, 21, 91–98.

Boeuf, M.G., & Doust, H., 1975 - Structure and development of the southern margin of Australia. Australian Petroleum Exploration Association Journal, 15, 33-43.

Bureau of Mineral Resources, Australia, 1976 - Gravity map of Australia, scale 1:5 000 000.

Cande, S.C., & Mutter, J.C., 1982 - A revised identification of the oldest sea-floor spreading anomalies between Australia and Antarctica. Earth and Planetary Science Letters, 58, 151-160.

Deighton, I., Falvey, D.A., & Taylor, D.J., 1976 - Depositional environments and geotectonic framework - southern Australia continental margin. Australian Petroleum Exploration Association Journal, 16, 26-36.

Eittreim, S.L., Hampton, M.A., & Childs, J.R., 1985 - Seismic-reflection signature of Cretaceous continental breakup on the Wilkes Land margin. Antarctica. Science, 229, 1082-1084.

Falvey, D.A., 1974 - The development of continental margins in plate tectonic theory. Australian Petroleum Exploration Association Journal, 14, 95-106.

Falvey, D.A., and Mutter, J.C., 1981 - Regional plate tectonics and evolution of Australia's passive continental margins. BMR Journal Australian Geology & Geophysics, 6, 1-29.

Fraser, A.R., & Tilbury, L.A., 1979 - Structure and stratigraphy of the Ceduna Terrace region, Great Australian Bight. Australian Petroleum Exploration Association Journal, 19, 53-65.

Griffiths, J.R., 1971 - Continental margin tectonics and the evolution of southeastern Australia. Australian Petroleum Exploration Association Journal, 11, 75-79.

Haxby, W.F., Karner, G.D., La Breques, J.L., & Weissel, J.K., 1983 - EOS, American Geophysical Union Transactions, 64, 995-1004.

Hinz, K., Willcox, J.B., Whiticar, M., Kudrass, G.-R., Exon, N.F., & Feary, D.A., 1986 - The Western Tasmanian margin: an underrated petroleum province? In Second South-Eastern Australia Oil Exploration Symposium Technical Papers, 395-412.

Konig, M., 1980 - Geophysical investigations of the southern continental margin of Australia and the conjugate margin of East Antarctica. Ph.D. thesis, Columbia University, New York (unpubl.).

Konig, M., & Talwani, M., 1977 - A geophysical study of the southern continental margin of Australia - Great Australian Bight and western sections. Geological Society of America Bulletin, 81, 1000-1014.

Lister, G.S., Etheridge, M.A. & Symonds, P.A., - Detachment models for the formation of passive continental margins (Tectonics, in press).

Mutter, J.C., 1978 - An analysis of the Mesozoic-early Cenozoic rifting history of Australia and Antarctica - implications for generalized geodynamic models of rifting. M.Sc. thesis, Univ. Sydney (unpubl.).

Nelson, R.G., Crabb, T.N., & Gerdes, R.A., 1987 - A review of geophysical exploration in the Polda Basin, South Australia. Australian Petroleum Exploration Association Journal, 26(1), 319-333.

Smith, A.g., & Hallam, A., 1974 - The fit of the southern continents. Nature, 225, 139-144.

Sproll, W.P., & Dietz, R.S., 1969 - Morphological continental - drift fit of Australia and Antarctica. Nature, 222, 345-348.

Talwani, M., Mutter, M., Houtz, R., & Konig, M., 1979 - The crustal structure and evolution of the area underlying the magnetic quiet zone on the margin south of Australia. In Watkins, J., Montadert, L., & Dickenson, P., (Editors) - Geological and geophysical investigations of continental margin. American Association Petroleum Geologists Memoir, 151-175.

Veevers, J.J., 1986 - Breakup of Australia and Antarctica estimated as mid-Cretaceous (95 $\pm$ 5 Ma) from magnetic and seismic data at the continental margin. Earth Planetary Science Letters, 77, 91-99.

Veevers, J.J., & Eittreim, S.L. - Reconstruction of Antarctica and Australia at breakup (95 $\pm$ 5 Ma) and before rifting (160 Ma) (in press).

Willcox, J.B., 1978 - The Great Australian Bight - a regional interpretation of gravity, magnetic, and seismic data from the Continental Margin Survey. Bureau of Mineral Resources, Report, 201.

Willcox, J.B., 1981 - Petroleum prospectivity of Australian marginal plateaux. In Halbouty, M.T., (Editor) - Energy resources of the Pacific Region. American Association Petroleum Geologists, Studies in Geology, 12, 245-272.

Willcox, J.B., Stagg, H.M.J., & Davies, H.L., 1986 - Structure, stratigraphy and tectonic development of the Great Australian Bight Basin. Bureau of Mineral Resources, Australia, Research Cruise Proposal, 1C.08.

Willcox, J.B., Stagg, H.M.J., & Davies, H.L. - Cruise Report: Structure, stratigraphy and tectonic development of the Great Australian Bight Region. Bureau of Mineral Resources, Australia Report (in press).

Williamson, P.E., O'Brien, G.W., Swift, M.G., Felton, E.A., Scherl, A.S., Lock, J., Exon, N.F., Falvey, D.A., & Marlow, M., 1987 - Hydrocarbon potential of the offshore Otway Basin. Australian Petroleum Exploration Association Journal, 27(1), 178-195.

Fig 1

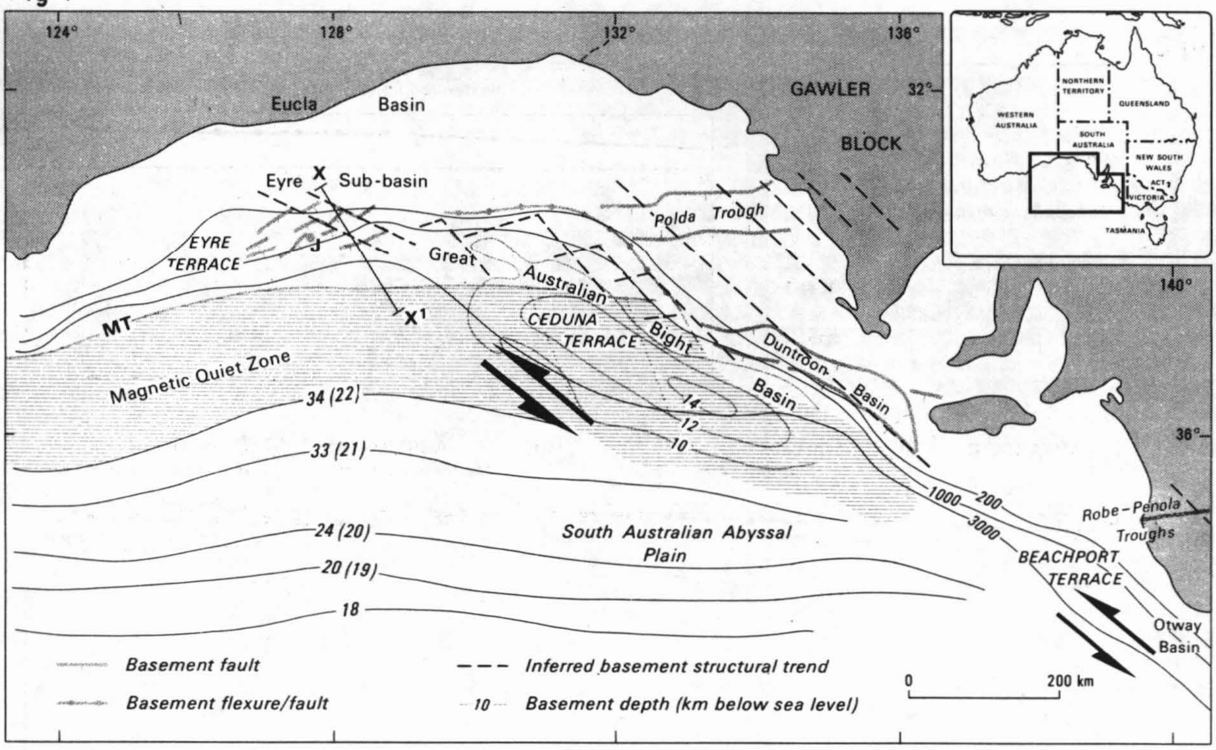

Figure 1. Map of the southern margin of Australia and part of the Southern Ocean, (modified from Lister, Etheridge & Symonds, in press), showing the major structural features of the passive margin, and the location of seismic profile (X-X') interpreted in Figure 3. Gross bathymetry shown by the 200, 1000, and 3000 m isobaths; numbered lines on the oceanwards side of the magnetic quiet zone (MQZ) (hachured) are seafloor-spreading magnetic anomaly traces interpreted by Cande & Mutter [1982] (previous interpretation by Weissel & Hayes [1972] shown in brackets). MT is the magnetic trough which defines the landward edge of MQZ. J is the location of Esso Jerboa-1 exploration well. Heavy arrows along southwest margins of Ceduna Terrace & Beachport Terrace (Otway Basin) indicate portions of the margin which are predicted to have been 'transpressional' or 'transtensional' zones during an extensional phase in the pre - Late Jurassic.

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Fig 2

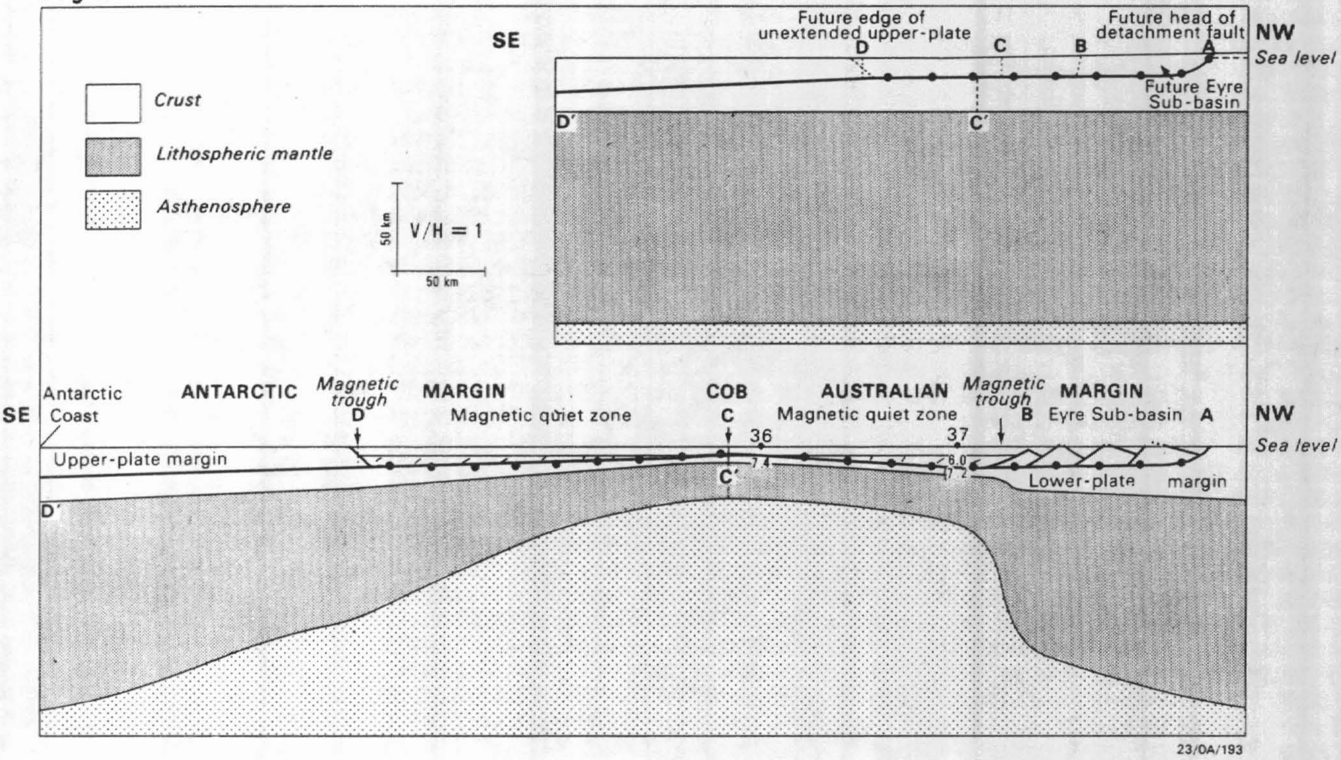

Figure 2. Balanced detachment model interpretation of the Eyre Terrace margin and its complementary Wilkes Land, Antarctica margin (below), with a palinspastic reconstruction (above) (from Lister, Etheridge & Symonds, in press). The detachment fault is shown by the beaded line. COB is the continent/ocean boundary; A, B, C, C', D and D' are control points common to both parts of the figure, and allow the effects of extension on various parts of the margin to be seen.

Fig 3

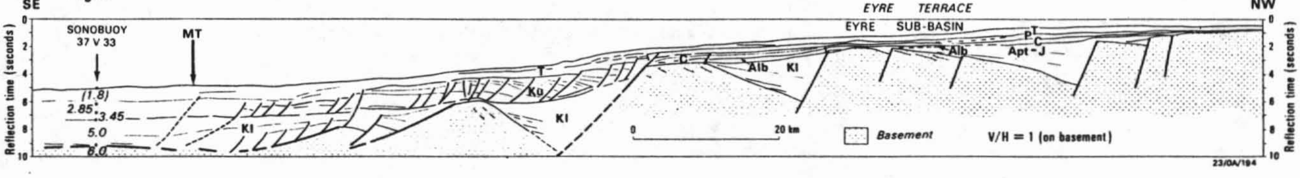

Figure 3. Line drawing from Shell seismic reflection profile X-X' across the Eyre Terrace and out onto the continental rise of the southern Australian margin (Location in Figure 1; from Lister, Etheridge & Symonds in press). It shows the ages of various sequences, T - Tertiary, P - Paleocene, C - Cenomanian, Alb - Albian, Apt - Aptian, Ku - Upper Cretaceous, Kl - Lower Cretaceous and J - Jurassic, based on stratigraphic ties to Jerboa - 1 (Bein & Taylor, 1981); refraction velocities from sonobuoy 37V33, MT indicates location of the magnetic trough.

Fig 4

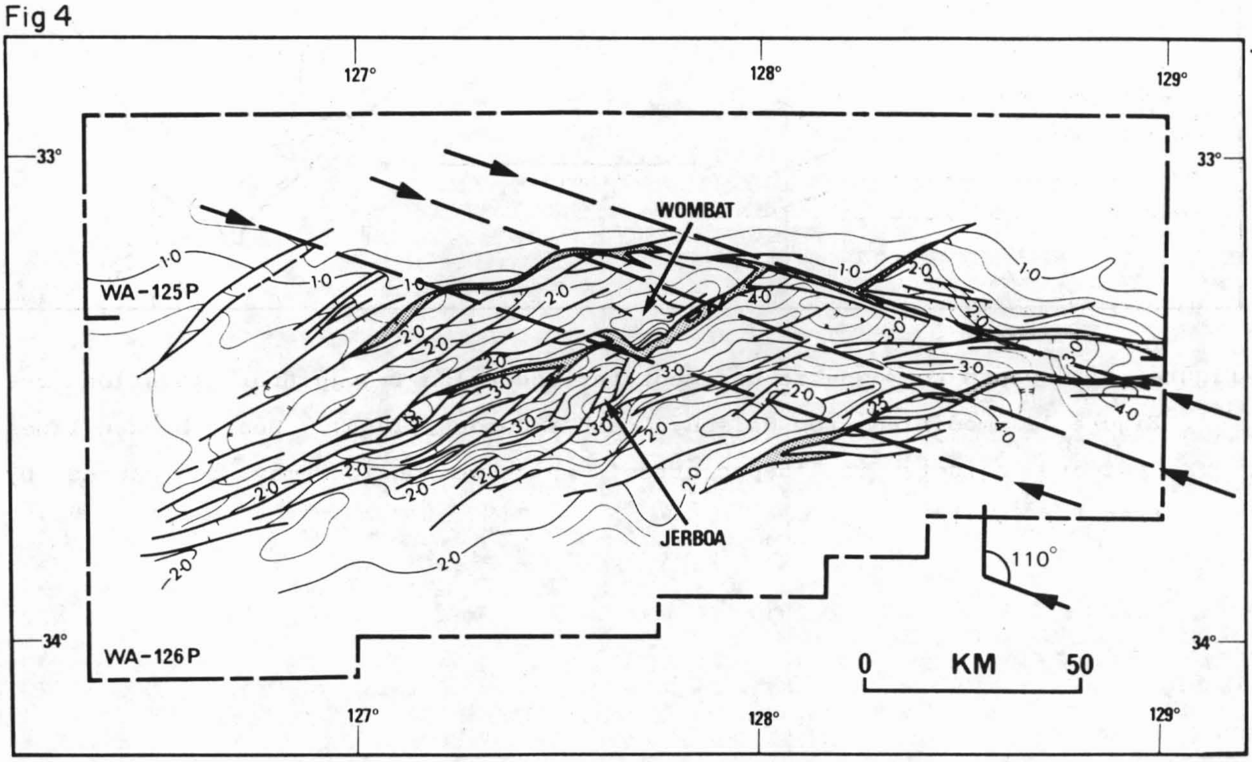

Figure 4. Time structure map on basement for the Eyre Sub-basin (modified from Bein & Taylor, 1981). Contour interval 0.25 seconds two-way reflection time. Heavy broken lines, trending 110°, are our interpreted transfer fault locations.

Fig 5

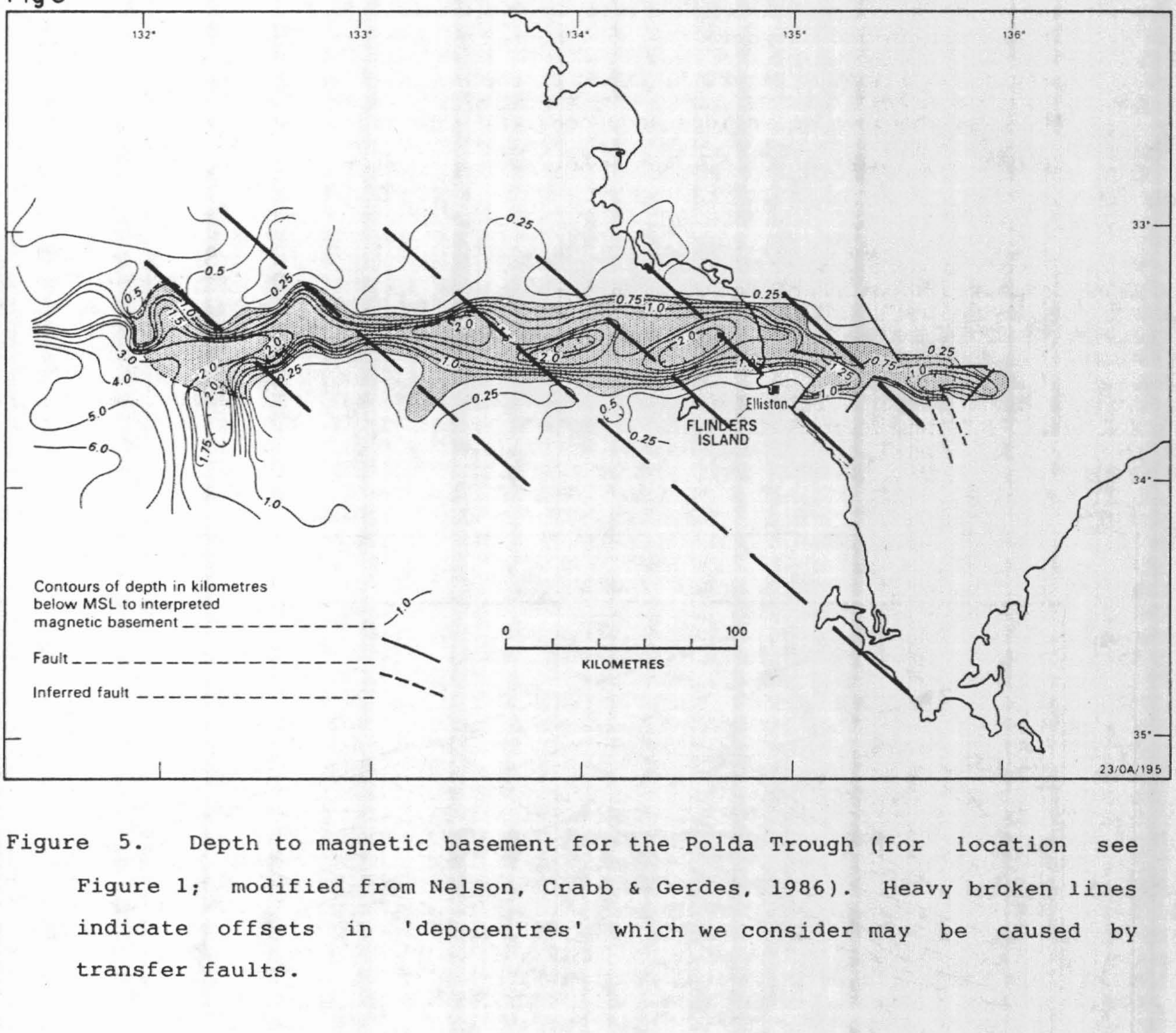

Figure 5. Depth to magnetic basement for the Polda Trough (for location see Figure 1; modified from Nelson, Crabb & Gerdes, 1986). Heavy broken lines indicate offsets in 'depocentres' which we consider may be caused by transfer faults.

Fig 6

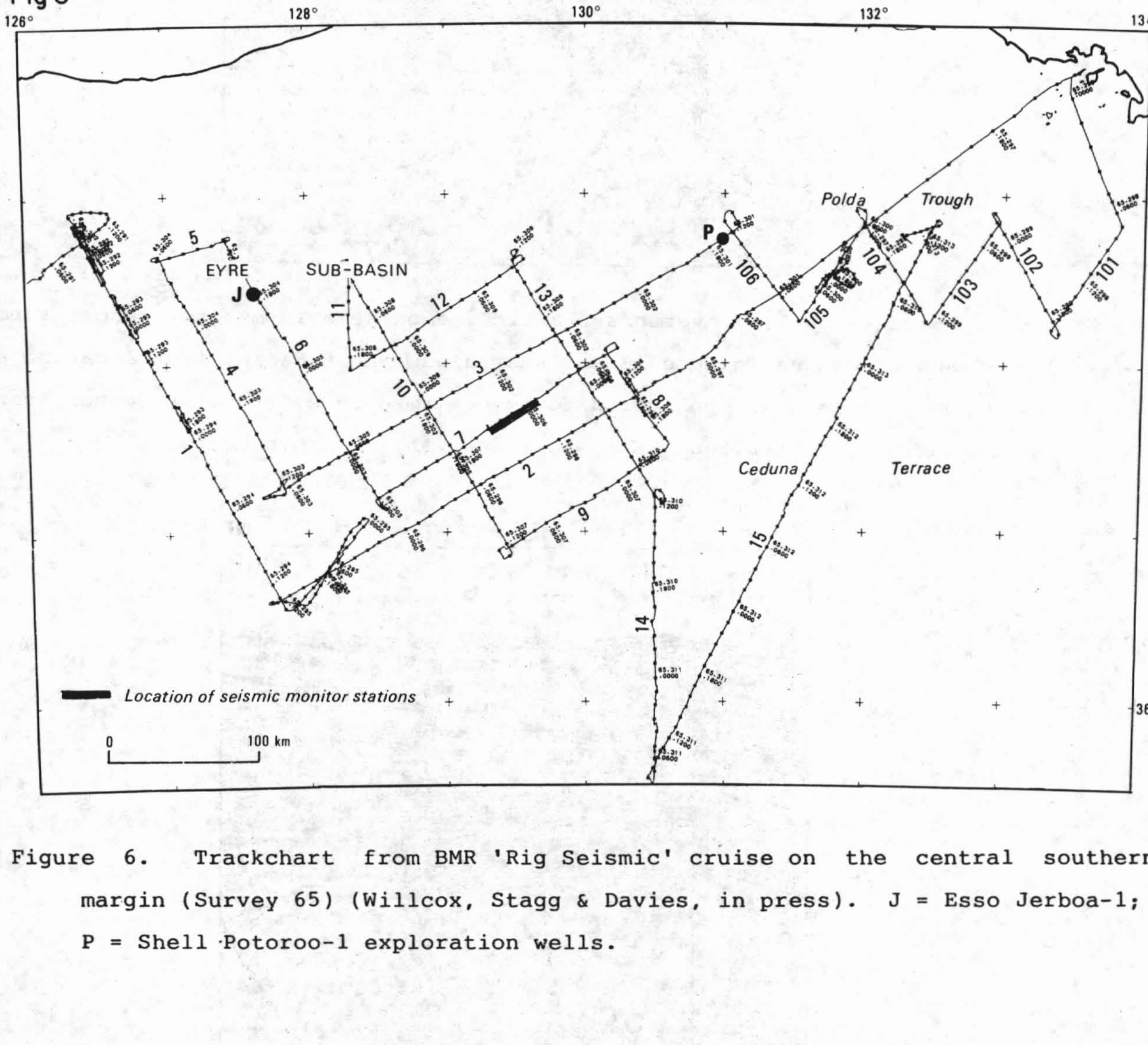

Figure 6. Trackchart from BMR 'Rig Seismic' cruise on the central southern margin (Survey 65) (Willcox, Stagg & Davies, in press). J = Esso Jerboa-1; P = Shell Potoroo-1 exploration wells.

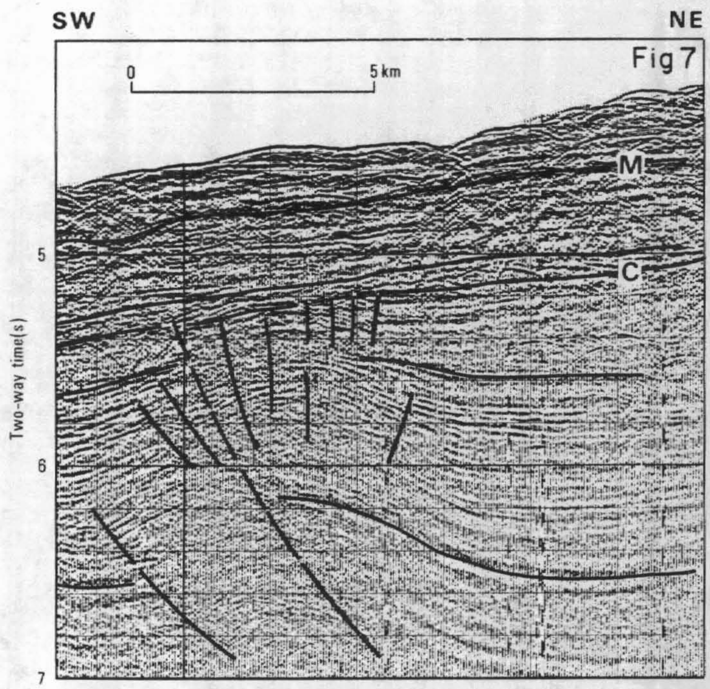

Figure 7. Faulted and (?) overthrust anticline on seismic monitor record across SW flank of Ceduna Terrace (Great Australian Bight Basin) (for location see Figure 6). This is considered to have formed in a 'transpressional zone' during crustal extension. M = Maastrichtian; C = Cenomanian.

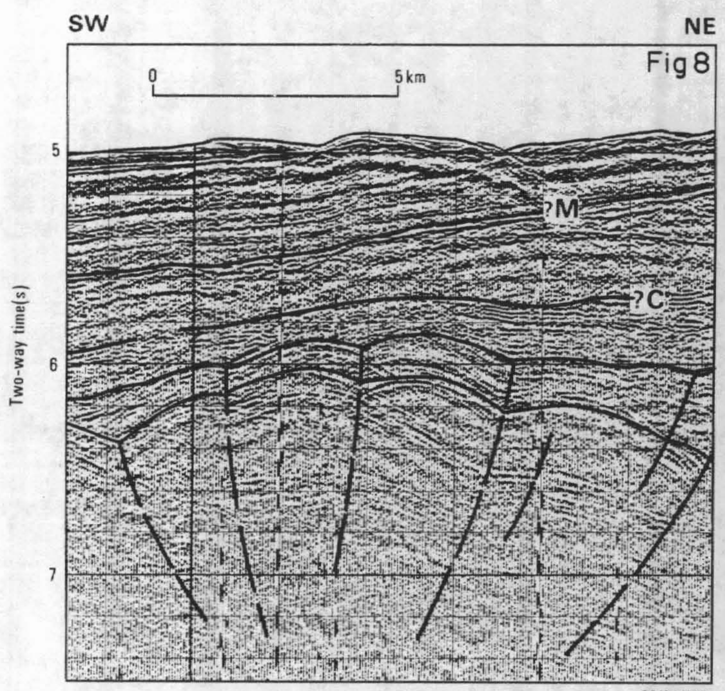

Figure 8. 'Flower structure' on seismic monitor record across SW boundary of Ceduna Terrace (for location see Figure 6). This is a typical wrench - related structure. M = Maastrichtian; C = Cenomanian.

Extensional Tectonics of the Offshore Otway Basin, Australia

P E Williamson, C D N Collins & M G Swift, BMR

The extensional tectonics of the offshore Otway Basin reflect the Cretaceous rifting history between Australia and Antarctica in the region. The rift trends reflecting that history and the extensional tectonics in the magnetic quiet zone close to the continent/ocean boundary were studied as part of the BMR Otway Basin project.

Two Lower Cretaceous rift phases have been interpreted for the Otway Basin: the first from approximately 140-120 MaBP culminating at Top Pretty Hill Formation time; the second from approximately 120-95 MaBP ending at Top Otway Group time. The first phase is retained as a 'fossil' rift system on the Crayfish platform in the west of the basin. The rift faults dip seawards and strike between east-northeast and east-west with associated south-southeast accommodation and transfer directions. The first phase rift directions are similar to those of the rift faulting of the southwestern Australian margin (Bein & Taylor, 1981) and are reflected in the coastline in that region. Their timing on the Crayfish Platform suggests that the initial rifting in the southeast was associated with the onset of seafloor spreading in the west at 120 MaBP (Cande & Mutter, 1982; Mutter & others, 1984). These first phase trends are in evidence over the central Southern Margin (Fraser & Tilbury, 1979; Whyte, 1978) and through Bass Strait as basement-involved faulting at the margins of the Bass Basin and minority trends within the Bass Basin (Williamson & others, 1987b) and as trends in the Gippsland Basin shown by Lowry (1987). This suggests that the initial rifting of the whole Australian Southern Margin could have been along the first phase trends.

The second phase of Lower Cretaceous rifting has a strike direction of 120° and an occasional transfer fault direction of 030° (Etheridge & others, 1987), the rarity of which may be the result of the use of reactivated first phase trends for along-strike geometrical accommodation during the second phase of rifting. Second phase faulting is the dominant faulting observed for the region and is associated with the final rift and drift of the offshore Otway Basin of the southeastern Australian margin. Post-Lower Cretaceous reactivation of these second phase faults has occurred, particularly in the west of the central Voluta Trough region of the Otway Basin (Williamson & others, 1987a).

BMR multichannel seismic reflection data have also been able to image the whole crustal section of the magnetic quiet zone of the offshore Otway Basin. This is due largely to the relatively thin post-rift sediment cover associated with this and other Australian starved margins. The quiet zone of the Australian Southern Margin has been discussed by Weissel & Hayes, (1972) and Mutter & others, (1984), and quiet zones associated with a number of continental margins by Rabinowitz, (1982). Seismic reflection, refraction and dredge data have, in the case of the offshore Otway Basin, allowed the extensional structures in the magnetic quiet zone to be imaged and interpreted. Features include:

1. An extensional duplex made-up of extended Lower Cretaceous crust separated by a detachment from extended basement and lower crust. The amounts of extension calculated for subplanar normal faulting in the upper and lower layers of the duplex are less than 40 percent and are broadly similar. It seems likely then that faulting and associated extension of the duplex was associated with Lower Cretaceous rifting, as is interpreted for similar faulting in the upper duplex landward of the quiet zone.

2. Shallowing seawards of the dips of extensional faults in both the top and bottom layers of the duplex imply greater extension towards the continent ocean boundary. The extensional faults are cut by later higher-angle normal faults which penetrate the whole crust and are interpreted as having occurred during subsequent thermal subsidence.

3. A total crustal thickness of 10 km comprising approximately 3 km of sediment and 7 km of basement and lower crust. The crustal thicknesses obtained from sonobuoy data and from results of Talwani & others, (1979) are consistent with the level of subsidence observed, calculated using the method of Hillinger & Sclater (1983). The crustal thicknesses cannot be arrived at via the observed faulting and require a previous episode of basement and lower crustal thinning. This probably involved simple shear (Wernicke, 1985) and/or penetrative ductile deformation (McKenzie, 1978). Parallel reflectors within the lower crust may indicate some component of penetrative ductile deformation prior to the observed faulting.

4. A possible thin veneer of upper plate basement retained over what would be classified as a lower plate lower crust (e.g., Lister & others, 1986). Dredging of ridges on the seaward side of the quiet zone has yielded basement sedimentary rocks of probable Palaeozoic age, and seismic

refraction velocities for basement and lower crust range from 5.6 to 7.0 km/sec, implying that both the top and bottom of the initial Lachlan Geosynclinal basement and lower crust (Finlayson & others, 1980) are represented.

5. An interpreted increase in the volume and incidence of igneous lithologies towards the continent/ocean boundary. The interpreted igneous features show high amplitudes and mounding, and diffract seismic energy. They are often located adjacent to fault planes.

6. Retention of up to 3 km of Lower Cretaceous Otway Group sediment even on the quiet zone at abyssal plain depth. These sediments appear to source oil on the Crayfish Platform of the Otway Basin (McKirdy & others, 1986) and its presence even in the quiet zone may indicate a deeper-water petroleum potential on the continental slope of the Otway Basin where burial is greater, faulting is less, and source maturity would have occurred later than on the abyssal plain, where the Otway Group outcrops and is heavily faulted.

References

Bein, J., & Taylor, M.L., 1981 - The Eyre Sub-basin : recent exploration results, APEA Journal. 21, 91-98.

Cande, S.C. & Mutter, J.C., 1982 - A revised identification of the oldest seafloor spreading anomalies between Australia and Antarctica. Earth and Planetary Science Letters. 58, 151-160.

Davidson, K., 1980 - Rotational displacements in southern Australia and their influences on hydrocarbon occurrence. Tectonophysics, 63, 139-153.

Etheridge, M.A., Branson, J.C. & Stuart-Smith, P.G., 1987 - The Bass, Gippsland and Otway Basins, southeast Australia : a branched rift system formed by continental extension. In Beaumont, C. & Tankard, A. (Editors) - Sedimentary basins and basin-forming mechanisms. Can. Soc. Petrol. Geol., Mem., 12.

Exon, N.F., Williamson, P.E., & others, 1987 - Rig Seismic Research Cruise 3 : offshore Otway Basin, southeastern Australia. Bureau of Mineral Resources, Australia Report 279.

Finlayson, D.M., Collins, C.D.N. & Denham, D., 1980 - Crustal structure under the Lachlan Fold Belt, southeastern Australia. Phys. Earth planet. Interiors, 21, 321-342.

Fraser, A.R. & Tilbury, L.A., 1979 - Structure and stratigraphy of the Ceduna Terrace region - Great Australian Bight. APEA Journal, 19(1), 53-65.

Hellinger, S.J., & Sclater, J.G., 1983 - Some comments on two-layer extension models for the evolution of sedimentary basins. J. Geophys. Res., 88, 8251-8270.

Lister, G.S., Etheridge, M.A. & Symonds, P.A. - Detachment models for the formation of passive continental margins. (Tectonics, in press).

Lowry, C.D., 1987 - A new play in the Gippsland Basin. APEA Journal, 27(1), 164-172.

McKenzie, D.P., 1978 - Some remarks on the development of sedimentary basins. Earth and Planetary Science Letters, 40, 25-32.

Mutter, C., Hegarty, K.A., Cande, C.S. & Weissel, J.K., 1984 -Breakup between Australia and Antarctica : a brief review in the light of new data. Tectonophysics, 114, 255-79

McKirdy, D.M., Cox, R.E., Volkman, K., & Howell, V.J., 1986 - Botryococcane in a new class of Australian non-marine crude oils. Nature, 320, 57-59.

Rabinowitz, D., 1982 - Gravity measurements bordering passive continental margins. In Scrutton, R.A. (Editor) - Dynamics of passive margins. Washington, D.C. American Geophysical Union, Boulder, Colorado - Geological Society of America.

Talwani, M., Mutter, J.C., Houtz, R. & Konig, M., 1979 - The crustal structure and evolution of the area underlying the magnetic quiet zone on the margin south of Australia. In Watkins J.S., Montadert, L., & Dickerson, P.W. (Editors) - Geological and geophysical investigations of continental margins. Am. Assoc. Petrol. Geol. Mem. 29, 151-75.

Weissel, J.K. & Hayes, D.E., 1971 - Asymmetric spreading south of Australia, Nature, 231 (5304), 517-521.

Wernicke, B., 1985 - Uniform sense normal simple shear of the continental lithosphere, Can. J. Earth Sci., 22, 108-125.

Whyte, R.K., 1978 - Shell's offshore venture in South Australia. APEA Journal, 18(1), 44-51.

Williamson, P.E., O'Brien, G.W., Swift, M.G., Felton, E.A., Scherl, A.S., Lock, J., Exon, N.F., Falvey, D.A. & Marlow, M., 1987a - Hydrocarbon potential of the offshore Otway Basin. APEA Journal, 27(1), 173-196.

Williamson, P.E., Pigram, C.J., Colwell, J.B., Scherl, A.S., Lockwood, K.L., & Branson, J.C., 1987b - Review of stratigraphy, structure and hydrocarbon potential of Bass Basin, Australia, American Association of Petroleum Geologists Bulletin, 71(3), 253-280.

Williamson, P.E., Collins, C.D.N., & Falvey, D.A. - Crustal structure in the magnetic quiet zone - Australian Southern Margin (in preparation).

Williamson, P.E., Swift, M.G., & O'Brien, G.W. - Two stage early Cretaceous rifting of the southeastern Australian Margin (in preparation).

Structural style of the Townsville Trough and its implications for the development of the northeast Australian margin

P A Symonds, BMR, D Capon, Santos Ltd., P J Davies, C J Pigram, & D A Feary, BMR

The Townsville Trough is an east-west - trending bathymetric depression within the northeast Australian continental margin, and separates the Queensland Plateau, to the north, and the Marion Plateau to the south (Fig. 1). It extends eastwards from the slope of the Great Barrier Reef and its junction with the north-northwesterly-trending Queensland Trough, to the complex area of troughs north of the Cato Trough. The Townsville Trough widens and deepens to the east with water depths at its centre ranging from 1100 to 2500 m. It is one of a number of troughs and marginal plateaux that comprise the northeast Australian margin (Fig. 3), and which are generally considered to be modified and subsided continental crust formed as a result of fragmentation of a northeastern extension of the Tasman Fold Belt (Gardner, 1970; Ewing & others, 1970; Falvey, 1972; Falvey & Taylor, 1974; Taylor, 1975; Mutter, 1977; Taylor & Falvey, 1977). This rift phase of development, which may have commenced in the Early Cretaceous, and was certainly in progress during the Late Cretaceous, preceded continental breakup and the formation of small enclosed ocean basins to the east (the Coral Sea Basin and Cato Trough) and to the south (the Tasman Basin). The Tasman sea-floor spreading commenced 80 MyBP - Campanian (Hayes & Ringis, 1973; Weissel & others, 1977; Shaw, 1978), and propagated northwards forming the Cato Trough and finally the Coral Sea Basin from 65 MyBP - Paleocene (Weissel & Watts, 1979). Sea-floor spreading ceased along the length of this system about 56 MyBP - early Eocene.

The western-most rift basin system consisting of the Bligh Trough, Osprey Embayment and Queensland Trough, and its possible southern extension into the Capricorn Basin (Figs 3 & 4), has a northwest trend which was thought to have been inherited from the Tasman Fold Belt (Mutter, 1977). However, the Townsville Trough, which joins with this rift system, has no clear relation to any known structure onshore, being roughly perpendicular to the main fold-belt trend, and its orientation has remained an enigma.

In 1985 BMR used its research vessel 'Rig Seismic' to collect 1300 km of 48-channel seismic reflection data along the Townsville Trough in a zig-zag pattern (Fig. 1) (Davies, Symonds & others, in press). The regional seismic line network that now exists over the trough has allowed a greatly improved understanding of its structural development. However, interpretation of the exact structural style, and even the trend of some major features, remains difficult because the regional coverage is inappropriate to the structural complexity of the region. A preliminary and somewhat schematic picture of the structural style of the Townsville Trough is shown in Figure 2.

The main structural element of the Townsville Trough is a central rift basin up to about 100 km in width. The rift basin appears to be split into two parts about a constriction at latitude 150°20'E. The eastern part has a general east-northeast trend and the western part has a west-northwest trend. The rift basin is flanked by 'basement' platforms which underlie the Queensland and Marion Plateau margins to the north and south, respectively. The platforms dip towards the rift basin and range in depth from about 1.7s to 3s seismic reflection two-way time below sealevel. The platforms themselves are disrupted by generally small-throw normal faults which tend to dip away from the rift basin creating small half-grabens. The platforms have been strongly eroded and towards the end of the seismic lines they have a planated appearance. The margins of the basement platform adjacent to the eastern part of the rift basin consist of northeast-trending (54°) and northwest-trending (315°) segments. The northern margin of the western part of the rift basin has a west-northwest trend, but there are indications that this trend is not primary and that it can be resolved into the northeast and northwest trends of the eastern basin, as indicated by the dots on Figure 2. This implies that the northern margin of the western basin is composed of a series of short segments and that basin architecture is more complex as it approaches the junction with the Queensland Trough. Both margins of the basement platform generally dip at a relatively low angle (about 25-35°) beneath the sediments of the rift basin. In some places this dip corresponds to the top of a tilt block (Fig. 5A); in other places it appears to represent a low-angle normal fault plane.

Structural trends within the rift basin are difficult to establish owing to the line spacing; however, the high-standing blocks in the eastern part of the rift basin appear to strike in a northeast direction, parallel to the most prominent basin-margin trend. Most are tilt blocks bounded by low-angle normal faults and half-grabens, although horst blocks are present in a few places. The low-angle faults are most commonly downthrown to the northwest. In contrast to the above, near the northern margin of the eastern part of the rift basin at about 150°25'E there are two tilt blocks which appear to have a northwesterly strike and to be bounded by faults downthrown to the southwest (Figs. 2 & 5A). The significance of these structures to basin development remains unknown at this stage; however, they may reflect increasing complexity in rift basin structure to the west, and in particular near the junction of the eastern and western parts of the basin. The tilt blocks are often associated with a syn-rift sediment package and are onlapped by late stage rift-fill sediments (Fig. 5A). The corners of some tilt blocks have been eroded and bevelled but to a much lesser extent than those of the Queensland Trough. Block corners are commonly associated with the development of gentle anticlines in the overlying post-Late Cretaceous sediment as a result of flexural drape, differential compaction and re-activation of the bounding faults (Figs. 5A & B). The processed BMR seismic data indicate that there are several areas within the trough where the sediments are more than 4s seismic reflection two-way time in thickness.

An important feature of the structure of the eastern part of the Townsville Trough rift basin is the presence of subtle northwest-trending transverse lineaments (Fig. 2), which align with the generally right-lateral offsets of the basin margin. These can be quite difficult to map in the centre of the basin owing to the spacing and orientation of the seismic grid. We have defined the transverse lineaments by terminations of basement highs and grabens, and along strike changes in the width, slope and character of these features, as well as changes in the direction of dip of the tops of blocks and their bounding faults. The transverse zones appear to correspond to structural complexities on the margins of the rift basin. One explanation of the transverse lineaments is that they are transfer faults, which essentially perform a similar function to oceanic transform faults within extended continental crust (Bally, 1981). The concept of transfer faults, which are accommodation structures that allow variations in the geometry of extension along the strike of the rift, has recently been applied to the North Sea by Gibbs (1984), to the Gippsland and Bass Basins by Etheridge & others (1984, 1985), and to passive margin evolution by Lister & others (1986; in press). The recognition of the northwest-trending transverse lineaments as transfer faults, and the associated northeast trend as the strike direction of low-angle normal faults implies that at least the eastern part of the Townsville Trough formed by northwest-southeast extension. For the western part of the trough to have a similar origin, its northern, apparently west-northwest - trending, margin would need to be resolved into northeast and northwest-trending components, as mentioned previously.

In summary, the preliminary structural picture that has emerged for the Townsville Trough, is one of high and low-angle normal faults trending at about 54°, and cross-cutting lineaments trending at about 315°, which have compartmentalised the eastern part of the rift basin (Fig. 2). The latter are probably transfer faults which make an angle of about 99° with the regional extension direction, and their presence suggests that at least the eastern Townsville Trough formed by slightly-oblique northwest-southeast extension. If this scenario is correct it has important implications for the development and structural style of adjacent parts of the margin. For example, if the Queensland Trough formed during the same episode of rifting/extension, it implies that the basin-forming structures beneath it should be the result of extreme oblique extension, and the trough will exhibit many of the features of transtensional basins. There is some evidence for this in the basement-fault pattern associated with the Queensland Trough (Symonds & others, 1984). Very-low-angle normal faults and highly-rotated tilt blocks occur in a number of areas in the Queensland Trough and on the adjacent Queensland Plateau (Fig. 5C) indicating that significant upper crustal extension has occurred in places.

The northwest-southeast - directed Cretaceous and ?older extensional event which appears to have produced the Townsville Trough is nearly perpendicular to the Paleocene-early Eocene opening direction of the Coral Sea Basin to the north, and the Cato Trough to the southeast. In fact the Townsville Trough runs parallel to the transform direction at the northern end of the Cato Trough. This could be taken to infer a transtensional strike-slip origin for the Townsville Trough; however, the structural style of the trough described above implies that the northwest-southeast extensional-stress-field which produced the Townsville and Queensland Troughs pre-dated, and was probably unrelated to, the stress field which finally resulted in break-up and seafloor spreading in the Coral Sea Basin.

Application of the detachment model of passive margin formation (Lister & others, 1986; in press) to this margin requires a complex branched or multiple detachment system. Upper crustal extension in the Townsville Trough and highly oblique extension in the Queensland Trough would have been accompanied by sub-detachment pure shear beneath both the troughs and the adjacent Queensland Plateau.

References

Bally, A.W., 1981 - Atlantic-type margins. In Geology of Passive Continental Margins, American Association of Petroleum Geologists, Education Course Note Series, 19, 1/1-48.

Davies, P.J., Symonds, P.A., & others, 1987 - Rig Seismic research cruise 4 and 5: northeast Australian margin. Bureau of Mineral Resources Australia, Report, 255.

Etheridge, M.A., Branson, J.C., Falvey, D.A., Lockwood, K.L., Stuart-Smith, P.G., & Scherl, A.S., 1984 - Basin-forming structures and their relevance to hydrocarbon exploration in Bass Basin, southeastern Australia. BMR Journal of Australian Geology & Geophysics, 9, 197-206.

Ewing, J.I., Houtz, R.E., & Ludwig, W.J., 1970 - Sediment distribution in the Coral Sea. Journal of Geophysical Research, 75, 1963-1972.

Falvey, D.A., 1972 - The nature and origin of marginal plateaux and adjacent ocean basins off Northern Australia. University of New South Wales (Ph.D. Thesis).

Falvey, D.A., & Taylor, L.W.H., 1974 - Queensland Plateau and Coral Sea Basin structural and time stratigraphic patterns. Australian Society of Exploration Geophysicists, Bulletin, 5, 123-126.

Gardner, J.V., 1970 - Submarine geology of the western Coral Sea. Geological Society of America, Bulletin, 70, 1399-1424.

Gibbs, A.D., 1984 - Structural evolution of extensional basin margins. Journal of the Geological Society, London, 141, 609-620.

Hayes, E.D., & Ringis, J., 1973 - Seafloor spreading in the Tasman Sea. Nature, 243, 454-458.

Lister, G.S., Etheridge, M.A., & Symonds, P.A., 1986 - Detachment faulting and the evolution of passive continental margins. Geology, 14, 246-250.

Mutter, J.C., 1977 - The Queensland Plateau. Bureau of Mineral Resources, Australia, Bulletin, 179.

Shaw, R.D., 1978 - Seafloor spreading in the Tasman Sea: a Lord Howe Rise - eastern Australia reconstruction. Bulletin of the Australian Society of Exploration Geophysicists, 9(3), 75-81.

Taylor, L.W.H., 1975 - Depositional and tectonic patterns in the western Coral Sea. Australian Society of Exploration Geophysicists Bulletin, 6(2/3), 13-29.

Taylor, L.W.H., & Falvey, D.A., 1977 - Queensland Plateau and Coral Sea Basin: stratigraphy, structure and tectonics. APEA Journal, 17(1), 13-29.

Weissel, J.K., & Hayes, D.E., 1977 - Evolution of the Tasman Sea reappraised. Earth and Planetary Science Letters, 36, 77-84.

Weissel, J.K., & Watts, A.B., 1979 - Tectonic evolution of the Coral Sea Basin. Journal of Geophysical Research, 84, 4572-4582.

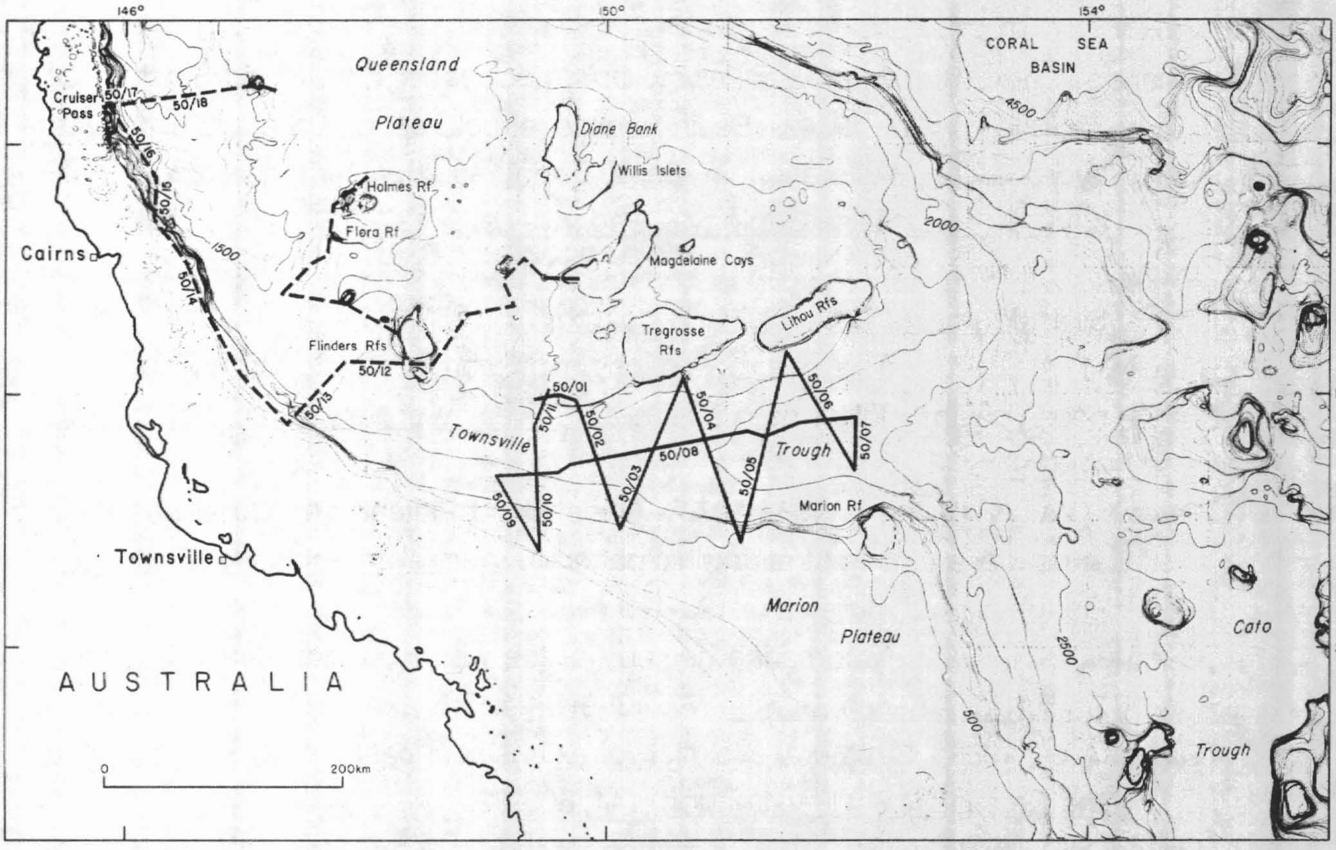

Figure 1. Bathymetry of the Townsville Trough region showing the location of the 1985 'Rig Seismic' data.

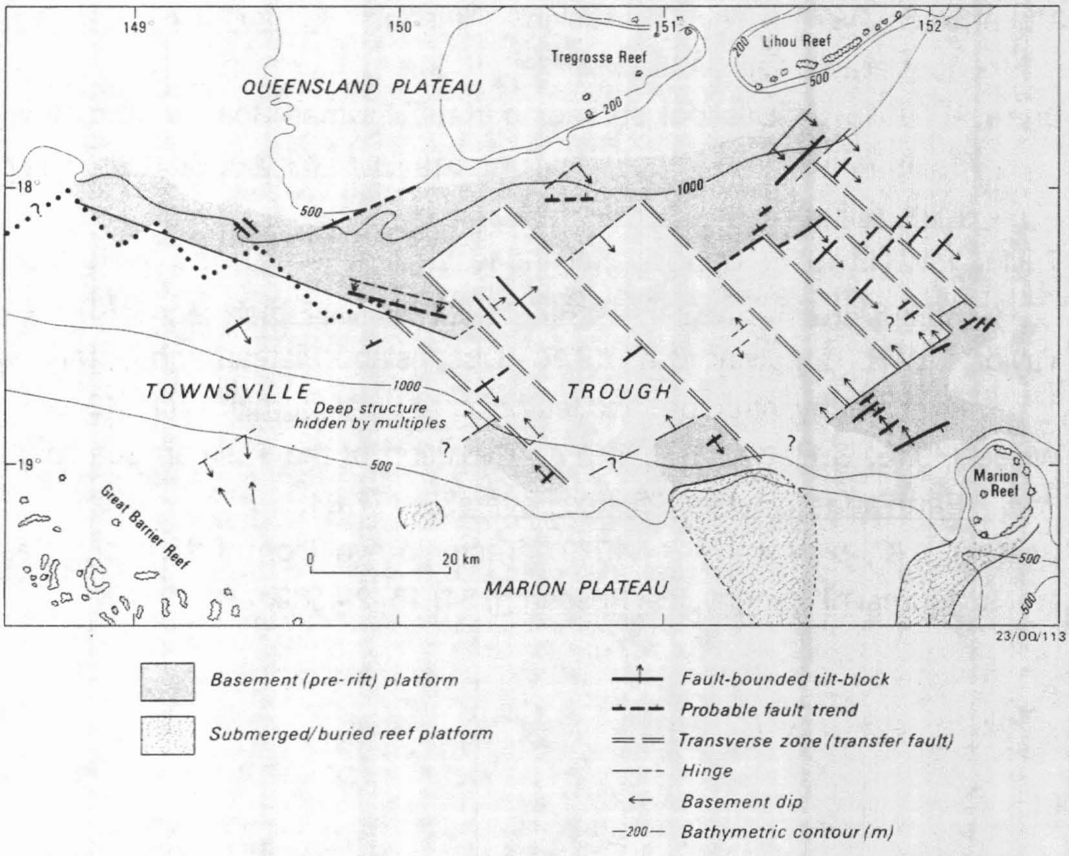

Figure 2. Preliminary schematic map of the basin-forming structures beneath the Townsville Trough (after Davies, Symonds & others, in press).

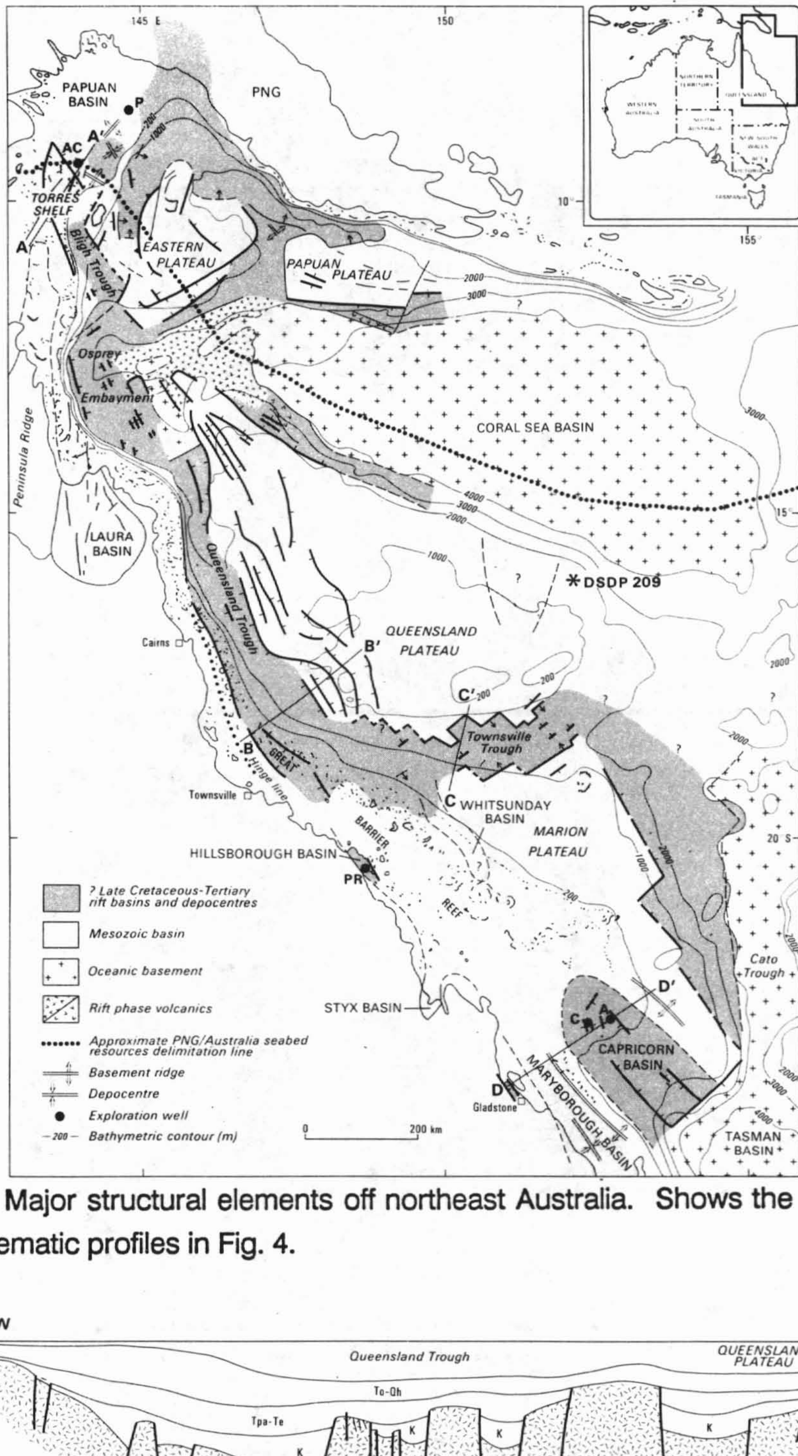

Figure 3. Major structural elements off northeast Australia. Shows the locations of the schematic profiles in Fig. 4.

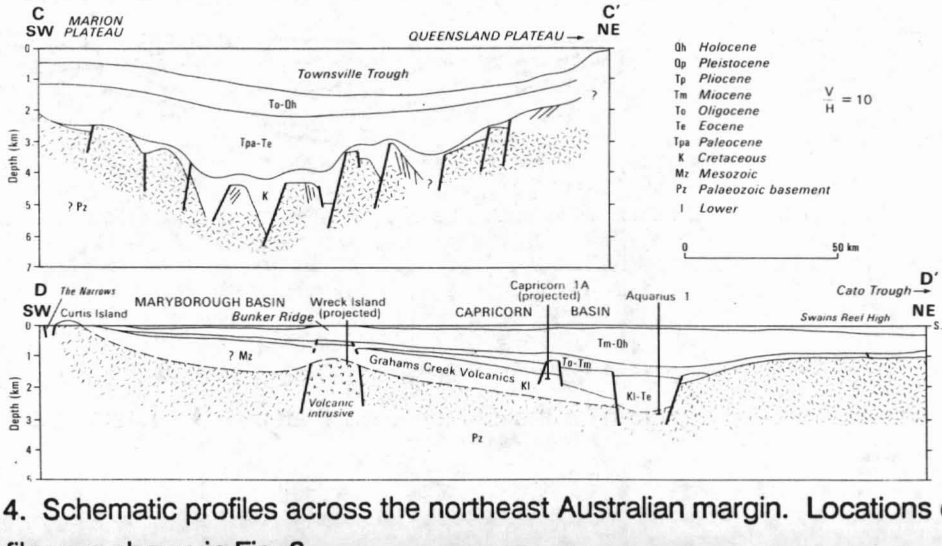

Figure 4. Schematic profiles across the northeast Australian margin. Locations of the profiles are shown in Fig. 3.

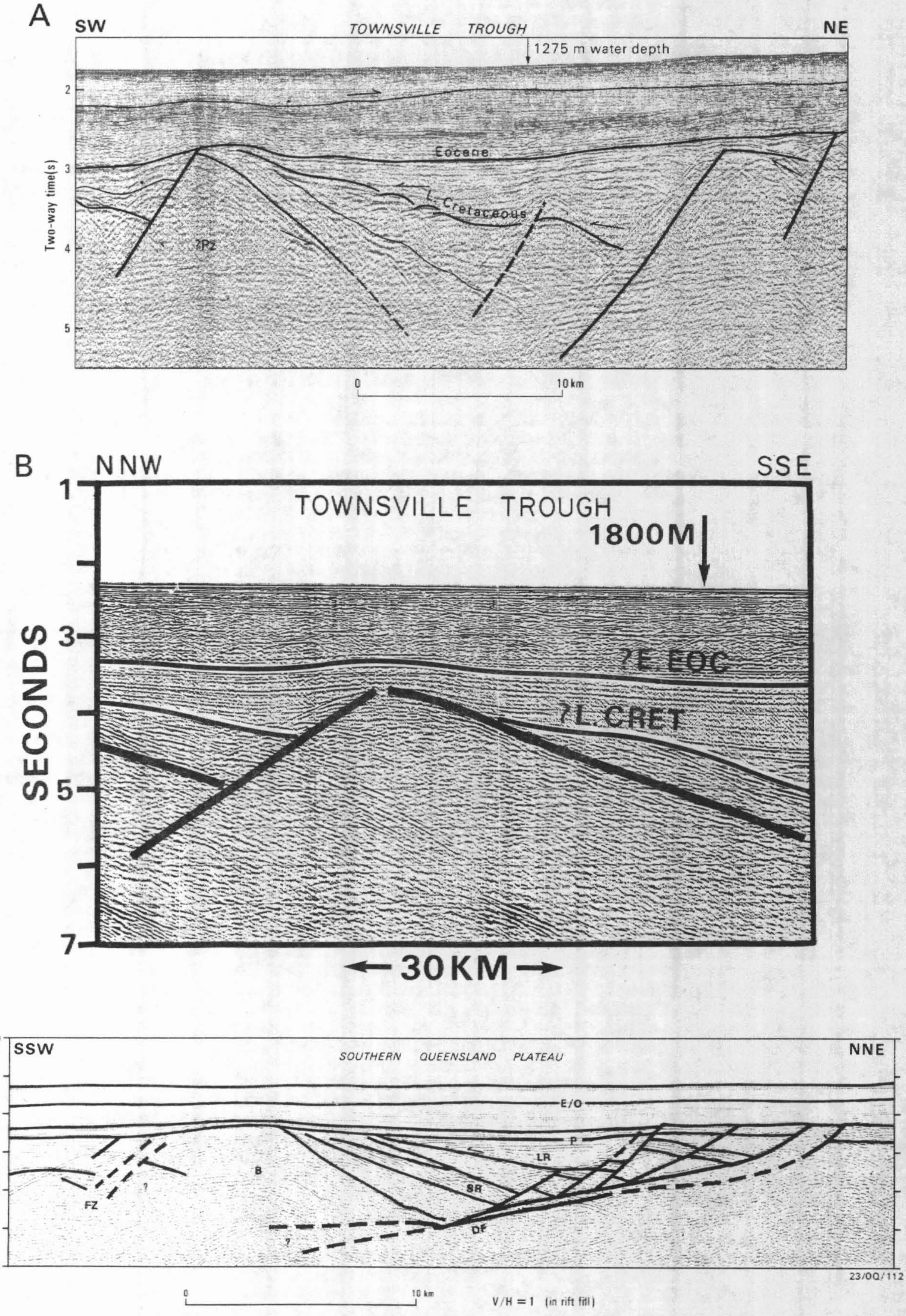

Figure 5. Portions of processed seismic data across the Townsville Trough (A and B) and the southern margin of the Queensland Plateau (C - after Lister & others, in press).

Tectonic framework of the north Perth Basin, Western Australia

J F Marshall & C-S Lee, BMR

The north Perth Basin is bounded by the Darling Fault to the east and the transform fault of the Wallaby Fracture Zone to the west (Fig. 1). Between these major tectonic features lies a series of ridges and troughs that collectively form the basin. The offshore portion consists of two major regions: the Edel Platform and the Abrolhos Sub-basin (Fig. 2). Three separate phases of rifting and extension are apparent in the offshore north Perth Basin, one in the Edel Platform and two in the Abrolhos Sub-basin. These three episodes of extension span the period from at least the Early Permian to the Early Cretaceous. However, it appears that relatively short periods of extension were punctuated by prolonged periods that were dominated by vertical tectonism, in the form of subsidence and steep normal faulting.

The earliest phase of rifting offshore occurs beneath the Edel Platform. Here, what have been interpreted as Early Permian or older sediments (Smith & Cowley, 1987) have been extended by a series of shallow-dipping rotational faults that dip landwards. Decollement on the faults is relatively shallow and at different levels. Extension increases to the east. There is relatively little dip-slip displacement or rotation on the westernmost faults which suggests that they could be oblique-slip faults. Synrift sediments of presumed Early Permian age form wedges within the half-grabens, which in turn are overlain by a conformable sequence of Late Permian and Triassic sediments. Drilling results from Edel-1 well suggest possible rift volcanism during the Early Permian.

In the Late Permian, the locus of rifting shifted from the Edel Platform to the Abrolhos Sub-basin. Here, the Early Permian sequence is displaced by low-angle, east-dipping extensional faults into a series of tilted blocks (Fig. 3). Synrift sediments within the half-grabens onlap the inclined tops of these blocks. The basal synrift sediments are considered to be equivalent to a marine facies of the Late Permian Wagina Sandstone, which in turn is overlain by the Early Triassic Kockatea Shale (Fig. 4). The attitude of these faults (i.e., easterly-dipping) would suggest that rifting had occurred along a roughly north-south axis landward of the present shelf. This is similar to the attitude of the rotational faults beneath the Edel Platform. However, the rifted blocks in the Abrolhos Sub-basin subside to the west and the sediments thicken in that direction also, suggesting that the major thermal anomaly occurred to the west of the rift axis.

During the Triassic and Jurassic, the Abrolhos Sub-basin underwent a phase of subsidence and normal faulting, whereas the Edel Platform remained relatively stable. High-angle faults, downthrown to the west by as much as one second (TWT), are present in the vicinity of the present shelf-break. This rapid phase of subsidence was accompanied by deposition of thick sequences of mainly continental sediments within the sub-basin (Fig. 4).

A third period of extension occurred prior to the onset of seafloor spreading during the Neocomian. The Late Jurassic sequence was extended by a series of rotational faults, dipping to the west (Fig. 3). Many of these faults show decollement onto the top of the Middle Jurassic Cadda Shale.

Seafloor spreading resulted in the development of the Wallaby Fracture Zone and numerous fracture zones within oceanic crust to the south (Veevers & others, 1985). These fracture zones resulted from oblique extension of the crust at the time of break-up. Within the margin, this oblique extension manifests itself in the form of re-activation of the tilted fault-blocks within the Abrolhos Sub-basin. This re-activation resulted in structures similar to wrench faults, with antithetic and "flower" structures that are quite complex in places. However, this style of faulting is believed to be related to strike-slip motion that is associated with oblique extension of the crust along the fracture zones. This horizontal displacement has also resulted in formation of anticlinal structures in many parts of the basin.

Examination of the fault pattern and the distribution of basement highs and lows within the north Perth Basin shows a strong similarity to regions that exhibit a substantial component of strike-slip motion and its associated oblique extension at the present time (e.g., the San Andreas Fault system). This suggests that for part of its existence the Darling Fault underwent more of an oblique-slip motion than the usually-assumed normal displacement. Similarly, the difference in structural styles between the Edel Platform and the Abrolhos Sub-basin suggest an oblique-slip boundary between these two provinces. This indicates that both oblique-slip and rotational normal faulting were operating simultaneously during rift phases within the north Perth Basin. It could be postulated also that the Darling Fault is the surface manifestation of the major detachment fault along the rifted margin of southwestern Australia.

References

Smith, G.C., & Cowley, R.G., 1987 - The tectono-stratigraphy and petroleum potential of the northern Abrolhos Sub-basin, Western Australia. APEA Journal, 112-136.

Veevers, J.J., Tayton, J.W., Johnson, B.D., & Hansen, L., 1985 - Magnetic expression of the continent-ocean boundary between the western margin of Australia and the eastern Indian Ocean. Journal of Geophysics, 56, 106-120.

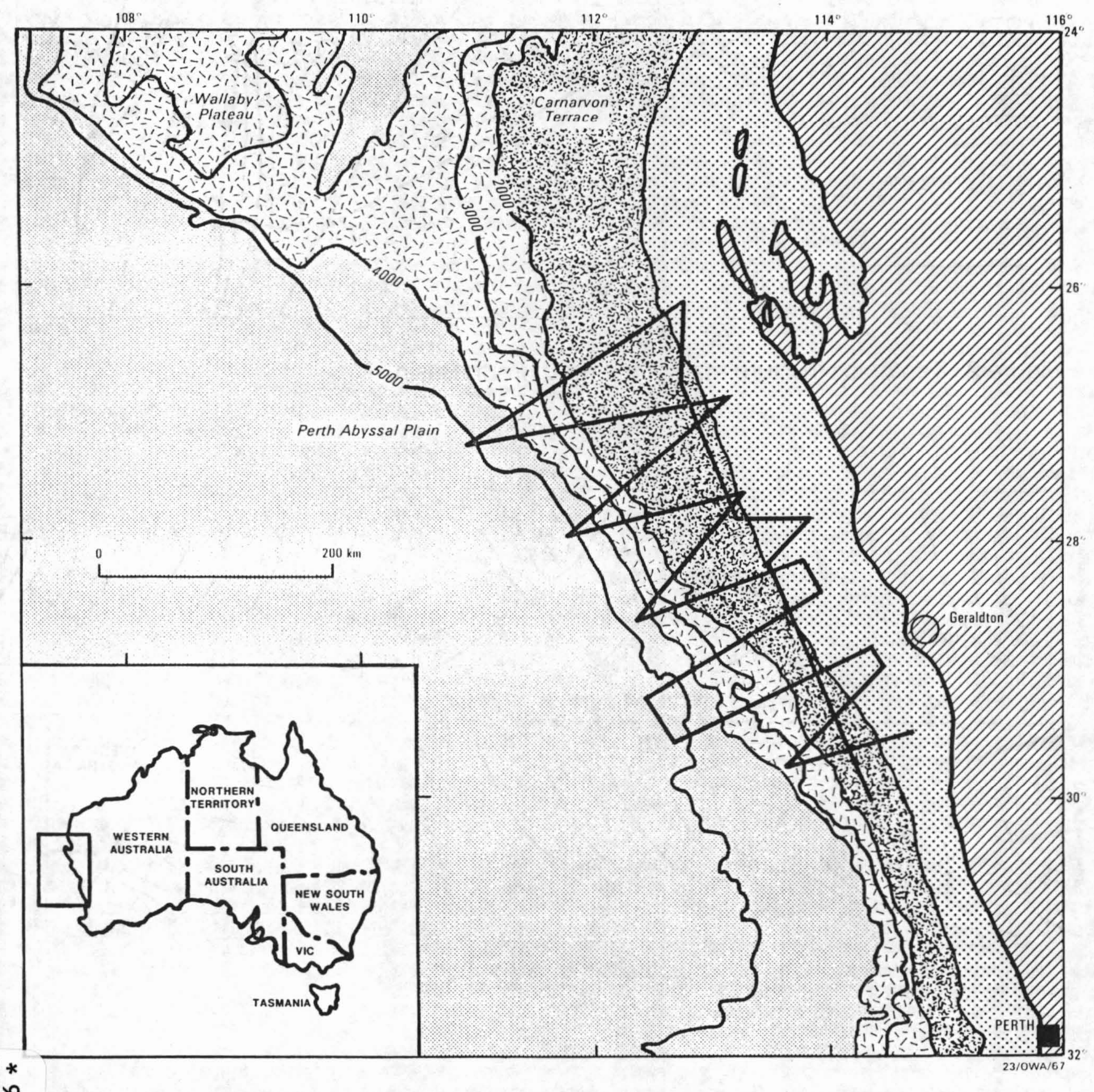

Figure 1. Bathymetry of north Perth Basin region showing the Wallaby Fracture Zone extending into the continental margin. Ship tracks are BMR seismic lines carried out in 1986.

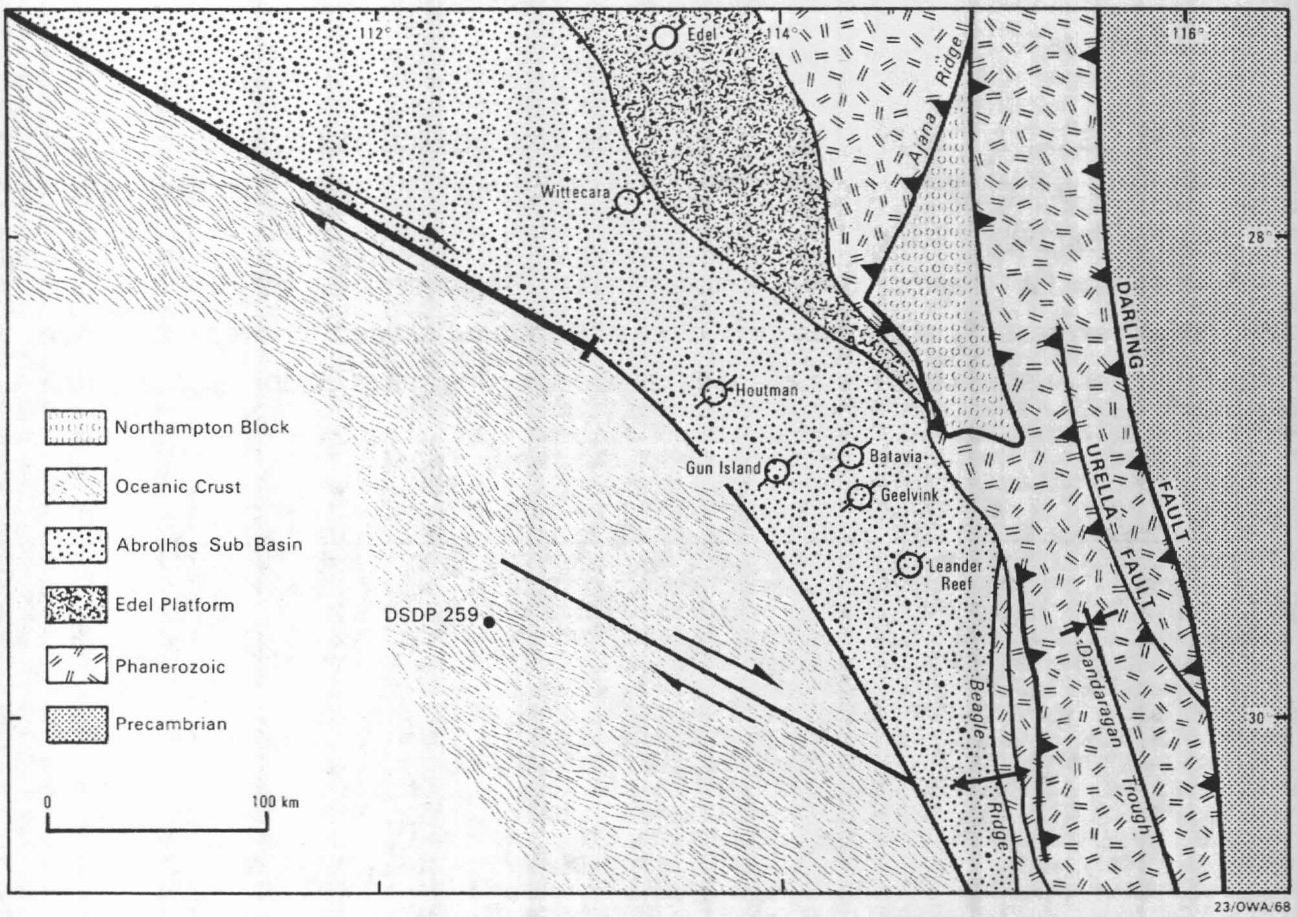

Figure 2. Tectonic structure map of the north Perth Basin.

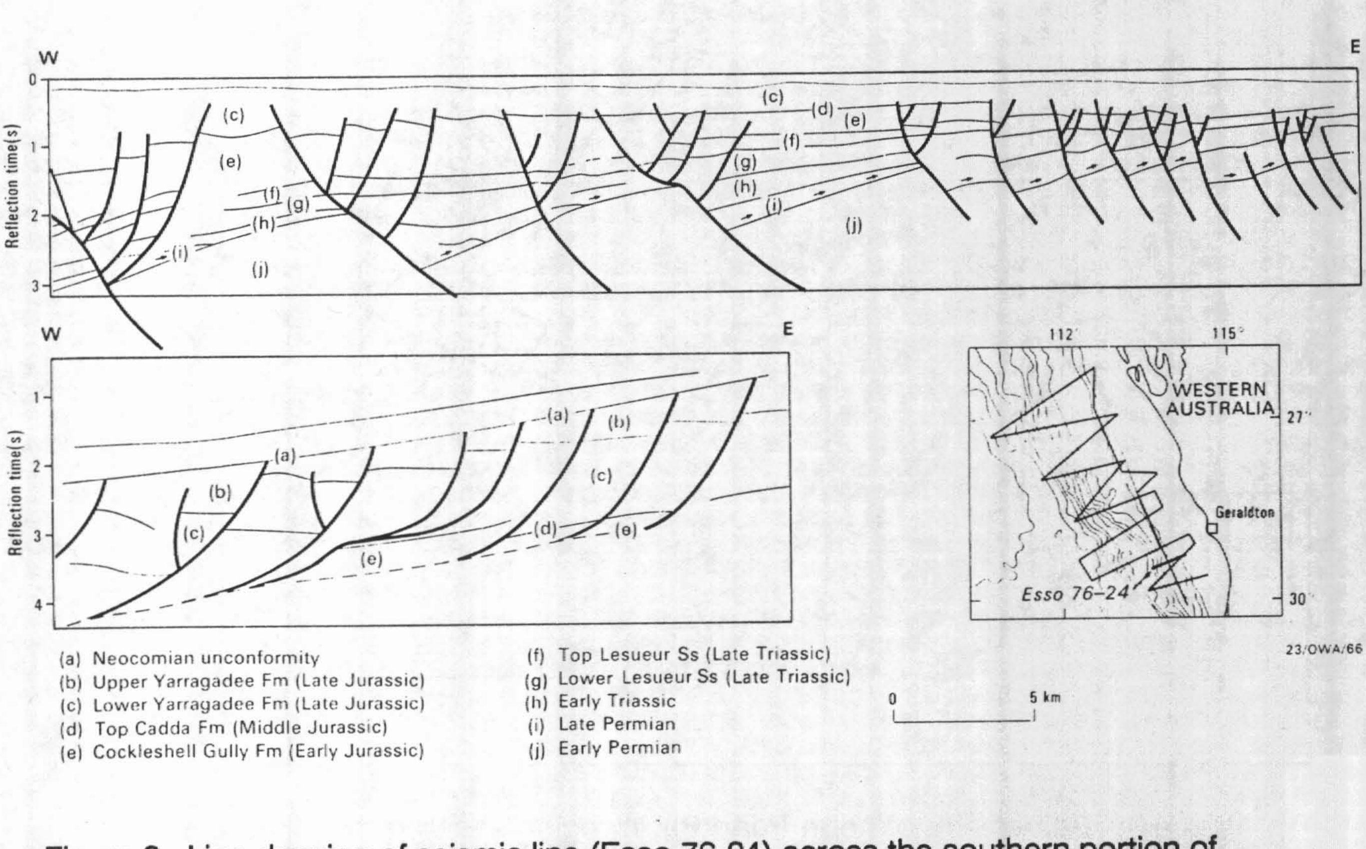

Figure 3. Line drawing of seismic line (Esso 76-24) across the southern portion of the Abrolhos Sub-basin.

PERIOD	EPOCH	FORMATION	LITHOLOGY	ENVIRONMENT
QUATERNARY	Pleistocene	Coastal Limestone		Marine
TERTIARY	Paleocene	Kings Park Shale		Marine
CRETACEOUS	Late	Gingin Chalk		Marine
		Osborne Formation		Marine
	Early	Warnbro Group		Marine to Continental
JURASSIC	Late	Yarragadee Formation		Continental
	Middle	Cadda Formation		Marine
	Early	Cockleshell Gully Formation		Marginal Marine to Continental
TRIASSIC	Late	Lesueur Sandstone		
	Middle	Woodada Formation		Marine
	Early	Kockatea Shale		Marine
PERMIAN	Late	Wagina Sandstone		Marine
	Early	Caryngina Formation		Mainly Marine
		Irwin Coal Measures		
		Holmwood Shale		Marine to Continental
		Nangetty Formation		
SILURIAN		Tumblagooda Sandstone		Continental
PRECAMBRIAN		"Basement"		

23/OWA/69

Figure 4. Stratigraphy of the north Perth Basin.

Igneous history related to extension in the Basin and Range Province, soutwestern USA

S J Reynolds, Arizona Geological Survey

(See earlier abstract on 'Continental extension in the Basin and Range Province, USA : structural and igneous aspects')

Extensional marginal basins and related magmatism of the southwest Pacific

R W Johnson, BMR

Evidence for extensional tectonics is widespread in and near the zone of convergence between the major Pacific and Indo-Australian plates in the southwest Pacific. Understanding this apparent paradox of extension in a region of convergence has been the main challenge in studies of the 15 or more marginal basins in this region (Fig. 1, Table 1). The nature of some of the basins is now known in some detail, but the origin of others - particularly the older ones - remains poorly understood. In general, however, the basins are known to be floored by oceanic crust similar to that in the major oceans (Karig, 1971; Packham & Falvey, 1971). Many of them are younger than the crust of the adjacent southwest-Pacific ocean, and their seafloor volcanic rocks are the same as, or rather similar to, mid-ocean-ridge basalts (MORB).

Marginal basins in the southwest Pacific that are the focus of particular volcanological, geochemical, and petrological interest in BMR are the active Woodlark and Manus Basins in western Melanesia (which are probably collision related), and the late-Mesozoic to early-Cainozoic basins to the east of Australia (see below).

Models of marginal-basin formation

Development of many marginal basins has been attributed to mantle upwelling and extension behind island arcs caused by either (1) subduction-related diapirism (Karig, 1971; Fig. 2A), or (2) secondary mantle convection induced by the down-going slab (Sleep & Toksoz, 1971; Fig. 2B). Crustal extension leads to splitting of the arc producing an inactive 'remnant arc' separated from the active arc (nearer the trench) by a 'back-arc' or 'inter-arc' basin. Difficulties with both of these models include: (1) the generally short-lived (10-20 Ma) nature of basin formation and the apparently multistage origin of some basins (why does back- arc extension cease though subduction continues?);(2) the absence of back-arc basins at many arc-trench systems; (3) removal during extension of the spreading axes from the proposed sites of diapirism above down-going slabs; (4) the predicted sites of extension by slab-induced convection are at least 100 km back from the junction between the subducting slab and the base of the overlying lithosphere where arc-rifting takes place; (5) some active marginal basins, such as the Woodlark Basin (see below), represent propagation of oceanic ridges into continents and are clearly not of back-arc origin.

Other investigations have been directed at evaluating marginal-basin development in terms of global interactions of the major plates, the model favoured by Packham & Falvey (1971; Fig. 2C). Plate-kinematic concepts reported in the literature include: (1) the seaward migration, or 'rollback' (Dewey, 1980), of trenches and the 'landward' migration of the overriding plate in an absolute reference frame, leading to behind-the-arc extension (Chase, 1978); (2) subduction of old, cold, lithosphere (western Pacific) causes rollback and behind-the-arc extension, whereas subduction of young, hot lithosphere (eastern Pacific) does not (Molnar & Atwater, 1978); (3) the state of stress in the overriding plate determines whether a marginal basin forms by extension in arc-trench systems of the 'Mariana type', or whether no basin forms in systems of the 'Chilean type' because the overriding plate is under compression (Uyeda & Kanamori, 1979); (4) changes in the direction of Pacific-plate motion control marginal-basin development in the western Pacific (Jurdy, 1979); (5) basins develop where both the overriding plate retreats from the trench in an absolute sense and the subducted plate undergoes a significant speed-up causing a stress 'surge' (Fein & Jurdy, 1986). Difficulties with the global plate-kinematic approach were listed by Taylor & Karner (1983) who included the problem of whether lithosphere can transmit forces from the down-going slab to the outer rise of the trench where rollback is thought to take place.

Some marginal basins appear to represent pieces of old ocean crust abandoned behind island arcs when subduction was initiated within an oceanic plate (the Bering Sea behind the Aleutian island arc is one reported example). Trapped basins of this type have not been recognised unequivocally in the southwest Pacific, but could be represented by the D'Entrecasteaux and North Loyalty basins.

Marginal-basin magmatism

Concepts of magma genesis in the marginal basins of the southwest Pacific run hand-in-hand with studies of the ridge systems of the major oceans where much more research has been undertaken. The basic petrological model for MORB generation by partial melting of a depleted peridotite mantle, remains viable for both ocean ridges and marginal basins, but mantle-source regions have compositional heterogeneities. The controversy over the composition and depth of generation of primary-MORB compositions continues to be debated. One claim is that MORBs containing 9-10 per cent MgO represent primary magma (e.g. Presnall & Hoover, 1984), whereas another is that the most primitive MORBs are derived by fractionation of picritic magmas (e.g. Green & others, 1979).

MORB sources affected by a subduction-related component are inferred for marginal-basin basalts characterised by, for example, enrichments in elements such as K, Rb, Sr, and Ba, by slightly different isotopic signatures, and by higher volatile contents and Fe^{3+}/Fe^{2+} values compared to normal (N-type) MORB (e.g. Hawkins & Melchior, 1985). However, the search for a distinctive 'back-arc-basin type' basalt remains elusive as these rocks in general differ only slightly (in different degrees) from N-MORB.

Regional, along-axis differences in chemistry from N-MORB towards ocean-island-type basalts have been identified on mid-ocean ridges. In addition, but at a more local scale, compositional differences have been found to be related closely to the segmented structure of ocean ridges. Some differences relate to termination of ridges at transform faults, to overlapping spreading centres, to propagating rift tips, and to axial bathymetric highs (e.g. Perfit & Fornari, 1983; Sinton & others, 1983; Thompson & others, 1985). Others, determined by closely spaced sampling, are found at small kinks, bends, and offsets on ridges (e.g. the 'devals' of Langmuir & others, 1986).

Hydrothermal processes leading to polymetallic-sulphide deposition on the ridge crests of marginal basins and mid-ocean rifts result as a consequence of magmatic processes that energise sub-seafloor hydrothermal convection, and tectonic processes which contribute to the permeability that allows convection. Exhalative sulphide deposits have been recorded for both the Lau (Stackelberg & others, 1985) and Manus (Both & others, 1986) Basins. The top of a flat-lying, 2-3 km-wide, magma chamber 3.5 km below the seafloor has been inferred to exist beneath the Valu Fa Ridge, in the Lau Basin, from multi-channel seismic reflection lines (Morton & Sleep, 1985).

Specific examples

Over 40,000 km^2 of Manus Basin seafloor were mapped in 1986 using the SeaMARC II side-scan sonar and bathymetry system, revealing some unexpected structures (Taylor & others, 1986). Three en-echelon sigmoidal spreading axes are recognised which link with a rhomb-shaped microplate bounded by transform faults, rift grabens, and a curving propagating spreading axis or sphenochasm. A newly-identified type of plate boundary consisting of a series of en-echelon spreading axes cut by a shear zone, is termed an 'extensional transform zone' (Taylor & others, 1986). Volcanic rocks dredged from 36 sites in the Manus Basin may be divided into (1) a low-K suite, from MORB to ferrobasalt and dacite, and (2) a basalt-dacite suite that appears to be of arc affinity even though sampled clearly from a marginal-basin setting. Tectonic relationships between the Manus Basin and New Britain island arc to the south remain obscure.

The N-MORB-generating Woodlark Basin runs at right-angles to the active Solomon island arc and is clearly not a back-arc basin. Its western apex is propagating into the continental crust of eastern Papua where there are late-Cainozoic volcanoes of both arc and continental-rift types. The eastern end of the basin is an unusual, if not unique, example of present-day subduction of an active ridge almost orthogonally beneath an island arc. This has produced unusual petrological effects, including arc-type volcanism close to, and in front of, the subduction zone (Johnson & others, 1987).

Intraplate volcanism of late-Mesozoic to Quaternary age extends down the 'highlands' of eastern Australia, along the western edge of the Coral Sea and Tasman Sea Basins. Felsic volcanoes make up a well-defined hot-spot trace, younging to the south, but the remaining, dominantly mafic, 'lava-field' volcanism has no obvious time-transgressive spatial pattern. A major effort is being made to establish what relationship both the hot-spot and lava-field volcanism have to highlands uplift and to the contemporaneous development of the basins to the east (Coral Sea, D'Entrecasteaux, North Loyalty, New Caledonia, Tasman Sea), in particular the Tasman Sea basin (see Wellman, 1987, for a review of highlands-uplift models). Results are to be published in the bicentenary year (Johnson & Taylor, in preparation).

References

Both, R., Crook, K., Taylor, B., Brogan, S., Chappell, B., Frankel, E., Liu, L., Sinton, J., & Tiffin, D., 1986 - Hydrothermal chimneys and associated fauna in the Manus back-arc basin, Papua New Guinea. Transactions of the American Geophysical Union, 67, 489-490.

Chase, C.G., 1971 - Tectonic history of the Fiji Plateau. Geological Society of America Bulletin, 82, 3087-3110.

Chase C.G., 1978 - Extension behind island arcs and motions relative to hot spots. Journal of Geophysical Research, 83, 5385-5387.

Connelly, J.B., 1976 - Tectonic development of the Bismarck Sea based on gravity and magnetic modelling. Geophysical Journal of the Royal Astronomical Society, 46, 23-40.

Davies, H.L., Symonds, P.A., & Ripper, I.D., 1984 - Structure and evolution of the southern Solomon Sea region. BMR Journal of Australian Geology and Geophysics, 9, 49-68.

Dewey, J.F., 1980 - Episodicity, sequence, and style at convergent plate boundaries. In Strangway, D.W., (Editor) - The continental crust and its mineral deposits. Geological Association of Canada Special Paper 20, 553-573.

Dubois, J., Dugas, F., Lapouille, A., & Louat, R., 1978 - The troughs at the rear of the New Hebrides island arc: possible mechanisms of formation. Canadian Journal of Earth Sciences, 15, 351-360.

Fein, J.B., & Jurdy, D.M., 1986 - Plate motion controls on back-arc spreading. Geophysical Research Letters, 13, 456-459.

Green, D.H., Hibberson, W.O., & Jaques, A.L., 1979 - Petrogenesis of mid-ocean ridge basalts. In McElhinny, M.W., (Editor) - The Earth: its origin, structure and evolution. Academic Press, London, 265-299.

Hawkins, J.W., & Melchior, J.T., 1985 - Petrology of Mariana and Lau Basin basalts. Journal of Geophysical Research, 90, 11,431-468.

Hayes, D.E., & Ringis, J., 1973 - Sea-floor spreading in the Tasman Sea. Nature, 243, 454-458.

Johnson, R.W., Jaques, A.L., Langmuir, C.H., Perfit, M.R., Staudigel, H., Dunkley, P.N., Chappell, B.W., Taylor, S.R., & Baekisapa, M., 1987 - Ridge subduction and forearc volcanism: petrology and geochemistry of rocks dredged from the western Solomon arc and Woodlark Basin. In Taylor, B., & Exon, N.F., (Editors) - Marine Geology and Geophysics of the Woodlark Basin-Solomon Islands region. Circum-Pacific Council for Energy and Mineral Resources Earth Science Series (in press).

Johnson, R.W., & Taylor, S.R., (Editors) - Intraplate volcanism in eastern Australia and New Zealand. Australian Academy of Science, Canberra, (in preparation).

Jurdy, D.M., 1979 - Relative plate motion and the formation of marginal basins. Journal of Geophysical Research, 84, 6796-6802.

Jurdy, D.M., & Stephanick, M., 1983 - Flow models for back-arc spreading. Tectonophysics, 99, 191-206.

Karig, D.E., 1971 - Origin and development of marginal basins in the western Pacific. Journal of Geophysical Research, 76, 2542-2561.

Karig, D.E., & Mammerickx, J., 1972 - Tectonic framework of the New Hebrides island arc. Marine Geology, 12, 187-205.

Kroenke, L.W., & Rodda, P., 1984 - Cenozoic development of the southwest Pacific. UNESCO Committee for Co-ordination of Joint Prospecting for Mineral Resources in South Pacific Offshore Areas Technical Bulletin, 6, 122 pp.

Langmuir, C.H., Bender, J.F., & Batiza, R., 1986 - Petrological and tectonic segmentation of the East Pacific Rise, 5°30'-14°30'N. Nature, 322, 422-429.

LaBrecque, J.L., Cande, S.C., & Dent, D.V., 1977 - Revised magnetic polarity time scale for the Late Cretaceous and Cenozoic time. Geology, 5, 330-335.

Lapouille, A., 1982 - Etude de bassins margineaux fossiles du sud-ouest Pacifique: bassin Nord-d'entrecasteaux, bassin Nord-Loyaute, bassin Sud-Fidjien. In Equipe de Geologie-Geophysique du Centre ORSTOM de Noumea. Contribution a l'Etude Geodynamique du Sud-Ouest Pacifique. Travaux et Documents de L' ORSTOM, 147, 409-438.

Lawver, L.A., Hawkins, J.A., & Slater, J.G., 1976 - Magnetic anomalies and crustal dilation in the Lau Basin. Earth and Planetary Science Letters, 33, 27-35.

Luyendyk, B.P., Bryan, W.B., & Jezek, P.A., 1974 - Shallow structure of the New Hebrides island arc. Geological Society of America Bulletin, 85, 1287-1400.

Malahoff, A., Feden, R., & Fleming, H., 1982 - Magnetic anomalies and the tectonic fabric of marginal basins north of New Zealand. Journal of Geophysical Research, 87, 4109-4125.

Molnar, P., & Atwater, T., 1978 - Inter-arc spreading and cordilleran tectonics as alternates related to the age of subducted oceanic lithosphere. Earth and Planetary Science Letters, 41, 330-340.

Morton, J.L., & Sleep, N.H., 1985 - Seismic reflections from a Lau Basin magma chamber. In Scholl, D.W., & Vallier T.L., (Compilers & Editors) - Geology and offshore resources of Pacific island arcs - Tonga region. Circum-Pacific Council for Energy and Mineral Resources Earth Science Series, 2, 441-453.

Packham, G.H., & Falvey, D.A., 1971 - An hypothesis for the formation of the marginal seas in the western Pacific. Tectonophysics, 11, 79-109.

Perfit, M.R., & Fornari, D.J., 1983 - Geochemical studies of abyssal lavas recovered by DSRV ALVIN from eastern Galapagos Rift, Inca Transform, and Ecuador Rift 1. Major element variations in natural glasses and spacial distribution of lavas. Journal of Geophysical Research, 88, 10,519-10,529.

Presnall, D.C., & Hoover, J.D., 1984 - Composition and depth of origin of primary mid-ocean ridge basalts. Contributions to Mineralogy and Petrology, 87, 170-178.

Sinton, J.M., Wilson, D.S., Christie, D.M., Hey, R.N., & Delaney, J.R., 1983 - Petrologic consequences of rift propagation on oceanic spreading ridges. Earth and Planetary Science Letters, 62, 193-207.

Sleep, N.H., & Toksoz, M.N., 1971 - Evolution of marginal basins. Nature, 233, 548-550.

Stackelberg, U., & Shipboard Scientific Party, 1985 - Hydrothermal sulfide deposits in back-arc spreading centers in the southwest Pacific. Bundesanstalt fur Geowissenschaften und Rohstoffe Circular, 2, 3-14.

Taylor, B., & Exon, N.F., (Editors), 1987 - Marine geology and geophysics of the Woodlark Basin - Solomon Islands region. Circum-Pacific Council for Energy and Mineral Resources Earth Sciences Series (in press).

Taylor, B., & Karner, G.D., 1983 - On the evolution of marginal basins. Reviews of Geophysics and Space Physics, 21, 1727-1741.

Taylor, B., Sinton, J., Shipboard Party, & Crook, K., 1986 - Extensional transform zone, sulphide chimneys and gastropod vent fauna in the Manus back-arc basin. Transactions of the American Geophysical Union, 67, 377.

Thompson, G., Bryan, W.B., Ballard, R., Hamuro, K., & Melson, W.G., 1985 - Axial processes along a segment of the East Pacific Rise, 10°N to 12°N. Nature, 318, 429-433.

Uyeda, S., & Kanamori, H., 1979 - Back-arc opening and the mode of subduction. Journal of Geophysical Research, 84, 1049-1061.

Watts, A.B., Weissel, J.K., & Larson, R.L., 1977 - Sea-floor spreading in marginal basins of the western Pacific. Tectonophysics, 37, 167-181.

Weissel, J.K., 1977 - Evolution of the Lau Basin by the growth of small plates. In Talwani, M., & Pitman W.C., (Editors) - Island arcs, deep sea trenches, and back-arc basins. American Geophysical Union, Washington D.C., Maurice Ewing Series, 1, 429-436.

Weissel, J.K., & Hayes, D.E., 1977 - Evolution of the Tasman Sea re-appraised. Earth and Planetary Science Letters, 36, 77-84.

Weissel, J.K., Taylor, B., & Karner, G.D., 1982a - The opening of the Woodlark Basin, subduction of the Woodlark spreading system, and the evolution of northern Melanesia since mid-Pliocene time. Tectonophysics, 87, 253-277.

Weissel, J.K., and Watts, A.B., 1979 - Tectonic evolution of the Coral Sea Basin. Journal of Geophysical Research, 84, 4572-4582.

Weissel, J.K., Watts, A.B., & Lapouille, A., 1982b - Evidence for Late Eocene seafloor in the southern New Hebrides Basin. Tectonophysics, 87,243-251.

Wellman, P., & McCracken H.M., (compilers), 1979 - Plate tectonics, 1:20 000 000-scale map. Bureau of Mineral Resources, Australia - BMR Earth Science Atlas of Australia.

Wellman, P., 1987 - Eastern Highlands of Australia: their uplift and erosion. BMR Journal of Australian Geology and Geophysics, 10, 277-286.

Table 1. Marginal basins of the southwest Pacific

Basin	Age	Comment	Selected References
New Guinea	nk	Sediment-covered, western part of Bismarck Sea. Nature unknown.	Connelly (1986)
Manus	0-nk	Active seafloor spreading and rifting, but date of inception unknown. Possible propagating rift.	Taylor & others. (1986)
Solomon Sea	Possibly early Tertiary	Being consumed by subduction on at least two sides.	Davies & others. (1984)
Woodlark	0-5	Propagating into Papua to west, and being subducted in east. Not of back-arc origin.	Weissel & others (1982a), Taylor & Exon (1987)
Coral Sea	56-64	Originated by rifting of eastern Papua from north- eastern Australia. Not of back-arc origin.	Weissel & Watts (1979)
D'Entre-casteaux	65-80	Remnant of an old basin.	Lapouille (1982)
North Loyalty	41-55	Possibly related to Coral Sea, opening afterwards as a result of change in plate convergence (Jurdy & Stephanick, 1983).	Weissel & others (1982b), Lapouille (1982)
Coriolis Trough	Plioc-ene-nk	Incipient behind-arc spreading. Difference of opinion on whether extension is active or not.	Karig & Mammerickx (1972), Luyendyk & others (1974), Dubois & others (1978)
North Fiji	0-10	Complex plate-boundary configuration with triple junction and major transforms.	Chase (1971), Malahoff & others (1982)
South Fiji	25-34	Complex plate-boundary configuration with triple junction.	Watts & others (1977), Malahoff & others (1982)
Lau	0-4	Classic 'back-arc basin' in the sense of Karig (1971).	Lawver & others(1976), Weissel (1977)
Havre	0-3	Southerly continuation of the Lau Basin.	Malahoff & others (1982)
Norfolk	nk	Nature and origin unclear.	Kroenke & Rodda (1984)
New Cal-edonia	Late Cretaceous - Paleocene	Sediment filled. Age, nature, and origin unclear.	Kroenke & Rodda (1984)
Tasman Sea	56-77	Small ocean basin. Not of back-arc origin. Intraplate volcanism in west and in eastern Australia.	Hayes & Ringis (1973), Weissel & Hayes (1977)

Most age ranges (Ma) taken from Taylor & Karner (1983) who inferred the ages from identified magnetic anomalies using the reversal time scale of LaBrecque & others. (1977). nk: not known.

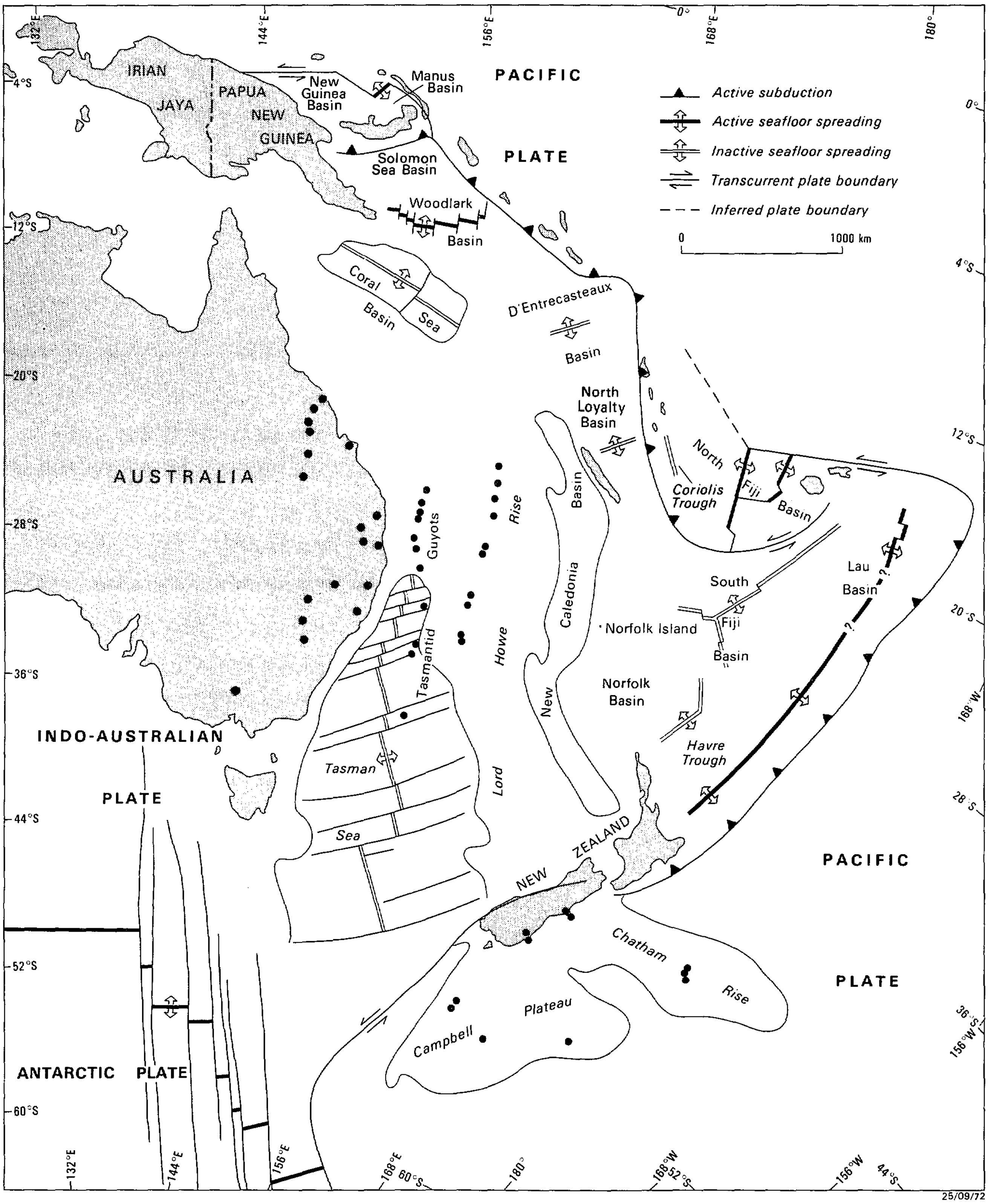

Figure 1. Marginal basins of the southwestern Pacific. Adapted from Wellman & McCracken (1979) and references listed in Table 1. Filled circles represent intraplate volcanoes (only felsic volcanoes shown for eastern Australia).

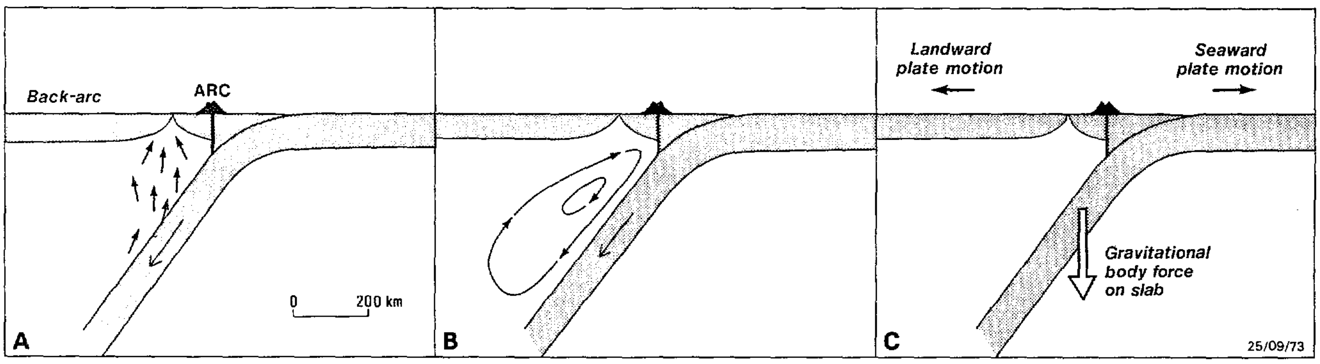

Figure 2. Models for the origin of marginal basins. (A) Diapirism resulting from heat or fluids (or both) from the downgoing slab. (B) Secondary convective flow induced by downward movement of the slab. (C) Global interactions of the major plates causing rollback of the trench and extension of the overriding plate.

Extensional magmatism in the Australian Proterozoic and its metallogenic implications.

L.A.I. Wyborn, BMR.

There are three major periods of basin-forming activity in the Australian Proterozoic: 2000-1860 Ma., 1800-1600 Ma., and 1100-600 Ma. Each basin sequence can be subdivided into rift and sag basins, and each period is separated by a period of deformation. The associated magmatism extended over 1500 Ma, and is dominated by suites of igneous rocks that are mafic, felsic, or bimodal; there are few suites of intermediate rocks (Figs. 1 and 2). The Late Proterozoic basins are often developed on top of earlier Proterozoic basins and, with time, the amount of igneous activity decreases in the superposed basins.

There are mainly two phases of igneous activity associated with the development of these sedimentary basins: a pre-rift phase which is characterised by major dyke swarms, and a rift phase which can be subdivided into two stages; a) rift-fill volcanism, and b) large-scale "anorogenic" granites which are coeval with, but emplaced some distance from, the main area of surface-rift development. There is very little igneous activity in the sag basins.

1. The pre-rift stage: Mechanisms of rift formation are divided into two classes: active mechanisms whereby ascending mantle convection thins the lithosphere, causing crustal doming and lithospheric failure; and passive mechanisms, whereby tensional stresses cause the failure of the continental lithosphere, and mantle diapirs penetrate to the base of the crust causing thinning (Turcotte & Emerman, 1983; Bott, 1981). For active rifts, the surface expression of these rising mantle plumes are the abundant dyke swarms that precede the earliest and Late Proterozoic basins at 2000 - 2300 Ma, and 1100 Ma respectively. In the lower crust, a percentage of these mantle-derived melts must underplate, as a considerable proportion of Proterozoic I- type felsic magmas have source ages coeval with the Early Proterozoic dyke swarms.

2. The rift stage: As a general rule the rift-fill volcanics tend to be mantle-derived, with the felsic, crustal-derived melts more prominent in the basement highs. Overall, the petrology of the igneous rocks is a very poor indicator of rift style compared with the more direct structural/ stratigraphic

features such as topography, faulting, and sediment fill (Sengör & Burke, 1978).

The variation in the composition of the mafic rocks is thought to reflect the degree and depth of partial melting in the mantle, which, in turn, is controlled by the nature and style of rifting (Dixon & others, 1981). Thus, depleted tholeiites, which are abundant in the Early Proterozoic basins, probably reflect a fairly rapid rate of extension in which there was a large degree of partial melting at shallow depths, rather than indicating that they were formed in an ocean-floor environment. This is reinforced by the fact that many of these depleted basalts are interbedded with fluviatile sediments, and in turn, are overlain by shallow-marine sequences. Conversely many of the Middle to Late Proterozoic alkaline volcanics could indicate a style of rifting that caused small degrees of partial melting at greater depth in the mantle. Where successive rifts develop on top of one another, there appears to be a progression with time from high-Mg basalts, to enriched tholeiites, followed by alkaline mafic volcanics.

The occurrence of felsic melts appears to depend on whether a suitable source material is available in the lower crust. All felsic melts are enriched in incompatible elements (in particular Zr, Nb, Y, U, and F), with the absolute levels varying systematically with each magmatic episode, so that each magmatic event has a unique composition (Wyborn & others, 1987). None of the felsic igneous rocks is derived directly from the mantle; all are derived from sources which have resided in the crust for at least 50 Ma, and in the case of the younger felsic melts, their source ages correspond to earlier Proterozoic extensional episodes. The felsic melts are uniform in composition over wide areas, often up to 4000 km^2, and must come from fairly homogeneous source regions. Thus, two major tectonothermal events are required to generate each felsic melt - one to form its source, and the second to produce the melt.

These felsic melts tend to be coeval with, but displaced from, the surface rift basins. However, as the detachment model of continental extension predicts, the thinnest lithosphere, and hence the maximum thermal anomaly, is displaced from the maximum area of surface-rift development. Lister & others (in press) predict that in this area of thinnest lithosphere, there will be mantle- derived underplating, which could lead to crustal melting (Fig. 3). This underplate could also become a viable source rock for felsic melts during future tectonothermal events (Wyborn & others, in press).

As a rule, it is difficult to recognise many of the classical structural extensional features in Proterozoic sequences because of subsequent deformation and metamorphism. Most Proterozoic tectonic models rely on the principle of lateral accretion and the concept of terrane analysis, with comparisons made to present-day analogues of continental and island-arc subduction and continental collision. However, by applying terrane analysis, whereby each fault-bounded package is considered as a separate entity, possible interrelationships between the various rocks packages that occurred during their formation may not be noticed, as the proviso is often applied that each package probably formed hundreds of kilometers away, and were brought together by collisional plate motions.

Nonetheless, with the detachment model for continental extension, there is produced a series of juxtaposed rock packages comprising the upper and lower plates, sag basin, rift basin, and core complex. Although each is usually separated by major fault systems, it is still possible to correlate across these packages and show that there is an interrelationship. Specific problems with the applications of modern analogues are that these Proterozoic "suture zones" (usually equivalent to the extensional detachment faults) never have ophiolites, and the faults in the "thrust duplex package" (usually the zone of listric faults) can be shown by sedimentology to be original growth faults which were present during sedimentation. Although the metamorphism is coeval with the so-called collision event, it is characterised by anticlockwise P-T-t paths, as would be expected in an extensional setting, as opposed to the clockwise P-T-t paths of the collisional terranes. Magmatism is always dominated by mafic and/or felsic rocks, such as is characteristic of extensional zones (Lipman & others, 1972). Mantle-normalised trace-element diagrams ("spidergrams") are always characterised by Sr depletion and Y non-depletion, inferring that during both the formation of the sources of the felsic melts and in their subsequent reworking into the upper crust, plagioclase, but not garnet, was stable. This implies that for both these magmatic events, melting took place at less than 150 mpa, at depths compatible with the high heat flow of extensional systems.

What is apparent from the magmatic systems of the Australian Proterozoic is that there is a series of synchronous continent-wide extensional events, and nowhere is there evidence for arc magmatism to suggest that these are back-arc extensional basins driven by subduction zones. The common magmatic dates, e.g. 1670 Ma, 1740 Ma, 1450 Ma, and 1100 Ma, are also prominent globally, as are the distinctive compositions. It can also be inferred that mantle dynamics is the driving force behind these continent-wide, if not global, extensional events. As an example of this, it can be shown that the earliest Proterozoic basins can be aligned in a pattern consistent with small-scale mantle convection (Wyborn, in press; Etheridge & others, 1987).

The dominance in the Proterozoic of extension tectonics has metallogenic implications. There is an abundance of rift-related, or structurally-controlled ore deposits, whilst Au- and Cu-porphyry systems are rare, presumably because of the scarcity of magmas similar to those found in subduction zones. Two important ingredients in these Proterozoic rift systems are the abundance of evaporite-rich carbonates in the sag basins, and the fluorite-rich anorogenic granites. When altered by either metamorphic, diagenetic, or surficial processes, these units tend to release a fluid that is oxidised, and either chlorine- or fluorine-rich, and which appears to have the capacity to mobilise two main groups of elements - U, Au, Pt, and Cu or Pb, Zn, and Ag.

References

Bott, M.H.P., 1981 - Crustal doming and the mechanism of continental rifting. Tectonophysics, 73, 1-8.

Dixon, J.E., Fitton, J.G., & Frost, R.T.C., 1981 - The tectonic significance of post-Carboniferous igneous activity in the North Sea Basin. In Illing, L.V., & Hobson, G.D., (Editors) - Proceedings of the second conference on petroleum geology of the Continental Shelf of north-west Europe. Institute of Petroleum, London, 121-137.

Etheridge, M.A., Rutland, R.W.R., & Wyborn, L.A.I., 1987 - Orogenesis and tectonic process in the early to middle Proterozoic of Northern Australia. In Kröner, A., (Editor) - Proterozoic lithospheric evolution. American Geophysical Union, Geodynamics Series, 17, 131-147.

Lipman, P.W., Prostka, H.J., & Christiansen, R.L., 1972 - Cenozoic volcanism and plate-tectonic evolution of the Western United States. I. Early and Middle Cenozoic. Philosophical Transactions of the Royal Society of London, Series A, 271, 217-248.

Lister, G.S., Etheridge, M.A., & Symonds, P.A. - Detachment models for the formation of passive continental margins. Tectonics (in press).

Sengör, A.M.C., & Burke, K., 1978 - Relative timing of rifting and volcanism on Earth and its tectonic implications. Geophysical Research Letters, 5, 419-421.

Turcotte, D.L., & Emerman, S.H., 1983 - Mechanisms of active and passive rifting. Tectonophysics, 94, 39-50.

Wyborn, L.A.I. - Petrology, geochemistry, and origin of a major Australian 1840-1880 Ma felsic volcano-plutonic suite. Precambrian Research (in press).

Wyborn, L.A.I., Page, R.W., & Parker, A.J., 1987 - Geochemical and geochronological signatures in Australian Proterozoic igneous rocks. In Pharaoh, T.C., Beckinsale, R.D., & Rickard, D.T., (Editors) - Geochemistry and mineralisation of Proterozoic volcanic suites. Geological Society of London, Special Publication, 377-394.

Wyborn, L.A.I., Page, R.W., & McCulloch, M.T. - Petrology and geochemistry of the post-1820 Ma granites of the Mount Isa Inlier: implications for the generation of Proterozoic "anorogenic" granites. Precambrian Research (in press).

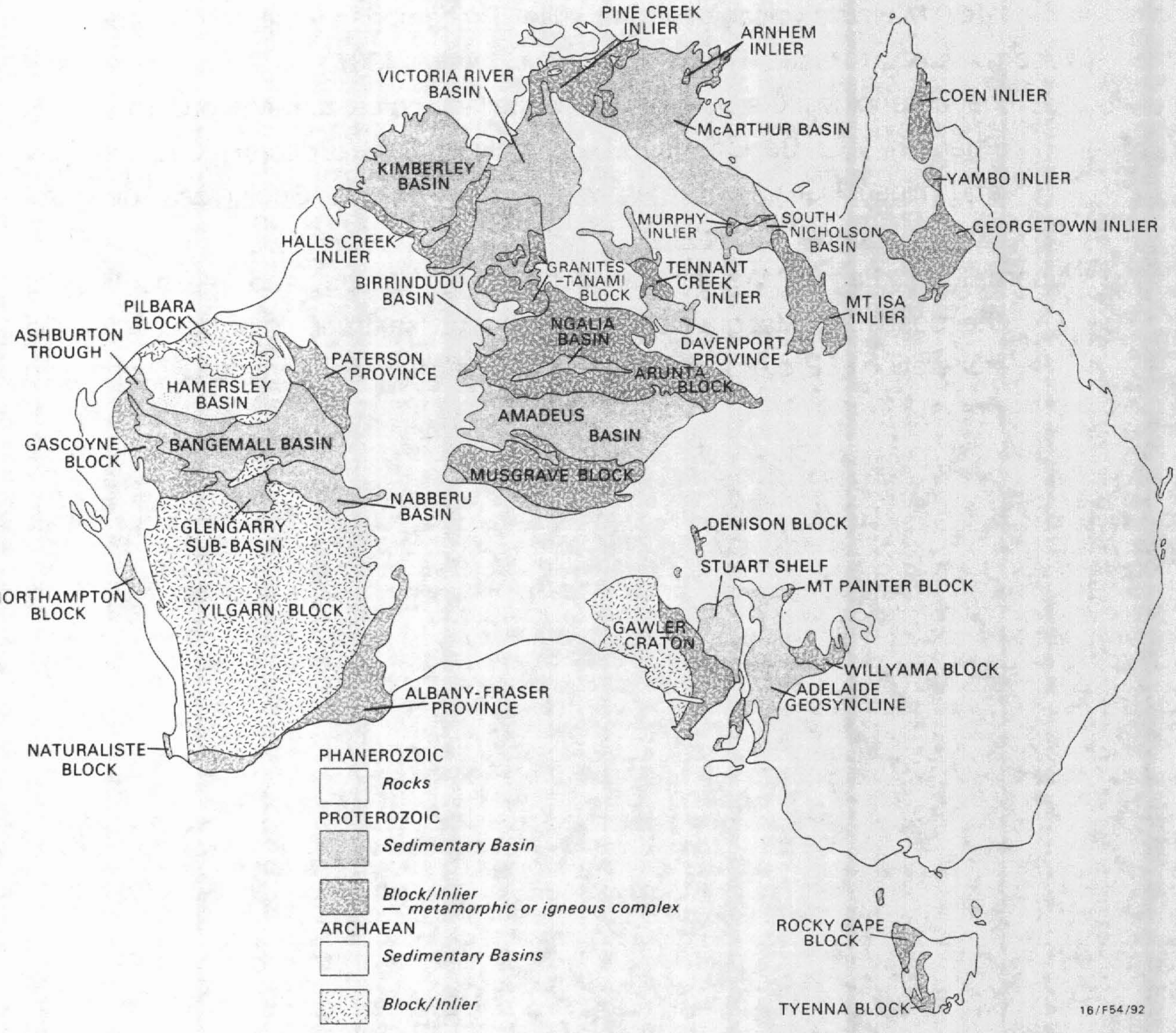

Figure 1. Distribution of Australian Proterozoic domains.

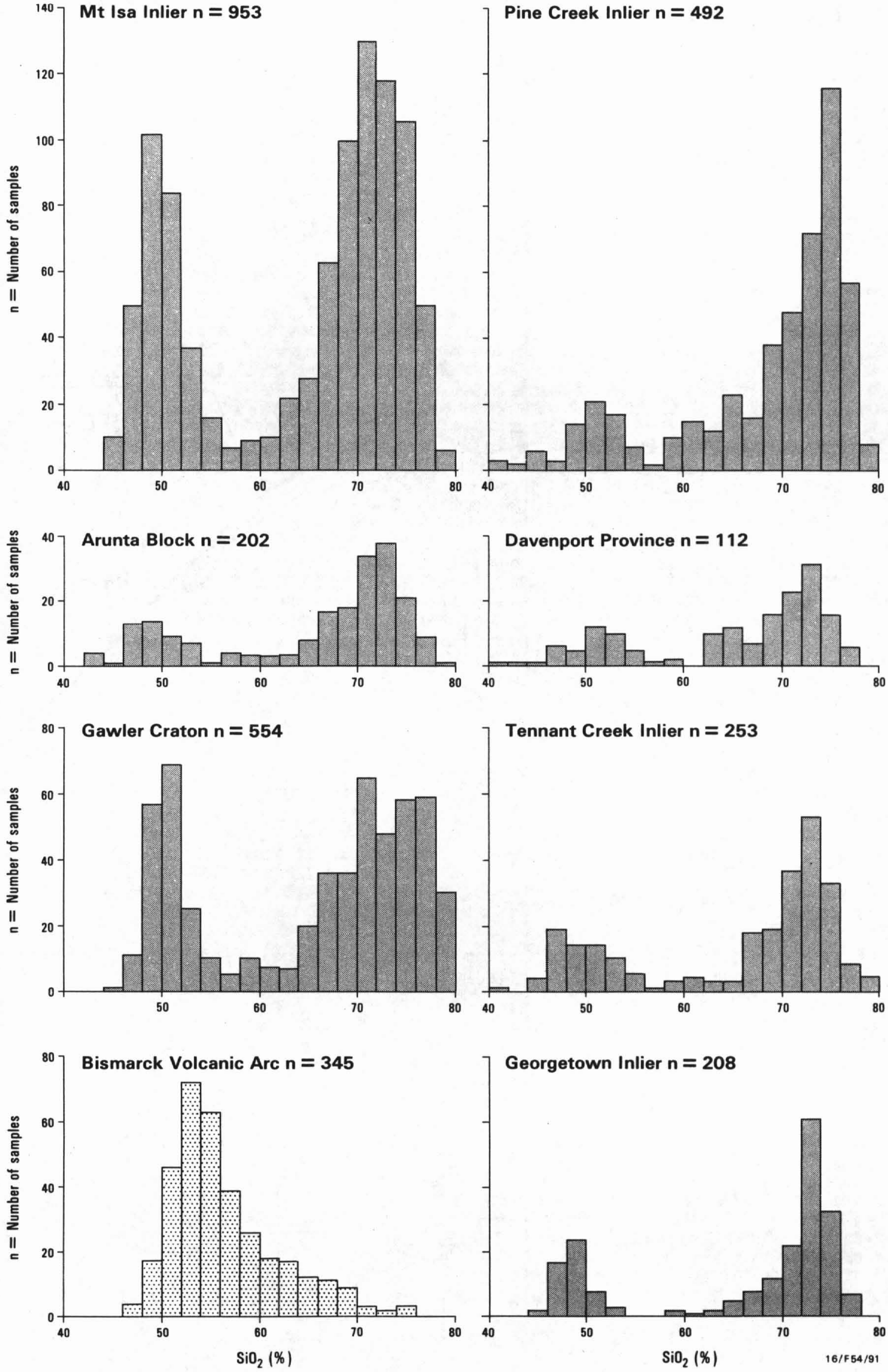

Figure 2. Frequency distribution histogram for SiO_2 for the Proterozoic domains of Australia. Comparison is also made with analyses of the Cainozoic Bismarck Volcanic Arc.

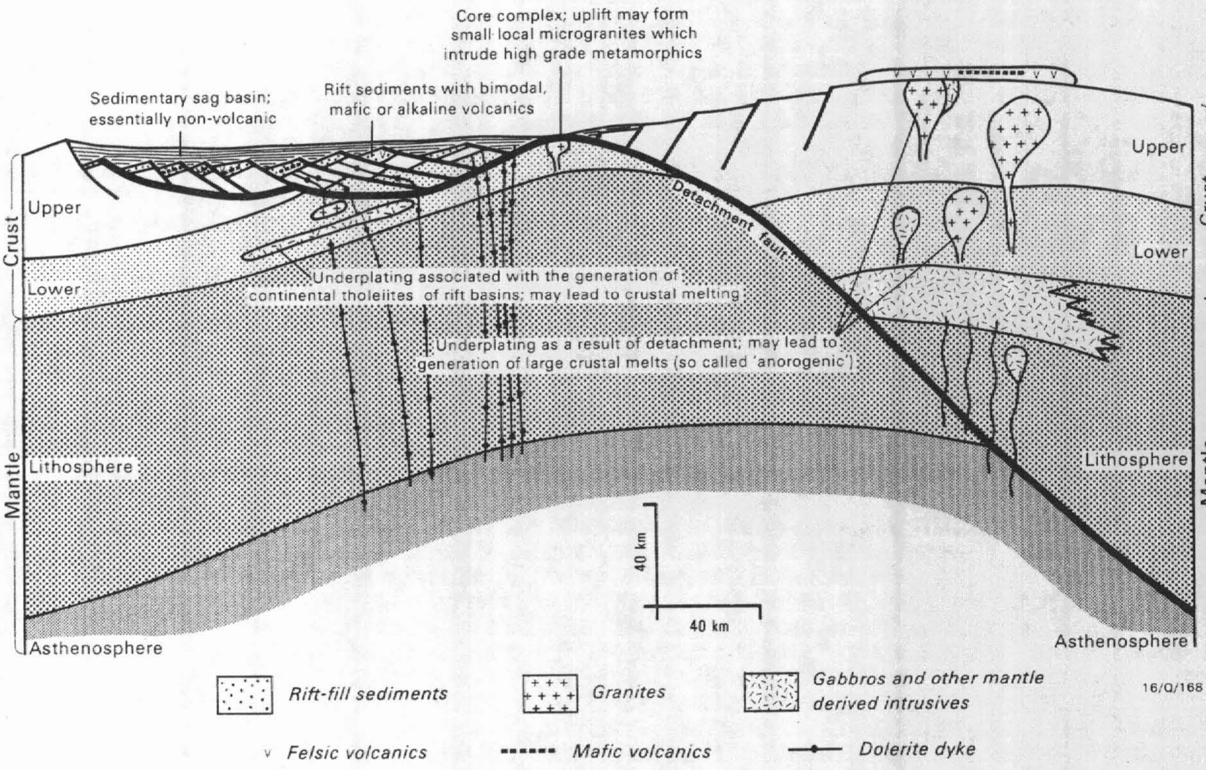

Figure 3. Schematic diagram showing the relationship of underplating and granite generation to asymmetric extension tectonics.

Hydrothermal processes during continental extension and their relationship to ore deposition

R W Henley, BMR

Rather like footprints in the sand, hydrothermal ore deposits provide evidence of past dynamothermal events in the evolution of the earth's crust. They record the coupling of key processes of lithosphere tectonics, metamorphism and plutonism and of massive fluid transfer through and within the evolving crust. Nowhere are these processes coupled so strongly as in the rifted environments of the crust. In these environments interactions of crust, mantle lithosphere and aesthenosphere occur, providing a critical interfacing of some of the Earth's major reservoirs of ore-forming components.

General relations between rifting, bimodal volcanism and ore formation have been discussed elsewhere (Sawkins, 1982; Sillitoe, 1982). A wide range of metals occurs as deposits in rifted settings and relates to periods of early or late magmatism, specific evolutionary paths for host magmas, and to the unique combination of elevated heat flow, tectonics, sedimentation and atmosphere-hydrosphere interaction which occurs in these dynamic environments. Robbins (1983) has focussed on specific sedimentary environments in crustal rifts, the occurrence of geochemically distinctive, often organic-rich rift lakes, local hydrologic controls, and rapid sedimentation. The latter is especially important as a process for rapid burial and preservation of sensitive ore environments as well as burial of distinctive geochemical facies (hematitic sandstones, metalliferous muds) which may subsequently control hydrothermal transport and deposition of metals.

For hydrothermal deposits, focussing of fluid flow and the total mass flux of fluids are dependent on structural control and the transient thermal gradient imposed on the crustal setting (Henley & Hoffmann, 1987). Understanding of the interdependent linkages between these factors and the geochemistry of fluids and host rocks is crucial to determining the origin of mineral deposits in rifted terranes.

Data for the solution chemistry of the major ore metals provide one of the keys to understanding of these linkages. Figure 1 shows the relationships between solution concentration and metal value with respect to formation of significant ore bodies in similar periods of time. Attainment of threshold values for economically significant metal transport is dependent upon the concentration of the major complexing ligands required for a given metal and, in turn is dependent upon environmental factors - the presence of enhanced salinity, fluids, sources of reduced sulphur, etc.

Although a much, much wider data base for hydrothermal chemistry is needed, reliable data are now available for the complexing of some of the major base and precious metals. For base metals (Pb, Zn) chloride is the most effective ligand for metal transport, with concentrations generally greater than that of seawater required (20,000 mg/kg). For gold, as a bisulphide complex, H_2S concentrations of about 100 mg/kg are required for significant transport but, in contrast to base metals, low salinity is necessary in all but the most specialist settings (discussed below). Silver and copper may be transported as either bisulphide or chloride complexes (Henley & Brown, 1986) but apart from Ag-Cl, complexing data are as yet unreliable. The bisulphide solubility contours for Ag_2S shown in Figure 2 are based on field data. Figure 2 illustrates the relative transport capacity of these metals at 250^{o} in H_2S-Cl^- solutions under mineral-buffered conditions and the range of compositions of fluids actually observed in the Earth's crust at the present day (Henley, 1986). Such fluids encountered during geothermal and hydrological investigations represent the high-level hydrologic regimes of the Earth's crust. Stable isotope studies show that for these, and their fossil equivalents, fluid recharge to the systems is dominated by the hydrosphere-meteoric, ground-ocean, waters or evolved continental brines (brines evolved through surface evaporation or dissolution of evaporites). As shown in Figure 2, convection of the latter two more saline fluids, in relatively reducing sequences, inevitably leads to base-metal transport; deposition requires specific structural/sedimentological traps in or near the surface. A good example is the McArthur Basin (Neudert & Russell, 1981; Eugster, 1985). The formation of sediment-hosted lead-zinc deposits is therefore the predictable (Markovian) consequence of crustal rifting in relatively arid regions ($\pm$ ~40^{o} latitude).

Figure 3 shows the strong effect of temperature on the transport of base metals in rock-buffered systems. Because of this, economic base metal deposits may be generated in rifted terranes where high-level convection of relatively less saline (ocean) waters occurs in response to the high heat flow. The resulting deposits are the Kuroko/Archaean massive sulphides (Cathles, 1984) and similar assemblages are, of course, now well-known from the rifts of the oceanic spreading centres (Rona, 1983). In these environments reduction of ocean sulphate at high temperatures may be the major source of reduced sulphur.

The high H_2S contents required for significant gold transport in terrestrial groundwater-recharged systems appear to require a deep lithosphere source (Henley & Hoffmann, 1987; Kerrich, 1986) perhaps in association with input of mantle-derived melts rich in gold relative to normal crust (2-3 g/kg). The active environments of the Taupo Volcanic Zone (New Zealand) provide a good example. Asymmetric growth of the Taupo Zone is reflected in an asymmetric distribution of gas-rich, gold-rich systems. In this case, the relatively-low salinity, rock-buffered high-level hydrothermal systems operate in response to elevated heat flow and magmatism, but their potential for gold transport is determined by the extent of such deep fluid source interactions. Epithermal (and perhaps shear-zone) gold deposits therefore provide a unique record of such shallow crust/deep crust/mantle interaction.

Temperature is a less critical variable with respect to gold solubility, than in the case of base metals. However high temperature in the deep ligand source regime may be essential and reflects metamorphic decarbonation and desulphidation reactions and/or subcrustal input of sulphur (Henley, 1986). Introduction of deep-sourced H_2S may provide the difference between gold-rich (such as at Buchans, Newfoundland) and gold-poor massive sulphide deposits (e.g. Kosaka, Japan).

Another ore environment specific to rifted or extensional terranes is the Picacho-type discussed elsewhere in this symposium (R Kerrich). Metal suites in these deposits are similar to those in high-level epithermal deposits but the deposits are closely related to detachment faults. Again, interaction of fluid regimes appears to be important in their generation (Kerrich & Rehrig, 1987). Perhaps crustal extension at rates similar to, or faster than, epithermal ore emplacement may account for this deposit style by synchronous dismemberment of an evolving hydrothermal system?

As noted above, highly-evolved sedimentary environments may occur in rift-settings. In a relatively arid region, extensive surficial oxidation occurs to produce sediments rich in oxidised iron. Under such conditions, convecting fluid must be relatively oxidised so that gold may be transported as a thiosulphate complex (Webster, 1986) or, if acidic fluids are developed, perhaps by chloro-species as indicated in Figure 2. There are few experimental data available to test the constraints on the latter, but it is interesting to observe the possibility for co-transport of platinum group elements under the same conditions. The Proterozoic rift basins (which host deposits like Coronation Hill, NT) of Australia and elsewhere (Etheridge et al, 1987) have potential for such processes to have occurred but only scant data are available for these environments. Transport of U and V is likely also to occur in such relatively-oxidising environments.

In summary, extension zones provide the setting for development of a wide range of permutations of interactions between fluid source regimes, crustal and subcrustal rocks, enhanced heat flow and the atmosphere. They are non-equilibrium, high-energy settings ideally suited to provide the feedstock for chemical (geo- and bio-) processing on the crustal scale and development of some of the world's major mineral deposits.

References

Etheridge, M.A., Rutland, R.W.R., & Wyborn, L.A.I., 1987 - Orogenesis and tectonic process in the Early to Middle Proterozoic of Northern Australia. In Kroner, A., (Editor) - Proterozoic lithospheric evolution. American Geophysical Union, 131-147.

Eugster, H., 1985 - Oil shales, evaporites and ore deposits. Geochim. et Cosmochim. Acta, 49, 619-635.

Helgeson, H.C., 1969 - Thermodynamics of hydrothermal systems at elevated temperatures and pressures. Am. Jour. Science, 267, 729-804.

Henley, R.W., 1986 - Ore transport and deposition in epithermal environments. In Herbert, H., & Golding, S., (Editors) - Stable isotopes and fluid processes in mineralisation. Geol. Soc. Australia Spec. Pub. (in press).

Henley, R.W., & Hoffman, C.F., 1987 - Gold: sources to resources. In Australasian Institute of Mining & Metallurgy. Pacific Rim Congress Proceedings, 159-168.

Kerrich, R., 1986 - Fluid infiltration into fault zones. Chemical, isotopic and mechanical effects. Pageoph, 124, 225-268.

Kerrich, R., & Rehrig, W., 1987 - Fluid motion associated with Tertiary mylonitisation and detachment faulting: $^{18}O/^{16}O$ evidence from the Picacho metamorphic core complex, Arizona. Geology, 15, 58-62.

Neudert, M.K., & Russell, R.E., 1981 - Shallow water and hypersaline features from the Middle Proterozoic Mount Isa Sequence. Nature, 293, p.284-286.

Robbins, E.I., 1983 - Accumulation of fossil fuels and metallic minerals in active and ancient rift lakes. In Morgan, P., & Baker, B.H., (Editors) - Processes of continental rifting. Developments in geotectonics 19. Elsevier, 633-658.

Rona, P.A., 1984 - Hydrothermal mineralization at seafloor spreading centres. Earth Science Reviews, 20, 1-104.

Sawkins, F.J., 1982 - Metallogenesis in relation to rifting in continental and oceanic rifts. Amer. Geophys Union Geodynamics Series Volume 8, 259-269.

Sillitoe, R.H., 1982 - Extensional habitats of rhyolite-hosted massive sulfide deposits. Geology, 10, 403-407.

Webster, J.G., 1986 - The solubility of gold and silver in the system Au-Ag-S-O_2-H_2O at 25^{O}C and at 1atm. Geochim. et Cosmochim. Acta, 50, 1837-1846.

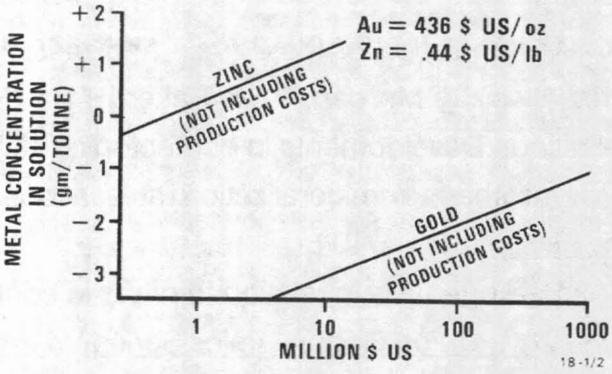

Figure 1. Relationships between deposit tonnage and solution chemistry and fluid flow rate as a function of the relative value of precious and base metals.

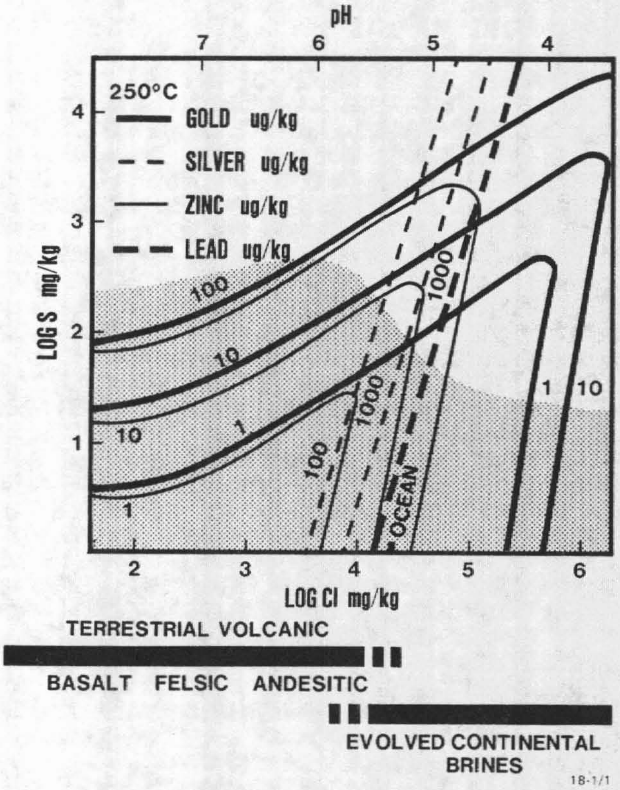

Figure 2. Relative solubilities of precious and base metals at 250°C in chloride-sulphide solutions whose pH and redox state are controlled by common mineral-fluid equilibria. Data from Henley (1986). Gold transport as a chloride complex may occur under high salinity relatively low pH conditions; solubility data for this environment are based on Helgeson (1969) and Henley (unpublished experimental data). The compositional range encountered in modern environments is shown by the stippled ornament and approximate salinity ranges indicated in the lower part of the diagram. H_2S is the dominant sulphide species in the compositional range shown but sulphate occurs under low salinity low sulphur conditions in the bottom left of the figure. (Halite saturation is ~5.54 log Cl).

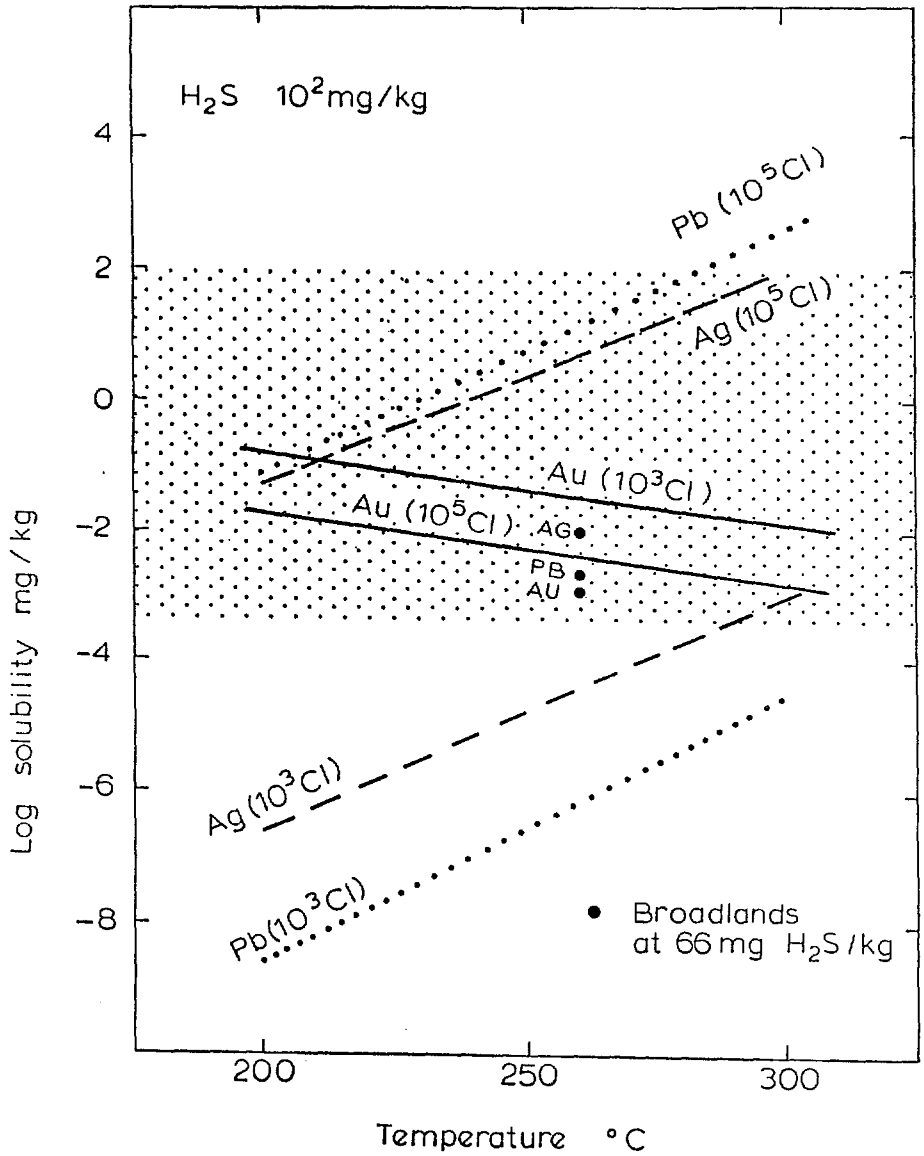

Figure 3. Solubility of precious and base metals as a function of temperature at fixed H_2S content in mineral buffered hydrothermal systems. Note the differing effect of temperature and salinity on base metal and gold transport. The stippled region emphasises the temperature-salinity-metal concentration range of principal economic interest (Henley 1986).

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Fluid dynamics in high heat flow extensional regimes

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From a fluid-dynamic point of view two kinds of high heat flow extensional regime must be distinguished: (1) Terranes where the crust is thinned by listric faulting or by movement on normal faults that bottom-out abruptly at ~15 km. In these terranes crustal extension tends to be distributed over broad regions, as in the Basin and Range Province in the USA, (2) Terranes where the crust is broken by through-going faults, and extension is focussed in ~60 km-wide grabens that are intruded by bimodal magmas as well as filled with sediments. Examples of the second kind of terrane include the Taupo Graben in New Zealand, the Salton Trough in the USA, and the Green Tuff Belt (GTB) in Japan.

The second kind of terrane, especially the GTB, is particularly instructive. The ~1500 km-long GTB cross-cuts the older geological grain. The areas where the belt formed were subaerial until close to middle Miocene time (~13 Ma), when the belt rapidly subsided to ~3.5 km below sea level, and ~350°C hydrothermal solutions debauched to the sea floor in isolated "mineral districts" producing the Kuroko massive sulphide deposits. The belt uplifted to near its former elevation by ~5 Ma, conveniently allowing the Kuroko deposits to be mined subaerially today.

The tectonic behaviour of the GTB, and the clustering of the Kuroko massive sulphide deposits in isolated mineral districts can be understood if the GTB formed as a result of an aborted rifting of the Japanese island arc. Figure 1 shows the hypothesised tectonic evolution of Japan. Figures 2 and 3 show how the aborted rift hypothesis can account for the GTB and its deposits.

The Kuroko deposits formed over ocean rift segments where heat was introduced massively to the crust by basaltic magmas. The fluids that formed the Kuroko deposits debauched to the sea floor at temperatures of 350°C. These temperatures are much less than the magma heat source, but very similar to the black-smoker vents on the seafloor today, and in fact very similar to the maximum temperatures in essentially every active and fossil (convective) geothermal system. Models of the convective cooling of intrusives by ground-water convection (e.g., Fig. 4) show that a rapid decrease in permeability with temperature above 300-350°C is required if fluids are to be vented at

~350^{o}C. Most (but not all) of the fluid must interact with the magmas at temperatures of ~350^{o}C.

The most likely explanation for the temperature regulation is a modified Lister thermal-cracking-front model in which, due to the rapid increase in rock dissolution kinetics at temperatures above ~300-350^{o}C, the points of contact that prop-open fractures and make them permeable are rapidly dissolved (and reprecipitated in pressure shadows) and the fractures healed. The healing is rapid enough that most of the fluid circulates through rock at temperatures of ~350^{o}C. This is not to say that some fluid does not enter the fracture the instant it is created and become heated to far higher temperatures. Such fluid will be removing heat from the magma in a heat pipe fashion: the hot end will produce vapour and the cool end condense it. Salt will be deposited at the hot end. The interplay of the production of low-salinity recondensed vapour and the dissolution of previously precipitated salt as the thermal-cracking-front migrates into the cooling intrusive could produce the salinity variations observed in the hydrothermal solutions associated with massive sulphide deposits and black smokers.

The significance of temperature control on permeability cannot be over-emphasised. In high-temperature systems (systems able to create their own permeability by thermal cracking) it is the main permeability control. The only other control of equal significance is the alteration of rock to plastic products, such as the alteration of basalt of serpentine.

Temperature-dependent permeability provides insight into the origin of gold deposits in massive sulphide districts. As Kerrich & Fyfe pointed out, lode gold deposits along breaks in greenstone belts are peculiar in containing very little base metal (Fig. 5). The selectivity of the gold mineralising solutions for gold and against base metals, and conversely the selectivity of seawater-salinity hydrothermal solutions for base metals and against gold (relatively) can be understood from the very strong inverse-salinity dependence of gold solubility in rock-buffered hydrothermal solutions (Fig. 6). Gold has very much greater solubility in low-salinity solutions because such solutions have higher pH and thus have much greater HS^- concentrations for a given rock-buffered level of H^2S. HS^- is the most important gold complexing agent. Conversely the base metals are mainly complexed by Cl^-, so their concentration increases with increasing salinity.

Putting the two observations (chemistry and temperature-dependent permeability) together, an explanation for the location of gold deposits along breaks near the amphibolite-greenschist facies boundary can be found (Fig. 7). Permeabilities will be low in the amphibolite facies because of its $\geq$400^{o}C temperatures. Thus geopressure is likely to develop when hydrated minerals break down, releasing low-salinity, gold-scavenging waters. These fluids will eventually accumulate and hydrofracture through to the more permeable greenschist facies. Decompression for a variety of reasons is likely to trigger mineral and gold precipitation.

If lode gold deposits do represent a record of fluid escape and decompression from geopressure, their study could provide insights useful to understanding the escape of hydrocarbon-bearing fluids from geopressured zones such as those in the USA Gulf Coast.

References

Cathles, L.M., 1986 - The geologic solubility of gold from 200-350^{o}C, and its implications for gold-base metal ratios in vein and stratiform deposits. Canad. Inst. Mining and Met. Sp. Vol. 38, 187-208.

Cathles, L.M., 1983 - An analysis of the hydrothermal system responsible for massive sulfide deposition in the Hokuroku Basin of Japan. Economic Geology Monogr., 5, 439-487.

Cathles, L.M., Guber, A.L., Lenagh, T.C., & Dudas, F.O., 1983 - Kuroko-type massive sulfide deposits: products of an aborted island arc rift. Economic Geology Monogr., 5, 96-114.

HYPOTHETICAL EVOLUTION OF JAPANESE ARC

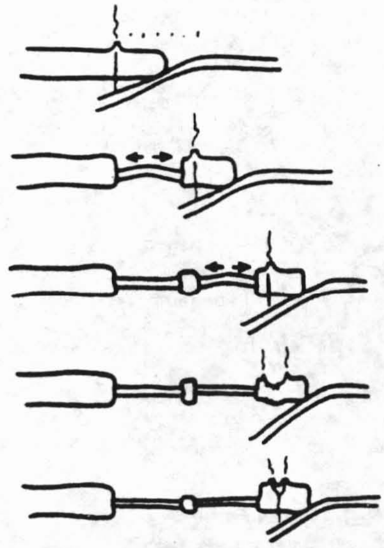

Figure 1: Hypothetical evolution of the Japanese island arc. Before 30 m.y. ago, Japan was attached to Asia and subduction was of the Andean type. Subsequently there were two successful riftings of the Japanese island arc. The first opened the Sea of Japan and the second, the Yamato basin (see Fig. 3). We associate the formation of Kuroko massive sulphide deposits in the Green Tuff Belt with a rifting event that occurred ~13 m.y. ago that tried unsuccessfully to open a third marginal basin. We call this aborted attempt a failed rift.

FLOATING WOOD BLOCKS

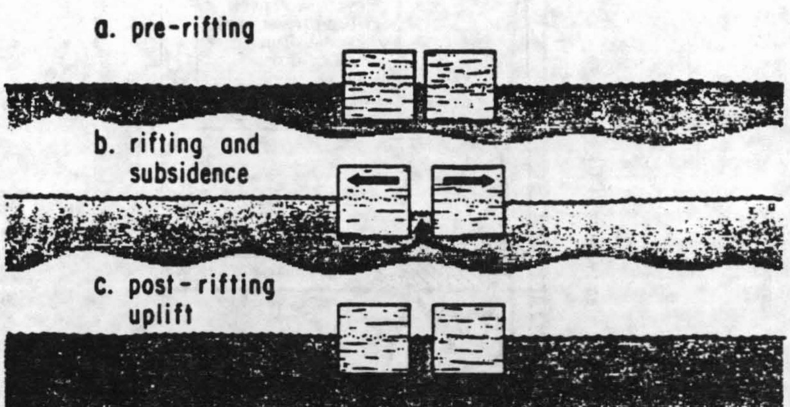

Figure 2: The premineralisation rift-related subsidence and post-mineralisation uplift observed in the Green Tuff Belt can be understood by analogy to the water level between two wood blocks floating in water. If the blocks are stationary, the water level is the same between the blocks as outside (a). When the blocks are moved apart, water must flow up between the blocks. In doing so it encounters viscous resistance and loses fluid head. Thus, while the blocks move apart, the fluid level between the blocks is depressed relative to the normal level (b). If the motion of the blocks is stopped, fluid level between the blocks returns to normal. The water is analogous to the fluid asthenosphere upon which the lithosphere floats in isostatic equilibrium. Rifting should produce subsidence of the magnitude and character that occurred in the Green Tuff Belt (see discussion text). Cessation of rifting will result in an uplift to near prerifting elevations.

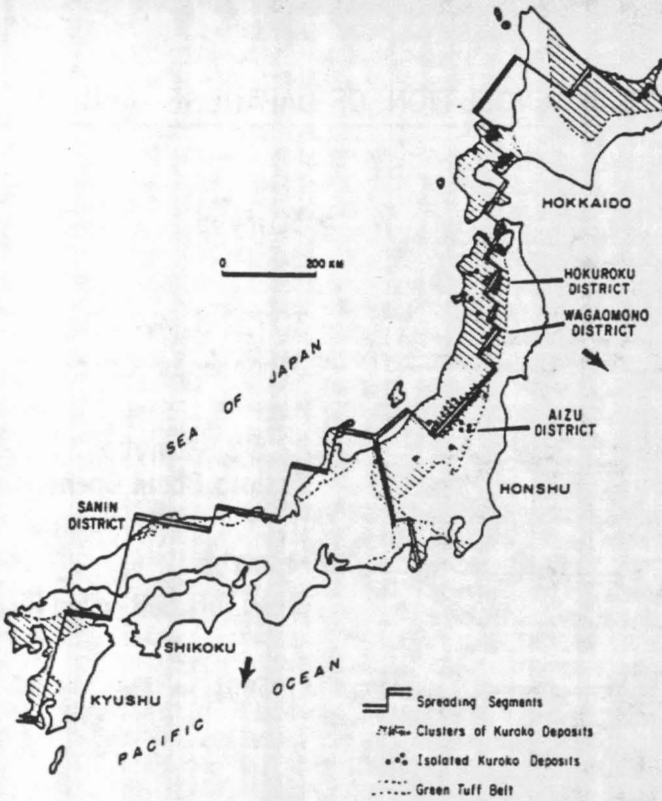

Figure 3: Mining districts within the Green Tuff Belt can be associated with a reasonable distribution of spreading centres offset by transform faults. The orientation of spreading segments in this illustration is chosen to be perpendicular to the presumed extension (heavy arrows). Divergence between the extension in northern Honshu, which follows the migrating Japan trench, and the extension in southern Honshu causes extension in the Fossa Magna area in central Honshu. The pattern of spreading segments and transforms is not the only pattern that can match the distribution of ore deposit clusters in northern Honshu. A pattern with spreading centres oriented northwest-southeast offset by northeast-southwest transform faults will do almost equally well; however, the pattern chosen accounts for the Green Tuff Belt in southern Honshu and the Fossa Magna area in a more natural way.

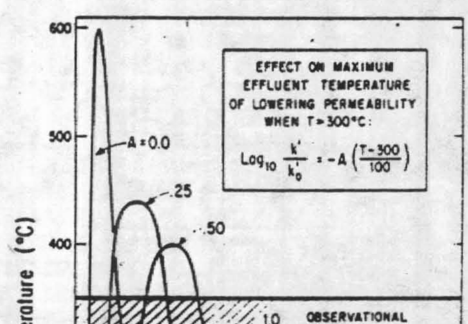

Figure 4: The calculated maximum temperature of vented solutions drops rapidly if the permeability of rock is decreased exponentially from its base level, k_0, at temperatures above 300°C. Such a temperature dependence on permeability is reasonable because the rate of rock dissolution reactions increases rapidly at T > 300°C, allowing removal of the topographic irregularities that prop fractures open and make them permeable (see discussion in previous section of text). If A in the formula shown is given a value of ~1.0, the calculated temperatures of venting fluids are within the range geologically observed.

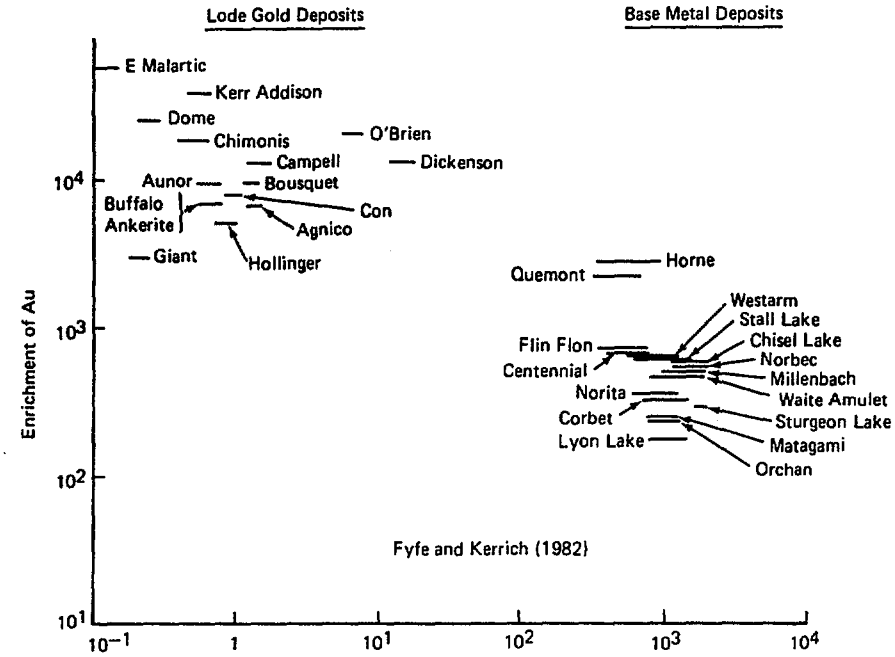

Figure 5: The enrichment of gold and Cu + Zn in a greenstone-hosted gold and base metal deposit relative to unaltered igneous rocks. The distinct bimodal populations require selection for gold and rejection of base metals in the lode gold deposits, and selection for base metals and relative rejection of gold in the massive sulphide deposits. Figure is from Fyfe & Kerrich (1983, Fig. 4).

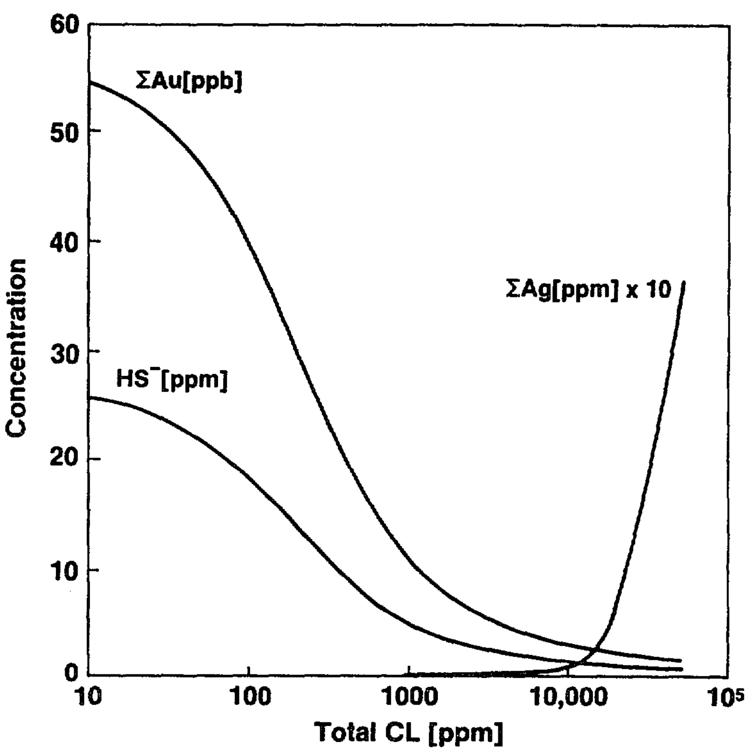

Figure 6: The geologic solubility of gold in 300°C solutions as a function of salinity. Concentrations calculated by EQ3. The solubility of silver as chloride complexes as shown to illustrate the general salinity dependence of metals for which chloride ligands are dominant. Recent studies of precipitates in geothermal wells in New Zealand indicates that complexes other than chloride must contribute to the solubility of silver (Brown, submitted). The silver curve should not therefore be taken literally but simply as a proxy for base metals that are carried as Cl complexes.

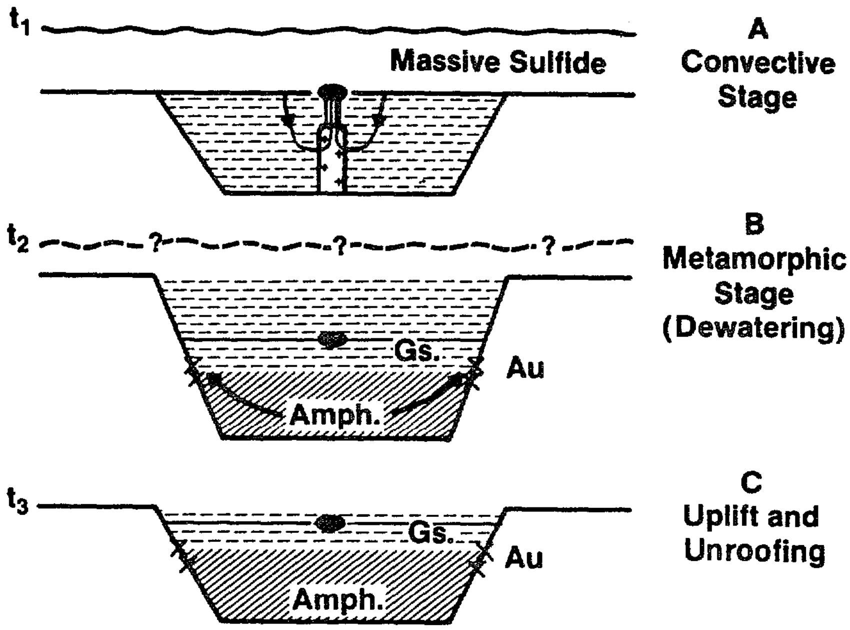

Figure 7: Cartoon illustrating the formation of massive sulphide and lode gold deposits in an evolving greenstone belt. During the period of most active rifting and extension, the belt is intruded by magmas, and seawater convection forms massive sulphide deposits on the sea floor (A).
Infilling of the basin with volcanics causes burial metamorphism. Hydrous minerals dewater, and lode gold deposits are formed where these low-salinity fluids decompress from lithostatic to hydrostatic pressures. For reasons discussed in the text this will most likely occur at border faults near the amphibolite-greenstone grade metamorphic boundary (B).
Finally uplift and erosion unroofs the deposits leaving them in a spatial configuration similar to that observed. The difference in salinity between seawater which precipitated the massive sulphide deposits, and the low-salinity metamorphic dewatering fluids that precipitated the lode gold deposits, accounts for the different metal contents of the deposits.

Epithermal mineralisation in the Western Pacific: essential characteristics, distribution and questions

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This brief review of epithermal gold mineralisation and its tectonic setting in the Western Pacific is divided into three parts: (1) a highlight of the basic attributes of epithermal systems (their geological but mainly geochemical characteristics), and the principal factors governing their formation; (2) the distribution of epithermal deposits in the Western Pacific and their relation to plate tectonic setting; and (3) questions that need to be addressed through problem-oriented investigations in order to elucidate the interconnection between the tectonic setting, source region and type of hydrothermal fluids, and subsequent mineralisation in the epithermal environment.

Characteristics of epithermal mineralisation

Epithermal mineral deposits are generally dominated (economically) by precious metals (though base metals can dominate, for example in the case of massive sulphides - a "special case" style of epithermal mineralisation). The epithermal environment of hydrothermal systems occurs near the surface at relatively low temperatures (to 350^{o}C) (Figs 1 and 2) and has a magma as its driving heat-source. In this shallow regime, often hosted by volcanic rocks in a subaerial to submarine setting, occur physical and chemical processes which are extremely conducive to metal precipitation, e.g., rapid temperature and pressure decrease, concomitant boiling with resultant gas loss, and mixing of the ascending mineralising fluid with ambient fluid. Given that such processes are common in this environment, it is necessary only for fluid flow to be focussed in order for deposition of significant grades to occur. However, the time necessary to form an ore deposit (and thus the probability of formation) will then be a function of the transport capability of the fluid, i.e., how much metal (e.g., gold) it can carry, assuming saturation.

The transport capability is essentially a function of the chemistry of the fluid, which has recently been clarified through the examination of geothermal systems (Henley & Ellis, 1983) and fluid inclusions of epithermal deposits (Hedenquist & Henley, 1985). Relatively gas-rich (CO_2 + H_2S) fluids favour the formation of gold-rich deposits, since gold is largely transported as a bisulphide complex (i.e., more H_2S in solution means more gold transported per unit mass of fluid), while relatively saline fluids favour the transport of chloride-complexed metals such as silver and base metals. Henley (1987) quantified this relationship between chemistry and metal content. In general, gold-rich deposits formed from fluids with <1 wt% NaCl and dominated by gas, whereas the silver and base metal (including massive sulphide) deposits formed from more saline (3-12 wt% NaCl), possibly lower gas, fluids.

For this reason, Hedenquist (1987) considered gold-rich and silver-rich (and base metal) divisions of epithermal mineralisation as indicating a first order (essential) variation in the fluid chemistry. Another essential division is based on the degree to which the oxidising and acid magmatic component of the epithermal fluid has been neutralised by interaction with host rocks. If interaction is large, the fluids are relatively reduced and have a near neutral pH; they may be dominated by low-salinity meteoric fluids, moderate-sulphidation fluids (i.e., reduced sulphur) and are equivalent to most geothermal systems (Fig. 2). The other extreme occurs when there is little primary neutralisation of the magmatic component and the fluids are relatively oxidised, acid and often saline. These are termed high sulphidation fluids, and are equivalent to near-vent volcanic discharges (Fig. 2); there is a large potential for discovery of this latter type in the Western Pacific (Hedenquist, 1987).

In the subaerial environment, low sulphidation systems are exemplified by Martha Hill (New Zealand), Kelian (Indonesia), Baguio (Philippines), Hishikari (Japan), McLaughlin (California), Creede (Colorado) and Fresnillo (Mexico) - the latter two related to evolved brines. The high sulphidation examples are Summitville (Nevada), Temora (Australia), Lapanto (Philippines), Chinkuashih (Taiwan), Mt Kasi (Fiji), and Nansatsu (Japan). Intermediate or overprinted styles include El Indio (Chile), and possibly Panguna (Papua New Guinea).

In a submarine hydrothermal setting, the meteoric component shown schematically in Figures 1 and 2 is replaced by sea water, resulting in higher salinities than observed for the gold-rich low-sulphidation systems; this explains the base metal and silver dominance in this environment. Examples of this type of mineralisation include the Kuroko deposits (Japan), Mt Lyell and Mt Morgan (Australia), and other massive sulphides (Ag/Au = 20-200), while a near-shore but still submarine environment is probable for mineralisation in the Drake area of Australia.

Questions related to source regions of hydrothermal fluids

Important questions to ask in relation to ore formation and therefore exploration concern the source of the components that eventually reach the epithermal environmental (e.g., chloride, CO_2, sulphur and metals). How does tectonic setting (i.e., style of subduction, type of magma, nature of crust, arc versus back-arc setting, subaerial versus submarine, etc.) relate to fluid chemistry? What causes gas-rich fluids to form in some environments and not in others? Henley (1987) has suggested that a variety of sources has the potential to supply important and diagnostic volatile components to epithermal systems associated with arc volcanism. These range from deep crustal metamorphism (CO_2, H_2S) to metamorphism and degassing of subducted oceanic crust and sediment (H_2S, CO_2, helium).

Sillitoe (1981) argues from geological and geochemical evidence that derivation of the metal suite characteristics of deposits related to volcanic areas is little influenced by either the maturity of the arc or the nature or composition of the crust. He concludes that these deposits, largely porphyry and epithermal, are closely related to magmas directly associated with the mantle wedge above subduction zones (though a more oxidised magma system may be related to island-arc rather than continental settings).

Figure 3 shows tectonic elements of the Western Pacific, and the distribution of many of the discovered epithermal deposits and prospect areas. The latter are dominantly hosted by products of calc-alkaline volcanism, though basalts host some mineralisation in Fiji, while mineralisation at Lihir (Papua New Guinea) is related to alkalic volcanism.

The tectonic system of mineralisation may be either continental (e.g., New Zealand, Kyushu (Japan), and eastern Australia, or oceanic (e.g., Vanuatu) in setting. The hydrothermal systems can be related either to magmatism in the arc or to back-arc (marginal basin) activity.

New Zealand and Kyushu also illustrate situations where oceanic-volcanic-arc and back-arc activity extend from a submarine to subaerial setting (Figs 4 and 5). This has the effect of replacing the higher-salinity-seawater circulation with meteoric-water circulation, and allows boiling to occur.

Research into the tectonic and magmatic environments of young epithermal mineralisation in the Western Pacific region becomes critical to the resolution of these questions and extension of epithermal concepts into older terranes.

References

Hedenquist, J.W., 1986 - Geothermal systems in the Taupo Volcanic Zone: their characteristics and relation to volcanism and mineralisation. Royal Soc. N.Z. Bull., 23, 134–168.

Hedenquist, J.W., 1987 - Mineralisation associated with volcanic–related hydrothermal systems in the circum–Pacific Basin. Trans. 4th Circum Pacific Energy and Mineral Resources Conference, Singapore, 1986 (in press).

Hedenquist, J.W., & Henley, R.W., 1985 - The importance of CO_2 on freezing point measurements of fluid inclusions: evidence from active systems and implications for epithermal ore deposition. Econ. Geol., 80, 1379–1406.

Henley, R.W., 1987 - Ore transport and deposition in epithermal environments. Proc. Symp. Stable Isotopes and Fluid Processes in Mineralisation, Brisbane, 1985, (in press).

Henley, R.W., & Ellis, A.J., 1983 - Geothermal systems, ancient and modern. Earth Sci. Reviews, 19, 1–50.

Karig, D.E., 1974 - Evolution of arc systems in the Western Pacific. Ann. Reviews Earth plan. Sci., 2, 51–75.

Kemp, P.J.J., 1984 - Neogene and Quaternary extent and geometry of the subduction Pacific Plate beneath New Zealand. Tectonophysics, 108, 241–266.

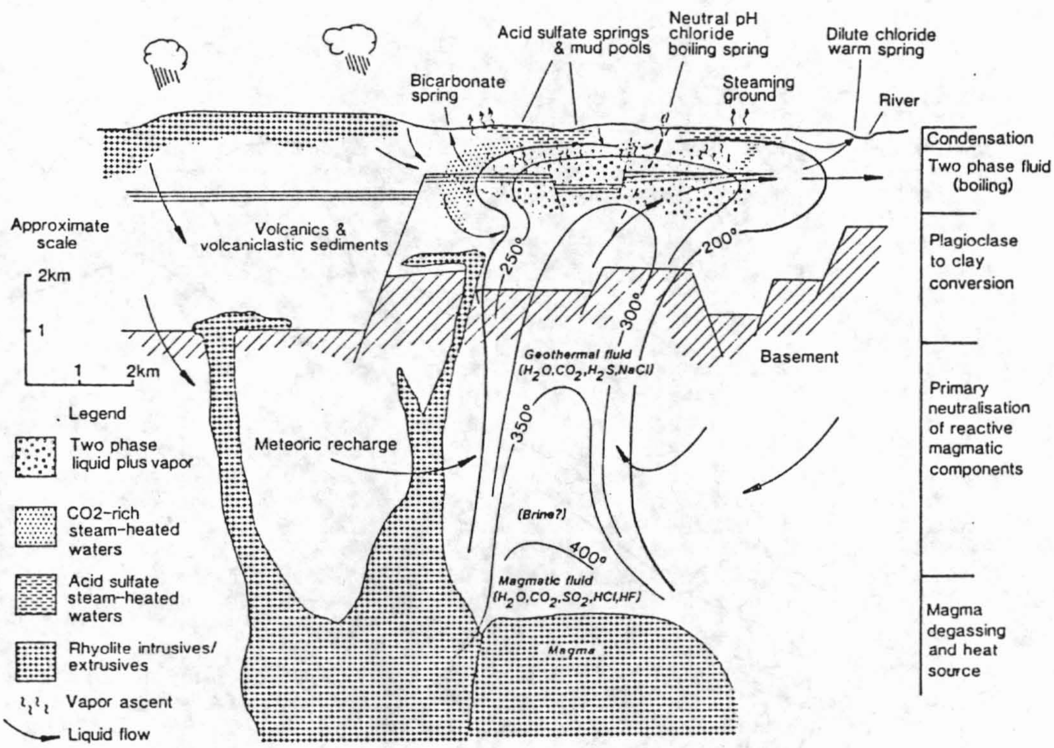
Sillitoe, R.H., 1981 - Ore deposits in Cordilleran and island–arc settings. Ariz. Geol. Digest, 14, 49–70.

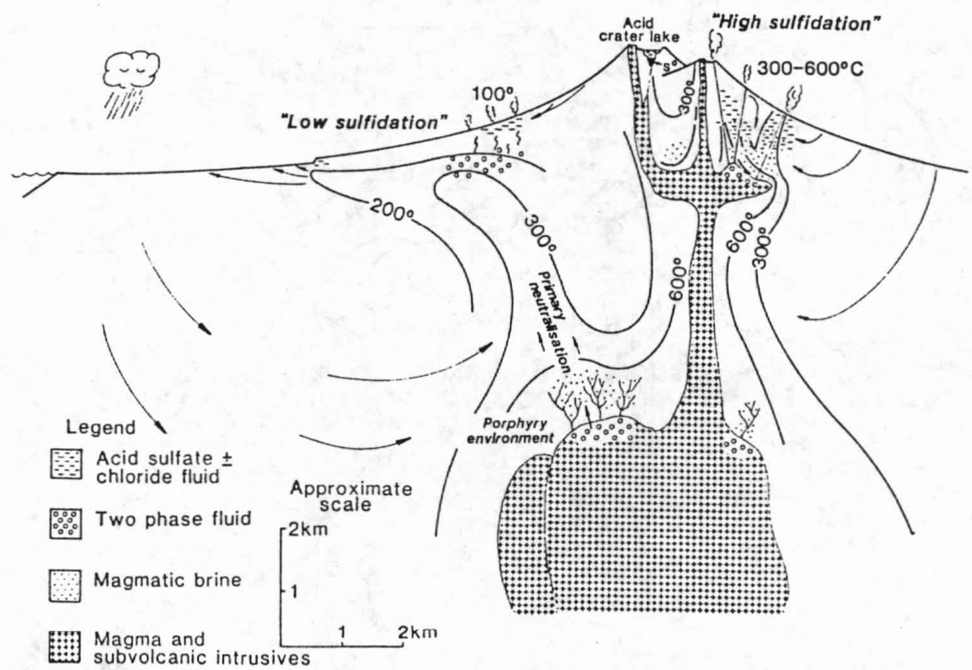
Sillitoe, R.H., 1987 - Copper, gold and subduction: trans–Pacific perspective. In Australasian Institute of Mining & Metallurgy, Pacific Rim Congress 87, Proceedings, 399–404.

Stern, T.A., 1985 - A back arc basin formed within continental lithosphere: the Central Volcanic Region of New Zealand. Tectonophysics, 112, 385–409.

Uyeda, S., 1987 - Active hydrothermal mounds in the Okinawa back–arc trough. EOS, Trans. Amer. Geoph. Union, 68(36), 737.

Uyeda, S., & Nishiwaki, C., 1980 - Stress field, metallogenesis and mode of subduction. Geol. Assoc. Canada Spec. Paper, 20, 323–340.





Figures 1 (top) and 2 (bottom): Schematic representations of fluid flow and epithermal processes in volcanic-hosted hydrothermal systems. Hosts are silicic and andesitic, with relatively low and high relief, respectively. The high sulphidation environment is located close to the heat source, and can have a large magmatic component. A little imagination allows one to envisage the variations in fluid flow (and source) and processes if these were submarine (i.e., seawater-dominated for the low sulphidation environment, and less lateral flow due to smaller hydraulic gradients). From Hedenquist (1986, 1987) and references herein.

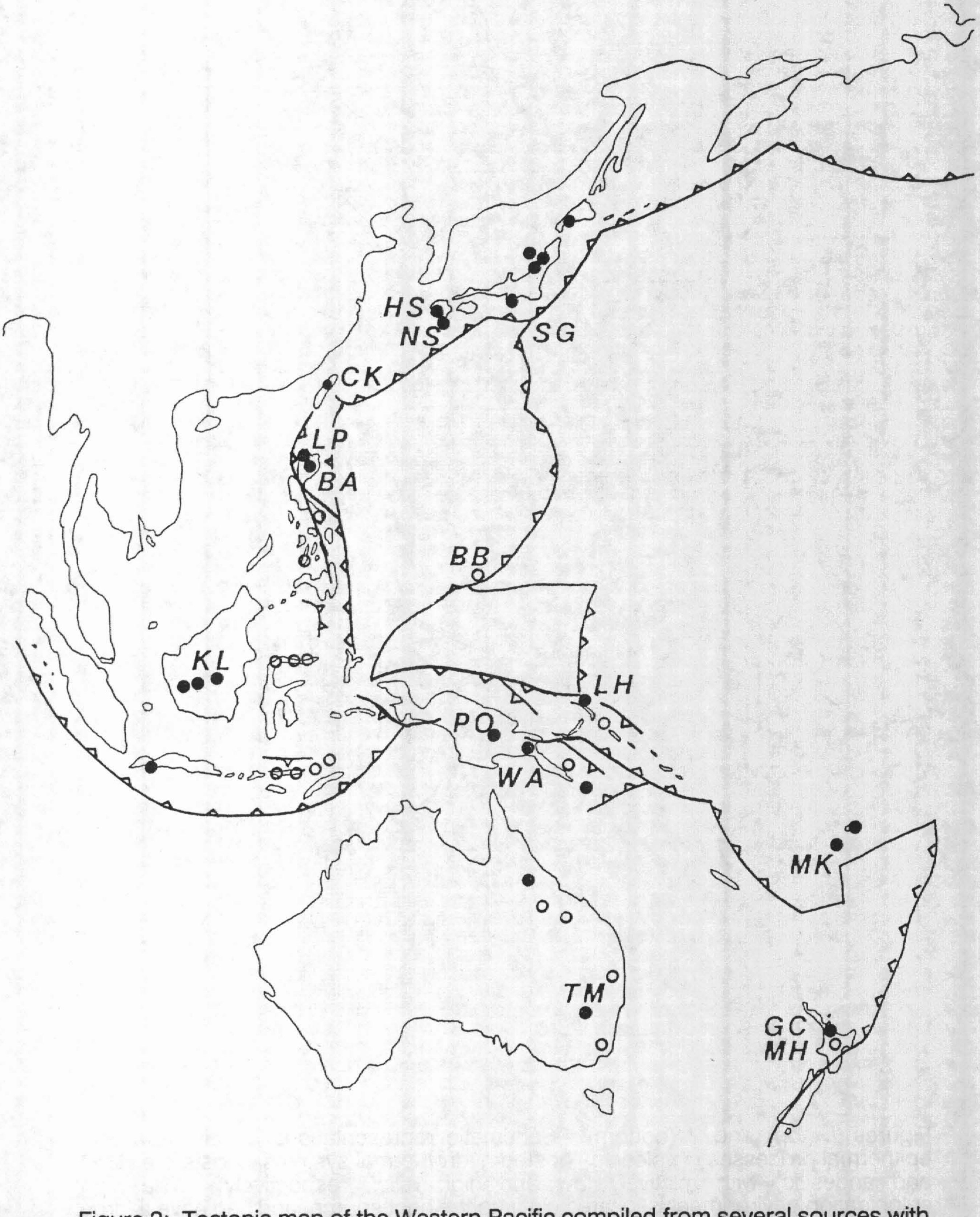
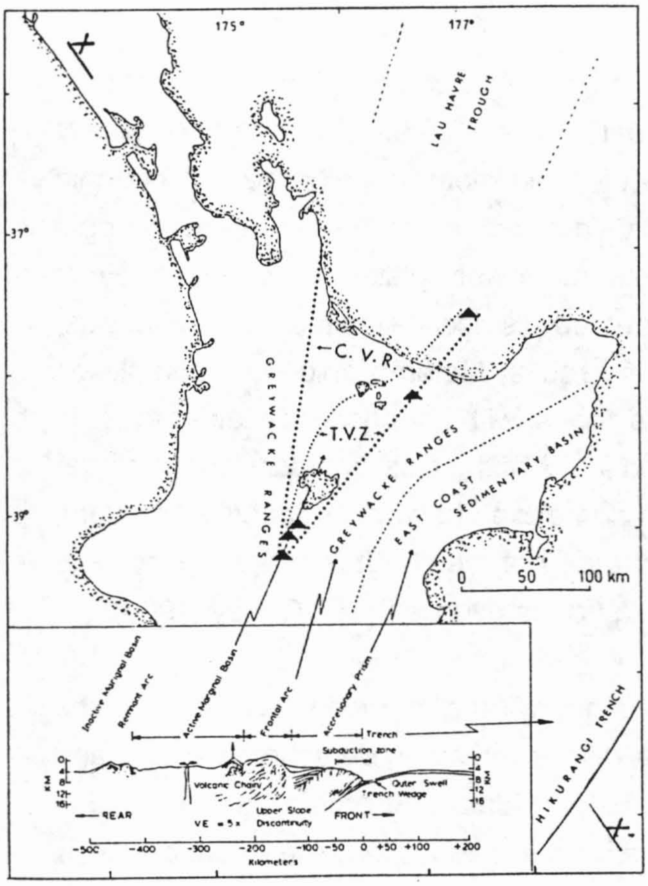
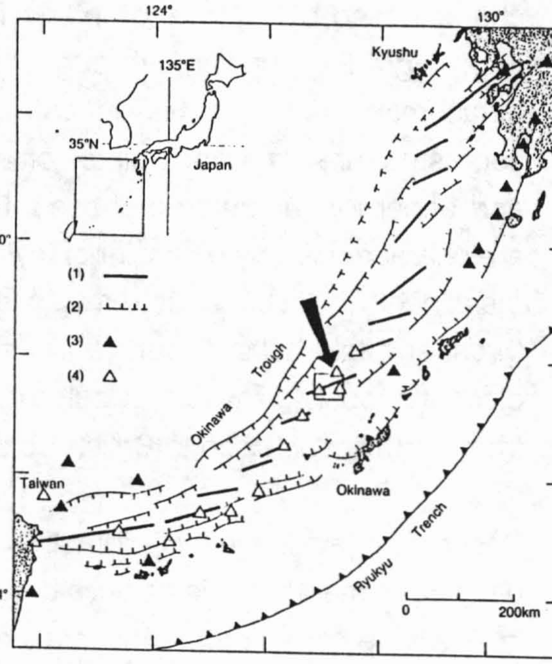


Figure 3: Tectonic map of the Western Pacific compiled from several sources with the location of several epithermal deposits (filled circles) and prospect areas (open circles) shown. Mineralisation is essentially confined to volcanics in a subduction-related volcanic arc - back-arc setting; this varies from Palaeozoic (eastern Australian) to active (Kyushu and New Zealand).





Figures 4 (top) and 5 (bottom): Trench-volcanic-arc - back-arc (marginal basin) sequence east to west for New Zealand (oceanic to north) and Kyushu (oceanic to south), respectively. Active and fossil epithermal activity is present in both areas related to arc and back-arc volcanism. Arsenic and molybdenum-bearing siliceous precipitates have been recently identified in the Okinawa Trough (arrow), the first directly observed hydrothermal activity in a back-arc basin (analogous to the Kuroko setting). From Stern, 1985, and Uyeda, 1987, respectively.

Fluid-metasomatic regimes in metamorphic core complexes

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Metamorphic core complexes constitute a distinct structural-lithological entity in the tectonic architecture of the North American Cordillera. They are geologically characterised by a 'core' of intensely-deformed metamorphic and (or) plutonic rocks, grading upward through gently-dipping mylonites to a low-angle detachment fault. This infrastructure, or lower plate, is overlain by a suprastructure, or upper plate, of attenuated, listrically-faulted, allochthonous, and generally-unmetamorphosed Tertiary rocks. Metamorphic core complexes are characterised by extensional mylonite zones, with younger-on-older faults in the allochthon, in contrast to mylonite and faults of the classic older-on-younger thrust and fold belts. Twenty-five such complexes have been identified to date, discontinuously arrayed in a belt along the axis of the North American Cordillera, from northwest Mexico to British Colombia (for an overview, see Coney, 1980).

Spectacular evidence for fluid activity accompanying deformation is present in the ubiquitous metasomatic chloritic breccias developed along detachment faults, and in locally pervasive potassic alteration of the allochthon. Sporadic Cu-Au and Fe, Mn-oxide mineralisation is associated with the chloritic breccias (Crittenden & others, 1980). Light stable isotope studies have been conducted to date on three metamorphic core complexes, the Buckskins complex in Arizona (Kerrich & Rehrig, 1987; Kerrich, 1987), the South Mountain complex of Arizona (Smith & Reynolds, 1985) and the Bitterroot Dome-Sapphire Block detachment zone, western Montana (Kerrich & Hyndman, 1986).

Mylonites

The Buckskins tectonic section features a pronounced upward trend in $\delta^{18}O$ values from 7.6 per mil in the infrastructure to 18.9 per mil in allochthonous Tertiary volcanic rocks, with an attendant decrease of estimated temperature from 550 to 150^{o}C (Fig. 4). Undeformed Oracle Granite (~1.6 Ga) of the lower tectonic plate retains near-magmatic mineral pair fractionations which correspond to isotopic temperatures of 500 to 600^{o}C, that are typical of slowly-cooled plutonic rocks (cf. Taylor, 1968). Mylonitic equivalents have whole rock $\delta^{18}O$ values in compliance with the granite protolith, but mineral-pair fractionations have been reset to 350-500^{o}C, which probably records the ambient temperature of ductile deformation. Mylonites have been interpreted as forming either under closed-system conditions, or in the presence of fluids isotopically buffered by a similar rock reservoir.

Ductile granite mylonites in the Bitterroot Dome formed at ambient temperatures of 550 +/- 50^{o}C in the presence of fluids having $\delta^{18}O$ values of 9 +/- 1 per mil, and likely of magmatic or metamorphic origin. Mylonites developed in metasedimentary rocks 25 km from the granite contact also retain isotopic concordance, which corresponds to temperatures of 450-500^{o}C. Granite and metasediment mylonites formed under conditions of low water/rock ratios as indicated by the isotopic compliance of minerals and whole rocks in mylonites with their undeformed protolith. Smith & Reynolds (1985) reported relatively uniform $\delta^{18}O$ values, and the presence of deep-seated fluids in the South Mountain granodiorite of Arizona.

Chloritic breccias

In the Buckskins core complex breccias are shifted by +2 per mil relative to mylonites or fresh granites: the breccias formed under open-system conditions at temperatures of 300 to 350^{o}C, based on fluid inclusion, isotopic, and mineralogical criteria. Hydrothermal fluids implicated in the Fe, Mg, Mn-metasomatism that produced the chloritic breccias have $\delta^{18}O$ values of 3 +/- 1 per mil (Fig. 4). Veined and hydrothermally-altered rocks of the middle plate record a progressive trend of ^{18}O enrichment up-section, at diminishing temperatures (220-330^{o}C), but the rocks exchanged with isotopically-similar fluids as involved in the lower chloritic breccias.

Granites and mylonite products in the Bitterroot Lobe have tightly clustered $\delta^{18}O$ values, whereas feldspar and biotite have preferentially shifted to low ^{18}O values and record disequilibrium fractionations. This effect is most pronounced in the chloritic breccias which are depleted by ~10 per mil relative to undisturbed mylonitic precursors: the breccias appear to have formed under conditions of lower temperature (250 to 370^{o}C) and effective confining stress, acting as aquifers for the infiltration of low ^{18}O surface waters ($\delta^{18}O$ = -7 to -12 per mil). Criss & Taylor (1983) demonstrated extensive O and H-isotope exchange of the Idaho batholith and Bitterroot Lobe with meteoric water, which occurred following tectonic removal of ~15 km of metasedimentary rocks in the Sapphire Block cover (Kerrich & Hyndman, 1986).

At present, it is not clear as to whether brittle fracturing accompanied by meteoric-water incursion was responsible for development of the chloritic breccias. Alternatively, the chloritic breccias could have been generated in the presence of more ^{18}O enriched fluids as was the case for the Buckskins complex, with subsequent isotopic overprinting by infiltration of meteoric water.

Tertiary alkali basalts in the allochthonous upper plate of the Arizona core complexes are characterised by pervasive hydrothermal alteration to an assemblage of K-feldspar, calcite, Fe, Mn-oxides and anhydrite. $\delta^{18}O$ values of K-feldspar are highly erratic, spanning 6 to 21 per mil. $\delta^{18}O$ and $\delta^{13}C$ values of calcite and K-feldspar are positively correlated, implying that some of the data spread can be accounted-for in terms of temperature variation. Fluid inclusion in quartz indicate the presence of variably saline brines. Potassic alteration of the volcanics appears to have been induced by formation brines stored within the upper plate, and derived largely from evolved meteoric water.

The trends of increasing $\delta^{18}O$ values in conjunction with diminishing temperatures have been interpreted in terms of dual fluid regimes. At lower structural levels there is a switch from closed-system conditions in mylonites and granites to open-system conditions in chloritic breccias, which reflect the expulsion of high temperature, crust-equlibrated and reduced fluids inducing Fe, Mg, Mn metasomatism: these fluids could have been magmatic, metamorphic or evolved formation brines. In contrast, cool, oxidised thermal waters, likely of evolved meteoric origin, were present in the upper plate. Upper plate fluids locally penetrated down to the middle plate, mineralogically and isotopically overprinting the chloritic breccias. Thus the tectonic section in core complexes records an upward transition from high to low temperatures, low to enhanced water/rock ratios, ductile to brittle deformation, and the conjunction of deep aqueous reservoirs at lithostatic pressure with shallower surface reservoirs under hydrostatic conditions.

References

Coney, P.J., 1980 - Cordilleran metamorphic core complexes: an overview. Geological Society of America Memoir, 153, 7-34.

Criss, R.E., & Taylor, H.P., Jr., 1983 - An $^{18}O/^{16}O$ and D/H study of Tertiary hydrothermal systems in the southern half of the Idaho batholith. Geological Society of America Bulletin, 94, 640-663.

Crittenden, M.D., Jr., Coney, P.J., & Davis, G.H., 1980 - Cordilleran metamorphic core complexes. Geological Society of America Memoir, 153.

Kerrich, R., 1987 - Fluid and thermal regimes in metamorphic core complexes. Rundschau Geologische (in press).

Kerrich, R., & Hyndman, D., 1986 - Thermal and fluid regimes in the Bitterroot Lobe-Sapphire Block detachment zone: evidence from $^{18}O/^{16}O$ and geologic relations. Geological Society of America Bulletin, 97, 147-155.

Kerrich, R., & Rehrig, W., 1987 - Fluid motion associated with Tertiary mylonitisation and detachment faulting: $^{18}O/^{16}O$ evidence from the Picacho metamorphic core complex, Arizona. Geology, 15, 58-62.

Smith, B.M., & Reynolds, S.J., 1985 - Oxygen and hydrogen isotope study of mylonitisation and detachment faulting, South Mountains, central Arizona. EOS, American Geophysics Union Trans., 66, 1138.

Taylor, H.P., 1968 - The oxygen isotope geochemistry of igneous rocks. Contrib. Mineral. Petrol., 19, 1-71.

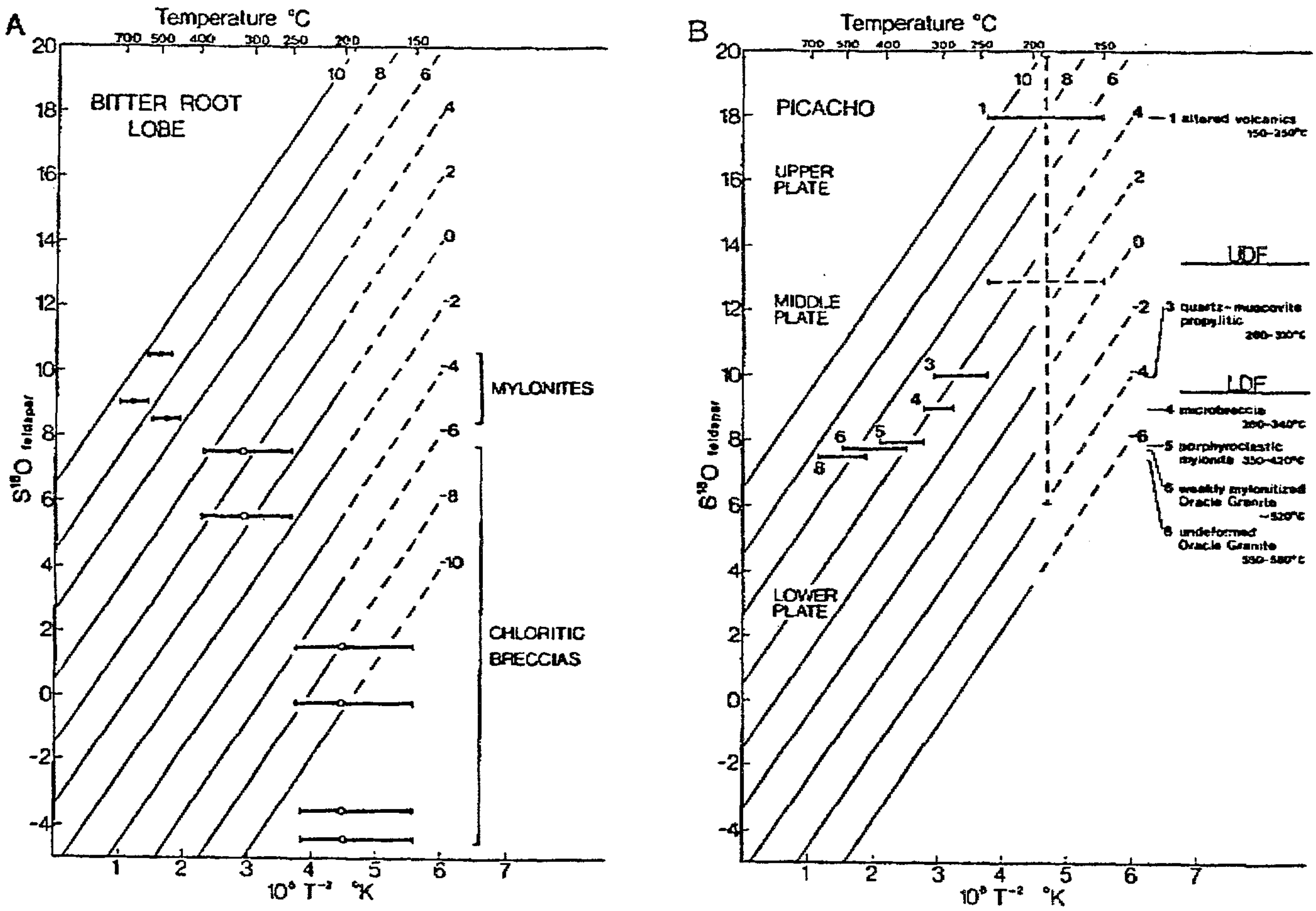

Figure 1. A. $\delta^{18}O$ values of feldspars (albite or Kfeldspar in granites, mylonitic derivatives and chloritic breccias from the Bitterroot Lobe–Sapphire Block detachment, with estimated equilibration temperatures (modified from Kerrich & Hyndman, 1986). The family of diagonal lines are compatible values of $\delta^{18}OH_2O$, temperature, and ^{18}O feldspar, constructed from the feldspar–water calibration of O'Neil & Taylor (1967).

B. $\delta^{18}O$ values of feldspars (albitic plagioclase or K–feldspar) from granites, mylonites, chloritic breccias, and altered Tertiary alkali basalts, from the Buckskins metamorphic core complex, Arizona, with estimated temperatures of equilibration (modified from Kerrich & Rehrig, 1987). The dashed bar corresponds to the range of whole rock $\delta^{18}O$ values and estimated alteration temperature, of altered Tertiary alkali basalts in the suprastructure of core complexes from Arizona (after Kerrich 1987). The family of diagonal lines is as in A.

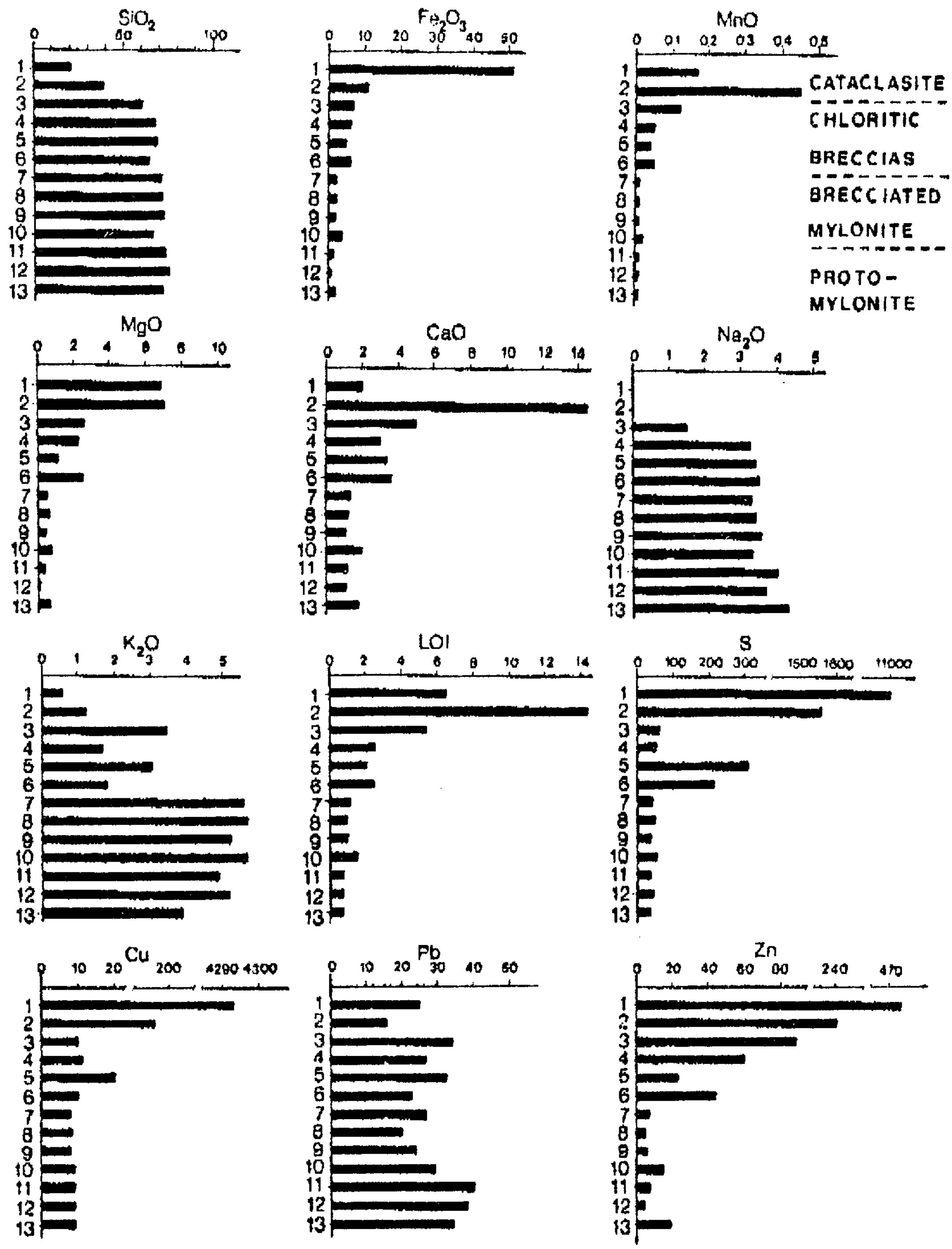

Figure 2. Abundances of selected major element oxides and trace elements, as a function of tectonic position at Copper Penny. Buckskins core complex, Arizona.

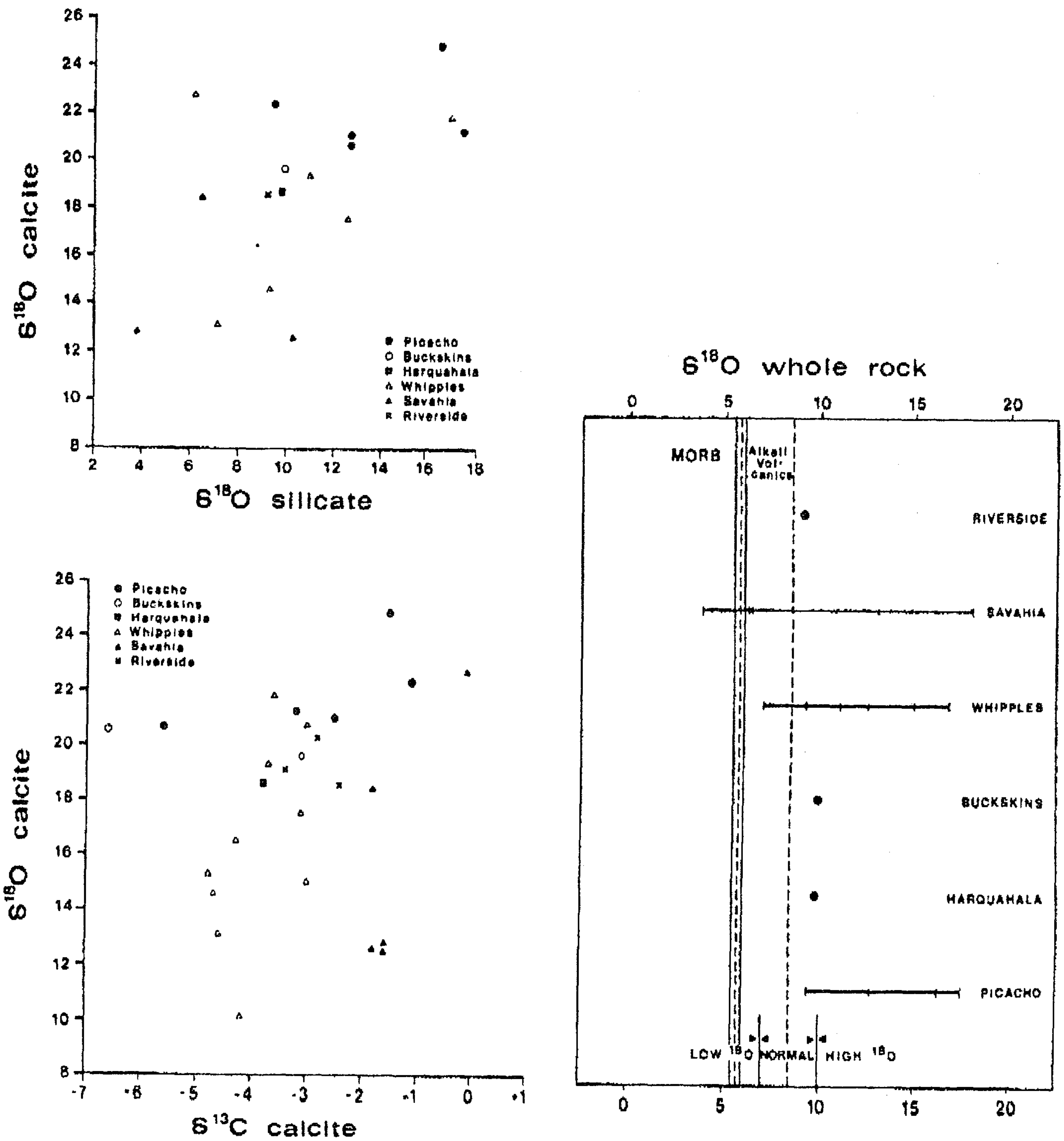

Figure 3. Isotopic data for altered upper plate volcanic rocks from core complexes in Arizona. Plot of $\delta^{18}O$ calcite versus $\delta^{18}O$ whole rock silicate (carbonate-free) [top left]. Plot of $\delta^{18}O$ calcite versus $\delta^{13}C$ calcite [bottom left]. Compilation of oxygen isotope data for carbonate-free whole rocks from specified localities. The solid vertical lines indicate the range of fresh mid-ocean ridge basalts ($\delta^{18}O = 5.7 \pm 0.3$), the dashed vertical line corresponds to the upper limit of $\delta^{18}O$ values for fresh alkaline volcanic rocks (after Kyser, 1986). Low - ^{18}O, normal - ^{18}O, and high - ^{18}O are categories of felsic plutonic rock, based on whole rock $\delta^{18}O$ values given by Taylor (1978) [right].

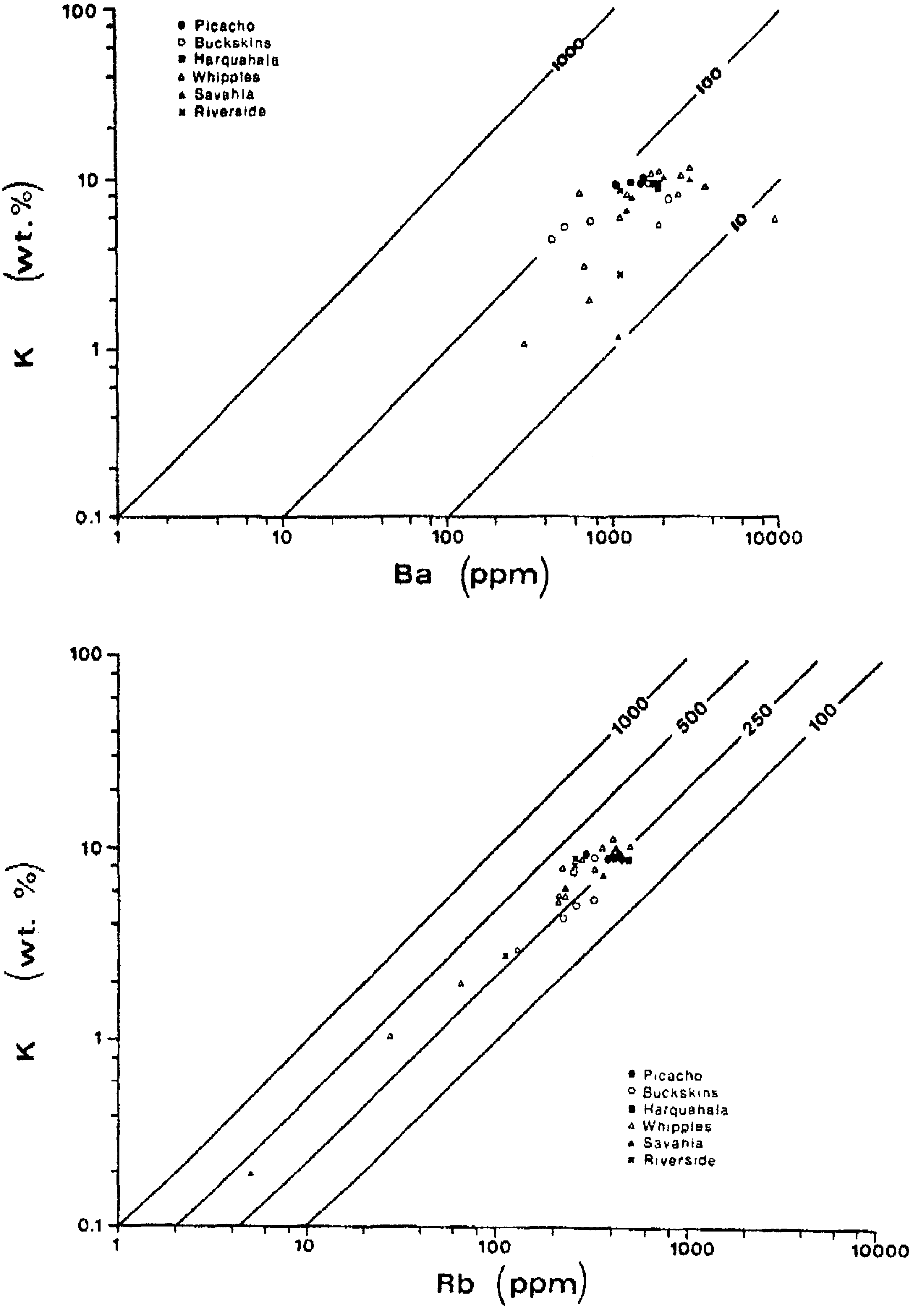

Figure 4. Plot of K versus Ba (A), and K versus Rb (B) for altered upper plate volcanic rocks, Arizona core complexes.

The significance of fluid activity during detachment faulting

G S Lister, Monash University, & H Green, University of California, Davis, USA

In metamorphic core complexes in Arizona and California, there is evidence for significant fluid activity during detachment faulting. The upper plate is extensively metasomatized, with addition of K, Mn and Fe. There is extensive chloritisation of breccias and microbreccias beneath the detachment fault. There is abundant veining with silica, calcite and quartz. This fluid activity is of interest because of two reasons: (a) the possible link between detachment faulting and economic mineralisation; (b) the importance of pressurised fluids to the operation of shallow-dipping faults.

Detachment-fault orebodies have received some attention in the last few years, but there is a complete absence of detailed information concerning their genesis. The orebodies often pre-date detachment faulting. This fact has been used to argue that the orebodies have nothing to do with detachment faulting. However, it has been shown that there is a protracted history of detachment faulting, although everything happens in a remarkably short time frame (<2-3 Ma). It is therefore only logical that orebodies (and any other structure one cares to mention) pre-date the last episode of detachment faulting. In many of the core complexes in Arizona and California, the (portions of) presently visible detachment fault are amongst the youngest structures present in the core complex. The only structures that are younger in the Whipple Mountains are isolated high-angle normal faults. This little-discussed fact has many implications for the mechanism of continental extension. In terms of the origin of so-called "detachment-fault" orebodies, it means that more sophisticated arguments need to be put forward concerning the timing of their formation. For example, it needs to be shown that the orebodies pre-date mylonitic deformation, which in many cases seems to be the earliest recognisable microstructural evidence of the Tertiary ductile deformation which took place when the detachment terranes began their evolution.

It has been suggested that detachment fault orebodies could form as the result of two processes. First, the formation of metamorphic core complexes is associated with intensive volcanic activity, and there are many examples of lower plate granitoid batholiths. These rocks introduce additional heat into the system. They also introduce fluids. Many crystallising batholiths were transected by ductile shear zones. Fluids expelled from the batholith migrate up these shear zones, because of enhanced permeability as the result of dilatancy induced during ductile deformation. These pressured pore-fluids rise along the shear zone, until under certain conditions, the rock ceases to flow in a ductile fashion, and instead is cut by brittle faults. Movements on these faults are kinematically coordinated with those in the ductile shear zones. The faults appear to be part of the same movement zone. In metamorphic core complexes such as the Whipple Mountains, California, or the South Mountains, Arizona, the transition from ductile to brittle behaviour appears to coincide with the arrival of such pressured pore-fluids. These fluids may be the reason for the formation of detachment-fault orebodies. The ore metals may be introduced as the result of mantle-derived igneous activity, or they may have been scavenged by the shear-zone fluid-system, as it was driven along the shear zone by hydraulic gradients induced by variations in the amount of dilatancy during deformation. The Precambrian mafic gneisses may have served as source rocks in this case.

The second process considers the effects of extension of the upper plate, which is heated from below by the introduction of igneous rocks, as well as by the juxtaposition of the relatively hot rocks of the lower plate. The upper plate is cut by normal faults which may have remained open to considerable depths. Ongoing extension results in the maintenance of permeability. The upper plate contains brines, and the fluids are highly oxidising. The fluid-pressure was probably hydrostratically maintained. The shear-zone fluids, emanating from the detachment zone, are reducing, or only slightly oxidising, and they are more highly pressured (close to lithostatic pressure in many cases). The zone where these two fluid systems mix (deep in the detachment system) is a zone where abrupt pressure changes may occur, and marked change in redox potential. It is therefore a zone of economic interest. The question is whether the orebodies found in such a zone are related to the upper plate fluid-system, which has percolated through intensely disrupted and altered volcanic rocks (basaltic, andesitic, dacitic and rhyodacitic), as well as other elements of the stratigraphy, or to the lower plate fluid-system, as described above.

The other question of interest concerns the origin of detachment faults. It is difficult to understand how a shallow-dipping normal fault can originate without intrinsic involvement of lithostatically-pressured fluids. For example, in a delta, Mandl & Crans have described how fluid pressure can abruptly rise from hydrostatic to lithostatic, with depth. This causes an abrupt decrease in strength, and a natural factor localising the formation of shallow-dipping faults. The presence of lithostatically-pressured fluids at the fault plane also has major implications in that it enhances the ability of the upper plate to move with only small shear stresses being applied. In an extending orogen, it is difficult to imagine how a shallow-dipping normal fault at depth would remain active without over-pressured fluid migrating along the fault plane. Otherwise large shear stresses would need to be applied, and unless the stress field has been rotated over large areas, so that σ_1 is no longer vertical, we would otherwise expect steeply-dipping normal faults to form during extension.

These questions concerning the role of fluids led us to examine fault rocks from the Whipple and the South Mountains, using optical and electron microscopy. The study commenced by comparing the so-called microbreccias (ultracataclasites) of the Whipple and South Mountains. There is no comparison. There are two types of microbreccia in the South Mountains. One type (generally grey in appearance) is a hydrothermally 'stewed' cataclasite. Chlorite is felted and equigranular. Quartzite masses have foam-textures, usually found in much higher-grade metamorphic rocks but in this case the result of annealing a cataclasite in the presence of abundant pore-fluids. There has been no deformation since the 'stewing' process got underway. The black microbreccia has had a complex history. Mylonitisation continued until comparatively high stress levels, with recrystallised grain-size decreasing to a few microns. Micro-cracking (in the presence of fluid) then began to dominate the deformation process, with sericite minerals growing in the micro-cracks. Conjugate orthogonal sets of micro-fractures formed, symmetrically inclined at approximately 45° to the mylonitic foliation. Minor faults also formed, subparallel to the foliation. Ultracataclasite bands then began to form, and in zones of ultracataclasite, rhomb shaped blocks have been plucked from the adjacent fracturing mylonite, and rolled in the ultracataclasite.

We can explain these observations if we adopt the following hypothesis. The movements on the detachment fault were episodic, and they took place in the presence of pressured pore-fluid. Episodic fluid migration took place, as indicated by differing La/Ce/Nd ratios in growing and fracturing epidotes. Dilatancy was induced prior to failure, and as the result of failure, macroscopic dilation took place. This dilatancy was relaxed by virtue of its conversion into penetrative micro-dilatancy, subsequent to the failure event. Micro-fracturing took place in a regime of vertical σ_1 as if deformation in the lenses immediately adjacent to the fault was coaxial. The angle between these conjugate micro-fractures is 90°, suggesting that micro-fractures developed without internal friction, such as would be the case for sub-critical crack propagation, below the level of stress at which macroscopic failure would take place. Fluids are sucked into the rock as fractures grow, and phyllosilicates grow parallel to the fracture planes. As deformation goes on, these phyllosilicates are rotated slowly towards the main foliation plane, while new micro-fractures form at higher angles.

Chloritisation, in discrete veinlets, occurred last in the geological history, as can be inferred from these rocks.

In comparison, the microbreccia from the Whipple Mountains had comparatively little of the geological history recorded in the microstructures. There is evidence only of a protracted history of cataclasis. Six generations of structures are evident in the one hand-specimen. Clasts exist within clasts, that exist within clasts. They all comprise randomly-oriented crystallites of quartz, with overgrowths of quartz contaminated by Fe-bearing clay minerals. Red and green microbreccias exist presumably related to the oxidation state during different generations of faulting. These crystallites grew (statically) until impingement, and then themselves were overgrown. Generally, less than five percent impurity is present in the overgrowths, so the cataclasites are essentially cherts. A history of repeated crushing then healing of the fault is indicated. Even at this time, the cataclasites have high permeability, mainly due to the leaching of carbonates. Presumably, an episode of fault activity was heralded by (locally) high fluid-pressures, and the pre-existing cataclasite was crushed until the matrix structure disintegrated. The matrix was highly mobile, and flowed freely into cracks in the adjacent rocks. This suggests it was super-plastic, and that high pore-fluid pressures were maintained during fault movement. Subsequently, induration occurred statically, as fluids flushed through the cataclasite, and crystallites nucleated and grew until impingement occurred. Then overgrowth took place. All this took place under static conditions. This history is of interest because it implies episodic behaviour, and cycles of fluid activity on the detachment system.

Some examples of gold deposits associated with extensional structures in southwest USA

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(No abstract)

Extensional and intrusive-extrusive structural interactions in the late Palæozoic of northeastern Queensland

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Extension of continental crust is commonly accompanied by a broadly cogenetic association of voluminous, predominantly dacitic to rhyolitic, ignimbrites and major passively-emplaced granitoid bodies, plus volcanic and volcano-tectonic subsidence structures (cauldrons, sensu Smith & Bailey, 1968) which might or might not be expressed surficially as physiographic calderas. Indeed, the widespread development of such an association might in itself be diagnostic of extensional tectonism (Sillitoe, 1982). However, while there has been much description and discussion of the stratigraphic structural consequences of extension (and transtension) in terms of sedimentary sequences (Gibbs, 1984; Bosworth, 1985; Harding, Vierbuchen & Christie-Blick, 1985; Lister, Etheridge & Symonds, 1986; Etheridge, 1987), there has been little explicit consideration of such consequences in the granitoid-ignimbrite association.

During late Palæozoic time, northeastern Queensland was a site of significant continental granitoid-ignimbrite magmatism, with concomitant development of subsidence structures which apparently evolved mainly in the subsurface (Branch, 1966; Oversby, Black & Sheraton, 1980). Ignimbrite-dominated extrusive activity might have extended discontinuously into Early Permian time in and adjacent to the Featherbed Volcanics (Fig. 1). In most areas, however, preserved Permian extrusive sequences are more heterogeneous, and contain markedly less voluminous ignimbrites, than their mid- to Late Carboniferous counterparts. Magmatism probably essentially ceased in the mid-Permian, and subsidence structures were apparently being inverted and eroded late in the Period. Parts of some structures also experienced Mesozoic and Cainozoic reactivation.

Overall orientations and alignments of dykes and other intrusive bodies, extrusive centres, and single to composite subsidence structures, suggest that the regional minimum principal stress (σ_3) was horizontal and approximately east-west-trending (with some local deviations) during main-phase granitoid-ignimbrite activity (cf. Nakamura, 1977). Lack of evidence for a region-wide compressional deformation of appropriate age implies that this σ_3 probably represented kinematic extension. The preferred interpretation of tectonic setting envisages the region as a passive continental margin of upper-plate type (Lister, Etheridge & Symonds, 1986) during late Palæozoic time (Oversby, in preparation).

The most conspicuous late Palæozoic structures in northeastern Queensland at current levels of exposure are steep fracture-intrusion systems of variable areal extent and lateral-vertical complexity. These are typically linear, radial, and/or concentric (annular). Well-developed cone-sheet-type systems are uncommon, although many intrusive bodies within extrusive sequences might be incompletely connected to steep counterparts in blind bell-jar-like geometries. These fracture-intrusion systems probably formed mainly in response to vertical stress variations caused by inflation and deflation of periodically over- and under-pressured (evacuated) magma chambers (Anderson, 1936; Koide & Bhattacharji, 1975; cf. also Withjack, 1979), although precise details might be debatable. Similarly, deformation of extrusive sequences, dominated by marginal inward steepening with local development of conical flexures and folds, can be most simply ascribed to cauldron subsidence on overall inwardly-inclined faults.

In contrast, "classical" extensional structures, such as listric normal faults with reverse drag, are not obvious in the late Palæozoic granitoid-ignimbrite association of northeastern Queensland. Asymmetrical (trapdoor) cauldrons are common however, implying that most marginal, and some internal, fracture-intrusion systems (particularly the concentric ones) are listric in part. Possible associated reverse drag of extrusive rocks has so far been detected only in the central and northern Newcastle Range and Bagstowe sequences (Fig. 1). In the central Newcastle Range, this deformation is interpreted from inclinations of eutaxitic foliation in ignimbrite towards apparent ductile structural contacts, suggesting that it took place before rocks had cooled and become brittle. Additionally, in the central Newcastle Range, a gently-dipping younger-on-older contact is interpreted as an early listric fault sole or internal detachment because of steeply discordant eutaxitic foliation in overlying ignimbrite remnants and incompleteness of local stratigraphy. Similar structures might be represented in Featherbed Volcanics (Fig. 1) by large-scale secondary flowage (rheomorphic) folding developed in some ignimbrite sheets (DE Mackenzie, pers. comm.), where there does not seem to be good evidence for initial accumulation on steep palæotopographic slopes.

Rocks affected by the postulated listric fault sole or detachment in the central Newcastle Range are separated from counterparts in an apparently complete stratigraphic sequence by a fault zone which is essentially orthogonal to the dominant northerly trend of other structural features in the area. This zone is probably equivalent to a second-order transfer fault (Etheridge, 1987). Other accommodation zones separating areas of contrasting structural geometry, particularly of marginal fracture-intrusion systems, and/or intrusive-extrusive stratigraphy, can be identified between and within component cauldrons of the first-order Newcastle Range volcano-tectonic subsidence structure (Oversby, Black & Sheraton, 1980). Such zones are commonly expressed as complex belts of fault blocks and slices; andesitic lavas seem to be preferentially localised along many of these belts, suggesting an unusually "leaky" behaviour. Equivalent accommodation zones can be postulated for situations comparable to the Newcastle Range, such as the outcrop area of Featherbed Volcanics. Some zones can be detected in adjacent basement as diffuse intruded and mineralised "lines" (cf. Bain & Withnall, 1980); at least one in the Newcastle Range is apparently shared with an ignimbrite-dominated sequence about 25 km to the west.

Collective and individual relationships in the late Palæozoic granitoid-ignimbrite association of northeastern Queensland suggest a regime of extensional and intrusive-extrusive structural interactions in which the amount and/or rates of magma-associated displacements (Fig. 2b) were so much greater than those of extension (Fig. 2a) that manifestations of the latter process were supressed and/or overprinted (cf. Fig. 2a and 2b) virtually out of existence (cf. Etheridge, 1986). Acceptance of this scenario has some implications for the wider regional context, such as identification of the semi-continuous batholith belt between the northern Newcastle Range and southern Featherbed volcanics, separating predominantly north-younging Newcastle Range stratigraphy and structures from south-younging fracture-intrusion systems of the Warby "ring-dyke complex" (Fig. 1) as being equivalent to a first-order transfer fault; and recognition that some major structures of pre-Late Palæozoic origin, including precursors of the present Gilberton, Burdekin River, and Clarke River Faults (Fig. 1) probably behaved as non-orthogonal accommodation features; and tentative correlation of some repeat distances between intrusive-extrusive features (cf. Vogt, 1974), as well as mismatch between cauldrons and subjacent batholiths suggested at least locally by gravity data (Oversby, Black & Sheraton, 1980; Fig. 6) with extensive mid- and/or upper-crustal detachment. Further investigations of granitoid-ignimbrite associations comparable to the one in northeastern Queensland could well lead to recognition of some special manifestations of extensional tectonics, possibly including pseudo-(structural) resurgence (cf. Fig. 2a), and unexpected spatial/temporal reversals of brittle-ductile response in and adjacent to ignimbrite-dominated sequences, which have not yet been clearly demonstrated.

References

Anderson, E.M., 1936 - The dynamics of the formation of cone-sheets, ring-dykes, and cauldron subsidences. Proceedings of the Royal Society (Edinburgh), 56, 128-157.

Bain, J.H.C., & Withnall, I.W., 1980 - Mineral deposits of the Georgetown region, northeast Queensland. In Henderson, R.A., & Stephenson, P.J., (Editors) - The Geology and Geophysics of Northeastern Australia. Geological Society of Australia, Queensland Division, Brisbane, 129-148.

Bosworth, W., 1985 - Geometry of propagating continental rifts. Nature, 316, 625-627.

Etheridge, M.A., 1986 - On the reactivation of extensional fault systems. Philosophical Transactions of the Royal Society of London, A, 317.

Etheridge, M.A., 1987 - The geometry and tectonic significance of transfer faults in continental extension terranes (abstract). Geological Society of Australia Abstracts, 19, 78-79.

Gibbs, A.D., 1984 - Structural evolution of extensional basin margins. Journal of the Geological Society London, 141, 609-620.

Harding, T.P., Vierbuchen, R.C., & Christie-Blick, N., 1985 - Structural styles, plate-tectonic settings, and hydrocarbon traps of divergent (transtensional) wrench faults. In Biddle, K.T., & Christie-Blick, N., (Editors) - Strike-slip deformation, basin formation and sedimentation. Society of Economic Paleontologists and Mineralogists Special Publication, 37, 51-77.

Koide, H., & Bhattacharji, S., 1975 - Formation of fractures around magmatic intrusions and their role in ore localisation. Economic Geology, 70, 781-799.

Lister, G.S., Etheridge, M.A., & Symonds, P.A., 1986 - Detachment faulting and the evolution of passive continental margins. Geology, 14, 246-250 (and comment by Bosworth, W., and reply, 1986 - Geology, 14, 890-892).

Nakamura, K., 1977 - Volcanoes as indicators of tectonic stress orientation - principal and proposal. Journal of Volcanology and Geothermal Research, 2, 1-16.

Oversby, B.S., Black, L.P., & Sheraton, J.W., 1980 - Late Palæozoic continental volcanism in northeastern Queensland. In Henderson, R.A., & Stephenson, P.J., (Editors) - The Geology and Geophysics of Northeastern

Australia. Geological Society of Australia, Queensland Division, Brisbane, 247-268.

Oversby, B.S., - Late Palæozoic magmatism in northeastern Queensland - incipient rifting at a passive margin of Gondwanaland? (abstract), 9th Australian Geological Convention, Brisbane (in preparation). Geological Society of Australia Abstracts.

Sillitoe, R.H., 1982 - Extensional habitats of rhyolite-hosted massive sulfide deposits. Geology, 10, 403-407.

Vogt, P.R., 1974 - Volcano spacing, fractures and thickness of the lithosphere. Earth and Planetary Science Letters, 21, 235-252.

Withjack, M., 1979 - An analytical model of continental rift patterns. Tectonophysics, 53, 59-81.

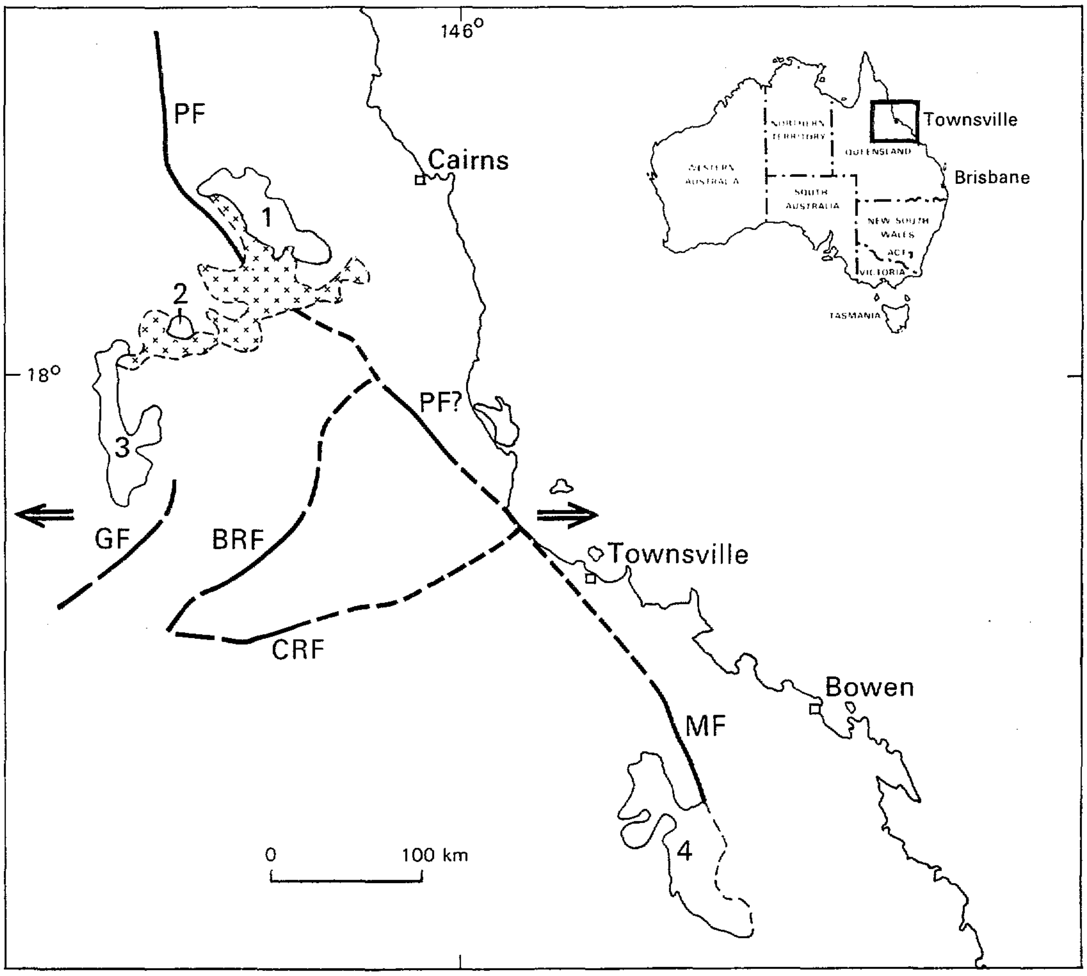

Fig 1: Major late Palaeozoic ignimbrite-dominated sequences in northeastern Queensland. 1 – Featherbed Volcanics; 3 – Newcastle Range Volcanics; 4 – Bulgonunna Volcanics. Other features mentioned in text: 2 – Warby "ring-dyke complex" to the north of granitoid batholith belt between 1 and 3; BRF – Burdekin River Fault; CRF – Clarke River Fault; GF – Gilberton Fault. Palmerville Fault (PF) and Millaroo Fault (MF) also shown for reference. 16/Q/173

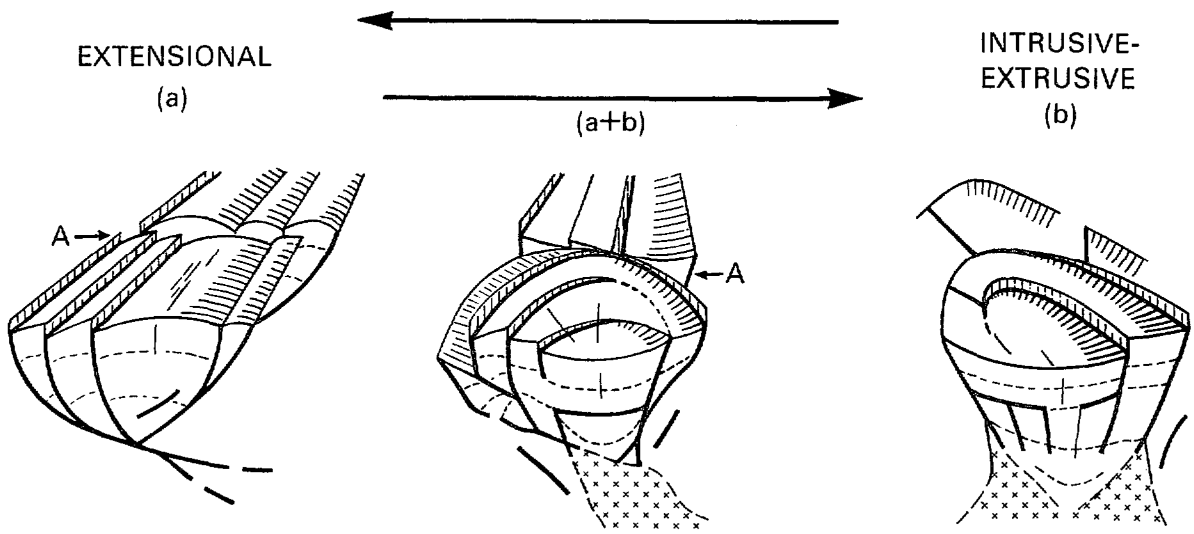

Fig 2: Schematic styles of end-member extensional (a) and intrusive-extrusive (b) structures, and a middle-member composite feature (a+b). Most late Palaeozoic features in northeastern Queensland probably lie between (b) and (a+b). Dominant extensional characteristics in an intrusive-extrusive association, at or close to (a), might be (incorrectly) attributed to structural resurgence. A – accommodation zone. 16/Q/174