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On the Reactivation of Extensional Fault Systems

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On the reactivation of extensional fault systems

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In terranes that have undergone substantial extension, three sets of faults dominate: (*a*) shallow- to steep-dipping, commonly rotational normal faults; (*b*) a really extensive, shallow-dipping, normal detachment faults; and (*c*) steep-dipping transfer faults that strike at high angles to the normal faults. These fault systems may extend through a large fraction of the crust. Reactivation of these fault systems will depend primarily on the relative strengths of the faults (shear zones) and their host rock, and their orientation in the prevailing stress field. It is concluded that reactivation is generally mechanically favoured, but that it will probably only take place when the fault–shear zones are in near-ideal orientations. Consideration of the tectonic setting of extended terranes and of the limited number of well described examples suggests that reverse (thrust) reactivation of the normal and detachment faults and wrench reactivation of transfer faults are the most likely styles. Examples of these styles are described from the Bass Strait Basins of southeastern Australia. Because extended terranes commonly underlie sedimentary basins (for example, on passive continental margins), reactivation of extensional faults may be a key control on the tectonic evolution of such basins (i.e. basin inversion).

1. Introduction

There is widespread evidence for repeated reactivation of ancient fault zones, particularly of the larger zones which extend through a significant fraction of the crust or lithosphere. Sykes (1978) pointed out that intraplate seismicity on the continents is commonly concentrated along ancient fault zones, especially where they are near the terminations of oceanic transform faults. In addition, field geologists have long recognized that the deposition and deformation of cover sequences may be influenced to various extents by reactivation of basement fault zones. Reactivation of fault zones is also predicted by basic mechanical considerations. Fault zones in the brittle régime are likely to have lower cohesive strength and sliding friction than surrounding intact rock (Donath & Cranwell 1981), and in the ductile régime flow strength will commonly be lower in previously active shear zones because of finer grain size or distinctive mineralogy (White 1976; Etheridge & Wilkie 1979).

Despite this widespread support for fault reactivation, it is not universal (Wernicke 1984; Wernicke *et al.* 1985), and should not be assumed simply because it is conceptually appealing. This paper will examine the potential for reactivation of extensional fault systems, with particular emphasis on those fault systems that accommodate substantial stretching of the continental lithosphere (Bally 1981; Wernicke & Burchfiel 1982). Reactivation of these fault systems is potentially important, because they commonly underlie sedimentary basins and passive continental margins that result from the lithospheric stretching, and may therefore control their subsequent tectonic evolution. This paper will first summarize the geometry of extensional fault systems, and then demonstrate via simple mechanical considerations that they have some, but not necessarily a high, potential for reactivation. The surprisingly limited

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published documentation of extensional fault reactivation will be reviewed. Some common styles for reactivation will be illustrated by examples from the Bass Strait Basins, southeastern Australia, where rejuventation of extensional fault systems has been an important control on petroleum accumulation.

2. Geometry of extensional fault systems

Recent studies of extensional fault systems have established a sound geometric framework that is consistent between continental terranes (Proffett 1977; Wernicke 1981; Wernicke & Burchfiel 1982; Davis *et al.* 1983, 1985; Miller *et al.* 1983), passive continental margins (Montadert *et al.* 1979; Le Pichon & Sibuet 1981; Bally 1981, 1982), and some intraplate basins (Gibbs 1984; Harding 1984; Barton & Wood 1984; Etheridge *et al.* 1984*a*, 1985). There are three main fault types in such extensional systems (see figure 1).

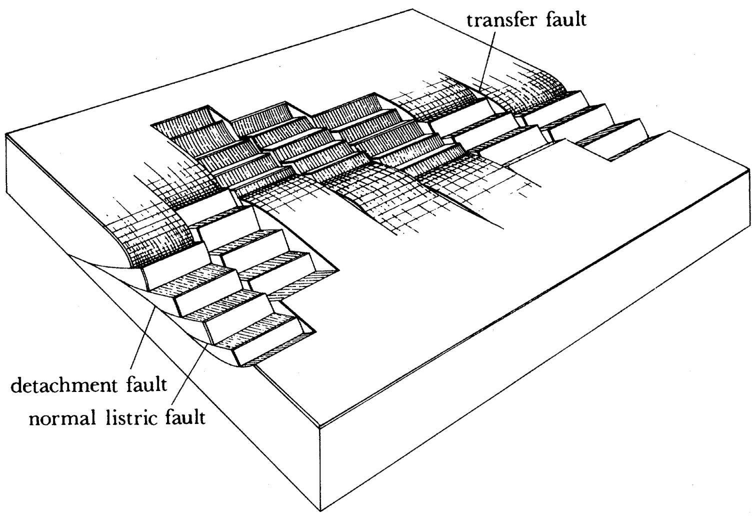

Figure 1. Schematic block diagram showing the main elements of an extensional fault system.

(i) Planar to listric normal faults that are commonly rotational, reaching low dips at large extensions. Several generations of faults may occur in a single extended terrane, and the later will have steeper dips than the earlier faults. In the Basin and Range province, the main range-bounding faults dip quite steeply, and are younger than the shallow-dipping faults that accomplished most of the extension (Proffett 1977). The rotational normal faults occur on a range of scales, from outcrop to more than 10 km in dip and strike extent.

(ii) Transfer faults, which are generally steep to vertical, and strike perpendicular to the normal faults. Transfer faults are analogous to oceanic transform faults, in that they accomodate differences in normal fault geometry along the strike of the extended terrane. They may terminate at any normal fault, but will commonly extend across the terrane, although not beyond its margins. Their depth extent will be the same as that of the normal faults whose geometric variations they accommodate. The largest transfer faults may extend for over 100 km along strike and through a large fraction of the crust, or even into the mantle.

(iii) Very large, essentially irrotational, shallowly dipping detachment faults, which commonly separate a strongly faulted, more brittle upper plate from a more ductile lower plate. Wernicke (1981) and Davis *et al.* (1983, 1985) regarded the detachment fault as a dipping planar dislocation or zone on which the lower plate was dragged out from beneath an extending upper plate. In contrast, Miller *et al.* (1983) considered that a sub-horizontal detachment fault represents the brittle–ductile transition, separating a brittle upper plate from a ductile lower plate, both of which extend in macroscopic pure shear. The Wernicke model is favoured by overprinting relations and sense of shear determinations within and adjacent to the detachment fault zone and is preferred here (figure 1). Detachment faults commonly have areas in excess of 10000 km^2, and may branch into multiple fault systems, especially at shallow levels (Lister & Davis 1985).

It is suggested that these three components (normal, transfer and detachment faults) are present in most extensional fault systems, and the resultant complex fault geometry (figure 1) will have a substantial effect on the potential for, and style of, reactivation in a range of tectonic settings.

3. Mechanics of fault reactivation

A detailed discussion of the mechanical principles of fault reactivation is beyond the scope of this paper. However, the following points provide background for subsequent discussion.

In the brittle régime, the potential for reactivation can be assessed in simple terms by comparing the Coulomb–Navier parameters cohesion and coefficient of internal friction between the fault zone and the surrounding rock mass. If the fault zone contains significant breccia or fault gouge, its cohesive strength (S on figure 2) will be significantly less than that of the intact rock mass, and reactivation is favoured. However, if earlier movement has exhumed a previously ductile shear zone, the fault rock may be relatively intact and have high cohesive strength. Hydrothermal alteration within a fault zone, especially if substantial silicification results, could also increase the cohesive strength of the zone, even to the point where it exceeds that of the surrounding rock. A specific case of the last phenomenon may be where an early fault zone in a slate sequence is the locus for extensive quartz veining or silicification, thereby increasing the strength of the zone. The coefficient of internal friction does not vary widely from one rock type to another, unless specific clay minerals are present in the fault zone (Byerlee 1968, 1978). Therefore, a pre-existing fault zone is unlikely to have a substantially different resistance to sliding from that of a new zone in the surrounding rock. However, if mineralogical alteration accompanied the earlier faulting, a significant change in internal friction is possible. Once again, weakening is expected, but an increase in friction is possible; for example, in a silicified fault zone within shales or acid volcanics. This range of mechanical behaviour in the brittle régime is summarized on a Mohr diagram (figure 2), illustrating the general tendency for frictional failure envelopes within fault zones to be at lower shear stresses than in intact rock. However, figure 2 also demonstrates that the fault zone can be stronger, and that reversals in strength between fault zones and intact rock can take place with increasing depth within the brittle régime. Brittle fault zones therefore generally have lower shear strengths than their host rock, but in many cases strength differences may be small.

In the ductile régime, faults give way to shear or mylonite zones in which deformation is governed by flow laws of quite different form to the Coulomb–Navier behaviour in the brittle régime (Goetze & Evans 1979; Sibson 1982; Chen & Molnar 1983). Under these conditions,

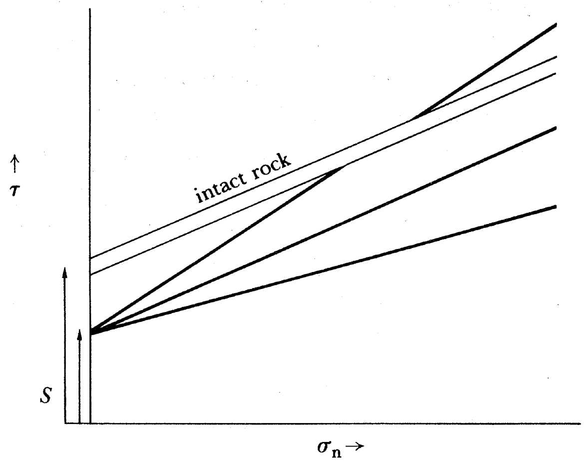

FIGURE 2. Mohr diagram for stress, with idealized Coulomb–Navier failure envelopes for intact rock (open lines), and for fault zones (solid lines) with a range of mechanical properties, showing the effect on shear strength of variations in cohesive strength (S) and coefficient of internal friction (u).

the relative shear strengths of the shear zone and the surrounding rock are governed largely by grain size, mineralogy, and structure (such as foliation orientation, crystallographic fabric) (White 1976; Etheridge & Wilkie 1979). The presence of finer grained, commonly retrogressed, highly foliated rocks in shear zones therefore favours reactivation in the ductile régime, but there are no data which enable the strength to be quantified. Ductile shear zones are also the locus of extensive fluid flux, and may therefore deform by mass transfer mechanisms at lower shear stress than the surrounding rock (Beach 1976; Etheridge *et al.* 1984*b*).

Through the brittle–ductile transition, an overlap of frictional and creep mechanisms will occur. The effects of this overlap will be complex, but are likely to do little more than smooth out the transition in behaviour. A significant effect may arise if a highly foliated ductile shear zone is exhumed to the brittle region, because the friction coefficient parallel to the foliation of the shear zone is likely to be significantly lower than in the surrounding rocks (Donath 1961).

In summary, simple mechanical considerations support the widely held view that reactivation of fault (shear) zones will commonly be favoured over new fault generation, provided that the zone is in a high shear-stress orientation. The role of fault orientation in the stress field will be discussed in the next section. It should be emphasized, however, that the difference in shear strength fault or shear zones and intact rock may be small. Hardening of fault zones may even result from extensive hydrothermal alteration within the zone, effectively reducing the tendency for reactivation. In fact, fluids within fault zones may well be the single most important factor influencing their tendency to be reactivated. High fluid pressures may substantially reduce sliding friction in the brittle régime, as will hydrothermal alteration to phyllosilicate-rich assemblages. In the ductile régime, enhanced reaction and mass-transfer rates, and hydrolytic weakening are all likely to reduce shear strength (Etheridge *et al.* 1984*b*).

4. Dynamics of extensional fault reactivation

In addition to the relative strengths of the fault (shear) zone and surrounding intact rock, the potential for reactivation is governed by the orientation of the zone in the local stress field. Donath & Cranwell (1981) derived the probabilities of new fault nucleation and reactivation of an existing fault in terms of cohesion, coefficient of friction, principal stress magnitudes and orientation of the fault in the stress field. Figure 3 shows the conditions for nucleation against reactivation for a single set of these parameters. Reactivation will take place in preference to nucleation if the pre-existing fault orientation lies in the shaded segment of the Mohr circle. Because of the number of parameters that control the size of this segment, it is difficult to present a simple quantitative picture of the likelihood for reactivation. However, the orientation range for reactivation diminishes rapidly at low strength differences between the zone and surrounding rock (figure 3).

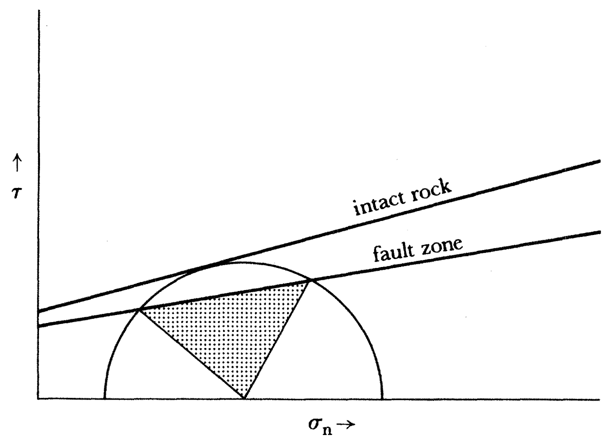

Figure 3. Mohr diagram illustrating the orientation range (shaded region) over which reactivation takes place in preference to nucleation of a new shear fracture in the surrounding intact rock.

Figure 4*a*,*b*,*c* qualitatively summarizes the potential for reactivation of various extensional fault sets in a range of common earth stress orientations (i.e. principal stresses horizontal and vertical). In simple terms, there is a strong tendency for low-angle normal (including detachment) faults to be reactivated as thrusts during subsequent horizontal compression at a high angle to the strike of the faults. Normal fault systems formed by substantial extension may have virtually identical geometries to thrust fault systems (Gibbs 1984; Boyer & Elliott 1982). Reactivation of extensional fault systems as thrusts and vice versa is therefore conceptually appealing, although well documented cases are rare. The other styles of reactivation that should be widespread are (i) normal faults, especially those with moderate to steep dips, with later normal movement (figure 4*a*), and (ii) transfer faults reactivated as wrench faults (figure 4*c*). Examples of these from the Bass Strait Basins will be shown later.

The simple schematic picture shown in figure 4 does not take account of oblique-slip reactivation. However, figure 4*c* does illustrate how transfer faults may play a significant role in suppressing oblique or strike movement on abutting normal faults, owing to the requirement for propagation of the fault into solid rock across the transfer fault.

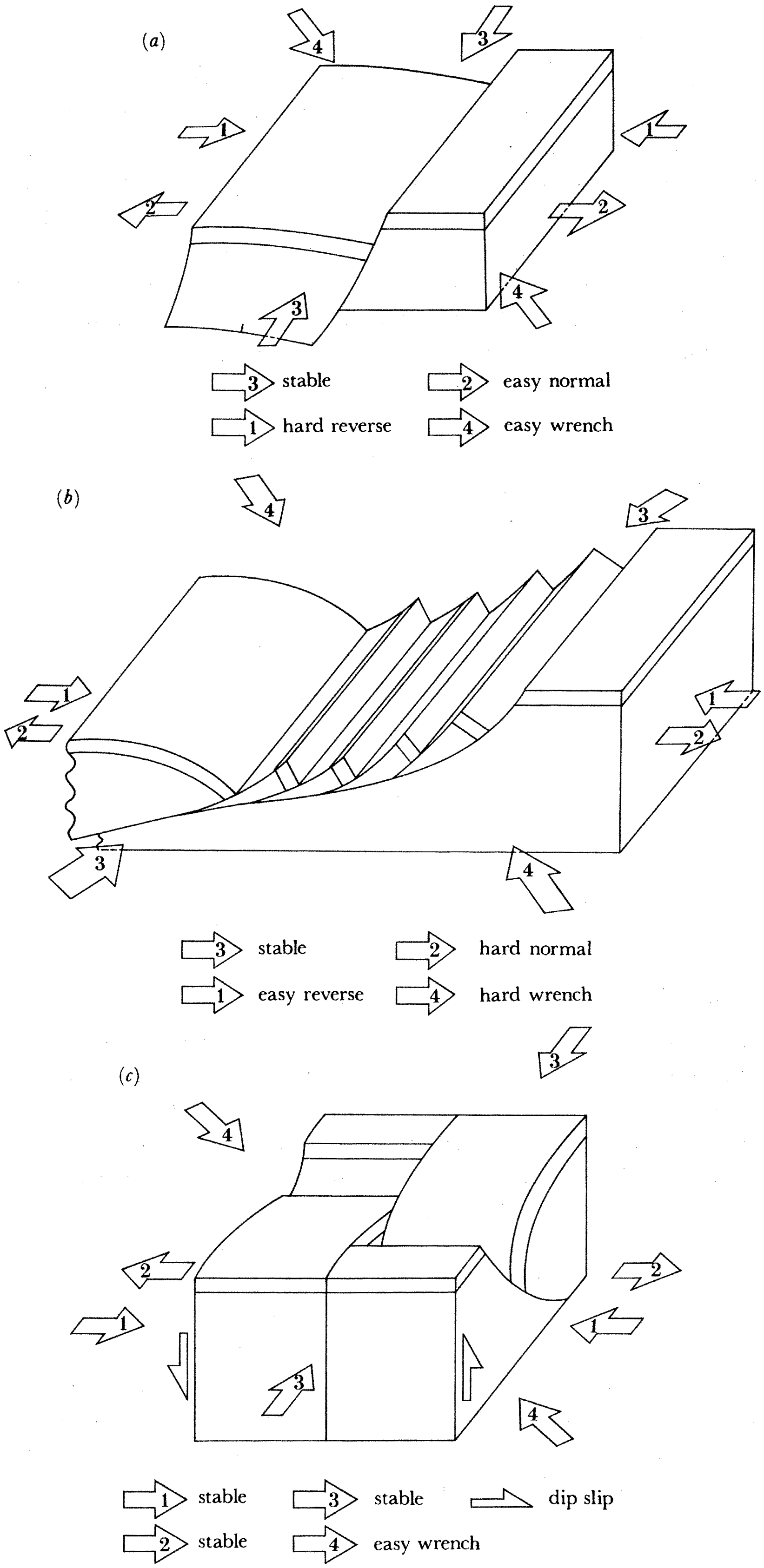

FIGURE 4. Block diagrams illustrating the role of principal stress orientation on reactivation potential for various extensional fault geometries. (*a*) Steep normal faults, (*b*) shallow rotational normal faults listric to a detachment fault, and (*c*) transfer fault separating oppositely dipping normal faults.

5. Discussion

(a) *Examples of extensional fault reactivation*

Fault reactivation is a very widely described phenomenon in the geological literature. For example, the reactivation of basement faults during both deposition and deformation of a cover sequence is virtually a tenet of continental tectonics. The lithospheric stretching model for the evolution of passive continental margins (McKenzie 1978; Le Pichon & Sibuet 1981; Dewey 1982) provides a mechanism for inducing widespread extensional fault systems in both basement and its sedimentary cover, which are available for reactivation during subsequent compressional orogeny. Dewey (1982, p. 399) has recently integrated the stretching model into an overall tectonic scheme for basin development and orogeny, and concluded that 'listric normal faults are convenient zones of weakness in stretched crust for thrust nucleation'. With this broad acceptance and the conceptual appeal of the phenomenon of extensional fault reactivation, it is surprising how few well documented examples have been published. In this section, some of the more recent and widely quoted cases will be reviewed and some new examples described, to assess both the true extent of the phenomenon and the criteria for recognition of reactivated extensional faults.

In strongly deformed terranes, the amount of strain and the structural complexity generally prevent recognition of early extensional structures. In less deformed regions, however, a number of reasonably well documented examples have been described. Stonely (1982) illustrated Mesozoic normal growth faults in the Wessex Basin of southern England that were reactivated as reverse faults during Cainozoic shortening. In this case, the amount of shortening is small (less than 10%), insufficient to completely reverse the original normal movement. The reverse movement is recognized from faults and monoclines in the Tertiary cover, and from the accentuated rollover structures adjacent to the faults. Published seismic sections that convincingly document reverse reactivation of normal faults are rare. Two excellent examples are from the Ronne Graben and the South Sumatra Basin. In the Ronne Graben, major basin-margin normal faults with throws of several kilometres have undergone hundreds of metres of reverse reactivation related to the Alpine Orogeny (Ziegler 1983). The South Sumatra Basin example is particularly convincing, with the original normal movement defined by both the displacement of the basement surface and rotation of basement and syn-rift sediments (Harding 1983). The reverse movement is defined by displacements of the post-rift sedimentary blanket. However, other examples of purported reverse reactivation or inversion of extensional faults given by Ziegler (1983), Plawman (1983) and Davis (1983) are less obvious, primarily because the original normal character of the faults has not been adequately demonstrated. The reverse faults in the Gippsland Basin illustrated by Davis are analogous to those shown and discussed below, but the work of Etheridge *et al.* (1985) suggests that Davis's normal faults are minor structures related to subsidence rather than primary extensional faults (compare figures 3.3–3.22 of Davis (1983) with figure 8 here).

Despite the paucity of well documented published examples of reverse reactivation of normal faults, it is clear that it is the single most important mechanism of basin inversion. Its importance in the inversion of many of the Mesozoic to Cainozoic basins of northwestern Europe was highlighted in a number of the papers presented in a symposium on '*Inverted Mesozoic basins in the Alpine Foreland*' (Ziegler 1985).

Perhaps the most widely quoted recent example is from the Zagros fold belt, where focal mechanism studies of crustal earthquakes overwhelmingly indicate reverse movement on basement faults that dip between 40° and 50° (Jackson 1980; Jackson *et al.* 1981). Jackson (1980) argued that the steepness of the reverse faults is a result of their origin as listric normal faults formed during the extension that accompanied development of an earlier passive margin in the area. Shortening of a previously stretched crust also accounts for the present thin to normal crustal thicknesses. It should be noted that no direct evidence was presented for the existence of early normal faults.

Winslow (1981) argued for reverse reactivation of normal faults in the southern Andes, primarily to overcome compatibility problems at the basement–cover interface in palinspastically restored cross sections. The apparent shortening in the basement was about 40% less than in the cover. Winslow explained this discrepency by appealing to 40% extension on normal faults in the basement before deposition of the cover sequence. Independent confirmation of this early faulting came from thickness variations of the lower cover units from one fault block to the next, implying an early horst and graben structure (figures 8 and 9 of Winslow 1982). Butler & Coward (1984) argued that their palinspastic sections across the Moine Thrust Complex contained no pre-existing extensional structures, despite the absence of evidence for substantial crustal thickening. However, it is possible that their sections contain only that part of the sequence deposited during post-stretching subsidence, and that reactivated extensional faults exist deeper in the section.

On a larger scale, major transfer and detachment faults may also be reactivated. Craig *et al.* (1984) described lineament patterns of various ages in the Canning Basin, Western Australia, and concluded that (p. 68) '...NW–SE transfer faults of the Palaeozoic stage operated as extensional faults during the Mesozoic. The NE–SW Palaeozoic extensional faults were reactivated as Mesozoic transfer faults.' It is not clear how originally shallow-dipping normal faults could be regenerated as transfer faults, which must dip steeply to avoid significant space problems, and vice versa. Again, no specific example of movement reversal on a fault is described. Detachment faults are ideally oriented for reactivation as major thrusts, but the concept of detachment faulting is relatively new (Wernicke 1981), and I am unaware of any documented examples of such reactivation.

(*b*) *Examples from the Bass Strait Basins, southeastern Australia*

In 1982, the Australian Bureau of Mineral Resources (B.M.R.) contracted a 3200 line-km reflection seismic survey in the Bass Strait region between mainland Australia and Tasmania. The survey was centred on the Bass Basin, but lines were also run across parts of the Otway Basin and the hydrocarbon-rich Gippsland Basin. The examples given here come from these last two basins, both from the B.M.R. and industry data. Structural interpretations of the data are given by Etheridge *et al.* (1984*a*, 1985) and Williamson *et al.* (1985). In summary, all three basins were initiated by early Cretaceous extension of about 50–80%, followed by thermal subsidence and, for the Otway Basin, continental break-up. The extensional structures are only well displayed near the basin margins, but major normal and transfer fault arrays have been mapped in the Bass and Gippsland Basins (figure 5).

In the Otway Basin, early Cretaceous extension was followed by further extension and subsidence associated with break-up in the late Cretaceous and early Tertiary. The protracted history of extension and subsidence led to normal reactivation of early normal faults up through the section (figure 6). This type of reactivation is distinguished from simple growth faulting

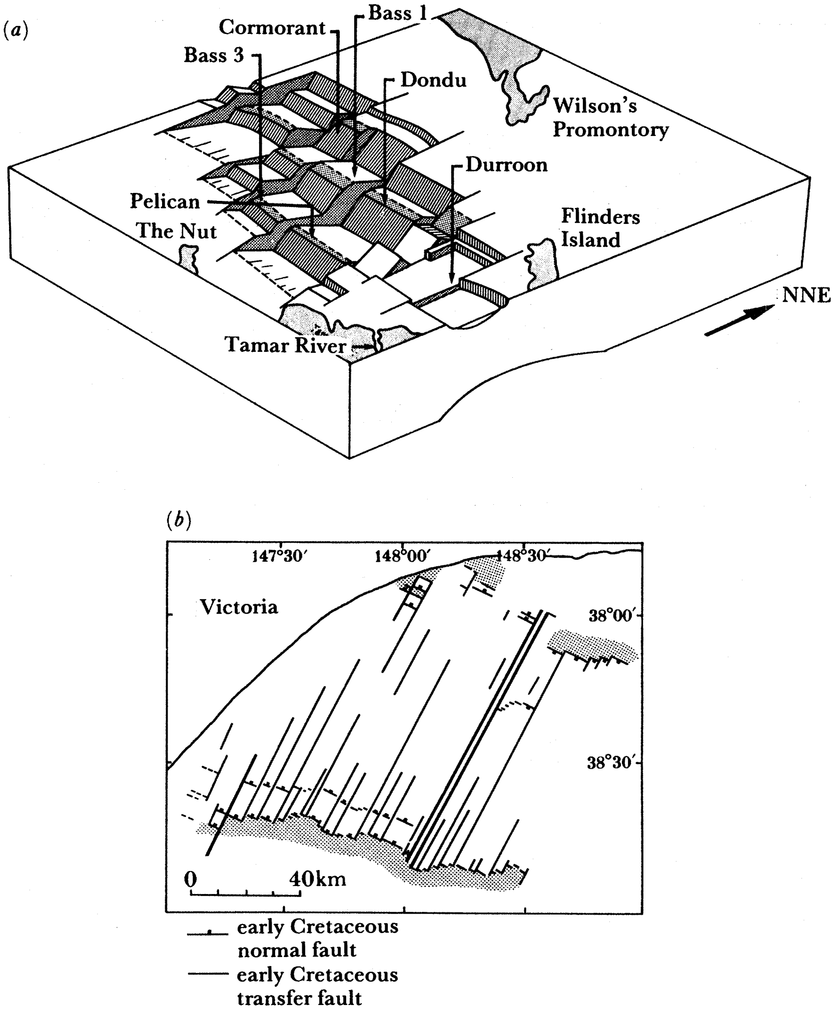

FIGURE 5. (*a*) Schematic block diagram of the extensional structures in the Bass Basin. Note the major transfer fault zones that extend across the whole basin and to depths of at least 10 km. (*b*) Map of extensional structures in the Gippsland Basin; the structures could generally only be mapped along the basin margins, but note the similarity in structural style to the Bass Basin.

by the presence of angular unconformities which represent substantial breaks in fault movement. Reactivation of normal faults during later extension will be limited by the amount of earlier rotation of the normal faults. Proffett (1977) showed that extreme rotation led to the deactivation of such faults, and the generation of a new, steeper set, which continued to rotate. Reactivation of shallow-dipping normal faults during thermal subsidence is also limited, because the fault geometry will not allow significant vertical motion without further extension.

In the Gippsland Basin, major normal faults with throws of over 1 km have been mapped along the basin margins. In the basin centre, the rift structures are too deep (greater than 4 s T.W.T.T.), and cannot be resolved on the available seismic data. The normal faults have short strike extents, with (usually) small offsets on steep-dipping orthogonal transfer faults (figure 5*b*). One transfer fault, which has been mapped on both margins and therefore extrapolated with some confidence across the basin centre, divides the basin into an eastern half in which

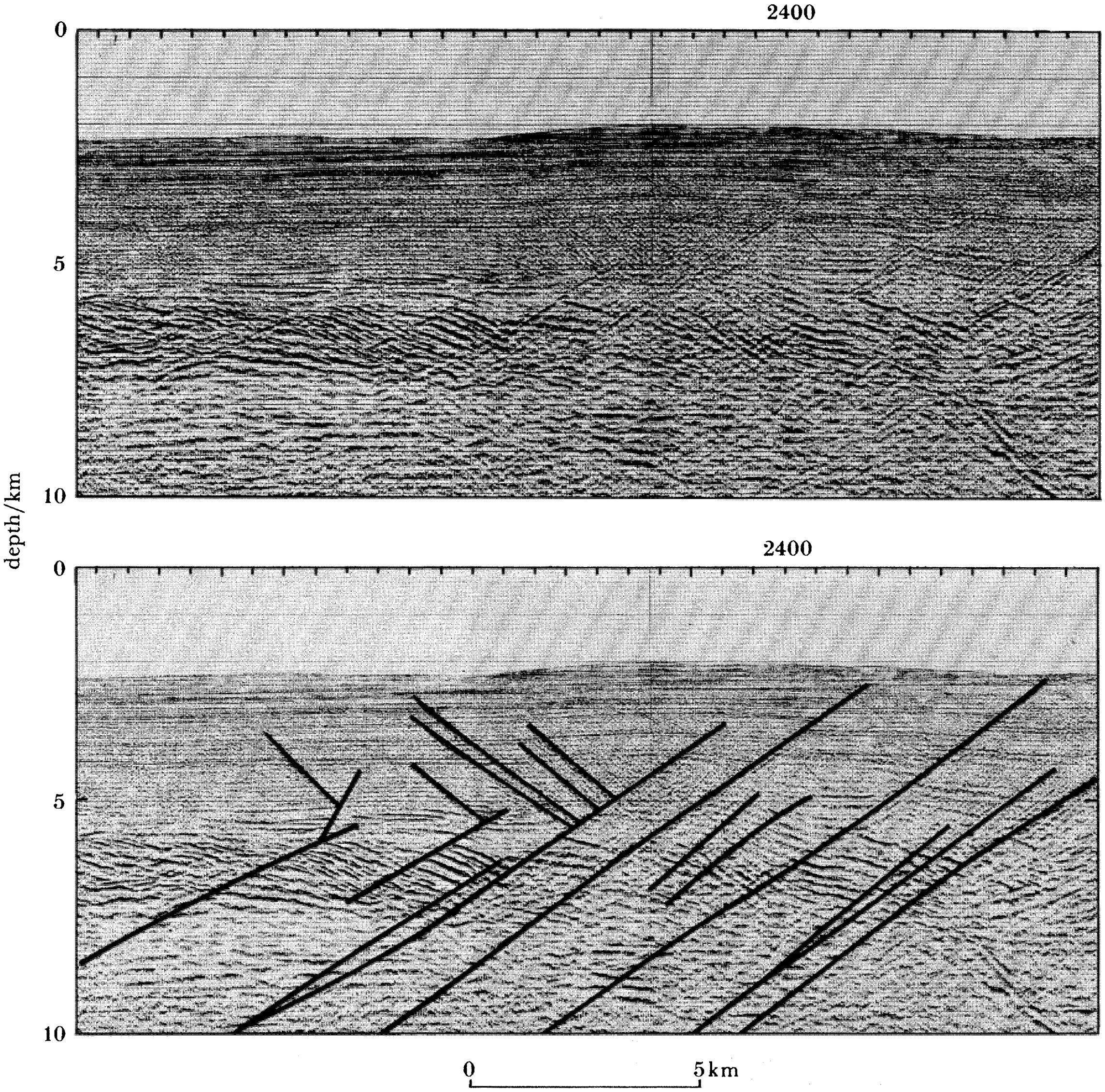

FIGURE 6. Seismic depth section of part of continental margin line 22 of the B.M.R. seismic survey (Etheridge *et al.* 1984*b*). Vertical exaggeration multiplied by factor of two.

all known extensional normal faults dip south southwest, and a western half in which the faults dip north northeast (figure 5*b*). Another major transfer fault is inferred near the current western shoreline (figure 5*b*). Fracture patterns in the onshore parts of the basin (Barton 1981), and stress measurements and fault plane solutions in the region (Denham *et al.* 1981; Gibson *et al.* 1981) indicate an approximately northwest–southeast compression in the late Tertiary to Recent. Reactivation of the extensional fault array of figure 5*b* in this stress field is shown schematically in figure 7. Normal faults should be reactivated as reverse faults, and the transfer faults should undergo wrench reactivation to give rise to anticlines, monoclines and flower structures (figure 7). In addition, the very shallow water depth in the basin throughout the period of reactivation provides the opportunity for the extensional fault patterns to influence subsidence-phase erosion and sedimentation.

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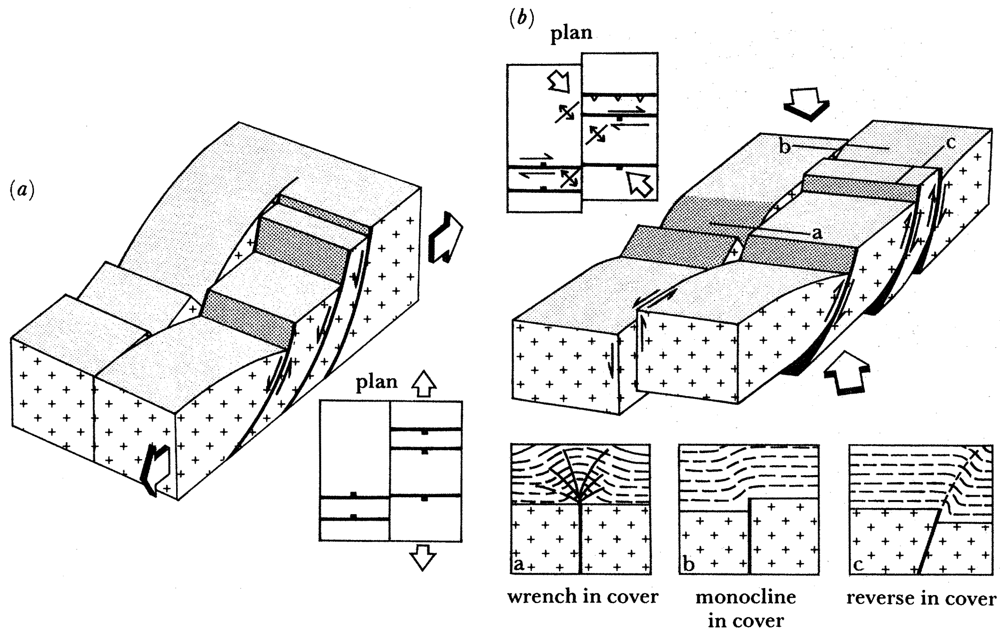

FIGURE 7. Schematic block diagrams of (*a*) typical normal and transfer fault geometry in Gippsland Basin, and (*b*) reactivation of these faults under the influence of a horizontal northwest–southeast maximum principal compression.

Later Tertiary to Recent compressive or wrench structures are present throughout the Gippsland Basin, and they contain the bulk of the known hydrocarbon reserves. The hydrocarbon-bearing structures are mainly northeast to east-trending anticlinal, reverse fault or unconformity traps, which have generally been related to an east-trending, right-lateral wrench zone along the northern margin of the basin (Threlfall *et al.* 1976). However, Etheridge *et al.* (1985) have suggested that most, if not all, of these structures were strongly influenced by reactivation of the early Cretaceous extensional faults.

Evidence for reactivation comes from both direct observation of movement reversal in seismic section and the spatial association of older and younger structures. Figure 8 shows an early Cretaceous normal fault that has been reactivated by reverse movement. The fault is clearly originally normal, because it separates early Cretaceous rift-fill, which dips towards the fault, from pre-Mesozoic basement. The reverse reactivation is best resolved in the overlying Tertiary sequence, and is only a small fraction of the original normal movement. This is presumably the style of reactivation envisaged by Jackson (1980) in the Zagros fold belt. In most of the Gippsland Basin, particularly beneath the hydrocarbon fields, the extensional structures are too deep to be resolved on the seismic data available. However, comparison of the locations of the hydrocarbon fields and some dry structures with the extensional fault array (figure 9) supports a clear correlation between the two sets of structures. In particular, structures and fields are aligned along the two major transfer fault zones described above, as well as along the northern basin margin. Even along the northern margin, there is evidence for reactivation of the bounding normal faults (figure 8), rather than formation of a new wrench zone. Because the amplitude and number of the younger structures decreases from north to south across the basin, the magnitude of the reactivating shear stress components must have decreased from north to south, either by reduction of deviatoric stress magnitudes or by rotation of principal stress

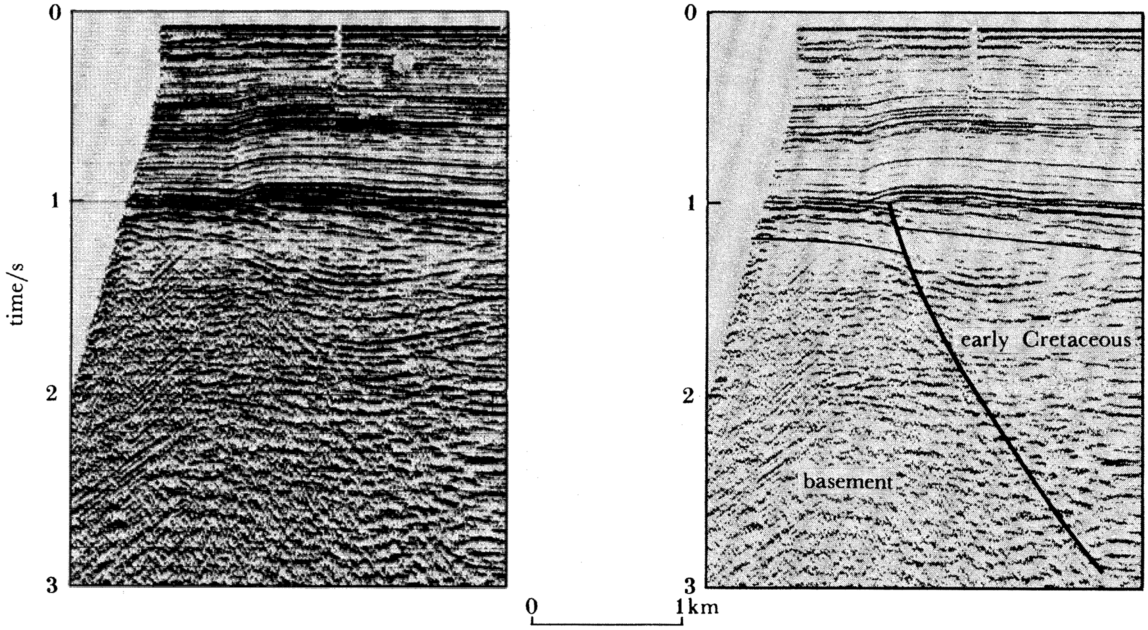

FIGURE 8. Seismic time-section from northeastern Gippsland Basin, illustrating reverse reactivation of a major early normal fault. Vertical exaggeration of *ca.* 1.5 at 3 s T.W.T.T.

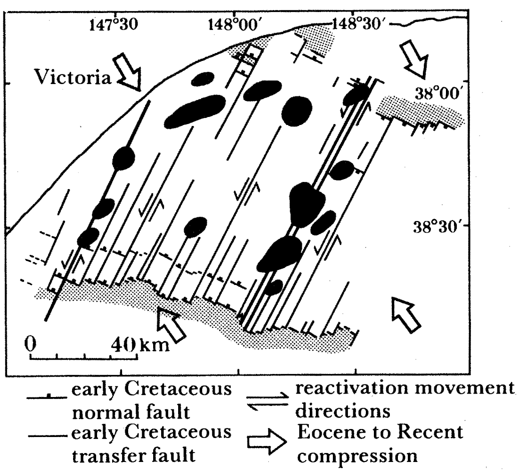

FIGURE 9. Hydrocarbon fields (mostly in late fold or reverse fault structures) and the larger dry structures superimposed on the Gippsland Basin extensional fault map. Note the association of the late structures with two major transfer fault zones.

trajectories. There is also some evidence that minor reactivation, especially of the transfer faults, continued to the present day in all three basins, giving rise to subtle facies variations in the overlying sediments and even influencing seabed erosion (figure 10).

(*c*) *Criteria for recognizing reactivation*

Until recently, extensional fault systems were thought to be dominated by steep-dipping normal faults. Their steep dip significantly reduced the potential for reactivation, especially

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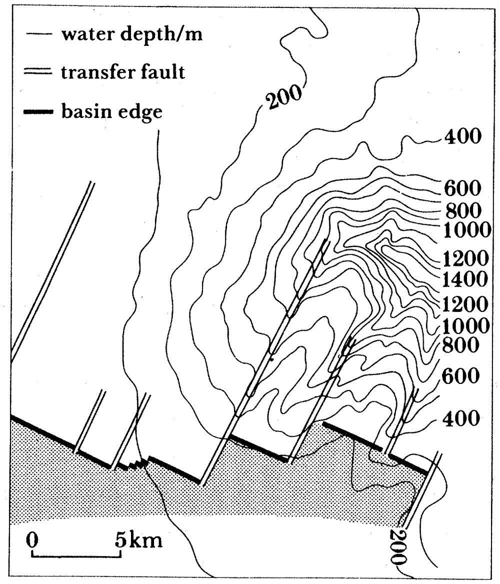

FIGURE 10. Bathymetric contour map of part of southeastern Gippsland Basin overlaid on transfer fault traces mapped at Middle Cretaceous unconformity. Note the preferential seafloor erosion along transfer fault trends.

during subsequent shortening. The elucidation of extensional fault systems that are geometrically analogous to thrust systems (Wernicke 1981; Gibbs 1984) substantially increases the potential for reactivation in a subsequent compressional orogen. However, recognition of extensional fault reactivation is likely to be difficult, especially where they have been overprinted by moderate to large strains.

The clearest criterion for recognition of reactivation is the differential offset of markers that respectively pre- and postdate the original extension. Because extension is commonly accompanied by high-level igneous activity, intrusive and extrusive bodies emplaced between displacement events may provide excellent markers. In the Basin and Range province, for example, separate extensional episodes are clearly marked by pyroclastic layers that are deposited unconformably on tilted fault blocks, and then subject to a later episode of faulting (Proffett 1977). As McKenzie (1978) pointed out, extension is commonly followed by thermal subsidence, and the rift unconformity between the extensional and thermal sag sequences (Gibbs 1984; Etheridge *et al.* 1985) represents an excellent marker. Reactivation of the extensional structures will be clearly recorded by displacements of the unconformity and the sequence above it (figure 8).

Where later deformation is large, these displacement criteria will be harder to recognize, and the construction of balanced cross sections is likely to be necessary to establish the earlier extension and reactivation. In this case, there will be an apparent difference in the shortening of the extended sequence and the overlying rocks (Winslow 1981). However, given the structural complexity of substantially extended terranes (Lister & Davis 1985), construction of sufficiently accurate balanced sections in those that have been subsequently deformed will be very difficult.

At deeper levels, where ductile shear zones predominate, criteria exist for determining the sense of shear in the zones (Simpson & Schmid 1983; Lister & Snoke 1984). In this way, the normal or reverse character, or sense of strike-slip of a zone can be determined. It may also be possible to recognize reactivation if one sense of shear is restricted to sub-zones that consistently overprint a separate set of sub-zones with the opposite sense of shear. However, a large number of observations must be made, because heterogeneous simple shear is the rule in many zones, and some apparent sense reversals are to be expected within zones that have undergone macroscopic shearing of only one sense.

Thrust reactivation of major detachment faults is a particularly appealing concept, because of both their orientation and their large depth and strike extents, and it may seem relatively easy to establish, Indeed, if the detachment fault is a single, planar dislocation, as envisaged by Wernicke (1981), reverse reactivation may be readily documented by careful analysis of movement history and shear sense. However, the evolving shear-zone model of Lister & Davis (1985) predicts a much more complex detachment fault geometry, upon which superimposed reverse movement is likely to be more difficult to recognize.

Transfer faults will generally undergo wrench reactivation, which will not be easy to determine within the extended terrane itself especially because it has a 50% chance of being in the same sense as the original displacement. However, wrench reactivation will give rise to characteristic structures in an overlying (thermal sag) sequence. In an extensional basin, the occurrence within the sag sequence of wrench structures (en echelon anticlines, flower structures) aligned across the basin may be diagnostic of this style of reactivation (figure 9).

In substantially extended terranes, major transfer faults or fault zones are likely to persist across the whole terrane and down through a significant fraction of the crust, or even into the upper mantle. They therefore represent excellent candidates for persistent and long-lived reactivation. Because most transfer faults are steep-dipping and close to planar, reactivation will result in long, narrow, straight features that may be recognized as classical lineaments with long histories of movement. Even where an extended terrane has undergone subsequent orogeny, transfer faults may be available for reactivation provided that they extended below the depth of significant disruption. In this case, the lineaments will tend to be at high angles to the depositional and structural grain. Recent reactivation of such long-lived structures will be mainly recognizable in the topography, with preferential erosion of fractured rock within the zone. Ancient reactivation may be fingerprinted by a number of features, including sedimentary facies boundaries, zones of repeated deformation and a concentration of igneous activity. It is suggested that reactivation of transfer faults which originated during lithospheric stretching may be a major contributor to the generation of long-lived lineaments.

The increasing recognition of extensional fault systems beneath passive continental margins has substantial implications for reactivation, because passive margins commonly become involved in a compressional orogen at a convergent margin later in the Wilson cycle. Dewey (1982) has coupled the concepts of lithospheric stretching and shortening in this way, and alluded to the potential importance of extensional fault reactivation during subsequent compression. In the passive-margin setting, extension is followed by subsidence, with several kilometres or more of sediment deposited on the extended terrane. Involvement of this thinned crust with its sediment pile in, say, a convergent orogen will mean that extensional fault reactivation is initiated at some depth, probably in the ductile régime. Continued shortening and uplift is likely to exhume the stretched portion of the crust and raise the extensional faults into the brittle régime. Throughout this shortening history, folding and dissection by faulting of the extensional structures may occur, severely limiting their potential for reactivation.

Recognition of extensional structures within a compressional orogen such as the Alps, Himalayas or Caledonides will obviously be extremely difficult, and unravelling their reactivation history may well be impossible.

6. Conclusions

(1) Continental crust which has been significantly stretched contains an array of extensional faults with a range of orientations. In simple dynamic or kinematic terms, there is therefore potential for reactivation in a variety of tectonic settings.

(2) The mechanical principles of reactivation are reasonably well established in both the brittle and ductile régimes, and pre-existing fault or shear zones have generally lower strengths than surrounding intact rock. However, the strength differences between a pre-existing zone and the intact rock may be small, limiting reactivation to those zones in particularly high shear-stress orientations. Hydrothermal alteration in fault zones is common, and may even lead to an increase in strength of the zone.

(3) A range of styles of extensional fault reactivation is predicted. Given the similarity between extensional and thrust fault systems, reactivation during shortening near parallel to the original extension direction is the most favoured, with simple reversal of movement on the faults. However, oblique-slip and extensional reactivation are also possible. Wrench reactivation of transfer faults is likely to be widespread.

(4) Coupling between successive extension and shortening is now recognized as a consistent feature of orogeny. In this scenario, sedimentary basins floored by stretched crust are subsequently shortened and uplifted, for example in a collisional orogen. Simple inversion of sedimentary basins is the most easily recognizable result of this process. Reactivation of major basement extensional faults as thrusts may therefore be a very important tectonic process, albeit difficult to recognize at higher strains. Transfer faults may be a key factor in the production of long-lived lineaments in orogenic terranes. Because of their large strike and depth extent, parts of them may survive the orogenic process to be available for repeated later reactivation.

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References

Bally, A. W. 1981 *Geology of passive continental margins*, pp.1–48. Am. Ass. Petrol Geol., Educ. Course Note Series **19**.
Bally, A. W. 1982 *Phil. Trans. R. Soc. Lond.* A **305**, 325–328.
Barton, C. M. 1981 *J. geol. Soc. Aust.* **28**, 333–340.
Barton, P. & Wood, R. 1984 *Geophys. Jl R. astr. Soc.* **79**, 987–1022.
Beach, A. 1976 *Phil. Trans. R. Soc. Lond.* A **280**, 529–604.
Boyer, S. E. & Elliot, D. 1982 *Bull. Am. Ass. Petrol. Geol.* **66**, 1196–1230.
Butler, R. W. H. & Coward, M. P. 1984 *Tectonics* **3**, 347–366.
Byerlee, J. D. 1968 *J. geophys. Res.* **73**, 4741–4750.
Byerlee, J. D. 1978 *Pure Appl. Geophys.* **116**, 615–626.

Chen, W-P. & Molnar, P. 1983 *J. geophys. Res.* **88**, 4183–4214.
Craig, J., Downey, J. W., Gibbs, A. D. & Russell, J. R. 1984 In *The Canning Basin, W.A.* (ed. P. G. Purcell), pp. 57–72. Perth: The Geological Society of Australia.
Davis, G. A., Lister, G. S. & Reynolds, S. J. 1983 *Geol. Soc. Am. Abs.* **12**, 311.
Davis, G. A., Lister, G. S. & Reynolds, S. J. 1985 *Geology* **13**, (In the press).
Davis, P. N. 1983 *Seismic expression of structural styles* (ed. A. W. Bally), pp. 3. 3119–3.3/24. Am. Ass. Petrol. Geol., Studies in Geology Series 15.
Denham, D., Weekes, J. & Krayshek, C. 1981 *J. geol. Soc. Aust.* **28**, 323–332.
Dewey, J. F. 1982 *J. geol. Soc. Lond.* **139**, 371–412.
Donath, F. A. 1961 *Bull. geol. Soc. Am.* **72**, 985–990.
Donath, F. A. & Cranwell, J. 1981 *Mechanical behaviour of crustal rocks* (ed. N. L. Carter, M. Friedman, J. M. Logan & D. W. Stearns), pp. 231–241. Washington, DC: American Geophysical Union.
Etheridge, M. A., Branson, J. C., Falvey, D. A., Lockwood, K. L., Stuart-Smith, P. G. & Scherl, A. S. 1984*a* *B.M.R. J. Geol. Geophys.* **9**, 197–206.
Etheridge, M. A., Branson, J. C. & Stuart-Smith, P. G. 1985 *APEA J.* **25**, 344–361.
Etheridge, M. A., Wall, V. J., Cox, S. F. & Vernon, R. H. 1984*b* *J. geophys. Res.* **89**, 4344–4358.
Etheridge, M. A. & Wilkie, J. C. 1979 *Tectonophysics* **58**, 159–178.
Gibbs, A. D. 1984 *J. geol. Soc. Lond.* **141**, 609–620.
Gibson, G., Wesson, V. & Cuthbertson, R. 1981 *J. geol. Soc. Aust.* **28**, 341–356.
Goetze, C. & Evans, B. 1979 *Geophys. Jl. R. astr. Soc.* **59**, 463–478.
Harding, T. P. 1983 *Seismic expression of structural styles* (ed. A. W. Bally), pp. 3.3/13–3.3/18. Am. Ass. Petrol. Geol., Studies in Geology Series 15.
Harding, T. P. 1984 *Bull. Am. Ass. Petrol. Geol.* **68**, 333–362.
Jackson, J. A. 1980 *Nature, Lond.* **283**, 343–346.
Jackson, J. A., Fitch, T. J. & McKenzie, D. P. 1981 *Spec. Publ. geol. Soc. Lond.* **9**, 371–379.
Le Pichon, X. & Sibuet, J. C. 1981 *J. geophys. Res.* **86**, 3708–3720.
Lister, G. S. & Davis, G. A. 1985 *J. struct. Geol.* (In the press.)
McKenzie, D. P. 1978 *Earth planet Sci. Lett.* **40**, 25–32.
Miller, E. L., Gans, P. B. & Garing, J. 1983 *Tectonics* **2**, 239–263.
Montadert, L., Roberts, D. G., De Charpal, O. & Guennoc, P. 1979 *Initial Report D.S.D.P.* no. 48, pp. 1025–1060.
Plawman, T. L. 1983 *Seismic expression of structural styles* (ed. A. W. Bally), pp. 3.3/1–3.3/2. Am. Ass. Petrol. Geol., Studies in Geology Series 15.
Proffett, J. M. 1977 *Bull. geol. Soc. Am.* **88**, 247–266.
Sibson, R. H. 1982 *Bull. seismol. Soc. Am.* **72**, 151–163.
Stoneley, R. 1982 *J. geol. Soc. Lond.* **139**, 543–554.
Sykes, L. 1978 *Rev. Geophys. Space Phys.* **16**, 621–688.
Threlfall, W. F., Brown, B. R. & Griffith, B. R. 1976 *Aust. Inst. Min. Metall. Petrol. Monogr.* **7**, 41–67.
Wernicke, B. 1981 *Nature, Lond.* **291**, 645–648.
Wernicke, B. & Burchfiel, B. C. 1982 *J. struc. Geol.* **4**, 105–115.
Wernicke, B., Walker, J. D. & Beaufait, M. S. 1985 *Tectonics* **4**, (In the press.)
White, S. H. 1976 *Phil. Trans. R. Soc. Lond.* A **283**, 69–86.
Williamson, P. E., Pigram, C. J., Colwell, J. B., Scherl, A. S., Lockwood, K. S. & Branson, J. C. 1985 *APEA J.* **25** (In the press.)
Winslow, M. A. 1981 *Spec. Publ. Geol. Soc. Lond.* **9**, 513–528.
Ziegler, P. A. 1983 *Seismic expression of structural styles* (ed. A. W. Bally), pp. 3.3/3–3.3/12. Am. Ass. Petrol. Geol., Studies in Geology Series 15.
Ziegler, P. A. 1985 *Terra Cognita* **5**, 103.

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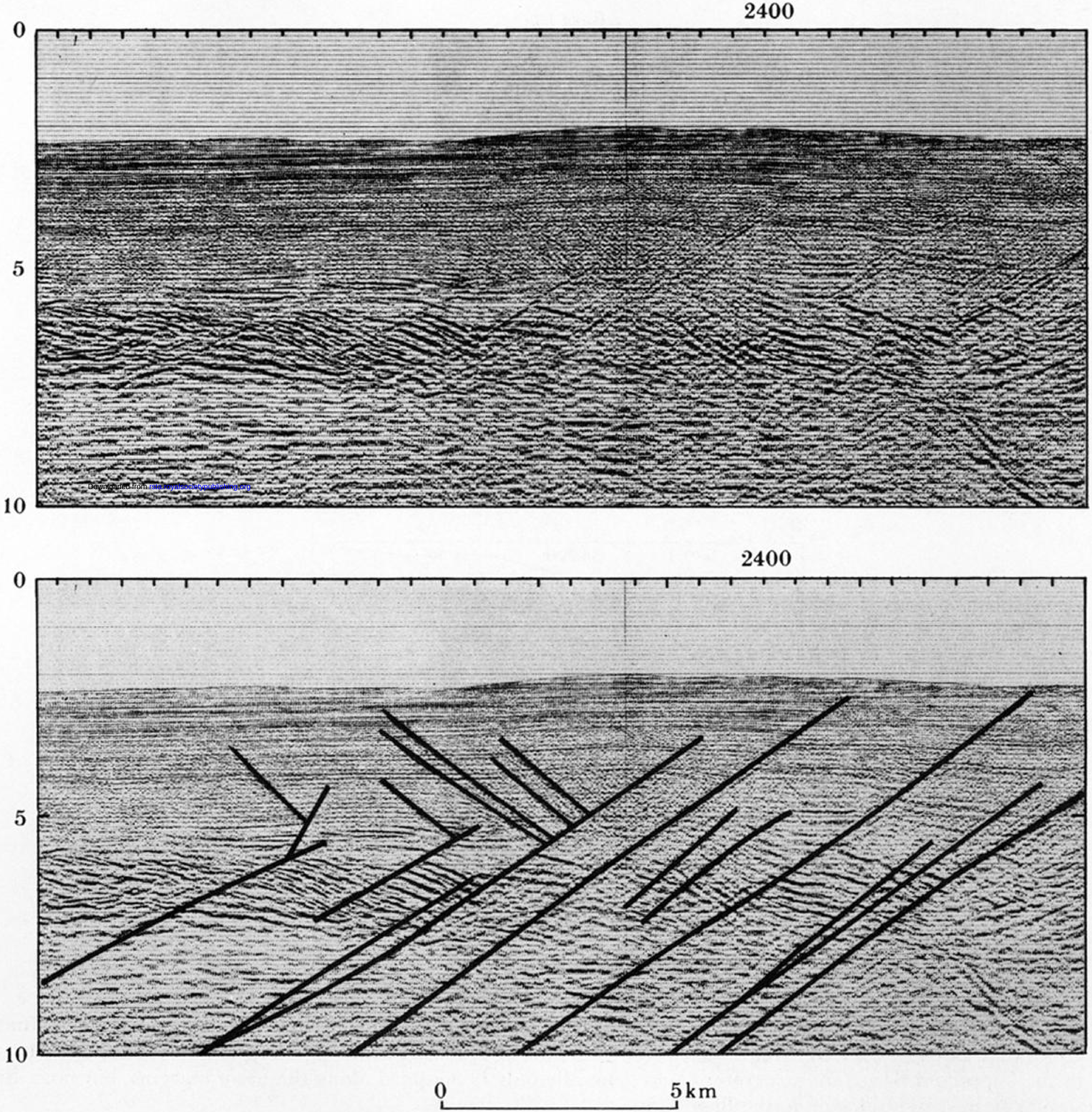

FIGURE 6. Seismic depth section of part of continental margin line 22 of the B.M.R. seismic survey (Etheridge *et al.* 1984*b*). Vertical exaggeration multiplied by factor of two.

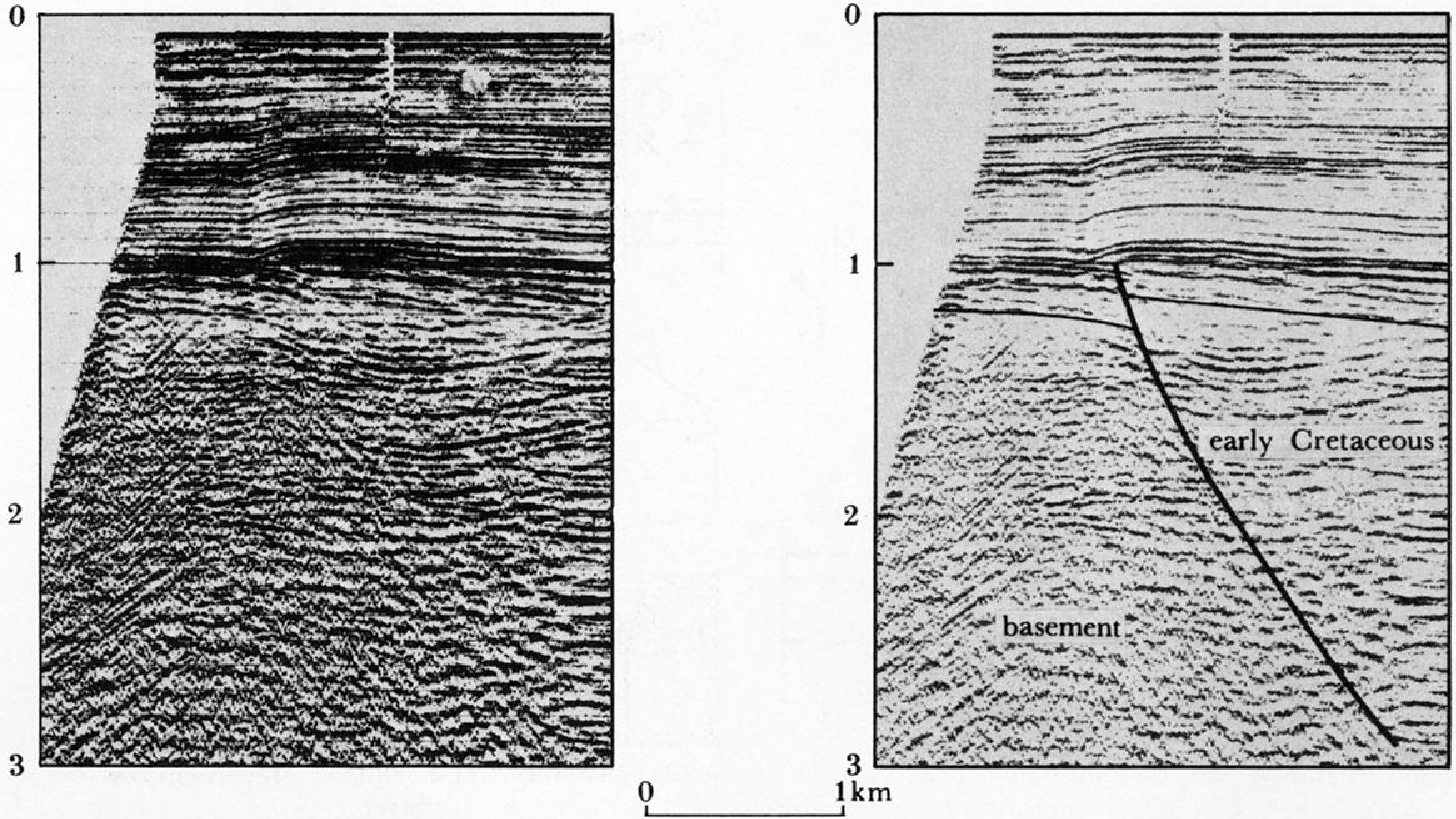

Figure 8. Seismic time-section from northeastern Gippsland Basin, illustrating reverse reactivation of a major early normal fault. Vertical exaggeration of *ca.* 1.5 at 3 s T.W.T.T.

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