Extensional subsidence, inversion and volumetric contraction in the Bass Basin of Australia: A seismic study

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DEDICATED TO MY WIFE REENA

STATEMENT OF AUTHENTICITY

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ABSTRACT

The Bass Basin, a Mesozoic-Cenozoic interior rift basin, has evolved through two rifting episodes, the Otway rifting and Tasman Sea rifting. Previous workers in the basin have variously interpreted the structural history and tectonic evolution of the basin with some recent views expressing the separate identity of two basins, the Durroon Basin and Bass Basin, corresponding to the southeastern and northwestern area of the previously known Bass Basin. The present study, which is mainly seismic based, was aimed to differentiate the influence of two rifting episodes on the basin and to characterise in detail, the structural style and subsidence history of the Durroon area and Bass area with an aim to ascertain the possible reasons why the two areas behaved differently in response to crustal extension. An additional objective of the study was to follow up the initial identification in 2D seismic data of a completely new type of fault system, a layer-bound polygonal fault system, by mapping the three dimensional characteristics of the fault structure using 3D seismic data and then delineate its possible extent in the basin from 2D data. In addition to the above, an attempt has been made to identify new possible direct hydrocarbon indicators in the basin while discussing the general hydrocarbon potential of the basin.

About 6000 line km of 2D seismic data and biostratigraphic and log information from 29 wells have been utilised to study the different fault patterns and the subsidence history in the basin. The Durroon and Bass areas have been considered separately because of variable density in the data set corresponding to the two areas. The interpretation was carried out using the seismic sequence stratigraphic approach following a modified version of the model proposed by Prosser (1993). While dividing the whole stratigraphy into its pre-rift, syn-rift and post-rift components, eight sequence boundaries were identified and correlated throughout the seismic data set. The extensional subsidence history in the two areas were studied in tandem, tracing the equivalent chronostratigraphic units in both areas. Although the structural style is distinctly different, it is suggested that the Durroon area and Bass area be referred as two separate sub-basins of the Bass Basin instead of current proposal for two separate basins as per the recent literature. It is newly proposed to use Durroon Sub-basin and Burnie Sub-basin respectively for the two areas.

The major erosional unconformity seen on seismic data near the Durroon-1 well within the Early Cenomanian succession has been interpreted from the overall seismic reflection geometry and internal seismic facies characteristics both above and below it, to be the rift-onset unconformity. This unconformity marks the onset of the Tasman Sea rifting phase (dated 96 Ma, *A. distocarinatus* zone). This major tectonic event may have been synchronous with the initiation of sea-floor spreading in the Southern Ocean.

The structural style in the Durroon Sub-basin is characterised by large, domino-style, tilted basement fault blocks. Fault throws are in the order of 4 - 5 Km and occur along rotational normal faults. Three major fault blocks, the Bark, Anderson and Boobyalla, have undergone extremely large tectonic subsidence related to extensional normal faulting. This is in complete contrast to the structural style in the adjacent Burnie Sub-basin to the northwest. Although two opposing major half-graben bounding-faults have accommodated initial extension with a large fault displacement (2-3 km) in this area, the predominance of sedimentary loading and syn-sedimentary faulting marks the structural style there with a large number of both planar and listric faults. The granitic nature of the basement in the Durroon Sub-basin as opposed to basement composed of metasediments in the Burnie Sub-basin is thought to have played an important role in controlling the way the two areas behaved in response to crustal extension.

The structural style in the Durroon Sub-basin has been shaped through three major extensional phases that are reflected in different fault trends. The initial response to extensional forces was the development of discrete fault segments across the area. The present day fault pattern subsequently developed through progressive propagation and linkages of the initial fault segments. The obliquity of Tasman rifting had a profound influence on the structural development. The offshore extension of the NE-trending Arthur Lineament probably acted as a buttress to limit the tensional stresses, resulting in the Late Cretaceous history in the Burnie and Durroon Sub-basins being very different. However, from the Early Tertiary, the two sub-basins became linked. Interestingly, the equivalent area corresponding to the Arthur lineament has been proposed as a transfer zone dividing the two sub-basins (Pelican-Squid transfer zone of Setiawan, 2000). It is suggested that perhaps, during the Otway rifting history, the area acted as a transfer zone that was also the limiting zone to the effect of tensional stresses during Tasman Sea rifting stage. However, all the earlier studies including this one suffer from the very poor data coverage in the intervening area between the two sub-basins corresponding to the proposed Pelican-Squid transfer zone (Setiawan, 2000; Das and Lemon, 1999) along the Arthur Lineament.

A simple shear-pure shear model of extension has been invoked to explain the major westerly-dipping normal faults seen in the Moore Basin west of the Lord Howe Rise (Stagg *et al.*, 1999) and the ENE-dipping basin-bounding faults in the Durroon Sub-basin. These two areas perhaps formed initially as conjugate pair in the southeastern Australian Continent during the rift valley development prior to opening of Tasman Sea. Otway rifting had already created major weakness in the crust in the area and superimposition of later Tasman Sea rift tectonics resulted in an apparently clean, sharp split in the continental crust seen in the vicinity of the shelf and slope area of the SE Australian margin.

The interpretation that the Durroon Formation (Smith, 1986) is confined to the Durroon area and has been deposited under the influence of the Tasman Rift and a later interpretation that the Durroon Formation is equivalent to the Golden Beach by Hill *et al.*, (1995, Figure 3) suggest that the extensional stresses due to Tasman Sea rifting event did not extend beyond the Durroon area. The depositional style of the Durroon Formation, i.e., sedimentation during active rifting on tilted fault blocks, is certainly analogous to the Golden Beach Group in the Gippsland Basin. On the basis of the present study, however, it is believed that although the extensional stresses of Tasman Sea rift had greater impact on the Durroon area in terms of massive displacement along normal extensional faults, leading to greater amount of fault related extension, the seismic data in the Bass area suggest that the impact of Tasman Sea rift had expressed quite differently. Chronostratigraphically, an equivalent unit to that of Durroon Formation could be correlated in the Bass area also based on seismic reflection characteristics but remains subjective in the absence of actual drilling information in this area. However, it is safe to predict that eventhough the two units might have been deposited during the same time interval (Early Cenomanian-Early Campanian), they would be lithostratigraphically quite different.

Although most of the previous workers in the basin have concluded that the extension amount in the Durroon area is less than the Bass area, the results of the present study do not support that view. A preliminary estimate of the amount of extension by measuring the pre-faulted basement length and the post-faulted basement length estimates extension of 1.36 in the Durroon Sub-basin compared to 1.22 in the Burnie Sub-basin.

Although inversion structures have earlier been identified from seismic data in the Burnie Sub-basin, this study has identified for the first time seismic evidence of inversion structures in the Durroon Sub-basin along some seismic lines. However, basin inversion has earlier been inferred from thermochronology data and erosion of strata on crests of tilted fault blocks in the Durroon area (Hill *et al.*, 1995). It is suggested that the inversion inferred from erosion of strata on uplifted footwall blocks, as seen near Durroon-1 well, should be concluded with caution because it can be explained by simple extensional model of hanging wall subsidence and footwall uplift related to thermo-mechanical behaviour of crust and subsequent erosion of the emergent crests of the fault blocks.

A new type of layer-bound fault system was identified on 2D seismic data in most parts of the basin. Details of this system have been drawn from a 3D seismic data set above the Yolla Field. This fault system in map view is composed of almost randomly oriented, high density, minor extensional faults organised in a polygonal network. The component faults are typically 500m-1000m long and have throws ranging from 10m to 40m. The average fault spacing ranges from 60m to 500m. In seismic sections, there are at least three units or stratigraphic tiers observed within the deformed interval showing a range of variation in the reflection characteristics and fault pattern development. The polygonal network of fault system shows a near equal distribution of fault strike orientations suggesting an isotropic stress regime during deformation of each unit within the deformed interval.

The polygonal fault system deforms the very fine-grained Late Tertiary calcareous clay and marl-dominated succession of the basal Torquay Group throughout almost the entire offshore Bass Basin. The seismic expression of this pervasively-deformed unit suggests that there is no displacement transfer to the basement structures and the stratigraphy overlying and underlying this sequence is undisturbed, characterised by continuous reflection patterns.

The development of polygonal fault systems by three-dimensional volumetric contraction of muddy sediments during early burial was first reported in literature from the central North Sea Basin (Cartwright, 1996). Layer-parallel volumetric contraction measured in seismic sections from the North Sea has recently been attributed to a process called *syneresis* of colloidal smectitic gels during early compaction history of sediments (Dewhurst *et al.*, 1999). Syneresis results from the spontaneous contraction of a sedimentary gel without evaporation of constituent pore fluid. This process is driven by inter-particle attractive forces in marine clays and is governed by the change of gel permeability and viscosity with progressive compaction. A similar process is attributed to the seismically-observed complex polygonal fault system in the Bass Basin.

Detection of direct reflection from flat fluid contacts unconformable with the surrounding rock reflections to indicate gas-filled reservoirs is now well established in the geophysical literature. Two new 'flat spot anomalies', one in the Durroon Sub-basin and the other in the Burnie Sub-basin have been identified. The 'flat spot anomaly' in the Burnie Sub-basin, near Koorkah-1 and Seal-1 corresponds to a possible gas-liquid contact within the reservoir sands of M. *diversus* zone of the early Tertiary Eastern View Coal Measures. The structure has been studied in good detail and shows an elongated anticline oriented NE-SW, with faulted margins to both the north and south. Traces of gas and insitu hydrocarbons in the sands at the *M. diversus* level in Seal-1 and traces of free oil in the Paleocene section in Koorkah-1 suggest the good prospectivity of the structure. The other 'flat spot anomaly' identified in the Durroon Sub-basin has not been detailed due to paucity of both seismic and nearby well data.

The Bass Basin holds very good potential for stratigraphic hydrocarbon play development. There are several possibilities in which the interdigitation of syn-rift sandstone-rich fluvio-deltaic reservoirs with source-rich deep water lacustrine facies in the footwall and hangingwall slopes of tilted basement fault blocks might hold good hydrocarbon trapping situation.

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CHAPTER 1 INTRODUCTION

1.1 General

The Bass Basin is one of the three Bass Strait basins in the southeastern Australia that have been explored for hydrocarbons since the mid sixties of the 20th century (Figure 1.1). Though the other two basins, the Gippsland and the Otway Basins, are now major oil and gas producers, the Bass Basin is relatively much less explored and is yet to have commercial hydrocarbon production.

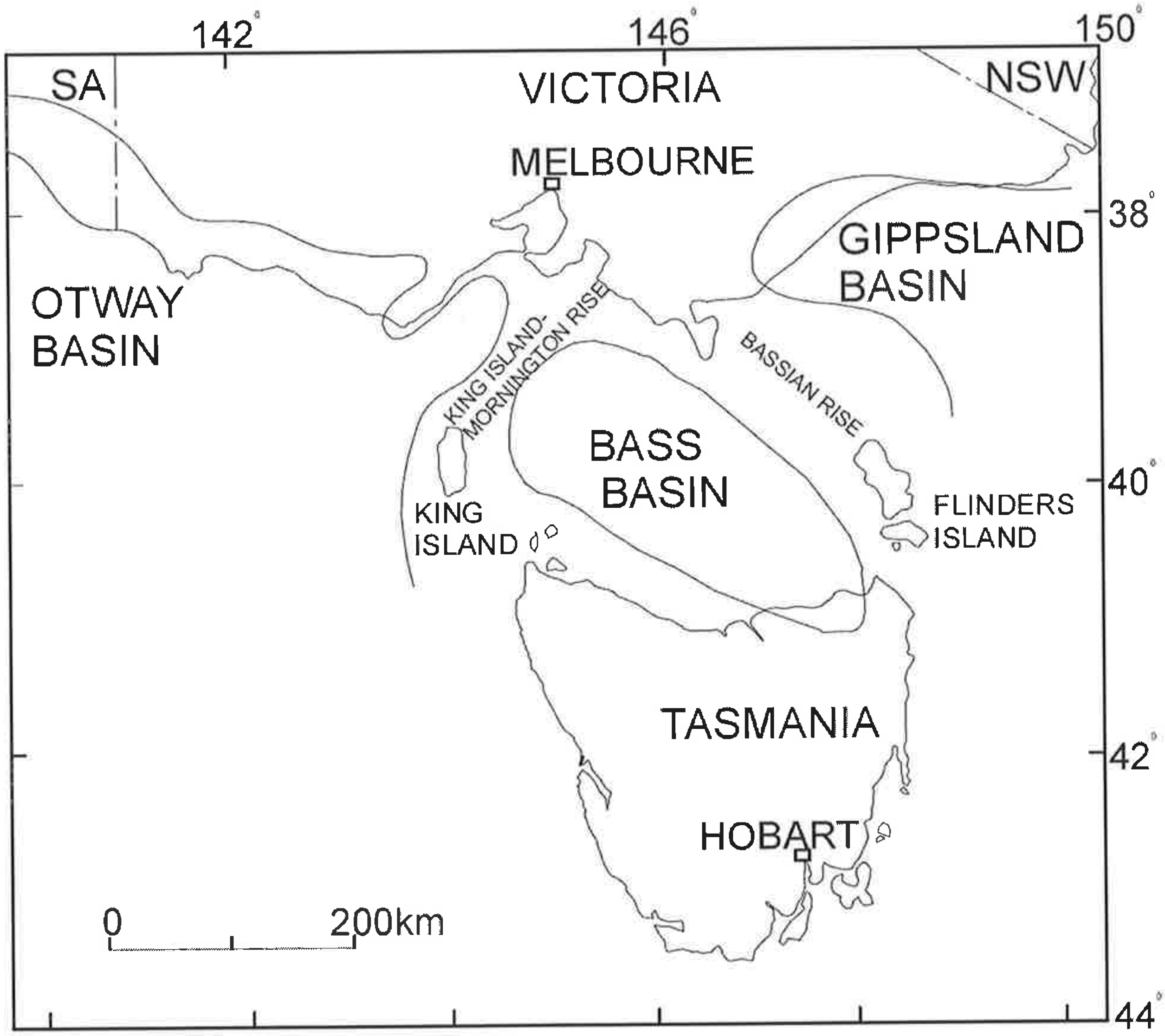


Figure 1.1 General location map for the Bass Basin.

Though the basin is thought to contain sediments ranging in age from Late Jurassic- Early Cretaceous to Recent, no well other than Durroon-1 has penetrated the Early Cretaceous section. Sampling of the Mesozoic section in the basin is very sparse with most of the wells targetting specific horizons in the Late Cretaceous-Early Tertiary part of the hydrocarbon prospective Eastern View Coal Measures Group. The literature on the Bass Basin suffers from a lack of consensus on the tectonostratigraphic evolutionary history of the basin. The initial outline of the Bass Basin (Weeks and Hopkins, 1967, Williamson *et al.*, 1987) is now considered to encompass two separate basins, the Durroon Basin and the Bass Basin corresponding to the southeastern and the northeastern part respectively (Baillie and Pickering, 1991; Gunn *et al.*, 1997b). A new terminology is required to remove overlap and ambiguity of the term 'Bass' Basin.

For most of this thesis, the terminology Durroon area and Bass area have been used to identify with the Durroon Basin and Bass Basin of the recently published literature. However, arguments have been proposed to define these two areas as two separate sub-basins of the complete entity of the Bass Basin. The Durroon area could be called the Durroon Sub-basin while the remaining area might be known as the Burnie Sub-basin, named after an important town on the northern coast of Tasmania. The two sub-basins have a similar history but the structural development shows influences from the east and west, which are more prominent in the sub-basin closest to those influences.

1.2 Objectives

The primary objective of the study was to gain a better understanding of the tectonostratigraphic evolutionary history of the Bass Basin. In particular, the study has focussed on mapping and analysing all the faults and fault patterns in the Bass Basin in relation to the subsidence history and its influence on sedimentation and hydrocarbon potential of the basin. The reason why the Durroon area and the Bass area behaved differently in response to extensional stresses was investigated. As a final outcome, it was thought important to clarify some of the existing disagreement about the broad tectonic and structural history of the basin and in particular to separate the influence of the Otway and Tasman Sea rifting episodes on the sedimentation history of the Bass and Durroon area. The study also aimed at investigating the occurence in the basin and nature of a recently recognised fault system, a polygonal fault system.

The study used a seismic sequence stratigraphic approach to integrate biostratigraphic, wireline and seismic data in analysing the structural style and stratigraphic evolution. A large amount of 2D seismic data formed the basis of the detailed interpretation of the structural history and the hydrocarbon potential of the basin. The only available 3D seismic dataset over the Yolla Field was utilised to analyse the nature of the polygonal fault system in the basin.

1.3 Rift tectonics and sedimentation

The manner in which the crust ruptures and rift basins develop in extensional terranes has been the subject of extensive research. Our understanding of the continental rifting process and the evolution of the extensional basins resulting from such crustal extension has been constantly improving with both high resolution sub-surface information and detailed study of outcrop geology.

1.3.1 Rift basin evolution models

The classic extensional model that recognises the evolutionary process from rifting to thermal subsidence was proposed by Falvey (1974). This model produces a rift system distributed symmetrically relative to the rift extension axis. McKenzie (1978) established the early model of extensional basin formation, assuming instantaneous rifting with uniform stretching (assuming the crust and lithosphere to extend by the same amount), thermal perturbation of the lithosphere and the decay of this perturbation in the post-rift stage after the rifting terminated. The model was further refined by the more realistic case of a finite duration for the rifting and its impact on the decay of any associated thermal anomaly (Jarvis and McKenzie, 1980). This model is commonly referred to as the pure shear model. Symmetric grabens defined by normal faults are formed from a simple trapezoidal block and allow isostatic subsidence of the block into the mantle. Following mechanical stretching, the upwelled asthenosphere should gradually cool, causing thermal contraction and regional subsidence that depends uniquely on β (the stretching factor) and characteristically takes the form of a negative exponential. Another crustal extensional model, accommodated by a major shear plane or crustal detachment, was proposed by Wernicke (1985) and based on a study of the Basin and Range Province, Western USA.

Morley (1995) has summarised the different characteristics of the various extension models. The low-angle detachment model or simple shear model for extensional faults proposed by Wernicke (1985) suggested that high and low-angled extensional fault systems could join into one, or a series of low-angled (less than 20^0) faults (brittle faults passing into narrow ductile shear zones) that traversed much or all of the crust. The main zone of thinning (faulting) at the surface would thus be offset tens of kilometres from the zone of lower crustal thinning (See figure 1.2c). This contrasted with the McKenzie (pure shear) model of rifting where ductile lower crustal thinning occurs below the surface rift (Figure 1.2a). Several authors have also suggested hybrids of the two models where low-angle faults may pass into a broad zone of bulk pure shear in the lower-middle crust (e.g. Rehrig & Reynolds, 1980, combined simple shear-pure shear model, Figure 1.2b)

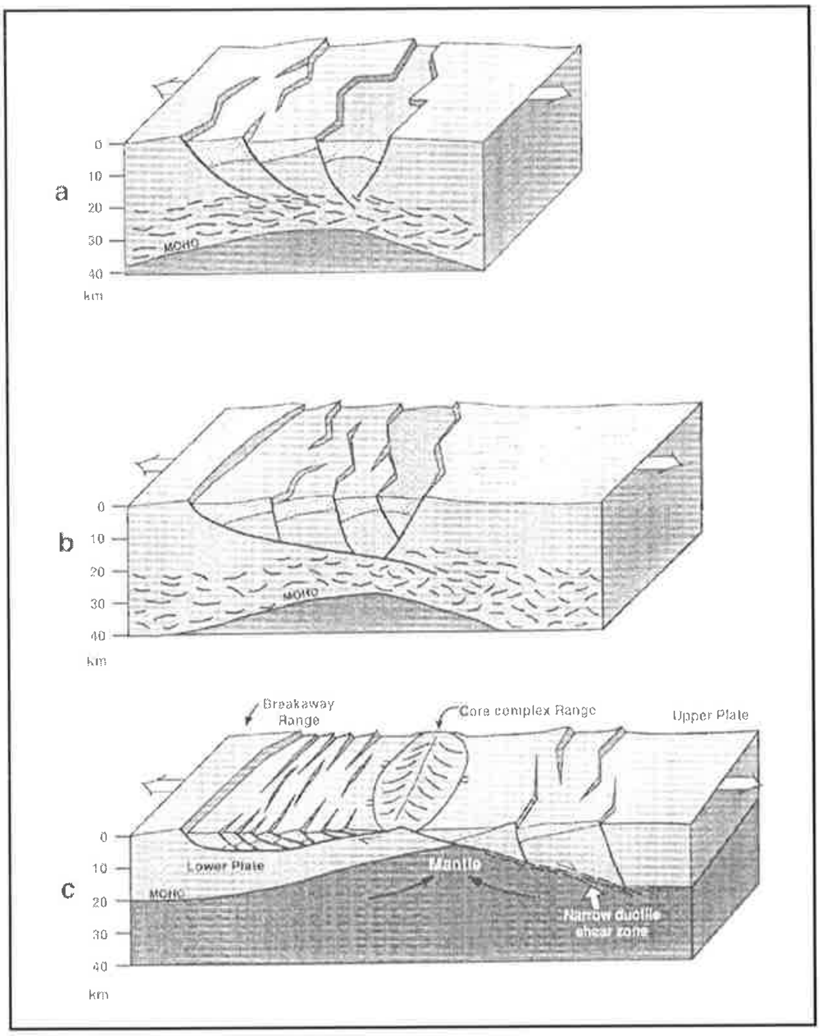


Figure 1.2 Diagram illustrating the main models of crustal extension. (a) Pure shear model (McKenzie, 1978). (b) Simple shear-pure shear model. (c) Simple shear model (Wernicke, 1981, 1985). Vertical scale=Horizontal scale. (After Morley, 1995).

1.3.2 Generalised rift architecture and structural style

Studies on continental rifts in the 1980s have established that they are dominated by asymmetrical half grabens rather than by symmetrical horsts and grabens (Rosendahl, 1987). The fundamental unit in a rift is a half-graben characterised by the boundary fault system at one side and flexural margin on the other side with both synthetic and antithetic faults (Rosendahl, 1987)(Figure 1.3a). In map view, it has been shown that some active rifts tend to be sinusoidal, such that the downthrown hanging walls of the asymmetrical half-gabens lie on the concave side of the boundary fault zone (Bosworth, 1985; 1987; Rosendahl, 1987). The sense of throw of these rift bounding fault zones thus reverses at inflections in the sinusoid or accommodation zones. These are commonly expressed by a region of strike-slip faults or a diffuse area of normal faults with opposite senses of throw (Figure 1.3b). Some authors have named the transverse structures separating oppositely-dipping half-graben as transform faults (Bally, 1982), transfer faults (Gibbs, 1984); transfer zones (Morley *et al.*, 1990). A large amount of research has been done on the nature and characteristics of these transfer zones (Chorowicz, 1989; Destro, 1995; Martin *et al.*, 1993; Moustafa and Abd, 1992; Nelson *et al.*, 1992 and Scott and Rosendahl, 1989).

The primary first-order structural control on sedimentation in rifts arises from the alternating polarity, half-graben/accommodation zone geometry (Lambiase and Bosworth, 1995) (Figure 1.3b). An additional factor of equal significance is the evolution of this geometry through time. Rifts undergo two general structural changes during their active extension histories. Fault geometries, on a regional scale, tend to become more interconnected and through-going. Alternating, opposed-border faults would be mechanically very inefficient, at least in the vicinity of acommodation zones. The extra energy expended at accommodation zones may be reduced through reversals of basin asymmetries with time, at a single position along the rift axis (Lambiase and Bosworth, 1995). Strata in the central half-graben of the model in Figure 1.3c initially dip to the right, opposite to strata in the two adjacent sub-basins. Later in its history, the central, left-dipping border fault blocks, and the border fault blocks of the adjacent half-grabens propagate across the central sub-basin's flexural margin. Eventually, one single right-dipping breakaway links the three sub-basins and, overall the system is simplified.

The second evolutionary aspect of rifts in their active extensional history concerns the increase of basin volume. As rifts extend, a basin forms and is filled with a combination of air, sediment and/or water. The sediment/air/water combination is generally less dense than the surrounding country rock, and will induce isostatic uplift to maintain a properly compensated lithospheric column, unless the lithosphere is infinitely strong (Wiessel and Karner, 1989). This uplift will be superimposed on thermally-induced effects of rifting, non-isostatically driven uplift of the footwalls of major faults and density changes attributable to necking of the base of the crust and lithosphere. Thermal phenomena are generally reflected in rift-fill sequences by the presence of volcanic rocks and/or elevated geothermal gradients.

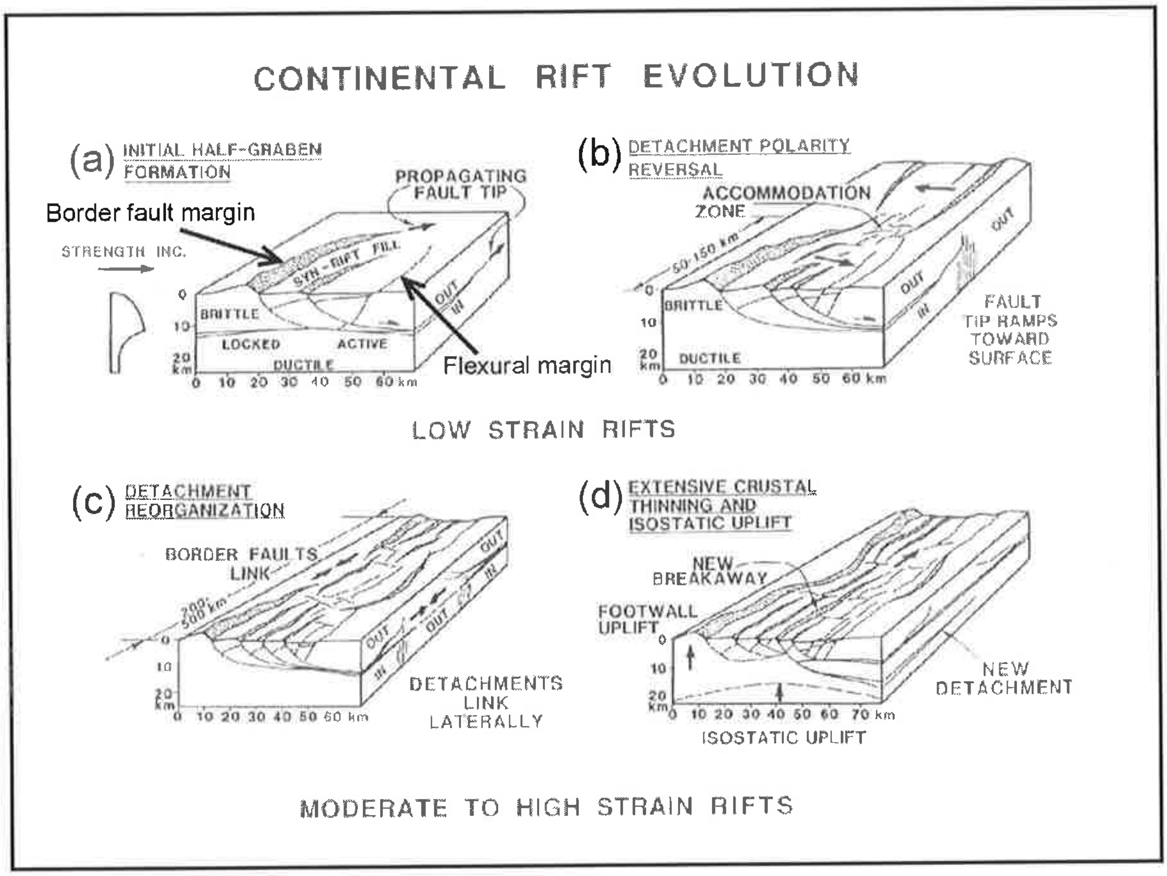


Figure 1.3 Kinematic model for continental rifting. Four main stages in the shallow crustal evolution of continental rifts can be recognised. (a) Rifts may initiate as gentle sags, but very quickly develop asymmetric half-graben cross-sections, probably by movement above a single detachment. (b) New detachments must then form, and in many cases, the dominant system dips in the opposite direction. Polarity reversal may arise due to random selection, or in some instances may reflect pre-existing basement structure. The junctures between detachment systems are referred to as accommodation zones. At higher strains, this mechanically inefficient arrangement of alternating detachments reorganises, eliminating faults that dip in one direction and linking border faults of the same polarity. (c) Crustal thinning leads to isostatic uplift and development of regional unconformities in the final stages of rifting, unless (d) thermal subsidence is of greater importance. Uplift results in abandonment of detachments and either late-stage high-angle normal faulting or formation of new detachments with new breakaways along the old rift axis. (After Lambiase and Bosworth, 1995).

Young rift sub-basins form with axis-parallel lengths of typically 50-150 km. The sub-basins very quickly acquire an asymmetric form and are separated along-strike by topographically high ground corresponding to the accommodation zones between individual crustal detachment systems (Lambiase and Bosworth, 1995). As strain accumulates some border faults or break-aways are abandoned, and fault systems coalesce into larger, more mechanically efficient geometries. This results in linking of sub-basins into single basins with length scale of several hundred kilometres. Accompanying these structural reorganisations, the rift undergoes isostatic uplift, such that although the low sides of fault blocks are subsiding and generating space for sediment, the high corners are simultaneously rising and being eroded (Figure 1.3d).

Thus the generalised picture of rift evolution is manifest at the surface by a number of recurring structural associations and phenomena. The most important of these are accommodation zones or transfer zones, uplift of rift shoulders and internal geometry of the asymmetric rift basin.

Transfer zones in rifts can be defined as a coordinated system of deformational features conserving extensional strain and commonly allowing changes in the boundary fault configuration to be accommodated in a variety of ways (Morley *et al.* 1990). Crustal extension results most importantly in the development of a network of normal fault systems in rift basins. Transfer fault zones refer to the conservation of the fault displacement in three dimensions and the term implies that the faults across which the change in geometry occurred were active at the same time and that extension was transferred from one fault or fault system to another along strike of an opening rift.

Continental rifts commonly display large-scale structural domains (tens of km) that have unique or consistent structural style. These domains are largely controlled by the geometry of, and interactions among, major displacement faults (heave greater than 1 km and throw sometimes a few km). Such major faults generally define either the rift boundary or large horst blocks. Large throws on faults usually generate isostatically uplifted footwalls, deformed by the flexural strain. Hence, as faults die out along strike, so do the footwall uplifts. Consequently transfer zones may commonly accommodate not only displacement but also elevation differences between adjacent footwall blocks.

Transfer zones in rift basins show a wide range of deformation including either discrete faults affected by normal slip, oblique slip or strike-slip or wide complex zones of pure normal faulting, transtension or broad warping. Transfer zones can occur at a variety of scales depending upon the size of the faults, and transfer systems of different scales may be nested one within another (Morley *et al.*, 1990). Again, they can be represented by a single fault to a broad area. A single fault acting as a transfer zone may link two normal faults, basins or sub-basins with different amounts of extension or different polarity or areas of different block rotation (e.g. areas with planar faults vs areas with listric faults)(Moustafa, 1997). On the other hand, transfer zones covering a broad area exist between half-grabens of opposite tilt directions or even between extended parts of the crust characterised by different structural styles (e.g. areas with predominantly horst and grabens vs areas with titled fault blocks). These zones help transfer the throw from one half-graben to the next as the rift propagates. Sometimes, the transfer zones are sites for major volcanic activity as seen in the southern Rio Grande rift (Mack and Seager, 1995).

There is still some current debate as to the nature of transfer zones. They may involve large scale cross faults orthogonal to the general rift trend with dominantly strike-slip or oblique-slip movement. There are some documented examples in the literature but in most of the cases, they have been rather assumed or inferred from an inadequate data coverage. The relay transfer fault systems in the region of overlapping fault terminations without cross faults may be more common. Some of the well documented strike-slip movement in the transfer zones have, however, been ascribed to the influence of the pre-existing structural grain in the basement on the distribution of the extensional strain depending upon the orientation of these pre-existing faults to the relative regional extension direction (e.g. Gulf of Suez; Moustafa, 1997).

1.3.3 Sedimentation patterns in rift setting

The pattern of sediment dispersal in a rift basin is strongly related to the rift architecture, as clearly illustrated in the publication by Leeder and Gawthorpe (1987) for continental half-grabens. Several other publications (Crossley, 1984; Rosendahl, 1987; Larsen, 1988; Morley *et al.*, 1990; Lambiase, 1990; Lambiase and Bosworth, 1995) have dealt with the spatial characteristics of sedimentation in an actively rifting basin. The basic structural unit in any rift is a half-graben and acts, during a major part of the rifting history, as a distinct structural and depositional centre. Half-grabens are generally 50 km wide and 100 km long with both sides bounded by major faults; on the down-dip side by the rift border fault and in the flexural margin side by another major fault. This gives the half-graben a characteristic asymmetry (Rosendahl *et al.*, 1986; Lambiase, 1990). The rift shoulders are topographically much higher than the corresponding sides of the half-graben unit and the border fault in the down-dip side has got the maximum throw or displacement. The basin floor is also faulted with other synthetic and antithetic faults giving a block size of 5-10 km.

The initial style of a rifted basin is a series of half-grabens linked in a network of oppositely-facing units with the linking governed by the transfer zones or accommodation zones between these half-grabens. The isostatic flexural / thermal response of the crustal extension will be uplift of the border fault margins and this is generally highest at the centre of the fault and dies out towards the tips of the faults. The elevation at the transfer zones is higher than the basin floor but not as high as the rift shoulders. The axial plunge inside a half-graben may either be in one general principal direction all along the rift strike or may be opposite. These factors together create a profound impact in the rift sedimentation as discussed below.

Rift shoulders deflect the regional drainage away from the basin, thereby limiting sediment influx (Frostick and Reid, 1989). This is more prominent on the down-dip sides of half-grabens where rift shoulders are highest and hence, much more sediment enters the rift from its up-dip margins. Generally, because no structural barrier marks the end of a rift system, sediment has relatively easy access into the system-terminating half-graben. In many systems, this axial transport contributes much of the sediment. The other entry point which is conveniently exploited by the sediments is along the transfer zones and large-scale river systems can follow these transfer zone route to dump clastic sediments into the rift.

Because of the asymmetric dip and plunge of the half-graben, a topographic low is formed that becomes a depocentre as drainage is focussed into it. However, fault bock topography on the basin floor impedes down-dip sediment transport by creating smaller depocentres in the hanging wall against the rift shoulder on an adjacent rift block. Whenever the rate of topographic rejuvenation exceeds the rate of deposition, most sediment moving down-dip becomes trapped in this manner, although some may bypass along the small-scale transfer zones between the small fault blocks. The up-dip ends of the topographically prominent blocks, including the uplifted margins, are always prone to erosion and thus can contribute to the sediment input to the system.

The typical half-graben geometry inhibits rift-axial transport of sediment both in the early stages and also in the later stages when the active faulting creates the subsidence in the hanging wall blocks and the flexural response creates the uplift of the margins in the footwall blocks. Therefore, each half-graben essentially becomes an isolated depositional basin and since these depocentres are almost confined from all sides, the sedimentation rate is slow and there is increasing tendency towards deep lakes and the deposition of lacustrine sediments with enriched source rocks at this stage.

The depositional patterns and facies distribution change when deposition rate exceeds the rate of tectonically-controlled topographic rejuvenation. And with progressive deposition of sediments, the fault block topography becomes buried, and basin-scale depositional units form. Eventually, rift axial transport becomes more dominant after the transfer zones have been buried, and finally, rift shoulders are eroded and no longer effect the depositional patterns.

A rift basin passes through several tectonic and structural phases during its evolution from initial rifting to post-rift subsidence, and is accompanied by a particular depositional style. The exact timing of the uplift of the rift shoulders is very important because sediment supply rate is altered significantly by development of rift shoulders. Before the shoulders form, regional drainage freely enters the basin, and sediment supply rate can be very high; after uplift, sediment supply is restricted to internal sources. Prior to this, however, block faulting occurs and although shallow lakes can form in topographic lows bounded by the fault blocks, the lacustrine deposits are laterally restricted and relatively thin. Most deposits are fluvial and alluvial sand and gravel, which form the basal unit of the generalised stratigraphic sequence. As subsidence along the basin-bounding faults continues, basin asymmetry develops rapidly relative to sedimentation rate.

Immediately after this stage, the transfer zones and the bounding faults make the stage set for large lake development. The climatic effect will however, decide to a large extent the lake extent and lake-level by controlling the availability of water. The cyclic fluctuation of the lake level based on the supply of water, in conjunction with the tectonically-driven basin subsidence, will determine the relative rise and fall of base level and thereby, control the transgression and regression cycles with respect to the lake level (Barr, 1987; Frostick & Reid, 1989). Here, another important consequence should also be noted that in marine-inundated basins, the similar effect would be governed by the relative sea level change instead of lake level change.

The time span from the initiation of rifting to large lake formation is quite rapid and the early basal fluvial units are comparatively thin and are followed by deep-water lacustrine facies. This stage is again important from hydrocarbons point of view because, the sediment supply rate is quite low and the large-scale deposition of rich source rocks in the deeper water is favoured. The rift tends to be sediment starved while the interplay between tectonics and sedimentation favours subsidence outpacing the sediment supply rate. At the ends of the rift, the end half graben generally forms a ramp into the surrounding region and allows sediments to enter the rift at either or both the end half graben limits.

In the next stage, either by better development of the regional drainage systems or by subsidence rate decreasing, the sediment supply rate will outpace the subsidence and the end half-grabens experience this earlier and fill up first. Another factor helps here because the end half-grabens have a larger drainage area than the others. At this stage, alluvial sediments fill the half-grabens without the lake while those with lakes with deltaic and pro-delta sediments. The sedimentation will breach the transfer zones after filling to the neck of each half-graben and will spill over to the adjacent half graben unit. The switching of depocentres can also occur with time if tectonic subsidence creates an imbalance between the adjacent half-grabens. The sandy sequence will dominate the area in the entry points along the transfer zones and further up-dip whereas the down-dip lows in the adjacent lake will be shale prone.

As the basin fills with sediments, the shallowing-up in the sedimentary sequence is a notable phenomenon. Also, once the sediment transport breaches the transfer zone to reach the next depocentre, the earlier filled half-graben becomes a flood plain area and acts as a sediment bypass zone despite the presence of the adjacent high rift shoulders.

This process continues till all the topographic barriers are submerged and the subaerial deposition by the through-going river system of alluvial/flood plain deposits is the case. If the regional sag phase is influenced by marine incursion, as the case in the Bass Basin of Australia, widespread shallow marine deposition will mark the post-rift sag phase. The Reconcavo Basin of Brazil has however, not experienced much post-rift sedimentation due to the fact that the cooling necessary to mark the phase was already over through heat dissipation during the long rifting history of the basin (Milani & Davison, 1988). This means, the way the tectonic subsidence marks the rifting phase in terms of the duration of the geological time over which the fault activity is spread, the thermal sag phase will be accordingly variable.

Where there is no marine influence, even in the late stage of the basin history, the erosion of the remaining rift shoulders and influx of sediments from outside of the basin, account for a thick alluvial/fluvial sequences and further regional subsidence of the basin results in a broad depression which gets filled by the swampy, paludal and shallow lacustrine deposition as in the case of the Sudan Basin (Schull, 1988).

To summarise the effect of tectonism and structural setting on sedimentation, it can be said that, although the movements on major boundary faults will have considerable long term effect on both the character and quantity of detritus supplied to the basin, environmental effects like lake level in a closed or barred basin, or eustatic level changes, will also control to a great extent the type and nature of sediments. Also, the lateral extent of the inundation resulting from transgression depends not only on the rise of water level, but also on topographic factors. In narrow, steep-sided rifts like Lake Tanganyika, a 5 m rise in water level would shift the shore line by 30 m whereas the same rise in water level in Lake Turkana would cause an average shift of 800 m because of the low angle of the boundary faulted margins (Frostick and Reid, 1989). This is again important from the petroleum exploration point of view because fine lacustrine or marine sediments would serve as lateral and top seal with the more permeable alluvial deposits with which they inter-digitate updip along the sloping fault block margins which experience transgressive sand deposition during lake-level highstand. Similarly, the lake-level lowstand will either create incised valley systems forming sand bodies along the longer and wider hangingwall block slopes and synchronously, the deep water turbidites would be concentrated along the deepest parts of the subsiding depocentres.

Lambiase (1990) has recognised four-stages in the general stratigraphic sequence in continental rifts. It begins with a basal unit that is dominated by fluvial and alluvial sands. This is overlain by lacustrine deposits with an abrupt transition from the fluvial to lacustrine environments. The deepest water conditions are represented by the lowermost lacustrine sediments. In the third stage, progressively shallower water deposits occur upward through the lacustrine interval, culminating in a gradual transition to fluvio-deltaic sedimentation. Eventually, the transition to subaerial environments is completed and subsequent deposits are primarily of fluvial and alluvial origin. This style of sedimentation continues through post-rift regional subsidence stage.

1.4 Seismic sequence stratigraphy

In the late 1960s, Peter Vail and several co-workers at the Exxon Production Research Company began to recognise unconformity-bound sequences on panels of seismic. By tying the seismic reflections to wells across a basin, they found that single reflections, formed as a result of the contrast in acoustic impedance between adjacent units, could be associated with differing lithologies at the basin margin and basin centre. He conjectured that seismic reflectors may present time equivalent or chronostratigraphic surfaces. Prior to this, seismic reflections were considered to be lithology or lithostratigraphic boundaries. Detailed analysis of well data showed this to be incorrect. Exxon further developed these ideas through detailed outcrop studies complemented by well and seismic data. The ideas were first published in 1977 as AAPG Memoir No. 26, 'Seismic Stratigraphy - Applications to Hydrocarbon Exploration' (Payton, 1977).

Seismic sequence stratigraphy is the study of stratigraphy and depositional facies as interpreted from seismic data (Mitchum *et al.*, 1977a). Seismic reflection terminations and configurations are interpreted as stratification patterns, and are used for recognition and correlation of depositional sequences, interpretation of depositional environment, and estimation of lithofacies.

Seismic sequence stratigraphic analysis is fundamentally based on the concept of depositional sequence which is defined as a stratigraphic unit composed of a relatively conformable succession of genetically-related strata and bounded at its top and base by unconformities or their correlative conformities (Mitchum *et al.*, 1977b). Because it is determined by a single objective criterion-the physical relations of the strata themselves-the depositional sequence is useful in establishing a comprehensive stratigraphic framework. It is not primarily dependent on determinations of rock types, fossils, depositional processes, or other criteria that generally are subjective and varied within a given sequence. Unconformities that bound depositional sequences are observable discordances in a given stratigraphic section that show evidence of erosion or nondeposition with obvious stratal terminations, but in places they may be traced into less obvious paraconformities recognised by biostratigraphy or other methods (Mitchum *et al.*, 1977b), The type of discordance is based on the manner in which strata terminate against the unconformable boundary of a depositional sequence (or a structural boundary).

Seismic sequence analysis subdivides the seismic section into packages of concordant reflections, which are separated by surfaces of discontinuity defined by systematic reflection terminations. These packages of concordant reflections (seismic sequences) are interpreted as depositional sequences consisting of genetically-related strata and bounded at their top and base by unconformities or their correlative conformities. Reflection terminations interpreted as stratal terminations include erosional truncation, toplap, onlap, and downlap (Mitchum *et al.*, 1977b).

Seismic facies analysis interprets environmental setting and lithofacies from seismic data. Seismic facies units are group of reflections whose parameters (configuration, amplitude, continuity, frequency, and interval velocity) differ from adjacent groups. After seismic facies units are recognised, their limits defined, and areal associations mapped, they are interpreted to express certain stratification, lithologic, and depositional features of the deposits that generated the reflections within the units (Mitchum *et al.*, 1977a). Lapout is the lateral termination of a stratum at its original depositional limit. Truncation is the lateral termination of a stratum as a result of being cut off from its original depositional limit. As mentioned earlier, primary seismic reflections follow chronostratigraphic (time-stratigraphic) correlation patterns rather than time-transgressive lithostratigraphic (rock-stratigraphic) units (Vail and Mitchum, 1977). Physical surfaces that cause seismic reflections are primarily stratal surfaces and unconformities with velocity-density contrasts. Stratal surfaces are major bedding surfaces and thus represent ancient depositional surfaces.

1.5 Tectonic systems tracts

Brown and Fisher (1977) defined the term 'systems tracts' as a 'linkage of contemporaneous depositional systems' and Posamentier and Vail (1988) described further the nature of idealised linked depositional systems that develop in response to a cyclical change in relative sea-level, controlled by fluctuations in eustacy, tectonics and sedimentation rates in a passive continental margin setting. They proposed the highstand, lowstand, transgressive and shelf-margin systems tracts to be basic components in the basin infill stratigraphy in basin analysis. Prosser (1993) suggested that in actively extending rift systems, it would not be possible to identify the eustatically controlled systems tracts of Posamentier and Vail (1988) and instead, the reflector dispositions and internal facies characteristics can be related to dominantly tectonic control. She coined the terminology 'tectonic systems tracts' to describe the various depositional systems in rift setting where tectonic control is the dominant influencing factor in sedimentation. While stating that the spatial distribution and temporal evolution of depositional systems in active-fault bounded basins is significantly influenced by active tectonics, Prosser proposed a four-fold division relating to rift initiation, rift climax, immediate post-rift and late post-rift stages of basin evolution to characterise rift basin infill stratigraphies.

Figure 1.4 shows schematically the **'rift initiation systems tract'**. The seismic expression is characterised by overall wedge-shaped reflection geometry with the thin end of the wedge lying high on the hanging wall dip-slope, and the internal reflector characteristics are generally hummocky and discontinuous. The depositional elements that constitute the systems tracts are (I) a longitudinal river system with channels and inter-channel-type deposition; (ii) small coarse-grained talus cones from the low relief fault scarps. The rate of sedimentation generally keeps pace with subsidence that is reflected in the wedge-shaped geometries that are superimposed upon one another, rather than formed through onlapping reflectors infilling an earlier-formed depression.

The 'rift initiation systems tract' may not be preserved always or if the seismic resolution is not good, it would be difficult to recognise. Alternatively, if the rifting is conspicuously at a very fast rate leading to creation of a faulted depression very early on in the basin evolution history, the second stage systems tracts may be more prominent.

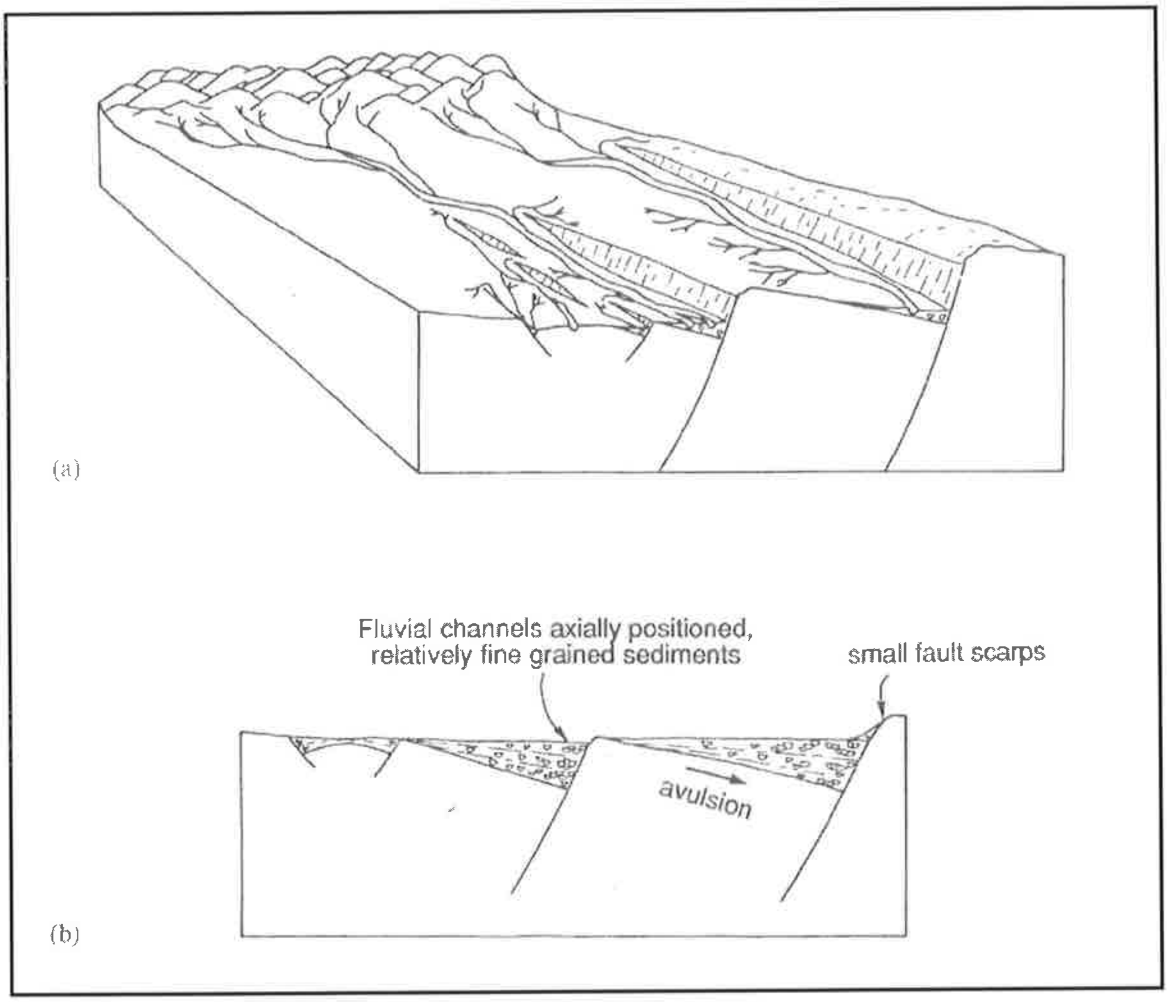


Figure 1.4 Rift initiation systems tract: (a) generalised block diagram; (b) schematic cross section (After Prosser, 1993).

The second stage division is the **'rift climax systems tract'** and the seismic expression is characterised by an increase in the amount of aggradation, together with the development of divergent forms related to continued tilting of the hanging wall during deposition (Figure 1.5). In high quality data, mounded forms associated with footwall-derived fans and talus may be recognised. Chaotic zones close to the footwall may represent coarse-grained rock falls and talus which have no ordered bedding and no chance of generating reflections. In more basinward positions, the contrast in lithology between fans and lacustrine/marine deposits means that interbedding of horizons with strongly contrasting lithologies is likely to occur and reflections should be generated. Although Prosser (1993) has further classified the rift climax systems tracts into early-, mid- and late- components, it is difficult to distinguish these individual systems tracts unless very high quality reflection data is available.

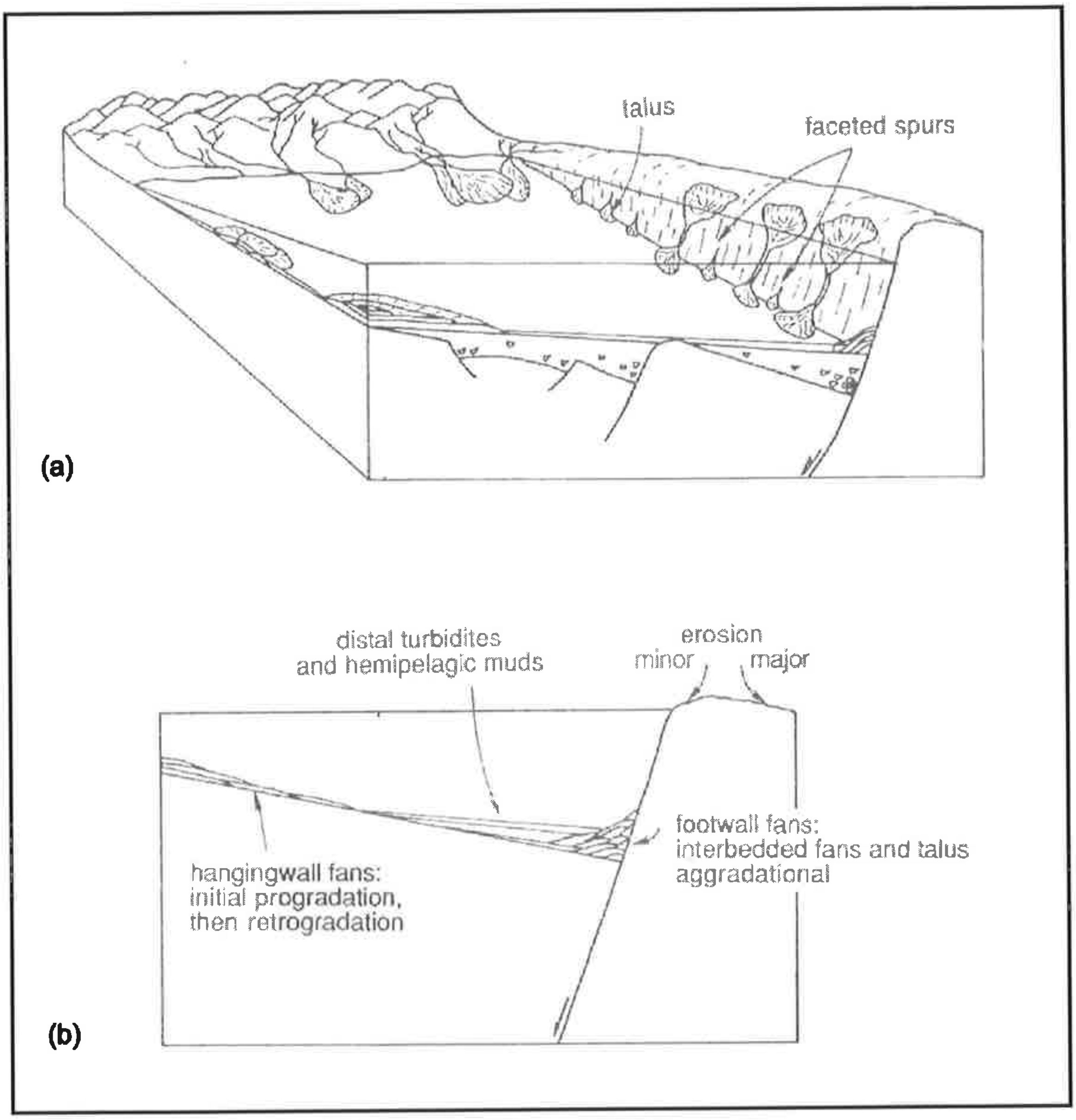


Figure 1.5 Rift climax systems tract: (a) generalised block diagram; (b) schematic cross section (After Prosser, 1993).

The **'immediate post-rift systems tract'** is signalled by an end to active tectonism and displacement on the main basin bounding faults, which means tilting of the hangingwall and differential subsidence across the fault plane ceases. As against Prosser's observation that progradation in the hanging wall slope and aggradation in the footwall slope would mark this stage, it is proposed here that the systems tract at this stage will be characterised by a continuous draping reflector that can be traced across the basin onto the adjacent footwall and hangingwall crests (Figure 1.6). The reason behind such a conclusion is that the faulting-induced displacement stops and there will therefore, be a continuous reflector from the individual footwall crest to the adjacent hangingwall crest and although the undulated topography created by the rifting till this stage will exhibit thickening of the strata towards the individual depocentres in the fault blocks. This is clearly seen in the seismic data in the Durroon area as discussed in Chapter-4.

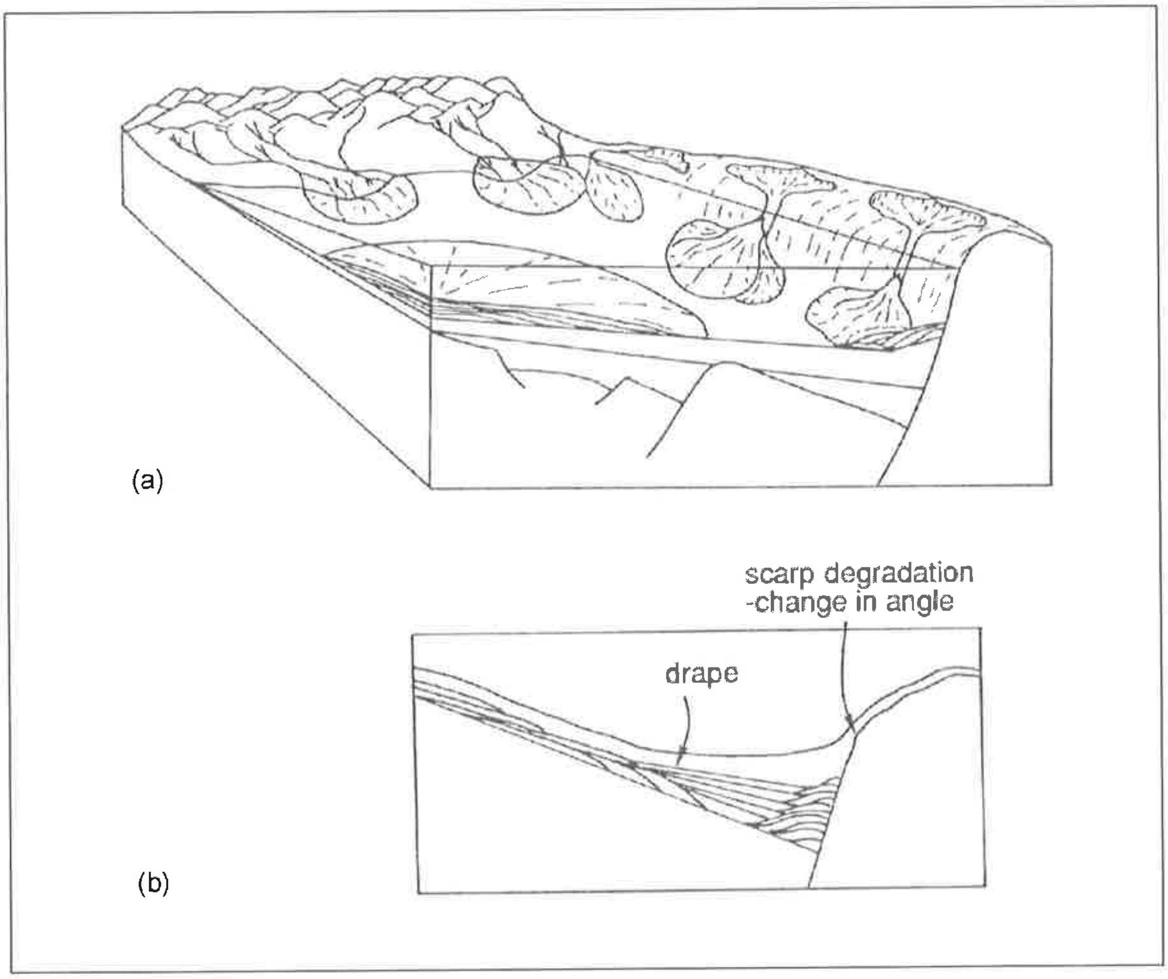


Figure 1.6 Immediate post-rift systems tract: (a) generalised block diagram; (b) schematic cross section (After Prosser, 1993).

The **'late post-rift systems tract'** develops as a response to the slow peneplanation of the faulted topography through gradual infilling. Seismically, it will be characterised by parallel reflectors that will be more continuous than the early post rift systems tract and will occupy positions through the entire cross section of the basin (Figure 1.7).

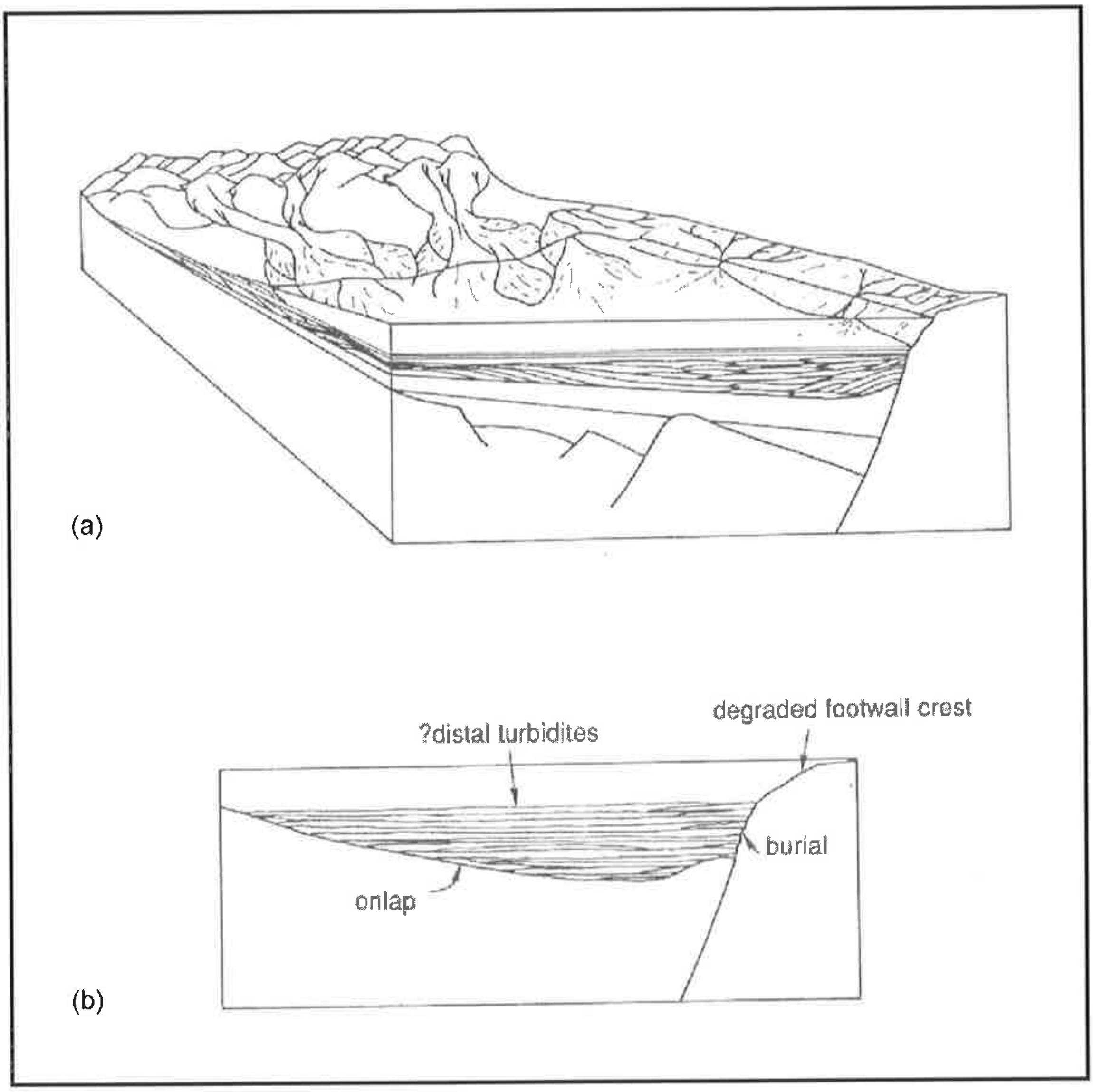


Figure 1.7 Late post-rift systems tract: (a) generalised block diagram; (b) schematic cross section (After Prosser, 1993).

Figure 1.8 shows an idealised line drawing of a seismic section across an ideal rift basin wherein all the systems tracts are clearly identifiable. But in practice, it may not be possible to clearly define each systems tract if it is an extremely thin component of the basin fill, or if local influences prevented its formation or preservation.

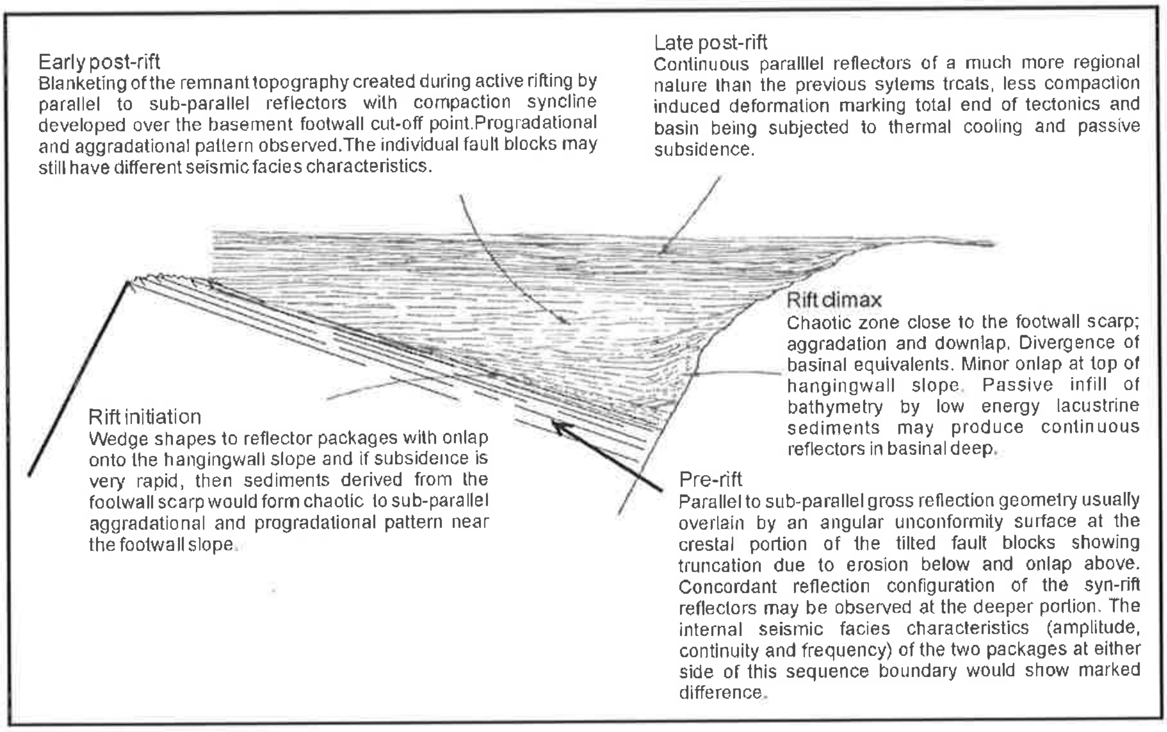


Figure 1.8. An idealised section of a line drawing of a seismic section through an ideal tilted basement fault block in a rift basin with characteristic seismic expressions (After Barr, 1991 and Prosser, 1993).

As discussed in Chapter-4, the stratigraphy in the Durroon area has been divided into tectonics systems tracts based on the model presented thus far. The rift initiation systems tract is not very clearly distinguishable and therefore, the two syn-rift sequences (syn-rift sequence I and syn-rift sequence II) constituting the syn-rift package can perhaps be comparable to the rift climax systems tract. The reason why the rift initiation systems tract is not clearly identified is perhaps it is not preserved or it is beyond current seismic resolution. In the SE Bass Basin, the early syn-rift or rift initiation systems tract is not very clearly distinguished and it is suggested that the two syn-rift sequences identified may both be comparable to the rift climax systems tracts.

1.6 Polygonal fault system

During the later half of 1990s, a new type of layer-bound polygonal fault system was recognised in the North Sea Basin data in the high-resolution 2D and 3D seismic data (Cartwright 1994a; 1994b; 1996). The group led by Cartwright made an extensive study of the polygonal fault system in the North Sea Basin and discovered the intricate relationship between clay mineralogy of the deformed rock types and polygonal fault system development. They noted that only 3D seismic data is suitable for mapping the very small-scale dimensions of the polygonal fault system. Although several other mechanisms have been proposed for their genesis, the polygonal fault systems are still considered to be caused by the colloidal mechanism called 'syneresis' (discussed in details in Chapter-5).

No detailed study on this new type of fault system has been carried out in the Bass Basin so far. However, midway through this project, a publication citing examples from 23 basins world-wide of the occurrence of this fault system was published by Cartwright and Dewhurst (1998) in which they have mentioned the possibility of this fault system in the Bass Basin based on their interpretation of the earlier published literature.

CHAPTER 2 PREVIOUS WORK IN THE BASS BASIN

2.1 Introduction

The Bass Basin is an intracratonic Mesozoic-Cenozoic rift basin covering an area of about 65,000 square km and is considered to have been formed during the rifting between the Australian and the Antarctic plates. The basin is geographically located between mainland Australia in the north and Tasmania in the south and lies almost wholly under Bass Strait waters with present day water depth ranging between 30 to 90 m (Williamson *et al.*, 1987). The eastern flank of the basin is occupied by the Flinders Island-Bassian Rise whereas the western flank is occupied by King Island and the King Island - Mornington Peninsula Ridge (Figure 1.1). Although this chapter is primarily based on the previous work by different workers in the basin on various aspects of geology, wherever applicable, comments as regards any variation from what is published in the literature, observed during this study, have been suitably noted.

2.2 Exploration history

The knowledge about the generalised tectonic setting, structural history and stratigraphic framework of the Bass Basin has developed over almost a period of 4 decades of petroleum exploration work by many different oil companies operating in the region. Exploration work in the basin began with a reconnaissance aerial magnetic survey flown on 12th December, 1960 followed by a major aeromagnetic survey in 1961 and regional seismic surveys in 1963. The first wildcat well, Bass-1 was spudded in July 1965 (Figure 2.1). The primary objective of this well was to test the potential of a seismic "build-up" anomaly originally interpreted to be a Tertiary carbonate reef, but upon drilling was found to be a sequence of Mid-Miocene pyroclastics (Weeks and Hopkins, 1967). In drilling Bass-1 to a sub-sea depth of 2178 metres, a non-marine fluvio-deltaic sequence of sands, shales and coals was encountered which closely resembled the then recently discovered hydrocarbon-bearing sequence (the Latrobe Group) in the adjacent Gippsland Basin. This led to a surge of exploration activity keeping the fluvio-deltaic sands of the Late Cretaceous - Late Eocene Eastern View Coal Measures as the primary target.

By 1982, 19 wells had been drilled in the basin. Of these 19 wells, significant hydrocarbon shows from sands within the Eastern View Coal Measures were recovered from Bass-3, Cormorant-1 and Pelican-1, -2 and -4. In Bass-3, RFT tests recovered 0.82 cubic centimetres of gas and 80 cubic centimetres of condensate from *L.balmei* sands. In Pelican-1 and -2,

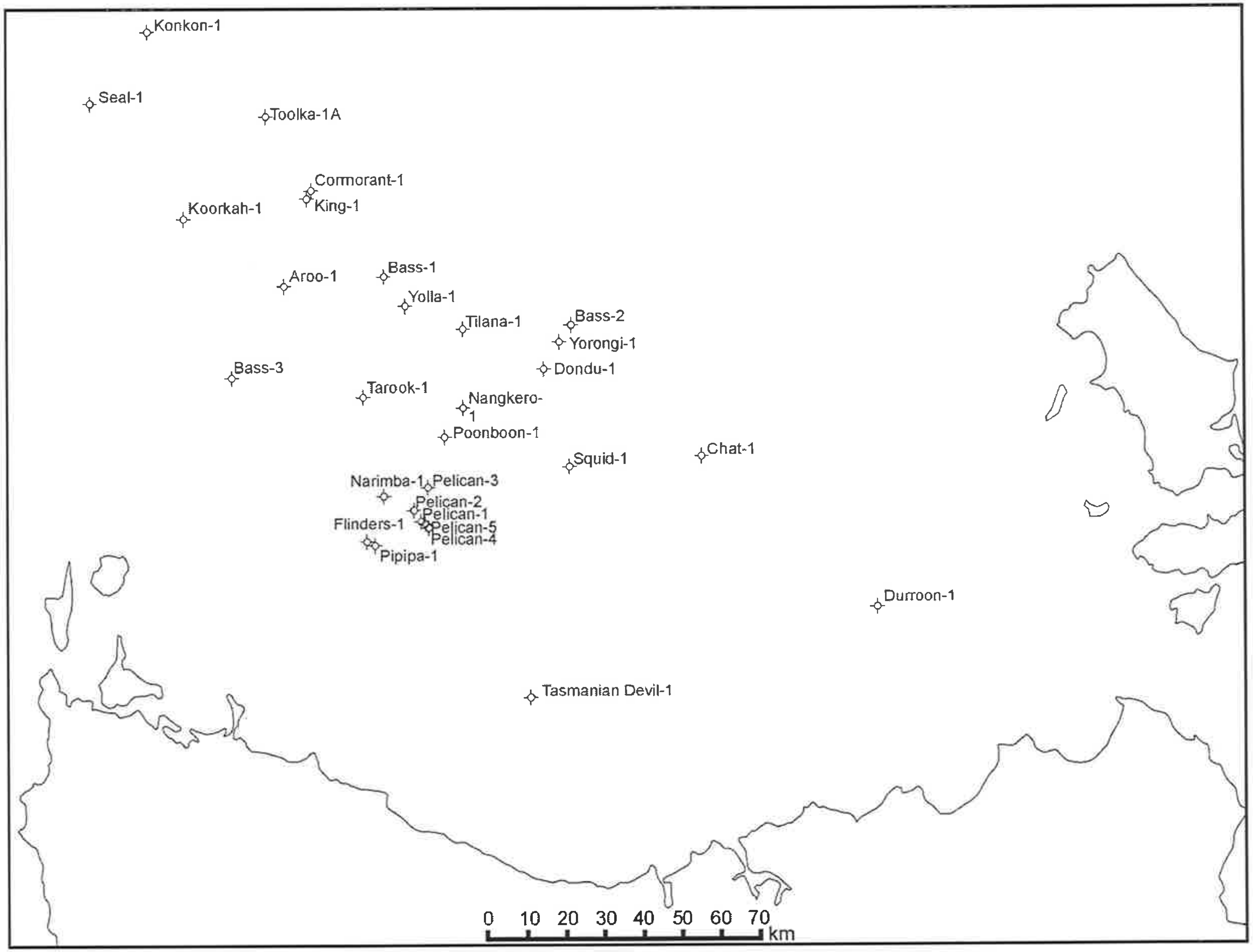


Figure 2.1 Location map showing the distribution of the 29 wells used in the study.

gas and condensate were recovered from *M.diversus* sands at various depths. The largest individual test volumes recovered were 3.9 cubic metres of gas and 6000 cubic centimetres of condensate and 1.05 cubic metres of gas and 750 cubic centimetres of condensate, from the Pelican –1 and –2 wells respectively. At Cormorant-1, 22,000 cubic centimetres of oil was recovered from *N.asperus* sands and minor gas from deeper zones (Brown, 1976).

Between 1982 to 1998, 12 more wells were drilled by subsequent operators together with a large amount of seismic data acquired over the basin. There were additional oil/gas shows found in some of these wells, viz. Poonboon-1, Tilana-1 and Yolla-1, apart from the confirmation of the sub-economic Pelican gas field. Of all these, Yolla-1, drilled to test an upper EVCM structural prospect in 1985, has been the most successful prospect, resulting in significant hydrocarbon recoveries. Here, a drillstem test at 1834 metres flowed 330,000 cubic metres of gas and 148 cubic metres of 44.4 API (American Petroleum Institute) gravity oil per day from Eocene sands. A further test at 2835 metres flowed 430,000 cubic metres of gas and 92 cubic metres of 51.2 API gravity oil per day from Palaeocene sands (Linder, 1986). Very recently, in 1998, two wells, Yolla-2 and White Ibis-1, were drilled that targetted reservoirs in the Late Cretaceous – Late Eocene Eastern View Coal Measures (EVCM) (Lennon *et al.*, 1999). Both wells encountered gas columns in the Palaeocene to Lower Eocene section of the EVCM (Intra EVCM). The estimated gas resource in the Yolla Field is about 450-600 BCF OGIP and the gas accumulation encountered in White Ibis is estimated at 85 BCF OGIP (Lennon *et al.*, 1999).

A huge amount of various geoscientific data including aeromagnetic, ground gravity-magnetic, multichannel seismic and subsurface well information have been accumulated over these years by different oil companies operating in the Bass Basin. The deepest subsurface information that is available in the basin is limited to 4168 metres sub-sea (Pelican-5). A number of papers have been published in the intervening years dealing with various aspects of the Bass Basin geology. Weeks & Hopkins (1967), Robinson (1974), Brown (1976), Smith (1986) and Williamson *et al.* (1987) have documented the exploration history, general basin geology and hydrocarbon potential of the basin. The tectonic evolution and structural development of the basin have been discussed by Davidson (1980); Davidson *et al.* (1984); Etheridge *et al.*, 1984 & 1985; Chamberlain (1988); Smit (1988); Young *et al.* (1991); Baillie and Pickering (1991); Brincat (1992); Hill *et al.* (1995); Gunn *et al.* (1997a, 1997b) and Setiawan (2000). Studies relating to subsidence, geothermal history and hydrocarbon maturation have been dealt with by Nicholas *et al.* (1981), Middleton (1982), and Williamson and Pigram (1986). The sedimentological and petrophysical analysis of the Bass Basin rocks has been carried out by many workers (Baillie and Bacon, 1989; Meszoly *et al.*, 1986; Cubitt, 1992; Cox, 1989; Baillie *et al.* 1991) and study on the source rock potential of the Bass Basin rocks has been published by Nicholas *et al.* (1981) and Miyazaki (1995).

Apart from these publications, a number of unpublished reports have been written by personnel from operating companies or by independent consultants at the request of the companies, many of which are available to the public under the terms of Petroleum Search Act. These include petrological work, sedimentological work, petrophysical work, geochemical evaluation work, palaeontological work, gravity-magnetic interpretation work and aeromagnetic interpretation work.

2.3 Tectonic Setting

Modern concepts of plate tectonics and sea floor spreading have revolutionised the understanding about the present day disposition of the continents and how the dynamic earth has evolved. It is now established that the earth is constituted of a number discrete lithospheric plates and that these plates are in constant motion, interacting at their boundaries in three main ways: a) divergent plate motion, i.e., plates moving away from each other (creating new ocean floor in between); b) convergent plate motion, i.e., plates moving towards each other (either collision between plate boundaries and/or subduction of one beneath the other); and c) transform movement between plate boundaries, i.e., plates moving past each other in a parallel sense without any consequent accretion or consumption of crust between the plates (Vine and Matthews, 1963; Wilson, 1965; Le Pichon, 1968; Issacks *et al.*, 1968; Heirtzler *et al.*, 1968; Dietz and Holden, 1970). The origin of sedimentary basins is intricately related to global plate tectonics. The classification and grouping of sedimentary basins is often done on the basis of their plate tectonic origin and their tectono-sedimentary history (Dickinson, 1975; Klemme, 1980; Kingston *et al.*, 1983). Continental drift is a necessary consequence of plate tectonics in that the continents would be passively rafted on the backs of the conveyor-belt-like crustal plates. The drift of the continents may be conveniently thought of as a summation of sea-floor spreading plus any over-all motion of the entire conveyor belt itself (Dietz and Holden, 1970).

Many authors have tried to reconstruct the universal continent of Pangaea as it would have been in the Permian. Early workers viz. Le Pichon (1968) regarded only six plates to explain global geotectonics whereas Dietz and Holden (1970) considered nine major plates to explain the reconstruction. Since then, more precision palaeomagnetic and radiometric data have been obtained to better delineate the plate boundaries, their relative motion in terms of magnitude (cm/year) as well as their vector direction of motion. Currently, the earth is considered to be constituted of at least 17 plates including major and minor plates (Lowman, 1981). According to a slightly different school of thought, Pangaea is supposed to be the amalgamation of two supercontinents of Gondwana and Laurasia during the Palaeozoic.

The Bass Basin is a product of pre-breakup continental extension in East Gondwana consisting of India, Antarctica and Australia. In terms of location, the Bass Basin is considered to be part of the southeastern end of the Australian Plate. Some authors have considered the Tasmanian sub-plate as a separate entity during the major part of the evolution of the region that helped formation of the Bass Basin (Carey, 1970; Davidson, 1980).

The present rift margins of Australia result from the break-up of India, Antarctica, Australia and New Zealand, Chatham Rise and the Campbell Plateau. Rifting began on the northwest margin and propagated in an anticlockwise direction around Australia to the northeast margin through five separate episodes of sea floor spreading (Falvey and Mutter, 1981). Consequently, in many of the basins, a simple Atlantic-style tectonic model in which a single syn-rift extensional phase is followed by a single post-rift thermal subsidence phase, is complicated by the effects of more than one rifting and drifting event. The timing of ocean floor initiation around the Australian margins is now reasonably well dated (Smith, 1986). For the southern margin, Cande & Mutter (1982) have proposed an age of Middle Cretaceous (100 Ma) and for the southeastern margin, Shaw (1978) has proposed Late Cretaceous age (80 Ma).

The onset of sea floor spreading (breakup) was preceded by extensive sedimentary basin subsidence with evolution of rift-grabens and syn-rift sequences. Smith (1986) has given a summary of the tectono-sedimentary history of the Bass Basin suggesting that the basin has evolved through two episodes of rifting and subsequent drifting; the Otway rifting and drifting followed by a rifting episode related to Tasman Sea opening). Though the Bass Basin is considered to be a failed rift, the rifting stage evolution of this basin is supposedly linked with the initial rifting and drifting of the Otway Basin Margin. A substantial amount of work has been carried out on the generalised tectonic setting of this region and the following is a brief description of the evolution of the continental margin in the surrounding region of the Bass Basin to give a broad contextual framework.

2.3.1 Pre-breakup extension and patterns of sea floor spreading

Courtillot and Vink (1983) suggested that the region into which the rift propagates ahead of continental breakup, must undergo some extension. Also, he suggested that the breakup would be diachronous along the margin, so that the oldest spreading anomaly adjacent to the margin would be progressively younger in the direction of the rift propagation. Since progressively younger magnetic anomalies are truncated to the east along the Australian and Antarctic margins, Mutter *et al.* (1985) proposed that the rifting of Australia and Antarctic progressed from west to east, as was earlier suggested by Boeuf and Dost (1975) using stratigraphic arguments. Mutter *et al.* (1985) also noticed that there is a west to east migration in the time at which initial subsidence ends and thermal subsidence begins from the subsidence patterns in the wells in the sedimentary basins along the southern margin. The thermally-controlled phase of subsidence is referred to as the drift stage of margin development, and the beginning of the drift stage coincides with the beginning of oceanic crust emplacement. They observed a west to east variation in the inferred time of changeover, from fault controlled (rift) to thermally controlled (drift) subsidence which is also compatible with the observation that the oldest magnetic lineations adjacent to the Australian and Antarctic margins show an eastward truncation that can also be explained by a west to east propagation of Australia-Antarctica breakup. While attempting to reconstruct East Gondwana before breakup to present day time, Powell *et al.* (1988) noted that there was considerable amount of pre-breakup continental extension. The probability of large Early Cretaceous continental extension in the Bass Basin was first noted by Etheridge *et al.* (1984, 1985) who interpreted seismic reflection profiles as being consistent with 60 – 100 % of crustal extension on a north-northeast azimuth. Using gravity and seismic profiles and estimates of crustal thickness between Tasmania and Victoria along a seismic profile (Johnson, 1973), Veevers and Ettreim (1988) showed that there was as much as 133 km of crustal extension coast-to-coast in the Gippsland and Bass Basins, and, further west in the vicinity of 130 degrees East in Australia, there was as much as 360 km of continental extension in the paired Australian and Antarctic continents reconstructed to their 96 Ma break-up position (Figure 2.2).

The rather large amount of continental extension prior to breakup is known to be confined to the Middle Jurassic (160 Ma) – Middle Cretaceous (96 Ma) interval. From the concentration of fission track ages of sphene, zircon and apatite in the volcanogenic fill of the Otway Basin in the range 130-96 Ma (Gleadow and Duddy, 1981), and from extensional faulting on the southern margin of Australia in the Early Cretaceous (Fraser and Tilbury, 1979; Williamson *et al.*, 1987), 22 % of the total extension has been apportioned to 160 – 132.5 Ma interval, and the rest to 132.5 – 96 Ma interval (Powell *et al.*, 1988).

Powell *et al.* (1988) have published the reconstructions showing the fragmentation of East Gondwana, first by continental extension and then by growth of the Indian and Southern

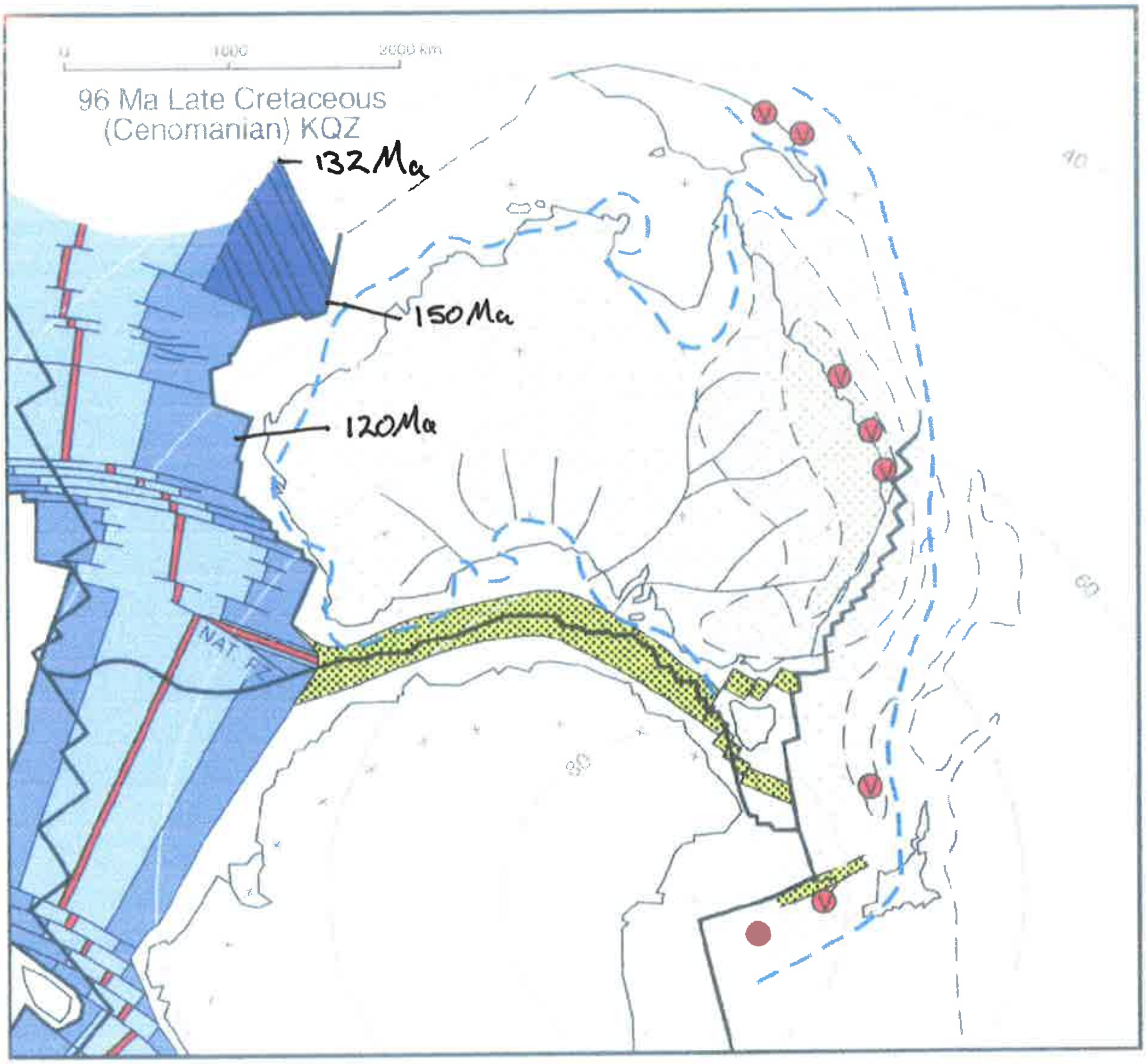


Figure 2.2 Reconstruction of the Australian plate margins during 96 Ma (After Veevers et al., 1991). The heavy line shows the start of the second phase of spreading in the Indian Ocean and of spreading in the southeast Indian Ocean and the Tasman Sea. The yellow stippled area is the mid-Cretaceous rift system. The closer stippled areas of eastern Australia mark uplifts. The broken black line indicates restored position of the areas outlined by the present 2 km isobath.

Oceans. Veevers *et al.*, (1991) reviewed sea floor spreading patterns around Australia. The reconstruction of Australia by Veevers (1991) for 96 Ma age (as shown in Figure 2.2) marks beginning of oceanic spreading in Tasman Sea at 96 Ma but most of the published literature (viz. Weissel and Hayes, 1977) have referred the Tasman Sea opening to be around A34 magnetic lineation which is at around 82-85 Ma. The present day arrangement of the plate margin is shown in Figure 2.3.

2.3.2 Subsidence and Thermal History of the basin

Middleton (1982) attempted to model the thermal subsidence history of the Bass Basin using three different approaches, viz. i) Thermal contraction of the lithosphere; ii) Crustal

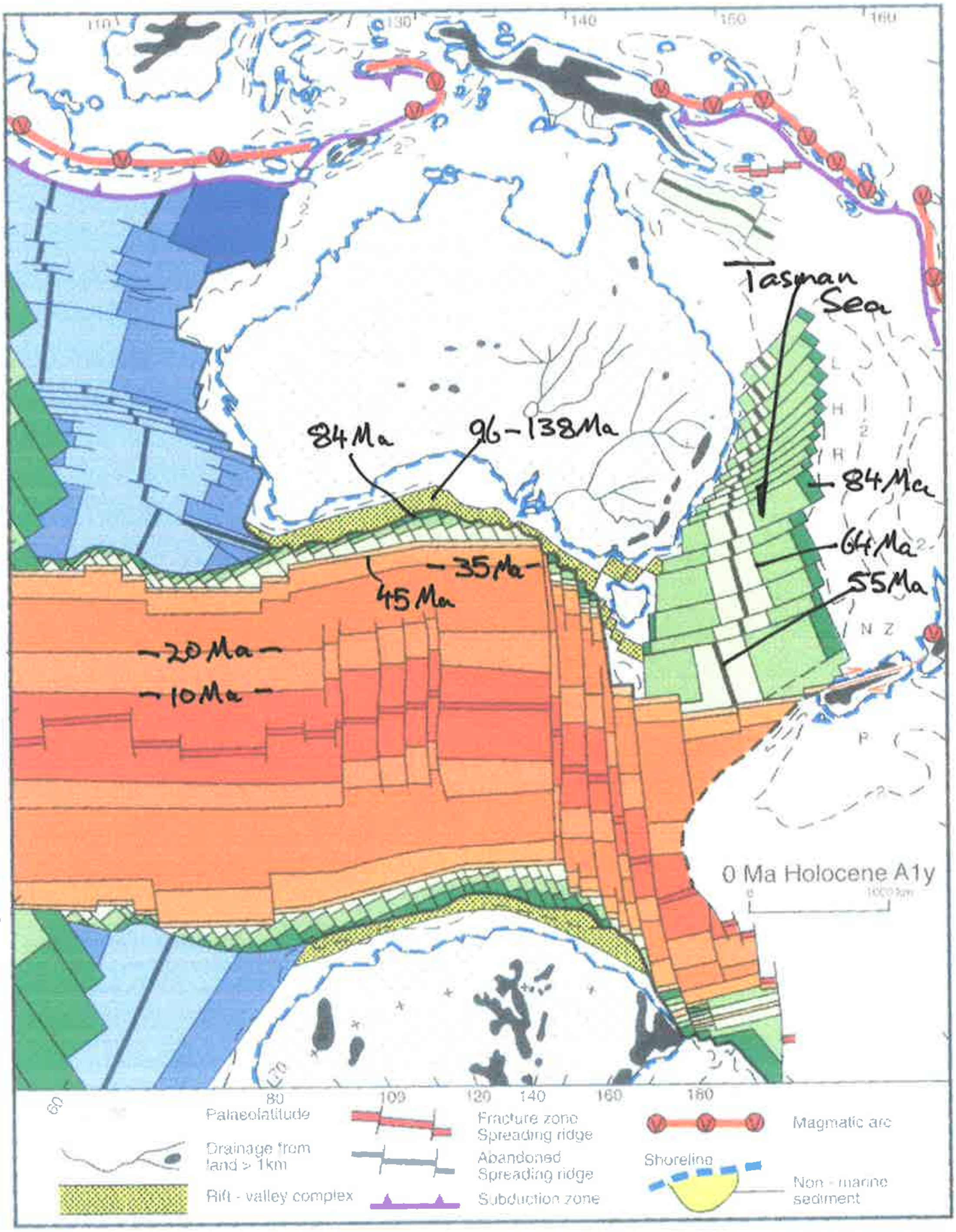


Figure 2.3 Plate reconstruction at present day (After Veevers et al., 1991). Plate divergence in the south and plate convergence in the north. LHR-Lord Howe Rise; NZ P-New Zealand Plateau.

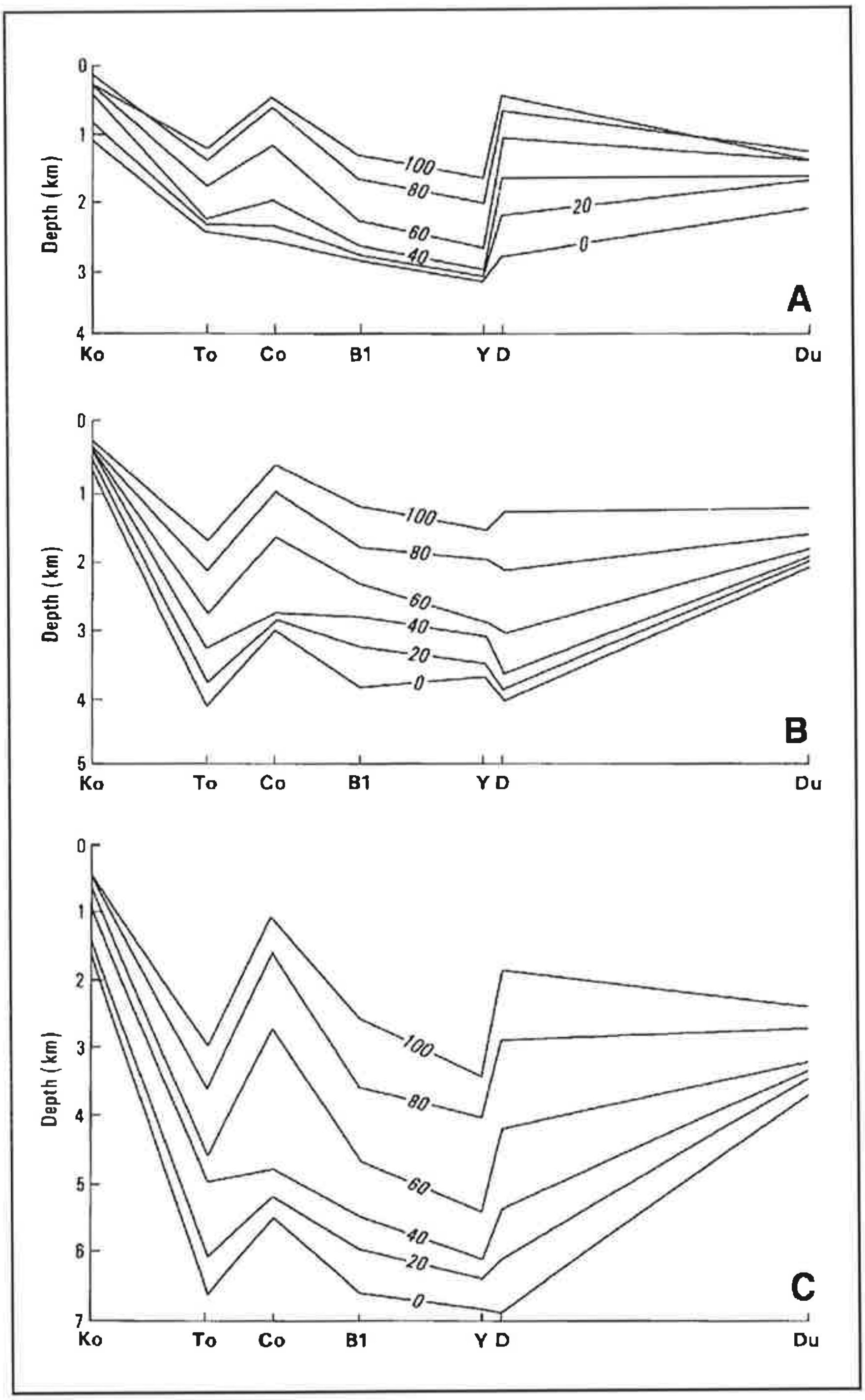
or lithospheric stretching; iii) Phase change mechanism. He concluded that continuous volcanism and basin subsidence observed in the basin could be best explained by the phase changes associated with deep crustal metamorphism. The main pointer to such a conclusion was derived from the fact that a cooling thermal history belies the continuous volcanism throughout the Tertiary.

Williamson and Pigram (1986) carried out geohistory analysis of 16 wells from the Bass Basin to investigate the thermal and subsidence history of the basin in relation to its depositional history, structural formation and maturation history. The geohistory plots of these wells fall into three groups comprising those wells sited over depocentres, basin margins or basement highs.

Group 1 wells situated over depocentres such as Dondu-1, Narimba-1 and Pelican-2 (Figure 2.1) show a pattern of high initial subsidence that lasts for up to 50 Ma before slowing to a rate that appears to be continuing at present. For Group 2 wells such as Bass-2, Bass-3, Durroon-1 and Konkon-1 located on the basin margins, the onset of subsidence is much later and rates of subsidence lower than for Group 1 wells. The wells situated over the basement highs (Group 3 wells) tend to be intermediate in character between Group 1 and Group 2 wells in having low initial rates of subsidence which then increase to rates comparable to those of the later stages of the Group 1 wells.

The reasons for this variable subsidence behaviour are related to the relative effects of sediment loading versus basement subsidence component at each site. Figures 2.4 and 2.5 are plots of the basement subsidence for two basin-wide cross sections. Figure 2.4 corresponds to a basin axis section extending from Konkon-1 to Durroon-1 while Figure 2.5 corresponds to a SW-NE section extending from Bass-3 to Bass-2. The effect of the sediment load at each site is calculated from the matrix weight of the sediment column (Part B of Figure 2.4 and 2.5). This produces a sediment loading curve through time which is then extracted from the total subsidence curve to give the thermal driven mechanism for the basement (Part A of Figure 2.4 and 2.5).

The relative effects of the basement subsidence and sediment loading component in basin subsidence are clearly illustrated in Figure 2.5 where Narimba-1, a Group 1 Well, and Bass-3, a Group 2 well, are juxtaposed. The total subsidence for the Bass-3 well is approximately twice that of the thermal component. They suggested that the basement highs initially subside in a manner comparable to the basin margins, that their subsidence is largely thermally driven. But having subsided to a point where these basement highs begin to be buried, sediment loading then assumed an important role and they then follow a subsidence regime similar to the Group 1 wells. Initial high rates of subsidence that characterise the Group 1 wells are caused by a combination of a high basement subsidence component following the rift phase of basin development and high rates of sediment input (the average sedimentation rate for the lower part of the Eastern View Coal Measures Group was approximately 250 m / m.y).



Wells in the figure:

Ko:-Konkon-1
To:-Toolka-1A
Co:-Cormorant-1
B1:-Bass-1
Y:-Yolla-1
D:-Dondu-1
Du:-Durroon-1

Figure 2.4 Basement subsidence plots for a basin axis cross section from Konkon-1 to Durroon-1 from 100 Ma to present in 20 million year increments (For location of wells, see Figure 2.1). Part A is the water loaded basement thermal cooldown subsidence. Part B is the sediment loading component of the basement subsidence and Part C is the total subsidence for the basement (A+B) (After Williamson and Pigram, 1986)

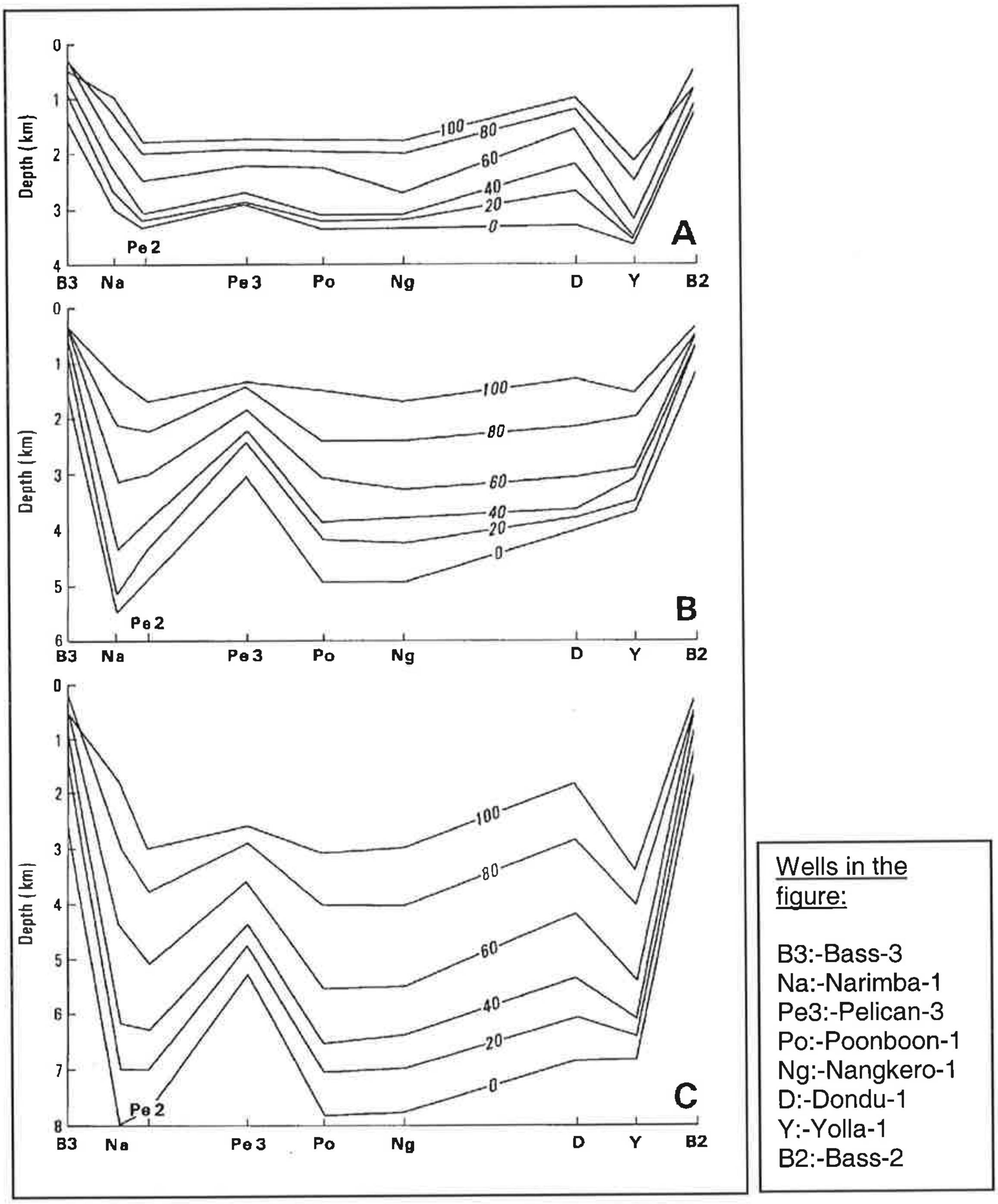


Figure 2.5 Basement subsidence plots for basin cross section from Bass-3 to Bass-2 (For location of wells, see Figure 2.1) Part A is the water loaded basement thermal cooldown subsidence. Part B is the sediment loading component of the basement subsidence and Part C is the total subsidence for the basement (A+B) (After Williamson and Pigram, 1986).

Although Williamson and Pigram (1986) recognise that the Bass Basin was initiated as a result of crustal rifting between Australia and Antarctica, they have failed to appreciate the tectonic subsidence caused by normal faulting during the crustal extensional phase in the basin. They have ascribed the high initial rates of subsidence to be the result of a mechanism predominantly driven by sediment loading. This study, as discussed in detail in Chapter-4, shows that the subsidence in the Durroon and Bass area during the rifting stage was driven by tectonic faulting though the effect of sediment loading was more pronounced in the Bass area than in the Durroon area, particularly after initial subsidence created by normal faulting with large displacement.

2.3.3 Flexural modelling

Setiawan (2000) has carried out flexural isostatic modelling to quantify the lithosphere behaviour in the Bass rift system by examining the stratigraphic response to continental extension. During extensional tectonics, the kinematic response of the lithosphere to deformation generates changes in the distribution of mass as a result of upper crustal faulting, lower crustal thinning and changes in the thermal structure of the lithosphere. The changes in mass distribution induce an isostatic response that controls the basin geometry, uplift and subsidence history, and the topographic slopes that control sediment systems.

Setiawan (2000) suggested that the Bass Basin rift system consists of a deep, narrow asymmetric half-graben system in the Durroon area (Durroon Sub-basin of Setiawan) and widens to the Bass area (Northern Bass Sub-basin of Setiawan), with a polarity reversal across a major 'Pelican-Squid transfer zone'. Flexural cantilever modelling conducted on both sub-basins revealed a larger crustal stretching in the Northern Bass Sub-basin compared to the Durroon Sub-basin. The average extension in the Durroon Sub-basin is only β=1.1, lower than the β=1.5 in the northern Bass Sub-basin. Setiawan tried to explain the different geometry in the two sub-basins from this difference in the value of β. From forward and reverse modelling in the two sub-basins, it was suggested that ultimate basin geometry of the Bass Basin is controlled by several aspects, viz. (i) fault geometry, including dip, (ii) the amount of extension, (iii) flexural rigidity and brittle seismogenic layer thickness, and (iv) multiple fault interaction, with corresponding load distribution. A best fit of observed and calculated basin geometry in seismic section BMR88-306 (Figure 4.1) was achieved using a T_e value of 1.75 that is relatively low and substantially less than the thickness of the cool, brittle upper crust (10 km in the model). The relatively low T_e value may suggest that flexural bending stresses of planar fault extension are sufficiently large enough to generate brittle failure within the upper crust. Therefore, it is not surprising to observe in seismic sections that brittle faulting is prominent in the Durroon Sub-basin. In contrast, brittle failure in the northern Bass Sub-basin is ambiguous. The best fit T_e for northern Bass Sub-basin, using seismic section TQH-49 (Figure 4.16), is 5 km with a 15 km brittle seismogenic layer thickness.

As will be discussed in Chapter-4, the extension estimates, denoted by stretching factor, derived by Setiawan are in conflict with those obtained by preliminary estimates of extension amount carried in this study by measuring the initial basement length between two points and the corresponding extended length after faulting-induced stretching.

2.4 Structural evolution

Though many authors have discussed some aspects of the structural history of the basin (Carey, 1970; Griffiths, 1971; Elliot, 1972; Falvey, 1974; Robinson, 1974; Boeuf and Dost, 1975; Brown, 1976; Davidson, 1980, Davidson *et al.*, 1984; Smith, 1986; Smit, 1988), Etheridge *et al.* (1984 and 1985) have investigated in some detail, the structural style of the basin in a regional scale using deep seismic information and classified the structures as per the different stages of basin development. Similarly, the most detailed study on structural model for the Late Tertiary has been carried out by Chamberlain (1988).

There is general agreement amongst these workers that the basin resulted in some way from the rifting between the Australian and Antarctica plates which began about the Late Jurassic - Early Cretaceous (140 Ma). Etheridge *et al.* (1984 and 1985) have elaborated the details of the structural style of the basin. According to them, the Bass Basin is considered to be one of the three Bass Strait Basins that resulted from contemporaneous extension or rifting. Two major sets of faults were developed during this stage – rotational normal faults and an orthogonal set of near-vertical transfer faults. The geometry of these structures indicates that the sub-basinal crust underwent extension of 60 to 80 per cent, prior to a thermal subsidence phase. Shallow-dipping normal faults bounding major Early Cretaceous half-grabens are seen in the basin, similar to the Otway and the Gippsland basins. While mapping the normal faults associated with the early extension phase, these authors noted the short strike extents of the normal faults and their abrupt termination against and displacement by a second set of extensional structures, namely the transfer faults.

Following the work of Bally (1982) and Gibbs (1984), Etheridge *et al.* (1984 and 1985) proposed that these transfer faults are not simple strike-slip faults, but are basically accommodation structures analogous to oceanic transforms, in that they allow variations in the geometry of extension along strike of a rift. Major normal faults can terminate against them (and vice versa) without the necessity for distributed strain throughout the rock mass. Etheridge *et al.* (1985) have noted that the normal faults were found to terminate abruptly, and be apparently displaced across transverse zones that spanned the width of the basin. They also proposed that major fault-bounded half-grabens are consistently offset by the transfer faults. Etheridge *et al.* (1985) also noted that the characteristics of these transfer faults are generally steep to vertical, and trend at high angles to the extensional normal faults. The location, spacing and even dip direction of the normal faults may change from one side of a transfer fault to the other.

The transfer fault mechanism in the structural model as proposed by Etheridge *et al.* (1984, 1985) seems to be quite exaggerated and without sufficient data support. This concept may have been derived in part by the wide spacing between the regional seismic lines. The interpretation work carried out and discussed in Chapter-4 does not lend credence to the rigidly followed orthogonal transfer fault system. Although they observed that transfer faults are deep-seated vertical fracture zones that provide potential paths for magma (and other fluid) migration and are therefore commonly the locus of concentrations of volcanic and high-level intrusive bodies, they have failed to provide any matching evidence.

Etheridge *et al.* (1985) classified another set of structures, the subsidence stage structures, different from the extensional stage structures based on the observation that they have a different gross geometry than the rift stage structures because the displacement field is near vertical rather than horizontal and also substantially less in magnitude. A third category of structures observed in the Bass Strait basins were considered to have been formed during the tectonic overprint stage, indicating a marked change from an extensional stress regime to a compressional stress regime. Many inverted structures have been observed in the eastern Otway & Bass Basin viz. Snail, Nerita, Konkon and Cormorant. It is however, characteristic that the reactivation is more or less confined to the northern portion of the basin. The magnitude diminishes considerably to the southern end (Etheridge *et al.*, 1985). A similar effect has also been observed in the Gippsland Basin where the intensity of reactivation reduces southwards. Although Etheridge *et al.* have noted little reactivation of the major normal faults and transfer faults in the central and southeastern part of the Bass Basin, the present study observed some clear cut inversion structures in the southeastern Bass Basin (Durroon area) as discussed in detail in Chapter-4.

Robinson (1974) and Brown (1976), who also studied the structural history of the basin, suggested the major structural trends to be aligned in the northwest-southeast direction. The predominant structural styles include: a) basement-involved fault blocks with onlap and differential compaction of sediment over the upthrown blocks; b) faulted anticlines; c) arches or folds caused by intrusive and extrusive igneous rocks. More importantly, they proposed a sequential development of the structures in the basin with respect to geological time. According to Robinson (1974), the structures in the southeastern area were developed in the Early Cretaceous with little or no structuring observed in this area in the Tertiary. On the other hand, the central and the northwestern areas show later structural development, mostly in the Tertiary. The structures include the Bass-3 faulted basement block which shows structuring throughout Miocene and the Early Eocene structures near Pelican-1, Ponboon-1, Bass-2 and Dondu-1. On the contrary, the Cormorant structure, which developed in post-Eocene time, was actually a pre-

Oligocene trough or depocentre that was uplifted during the Oligocene and Miocene, a prominent feature noted by many other authors (Chamberlain, 1988).

Moore *et al.* (1984) proposed that the Boobyalla Plains area in the northeastern Tasmania constitute the onshore extension of the Bass Basin. Based on gravity-magnetic data, lithofacies of the sections from two shallow wells (Boobyalla -1 and -2) and the biostratigraphic information, they suggested that the sediments encountered in these two wells were actually the proximal (near source) facies equivalent of the Eastern View Coal Measures found in the Durroon-1 well in the southeastern part of the basin. The sediments are dolerite boulders, pebble conglomerate and poorly sorted ferruginous sandstones deposited in a steep, fault-bounded trough of Cretaceous age. The structural style is characterised by NW-SE, N-S and NE-SW trending faults in the area.

On the contrary, Baillie and Pickering (1991) proposed a completely different tectonic and structural evolution during the rifting stage for the southeastern area of the Bass Basin and named it the Durroon Basin. They suggested that the two basins, though initially evolved independently, started to act as a single basin in the thermal subsidence phase. Brincat (1992), however, considered that the Durroon area to be a sub-basin of the Bass Basin and the Boobyalla, Bark and Anderson blocks to be separate half-grabens in the Durroon Sub-basin. As discussed in detail in Chapter-7, the Durroon area and the Bass area have been designated as Durroon and Burnie Sub-basin respectively and the Bark, Anderson and Boobyalla half-grabens (Bark, Anderson and Boobyalla Sub-basins of Baillie and Pickering, 1991) of Brincat (1992) have been defined as Bark, Anderson and Boobyalla fault blocks.

Chamberlain (1988) has made a very detailed study of the post Late Eocene structural development of the Bass Basin based on a large seismic data base and other relevant data spread over the entire basin. The basin, according to him, has undergone a complicated tectonic history while sagging in the Late Tertiary. The Late Tertiary structural features followed the earlier developed structural grain of the basin and the rejuvenation of pre-existing structures and the structural inversion of several features are important factors in the later stage structural development of the basin.

Two types of structural provinces have been interpreted by Chamberlain (1988) on the basis of structure contour mapping and fault trend analysis. The southeastern and the northwestern portions of the basin were mainly influenced by extensional tectonics with normal faults dominant whereas in the central portion of the basin, more complex structural features, including reverse and wrench faults associated with compressional tectonics, dominate the area. Of the two major periods of Late Tertiary tectonic activity, the Late Oligocene period experienced basin-wide structural development whereas the Miocene tectonic activity has been more restricted to the northern half of the basin. Anticlinal closures have been observed at the Mid Miocene level (for example Cormorant structure) apart from the Top Eastern View Coal Measures Group (Late Eocene) and Top Oligocene level which are the periods affecting the southern half of the basin (for example in the Pelican field).

2.5 Stratigraphic Framework

Four major lithostratigraphic units are recognised in the Bass Basin; the early Cretaceous Otway Group, the Late Cretaceous to Late Eocene Eastern View Coal Measures, the Late Eocene Demon's Bluff Formation and the Oligocene to Recent Torquay Group. Some authors have tried to draw a comparison between the stratigraphic columns of the three Bass Strait basins, namely, the Otway, Bass and Gippsland Basins (Williamson *et al.*, 1987; Davidson and Morrison, 1986). Earlier work by Weeks and Hopkins (1967), Robinson (1974), Brown (1976) and Smith (1986) have also dealt with the Bass Basin stratigraphy. Essentially, the rock units have been biostratigraphically sub-divided based on the spore pollen assemblage zones for the non-marine Early Cretaceous to Late Eocene succession and planktonic foraminiferal zonation for the Late Eocene to recent marine section (Taylor, 1971; Partridge, 1979 and Morgan, 1987).

The stratigraphic column of the Bass Basin is shown in the Figure 2.6 and a comparative chart of the stratigraphic columns of the three Bass Strait Basins is shown in Figure 2.7. Though most of the basin lies underneath water, the surface exposures in the surrounding regions viz. the Otway Ranges of the Torquay Embayment, King Island, Flinders Island and northern Tasmania provided a good support for the understanding of the rock types that may lie in the Bass Basin subsurface, particularly the basement rocks and the pre-rift stratigraphy. A more detailed description of the different lithostratigraphic rock sequences is given below:

Basement

Information from deeper depth in the basin is very limited. Of the 32 wells, only 8 have reached the Cretaceous and only 2 wells have encountered basement so far (Williamson *et al.*, 1987).

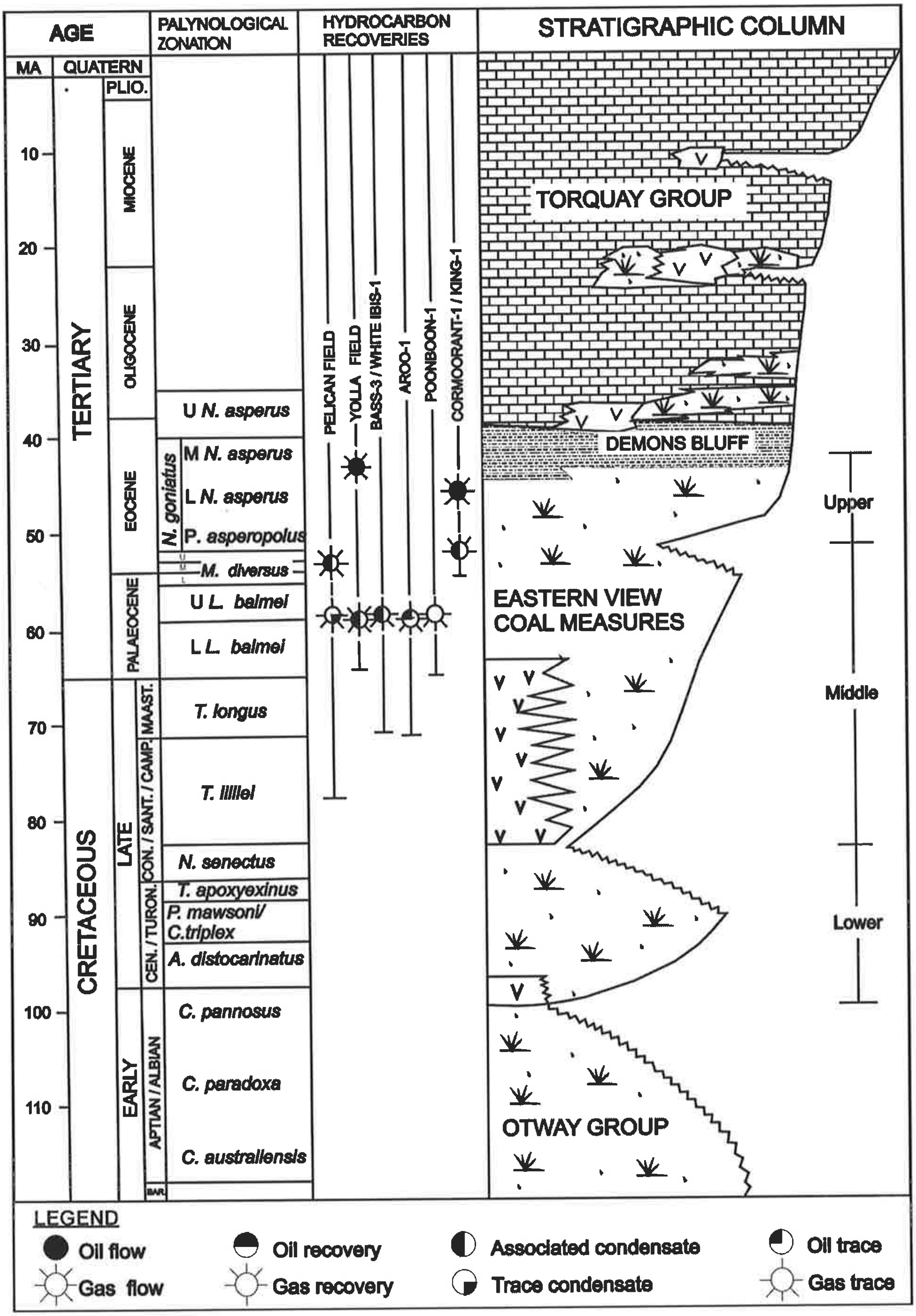


Figure 2.6 Generalised stratigraphy of the Bass Basin (After Lennon et al., 1999).

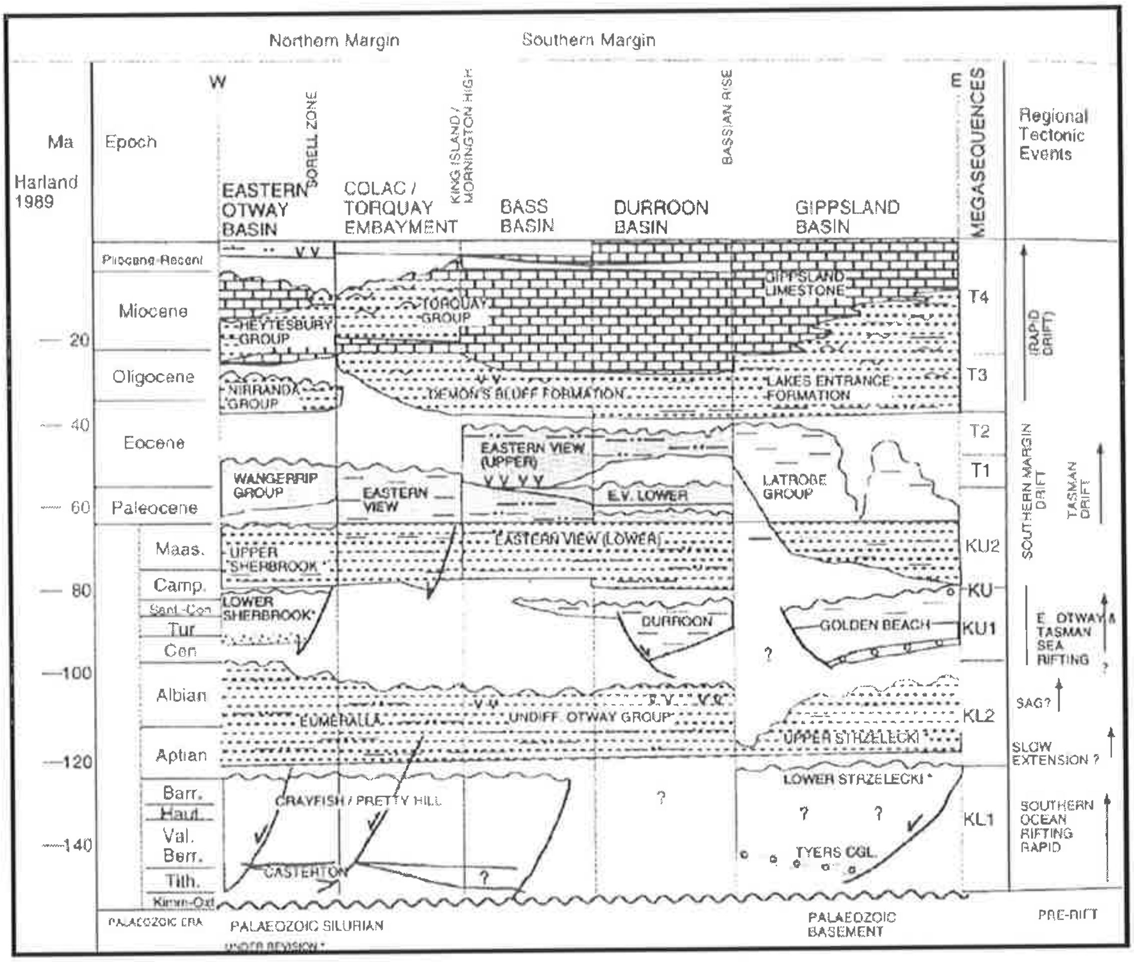


Figure 2.7 Comparative stratigraphic chart between Eastern Otway Basin, Colac Trough, Bass Basin, Durroon Basin and Gippsland Basin (After Hill et al., 1995).

Basement for the Bass Basin is considered to be the rocks of Late Jurassic age or even older. Along the eastern margin of the basin, basement outcrops on the islands associated with the Bassian Rise consist predominantly of the lightly metamorphosed Early Palaeozoic Mathinna Beds and Devonian to Carboniferous granites which belong to the Lachlan (Tasman) Fold Belt and have a dominant north-south structural alignment. Along the western margin of the basin, basement outcrops on the islands associated with the King Island Rise and consists predominantly of Late Proterozoic to Early Paleozoic metamorphics and metasediments (Spry and Banks, 1962). The Precambrian rocks which outcrop on King Island and along the west coast of Tasmania may also subcrop on basement highs within the basin.

Within the basin proper, basement has been encountered in two wells, Bass-2 and Bass-3. In Bass-2, located on the shoulder of the eastern margin of the basin, basement is recognised as highly fractured tuffaceous mudstones and has been radiometrically dated as Early Palaeozoic, possibly as old as Cambrian (Brown, 1976). In Bass-3, located on a basement high near the western margin of the basin, basement rocks in core include interbedded quartzite, recrystallised siltstone, fine grained sandstone and lightly metamorphosed shales (Esso, 1967; Brown, 1976). It is also suggested that Permian and Triassic sediments known from mainland Tasmania may underlie the known Bass Basin sediments, but no well has been drilled through such a section. Extensive Jurassic basalts, dolerites and volcanolithic sediments are extruded over Tasmania and may occur beneath the Bass Basin.

The basement rocks are unconformably overlain by the first order east-west to northwest-southeast oriented Cretaceous to Recent sequences as in the other rift basins in southeastern Australia. These rift sequences cut across the previous north-south structural trends. The angular unconformity between the pre-Jurassic sequences and the Lower Cretaceous sequences marks a major tectonic event in the geologic history of south-eastern Australia and is taken here to establish the beginning of the Bass Basin (Smith, 1986).

Rift onset was preceded by the Jurassic volcanics and may have given rise to uplift and erosion of the pre-rift sequences due to thermal expansion prior to and in the early stages of rifting, depending on the crustal thickness (Smith, 1986).

Otway Group

Rocks of the Early Cretaceous Otway Group are the oldest known sediments of the Bass Basin. They have been encountered in two wells within the basin, Konkon-1 and Durroon-1, but are well known from the Torquay Sub-basin. In Konkon-1, located near the northern margin of the basin, 40 metres of Otway Group sediments were encountered. The unit consists of undated lithic sandstone, claystone, siltstone, weathered volcanics, and minor coal, which was assigned to the Otway Group, with the Otway Group classification being inferred via lithological comparison with the Otway Group sediments of the Torquay Sub-basin (Brown, 1976; Williamson *et al.*, 1987). However, in the absence of any palynological evidence, the assigning of the Otway Group to the 40m section of the Konkon-1 could be wrong. The seismic evidence as interpreted in this study also does not agree with the observation as noted by Brown (1976) and Williamson *et al.* (1987).

In Durroon-1, located on a basement fault block in the southeastern corner of the basin, over 1400 metres of Otway Group sediments were encountered. Here, the sediments consist of rare conglomerates, immature lithic sandstones, minor interbeds of siltstones, calcareous shales and rare minor coals (Brown, 1976). Overlying the Otway Group sediments in this well, a 100 metre of undated and highly altered vesicular olivine basalt was encountered. The sandstones are typically fine to coarse grained, poorly sorted and highly immature. The upper 16 m of the sequence in the Durroon-1 well is of Late Albian age (C. *pannosus* zone) and the oldest palynological age obtained for the Otway Group sediments in Durroon-1 is Late Aptian (*C.hughesi* zone)(Brown, 1976).

Within the Bass Basin, the depositional environment for the Otway Group sediments is poorly understood due to the limited sampling of this unit. However, the Otway Group of the Otway Basin (Ellenor, 1976) and the corresponding equivalent Strzelecki Group of the Gippsland Basin (Threlfall *et al.*, 1976) have been more thoroughly investigated, and it is generally thought that the Otway Group sediments of the Bass Basin were deposited under similar conditions. Thus it is interpreted that the Otway Group sediments of the Bass Basin represent rift fill alluvial fan, alluvial plain and lacustrine complexes. Otway Group sediments are not recognised from Bass-2 or Bass-3.

Smith (1986) has pointed out that the Otway Group sediments have undergone considerable burial diagenesis. The lithic components are labile, showing physical and chemical breakdown. Cementation and overgrowth by quartz, carbonate, felspar, zeolites, chlorites and other clays are common. Consequently, the sands rarely show good visible porosity, and measured porosities and permeabilities for good sands are mostly in the range 8-20 % and 0-100md respectively, while the majority of the sandstones have poorer reservoir properties.

On the basis of Durroon –1 well data and nearby seismic data around this well, Smith (1986) has interpreted about 171 m of drilled section (between 1374m and 1545m) in Durroon -1 as due to the second episode of crustal extension related to the Tasman Sea Rifting phase and is dated as belonging to the *C.tiplex* palynological zone with the older *A.distocarinatus* and the overlying *T.apoxyexinus* zone absent. Lithologically, the unit consists of monotonous non-marine, dark-grey, slightly carbonaceous mudstones or shales and this sequence has been named as the Durroon sequence by Smith (1986). The apparent dominance of the argillaceous sediments and lack of arenaceous facies in this Durroon sequence suggests that the underlying Middle Cretaceous volcanics blanketed much of the Durroon area and possibly much of the Bass Basin. The volcanics formed the source of the fine-grained sediments in this sequence with only minor erosion of exposed pre-rift and Otway Group acting as a source of arenaceous detritus.

As has been discussed in Chapter-4, the lithologic description of the entire Durroon Formation as mudstone-dominated by Smith (1986) has been disputed by Baillie and Pickering (1991) and the wireline log character of the Durroon-1 well suggests that the lower part of the Durroon Formation is actually a sandy unit. Hill *et al.* (1995) suggested that during the period from the Cenomanian until Campanian, there was hiatus in sedimentation (their figure 3 and presented as figure 2.7) in the Bass area when during this period, the Durroon Formation was being deposited in the

Durroon area. It is believed that since there is no well penetrating the deeper part of the stratigraphy in the basinal depocentres of the Bass area, that could have intersected the age-equivalent sequences of the Durroon Formation, it is premature to conclude in such a manner.

Eastern View Coal Measures Group

Sediments of the Late Cretaceous to Late Eocene Eastern View Coal Measures (EVCM) unconformably overlie the Otway Group sediments. The EVCM has again been divided into two major lithostratigraphic units at the boundary between the upper and lower *M.diversus* palynological markers (Brown, 1976; Smith, 1986), the Lower and the Upper EVCM.

The lower EVCM consists of interbedded sandstones, siltstones, shales and thin coals. Based on available well data, the sediments become finer grained northwards, with the coarser sands being encountered in Durroon-1 and the Pelican wells and a greater percentage of shale and coal encountered in the Cormorant-1 and Konkon-1 wells. Volcanism was also contemporaneous with the deposition of sediments of this age, as over 400 metres of basalts were encountered in Aroo-1 interbedded with the sediments of *L.balmei* and older ages. Moore *et al.* (1984) described fanglomerate deposits from the onshore Boobyalla Sub-basin located in the northeastern Tasmania plains, which are palynologically equivalent to the lower Eastern View Coal Measures. Robinson (1974) suggested that one of the major sources of sediments was northern Tasmania. The environment of deposition for the lower Eastern View Coal Measures Group is considered to be an essentially fluvial regime, ranging from alluvial fan and braided streams (in the southern Bass Basin) to meander channels and deltaic environments with associated lacustrine and floodplain complexes in the remainder of the basin.

The upper EVCM have a blanket-like distribution over the basin. This sequence is characterised by clean channel sands and thick coal seams (25 metres thick coal seam in Dondu-1). Again, these sediments become finer grained to the north. Robinson (1974) interpreted the environment of deposition as deltas and coal swamps.

All significant oil recoveries have been obtained from sands within the Eastern View Coal Measures Group, specifically, sands within the *L.balmei, M.diversus* and *N.asperus* palynological zones (Figure 2.6).

Brown (1976) has suggested that the Narimba-1 and Pelican-1 areas were Early Tertiary depocentres including the Tarook-1 area which encountered 911 metres of *M.diversus* sediments. A comparison of well data suggests that the Narimba-Pelican-Tarook area was a more significant depocentre in the Early Tertiary than the Poonboon-Nangkero area, which is centrally located in the basin and contains a thicker section of Late Tertiary sediments.

Similarly, the Cormorant area was also an Early Tertiary depocentre. It means that the Poonboon-Nangkero area was more stable than the Cormorant and Narimba-Pelican-Tarook area of the basin during the Early Tertiary. This is supported by the well data which shows a thinner Late Eocene to top Cretaceous interval in the Poonboon-Nangkero area. Furthermore, the cleanest sands in this interval were encountered in these wells (Esso, 1974a) suggesting the stronger stream energies and/or greater reworking of sediments are associated with more stable areas. However, it is not clear whether the Poonboon-Nangkero area was situated over a Cretaceous palaeo-high or whether the Cormorant and Narimba-Pelican-Tarook areas subsided at a faster rate during the Early Tertiary to account for the above observations. It is believed that the structural style of Poonboon suggests that the structure developed by sedimentary drape over a palaeo-high which existed prior to *L. balmei* deposition (Esso, 1974b).

In a recent publication, Lennon *et al.* (1999) further subdivided the EVCM into Lower, Middle and Upper sequences separated by intraformational unconformities or disconformities (see Figure 2. 6). The Lower EVCM, bound by unconformities at lower A. *distocarinatus* and within the N. *senectus* Zone, has not been penetrated in the basin except at Durroon-1 where the sequence has been described as Durroon Formation (Smith, 1986) and is considered to be the stratigraphic equivalent of the Golden Beach Group of the Gippsalnd Basin. Lennon *et al.* (1999) have opined that the Lower EVCM with volcanics acting as sealing unit may provide as a hydrocarbon play akin to that encountered along the northern margin of the Gippsland Basin, for example at Kipper oil and gas field (Lowry, 1987; Sloan *et al.*, 1992). The Middle EVCM is bound below by the intra-N. *senectus* unconformity and above by another at the base of the upper M. *diversus* Zone. Deposition within the T. *longus* and L. *balmei* Zones was controlled by both compaction-induced subsidence and growth faulting. The upper part of this sequence has been penetrated by several wells (White Ibis-1, Bass-3 and Aroo-1) and contains a major gas accumulation at Yolla-1. Regional paleogeographic studies utilising core, wireline log and palynological data suggest the sediments of the Middle EVCM were deposited in a major fluvial system and associated wave dominated delta system prograding into a restricted embayment or large lake (Suttill *et al.*, 1987, Suttill, 1995). The Upper EVCM lies between the unconformity at the base of the upper M. *diversus* Zone and the top of the EVCM (lower to middle N. *asperus* Zone). These sediments accumulated during the sag-phase of the basin evolution and the depositional environment at this level is dominantly upper delta plain with some marine influence at the top of the sequence (Baillie and Bacon, 1989).

Demons Bluff Formation

The Demons Bluff Formation conformably overlies the Eastern View Coal Measures and represents a change in the depositional regime in the Bass Basin from non-marine to marine.

Throughout much of the basin, the Demons Bluff Formation is a sequence of interbedded soft chocolate-brown shales and minor greyish siltstones. These shales and siltstones are fossiliferous, glauconitic, calcareous, pyritic, micaceous and contain sporadic dolomite bands. At Nangkero-1, minor lignite bands are present. Minor sand interbeds are encountered in the wells located near the basin margin (Bass-2 and Dondu-1). At Durroon-1, the Demons Bluff Formation is represented entirely by a white coarse grained marine sandstone.

The environment of deposition for the Demon's Bluff Formation is interpreted to be a low energy restricted shallow marine 'embayment'. This is evidenced by the large number of arenaceous benthonic foraminifera contained within the formation (Esso, 1974c). The marine encroachment which deposited the Demons Bluff Formation originated in the north and spread rapidly southwards (Partridge, 1979). The top sands of the Eastern View Coal Measures, which underlie the Demon's Bluff Formation, have a marine character which probably represents winnowing of the fluvio-deltaic deposits during marine incursion.

To the north, in the Torquay Sub-basin, the Demons Bluff Formation attained a greater thickness (911 ft in Nerita-1), with the contact between the Demons Bluff Formation and the underlying Boonah Sandstone (equivalent to the top EVCM) being more abrupt. Partridge (1976) suggested that the sea level was raised on the adjacent continental shelf to first breach the Cape Otway-King Island High and then still further to breach the King Island-Mornington Peninsula ridge, causing flooding of the low lying topography of the Torquay Sub-basin and the barred Bass Basin.

Torquay Group

Conformably overlying the Demons Bluff Formation are the Oligocene to Recent marine sediments of the Torquay Group. The Demons Bluff Formation and the Torquay Group have been subdivided into eleven foraminiferal zones by Taylor (1971). The K zone contains the Demons Bluff Formation, with the J to A zones sub-dividing the Torquay Group sediments.

The base of the Torquay Group in many wells is represented by a thin, upward-fining sandstone. This light brown sandstone is fine grained, fossiliferous and slightly dolomitic. The sandstone grades into a greyish siltstone (and mudstone) which persists throughout much of the Oligocene section of the Torquay Group. The siltstone is soft, argillaceous, fossiliferous, glauconitic and calcareous. Pyritised shell fragments and worm burrows are common, particularly in the lower Oligocene section and sedimentary bedding features are lacking. Sandstone stringers are common in the wells located near the basin margin.

An environmental interpretation for the Late Tertiary section of the Bass Basin based on the foraminiferal data collected from Aroo-1 suggests that the Eocene-Oligocene boundary is marked by a pulse of planktonic foraminifera, which has been interpreted as a period of oceanic flooding caused by sea level highstands on the adjacent continental shelves. Three such cycles were penetrated in the Oligocene section at Aroo-1. However, for most of the Oligocene time, the environment of deposition is interpreted to have been a low energy, restricted shallow marine regime.

From the beginning of the Miocene to the Mid-Miocene, the sea level continued to rise in the Bass Basin, causing conditions to become more open and higher energy. Under the more open marine conditions, a greater number of planktonic foraminifera entered the basin, resulting in a greater deposition of biogenic carbonate (planktonic foraminifera) (Ellenor, 1976). This caused the siltstones and mudstones to grade into marls. At the basin margins, bioclastic limestones were being deposited during this time interval at Konkon-1 and Durroon-1.

The Miocene sea level peaked within the E foraminiferal zone, from which time the sea level fell until the Pliocene. In this period, bioclastic limestones were deposited. These limestones (cocquinas) contained an abundance of bryozoan fragments. Foraminifera and molluscan shell fragments were also common. Jones and Holgate (1980), using shallow seismic data, interpreted an apron of Miocene outcrop on the sea floor occurring around the western margin of the Bass Basin.

In the Early Pliocene, the sea level rose again in the Bass Basin. On Flinders Island, a low angle unconformity is recognised between the Miocene and Pliocene sediments (Spry and Banks, 1962). Jones and Holgate (1980) also mapped an unconformity between the Miocene and Pliocene sediments along the northeastern portion of the Torquay Sub-basin and up to the King Island-Mornington Peninsula Ridge. Within the central portion of the basin, the Miocene and Pliocene appear conformable. Bioclastic limestones are the prevalent rock type for the Pliocene to Recent section.

The well data along a west-east cross section of the basin suggest that the major depocentres since post Late Eocene time are located in the Poonboon-Nangkero and Aroo areas of the basin. A section in the northwest-southeast direction suggests a dramatic thinning of the Oligocene and Miocene section in the Toolka-Cormorant areas. It means that the areas where early Tertiary subsidence was greatest (Narimba-Pelican and Cormorant areas) did not continue to be the major depocentres in the Late Tertiary. In the Late Tertiary basin, subsidence appears to have been greatest in the Aroo, Poonboon and Nangkero areas. These changes in basin depocentres suggest a change in tectonic regime controlling the fault block movements and / or a change in the geothermal patterns controlling subsidence rates.

Igneous Rocks

There is strong evidence that there has been extensive volcanic activity in the Bass Basin throughout most of its evolutionary history. Brown (1976) has given an account of the type and ages of different igneous rocks encountered in about 8 wells that cover a wide range of composition and a broad geological range. Olivine basalts were encountered in Durroon-1 (Late Cretaceous) and Tarook-1 (Early Eocene or younger). Olivine gabbro was found in Toolka-1 and Cormorant-1 (Early Eocene or younger). Tuffs were found in Bass-1 of both Mid Miocene and Late Oligocene age. Basalts of Palaeocene and older age were encountered in Aroo-1 well and weathered tachyte of Mesozoic (?) age was found in Bass -2. Weathered basalts of Early Cretaceous (?) age were encountered in Konkon-1 well.

From the Late Eocene to Recent, three periods of igneous activity have been recognised from seismic and well data (Chamberlain, 1988) which are: (a) Late Oligocene (H foram zone), (b) Mid Miocene (E-F Foram zone) and (c) Late Miocene (C Foram zone).

Smit (1988) and Chamberlain (1988) have detailed the three main pulses of igneous activity in the thermal sagging phase of the basin. These belong to the Late Oligocene, Mid Miocene and Late Miocene (Figure 2.8). The Late Oligocene volcanics are the most extensive, occurring particularly along the northeastern and southwestern marginal areas and in the Bass-1 area within the basin. These volcanic centres appear to be aligned along northwesterly trends. The Miocene volcanics are less extensive and occur as more isolated features within the basin, but they appear to be aligned along northerly trends. The igneous intrusives occur predominantly along the central northwesterly axis of the basin. The emplacement of the intrusives was probably associated with volcanic activity elsewhere in the basin and does not represent separate igneous events. The intrusives encountered in the Cormorant-1 are dated at 22.5 Ma whereas the sediments hosting them are of *M.diversus* age (50 Ma).

Another feature of the volcanics is that they occur where there is greater structural deformation whereas the intrusives are associated with areas that have subsided more passively.

Reactivation of older faults during active tectonism in the Late Tertiary might have helped both structural growth as well as distribution of volcanics. The apparent lack of magnetic response of the Late Tertiary volcanics is an intriguing fact in the basin (Chamberlain, 1988) but he has proposed that the grid spacing may not have been sufficient to pick up the response from the igneous bodies. The northeastern and the southwestern areas have, however, clear detailing of the igneous bodies even though the shallow basement is there.

The volcanic rocks consist of olivine basalts and lapilli tuffs. On mainland Tasmania, volcanics of Oligocene and Miocene age outcrop statewide, but are concentrated along the northwest coastline region and are observed on seismic data to be further concentrated in the southwest to western marginal areas of the Bass Basin. Another large concentration of similarly aged volcanics is observed on seismic data to occur in the northeastern corner of the basin.

CHAPTER 3 METHODOLOGY

3.1 Rationale of the study

As has been mentioned in section 1.2, the primary objective of this study was to gain a better understanding of the tectonostratigraphic evolution of the basin by looking at the detailed geometric and kinematic evolution of the rift structure and its influence on the stratigraphy of the basin.

In spite of more than three decades of exploration activity, a review of the literature suggests that there still remains a lot of disagreement amongst various workers about the broad tectono-stratigraphic framework of the basin. For example, Robinson (1974) suggested that there are three structural provinces, SE, Central and NW area and the structural growth started in Early Cretaceous in the SE area and progressed to the Central and the NW area in the Early and Late Tertiary respectively. Etheridge *et al.* (1984, 1985) proposed simple NE-SW extension as the basin forming mechanism with shallow to moderately dipping rotational normal faults offset along vertical transfer faults. Smith (1986) defined for the first time a new stratigraphic unit (Durroon Formation) which he suggested was deposited during the Tasman rifting stage. Smit (1988) proposed two phases of extension forming a NW-trending rift with steeply-dipping normal faults and associated transfer / wrench zones in the N-S and E-W directions. Baillie and Pickering (1991) named the SE area as a new basin, the Durroon Basin, on the basis of a different structural history to that of the adjacent Bass Basin. Though they mentioned the structural history of the Bass Basin to be different from that of the Durroon Basin, they failed to provide the details of the mechanism or the specific differences in the geometric and kinematic aspects of their structural history. Their study suffers from a lack of supporting data in the NW Bass Basin to enable them to conclude that the Cretaceous period structural history of the SE part (proposed as Durroon Basin) is different from the NW part (suggested as the Bass Basin proper). Though stated in brief in their paper, the effect of Otway rifting and Tasman rifting is still not clear and this study was aimed to address that problem in more detail applying a seismic sequence stratigraphic approach to basin analysis.

Recently, based on structural interpretation of seismic data and flexural modelling, Setiawan (2000) concluded that the Bass Basin is segmented into three provinces separated by two major transfer zones. He named the three provinces as the Northern Bass segment, the Durroon segment and the Tasmania Island segment. He named the intervening transfer zone between the Northern Bass segment and the Durroon segment as the Pelican-Squid transfer zone and that between the Durroon segment and the Tasmania Island segment as the Northern Coast Tasmania transfer zone.

Clearly, there remains some disagreement about the structural and stratigraphic development of the basin. More specifically, though the basin is known to have been influenced by the two rift-drift histories related to the Otway and Tasman Sea rifting phases, the exact details of how these two main tectonic events have shaped the basin history is still unclear. This study was therefore, aimed to address that problem in more detail by applying a seismic sequence stratigraphic approach to basin analysis.

In order to meet this objective, a large database consisting a large number of regional scale seismic lines were collected over the entire basin along with data from 32 wells (section 3.2). The data density in the Bass area (NW part) and the Durroon area (SE part) is conspicuously different because of perceived hydrocarbon prospectivity of the NW part. Also, in spite of best efforts, seismic data from the SE part could not be obtained in better measure compared to the NW part. It was decided to divide the entire area into two separate enitities for detailed interpretation work. However, sequence stratigraphic subdivision of the sedimentary column was done based on a model proposed by Prosser (1993) by defining the 'broad tectonic systems tracts' (section 1.5) in both the areas by tying with relevant well data. These seismically-defined sequences were then mapped in the respective areas to analyse the basin evolution. The main focus of the study was to understand the evolution of the fault systems both in section and plan view and to relate the results to a broader tectnostratigraphic framework by integrating with the seismic sequence and seismic facies analysis results. Chapter 4 contains all the results from this study

Initial interpretation of 2D seismic data led to an observation about the small-scale layer-bound fault system in the lower part of the Torquay Group which was then decided to be mapped in further details. Though most of the work on layer-bound polygonal fault system has been carried out in the Late Tertiary section of the North Sea Basin (Cartwright, 1994a, 1994b; Cartwright, 1996; Lonergan, Cartwright & Jolly, 1998), a couple of papers have been published on a similar polygonal network of faulting in the Cretaceous siliciclastic sequences of the Eromanga Basin of Australia (Cartwright and Lonergan, 1997; Watterson *et al.*, 2000). There exists only one reference to a possible layer-bound fault system in the Bass Basin (Cartwright and Dewhurst, 1998). Hence, it was decided to make a detailed study on the existence and nature of this layer-bound polygonal fault system by analysing a 3D seismic dataset available in the basin (The Yolla 3D seismic dataset). The result of this analysis is included in Chapter 5.

Flat spot anomalies in the large 2D seismic dataset were investigated in order to assess some of the hydrocarbon potential of the Bass Basin. While discussing about the general hydrocarbon potential of the basin based on previous work, an attempt was made to list the direct hydrocarbon indicators (seismic flat spot anomalies) and the details have been presented in Chapter 6.

3.2 Database

This study utilised data from 29 wells distributed throughout the Bass Basin (Figure 2.1). Biostratigraphic reports contained within the well completion reports are based on spore & pollen assemblages from the Cretaceous section and foraminifera samples collected from sidewall cores and ditch cuttings and are of variable quality. Table 3.1 lists the wells that have been incorporated in this study. Although recently two wells have been drilled, Yolla-2 and White Ibis-1 (Lennon *et al*, 1999), the data were not open file and hence could not be included in the study.

Table- 3.1

List of the wells used in the study

Aroo-1	Bass-1	Bass-2	Bass-3	Chat-1
Cormorant-1	Dondu-1	Durroon-1	Flinders-1	King-1
Konkon-1	Koorkah-1	Nangkero-1	Narimba-1	Pelican-1
Pelican-2	Pelican-3	Pelican-4	Pelican-5	Pipipa-1
Poonboon-1	Seal-1	Squid-1	Tarook-1	Tasmanian Devil-1
Tilana-1	Toolka-1A	Yolla-1	Yorongi-1	

Seismic data were supplied by Origin Energy Limited and Australian Geological Survey Organisation. The long regional seismic lines were from the, AGSO40, AGSO148, B70A, B72A, BBS81, BMR88, TNK4 and TQH5 series and some prospect level campaigns viz. HB75, HB77A, S92A were included. The aim was to distribute a grid of seismic lines at an average line spacing of 5-6 km in the Bass area where access to good data coverage from company data was available during the project and an average line spacing of 20-25 km in the Durroon area where data could not be arranged to obtain better data density. Table 3.2 lists the seismic lines that were used in the main seismic interpretation phase of the study including both the Durroon and Bass area.

Apart from the 2D seismic data, the Yolla 3D seismic dataset was utilised for mapping the polygonal fault system in the Yolla area. This dataset was also supplied by Origin Energy Limited. A composite location map showing the distribution of all the 2D seismic lines used in the entire study along with the Yolla 3D survey area has been prepared (Figure 3.1) although detailed 2D line names can be seen in the location maps presented in the corresponding sub-chapters 4.1 and 4.2.

Table- 3.2

List of the 2D seismic lines used in the study

148-04	148-05	40-11A	40-11B	40-11C	40-19	40-19A
40-20	40-20A	B70A-6A	B70A-6B	B70A-9	B70A-21A	B70A-21B
B70A-24	B72A-89	B72A-106	B72A-114	BBS81-6	BMR88-301	BMR88-302
BMR88-303	BMR88-304	BMR88-305	BMR88-306	BMR88-307	BMR88-308	BMR88-309
BMR88-310	HB75A-200	HB77A-301	HB77A-305	HB77A-309	HB77A-314	S92A-108
S92A-119	S92A-127	S92A-135	S92A-139	TNK4-01	TNK4-03	TNK4-05
TNK4-07	TNK4-09	TNK4-11	TNK4-13	TNK4-16	TNK4-19	TNK4-20
TNK4-22	TNK4-25	TNK4-32	TNK4-34	TNK4-35	TNK4-37A	TNK4-38
TNK4-39	TNK4-40	TNK4-44A	TNK4-49	TNK4-50	TNK4-54	TNK4-55
TNK4-59	TNK4-75	TNK4-79	TNK4-81	TNK4-85	TNK4-89	TNK4-99
TQH5-1	TQH5-10	TQH5-100	TQH5-102	TQH5-103	TQH5-107	TQH5-11
TQH5-111	TQH5-115	TQH5-121	TQH5-125	TQH5-129	TQH5-13	TQH5-135
TQH5-139	TQH5-145	TQH5-149	TQH5-15	TQH5-151	TQH5-157	TQH5-161
TQH5-16A	TQH5-17	TQH5-18	TQH5-19A	TQH5-2	TQH5-21	TQH5-22
TQH5-23	TQH5-25	TQH5-28	TQH5-29A	TQH5-3	TQH5-31	TQH5-33
TQH5-35	TQH5-37	TQH5-40	TQH5-41	TQH5-44	TQH5-48	TQH5-49
TQH5-5	TQH5-54	TQH5-55	TQH5-56	TQH5-58	TQH5-59	TQH5-6
TQH5-60	TQH5-62	TQH5-63	TQH5-64	TQH5-66	TQH5-68	TQH5-7
TQH5-72A	TQH5-73	TQH5-8	TQH5-80	TQH5-82	TQH5-83	TQH5-89
TQH5-9	TQH5-90A	TQH5-94	TQH5-95	TQH5-96	TQH5-99	TQH5153A
TQH554-1	TQH5561A	TQH559-1				

3.3 Biostratigraphic and wireline log interpretation

Well completion reports for 32 wells in the Bass Basin were examined to extract biostratigraphic information available from each well. The biostratigraphic reports for the wells contain palynological data for the Cretaceous section of each well whereas for the Tertiary section, biozonation according to the foraminiferal assemblages is given in the reports. The biostratigraphic reports identify missing biozones and some reports give an interpretation of the depositional environment. In the wells that were drilled in the early period between 1960s and early 1980s, the biostratigraphic subdivision of the section was poor. For example, the initial biostratigraphic report for the Durroon – 1 (Taylor, 1973) recognised no missing section in the Mid Cretaceous whereas the later palynology report (Morgan, 1991) of the same well has identified two major missing biozones, namely the A. *distocarinatus* and C.*pannosus* zones. When the palynology was repeated in the well to closely examine the biostratigraphic data from the core and cutting samples, this discrepancy was noted and it is believed that the older biostratigraphic subdivision that exists today may undergo major revision if the palynology is repeated with more caution and care.

Most of the wells contain a suite of wireline logs consisting of resistivity, density, neutron, gamma and sonic logs. The wireline log character was integrated with the biostratigraphic subdivision from the wells to subdivide the section into different sequences bounded by unconfomities. Biostratigraphic reports were used to determine the timing and extent of missing section across the basin. Many of the biostratigraphic hiatuses were associated with pronounced log breaks which facilitated ease of correlation across the basin.

The density and sonic logs were utilised to create synthetic seismograms for 15 wells to give a good network of control tie points for the seismic interpretation across the basin. The GEOLOG software was utilised to prepare the synthetics. The Durroon area suffers from good well control and the main tie point was at Durroon-1 though the wireline log depth data from Chat-1 was tied with the help of checkshot data alone because there were no satisfactory sonic and density logs for the well. The Bass area has however, been well constrained by the construction of 14 synthetic seismograms (Table 3.3). Figure 3.2 shows a typical synthetic seismogram constructed for the Narimba-1 well.

Table- 3.3

List of the wells for which synthetic seismograms were prepared

Aroo-1	Bass-1	Bass-3	Cormorant-1	Dondu-1
Durroon-1	Koorkah-1	Nangkero-1	Narimba-1	Pelican-4
Pipipa-1	Poonboon-1	Tarook-1	Toolka-1A	Yolla-1

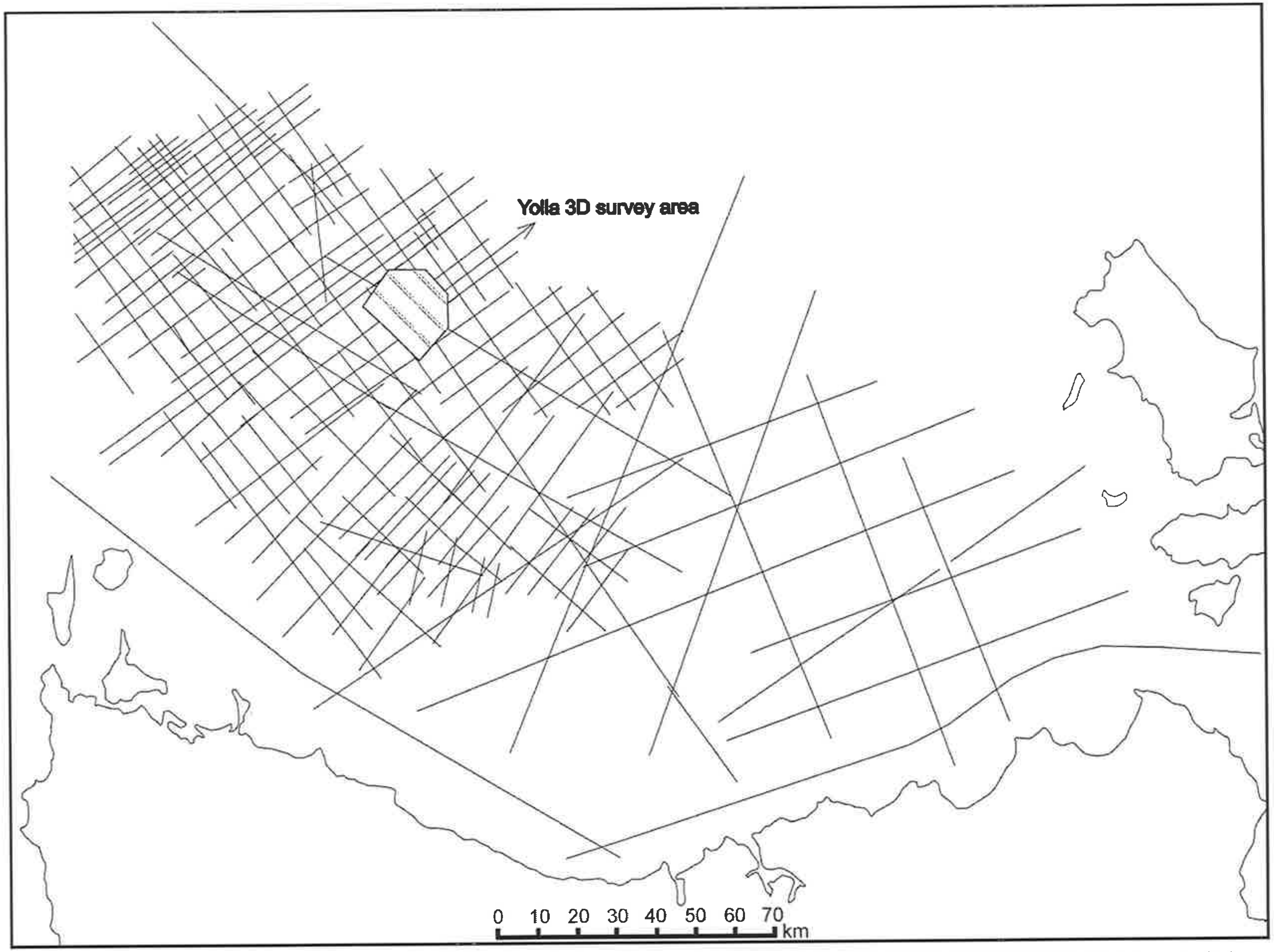


Figure 3.1 Composite location map showing the distribution of all the 2D seismic lines and the Yolla 3D survey area used in the present study. Detail of the Durroon and Bass area with individual line names are shown in Figure 4.1 and 4.16 respectively.

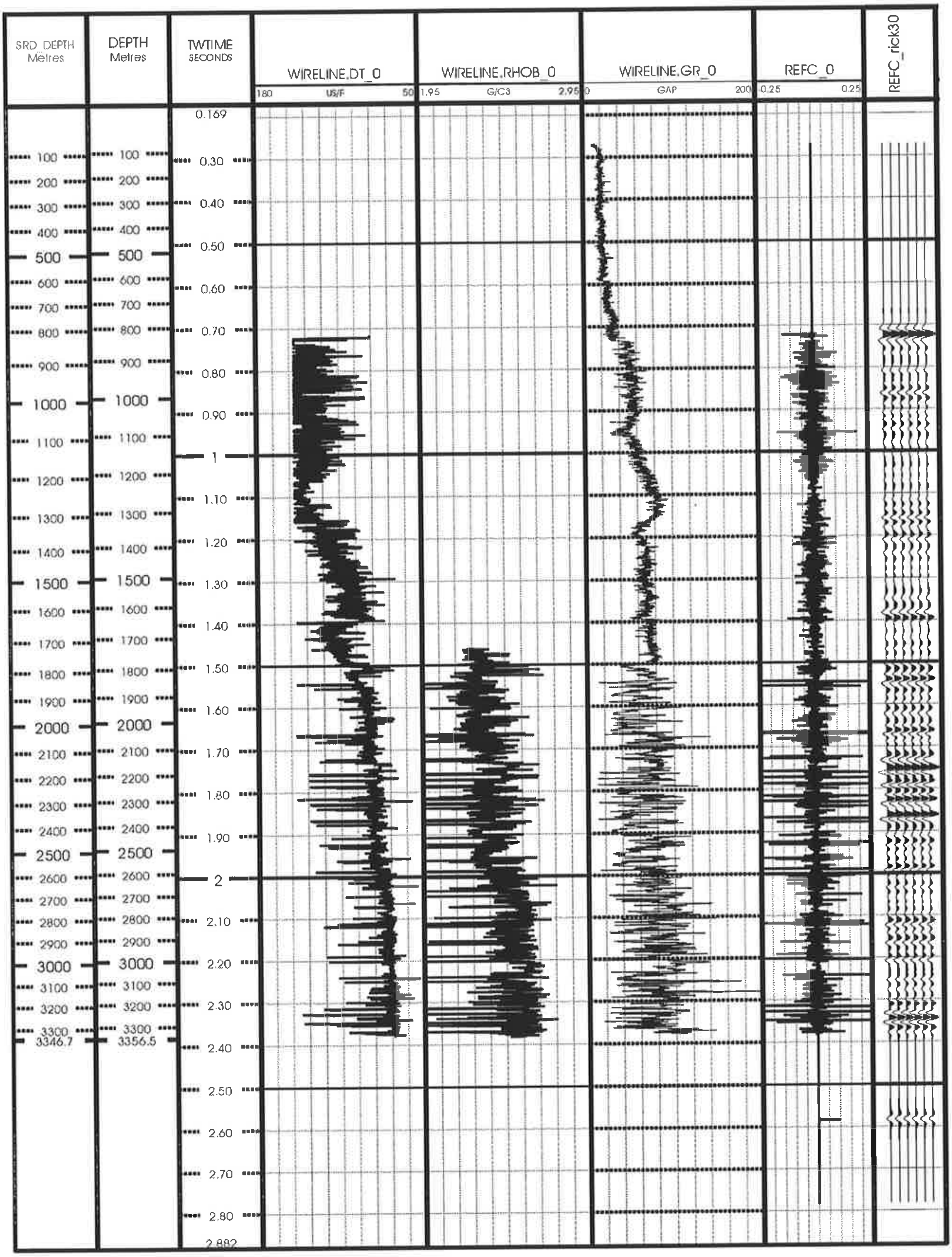


Figure 3.2 Typical synthetic seismogram constructed for the Narimba-1 well.

3.4 Seismic interpretation

A regional grid of seismic data was loaded onto a 2D workstation for interpretation using the seismic interpretation software SEISX (Paradigm Geophysical™, 1997). Wireline log data were converted to two-way time by constructing synthetic seismograms and loaded into the seismic database along with the seismic sequence stratigraphic picks defined during the wireline and biostratigraphic analysis.

Wireline logs and the associated sequence stratigraphic picks were posted on the seismic display window in order to identify the seismic response of events. A seismic pick was made at the nearest trough, peak or crossover on the seismic wavelet. This event was then mapped throughout the immediate area to ensure it corresponded with the other wells in the area. The seismic picks were adjusted and the wireline pick was re-evaluated until a consistent signature for the sequence stratigraphic event could be identified on the seismic. The event was then mapped throughout the extent of the seismic database using the standard seismic interpretation techniques (Mitchum and Vail, 1977) and information from all available wells to constrain the interpretation. Stratal termination patterns were also used to constrain the position of the picked horizon.

A series of two-way time and isochron maps were generated to investigate the nature of sediment deposition and basin development. By gridding the horizons mapped on the 2D seismic lines, it was possible to generate a two-way time map for each horizon to illustrate the present day configuration of the sediments. During the study, 8 events were identified and their spatial distribution mapped throughout the seismic database. These two-way maps were useful in defining the structural trends in the basin.

Since the data quality in the Pelican Trough area within the broader Bass area is of questionable integrity particularly beyond 2.5 seconds (as shown in Chapter-4), emphasis was not given on construction of two-way time maps or isochron maps in the Bass area. This remained a lacunae in the present study which should be improved upon with due care in future studies in the basin.

In the Durroon area, the line spacing was quite broad. This line spacing (up to 25 km) caused problems when gridding the horizons to generate the two-way time and isochron maps. The gridding process in SEISX works by defining a grid cell size and maximum gridding distance. The gridding distance is the number of adjacent cells used to calculate a grid cell value. Thus a combination of the grid cell size and gridding distance must be greater than half the largest line separation in order to interpolate between lines. A smaller grid cell size will give a smoother appearance to the resultant grid. However, if the grid cell size is too small, it will require a large gridding distance to interpolate between adjacent lines and this will tend to smooth out the localised features in the map. To address this problem, the maps were initially generated using a grid cell size of 650 × 650 m and a maximum gridding distance of 30 grid cells giving a range of 19.5 km over which the gridding calculation took place. This gave a coarse resultant grid which was improved by re-gridding using a grid cell size of 250 × 250 m and with a maximum gridding distance equal to the ratio of the original grid size to the new grid cell size, in this case a ratio of 2.6.

It was however, due to some glitch in the SEISX software (which could not be overcome during the course of the study), observed that the two-way time maps were not incorporating the fault picks along with the horizon picks. This could possibly be due to extremely large range of fault throws in the seismic sections that were to be incorporated in the two-way time maps. This forced for an alternative procedure to go for individual fault picking at different horizon level manually and drawing all the fault picks in a Corel Draw package to have an areal mapping.

All the faults were identified in the seismic data from reflection terminations and or fault plane reflections and all the faults were correlated in the plan view. As has been mentioned earlier, the fault pattern maps were prepared separately for the Durroon area and the Bass area. The fault pattern maps were analysed to understand the structural evolution of the basin. The results are discussed in detail in Chapter-4.

For the purpose of mapping the polygonal fault system, seven horizons were mapped within a part of the Yolla 3D area (10km × 8km). Two horizons are within the undeformed interval and 5 are within the deformed interval. The two-way isochron maps were prepared from the 3D data volume by utilising an 'autotrack' correlation technique to analyse the fault system and the results are discussed in detail in Chapter-5.

Of the two seismic flat spot anomalies identified in the basin during the study, one was investigated in detail to look at the prospectivity of the associated structure by preparing a two-way time structure map and a stratigraphic cross section. The results are presented in Chapter-6.

A preliminary attempt to estimate the amount of extension was made by measuring the pre-faulted length of the basement and the post-faulted (or present day) basement length in both the Durroon and Bass areas. Results are discussed in Chapter 4.3 and in the concluding Chapter-7. The estimate of extension derived does not represent the exact amount of extension due to crustal extension (reflected in normal faulting) because the present day situation would also contain a component of loading and/or compaction-related subsidence even along the pre-existing normal faults. However, this estimate has been considered as a guide to explain some of the other seismic interpretation results. Cummings *et al.* (2001, in preparation) have looked at a thorough estimate of extension by taking into the above factors utilising the Midland Valley 2Dmove software. The method they have adopted involves decompacting the sediments at several stages and backstripping each time to arrive at a correct estimate of the amount of extension.

CHAPTER 4 SEISMIC INTERPRETATION RESULTS

4.1 Durroon area (Durroon Sub-basin)

As was noted in section 3.1, variable data density forced the Durroon area and the Bass area to be considered separately for detailed interpretation work.

4.1.1 Database

Figure 4.1 is the location map of the Durroon study area showing the distribution of seismic lines and wells used in the study. In total, 17 seismic lines (a total of approximately 1580 line km of seismic data) have been used in the study of this part of the basin. These data correspond to mainly regional seismic lines in the area shot with an aim to investigate the deeper subsurface including crustal features. The list of the seismic lines is given in Table 1.

Table-4.1

List of the seismic lines used in the Durroon area

BMR88-301	BMR88-302	BMR88-303	BMR88-304	BMR88-305	BMR88-306
BMR88-307	BMR88-308	BMR88-309	BMR88-310	148-04	148-05
40-11B	40-11C	40-19A	40-20	40-20A	

There are 3 wells in this area. They are Durroon-1, Chat-1 and Tasmanian Devil-1. Though there are other seismic campaigns in the area, particularly those shot by oil companies over prospective targets, they are not regional lines. However, in spite of best efforts, these lines could not be accessed for inclusion in the present study.

4.1.2 Data quality

The main seismic campaign in the area was the BMR88 series that has a good coverage of the entire Durroon area at an average line spacing of 20 km. The quality of data is fair to good. The seismic line BMR88-306 is considered as the pivotal line in the series. This line, which runs through the Durroon-1 well, exhibits relatively the best quality data in the entire series of BMR-88. It is probably due to the fact that the average dip of the seismic reflectors and the shooting direction of the line are almost orthogonal to each other, particularly in the two main fault blocks west of Durroon–1 well (Bark and Anderson Sub-basins of Baillie and Pickering, 1991). Hence, it has been considered to be the key line to build the seismic stratigraphic model for the area of study and a full interpretation of this line has been presented below in good details.

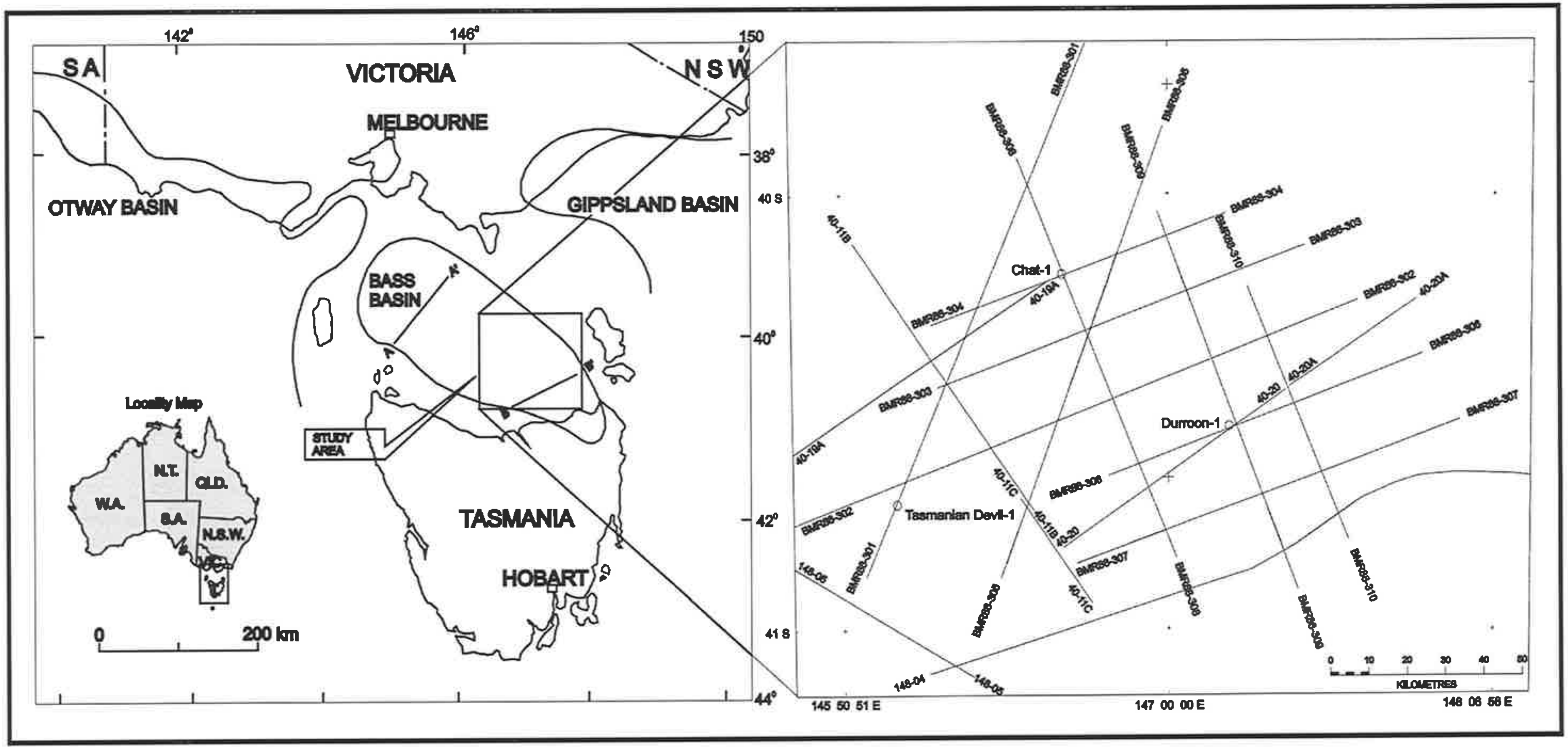


Figure 4.1 Location map of the Durroon area showing the distribution of the seismic lines and wells used in the present study.

Seismic ties at line intersections are within acceptable limits (10-20 milliseconds) except at places where the reflectors are dipping at very large angles which suggest that migration has not been able to perfectly shift the dipping reflectors from their apparent positions to their true depth locations in the subsurface. The artifacts from this problem at few places have been noted qualitatively while carrying out the interpretation work in the area. Quantitative elimination of the misties was considered to be unwise since static shifts, if applied to horizons based on mistie values at line intersections, would jeopardise the reflector positions where the dip is almost flat to low angle dips along strike lines.

The overall quality of the line BMR88-306 is moderate to good. There are few interference effects due to overmigration producing migration smiles, diffractions from the fault planes, out of plane reflections (sideswipes) and multiples both from sea floor and some horizons within the sedimentary section resulting in cross-cutting of reflectors at few places along the line. The problems from the events above sometimes combine to make interpretation more subjective but care has been taken by maintaining the consistency in approach while doing both structural and stratigraphic interpretation of the seismic data. Poor velocity analysis and hence poor stacking could be the reason for loss of reflector continuity at a few other places. The data from the depth (beyond 2.5 seconds) is generally poor to bad.

4.1.3 Methodology

The seismic reflectors have been tied through a synthetic seismogram using the sonic and density wireline logs from the Durroon-1 well and then age-dated by using the available biostratigraphic information from the well (Figure 4.2). The major sequence boundaries have been picked by identifying the reflection terminations following mainly the onlap and truncation patterns. Based on the model proposed by Prosser (1993), the sedimentary section at Durroon-1 well has been divided into its rift-related depositional sequences. The seismic sequence boundaries are dated using the biostratigraphy from the well and a synthetic seismogram constructed from the sonic and density data. The major angular unconformity showing truncation of reflectors below and onlap above has been identified as the rift-onset unconformity (Top Otway horizon) and is dated about 96 Ma. The Early Cretaceous A. *distocarinatus* and C.*pannosus* zones are missing in this well indicating flexural / thermal uplift of footwall block and substantial erosion at the beginning of the Tasman Sea rifting phase.

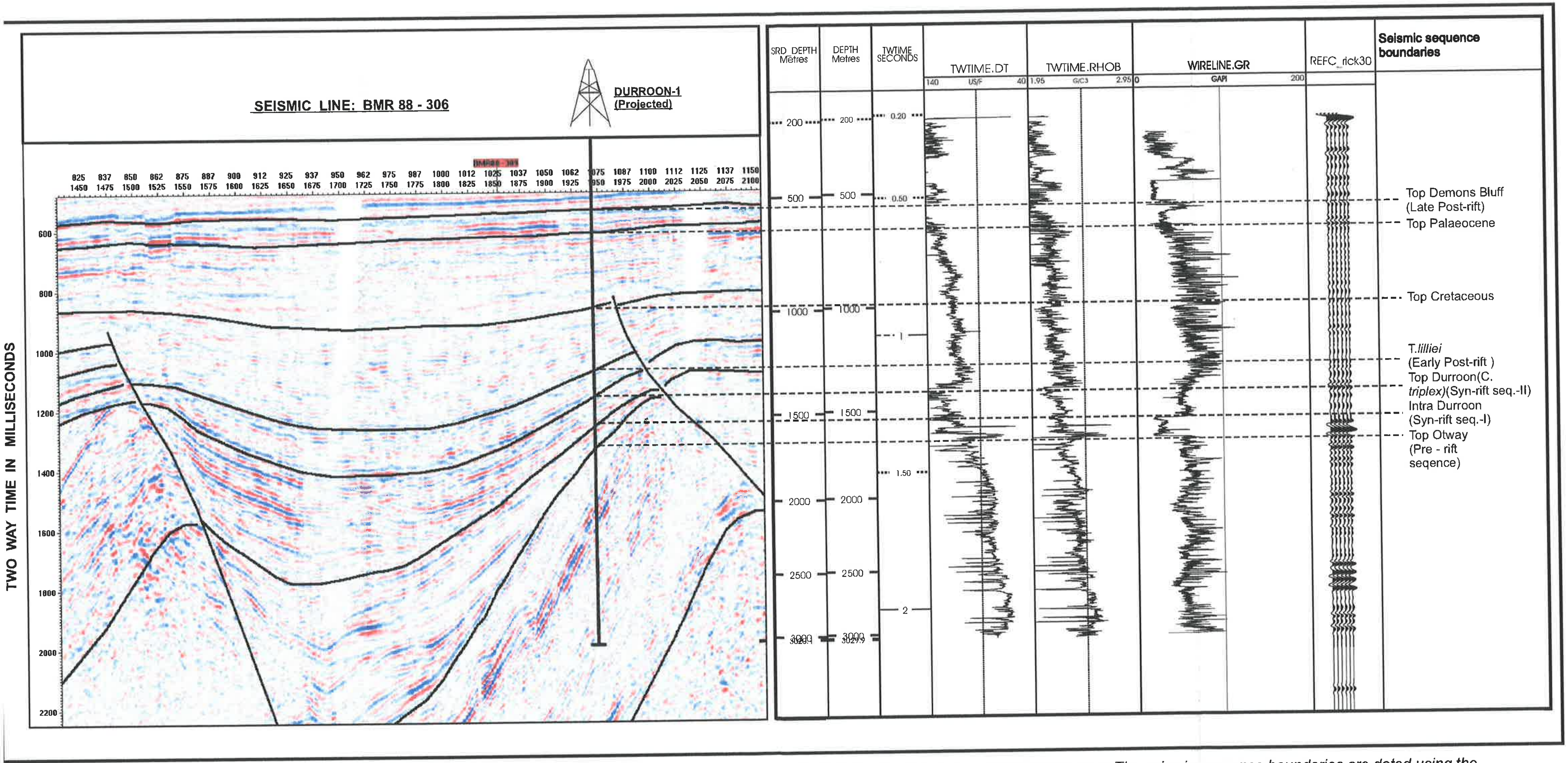


Figure 4.2. Based on the model proposed by Prosser (1993), the sedimentary section at Durroon-1 well has been divided into its rift-related depositional sequences. The seismic sequence boundaries are dated using the biostratigraphy from the well and a synthetic seismogram constructed from the sonic and density data.

The other sequence boundaries have been identified on the basis of onlap pattern of the reflectors combined with the wireline log character at Durroon-1 (Figure 4.2). In all, eight sequence boundaries have been considered for mapping. They are, from oldest to youngest, Basement, Top Otway, Syn-rift sequence I (Intra Durroon), Syn-rift sequence II (Top Durroon, C *triplex* zone), T. *lilliei* (Early post-rift), Top Cretaceous, Top Paleocene, Top DemonsBluff Formations.

The overall wedge shape of the two syn-rift sequences towards the fault surface is clearly seen. The early post-rift sequence shows gross parallel reflectors covering the topography with a constant thickness of sediments and drowning the crestal portions of both the footwall and hanging wall blocks but confined to the individual tilted basement blocks. There is however, some displacement observed due to faulting from sediment loading or compaction. The late stage post-rift sequence marks the complete quiescence of the basin from any tectonic movement and the broad parallel reflection configurations of a regionally continuous nature suggests thermal cooling and passive subsidence.

The prominent reflectors near the biostratigraphic boundaries as correlated at the Durroon-1 well have been picked up for seismic correlation throughout the study area. A tie at the Chat-1 well has improved the confidence of the correlation of the seismic horizons. The shallow Tasmanian Devil-1 well has been helpful for tying the Upper Tertiary sequences only. The information from the checkshot data in Chat-1 and Tasmanian Devil-1 has been used for depth conversion of seismic data because synthetics could not be made from the poor quality sonic and density curves from these two wells. However, the synthetic and checkshot data at the Durroon-1 well give palatable depth matching within about 10 milliseconds and it has been considered an acceptable limit for the overall interpretation confidence in the entire area by using time to depth conversion from checkshot data in the Chat-1 and Tasmanian Devil-1 wells.

4.1.4 Seismic sequence stratigraphic interpretation

As discussed in section 1.4, the recent advancement in seismic sequence stratigraphy as a fundamental tool in regional basin analysis has helped geologists link the depositional history of a sedimentary basin in a chronostratigraphic framework. Although, the earlier sequence stratigraphic concepts developed by Vail *et al.* (1977) were suitable for passive margin settings, the significant influence of the tectonic control on the sequence development in a rift basin setting was established by Prosser (1993) as discussed in section 1.5.

The Bass Basin has been affected by both Otway and Tasman Sea rifting without eventually developing into breakup stage and formation of a passive margin. All along its evolutionary history, it has been a barred basin though intermittent marine influence has been observed in the stratigraphic section, ranging from marginal marine in the later stages of rifting to nearly open marine shelfal conditions in the post-rift thermal subsidence phase of its sedimentation history. The entire sedimentary section in the Durroon area has been divided into several major sequences (8 mega sequences) following the methodology outlined by Prosser (1993).

The truncation of reflectors below a surface seen near the crestal portion of the hangingwall of the Anderson block (Footwall of the Boobyalla block) and mapped as Otway Top horizon show considerable erosion of the strata below (Figure 4.3). This horizon represents the top of the Otway Group and biostratigraphically there is a missing section in the A. *distocarinatus* and C.*pannosus* zones immediately above this surface (Morgan, 1991). Dated as Early Cenomanian (about 96 Ma), this unconformity surface can be correlated with the onset of the Tasman Sea rift.

The onset of Tasman rifting is evidenced by the fact that there is a sudden and clear change in the dip of the strata across this surface apart from the overall reflection configuration both below and above this surface of unconformity. The reflection configuration below this horizon is parallel to sub-parallel, moderate to high amplitude and medium to high frequency content whereas the reflector configuration above this surface is high amplitude, parallel to sub-parallel and medium frequency content having a progressive onlapping pattern (Figure 4.3). The Otway Group of rocks is the pre-rift sequence for the Tasman rifting phase as per Prosser's classification and consists of lithic sandstones with subordinate finer-grained sediments including coal of Albian-Aptian age (Baillie and Pickering, 1991).

Although this prominent unconformity surface has been considered as the Top Otway unconformity by all the previous workers in the area, the typical reflection configurations of this package allow it to be correlated as the pre-rift sequence to the beginning of Tasman Sea rifting. This is one of the major findings of the present study since the biostratigraphic information, the seismic data, wireline log character all fit into this Tasman Sea rift model quite well.

BMR88-306 (Anderson Block)

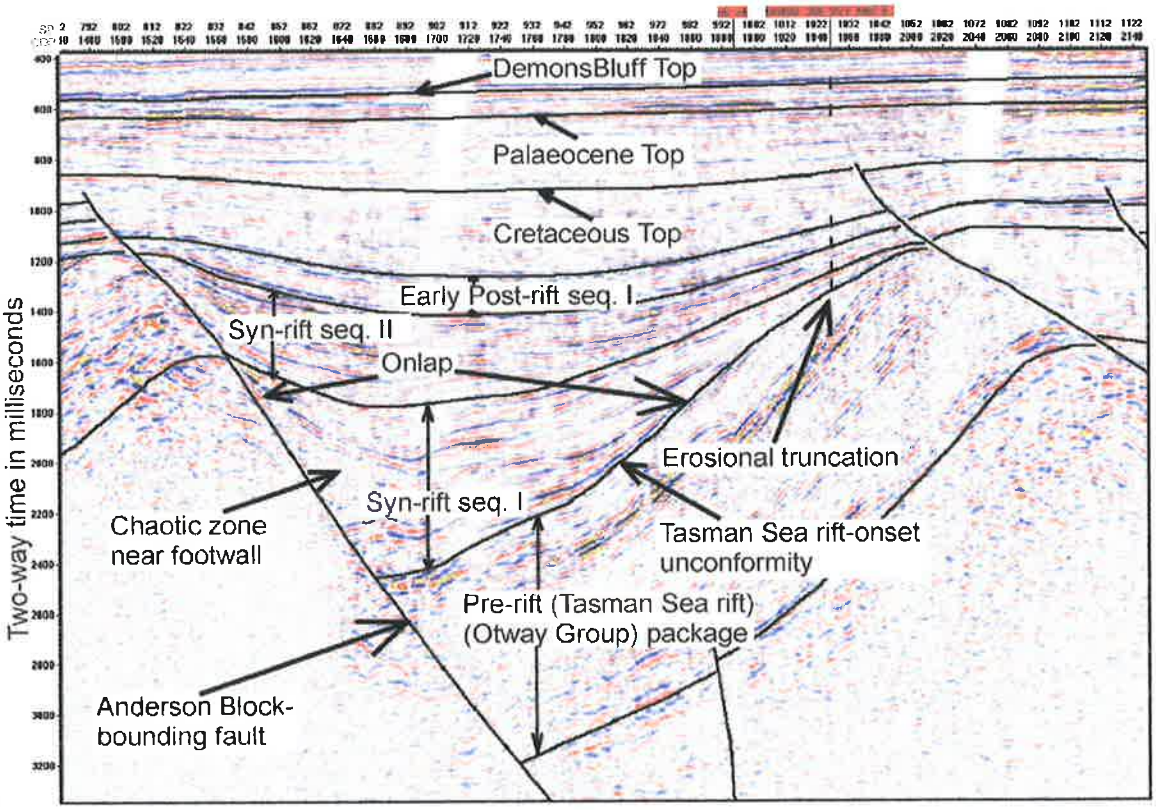


Figure 4.3 Reflection termination patterns and the major rift-related sequences. Note the wedging of the two syn-rift sequences towards the block-bounding fault and the early post–rift sequence is marked by reflectors draping the footwall and hanging wall crests.

Baillie and Pickering (1991) proposed the name Durroon Basin for the southeastern part of the previously known Bass Basin (according to earlier authors viz. Weeks and Hopkins, 1967; Williamson *et al.*, 1987) and sub-divided the basin into three sub-basins, Bark, Anderson and Boobyalla. However, it is considered here more appropriate to use the term 'fault blocks' in place of sub-basins and also the term Durroon Sub-basin for the Durroon area in place of the Durroon Basin. The reasons are obvious. The fundamental unit of any rift is a half-graben consisting of one or more fault blocks (Rosendahl, 1987) and is generally characterised by the rift border fault at one side and the flexural margin on the other side (Figure 1.3a) The interaction between these fundamental units (half-grabens) marks the geometry of any rift system (section 1.3.2). Following the terminology of Rosendahl (1987) the Durroon area is defined as the Durroon Sub-basin consisting of the Bark, Anderson and Boobyalla fault blocks (Figure 4.4).

The early rifting of any sedimentary basin is usually accompanied by normal faulting with flexural/isostatic/thermal uplift of the footwall blocks of the major tilted domino-style basement fault blocks. There is usually erosion at the crestal portions of these footwall

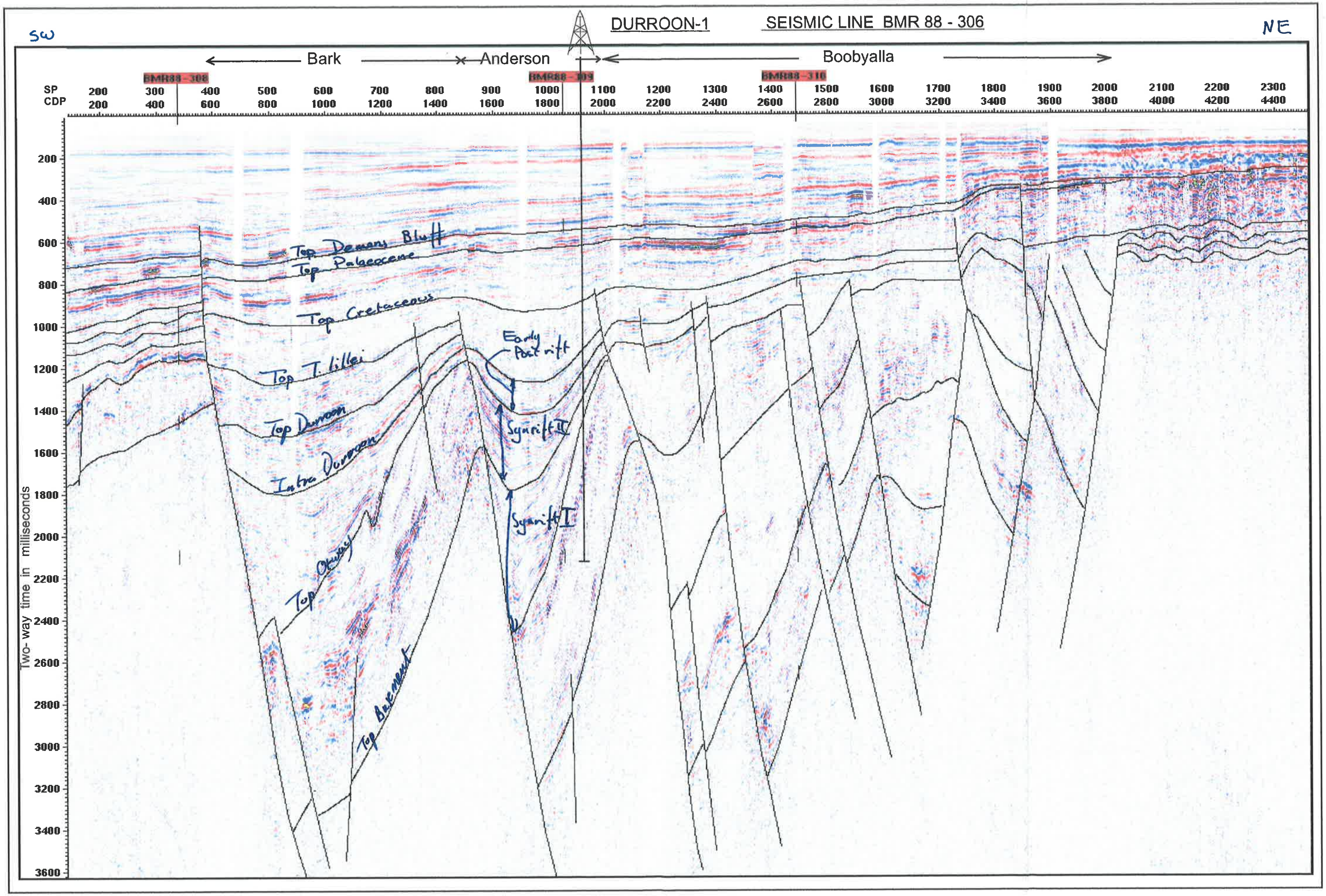


Figure 4.4 Interpreted seismic section showing the structural style in the Durroon area which is characterised by large tilted basement fault blocks with throws up to 4-5 km. The outline of the Bark, Anderson and Boobyalla fault blocks are marked at top. Line length is 89.5 km.

blocks as they become emergent in response to active extensional normal faulting (Barr, 1987; 1991; Cartwright, 1991; Roberts and Yielding, 1991;Yielding and Roberts, 1992; Roberts *et al.*, 1993). This is very clearly seen at the Durroon-1 well (footwall block of the Boobyalla fault block) as well as in the footwall block of the Anderson fault block (Figure 4.15a).

The Boobyalla block as a whole (not excluding the many intervening synthetic and antithetic faults) has some interesting features. As seen in the Figure 4.5, which is part of the seismic line BMR88-307, the pre-rift Otway Group thickness decreases rapidly towards the northeast (right of the line). The reason perhaps is due to the very pronounced emergence of this block at the beginning of Tasman Sea rifting (Roberts and Yielding, 1991; Barr, 1987) and consequent erosion of the Otway strata. Barr (1987) has observed from forward modelling half-graben sedimentary infilling that the larger the fault block, the higher is the emergence. Also, clearly, there is a dramatic difference in the reflection characteristics (seismic amplitude, frequency and continuity) between the Otway Group and the syn-rift sequence I. The quality of the seismic data within the Boobyalla fault block is relatively poor compared to the Bark and the Anderson fault blocks. Moreover, there is also a lack of well constraint in the Boobyalla area.

The reflectors below this unconformity surface in the Otway Group are parallel to sub-parallel with moderate amplitude and fair reflection continuity, clear from both the Bark and the Anderson fault blocks (Figure 4.3 and 4.4). Though there are many onlap and downlap surfaces and some small clinoforms within the Otway sequence, as seen from the overall reflection configurations, the Otway sequence is a totally different mega-sequence as opposed to the sequence above it. The second evidence suggesting the Otway sequence to be different from the syn-rift sequence is that the faulting within the hanging walls of these two blocks is almost non-existent. There are only a couple of faults affecting the lower part of the Otway sequence apart from the major basin bounding / tilted basement block bounding faults which, though they existed during Otway rifting phase, were reactivated during the Tasman rifting phase. These large basement-related faults show displacement of 4-5 km (calculated for both the Basement and the Otway Top horizons by taking average velocity and time difference for observed in the footwall and hangingwall blocks). The reflectors in the hangingwall block of both the Bark and Anderson fault blocks clearly onlap this unconformity surface whereas deeper in the section, the reflectors are more or less conformable with this rift-onset unconformity (Figure 4.3).

The syn-rift sequences I and II show wedge-shaped geometry with increase in thickness towards the fault plane though within the sequences themselves, there are clear

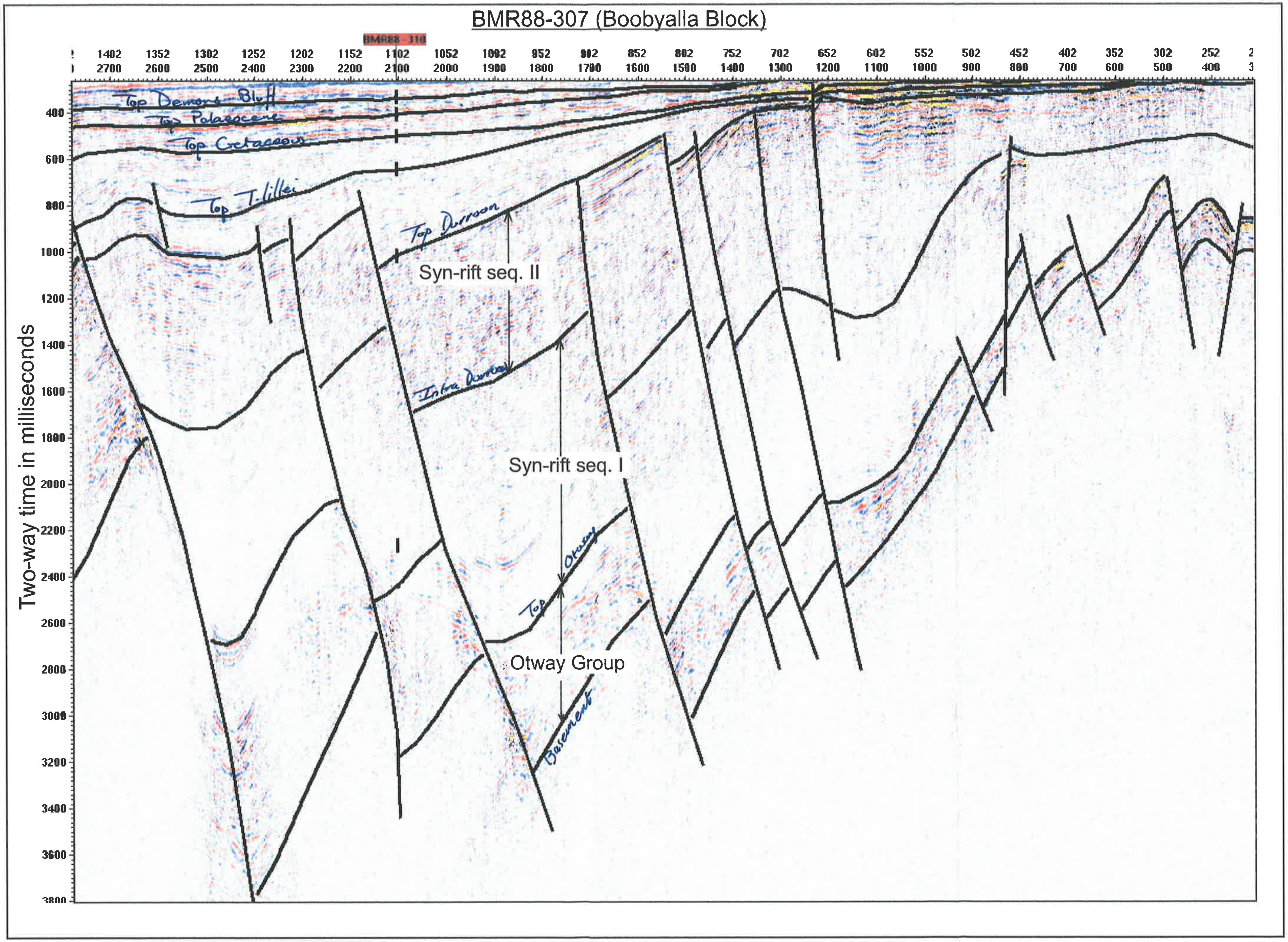


Figure 4.5 Part of the interpreted seismic section BMR88-307 corresponding to the Boobyalla Block. Note the thinning of the Otway Group to the east and possible uplift and erosion at the beginning of Tasman Sea rifting. Line length is 47 km.

indications of onlapping geometry of reflections onto the hangingwall slope and footwall slope (Figure 4.3 and 4.4). The Boobyalla main fault block area seems to have been affected by more faulting activity and is areally larger, dissected by number of faults with relatively lesser throws (Figure 4.4 and 4.5). There is a reversal of the dip direction of strata from west dipping to east dipping across a small graben-like zone within this sub-basin. The reflector continuity and amplitude are poor compared to the other two main fault blocks, the Bark and Anderson blocks, in their hangingwall blocks, the reason is not very clear. There are indications of reflectors onlapping onto the hangingwall blocks above the top of the Otway Top unconformity (Figure 4.5) within the Boobyalla block. A mainly divergent reflection configuration is predominant within this fault block with an increase in thickness in the syn-rift sequences towards the fault planes as opposed to the more sub-parallel nature of the onlapping reflectors in the other two main fault blocks. Also, there is a dramatic increase in the thickness of the syn-rift sequence I towards the major block- bounding fault in the Boobyalla block suggesting that more displacement has been taken up during this time compared to the other two main Bark and Anderson fault block-bounding faults.

The progressive onlapping pattern of the reflection configuration indicates lake highstand whereas there is indication of small downlap surfaces within the syn-rift sequences as well (Figure 4.3). Though accurate stratigraphy and age dating is difficult to determine without any well control, a qualitative interpretation of the facies is feasible.

The eroded footwall scarp material, consisting probably of coarse clastics, shows an aggradational pattern near to the fault plane, particularly in the Bark and Anderson fault blocks. The patchy reflectors with moderate amplitude and low reflector continuity have an overall mounded reflection configuration, indicating a high-energy environment of deposition close to the fault planes. These reflectors appear to be onlapping on successive horizons in the hangingwall blocks. The good reflector continuity of the successively onlapping reflectors in the Bark and Anderson blocks away from the fault planes indicates quiet low energy conditions in the depositional environment. It suggests that very rapid faulting activity created fault scarp relief / topography and conditions for deep water. The deep water developed in response to large fault displacement in the two fault blocks seems to have been passively infilled by eroded material from the hangingwall and footwall blocks (Leeder and Gawthorpe, 1987; Barr, 1987). Also, sediments brought in by antecedent drainage patterns in the evolving rift (Lambiase, 1990; Frostick and Reid, 1989) would tend to deposit deep-water lacustrine facies in the more basinal depocentres with coarser clastics in the proximal areas, particularly in the high areas. The evidence for this is seen in the wireline log character in the Durroon-1 well which is an arenaceous (more sandy facies) immediately above the Otway Top horizon (Figure 4 2). The shaly section succeeding this sandy facies (Durroon Formation of Smith, 1986) in the Durroon-1 well is deposited in a lake highstand which is palynologically of the C.*triplex zone.* The Durroon

Formation has marine dinoflagellates in two horizons within this formation (Morgan, 1991) which suggest that at least marginal marine conditions invaded the area during this period.

Active rifting was probably very rapid and supposedly in two or possibly three tectonic pulses of extension. This has been interpreted from the overall fault trend in the area. As discussed in the structural evolution of the area, a three stage extensional model is proposed. The Otway rifting was already active during the Late Jurassic-Early Cretaceous (evidenced by Jurassic dolerites in the Tasmanian onland margin areas and also in the Victorian coast (Baillie, 1989), the accommodation space being created by extensional normal faulting. It was during the transitional stage from rifting to drifting of the Otway rift, perhaps that the Tasman Sea rifting became active. The second stage of movement, the Tasman Sea rifting eventually created oceanic crust in the Tasman Sea between 82-60 Ma (Magnetic Anomaly 33 and 24) (Weissel and Hayes, 1977). Some of the existing faults of the previous Otway rifting phase in the area were reactivated with large displacement.

The reason why the second stage of rifting commenced immediately after the rifting history of Otway stage, is not clear. This area therefore, would make a good case for research into the very causal mechanism of any rifting event at any place, particularly the conditions that make rifting to begin. There is no study in the literature on whether the rift style has been active or passive in either the Otway or Tasman Sea rift events (Landon, 1994).

There is no well information available in the center of each basin where full stratigraphy is expected to be preserved to link the Otway rifting phase to the Tasman rifting phase. However, a model has been proposed to show what facies could be expected both laterally and vertically in such lacustrine-dominated sub-basins in the area. Lacustrine shales and silts have excellent scope for very good hydrocarbon source potential (Morgan, 1991) exists in the syn-rift section of the Tasman rift phase (Durroon Formation of Smith, 1986; Morgan, 1991). Closer to the footwalls and the near source areas of the hangingwall tilted fault blocks should provide good coarser clastics in fluvio-lacustrine and / or deltaic environments, suitable for petroleum reservoir rocks.

4.1.5 Structural interpretation of the seismic data

In total, 17 seismic lines (a total of approximately 1580 line km of seismic data) have been used to interpret the structural style of the area. The major faults have been interpreted in the seismic sections both from reflection terminations and fault plane reflections and then mapped in plan view. The two-way time structure contour maps at the five levels (Basement, Otway Top, Syn-rift sequence-I and Syn-rift sequence-II and Post rift sequence) have been prepared to get an understanding of the evolution of the structural trend in the area at these stages.

Major basin-bounding fault that developed early on as seen in line BMR88-306 (fault fa – Figure 4.4) did not propagate right across the basin during the first stage of movement. The fault segments originally active during the Otway phase of rifting were reactivated. Analysing the fault trend in the area in conjunction with the seismic stratigraphic analysis, a three-stage extensional history for the area is envisaged. A brief description of the structural history is presented below.

4.1.5.1 Stage-1 (Late Jurassic-Early Cretaceous)

The SE Bass Basin (Durroon area) was part of the Otway rift (continental extension in the eastern Gondwana and as a precursor to the separation of the Australian and Antarctica plates) affected a very large area in the SE Australia region (Griffiths, 1971; Brown, 1974). In this first phase, extensional stresses in a NE-SW direction (Etheridge *et al.*, 1984, 1985), aligned the normal faults in a NW-SE direction with subsequent infilling of the developing faulted depressions (Figure 4.6). The initial response to the extensional stresses was development of large number of faults with small strike extent all across the basin (Figure 4.6). There are clear indications of a high density of faulting during the Otway phase of rifting. The faults have not affected the complete Otway section in most of the cases being restricted to probably early stages of rifting. This is seen from the fact that all the strike and dip lines show high density faulting in the lowermost part of the Otway section (Figure 4.4). It indicates that after the initial rifting and associated faulting, the area was into regional subsidence phase where fluvial-dominated sedimentation took place. It is interpreted here that, being the end member half-graben unit (Rosendahl *et al.*, 1986; Ebinger; 1989; Lambiase, 1990) of the Otway rift, the Durroon area was experiencing the typical third stage / cycle of sedimentation dominated by a fluvio-lacustrine depositional environment. Sediments were probably sourced from the Tasmanian and Australian mainland side and also from the Bassian rise side of the basin. A succession of lithic sandstones, shale and coal, as interpreted from wireline log character and as reported by Baillie and Pickering (1991), constitute the Otway Group. The seismic character within this succession, as mentioned earlier, is moderate to high amplitude, parallel to sub-parallel reflectors with some deltaic progradational and onlap geometry seen. The high amplitude reflections may be from coal seams deposited in flood plain / swampy conditions during the waning phase of Otway rifting. There is a typical increase in thickness of the overall Otway package towards the major bounding faults of the tilted basement fault blocks. The sediment transport direction was most likely parallel to the faults although the sediment is likely to have been derived from the Bassian Rise to the east. Extension taken up by the faults during the Otway rifting stage is reflected in the thickening towards the faults and this provides a basis for determination of the activity of each fault through time. Time-thickness maps were produced where possible to support the identification of fault activity but computer problems prevented map production for some horizons. Identification of fault activity was largely restricted to measurements from seismic sections.

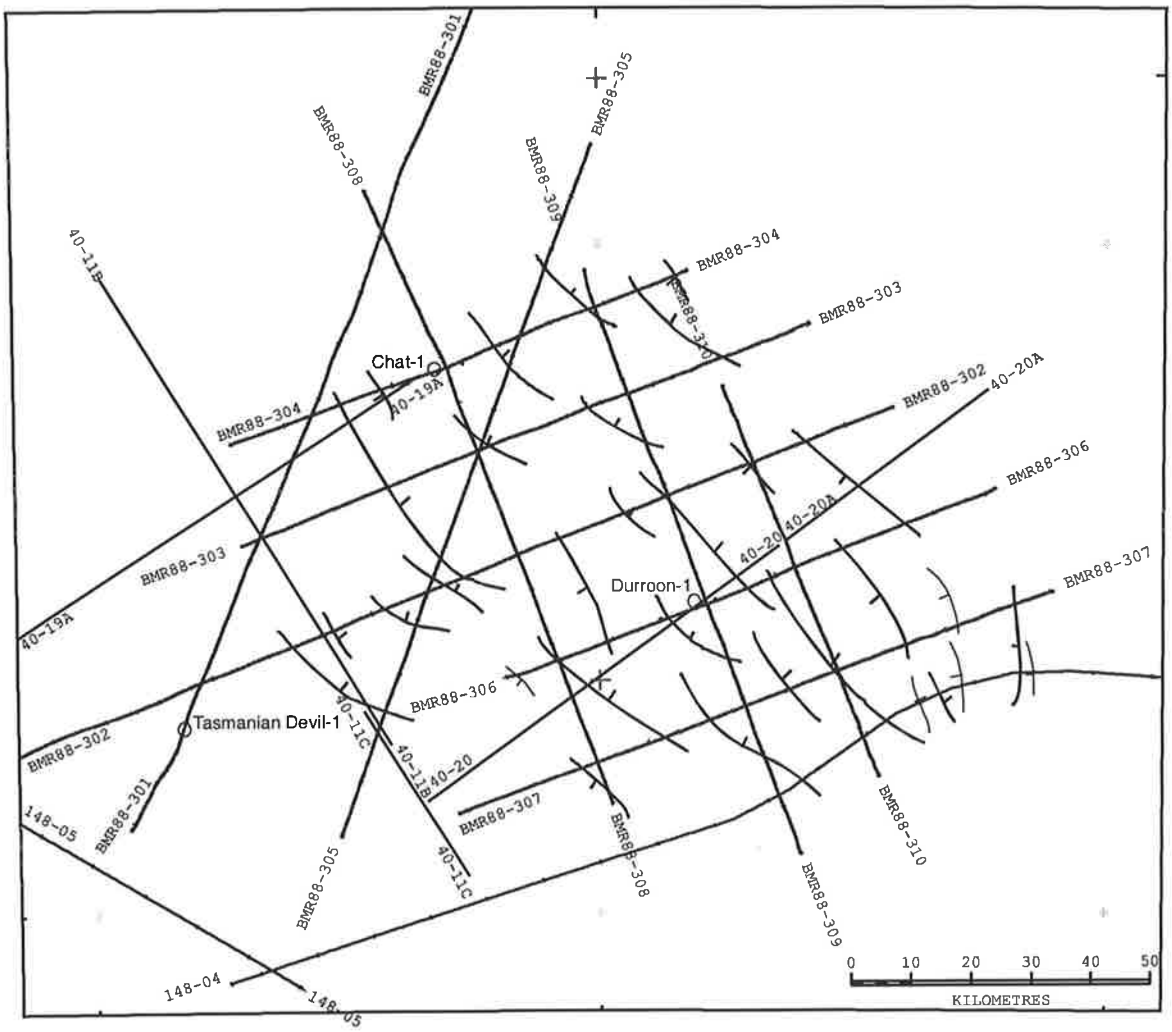


Figure 4.6 Fault pattern in stage 1rifting (Late Jurassic – Early Cretaceous) characterised by a large number of faults with small strike extent all across the basin in the NW-SE direction.

4.1.5.2 Stage-2 (Early Cenomanian)

The Tasman Sea rifting, initiated during the Early Cretaceous (Early to Mid-Cenomanian), accentuated some of the faults of the earlier system and reactivation with huge throws (up to 4-5 kms) shaped the structural history of the area to its present day pattern. In this second phase of extension, faulting was very rapid with almost all the faults of the first stage being reactivated to varying degrees with propagation along their tips in both directions. But notably here, the predominant trend of the faults was N-S. This was due to the E-W orientation of stresses during this phase of extension (Figure 4.7). The change in the overall stress direction from NE-SW direction in the Otway rifting stage to E-W in this

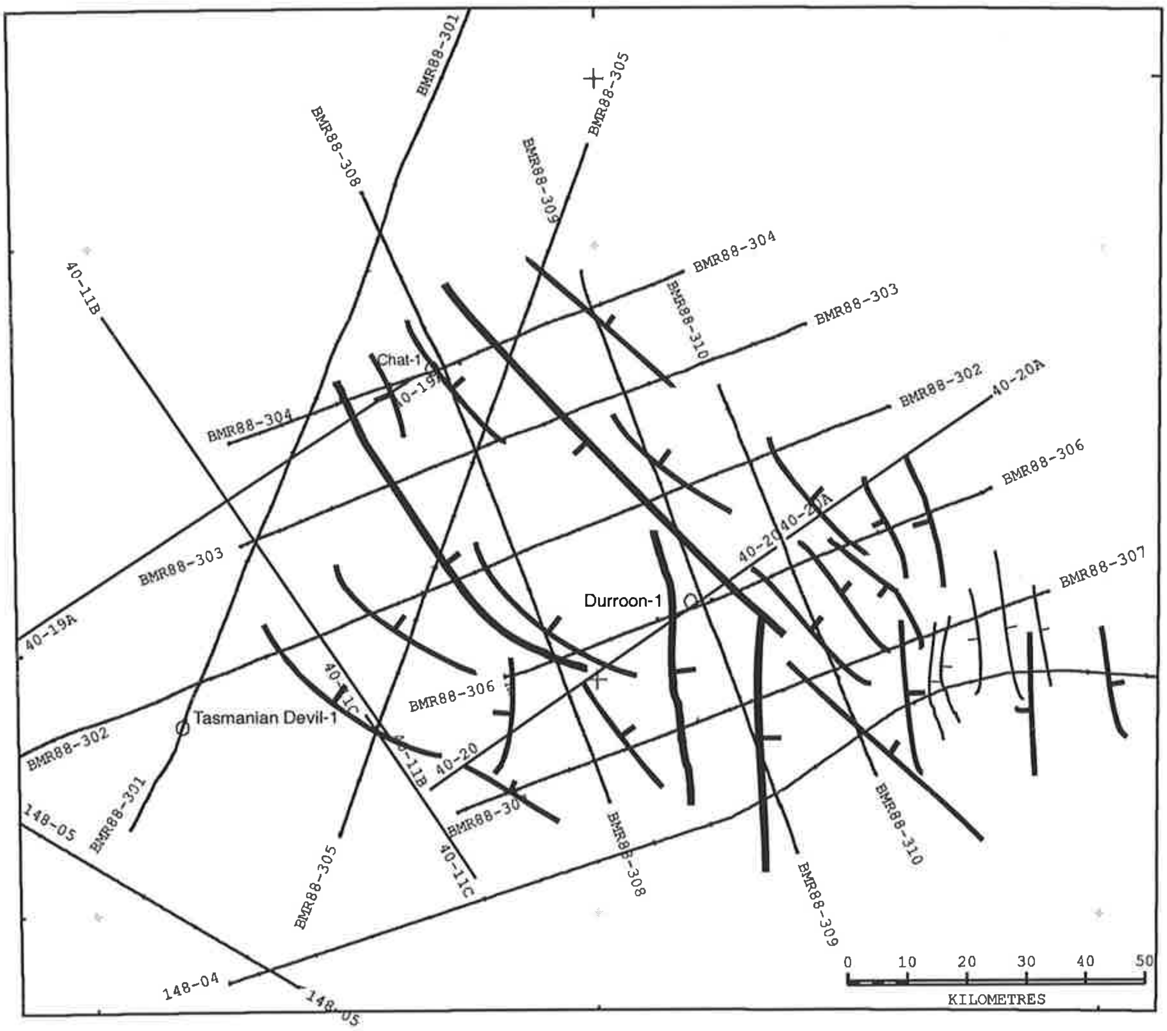


Figure 4.7 Fault pattern in the stage 2 rifting (Early Cenomanian). Mainly N-S fault trend due to E-W orientation of tensional stresses during Tasman rifting phase.

Tasman rifting phase had a strong influence on the alignment of the fault trends. The strongest evidence of this comes from the strike direction of two major faults; one major fault bounding the Anderson block to its west and the other one is the southern extent of the major bounding fault of the Boobyalla block. Both these major faults trend N-S even at their present day position.

The fault alignment suggests that the faulting activity has been very intense mostly in the Boobyalla block i.e., towards the south and the southeast. The very high fault intensity in the extreme SE part of the Durroon area suggests that it is perhaps indicative of the proximity to the future spreading axis of the Tasman Sea rift. The initiation of Tasman Sea rifting was probably synchronous with the initiation of Southern Ocean spreading (dated at 96 Ma approx., Powell *et al.*, 1988). During this phase of major extension, the three fault blocks (Bark, Anderson and Boobyalla) became prominent due to massive fault throw (4-5 kms) along the bounding faults of these blocks.

There are some indications of erosion of the Otway (Top) horizon in the Boobyalla block. However, the overall conformable nature of the strata both above and below the Otway top horizon suggest that the uplift of this end of the Durroon Sub-basin (flexural margin) of the Tasman rift as opposed to the border fault margin has been less prominent. It can therefore be argued perhaps the Otway rifting and Tasman rifting, being contiguous and the area being close to the future Tasman Sea opening margin, the area became more affected by faulting. The reflection character is very patchy with poor continuity, low to medium amplitude in the lower part of the syn-rift section as well as the Otway sequence.

Another interesting fact can be noted that the Otway Group thickness increases towards southwest along the line BMR88-306 suggesting massive uplift and erosion of the area encompassing the Boobyalla block before Tasman rifting began. Alternatively, the depocentre, during Otway Group deposition, was towards the Durroon-1 well location and to the southwest.

4.1.5.3 Stage-3 (Early Campanian)

The beginning of oceanic spreading in the Tasman Sea (82 Ma) was another major tectonic event accompanied by extensional forces. The faults already existing from the two earlier major tectonic phases of extension were reactivated further and the major tilted basement blocks created during stage-2 were subjected to rotational motion along the faults, thereby widening the basin extent considerably (Figure 4.8). The oblique spreading in the Tasman Sea influenced profoundly the present day NW-SE trend of the major faults. The fault strands propagated along strike further and coalesced rapidly as the spreading in the Tasman Sea continued. It is however, observed that during the later part of Tasman rifting and or during the beginning of spreading in the Tasman Sea, large extensional stresses were passively generated and taken up by the throughgoing fault bounding the Boobyalla block. This is interpreted from the dramatic increase in thickness of the late syn-rift sedimentary packages in the Boobyalla block towards the fault 'fg' (Fig. 4.5) seen from the wedge shaped reflection configuration of the sequences. It suggests rapid tectonic subsidence occurred along the large number of northeast dipping rotational normal faults and sedimentation kept pace and even surpassed the rate of accommodation generation.

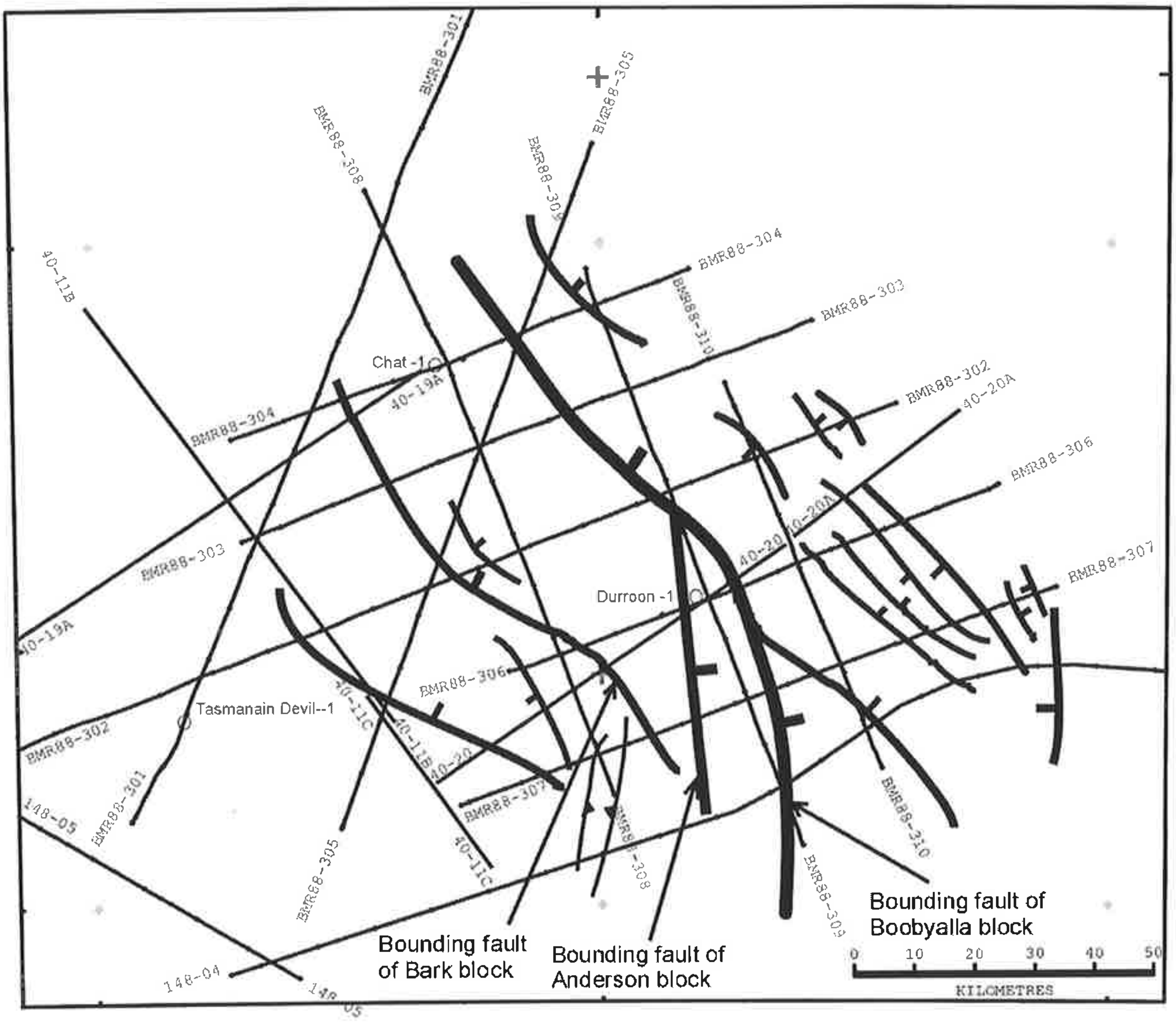


Figure 4.8 Fault pattern in the stage 3 extension (Early Campanian). Fault tip propagation and coalescing of faults during the beginning of oceanic spreading in the Tasman Sea brought out the present day trend in the NW-SE direction.

Figure 4.9 shows analogue modelling results of four stages of extension in case of an oblique rift. Here, the rift zone is at an angle of 60^0 to the extension direction and the figure shows the successive stages of development of the fault system (McClay and White, 1995). The rift border fault and the other interior faults developed at the first stage of extension. The faults developed with small strike extent and the displacement along the faults was small. The fault system was marked by intervening transfer zones or accommodation zones. At the successive stages of extension, the fault system in the rift became quite prominent with large displacements and the small faults of the earlier stage coalesced to form larger or longer faults.

The Otway rifting event had created the NW-SE-oriented rift zone but the direction of extension during the Tasman rift event was E-W. As the opening of the Tasman Sea was oblique to the faults already created in the sub-basin, it is expected that the fault pattern might follow that of the analogue model (Fig. 4.9). Unfortunately, not enough faults could be mapped to prove this point conclusively but some comparison can be made with Figures 4.7 and 4.8.

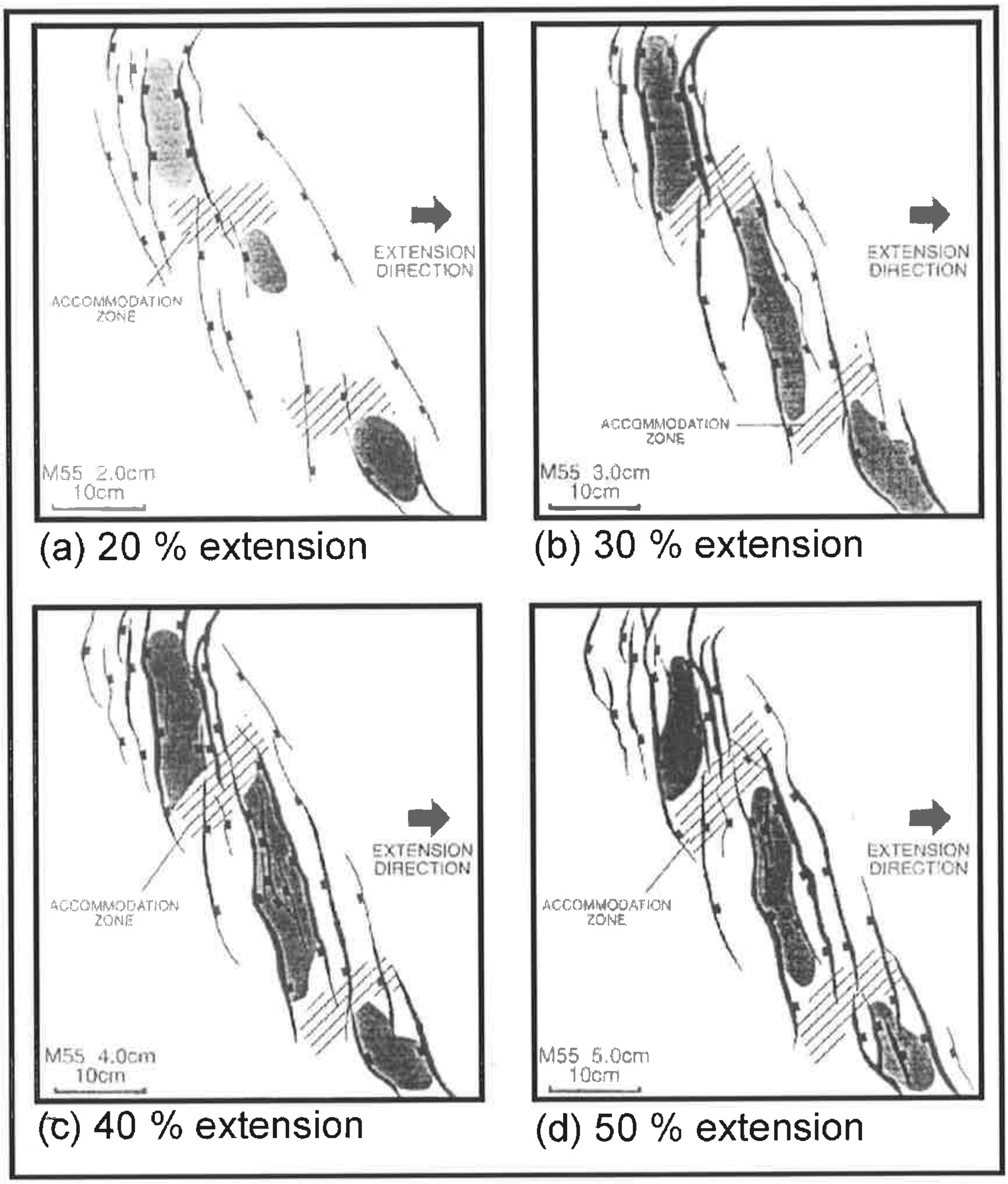


Figure 4.9 Analog sand box modelling results for an oblique rift model at four stages of incremental increase in extension amount. (After McClay and White, 1995)

The rift border fault is segmented and the fault orientation is oblique to the rift zone. The rift interior faults are also not very long, suggesting obliquity of the rift zone to the extensional direction. The fact that distinct depocentres developed during such rifting separated by accommodation zones is also clearly exemplified by the time structure maps at basement and Otway top level (Figures 4.10a and 4.10b).

Cartwright *et al.* (1996) suggested that sub-surface mapping of faults as a single, continuous structures may sometimes be erroneous because on closer scrutiny, sometimes by 3D seismic data (as shown by Needham *et al*, 1996), single traces of faults could be re-interpreted as zones composed of fault segments connected by relay structures or accommodation structures. Given the fact of very broad 2D line spacing in this study area (up to 20-25km), the resolution of the transfer zones or accommodation zones may not be possible except by much closer data coverage. In terms of very large-scale features, viz. Crustal-scale structures, it is felt that this Durroon area is characterised by tilted basement block style structures compared to horst-graben structures in the Bass area and the intervening area perhaps acted as a crustal-scale transfer zone allowing different styles of subsidence to accommodate extension.

Of the three major block-bounding faults (Bark, Anderson, Boobyalla blocks), the north-south trending fault of the Anderson block did not propagate too far and was blocked by the criss-crossing Boobyalla block-bounding fault and the effect is also seen apparently by more rotation of the pre-rift package of this Anderson block (see Figure 4.4). The three major faults in the present day structural pattern (Figure 4.8) seem to have developed to their present day extent in stages (compare with Figures 4.6 and 4.7) by fault tip propagation and coalescing of fault segments. The block bounding fault of the widest block, the Boobyalla block (Figure 4.8), appears to have taken most of the displacement since the Early Campanian. The extension is attributed to far field stresses generated by the beginning of spreading in the Tasman Sea.

Laird (1994) has presented evidence for four separate extensional events on the New Zealand margin of the Tasman Sea during mid to late Cretaceous. They are I) mid to late Albian (about 105 Ma); ii) latest Albian to early Cenomanian (95-100 Ma); iii) early Campanian (80-84 Ma) and iv) late Maastrichtian (70 Ma). From seismic data interpretation and fault trend analysis as discussed in the previous paragraphs, it can be suggested that the Durroon area and the area corresponding to the Moore Basin west of the Lord Howe Rise (Wilcox and Symonds, 1988; Stagg *et al.*, 1999) perhaps formed initially as conjugate pair in the southeastern Australian Continent during the rift valley development prior to opening of Tasman Sea. The good correlation of the three stages of extensional model with the three or possibly four stages of extension on the New Zealand margin of the Tasman Sea (Laird, 1994) supports the sequential development of the rift zone in this area.

A simple shear-pure shear model of extension (Kusznir *et al.*, 1991; Lister *et al.*, 1991; Kusznir and Zeigler, 1992;) has been invoked to explain the major westerly-dipping normal faults seen in the Moore Basin west of the Lord Howe Rise and the ENE-dipping basin-bounding faults in the Durroon Sub-basin.

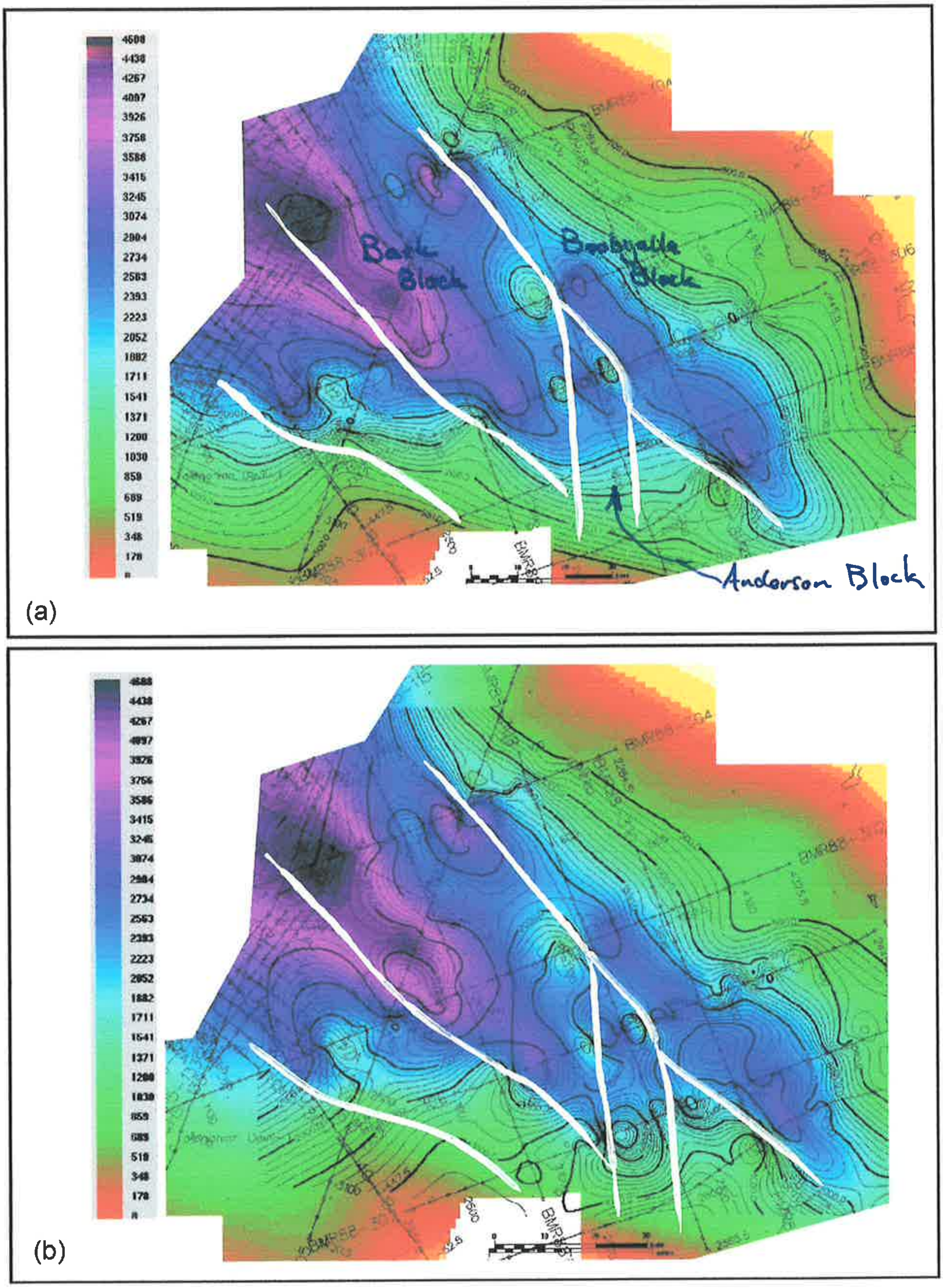


Figure 4.10 Two-way time structure maps at basement level (a) and Otway Top level (b).

(Grid location see Fig. 4.1).

Jongsma and Mutter (1978) hypothesised non-axial breaching of the pre-Tasman Sea rift valley on the limited evidence of pre-breakup tectonism in the southeastern Australian margin suggesting that the whole rift valley remained attached to the central part of the Lord Howe Rise. They had posed two unanswered questions as to whether the Australian crust was too weak before Tasman Sea rupturing and whether the crust could weaken over a long period of time without developing structures which reflect weakening. The observations during the course of the study in the seismic data in the Durroon area clearly indicate that there were large-scale normal fault systems with sediment infilling prior to breakup during Tasman Sea rifting stage (as discussed in the previous paragraphs). But the area that got affected is perhaps the whole Bass Basin area, which is into the hinterland compared to the present day eastern Australian margin. It is proposed that Otway rifting had already created major weakness in the crust in the area and superimposition of later Tasman Sea rift tectonics resulted in an apparently clean, sharp split in the continental crust seen in the vicinity of the shelf and slope area of the SE Australian margin

4.1.5.4 Inversion structures

Although it is now generally established that the Cormorant structure is an inverted structure formed by uplift in the Oligocene-Miocene of a pre-Oligocene trough (Robinson, 1974, Etheridge *et al.*, 1985 and also briefly presented in section 4.2), Hill *et al.* (1995) studied inversion around Bass Basin and noted mid-Cretaceous inversion in the Durroon area. They cited erosional truncation of the Aptian-Albian sequence on the footwall highs during Cenomanian as evidence and supported their view by observations at the Durroon-1 well location. As has been explained in the previous two sections, however, the Tasman Sea rift initiation would entail uplift of all the footwall blocks of any evolving rift half-graben (Barr, 1987, Roberts and Yielding, 1991). The alternative explanation for the erosion of the Aptian-Albian beds (which are proposed here as the pre-rift sequence to the Tasman Sea rift) and the formation of a truncational unconformity better suits the geotectonic framework in the region at that time.

Although Hill *et al.*, (1995) mentioned briefly about Late Cretaceous-Paleogene inversion in northeastern Tasmania and along the Tasmanian margin, particularly during the Late Paleocene, the hard evidence for such an uplift and erosion event is presented in the following paragraphs. The seismic data clearly show that Late Cretaceous and Paleocene beds have been uplifted and eroded perhaps during Late Paleocene-Early Eocene. The evidence of later stage regional uplift and erosion (truncation of reflection) at late Palaeocene-early-Eocene time (M. *diversus* palynological zone) in the line BMR88-306 at its eastern end (between shot points 1670-1800) (Figure 4.11) confirms the missing section in Durroon-1 well at M. *diversus* level. The surface corresponding to the M. *diversus* palynological zone is a prominent unconformity in the Bass Basin around many well locations (Figure 3.3). It is suspected that a large area in both constituting the Durroon and Bass areas was uplifted around this time to develop the M.*diversus* unconformity around many well locations.

Poor stratigraphic well control makes it difficult to define the age but the reflection character in the eastern part of this line (between shot points 1670 – 1800) clearly suggests reverse faulting, folding and perhaps thrusting along some mobile planes within the Otway section and the fault 'fr' (Figure 4.11) clearly shows the sequences from syn-rift sequence II up to Paleocene Top level have been thrusted along this fault 'fr' and the timing is perhaps late Paleocene – Early Eocene.

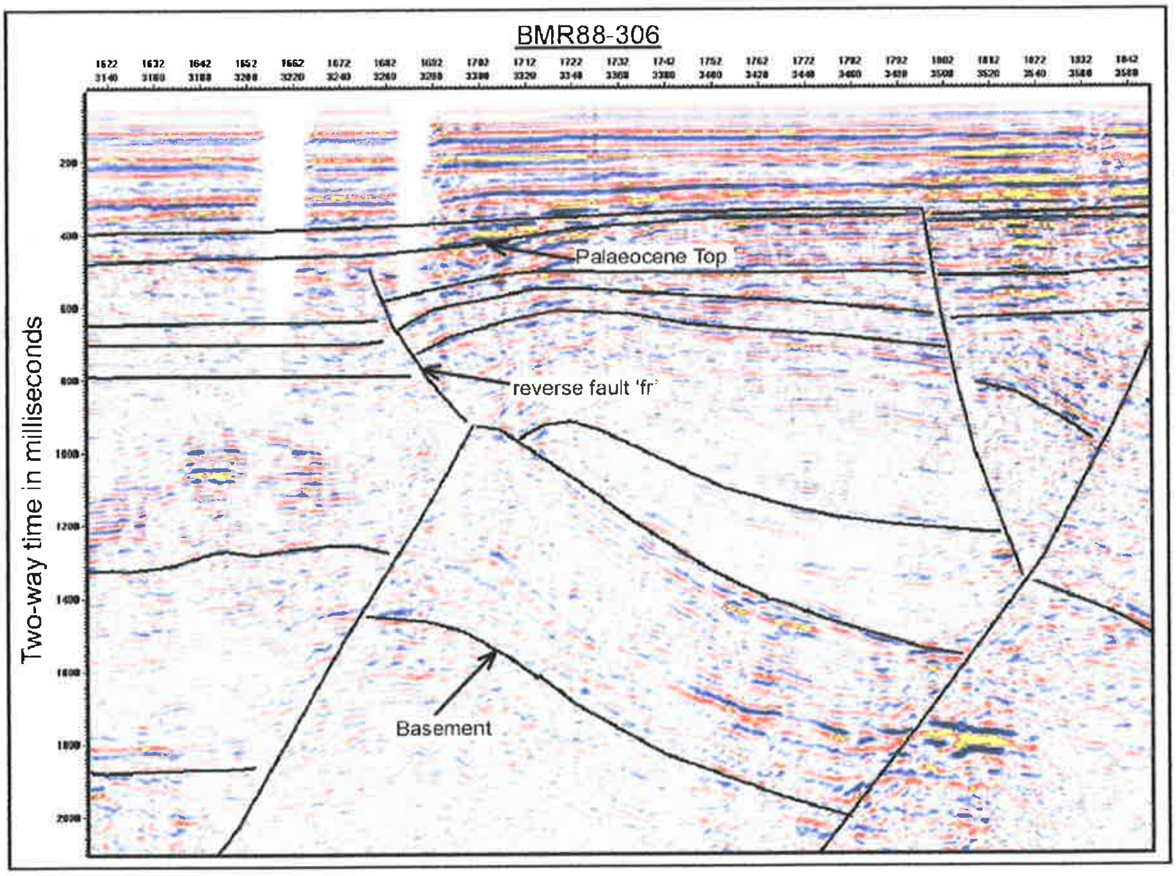


Figure 4.11 Part of the interpreted seismic section BMR88-306 showing the evidence of reverse faulting. Line length is 9.2 km.

The thinning of the Paleocene strata by uplift and possible erosion is clearly seen. It is believed that perhaps the fault 'fr' acted as a mobile detachment plane within some shale layer within the Otway Group (C. Morley, 2001, personal communication). No study thus far on the Bass Basin has observed any reverse faulting or thrusting in this area. Although it is seen only in this line, it is expected that other seismic lines in this area but not considered in this study, will also exhibit this feature. The data quality is not very good in this area and the deterioration of data quality may be an effect of tectonic disruption of the strata. The eventual uplift and emergence of the Flinders Island-Bassian Rise may have initiated during this time.

There are clear indications of reverse faulting in the southwestern part of the Durroon area as seen in seismic line BMR88-307 between SP 2000 and 2300 (Figure 4.12). This area shows reactivation of the pre-existing normal faults at the time of deposition of both the Otway Group and the early syn-rift sequences. The folding associated with the reverse faulting show compressional tectonics were active during the later stage of the basin evolution, probably during Paleocene-Eocene. The important aspect observed along this line is that the Otway Group does not extend beyond SP 2300 that was perhaps the limit of Otway Group deposition (Figure 4.12).

4.1.6 Comments about the break-up unconformity

From palynological analysis, Morgan (1991) recognised two levels in the Durroon-1 stratigraphy at which some palynological zones were missing. He observed A. *distocarinatus* and C. *pannosus* zones of the Early Cretaceous section missing at one level (depth about 5560 ft) and at another (depth 4620 ft), the N.*senectus* and T. apoxyexinus zones were missing. Morgan (1991) ascribed these two major zones of missing section to be unconformities and termed them as break-up unconformities associated with the Otway break-up and Tasman Sea break-up respectively.

Through sedimentary modelling, Barr (1987) showed that for a given fault dip and half-graben load in any evolving half-graben, the footwall crests of fault blocks exceeding a threshold size will become emergent upon extension. He also found that the wider the fault block, the higher is the predicted emergence. As a result of footwall emergence, an erosional unconformity develops at the top of the pre-rift sequence and is generally seen at the fault block crests. No 'break-up' unconformity, however, develops above the syn-rift sequence. The syn-rift section unconformably lies on the pre-rift section (particularly on the eroded crests of the emerging footwall blocks) as extension proceeds in any rift basin. Furthermore, Roberts *et al.* (1993), in a very detailed study of the Brent / Stratfjord fault block in the North Sea, modelled the sedimentary section in the fault block and they have shown that in a modified domino model involving two phases of extension, a prominent syn-rift unconformity also develops. This modified domino model then fits the North Sea example, as there were two major pulses of extension in the North Sea area during the Triassic and Jurassic.

The Durroon Sub-basin has been influenced by two rifting events related to the Otway rift stage and the Tasman Sea rift stage. It is proposed, in the light of the North Sea Basin modelling

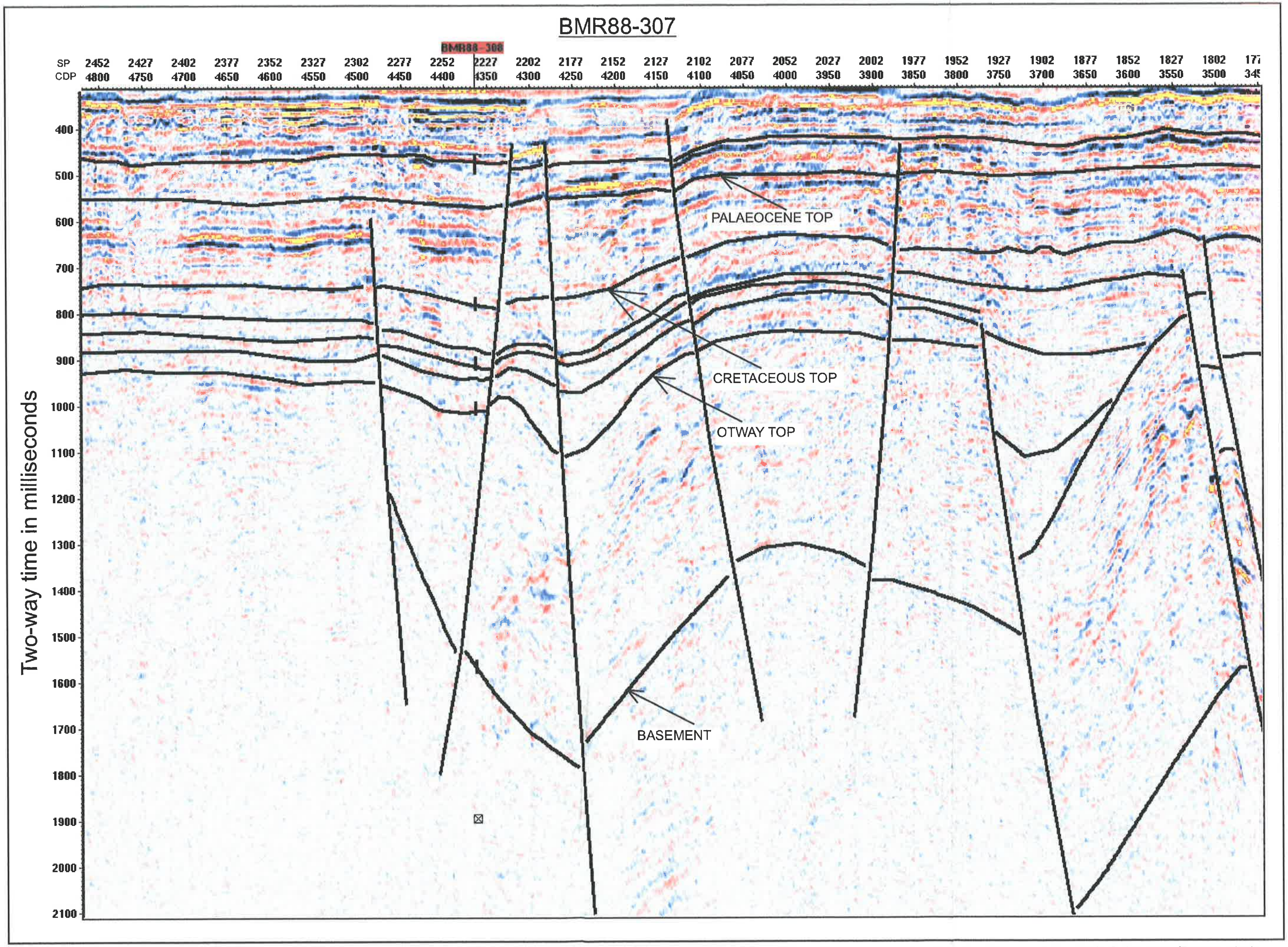


Figure 4.12 *Part of the interpreted seismic line BMR88-307 showing evidence of compressional structures as reactivation of earlier existing normal faults. Note the thinning of the Early Paleocene-Eocene strata by uplift and perhaps erosion. Line length is 26 km.*

results, that the Durroon Sub-basin be interpreted along the lines of a modified domino model. The fact that there are two prominent unconformities as per Morgan (1991) can be explained as follows. Near the Durroon-1 well which lies on an emergent crestal fault block in the SE Bass Basin, the syn-rift section lies unconformably on the pre-rift section. This first unconformity (described by Morgan, 1991 as the Otway break-up unconformity) can be more appropriately defined as the **'rift-onset unconformity'** related to the beginning of Tasman Sea rift stage. Similarly the major sequence boundary corresponding to the other major unconformity (top of the syn-rift sequence I) can be compared with the syn-rift unconformity (as seen in the Brent / Stratfjord fault block) and can be considered as having developed within the syn-rift phase of sedimentation in a modified domino model fashion. A similar pattern of movement encompasses the Golden Beach Group in the Gippsland Basin (Hill *et al*, 1995)

4.1.7 Influence of the basement grain and architecture

Gunn *et al.* (1997a) carried out detailed analysis of the aeromagnetic data over the Bass Basin and magnetic and gravity data of the onshore and offshore Tasmania (Gunn *et al.*, 1997b). They have interpreted the basement elements of the region in good detail. Figure 4.15 shows the basement elements of Tasmania and the adjoining offshore areas.

There are a few important elements of the basement architecture that seem to have influenced strongly the Late Jurassic-Early Cretaceous basin evolution. (i) The regionally extensive granitic mass to the east of the entire Tasmania and Bass Basin region. (ii) The massive Mesozoic granitic intrusion in the central part of the Bass Basin (iii) The Arthur metamorphic complex which extends from northwestern Tasmania well into the offshore Bass Basin.

The well completion report of Bass-3 suggests that the basement encountered at this well consists of metasediments which support the view that the nature of the basement in the western and northwestern part of the Bass Basin is made up of metamorphosed sediments as against the granitic basement to the east and southeast. Hence, mineralogically, the nature of the basement is different in the Bass area compared to the Durroon area. This difference in the type of basement rocks might have influenced the way the upper crust behaved in response to the extensional stresses. We observe that the upper crust behaved in a more brittle manner in the Durroon area where the upper crust ruptured by few large-displacement faults leading to large tilted basement blocks. In the Bass area, however, the upper crust was ruptured by large number of small throw faults except the two major half-graben bounding faults (Compare figures 4.4 with 4.17).

The Arthur metamorphic complex and its offshore extension perhaps had different influence on the overall structural evolutionary history of the area during the two stages of rifting (Otway and Tasman). It is proposed that the Arthur metamorphic complex acted as a transfer zone or

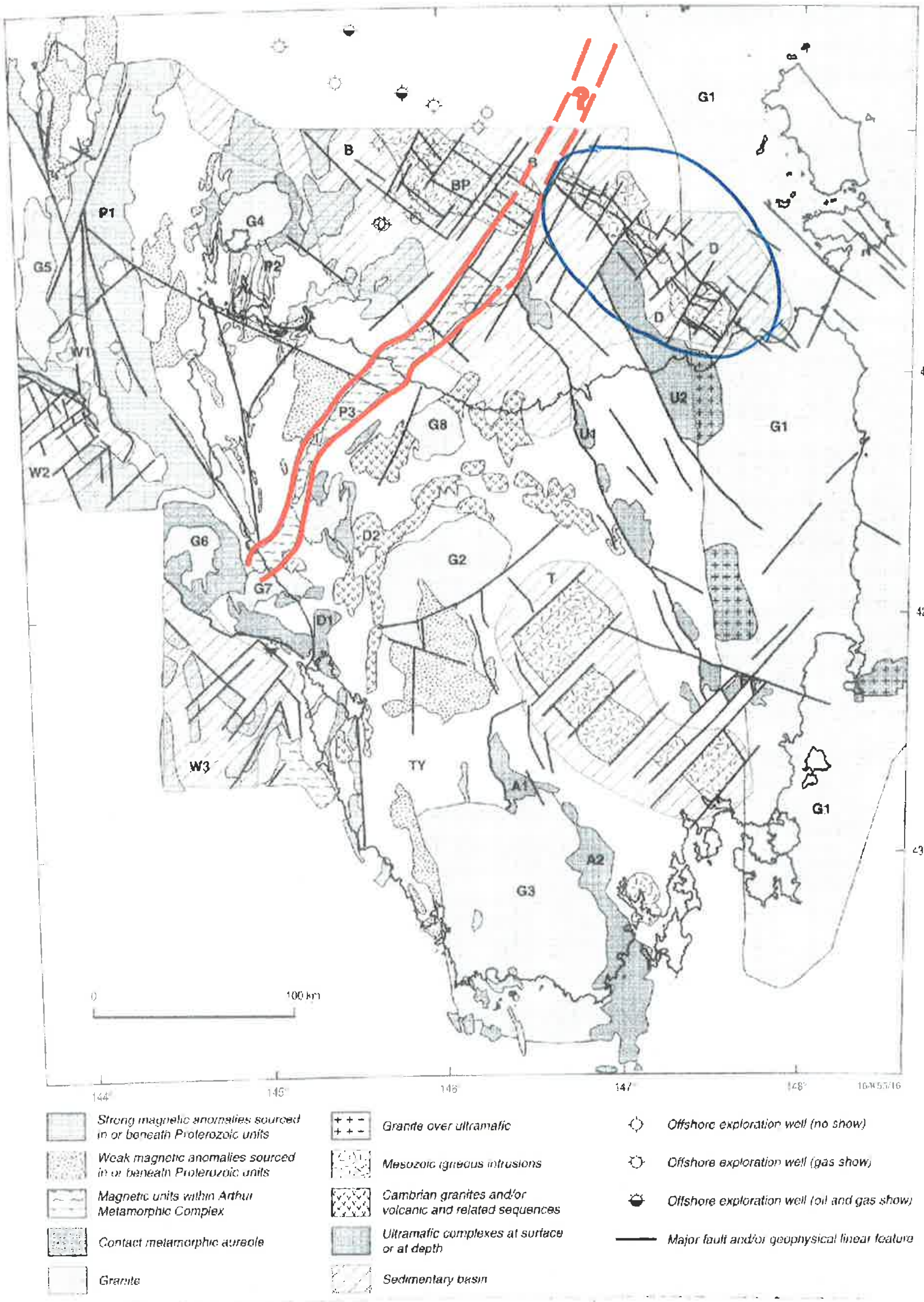


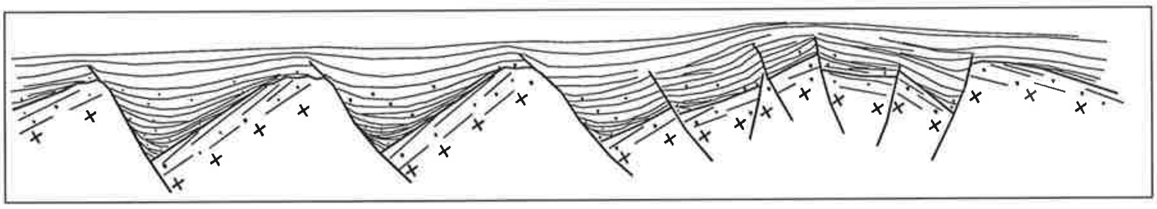
Figure 4.13. *Basement elements of Tasmania and adjoining areas (after Gunn et al., 1997b). Note the Arthur metamorphic complex and its offshore extension marked in red. The fault pattern in the Durroon Sub-basin (circled) derived from seismic interpretation broadly agrees with the NW-SE set but does not support the NE-SW offsets interpreted from the magnetics.*

accommodation zone between the Durroon and Bass area during the Otway rifting stage. The two areas developed different styles of faulting during this phase of extension as explained above. This ultimately resulted in the different extensional subsidence styles in the two areas (the Bass area showing horst and graben styles and the Durroon area showing the tilted basement block style) with the transfer/accommodation zone acting as a mechanism to separate the two styles.

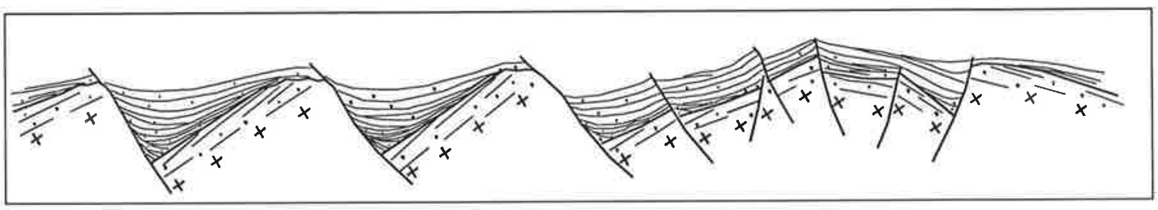
In the second stage of rifting (Tasman rifting), the extensional stresses were perhaps restricted to the Durroon area only. Smith (1986) has suggested that the Durroon Formation is unique to the Durroon area and was deposited during Tasman rifting stage. The prominent tilted basement blocks of the Durroon area and the rift-related depositional systems tracts identified clearly in this area point to this fact. The time equivalent to the Durroon Formation in the Gippsland Basin, the Golden Beach Group, was also deposited during active rifting on tilted basement blocks.

4.1.8 Tectonostratigraphic model showing progressive geological evolution of the Durroon Sub-basin from Palaeozoic to Recent

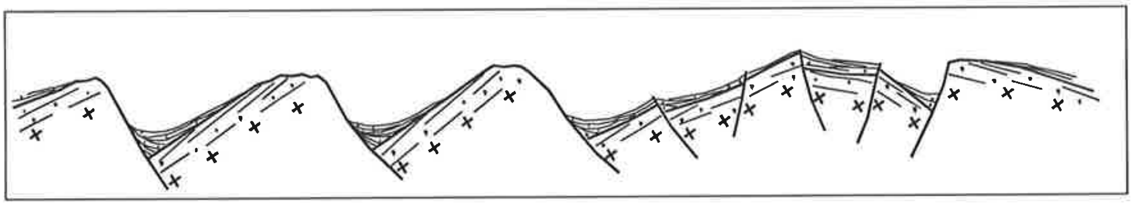
Figure 4.14 shows a conceptual model showing the progressive tectonostratigraphic evolution of the SE Bass Basin (Durroon Sub-basin) from Palaeozoic to Recent. The undulating Palaeozoic basement initially filled with sediments of Permo-Triassic or earlier age. During the Late Jurassic-Early Cretaceous, Otway rifting commenced in the eastern Gondwana creating faulted depressions. Volcanoclastic sediments were deposited in these faulted depressions. This was followed by a relatively quieter phase during which the influence of Otway rifting was subdued and is marked by parallel-sub parallel reflectors. Fluvio-lacustrine and deltaic facies were deposited with prominent coaly facies.



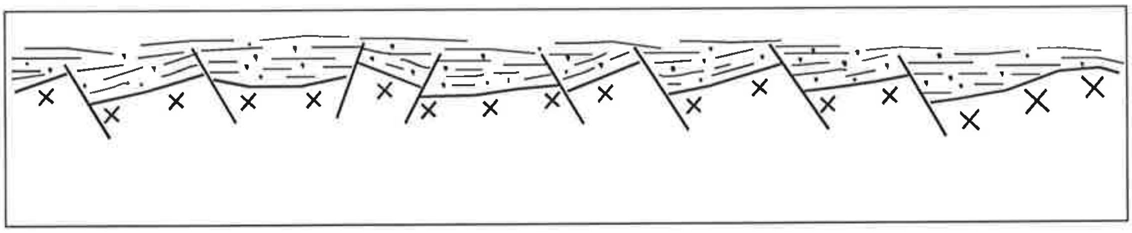
Stage - VI (Eocene-Recent) Complete cessation of rifting and passive subsidence during oceanic spreading in both Otway and Tasman rifts was accompanied by regional sedimentation of silliciclastics. Since Oligocene, open marine influence caused self carbonates to be deposited.



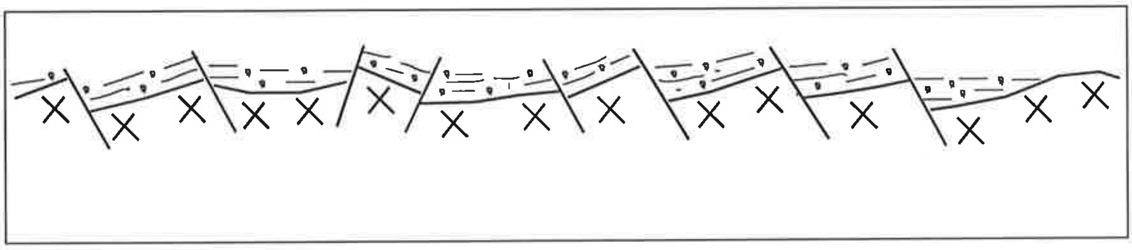
Stage - V (Late Cretaceous-Paleocene) Passive fill of the depressions interpreted from the onlapping reflectors onto the hanging wall and footwall blocks. Individual tilted blocks may be characterised by different lithofacies. Draping of the crests by continuous reflectors indicate end of faulting/rifting.



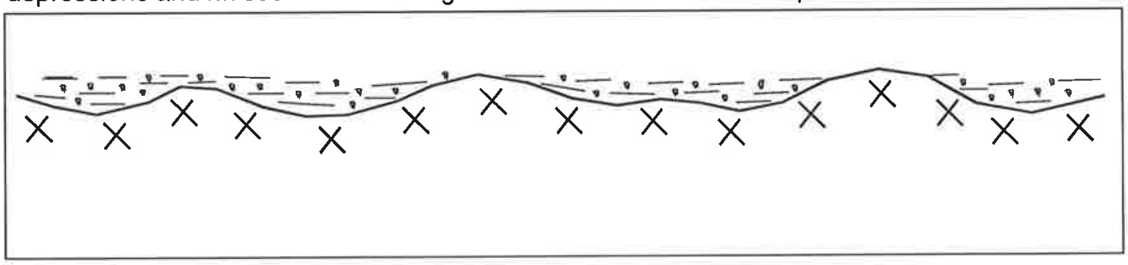
Stage - IV Early Cenomanian Tasman rifting reactivated the faults with large displacements. The Bark, Anderson and Boobyalla blocks became prominent. Fault scarp material deposited along both hanging wall and footwall slopes.



Stage - III Waning period of Otway rifting (Aptian-Albian) saw sedimentation spread, represented by parallel-subparallel reflectors of the Otway Group. Fluvio-lacustrine and deltaic facies with coal measures deposited.



Stage-II The Late Jurassic - Early Cretaceous faulting due to Otway rifting created faulted depressions and rift sedimentation began. Volcanoclastic sediments predominate.



Stage-I The undulating Palaeozoic basement filled locally with sediments of Permo-Triassic or earlier age.

Figure 4.14 Conceptual tectonostratigraphic model of the Durroon area (Durroon Sub-basin).

In the next stage, the initiation of the Early Cenomanian Tasman rifting reactivated the existing faults of the Otway rifting phase with massive displacement, leading to prominent Bark, Anderson and Boobyalla fault blocks. The fault scarp material derived from the emerging footwall blocks was deposited along both the hanging wall and footwall slopes. During the Late Cretaceous – Paleocene, the enormous depressions created by faulting were passively filled by lacustrine facies. This is interpreted from the reflectors onlapping the hangingwall and footwall blocks.

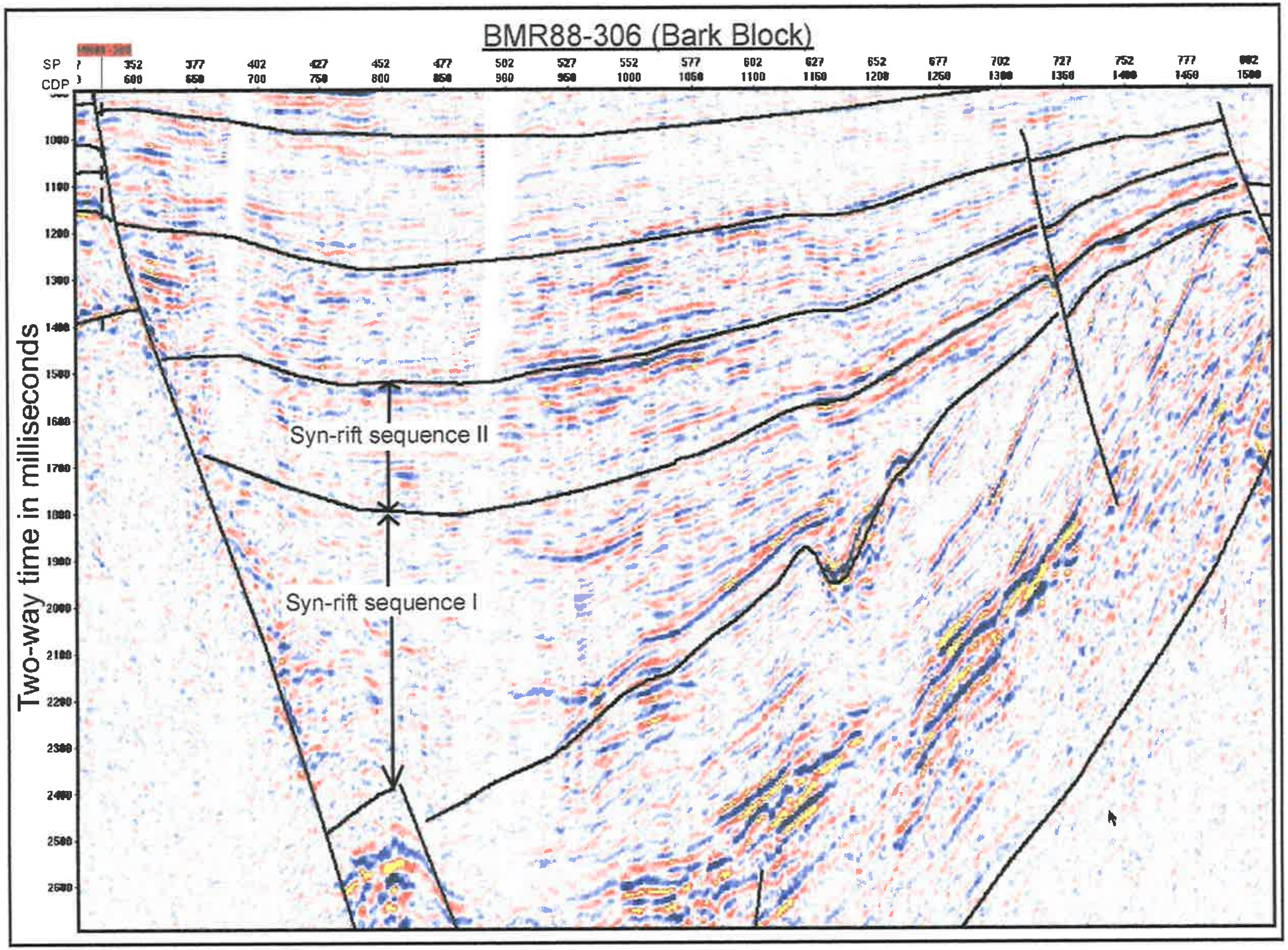


Figure 4.15 a Part of the interpreted seismic section BMR88-306 (Bark block). Line length is 18.7 km.

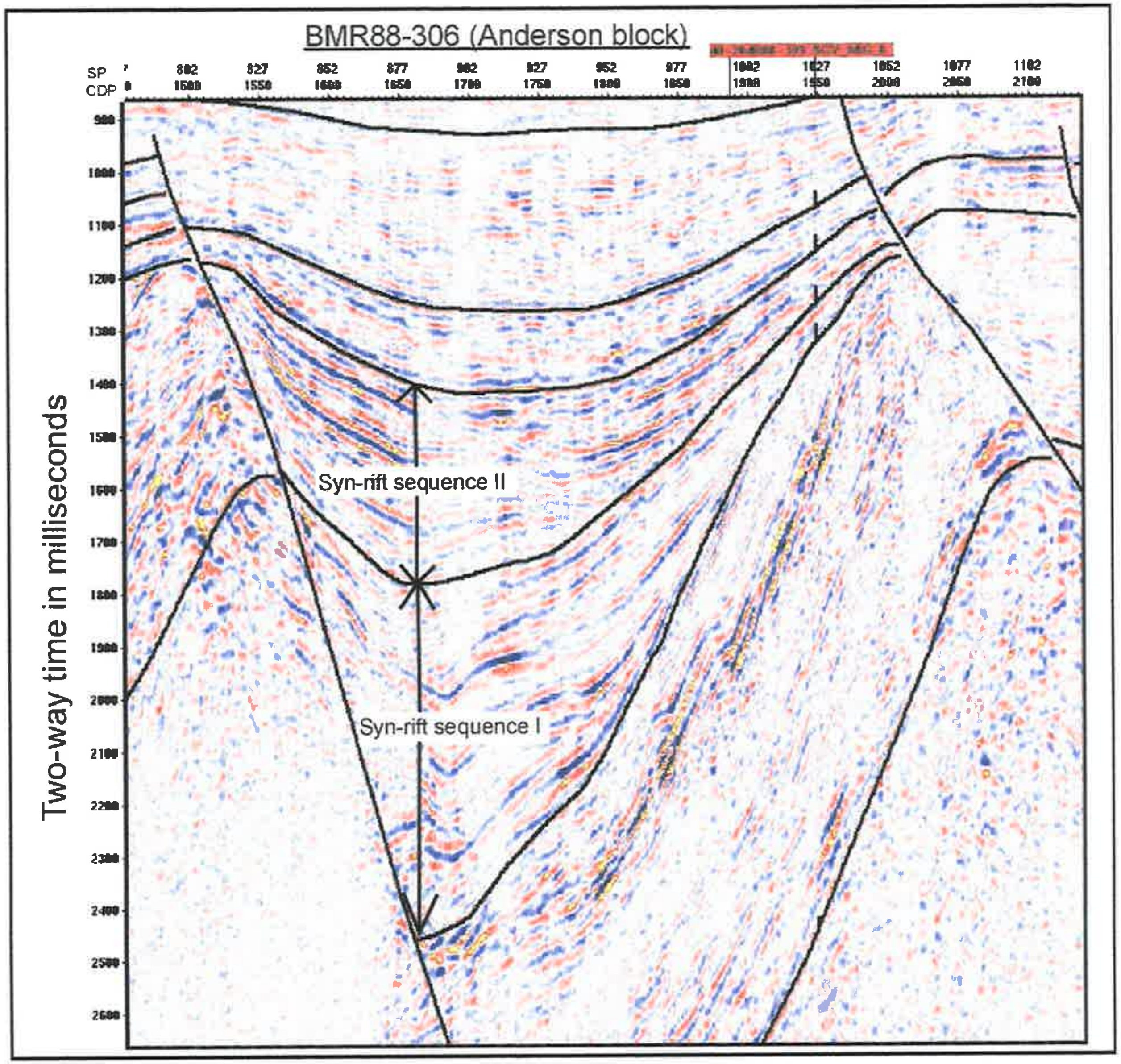


Figure 4.15b Part of the interpreted seismic section BMR88-306 (Anderson block). Line length is 14 km.

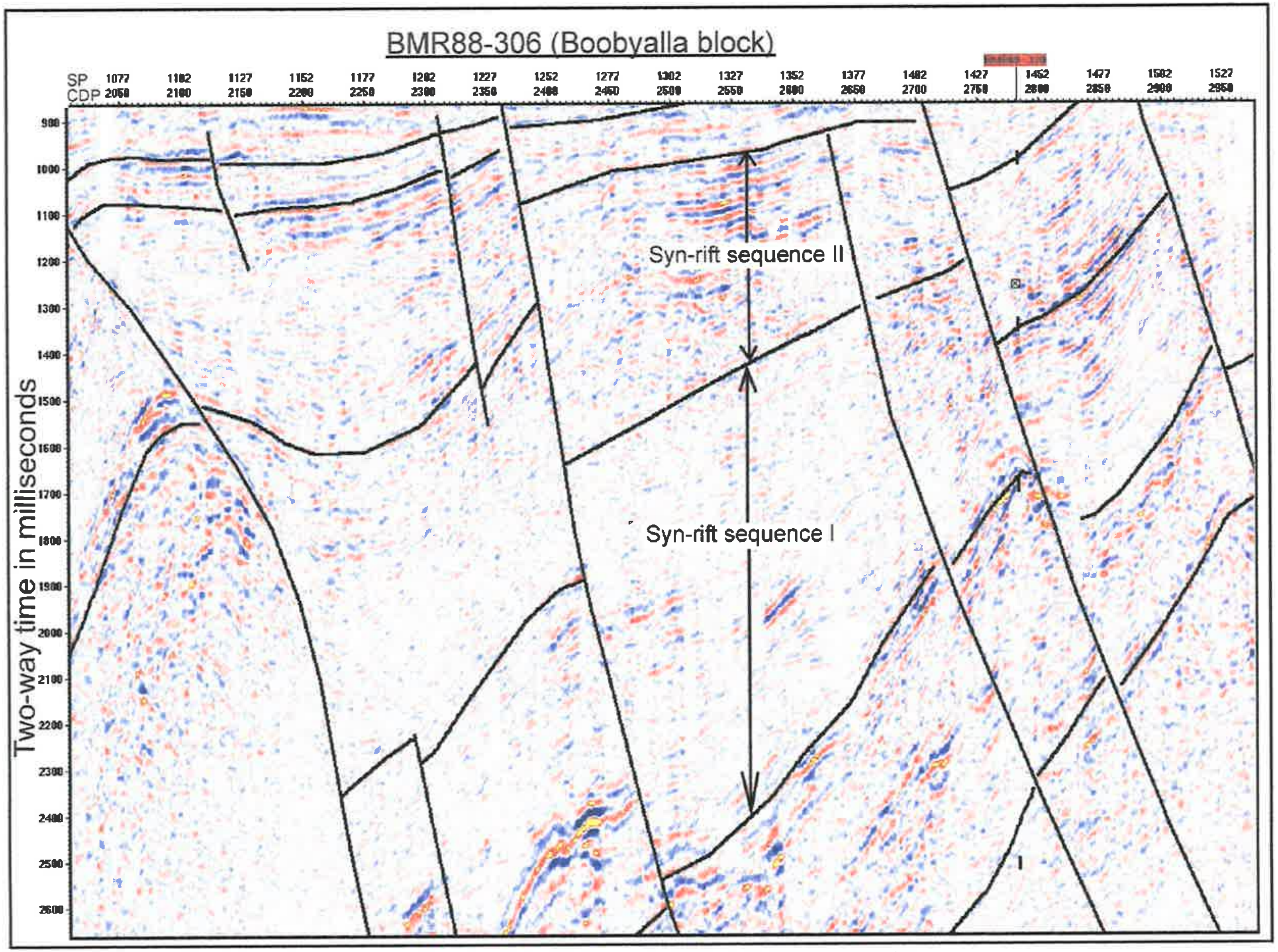


Figure 4.15c Part of the interpreted seismic section BMR88-306 (Boobyalla block). Line length is 18.7 km.

Here, it is important to note that the sedimentary fill on top of individual fault blocks may be characterised by different lithofacies. The nature of the reflector packages in terms of amplitude, frequency and continuity in the three major fault blocks is seen to be different (Compare the three figures 4.15a, 4.15b and 4.15c). It is quite obvious from the seismic facies that the lithofacies within these three independent depocentres is not exactly the same and any attempt to correlate the stratigraphy in a lithostratigraphic manner would be erroneous. The reason could be the nature of the source composition, the rate of subsidence in the individual blocks and hence rate of sedimentation in conjunction with the nature of fluctuation of lake-level or base level in the area. Ravnas and Steel (1998) showed that facies might change from block to block as sediment is channeled through by-pass zones along relay ramps. Although the seismic facies in the syn-rift sequence I in Bark and Anderson blocks look similar and are characterised by medium amplitude-medium frequency with good continuity, the continuity in the Anderson block is better than in the Bark block. It is interpreted from seismic analysis alone that the lithofacies in the Anderson block may be having good sand/shale layering to give rise to good contrast and hence better continuous reflections. This indicates the source composition or the environment of deposition may be somewhat different. The other factor responsible for picking good reflection continuity is that the energy conditions for deposition of this sequence was perhaps quiet and stable. The syn-rift sequence in the Bark block may be more arenaceous and hence lack of possibility of good sand/shale contrast. The syn-rift sequence I in Boobyalla block is however, marked by very poor reflection continuity and the amplitude is very low except some patchy reflectors. The energy condition and also the source composition in this block may be totally different compared to the other two blocks.

Draping of the crests by continuous reflectors indicates an end to the faulting / rifting. This will not necessarily be at the same time on each fault block. The details of all these factors can be better worked out only when more drilling results are available from the individual fault blocks. In the last phase, i.e. during Eocene-Recent, there was regional sedimentation of siliciclastics deposited in response to the cessation of rifting and passive subsidence during oceanic spreading in the Otway and Tasman rifts. During this period, the area was in a thermal cooling phase with minimal if any, tectonic activity and therefore, a time of deposition of regionally continuous strata. Since the Oligocene, an open marine influence is marked by deposition of shelf carbonates in the area.

4.2 Bass area (Burnie Sub-basin)

4.2.1 Database, data quality and methodology

As already discussed, the Bass area or the Burnie Sub-basin has been considered separately during the seismic interpretation work. The data density is much better in this area (average line spacing is about 5 km) and also the data quality is better for a majority of the lines particularly those of the TQH5 and TNK4 series. The data volume corresponds to all the lines in the Table 3.2 sans those lines in Table 4.1. Figure 4.16 shows all the 2D seismic lines used in the Bass area for interpretation.

The oldest rocks encountered in this part of the basin correspond to the Bass-3 well where the basement was encountered at a depth of 2350m. The basement is overlain unconformably by Upper Cretaceous sediments. The few wells that encountered Upper Cretaceous sediments in this part of the basin are Aroo-1 (Upper T. *longus* palynological zone), Bass-3 (Lower T. *longus)*, Dondu-1 (Upper T. *longus*), Konkon-1 (Upper T. *longus*), Koorkah-1 (T. *lilliei*), Pelican-5 (T. *lilliei*), Poonboon-1 (Lower T. *longus*) and Tilana-1 (Upper T. *longus*). There is no well that encountered Lower Cretaceous sediments in this area.

Seismically, as discussed in 4.2.2, the Otway Top has been clearly recognised as the rift-onset unconformity corresponding to beginning of Tasman Sea rifting. The Otway Group is considered as the pre-rift package. The whole stratigraphy in this part of the basin has been divided broadly into several major sequences utilising the available biostratigraphic information and correlated to seismic reflection surfaces through synthetic seismograms at several well tie positions. The broad divisions are basement, top of pre-rift package (Otway Top, although no actual age dating from drilling samples is available), the top of syn-rift package (the exact palynological age dating has not been possible due to lack of deeper well information), early post-rift package (Top Cretaceous), top Palaeocene and late post-rift package (two sequence boundaries corresponding to Top Late Eocene and Middle Miocene have been correlated and mapped here).

The seismic imaging of the deeper section of the stratigraphy is poor and the reflection quality deteriorates with both amplitude diminishing and continuity being very poor. Particularly beyond 2.5 seconds in the major trough areas, the data suffers from very low amplitude, low frequency and poor continuity. The composite line shown in Figure 4.17 consisting of parts of three separate lines (TQH5-33, TQH5-16A and TQH5-73) goes across the basin from the west side to the east side, cutting across the general basin strike (see Figure 4.16). This line presents the best possible quality of data in the entire Bass area in terms of deeper imaging quality. Although no well sits close to this line orientation, the wells Bass-3 and Aroo-1 have been tied with the nearest lines TQH5-35 and TQH5-29A (see Figure 4.16) and loop tied in the area. Full interpretation of this composite line with sequence boundary identification is discussed in detail in 4.2.2.

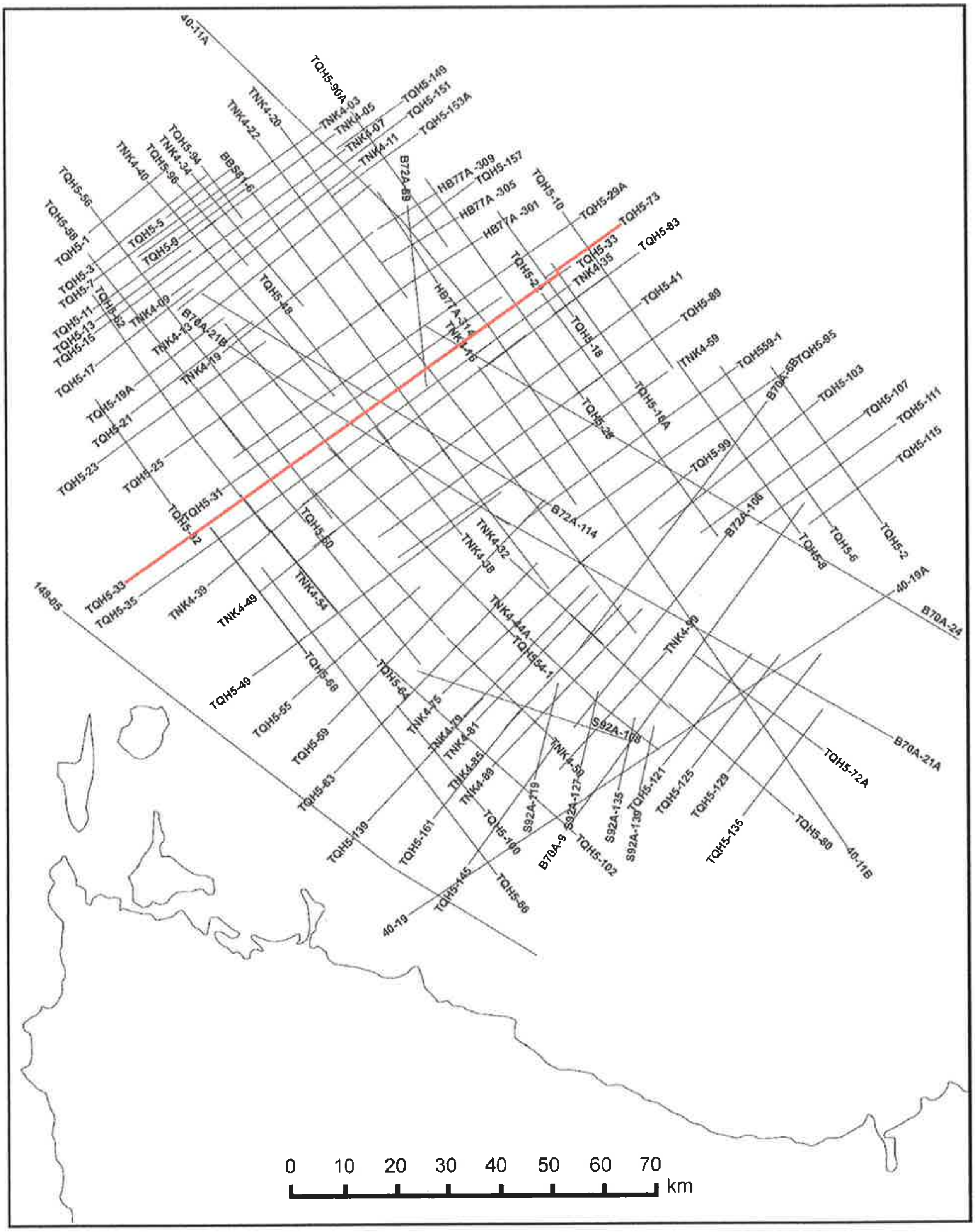


Figure 4.16 Location map showing the distribution of the 2D seismic lines used in the Bass area. The composite seismic line shown in Figure 4.17 is marked in red here.

4.2.2 Seismic sequence stratigraphic interpretation

The good quality data displayed on Figure 4.17 illustrates the mapped packages. The stratigraphy has been subdivided into several seismically identifiable sequences. Unlike in the Durroon area, the clear-cut identification of different syn-rift sequences is extremely difficult for three main reasons. The seismic data quality of the rift related syn-rift sequences is not very good as these units occur below 2.5 seconds two-way time. There is no well penetration to build confidence in the seismic picks and the clear, simple tilted domino fault block style structuring as observed in the Durroon area is not present. However, the top Otway unconformity has been picked with fair confidence following the reflection geometry and configuration in the Otway Group strata compared with the overlying sequence. The nature of reflection terminations of the syn-rift sequence strata is onlap onto the hangingwall with downlap on the footwall slope in the Yolla Trough (Figure 4.18a) and onlap onto the footwall slope in Cormorant Trough (Figure 4.18b). The syn-rift strata in the Cormorant Trough do not show onlap onto the hangingwall slope. Rather, there is evidence of continuous sedimentation onward from Otway time although wedging out onto the bounding fault is noted. It suggests that a large body of water in lake systems probably did not develop in the Cormorant Trough during this syn-rift stage as is very clearly evident in the Durroon area. The amplitude and frequency content are both low, indicating further that syn-tectonic sedimentation continued in the Cormorant Trough area. This is different in the Yolla Trough as observed from the seismic facies characteristics. The wedging of the syn-rift sequence towards the major block-bounding faults (Fault F3 in the Cormorant Trough and Faults F1 and F2 in the Yolla Trough) is also clearly evident (Figure 4.18a and 4.18b). Displacement patterns along Fault F4 in Figure 4.18b suggests that movement along this fault did not start until the Cretaceous sediments were deposited because both the Otway and syn-rift packages do not show evidence of any growth of strata, unlike Fault F3.

There is evidence of erosional truncation of the Otway Group strata (Figure 4.19) in the Yolla Trough. The Otway Group strata show parallel to sub-parallel internal reflection geometry and overall they are parallel to the basement which suggest that the Otway strata predate the Tasman rifting episode as discussed in the Durroon area (Figure 4.19).

The available palynology data for all the wells supplied by Mineral Resources, Tasmania do not show any missing section on the Cretaceous-Tertiary boundary (Aroo-1, Bass-3, Koorkah-1, Pelican-5, Poonboon-1 and Tilana-1). The reflection surface corresponding to the Cretaceous Top boundary correlated from Aroo-1 synthetic data suggests very clear truncation of some tilted strata (see Figure 4.20) at this level which means that definitely there is missing section at this palynological boundary. The biostratigraphic information is therefore confusing and is very similar to the problem that was discussed for Durroon-1 well (section 3.3). Revision of palynology should therefore clarify such problems. There is however, one mention in the well completion report of Koorkah-1 (Amoco, 1986) about a Maastrichtian angular unconformity which correlates well with what has been observed from seismic data near Aroo-1 well.

By T. *lilliei* age, the Durroon area reached the early post-rift stage, indicating relative quiescence from major tectonic activity. Here in the Bass area, however, it is clear that up to Maastrichtian age, there is tilting and erosion of some strata indicating thereby that the tectonic activity continued to this time. It is proposed that rapid sedimentation along the fault-bounded depocentres initially created by fault movement continued as sediment loading added to the now waning extension-driven subsidence. This was sufficient to continue tilting of the strata with subaerial erosion still possible on the high points of the blocks.

The other possibility is that the by end of Late Cretaceous / Early Paleocene (Maastrichtian age), the whole region including these basin depocentres (around Aroo-1 well) were peneplaned following uplift and denudation. There is evidence of regional Palaeocene / Maastrichtian peneplanation (Hill *et al.,* 1995; Wilcox *et al.*, (1992) and Duff *et al.*, 1991) of the Late Cretaceous sequences overlain by a basal Tertiary sequence of coastal plain to nearshore barrier systems (Bodard *et al.*, 1985). These authors have suggested that the regional peneplanation may have been associated with Late Cretaceous-Palaeocene compressional deformation recorded in the Otway Basin, west of the Sorrel Fault (Hall, 1994).

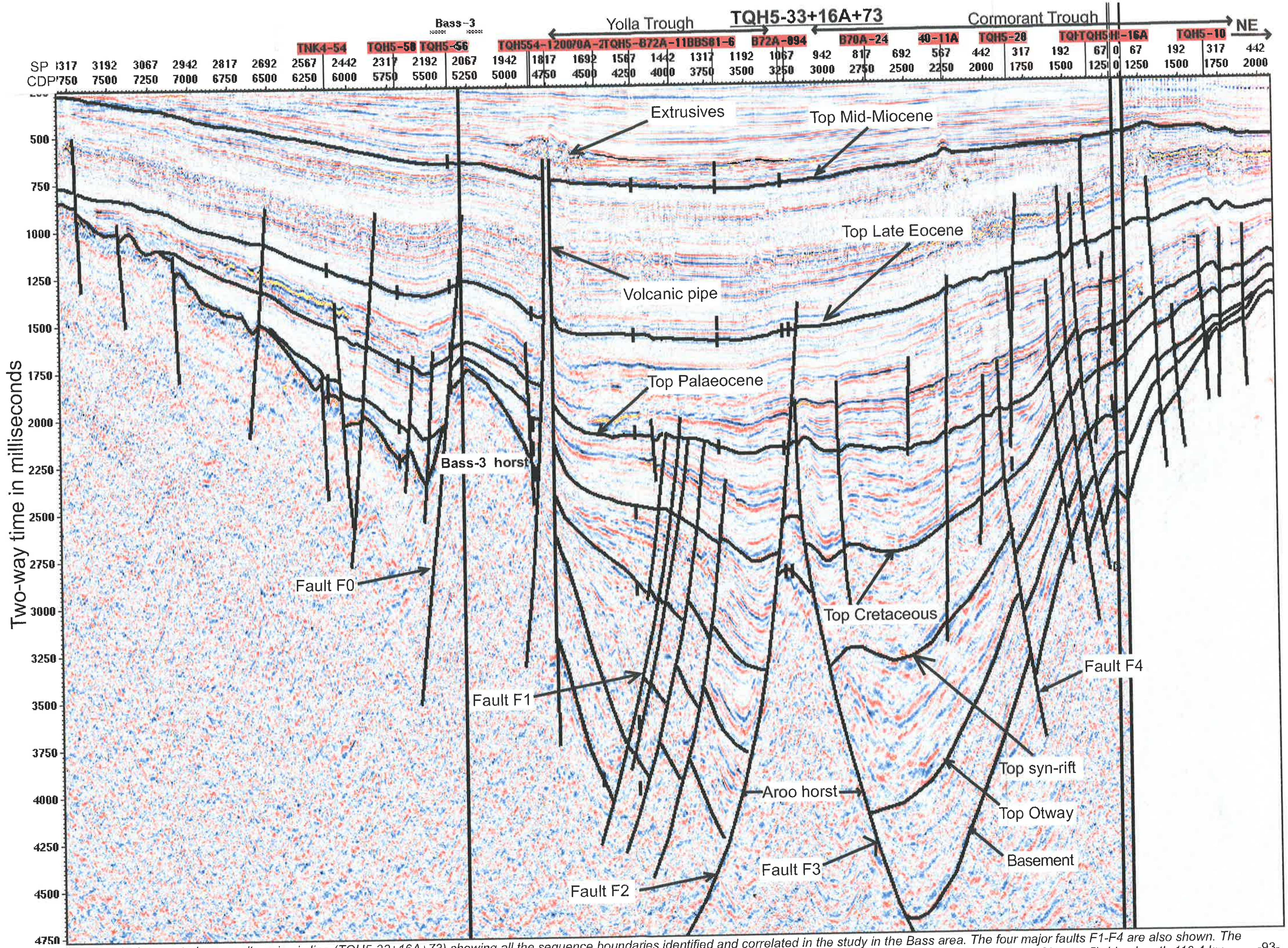


Figure 4.17 *Interpreted composite seismic line (TQH5-33+16A+73) showing all the sequence boundaries identified and correlated in the study in the Bass area. The four major faults F1-F4 are also shown. The sequence corresponding to Late Oligocene to Mid-Miocene shows layer-bound polygonal fault system of extreme high density but have not been marked here (discussed in Chapter-5). Line length-116.4 km.*

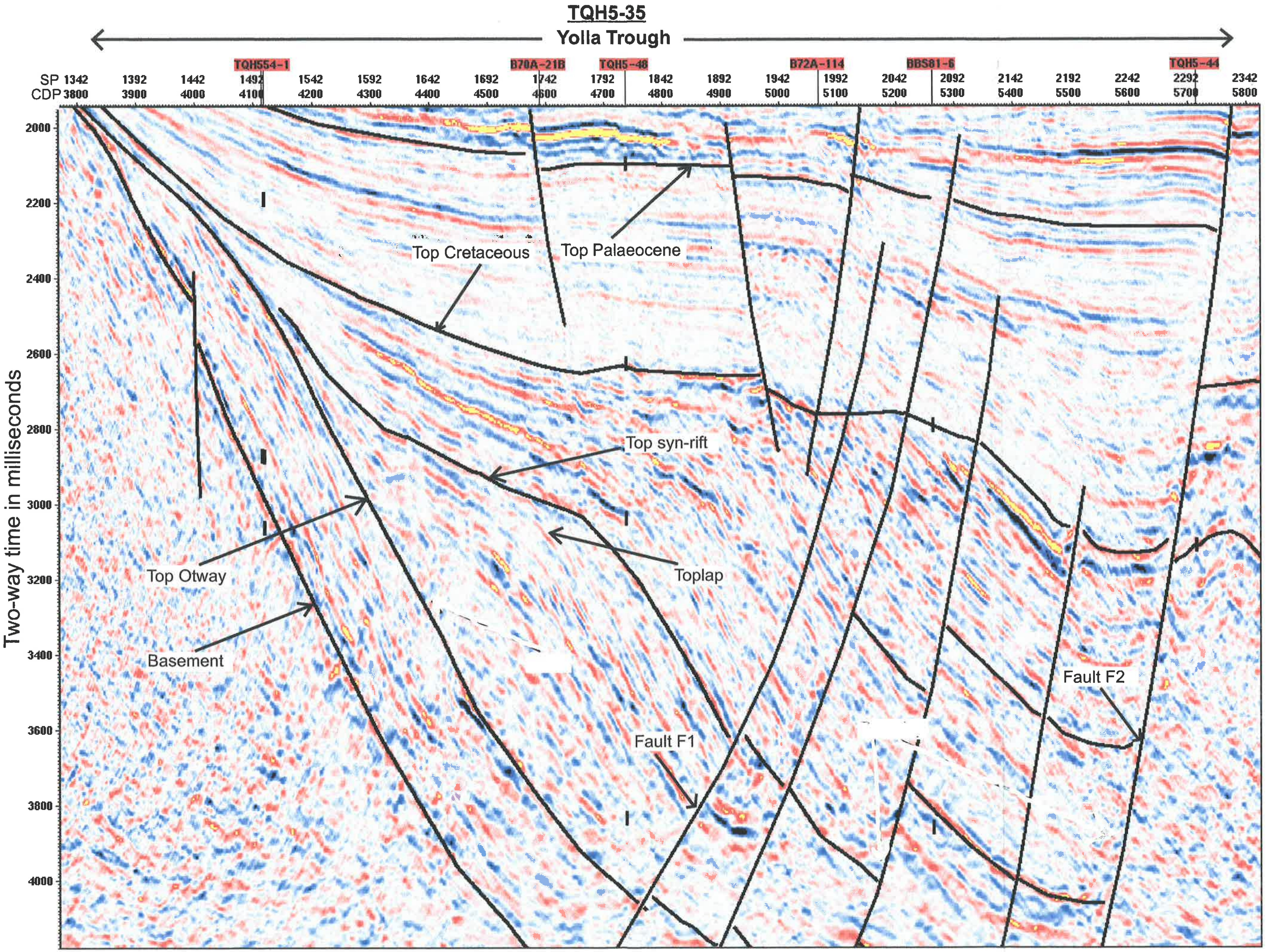


Figure 4.18a *Part of the interpreted seismic line TQH5-35 showing reflection termination patterns in the Yolla Trough. Note the wedging of the syn-rift package against faults 'F1 and F2. Line length - 30.8 km.*

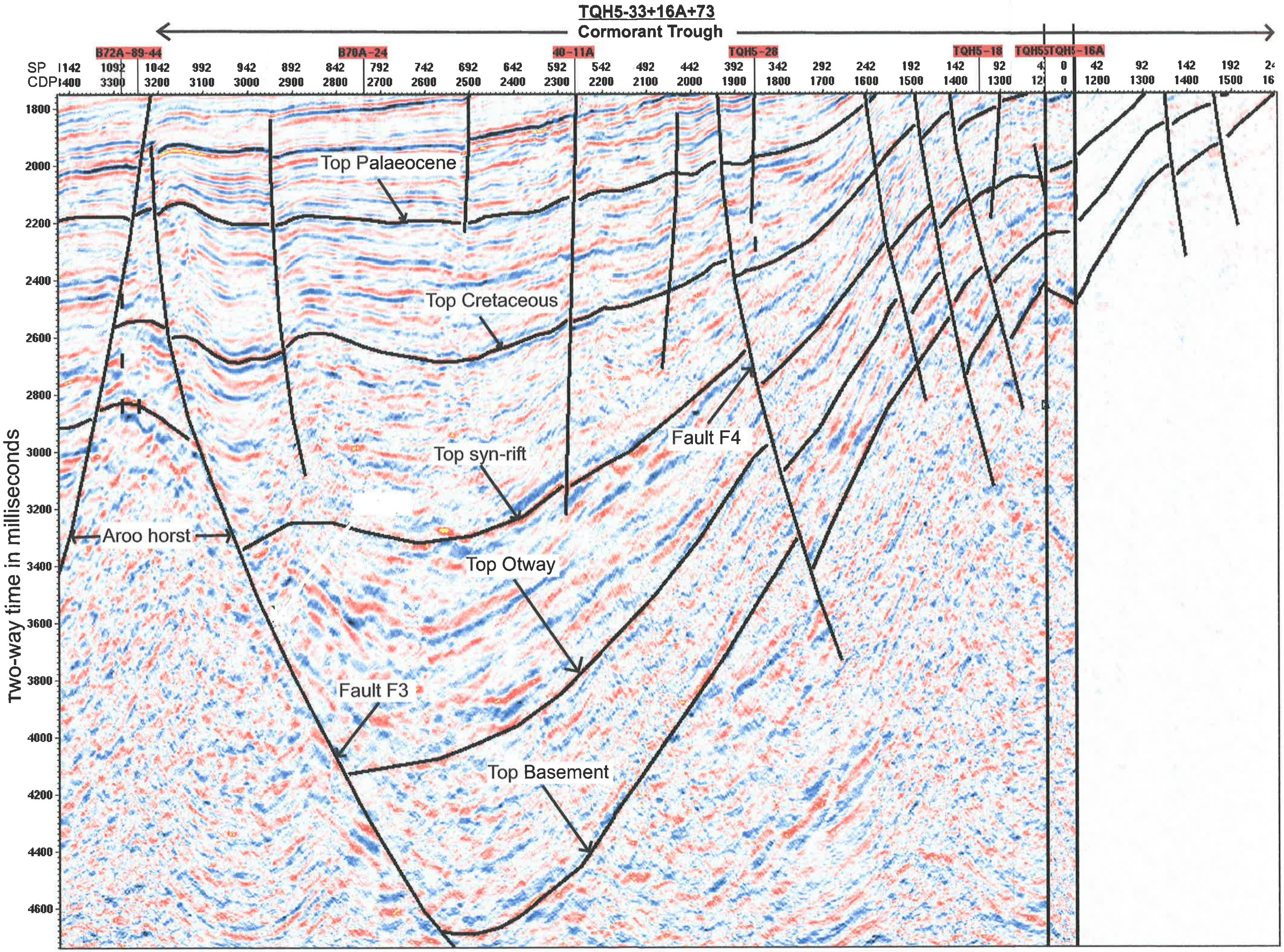


Figure 4.18b *Part of the interpreted composite seismic line (TQH5-33+16A+73) showing the seismic reflection termination patterns in the Cormorant Trough. Note the wedging out of the syn-rift and Otway sequence against the main bounding fault F3. Line length- 40.2 km.*

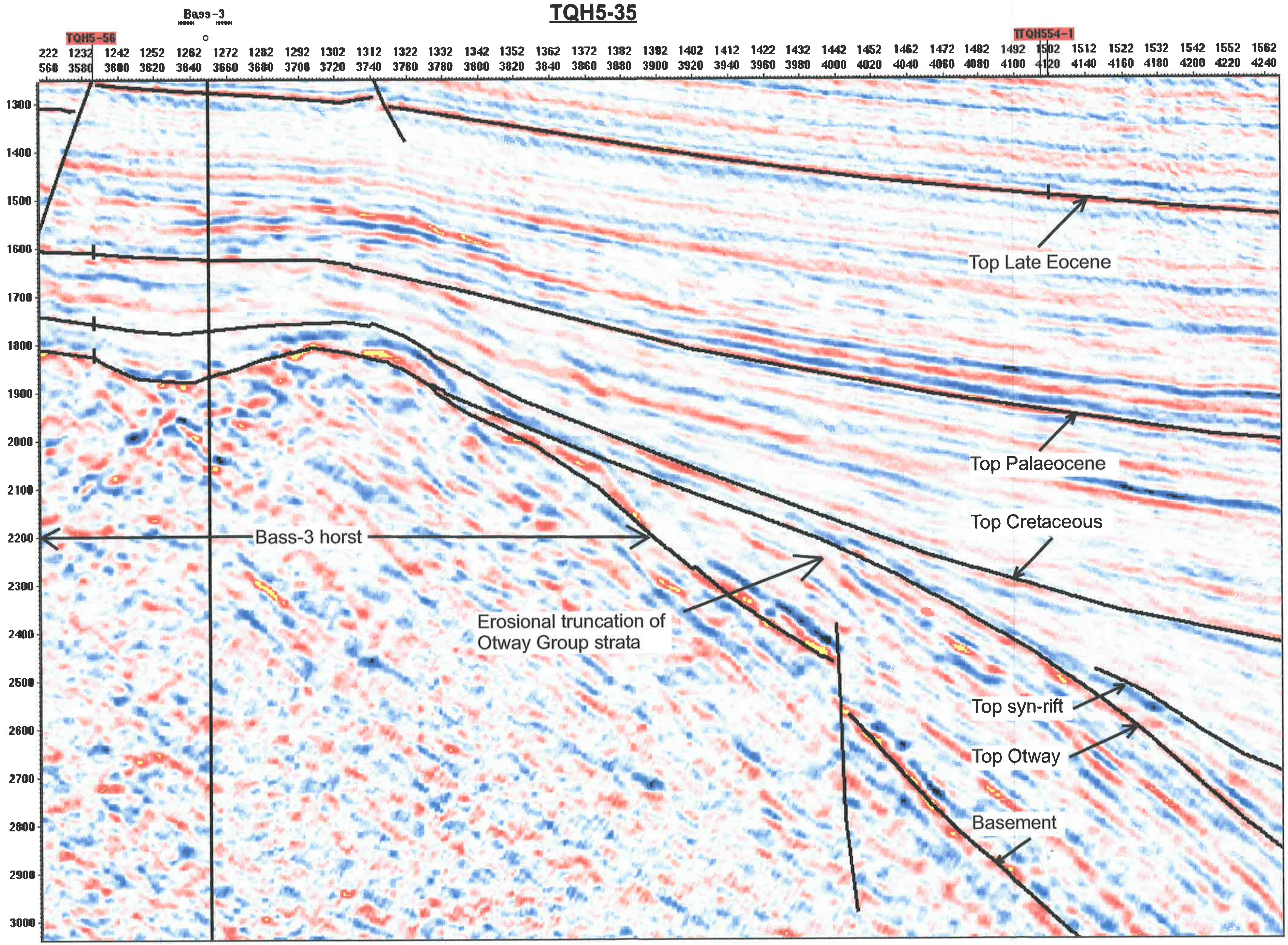


Figure 4.19 *Part of the interpreted seismic line TQH5-35 showing erosion of Otway Group strata. Also note the parallel to sub-parallel nature of the Otway Group strata parallel to the basement. Line length-10.3 km.*

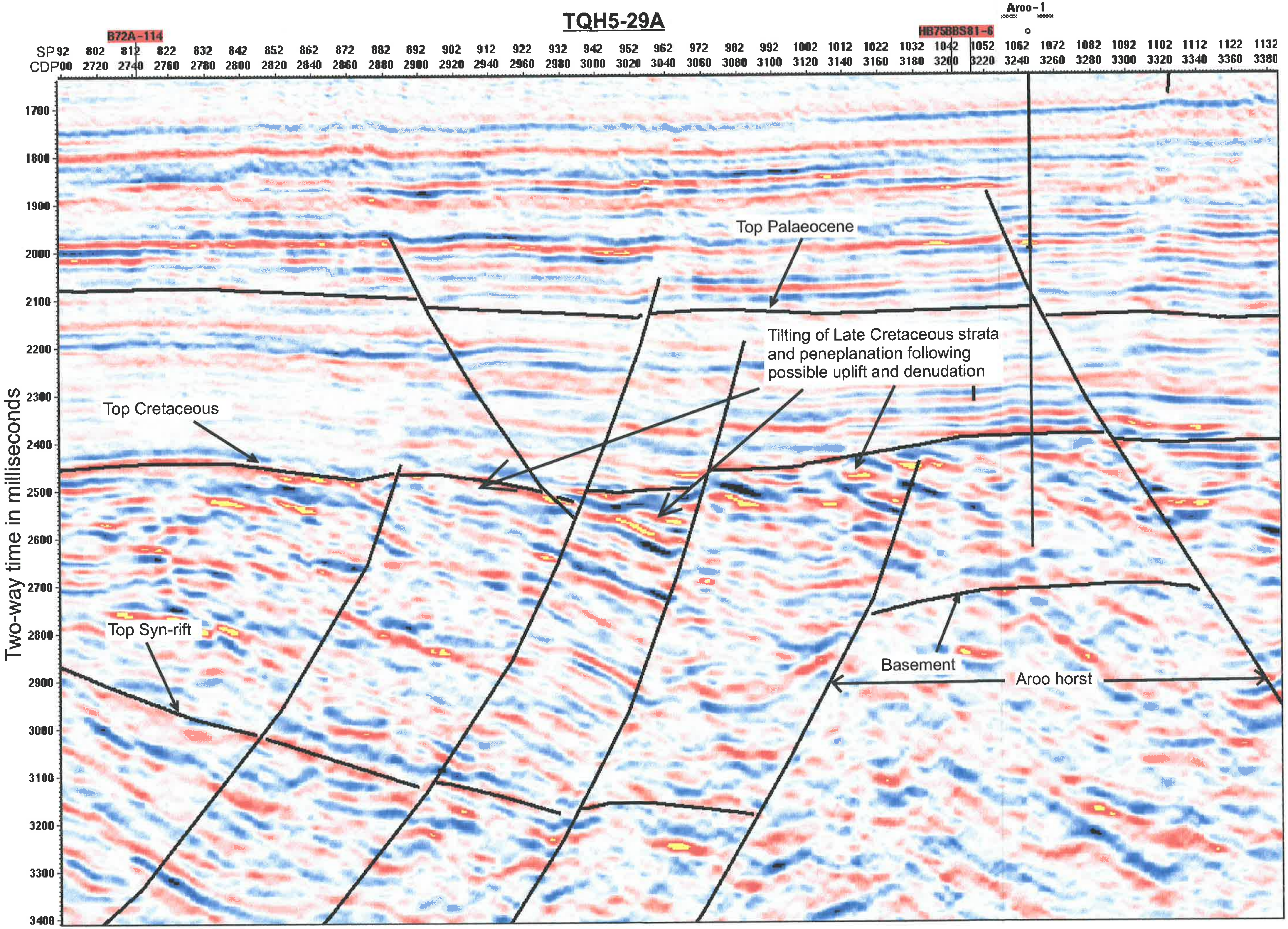


Figure 4.20 *Part of the interpreted seismic line TQH5-29A showing tilting of Late Cretaceous strata following possible uplift and denudation. Note absence of Otway or syn-rift strata on top of basement on the Aroo horst similar to that observed for Bass-3 horst. Line length-10.3 km.*

The dramatic increase in thickness of sediments deposited between the initiation of the Tasman rift (Otway Top unconformity) and Top Cretaceous unconformity as seen in Figure 4.18a within the Yolla Trough (between the Bass-3 and Aroo horsts) is ascribed to the effect of Tasman rifting as well as due to huge sediment load caused by rapid sedimentation in a fluvio-deltaic environment. There is no study in the literature detailing whether far field stresses generated by oceanic spreading can cause development of accommodation space along existing extensional normal faults in the adjacent rift valley. The Tasman Sea started opening during the Campanian (around 82 Ma) and it is observed that the sequence deposited in Cretaceous-Paleocene time (as seen in Figure 4.18a against the main half-graben bounding fault F2) shows evidence of increase in thickness. It is proposed here that the rapid sediment load in the depocentre during both Late Cretaceous and Paleocene periods caused reactivation of the normal fault F2, creating further accommodation space. This effect is more pronounced in the Bass area than the Durroon area (see Figure 4.4) probably because the Bass area in general was experiencing cooling at a faster and of greater magnitude compared to the Durroon area leading to crustal flexure and more ductile behaviour of the crust.

Another major unconformity is clearly seen in the seismic data with parts of tilted strata terminating as toplap (see Figure 4.18a) but there is no palynology data available anywhere in the entire Bass area to corroborate the age dating of this sequence boundary. To keep the chronostratigraphic event associated with this unconformity surface comparable with the Durroon area, it is dated as top of syn-rift sequence II. The wedging of the syn-rift sequences against the main bounding fault (Figure 4.18a) help in determination of the top of the syn-rift sequence unconformity. The Durroon area, however, was marked by reflection onlap surfaces rather than any toplap at this level.

There is also a marked difference in the reflection characteristics in the Otway sequence as well as the syn-rift sequence within the two half-graben like depocentres corresponding to the two troughs, the Yolla and Cormorant Troughs (compare Figure 4.18a and 4.18b). Neither of the sequences have any well control and interpretation is more subjective, based on reflection characteristics alone. On the western side of the Aroo horst, corresponding to the Yolla Trough, the Otway sequence has parallel to sub-parallel internal bedding geometry with erosion at the crestal portion attached with the hangingwall side (towards Bass-3 horst block) (see Figure 4.19) and the dip of the Otway strata is distinctly different to the dip of the syn-rift sequence strata, giving rise to a clear-cut angular relationship. The correlative Otway sequence in the Cormorant Trough on the eastern side of the Aroo horst, i.e., in the Cormorant Trough, however, has no indication of erosional truncation of Otway strata. The dips of the reflectors within the Otway sequence and syn-rift sequence do not have any appreciable difference between the Cormorant Trough and the Yolla Trough. The reason is not very clear. It is somewhat similar to observations of the attitude of reflections in the Boobyalla Block of the Durroon area.

There is indication of thickening of the strata (wedging) against the main bounding fault within the Otway and syn-rift sequences in the Cormorant Trough. Although the angular relationship between Otway sequence strata and syn-rift sequence strata is not discernible, there is however, some marked differences in the frequency content and reflector continuity between the two sequences (Figure 4.18b) within the Cormorant Trough. Setiawan (2000) has interpreted absence of any Otway Group in the Cormorant Trough while Williamson *et al.*, (1987) have interpreted presence of Otway Group sediments in the Cormorant Trough area. It is believed that although the sedimentation style and fault movement during sedimentation of the Otway Group might have been different in the two troughs, time equivalent stratigraphic intervals may be present. Actual well data, when available, would confirm these observations. It is difficult to jump-correlate seismic boundaries between two troughs, particularly for the Otway and syn-rift sequences. A degree of speculation still exists in the interpretation based on seismic facies characteristics in the absence of well data.

Unlike the Cormorant trough, the Yolla Trough shows indications of either deltaic or shoreline progradation (Figure 4.21). The delta appears to prograde to the east with the bottom set terminating against the bounding fault of the Yolla Trough (Fault F1 and F2, Fig.4.18). While the geometry could also be a result of onlap which later rotated, I calculated the "prograde" height to perhaps discriminate between the two possibilities.

Measuring the height of the apparent progradation in two-way time and converting it by utilising stacking velocity of 3000 m/sec, the actual height is about 200m-220m. Generally delta prograde height is about 100m in passive margin setting. It is therefore, believed that perhaps during the delta progradation in the Yolla Trough area against an actively extending fault, the apparent height is relatively more.

The difference in the reflection characteristics is indicative of the important role the Aroo horst played in controlling the sedimentary architecture within the two adjacent troughs. The initial extensional history in the two troughs has been different. The half-graben encompassing the Yolla Trough shows very large tilting of the Otway strata indicating rotational motion along the bounding fault with large displacement (2-3 km) creating areally large depression in which fluvio-deltaic sedimentation occurred. Sediment source was from the west. The major extension occurred in a very short time within the Tasman Sea rifting episode. The Otway Group does not show any major thickening against the bounding fault although there is a small increase in thickness downdip. Either the bounding fault existed during Otway period and was reactivated with large displacement or it was initiated only during the Tasman Sea rifting episode.

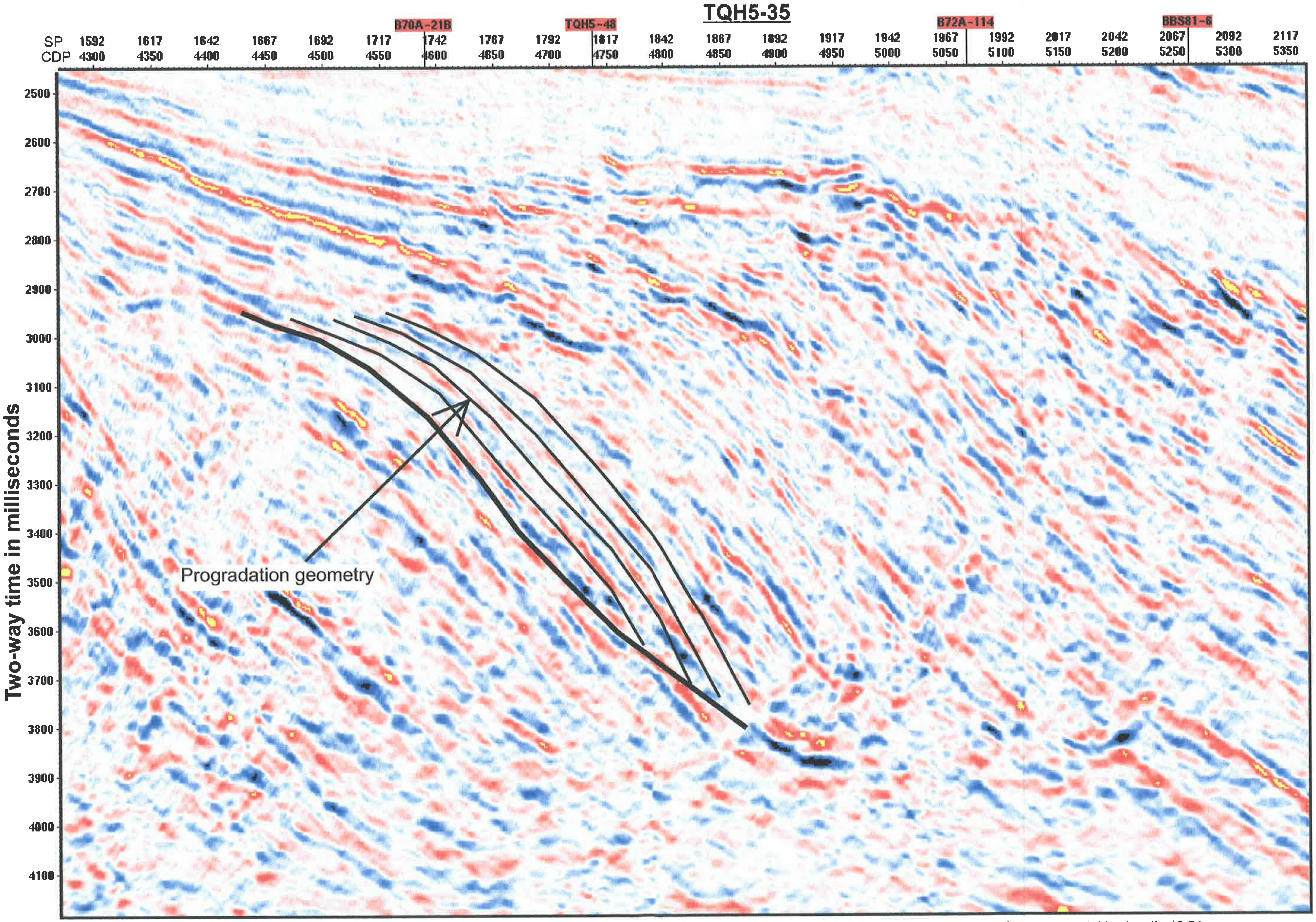


Figure 4.21 *Part of the seismic line TQH5-35 showing clear evidence of progradation geometry of the syn-rift sequence indicating either delta front build up or shore line movement. Line length- 16.5 km.*

With present understanding of the seismic data in the entire basin, it is believed that the major half-graben bounding fault to the west of the Aroo horst (Fault F2 in Figure 4.18a) and also Fault F1 (Figure 4.18a) already existed during the Otway rifting period and were further reactivated during Tasman rifting. In contrast, the sedimentation and structuring in the Cormorant Trough was different. The Otway Group shows evidence of thickening against the bounding fault to the east of the Aroo horst (Fault F3 in Figure 4.18b) which means that the fault was active during that time also. The syn-rift sequence deposited during the Tasman Sea rifting episode shows rapid thickening against the bounding fault indicating that a large amount of extension was taken up by this fault. The different seismic facies observed within the two troughs suggest that sediment source/composition and sediment transport direction might be different in the two depocentres apart from the fact the timing or style of fault movement has also been distinctly different. It is however, suggested from the reflection geometry in the two troughs, that rapid sedimentation into the troughs created during initial extension with sediment load adding significantly to subsidence, creating accommodation along with synsedimentary fault growth. That means the predominant forces for accommodation space development in the Bass area were sediment load and syn-sedimentary fault growth unlike in the Durroon area where passive infill of the holes created by large fault displacement in the tilted basement fault blocks built the sedimentary architecture in that area.

4.2.3 Structural interpretation of seismic data

The composite line as presented in Figure 4.17 shows the typical structural style in the Bass area. As is seen, the structural style is dominated by horst and graben style with two conspicuously dominant horsts (Bass-3 and Aroo horsts) emergent during the main rifting history of the basin. Although it is observed that the western half-graben (Yolla Trough) shows huge rotation of the Otway strata with tilting and syn-rift sedimentation occurring in the evolving depocentre during Tasman rifting. The eastern side, corresponding to the Cormorant Trough, shows progressive tilting during both the Otway and Tasman rifting episodes. The resultant basement topography in the Bass area is however, horst–graben style in contrast to the Durroon area where tilted domino basement fault block style is observed (discussed in 4.1.5).

The major fault (Fault F0), close to the western side of the Bass-3 horst, did not record any movement during the Otway and Tasman rifting episodes except perhaps towards end of the Late Cretaceous period creating some accommodation space for sedimentation. This means the movement along the Fault F0 is not age equivalent to the movement along the two other major extensional normal faults, Fault F2 and F3, which control sedimentation in the Yolla and Cormorant Troughs. This is in contrast to what happened in the Durroon area where probably all the major block bounding faults moved in unison. As is clearly seen from Figure 4.17, the faulting style during the early post-rift history and later thermal cooling stage is marked by a large number of faults with relatively less displacement caused by syn-sedimentary fault growth and due to effect of burial and compaction. The major block bounding faults F2 and F3 are seen to have reactivated quite late, i.e., up to Late Eocene. This reactivation is again interpreted to be due to sediment loading and/or compaction. The fact that the two troughs are separated by the Aroo horst, suggests the nature of sediments in the corresponding depocentres may have different lithofacies in the equivalent time interval. The rift onset unconformity is the Otway top unconformity and corresponds to the beginning of the Tasman rifting episode, similar to the Durroon area. The fact that Smith (1986) interpreted the Durroon Formation to be restricted to Durroon area only makes some sense only when is considered in terms of the sediment character rather than the age equivalence. It is believed that Tasman rifting caused substantial sedimentation in these two troughs but exact age correlation can only be established when deeper drilling information is available.

From the description of the seismic stratigraphy, it is safe to conclude that the hydrocarbon play types in the various depocentres in the Bass Basin would be controlled by separate petroleum systems because the source-rich sediments in the deeper water lacustrine deposits would be distinct from each other. However, this can be corroborated only with further study and new drill data.

By Late Eocene time, the basin had moved into the passive subsidence phase with the reflectors showing regional extent and covering the entire breadth of the basin. This is similar to what happened in the Durroon area.

In terms of the fault pattern in map view, it is observed that the Burnie Sub-basin behaved in a somewhat different manner to that of the Durroon Sub-basin in the initial rifting history. There were very few extensional normal faults in the entire area during the Otway rifting period (Late Jurassic-Early Cretaceous) only in the central part of the Burnie Sub-basin (in the Yolla and Cormorant Trough areas) and possibly in the Pelican Trough area (Figure 4.22). The fault trend in general remained WNW-ESE to E-W in the Yolla and Cormorant Trough areas and N-S in the Pelican Trough area. The beginning of Tasman rifting during Early Cenomanian in the second stage saw reactivation of some of the existing faults with large displacement, particularly the Faults F1-F3 as already discussed above (Figure 4.23). The existing faults propagated along their tips both vertically and in plan. The third stage saw the progressive propagation and coalescing of the faults further more with some new faults being created like Fault F0, and the trend of the faults associated with this event remained almost NNW-SSE to N-S (Figure 4.24). Interestingly, the faults created during the third stage encompass a much larger area of the sub-basin, particularly in the eastern parts, and it is suspected that these faults were all activated during the major tectonic event related to the initiation of oceanic spreading in the Tasman Sea. This is interpreted from the fact that all the seismic sections in the eastern part of the area show no growth of sedimentary thickness across the individual faults related to the previous two events. The clear-cut evidence of the Maastrichtian unconformity in this area from seismic data as already discussed suggests that perhaps a fourth stage of extension affected the basin. This is consistent with what has been observed by Laird (1994) in the New Zealand region. The fault trends remain the same but the most active segments of the faults are those parts perpendicular to the new direction of extension.

There seems to be a major inconsistency in the fault pattern maps prepared by Setiawan (2000) wherein he has proposed all the major faults in the Burnie Sub-basin to be dipping to the southeast and all the major faults in the Durroon Sub-basin to be dipping to the northeast (from which he has inferred a polarity reversal across the Pelican Squid transfer zone). But during this study, such a clear cut distinction in the dip of the fault planes has not been observed, particularly when the northeast dipping fault F3 just east of Aroo horst fault is seen to be a very large tectonic element in this part of the basin and has been active right from the period of Otway rifting up to Paleocene age. Apart from this major fault, there are a large number of faults with small throws that exist in the Burnie Sub-basin that dip to northeast which however, were active at a relatively later stage in the basin history.

The quality of the data in the Burnie Sub-basin becomes exceedingly poor in the Pelican Trough region (see Figure 4.25 and compare with Figure 4.17) and was the reason why detailed palaeogeographic map construction in the Burnie Sub-basin was not thought to be appropriate. To get a comprehensive picture of the sediment distribution patterns corresponding to each individual sequence interval, future work should begin with enhancement of data imaging in the main Pelican Trough area and the Pelican-Squid transfer zone area and then attempt to determine a realistic interpretation of the paleogeographic reconstruction.

Figure 4.22 *Fault pattern map of the Bass area at the first stage of rifting (Otway rifting: Late Jurassic-Early Cretaceous).*

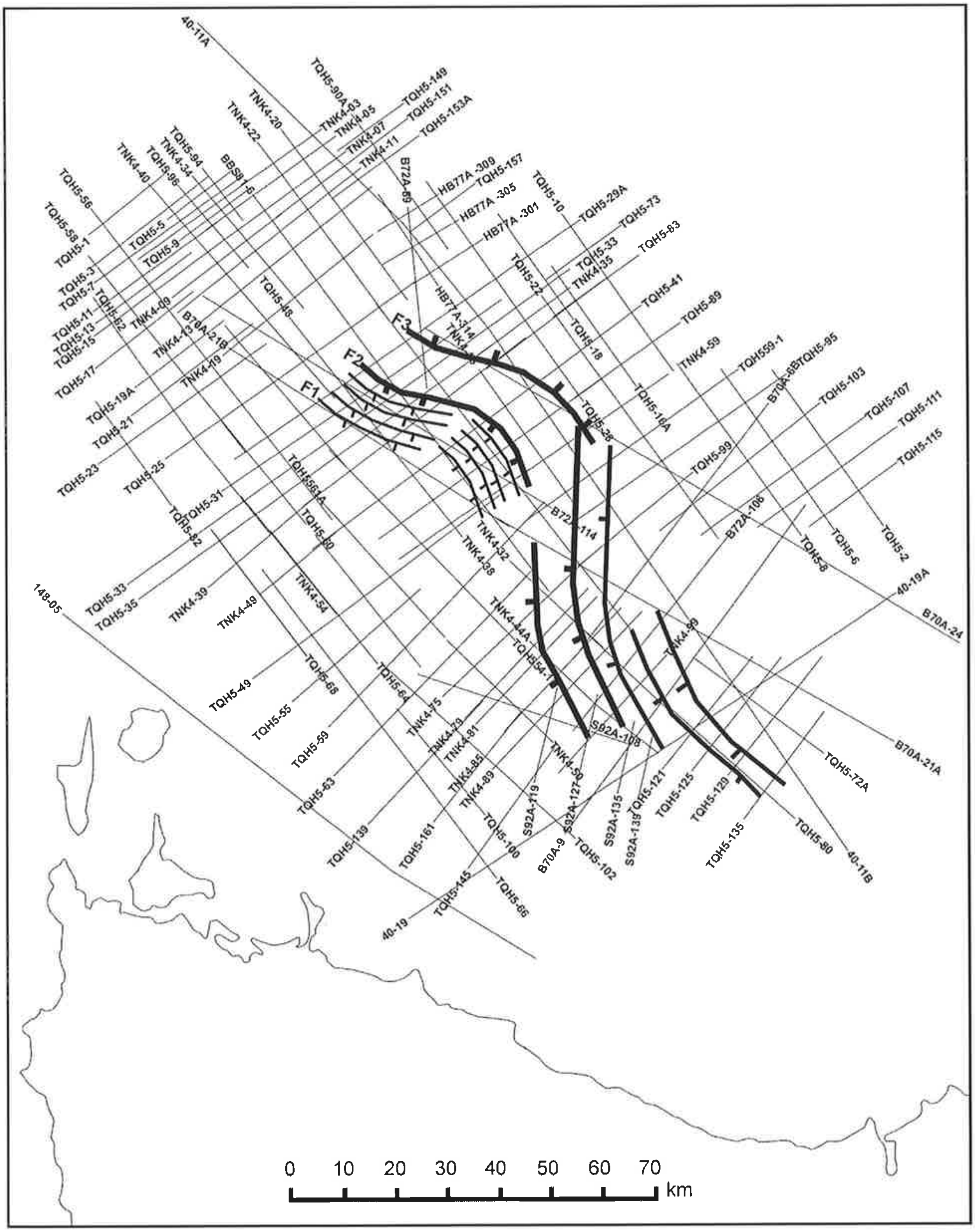


Figure 4.23 *Fault pattern map at the second stage of rifting (Tasman rifting: Early Cenomanian). Some of the earlier existing faults were reactivated with large displacement and progressed along tips to coalesce. There were few new faults both in the Yolla and Pelican Trough areas.*

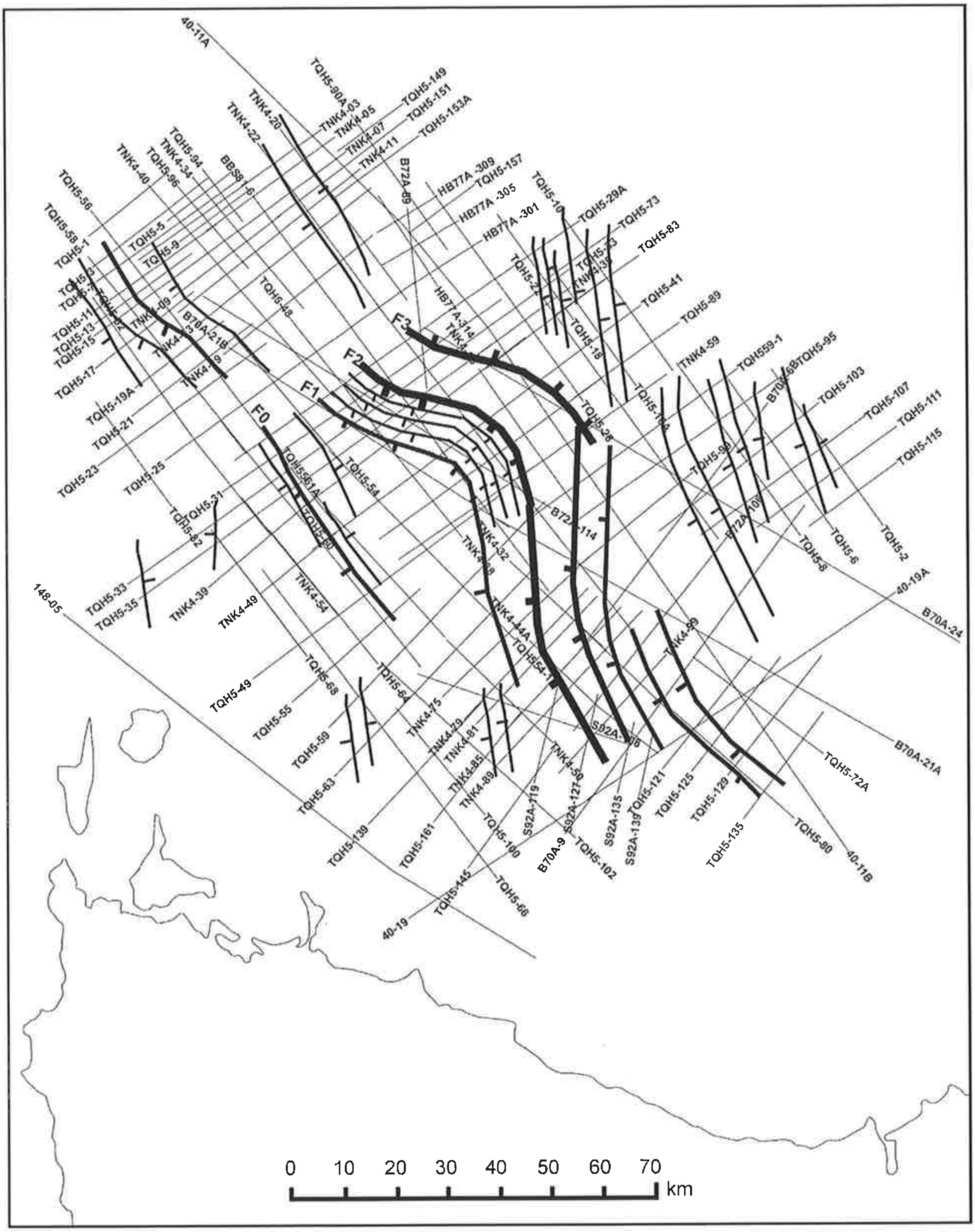


Figure 4.24 *Fault pattern map at the third stage of extension (Early Campanian). Progressive propagation and coalescing of existing faults with development of large number of new faults all across the area. Fault F0 recorded movement during this period.*

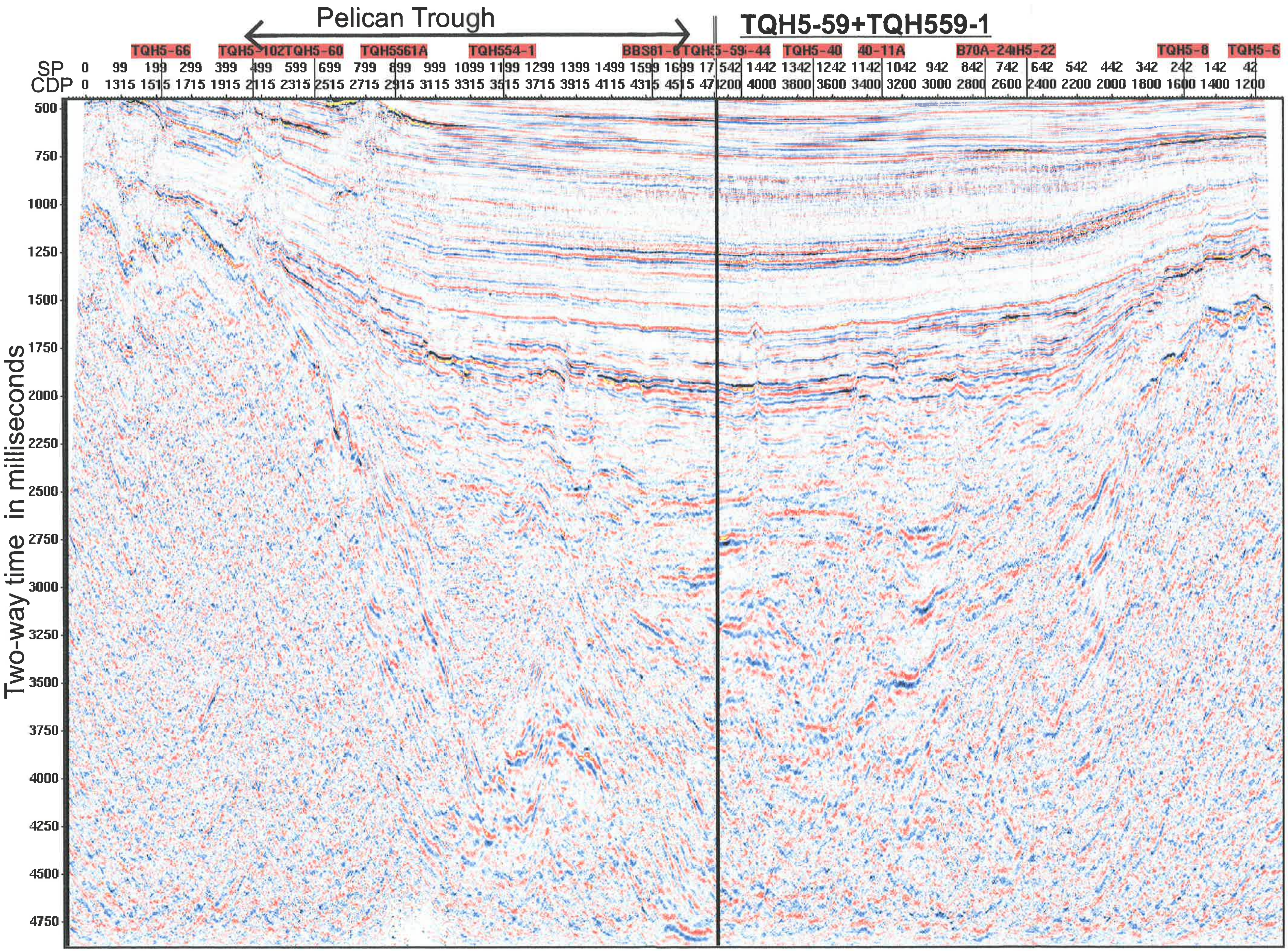


Figure 4.25 *Uninterpreted composite seismic line TQH5-59+TQH559-1 to show the deterioration of seismic data quality beyond 2.5 seconds particularly in the Pelican Trough area (Compare with data of Figure 4.17). Horizon correlation in the rift related sequences is highly unreliable. Line length-103.2 km.*

4.2.4 Compressional reactivation

Figure 4.26 shows a classical inverted structure (Cormorant structure) observed in the Bass area along the north-south oriented seismic line B72A-89. There are few important observations that can be made from this figure. The large anticlinal Cormorant structure was formed during Late Oligocene-Early Miocene period. Till the Late Eocene beginning from Upper Cretaceous, it can be easily seen how the sedimentary thickness increases towards Cormorant area, from south to north. It was during Late Oligocene-Early Miocene period, when the Australian Plate collided with the Indonesian archipelago to its north, the resulting compressional stresses propagated across the continent to have a profound effect in the northern part of the Bass Basin inverting the pre-Oligocene Cormorant Trough into its present day anticlinal structure. There were other continental collisions and realignments at this time, which were also close and could have a great influence on this area. E-W compression has been observed in New Zealand, the Gippsland Basin, the Eastern Otway Basin and the St Vincent Gulf Basin around Adelaide. It could be that near E-W compression caused some wrenching along the southern edges of Australia to build this structure (Young *et al.*, 1991; Davidson, 1980). The effect of this inversion is also seen in the faults which acted as extensional normal faults during deposition of Cretaceous-Late Eocene sediments but show a reverse sense of throw towards the upper tips clearly (Fault 'Fr' in Figure 4.22). There is clear evidence of thinning of strata and possible erosion at Late-Oligocene–Miocene level from the reflection seismic data.

Another important structural implication that can be noted from Figure 4.22 is that during the period from the beginning of sedimentation during Otway till Cretaceous time, the depositional dip was from north to south along this line towards the Aroo horst. During the Cretaceous, the sense of depositional dip changed towards the opposite direction along this line, which then became from south to north with possible basinal tilt or uplift. This is consistent with what was described in the section 4.2.2 regarding the Maastrichtian unconformity showing evidence of peneplanation following uplift and denudation in the basinal depocentre. It means that the Cretaceous-Tertiary boundary is a very prominent tectonic event in the structural history of the basin.

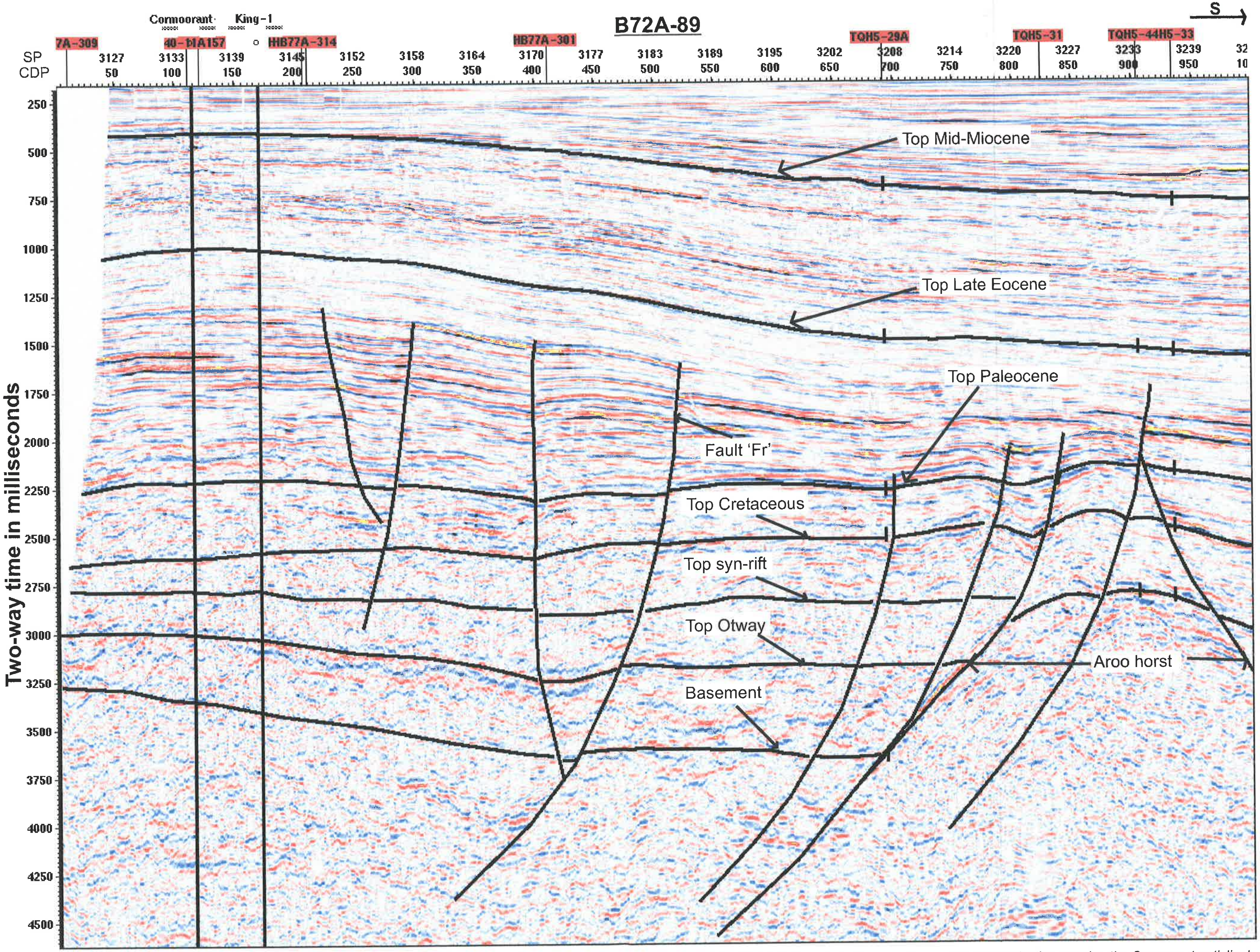


Figure 4.26 *Interpreted seismic line B72A-89 showing the Cormorant inverted structure. Note the thickening of the Cretaceous-Late Eocene section and later reactivation causing the Cormorant anticlinal structure to form. The Cretaceous-Tertiary boundary is also a prominent tectonic event in changing the depositional dip from south to north along this line. Line length-33 km.*

4.3 Preliminary measurement of extension

A preliminary estimate of the amount of extension by measuring the pre-faulted basement length and the post-faulted basement length estimates extension of 1.36 in the Durroon area compared to 1.22 in the Bass area.

The seismic line BMR88-306 was considered for the Durroon area and the composite seismic line (TQH5-33+16A+73) was considered for the Bass area. The total initial basement length was measured by taking each individual fault cut off position and physically measuring the length of each segment and then adding them together. For the BMR88-306 line, the initial length was 51.859km and final length (present day position) is 70.513km. The stretching factor was thus determined by subtracting the initial length from final length and dividing the result by initial length, which gave a factor of 0.36. Similarly for the composite line (TQH5-33+16A+73), the initial length was 96.153km and the final length was 117.460km and the stretching factor was calculated to be 0.22.

Although most of the previous workers in the basin have concluded that the amount of extension amount in the Durroon area is less than the Bass area (Etheridge *et al.*, 1985; Setiawan, 2000-see section 2.3.3), the results of the present study do not support that view. However, the exact amount of extension can be worked out by considering the effect of compaction and sediment loading by the method of 'backstripping' each layer of sediments onward from the time when the first rift sedimentation started. In a recent work in progress, Cummings *et al.* (2001) have considered all the above factors while attempting to measure the amount of extension along two regional seismic lines, one across the Bass area and another across the Durroon area. The preliminary results of their work suggest an extension amount 1.12 for the Bass area and 1.26 for the Durroon area. It was expected that the reconstruction of seismic section by depth conversion, decompaction and backstripping each layer would give proportionately a lesser absolute value for extension amount because of the additional effect of sediment load on the fault displacement as observed at the present day.

The very large amount of extension (60% – 80 %, as mentioned in section 2.4) in the Bass Basin suggested by Etheridge *et al.*, (1985) appears to be an unreliable estimate. The earlier workers perhaps wrongly interpreted the large post-rift thickness of sediments in the Bass area to be result of extension thus giving relatively higher value for extension amount for the Bass area compared to the Durroon area.

It is proposed that the additional subsidence in the Bass area is achieved by sediment loading and isostatic subsidence of a crust that is weaker than in the Durroon area.

CHAPTER 5 LAYER-BOUND POLYGONAL FAULT SYSTEM

5.1 Introduction

The existence of polygonal fault systems was first documented in the Lower Tertiary mudrocks from the North Sea Basin (Cartwright, 1994a, b). The fault system there is characterised by densely packed, layer-bound, minor extensional faults exhibiting a unique polygonal map geometry in plan view. Vertically, the polygonal fault system is organised in tiers of stratigraphically-bound layers of faults that have a distinctive structural style. Separate tiers have distinct fault spacing, orientations and fault trace shapes (Cartwright, 1996; Lonergan *et al.*, 1998). A similar polygonal network of faulting has been observed in the Cretaceous siliciclastic sequences of the Eromanga Basin of Australia (Cartwright and Lonergan, 1997; Watterson *et al.*, 2000).

Cartwright and Dewhurst (1998) published a paper midway through the current project in which they have mentioned the possibility of a layer-bound compaction fault system in the Bass Basin based on seismic observations from earlier published literature. However, there has been no other mention in the literature of Bass Basin about this newly recognised polygonal fault system. The initial interpretation work of this project observed that the shallow section corresponding to the Torquay Group is highly fractured, with faults having very small throws and restricted to one particular layer in the stratigraphy. This initial observation led to the detailed study of the polygonal fault system in 3D seismic data as it was obvious from the nature of the fault system that 2D seismic data is not sufficient to map the fault features (Das *et al*, 2001).

A detailed literature survey was made on this fault style and it was felt very important to carry out some detailed investigation into this fault system of the basin. The Yolla 3D seismic dataset was utilised for this purpose. While the 2D seismic data is not sufficient to adequately map a fault set of this nature, it does indicate its widespread existence throughout much of the Bass Basin. Though there is still a hot debate as regards the genesis of such fault systems, a synoptic discussion on the genesis is presented here with reference to the published literature. Though no direct study has been made regarding the hydrocarbon implications of the polygonal fault system in the basin, the implications for reservoir geometry and seal capacity in petroleum exploration are discussed with reference to published literature.

5.2 Stratigraphic setting of the polygonal fault system

The polygonal fault systems as seen in the Bass Basin affect the middle Oligocene to late Miocene section of the Late Tertiary basal Torquay Group. The lithology of the deformed units is dominated by calcareous clay and marl deposited in slope and basin plain depositional environments (Williamson *et al.*, 1987) in a barred marine basin setting. The region over which the polygonal fault system is developed is illustrated in Figure 5.1.

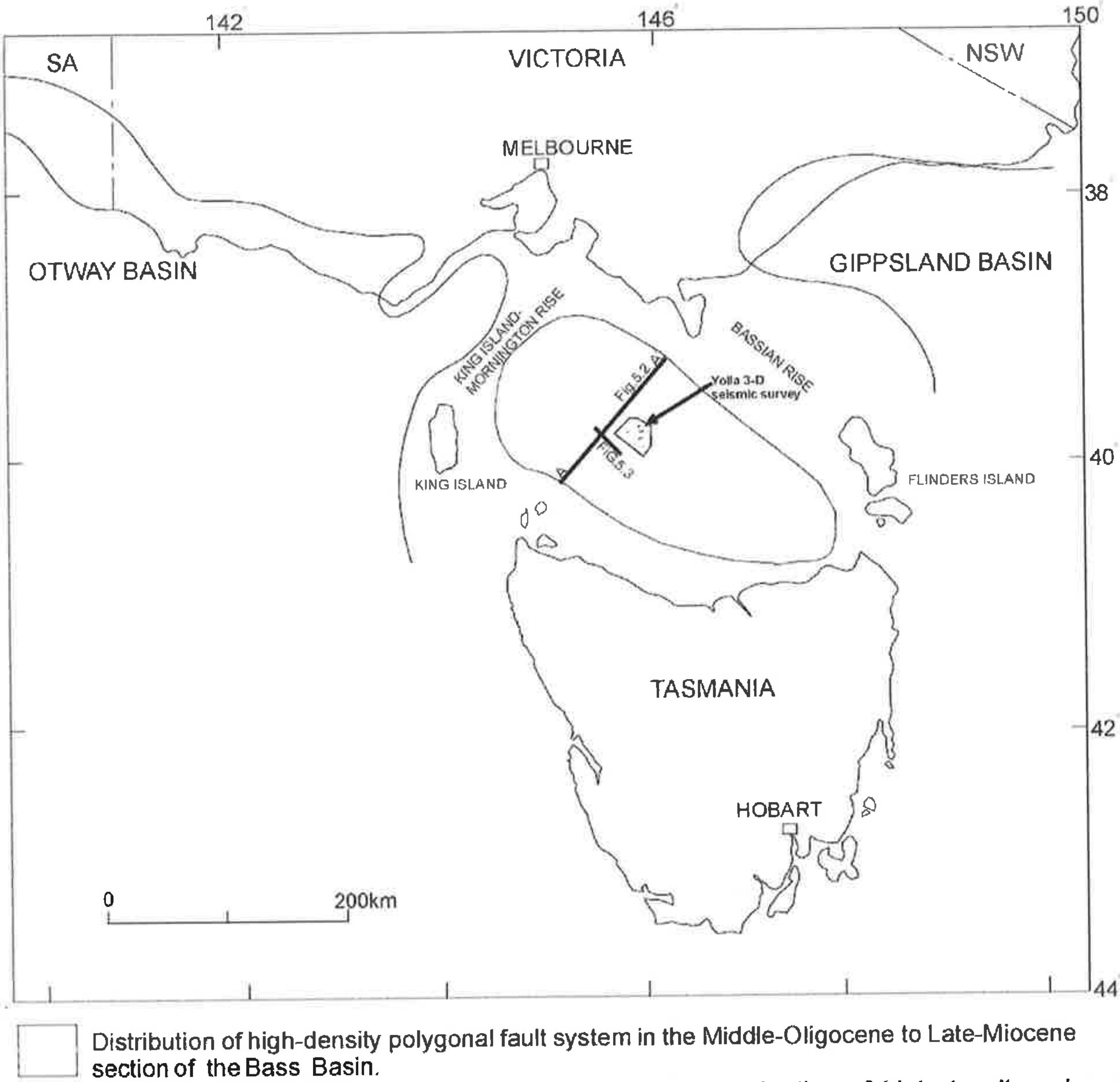


Figure 5.1 Regional map of the Bass Basin showing the distribution of high-density polygonal fault system over almost the entire offshore Bass Basin. The location area of the Yolla-3D seismic survey is also shown along with the locations of figures 5.2 and 5.3.

The Bass Basin was a large post-rift continental sag basin by the Late Eocene flanked by significant regional uplift on at least three margins: the King Island - Mornington rise in the west-NW and the Flinders Island - Bassian rise to the east-NE encompassing the basin in an elliptical fashion (section 2.5). The representative regional geo-seismic cross section across the central portion of the basin presented in Figure 5.2 shows that by Late Eocene, the basin went into passive thermal stage subsidence.

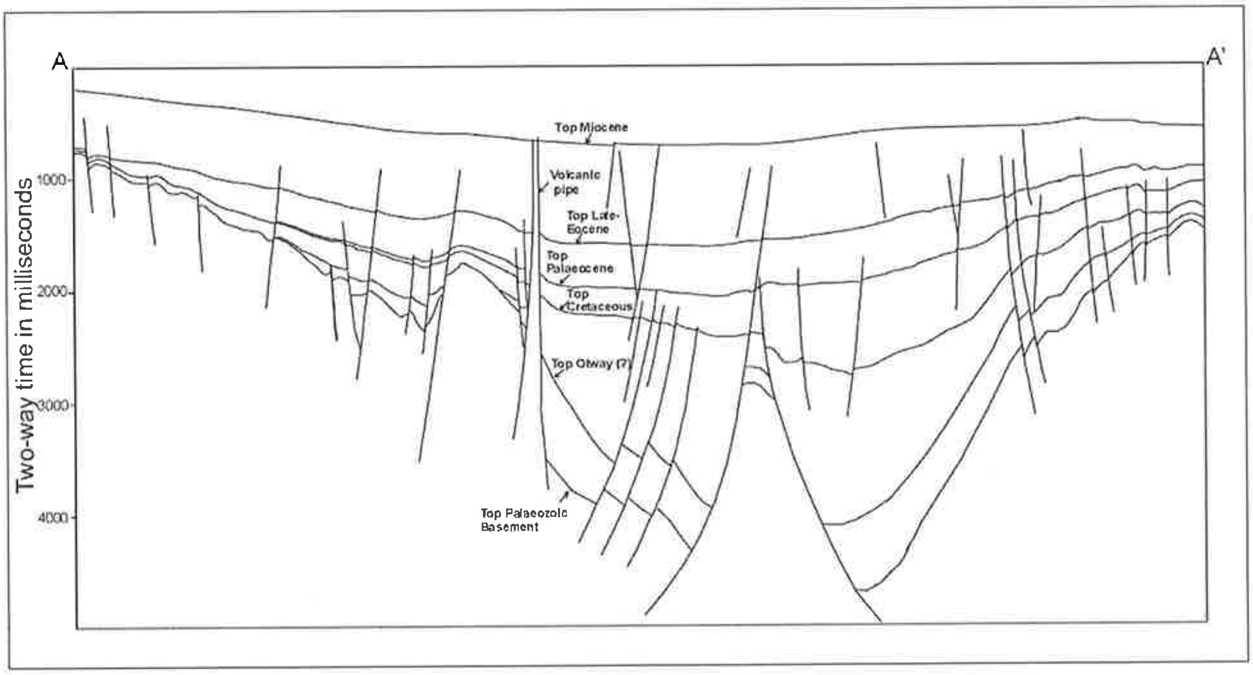


Figure 5.2. Regional geo-seismic cross-section across the central portion of the Bass Basin (for location see figure 5.1) showing the major sequence boundaries and their two-way time depths. Only the major tectonic faults are shown not the polygonal faulting which has affected the Middle Oligocene - Late Miocene succession.

The late Tertiary basin fill is characterised by very fine-grained depositional systems. The provenance for the sedimentation was from the east, northeast and south.

The composition of the sediments near Yolla-1 is Volcanoclastic (Wheeler and Kjellgren, 1986) and it suggests that the fine-grained clay and mudstone-dominated sediments were smectite-rich. Though most of the sedimentological/petrological studies to date in the basin are confined to the hydrocarbon prospective Late Cretaceous-Late Eocene Eastern View Coal Measures, there has been practically no detailed sedimentological or petrological study in the basin of the Torquay Group. Structural studies in the Bass Basin so far have mostly focussed on basin-forming mechanisms and hence on broad tectonic features like rift-related basement structures (Etheridge *et al.*, 1985; Williamson *et al.*, 1987; Das and Lemon, 2000a). As was mentioned earlier, the small-scale extensional faulting confined to only a particular stratigraphic interval in the Bass Basin has received almost no mention in the literature.

The regional section (Figure 5.2) shows that there have been no major tectonic movements in the basin since the Late Eocene except for minor reactivation along the pre-existing normal extensional faults. The basin has experienced a major Early Miocene compressional event producing inversion structures viz. Cormorant (Williamson *et al.*, 1987). The basin has largely undergone passive subsidence since the Late Eocene as a response to thermal relaxation after the initial rifting events formed the basin.

5. 3 Expression of the layer bound fault system in the 2-D seismic data

Figure 5.3 shows a 2-D seismic line from the central Bass Basin illustrating the key features of the seismic expression of the Late Tertiary succession in the basin. The middle Oligocene to Late Miocene section of the Late Tertiary has been pervasively deformed by minor extensional faulting characterised by fault spacing ranging between 60m-500m and with fault throws between 10m-40m. The deformation is clearly limited to a particular stratigraphic interval with the overlying and underlying intervals showing highly continuous reflections.

The pervasive discontinuous reflection character may easily be misinterpreted as a depositional response to a high-energy environment producing chaotic seismic facies, sometimes with mounded reflection geometry. Under close scrutiny, an alternative interpretation of the small-scale reflection terminations can be made incorporating high density but small scale extensional faulting affecting the particular layer. The wide spacing of 2-D seismic lines may not be suitable to fully resolve the small-scale extensional faulting as many of the faults may be oblique to the line orientation and may not be imaged properly. The very close spacing between the individual faults may mask the reflection image of these small-scale fault structures by spatial aliasing problems.

There are three units A, B and C, distinguished by amplitude-frequency characteristics of the reflection packages. Units A and C are of high amplitude and high frequency content whereas the middle unit B is characterised by low amplitude, low frequency reflections. The deformation style of all the units may be similar but the loss of frequency in the middle unit B may be the reason for lack of clear definition of the style of faulting. The lower unit seems to have been intensely faulted at central and right hand side of the section with the reflection patterns exhibiting continuous reflections on the left of the section. This may well be related to the variation in the grain size distribution and the mineralogical composition of the corresponding lithology that directly correlates with existence or non-existence of polygonal faulting. Dewhurst *et al.* (1999c) have established a direct correlation between high intensity polygonal faulting and the ultrafine grain-size of the clay particles and a relatively high smectite content. The higher the smectite content and percentage of ultrafine grain-size particles in the rock mass, the greater is the intensity of polygonal faulting and vice versa. The features observed in this seismic section are also reflected in the 3-D seismic data (next section). Possible causes are discussed in detail later in this chapter.

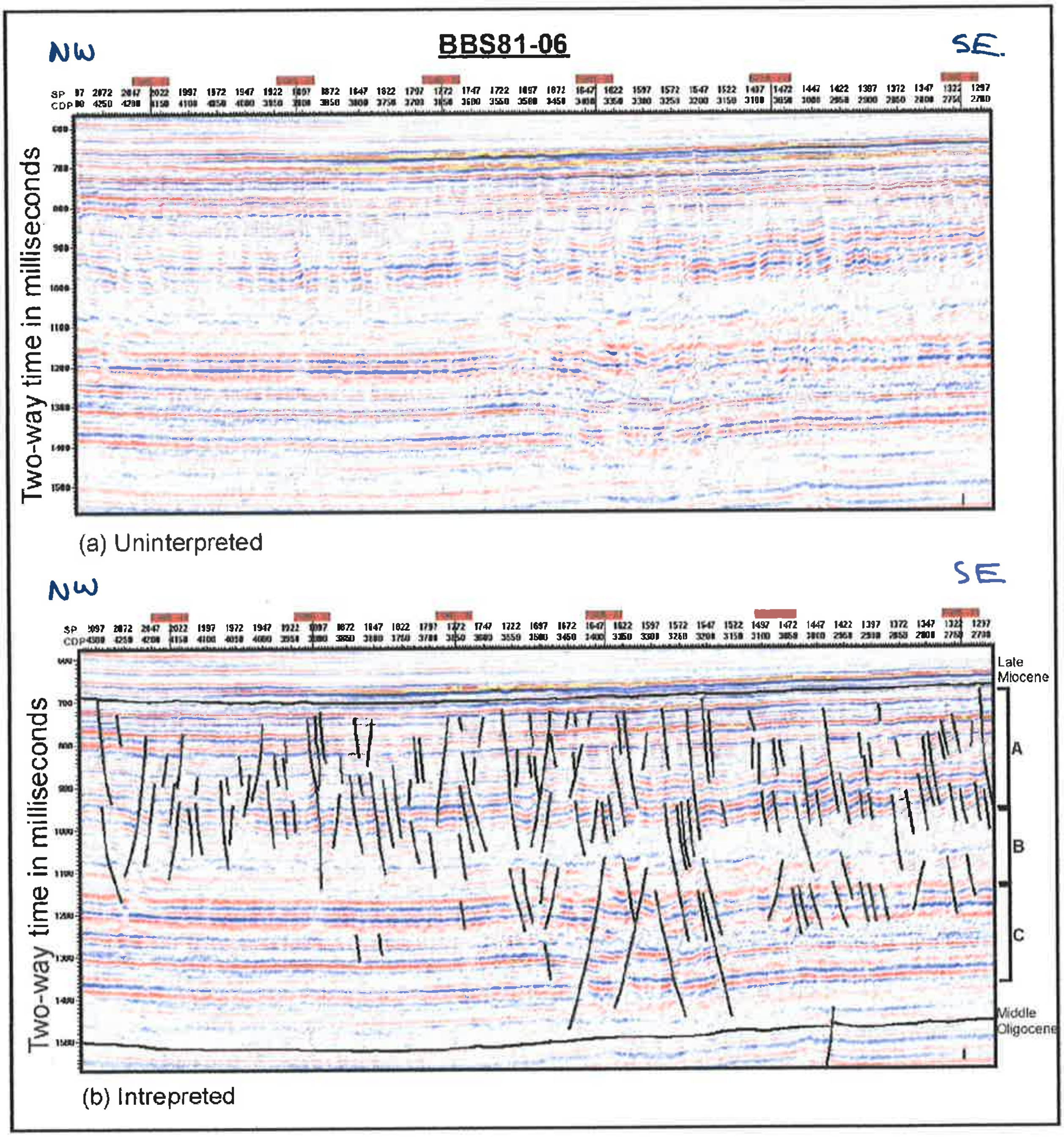


Figure 5.3 2-D seismic expression of the Middle Oligocene - Late Miocene interval showing highly discontinuous pattern of reflections, bounded above and below by extremely continuous reflection intervals. The seismic line is from the central Bass Basin (for location see figure 5.1).

5.4 Analysis of the polygonal fault system in the Yolla 3-D seismic data

The Yolla 3-D seismic survey was designed to cover specifically the area around the first major oil/gas discovery well in the basin, Yolla-1, in order to define the structural closure of the Yolla Field in greater detail. It covers an area of approximately 260 km^2. The survey was conducted in 1994 with a 12.5m $\times$ 25m bin size and comprises 30-fold data (Lennon *et al.*, 1999). For the present study, an area equivalent to 10km $\times$ 8km has been considered for horizon mapping

(Figure 5.4). The area is in the central part of the basin where the total stratigraphic thickness is near to the maximum in the basin.

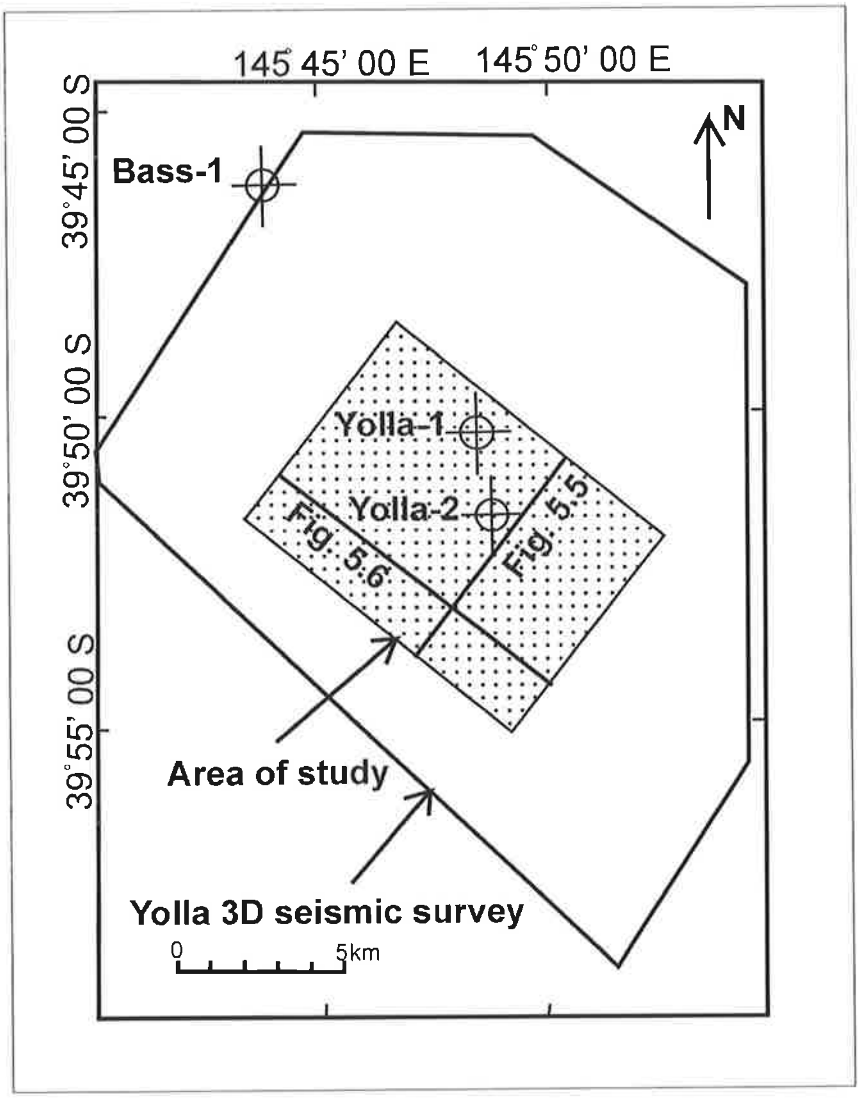


Figure 5.4 Location map of the Yolla-3D survey and the area considered for mapping of the polygonal fault system in the Bass Basin in the present study is indicated.

Though the high-resolution 2-D seismic data show the type of faulting in the deformed interval in cross section, it is almost impossible to map the fault pattern consistently from conventional 2-D seismic data. However, the 3-D seismic data show the faults to be much more clearly resolved and interpretation is much easier with less chance of spatial aliasing. Figure 5.5 is an inline seismic section showing the 7 horizons that have been mapped for the present study. The horizons 2-6 represent the polygonally-faulted horizons at successively shallower levels whereas horizons 1 and 7 represent undeformed layers, one underlying and the other overlying the deformed interval. The faults displacing the horizons in a normal extensional style are clearly seen. The crossline presented in the Figure 5.6 also clearly shows the structural style of the polygonal fault system. The faults are all extensional with small throws 10-40ms (approximately equivalent to 10-40m). Many of the faults dip in one direction, giving the appearance of rotated domino fault blocks. Some faults dip in the opposite direction and, when combined, give an impression of mounded reflection geometry. Individual faults are generally restricted in vertical extent to only a part of the entire deformed interval while some faults crosscut most of the interval.

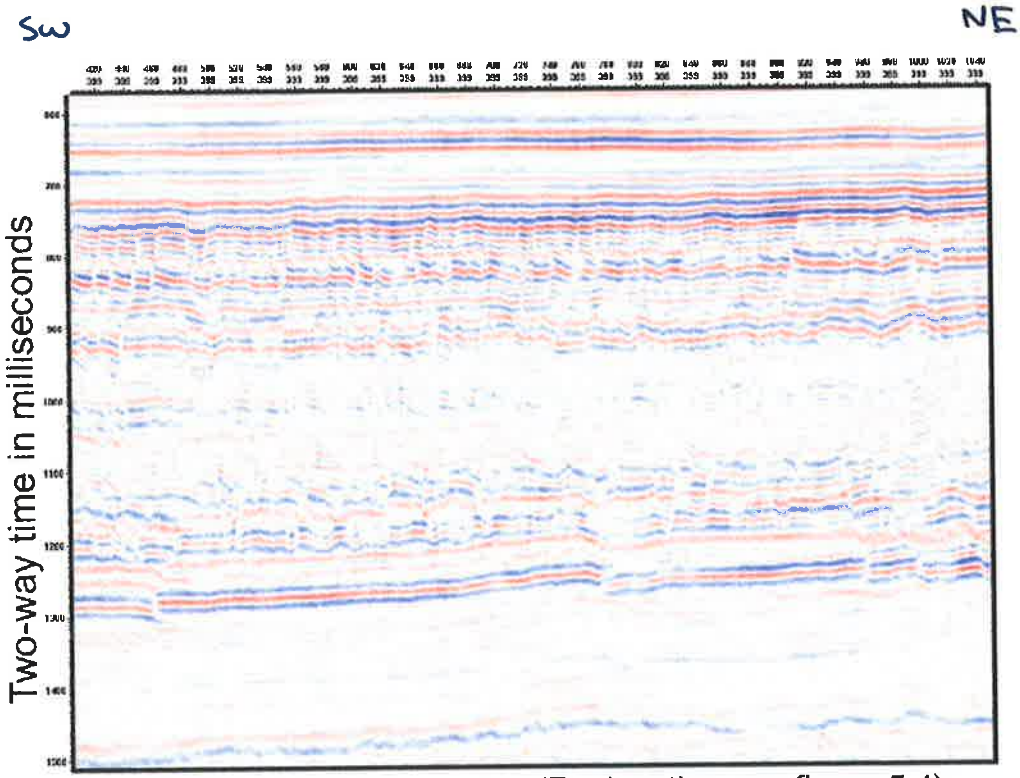
The pattern of faulting is seen to vary from one stratigraphic level to the next, but the extent of mechanical coupling between different sets of faults is not clear, since changes often occur within the low reflectance intervals. This packaging of sets of faults within stratigraphic units represents the tier structure (Cartwright, 1994a, 1994b, 1996). A similar structural style is also seen in the orthogonal lines (Figure 5.6). It is obvious from the two representative 3-D seismic sections shown here that each of the faulted units (A, B & C) has different faults affecting the layers in characteristic fashion though some of the faults crosscut from one unit to another.

Although it is much easier to interpret faults on the 3-D data than the 2-D data, the structure is so complex that it is difficult to make consistent fault interpretations on a section-by-section basis. Two approaches to mapping complex networks of small faults are possible using 3-D seismic data (Cartwright and Lonergan, 1996). A simple approach is to take time slice images and identify faults from offsets in the strike of bedding. Faults interpreted as time slice lineaments can be calibrated with the cross-sectional data to verify the interpretation. The second approach is to conduct a full manual horizon interpretation and to map closely-spaced horizons and to use these to analyse the 3-D structure in the 3-D seismic dataset. In the present study, 400 inlines were used for horizon picking using an 'autotrack' correlation technique while interpreting each line one by one. Figures 5.7-5.11 shows the horizon maps for the five successively deeper but polygonally-faulted horizons correlated within the Early Miocene deformed interval. The polygonally-organised faults are clearly recognisable as discontinuities in the two-way travel time values (colour gradations) to the horizon. There are some solid blacks on the contour maps along the fault traces which is a measure of fault heave. Irregularities in the fault traces and width are a reflection of the resolution accuracy in interpreting fault plane positions. As is seen in each of these maps, the fault pattern contains many triple and quadruple fault intersections typical of polygonal fracture networks (Lonergan *et al.*, 1998). To give a total picture for the area, two horizon maps, one for a horizon immediately overlying and the other immediately underlying the deformed interval are also presented (Figures 5.12 & 5.13). Some of the tectonic fault pattern is obvious from Figure 5.13 and this pattern penetrates into the lowermost polygonally-faulted horizon (Figure 5.11) which bears a strong resemblance with Figure 5.13 in so far as the tectonic fault pattern is concerned. It is seen clearly that the undeformed layer (horizon 7) represented by Figure 5.12 shows that absolutely no faulting activity has affected this layer.

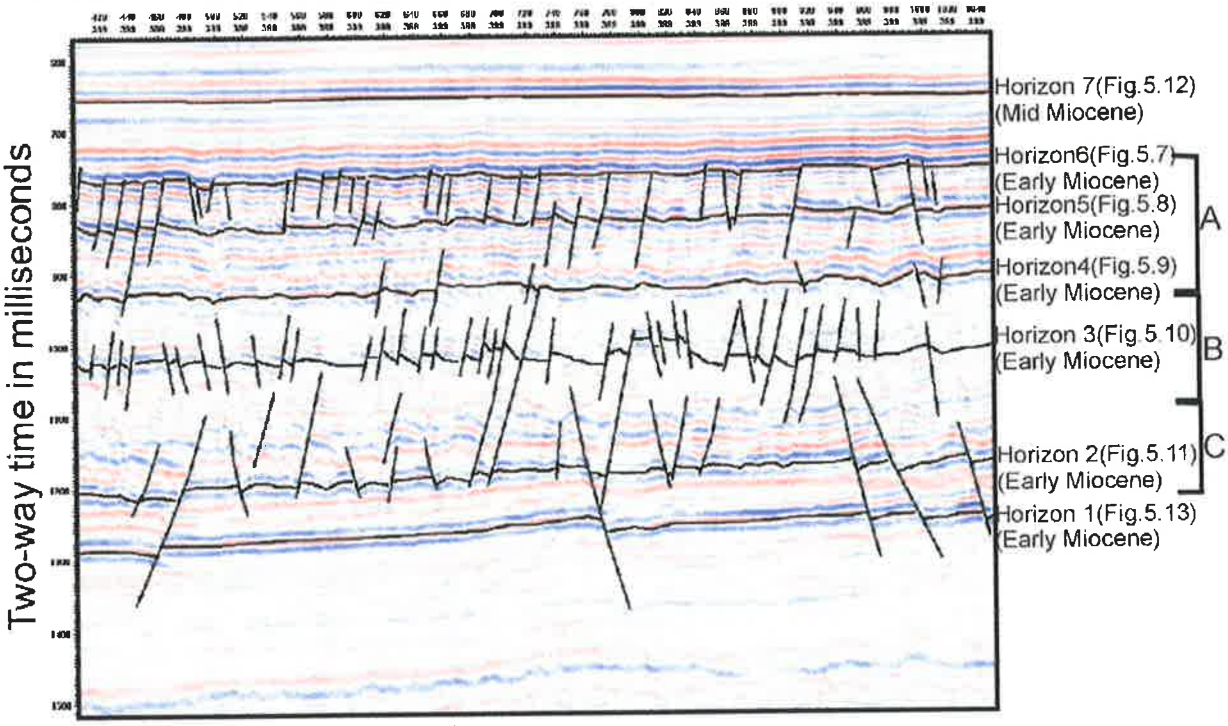
The 3-D seismic data in the Yolla area (Figures 5.5 and 5.6) show Unit A and Unit C have a marked high reflectivity. Unit A has a much higher frequency content than Unit C. Unit B in the middle suffers from poor reflectivity which directly indicates that the acoustic contrast necessary for good reflections does not exist. The two-way time structure map for the horizon 3 within the middle unit B (Figure 5.10) shows that the auto-tracking method of horizon picking in this case has been very less efficient because of the very poor reflectivity of the Unit B. The sharpness of the polygonal fault pattern as seen in this map (observed all through the area except near Yolla-1 and its close surroundings) is lost compared to all the other polygonally faulted horizons (horizon 2, 4-6). It is suggested here that the middle unit B is overpressured. It is hypothesised that dewatering of the upper unit A should be facilitated through polygonal faults into the unit above while the bottom unit C might use a similar escape route to dewater into the underlying unit. The middle unit perhaps does not have access to escape routes for dewatering, leading to overpressure in that mudstone layer. More complete dewatering and hence compaction might improve reflectivity in the layering while a lack of full dewatering might build up overpressure, where the contrast needed for good reflection does not exist. Any future research into overpressured zones and their reflection characteristics and detection of overpressures from reflections would perhaps be well suited in this area.

Though there has not been any investigation into the nature of the middle unit in terms of lithofacies, mineralogy and pressure state etc, it is suggested that this middle unit is overpressured. Some of the examples shown in literature from the North Sea also indicate a similar lack of reflectivity in the middle unit surrounded by intervals with better reflectivity.

The interval velocity log for Yolla-1 shows a dramatic decrease of velocity against the stratigraphic interval corresponding to the one showing polygonal fault development. The reason for that is perhaps because the low velocity clay-dominated section in the lower part of the Torquay Group which exhibits polygonal faulting is probably less consolidated in general than the underlying Demons Bluff Formation and also because the upper part of the Torquay Group is more carbonate-rich than the lower clay-dominated section. The interval velocity curve right against the middle unit shows an abrupt decrease of velocity that is indicative of the fact the unit is perhaps overpressured. Although no pressure study of this interval has been made, these indirect geophysical observations point to the fact that the development of polygonal fault system is intricately related with overpressure development (Cartwright 1994a, b).



(a) Uninterpreted inline section (For location see figure 5.4).



(a) Interpreted inline section

Figure 5.5 Interpreted inline section showing the 7 horizons mapped (3 in Unit A, 1 each in Unit B & C of the deformed interval) and 2 in undeformed intervals. The polygonal faults have affected all the three Units defined in the deformed interval and the lowermost horizon 1 has only tectonic faulting and no polygonal faulting (refer to figure 5.13) whereas the uppermost horizon 7 does not have any signs of faulting activity (figure 5.12). The faults have affected the three Units differently although there are some common faults between the Units A & B and also between Units B & C.

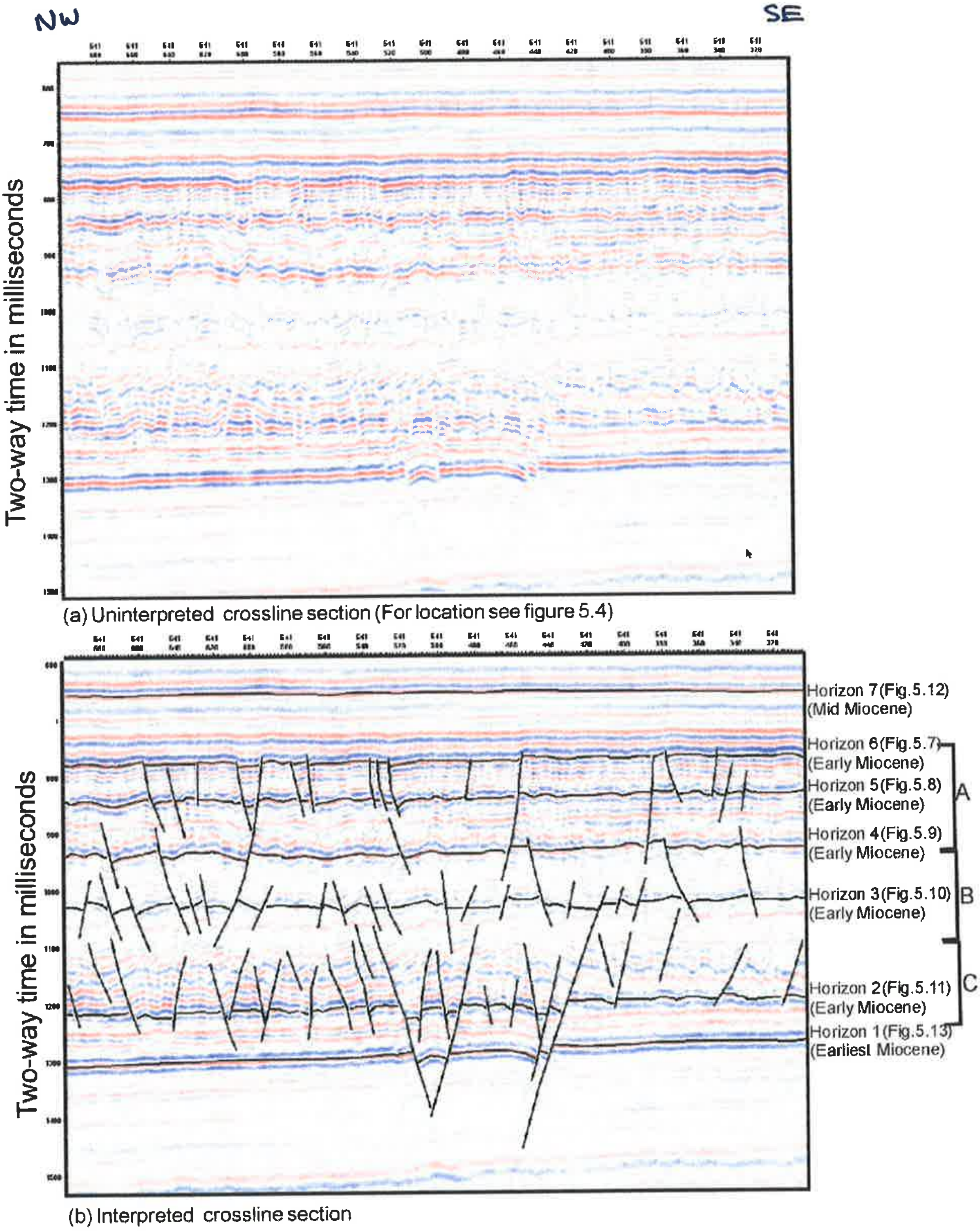


Figure 5.6. Interpreted crossline section showing the 7 horizons mapped. The polygonal faults have affected all the three units defined in the deformed interval and the lowermost horizon 1 has got only tectonic faulting. The uppermost horizon 7 does not show any faulting activity. The faults have affected the three units differently. There are some faults in common between units A & B and between units B & C.

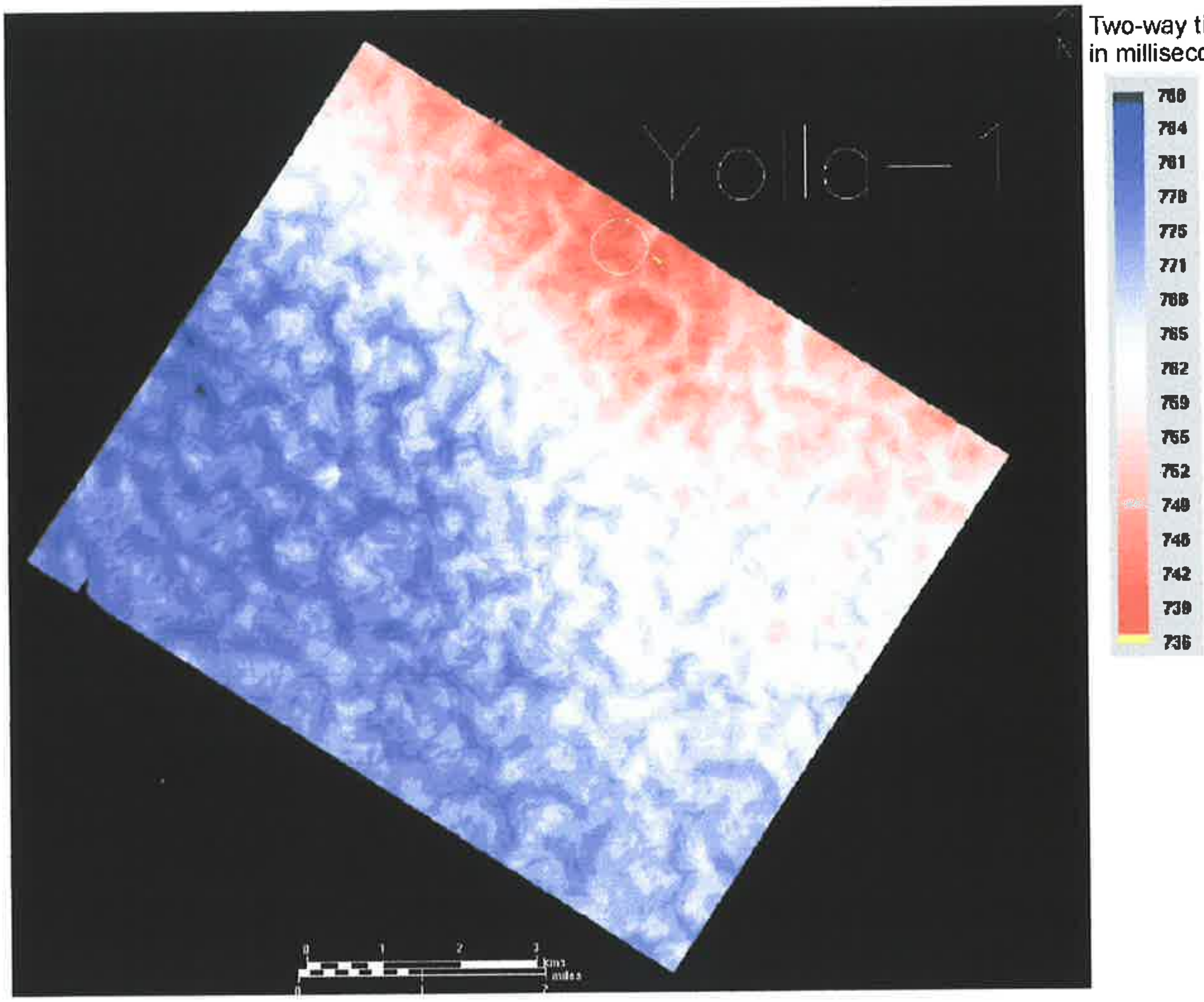


Figure 5.7 Two-way time structure map of horizon 6 within the Early Miocene (Unit A) interval. The horizon is deformed by small extensional faults that are organised in a polygonal pattern. Fault intersections are a mixture of orthogonal and non-orthogonal types involving 3, 4 and occasionally 5 fault segments.

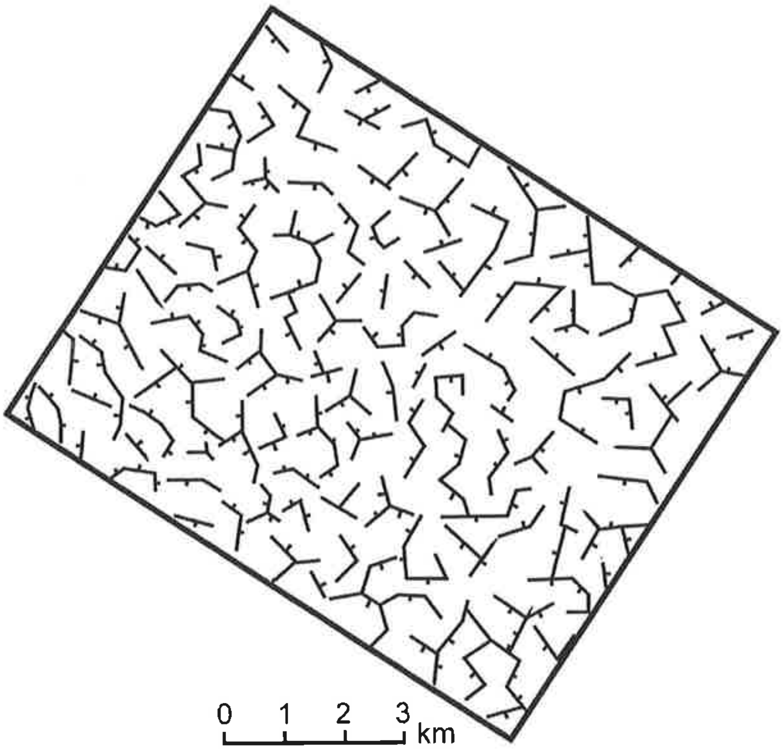


Figure 5.7a Fault trace map constructed from the two-way time structure map of horizon 6 within the Early Miocene (Unit A) interval (Figure 5.7). The fault map shows typically 'irregular' type polygonal fault geometry involving orthogonal and non-orthogonal fault intersections with up to 5 fault segments.

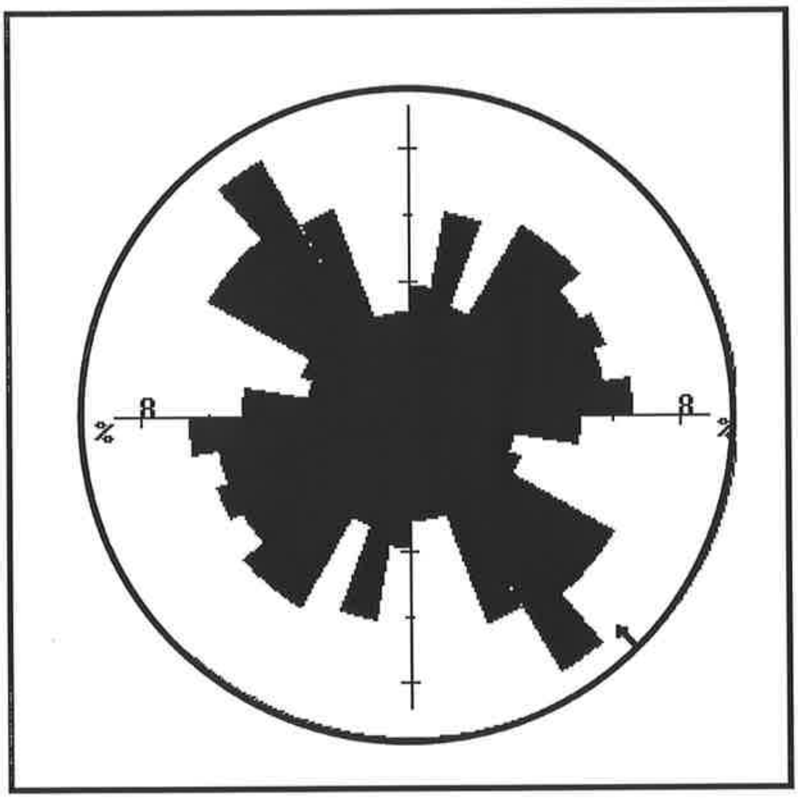


Figure 5.7b Rose diagram showing the fault strike distribution for all the faults (total 216 faults) as observed in figure 5.7a. Random distribution of fault strike is clearly seen.

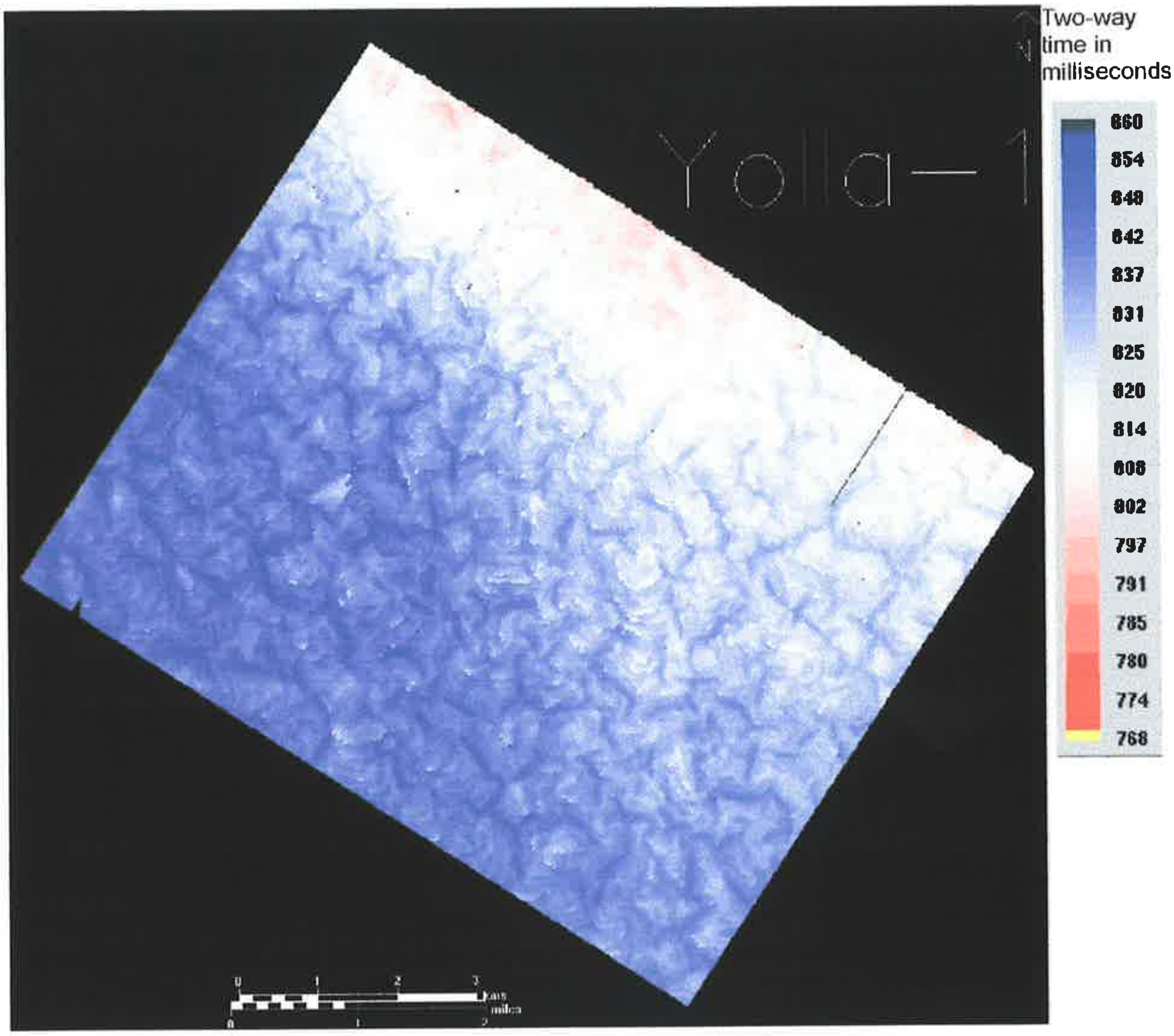


Figure 5.8 Two-way time structure map of horizon 5 within the Early Miocene (Unit A) interval. The horizon is deformed by small extensional faults organised in a polygonal network. Fault intersections are a mixture of orthogonal and non-orthogonal types involving 3, 4 and occasionally 5 fault segments.

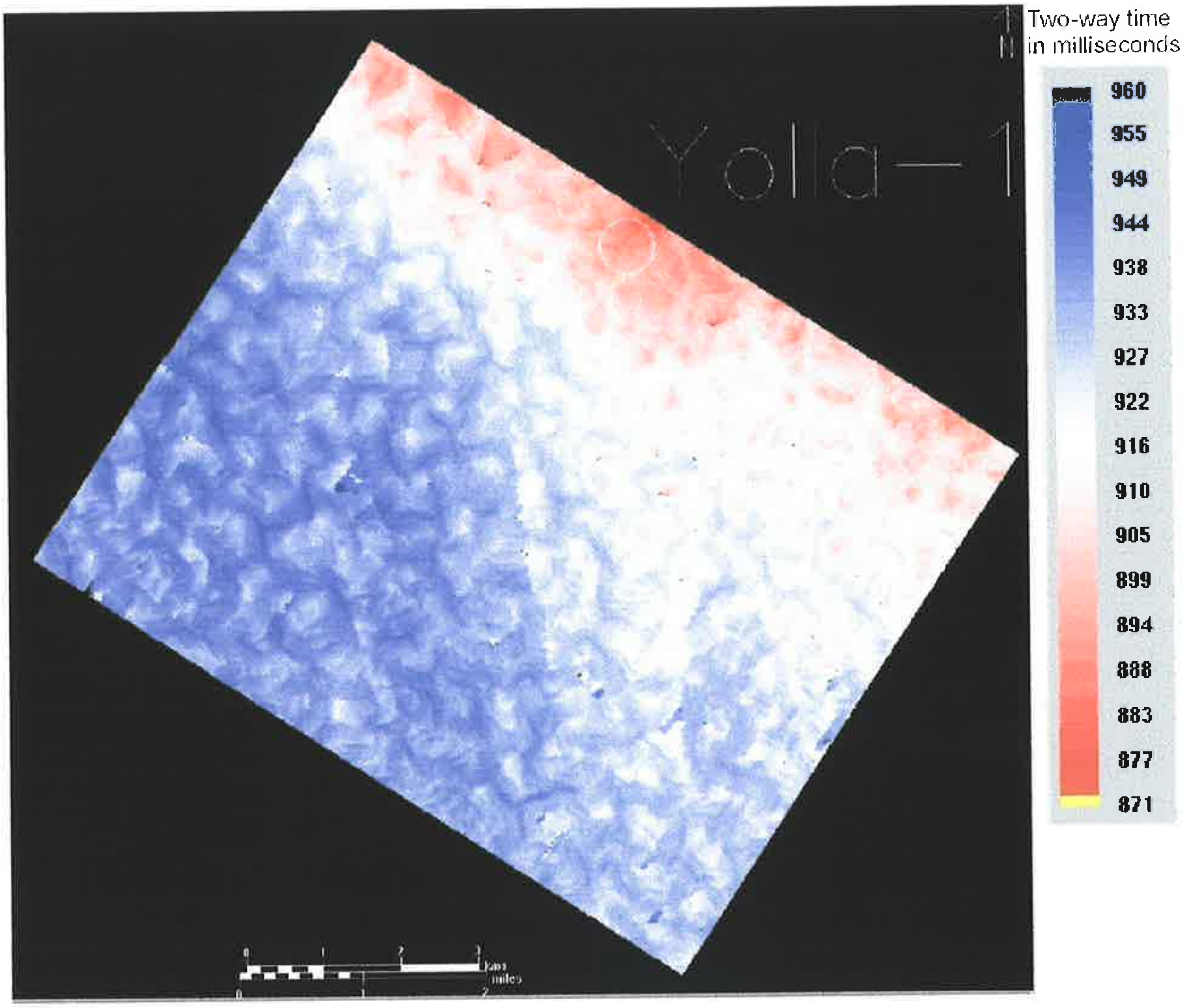


Figure 5.9 Two-way time structure map of horizon 4 within the Early Miocene (Unit A) interval. The horizon is deformed by small extensional faults organised in a polygonal network. Fault intersections are a mixture of orthogonal and non-orthogonal types involving 3, 4 and occasionally 5 fault segments.

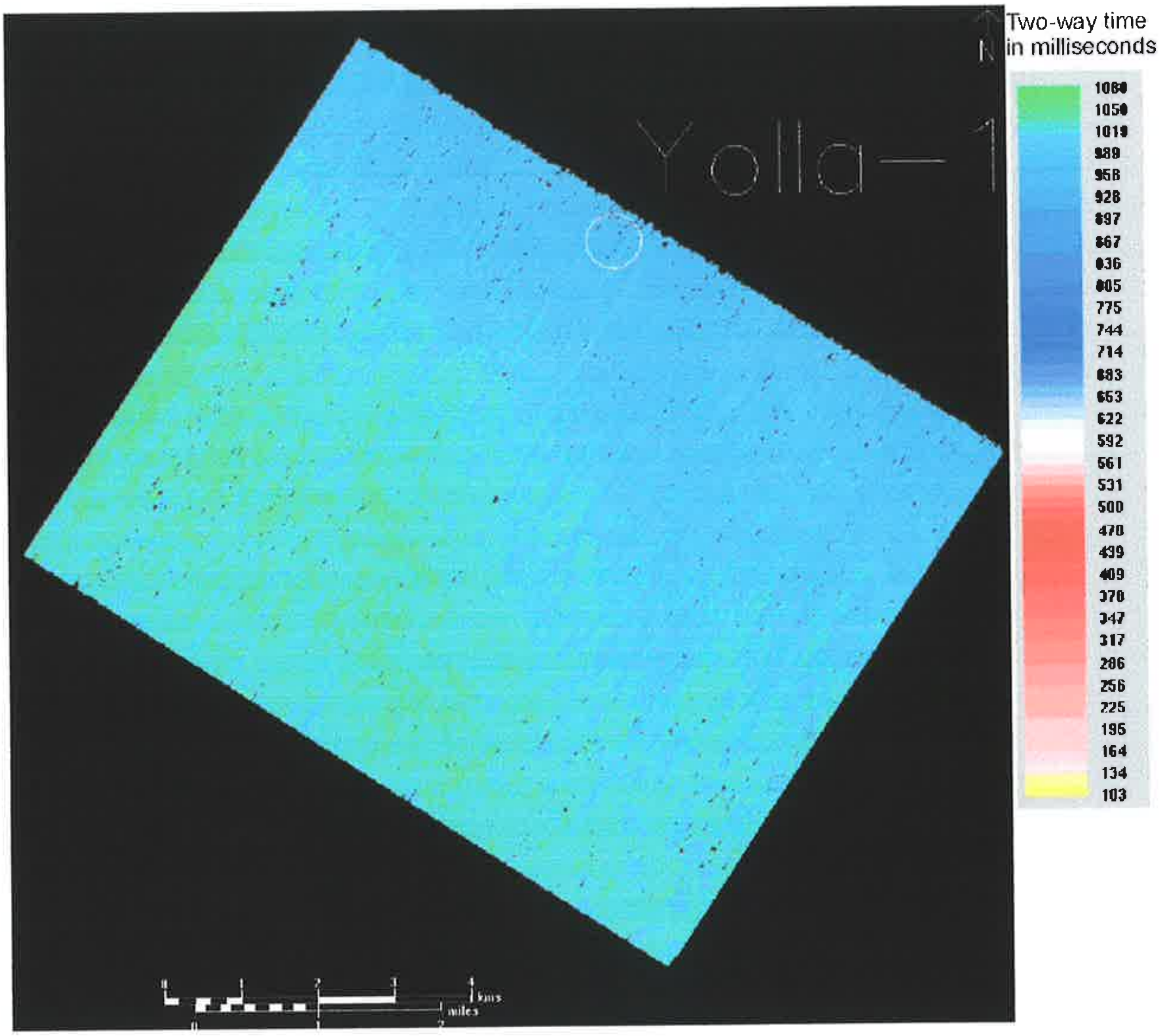


Figure 5.10 Two-way time structure map of horizon 3 corresponding to the Early Miocene (Unit B) interval. The polygonal faulting is seen in the entire map except near Yolla-1 and its close surroundings. The polygonal fault pattern is not very sharp in this map since the very low reflectivity of the Unit B has made the auto-tracking method of horizon picking much less efficient.

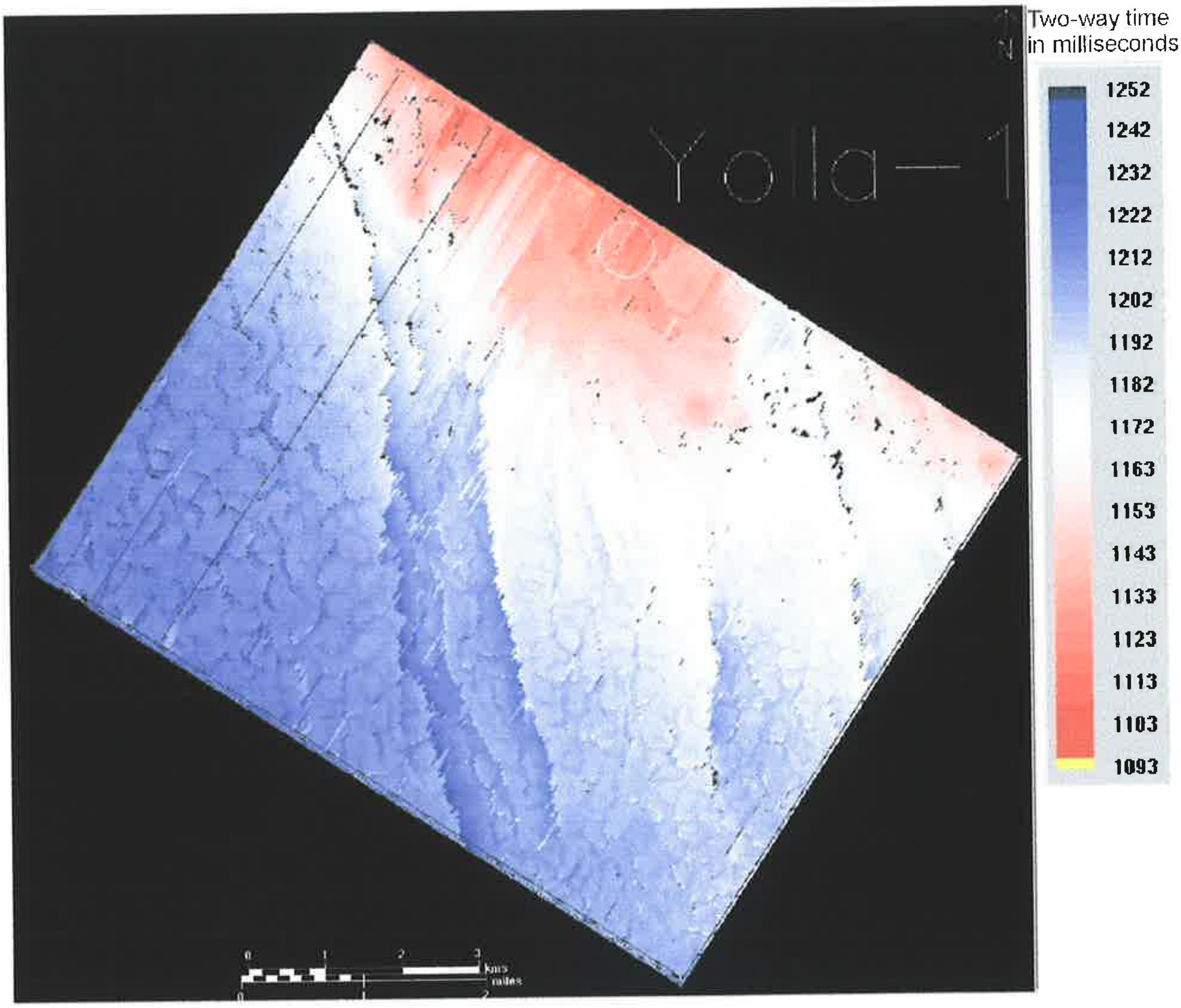


Figure 5.11 Two-way time structure map of horizon 2 corresponding to the Early Miocene (Unit C) interval. There is no polygonal faulting near Yolla-1 or to its immediate west, south and southeast. The broad tectonic fault pattern is quite evident in this map similar to the figure 5.13 in the eastern half of the map.

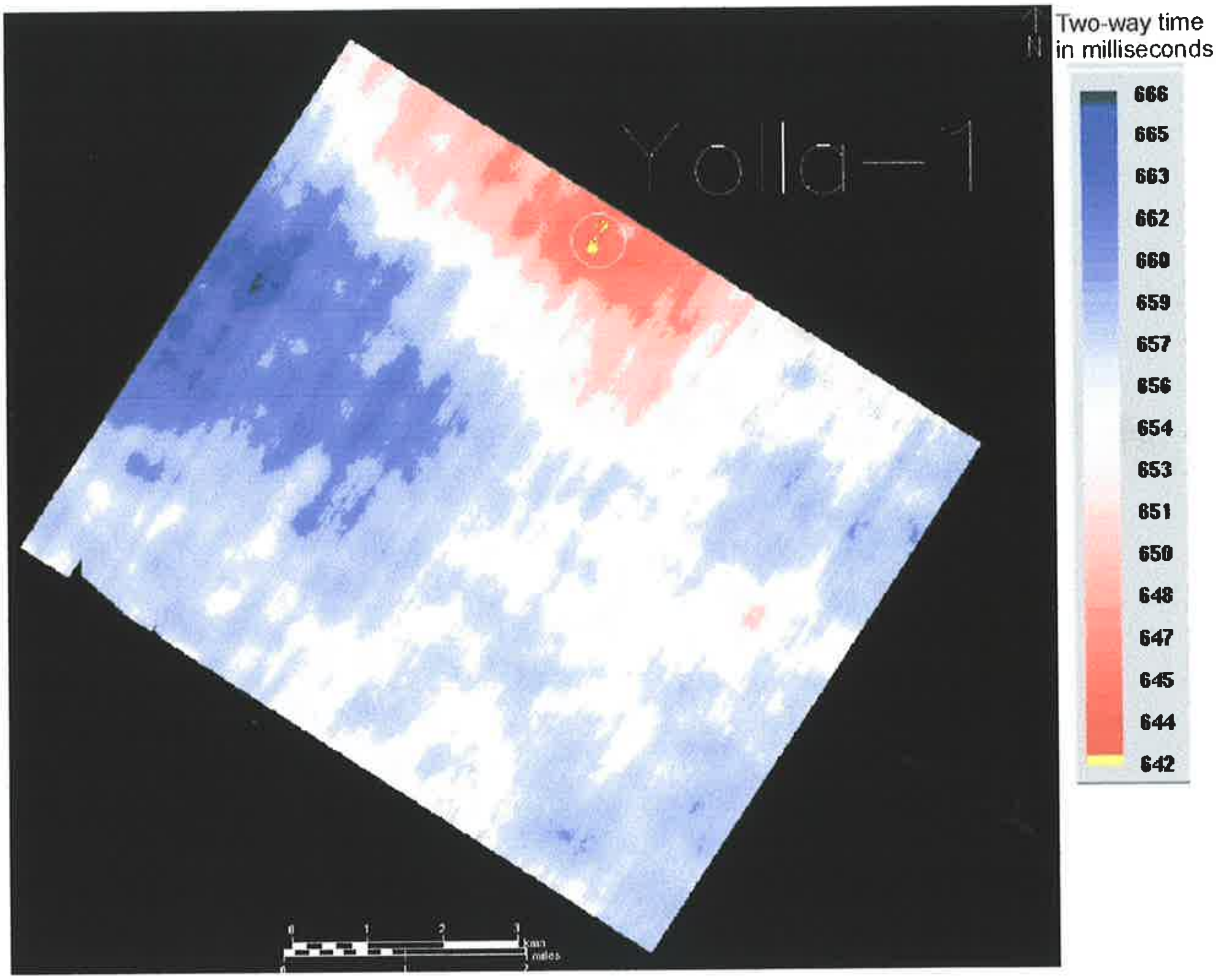


Figure 5.12. Two-way time structure map of horizon 7 corresponding to the zone above the Mid-Miocene deformed interval. Note the absence of any polygonal faulting in the horizon map. There is also no sign of any tectonic faulting at this horizon level.

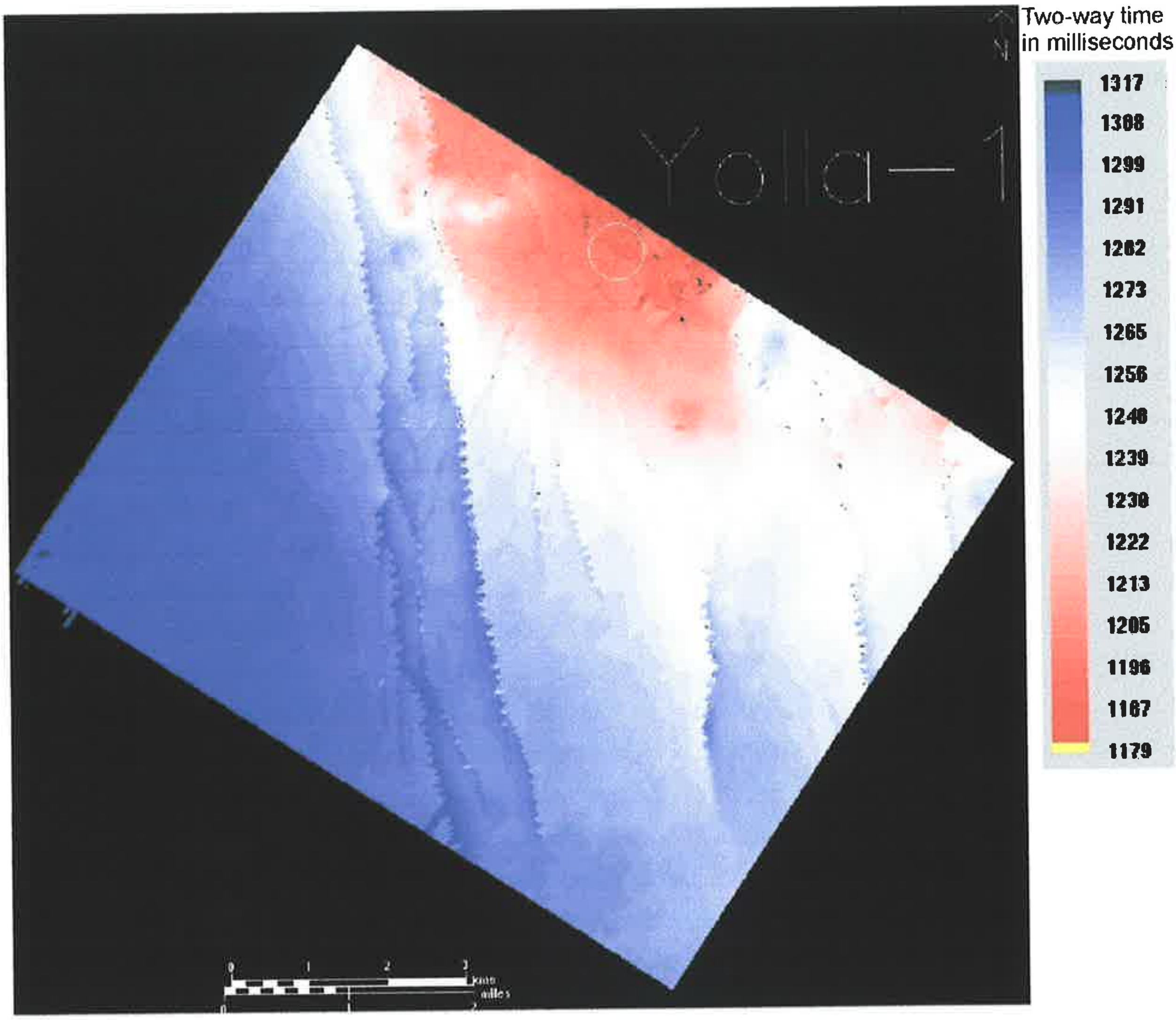


Figure 5.13 Two-way time structure map of horizon 1 corresponding to earliest Miocene, below the deformed interval. Note the absence of any polygonal faulting in the horizon map while the broad tectonic fault pattern is quite clear in the central and eastern part of the map with N-S fault trend. Note the similarity of the tectonic fault pattern between this figure and the figure 5.11.

The outstanding feature of the fault pattern seen in the Figures (5.7-5.11) is the polygonal geometry of the fault system with polygons, on average, 400m across. The fault trace map (Figure 5.7a) has been prepared for the horizon 6 within Unit A (Figure 5.7) by tracing each fault from the two-way time map. This map clearly suggests the random distribution of the fault strike orientations. The pattern does not resemble typical tectonic fault systems. There is no systematic offset of one fault trend by another nor there is any particular bias towards any fault strike orientation that is typical of a general tectonic fault system. There is a much broader range in fault orientations than would be expected for a tectonic fault set and the fault strike distribution is almost uniform in all directions. A rose diagram has been constructed on the basis of 216 small faults in the Figure 5.7a and has been presented in Figure 5.7b. It can be seen that the fault strike distribution has a very broad range although a small bias in the NW-SE direction could be seen. This small bias may be due to the general NW-SE strike of the Basin in post-rift time. The closest analog for the randomly oriented polygonal fault pattern is desiccation cracks on mud flats (Cartwright, 1996) but these are usually much smaller structures (cm- to m- scale) rather than the km-scale structures seen in all these maps. Mud cracks are vertical tensile fissures, open at the surface, whereas the structures in Figure 5.7-5.11 are normal faults in cross section with dips ranging from 20º - 50º. Though the polygonal fault patterns on each of these horizons (Figures 5.7-5.11) look to be similar in terms of fault trace length, spacing, intersection style and orientations, the entire deformed interval appears to have three different tiers that are independent of each other (Figure 5.5 &5.6). The tiers have been marked by separate styles of faulting though some faults cross cut from one tier to another. The 3-D geometry of the complex fault networks recognised on this 3-D seismic data set can be described as sets of interlocking skewed prisms, pyramids and cones. This polyhedral geometry is summarized in the Figure 5.14.

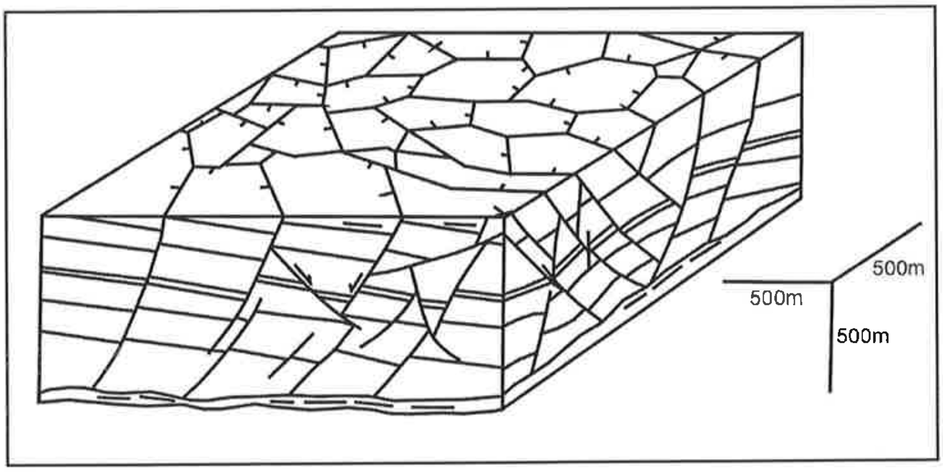


Figure 5.14. Block diagram showing the three-dimensional fault geometry typical of individual tiers in the Late Tertiary of the Bass Basin. Any orientation of cross-section through this polyhedral fault network would comprise sets of minor-displacement faults with extensional sense offsets of stratal reflections.

The bottommost horizon seems to have not been affected by polygonal faulting in the area close to Yolla-1 well and to its immediate vicinity in the southeast. The area to the north and east of the Yolla-1 well contains extensive Oligocene to Miocene volcanics. Varying acoustic velocities within the volcanics are responsible for both velocity pull-up and velocity push-down of seismic reflections (Lennon *et al.*, 1999). The ambiguous seismic response of the volcanics and intrusives causes localised perturbations in interval velocities. All this has added uncertainty and difficulty to interpretation for the lowermost polygonally-faulted horizon (horizon-2) but from close analysis of the seismic data, it is apparent that the polygonal fault system has not affected the area close to the Yolla-1 well and to its immediate southeast.

5.5 Comparison between the North Sea Basin and the Bass Basin

Figure 5.15 shows fault trace maps for two polygonally faulted horizons mapped by Cartwright and Lonergan (1996) from a 3-D seismic dataset in the central North Sea Basin. The similarities in structural style, scale, throw and orientation of the polygonal fault system between the North Sea Basin (as reported in a number of publications) and that of the Bass Basin (Figures 5.7-5.11) are striking. A set of lithological and stratigraphic parameters has been compiled for the deformed intervals in the two basins using published data and is listed in Table 5.1. In both basins, the dominant lithology of the deformed interval is mudstone. The depositional environments are similar: slope to basin floor environments in a restricted marine basin setting for both the basins. Although detailed sedimentological analysis has not been done for the Bass Basin interval, it is postulated here that the clay mineralogy in both cases should be similar, i.e., the Bass Basin mudrocks would also be smectite / montmorillonite-rich just like that in the North Sea Basin.

Table 5.1

A comparison between the depositional parameters of Lower Tertiary of North Sea and Late Tertiary of the Bass Basin

Basin	Lithology	Sedimentation rate	Depositional Environment	Age	Clay Mineralogy
North Sea	Mudstone (>70 %)	5-100m / Ma	Marine slope / basin floor	Eocene to Oligocene	Smectite
Bass Basin	Mudstone and marl	Low	Restricted marine basin	Oligocene to Miocene	Not known

Sources of data: Cartwright and Dewhurst (1998) and Williamson *et al.* (1987)

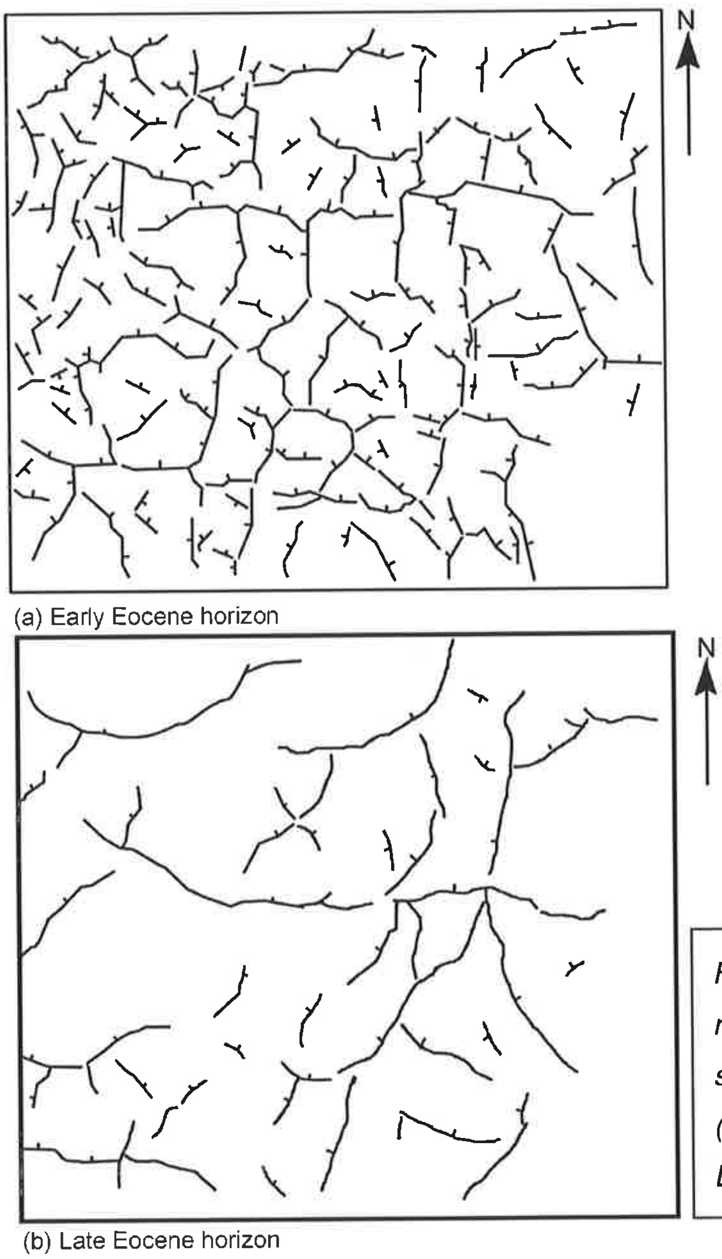


Figure 5.15 Fault trace map from two mapped horizons based on a 3-D seismic survey in North Sea Basin (Modified after Cartwright and Lonergan, 1996).

Lonergan *et al.* (1998) mapped over 30 stratigraphic horizons within six different 3-D seismic datasets in the central North Sea and categorised the fault map styles into four main types based on fault trace orientation, intersection angle, spacing, and fault trace linkage or connectivity:

I) Regular-Rectangular polygonal pattern, where majority of intersection angles are orthogonal (Figure 5.16a);

II) Curved polygonal pattern dominated by curved fault traces with a broad spread of intersection angles between 90^0 and 140^0 (Figure 5.16b);

III) Irregular well-connected polygonal fault pattern, with linear and curved fault traces and a predominance of orthogonal intersection angles (Figure 5.16c);

IV) Irregular, poorly-connected fault pattern with short fault traces, which tend to be grouped or aligned in clusters (Figure 5.16d).

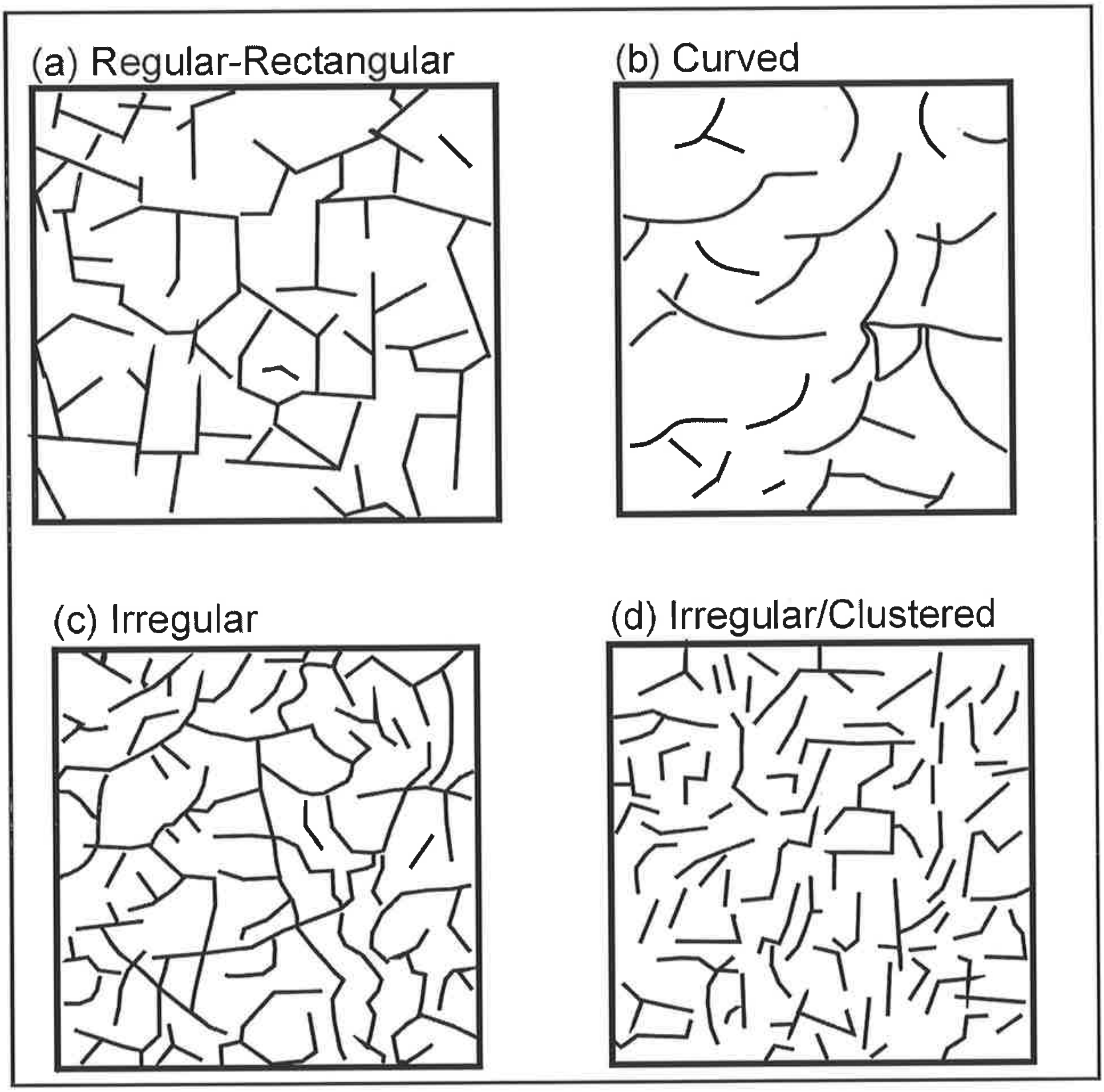


Figure 5.16 Classification of polygonal fault geometry based on North Sea datasets (Modified after Lonergan et al., 1998)

If the fault pattern obtained from the Bass Basin is compared with those from the North Sea Basin, the fault pattern most closely matches Type 3 i.e. an irregular well-connected polygonal fault pattern with linear and curved fault traces. For representation, an individual fault trace map based on the horizon map has been constructed for the topmost polygonally-faulted horizon. The fault trace map presented in Figure 5.7a has been constructed from the horizon shown in Figure 5.7. There is a good similarity in the fault pattern noted in the Figures 5.7a, 5.15a and 5.16c. It should also be remembered that only one 3-D survey data has been analysed in the basin so far and a full range of possible geometries could be observed in greater detail when more 3-D survey data are available in the basin.

5.6 Genesis of the layer-bound polygonal fault system

A discussion on the possible genesis of the layer-bound polygonal fault system in the Bass Basin is presented below drawing mainly upon the previous publications dealing with the North Sea Basin data. It is evident from the literature that over a period of few years, understanding about the mechanism of formation of this typical fault system has grown from pure kinematic and mechanical constraints to a genetic model linking the fault development to the very process of compaction and burial history of the muddy sediments.

5.6.1 Hydrofracture mechanism

Cartwright (1994a, b) proposed a model for the genesis of layer-bound fault system in the lower Tertiary mudrocks of the North Sea Basin by suggesting that episodic collapse of basin-scale overpressured shale compartments was responsible for the distribution of this typical fault system. It involves periodic build-up and release of high pore fluid pressures in the fine-grained successions, and the faulting is linked to movements induced by density inversion between under-compacted and normally compacted shale layers. This hydrofracture mechanism has been considered unlikely (Cartwright and Dewhurst, 1998) because all faults seen are not vertical. Also, it is not possible to generate a polygonal fault network by this overpressure-related mechanism with significant apparent extensional strain observed in the deformed units.

5.6.2 Volumetric contraction model

From detailed three-dimensional seismic mapping of the layer-bound polygonal fault network in the North Sea mudrock interval, and by doing detailed strain analysis on the 3D seismic data, Cartwright and Lonergan (1996) found that the apparent extensional strain on any given polygonally-faulted horizon was approximately uniform in all directions. Also, there is no preferred orientation of the fault strike of the individual fault segments and the overall fault strike distribution is random, suggesting that the stress condition remains isotropic during the formation of this fault system. The large apparent extensional strain (6% to 19%) affecting 250,000 square km area in the North Sea Basin would have required the basin margins to extend appreciably (almost 20 km along any line of 100km length) to accommodate the apparent extensional strain observed along the seismic sections (Figure 5.17a). There is, however, no evidence of such basement faulting showing contemporaneous basement extension and it is possible to invoke a general condition of pinned sidewalls (Figure 5.17b). These observations made them conclude that the fault system was not tectonic in origin. They proposed a layer-parallel volumetric contraction model for the development of polygonal faulting

(Figure 5.17b). The faulting develops in response to fluid expulsion from the mudrocks during its early burial and compaction history. The conditions for failure may be achieved through increased pore-fluid pressure or through tensile stresses generated as a result of pore-fluid loss, or a combination of these two processes (Cartwright and Lonergan, 1996).

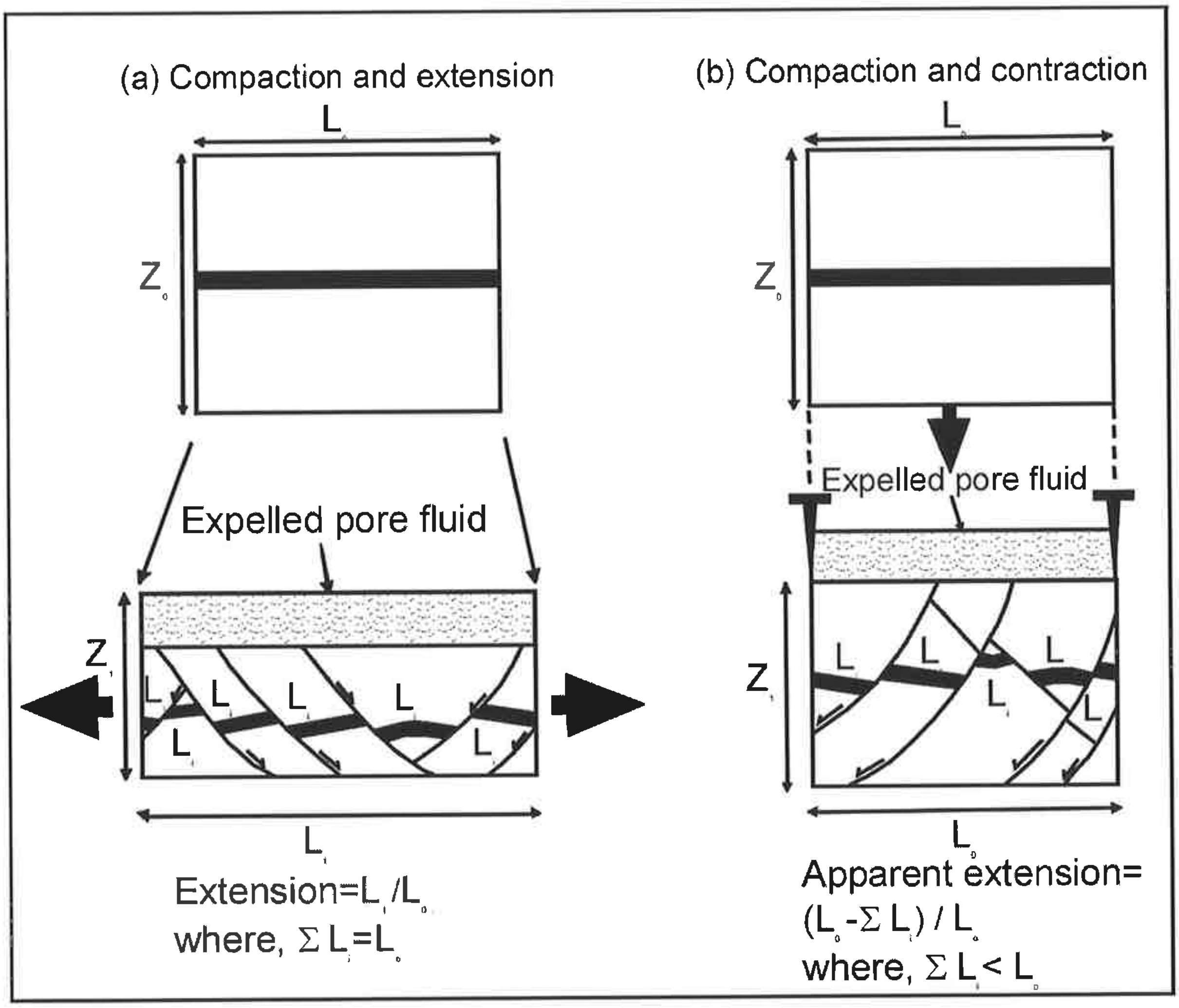


Figure 5.17 Alternative strain paths to explain the bulk strain observed in polygonal fault systems (Modified after Cartwright and Lonergan, 1996). The strain paths are illustrated two-dimensionally, since the polygonal fault network results in all cross sections having similar structural relationships. (a) A pure-shear strain path combined with one-dimensional flattening due to normal gravitational consolidation. Line balancing (preservation of original bedding lengths) requires that sidewalls move outward to accommodate the extensional strain. (b) An alternative strain path when the sidewalls are pinned, i.e., if the deforming layer is confined laterally on the scale of the basin, e.g., in the Late Tertiary interval of the Bass Basin. In this case, the only way to account for the development of normal faults is if the original bed length measured along stratal surfaces contracts during compaction. This contraction of bed length balances the extensional strain expressed in the faults. Vertical flattening strain is thus accompanied by lateral contraction strain of the sediments to produce a three-dimensional compaction (volumetric contraction).

It results from lateral contraction of stratal surface during faulting. As pore fluid is expelled, the sedimentary particles contract inwards as well as compact vertically, and this lateral contraction creates a space problem controlled by the lateral basin margin constraints. Since it is not mechanically feasible for large voids to be open at depth, the space problem is instead solved by the development of polyhedral arrays of normal faults, allowing lateral contraction to proceed. It was claimed that this model specifies a fully three-dimensional mode of compaction and therefore, is a radical departure from the conventional models of compaction of fine-grained sediments that are based on one-dimensional gravitational consolidation achieved through essentially vertical particle motion (Cartwright and Lonergan, 1996).

5.6.3 Syneresis as a physical process behind volumetric contraction

Citing examples of layer-bound faulting in 28 basins worldwide, and having compiled detailed data including lithology, dominant mineralogy, depositional environment and grain size from all these 28 basins, Cartwright and Dewhurst (1998) found that the typical layer-bound faulting is developed in depositional units composed dominantly of ultrafine grain sizes. While building on the model of volumetric contraction of Cartwright and Lonergan (1996), they suggested a suitable physical mechanism to explain why certain sedimentary units would contract volumetrically during compaction and not through the universally accepted model of compaction based on one-dimensional gravitational consolidation achieved through essentially vertical particle motion. They proposed a colloidal mechanism for the development of a layer-bound fault system. Colloidal materials are usually defined as granular materials with a dominant particle size of a few microns or less. Colloidal sediments undergoing compaction during burial can form gels, defined as 'a continuous fine-grained solid network immersed in a continuous liquid phase' (Cartwright and Dewhurst, 1998; van Olphen, 1977). Volumetric contraction during expulsion of pore fluid is a widely recognised process affecting gels of many different compositions. The process by which a gel contracts spontaneously without the evaporation of solvent is referred to as syneresis (Scherer, 1989a). White (1961) has observed fracturing induced during syneresis of clays and powdered limestones and noted that syneresis cracks were randomly oriented and inclined with shrinkage taking place in all directions. Syneresis is a highly complex process that can occur in colloids of vastly contrasting structure and composition and no simple generalization can be applied regarding the nature of particle-solvent interactions during syneresis (Scherer, 1989a, Cartwright and Dewhurst, 1998). It has been suggested that syneresis is driven by surface energy, since gels have a very large interfacial area, but it is also known that some colloids shrink as a result of chemical triggers, e.g., in metal hydroxide gels, where syneresis is often driven by condensation reactions involving hydroxyl groups (Cartwright and Dewhurst, 1998; Scherer, 1986; Brinker and Scherer, 1990). In a contrasting group of colloids, including organic gels such as cheese curd, however, it is thought that Van der Waals forces and hydrophobic attractions between protein molecules play a significant role in syneresis (van Vliet *et al.*, 1991).

Cartwright and Dewhurst (1998) suggested that different types of physico-chemical effects could promote syneresis during early stages of burial and compaction. In the claystone-dominated lithologies in layer-bound fault systems, high smectite content could provide the trigger because of the high values of specific surface area and cation exchange capacity of this mineral group. In deformed intervals consisting largely of pure nanofossil chalks, however, syneresis could be induced by diagenetic reactions taking place during early cementation. Dewhurst *et al.* (1999) carried the hypothesis of syneresis further by determining a positive correlation between high smectite content and ultrafine grain size with development of polygonal faulting. Studying a number of wells from different locations in terms of depositional setting (viz. basin floor and upslope in basin margin), they pointed out that when the lithofacies is clay-rich with a high smectite content, the faulting is intense and when it is relatively coarse-grained and smectite poor, the faulting activity diminishes.

Based on the work of Jungst (1934), White (1961), Burst (1965), van Olphen (1977), Nelson (1979), Scherer (1986, 1989a, 1989b, 1996), Brinker and Scherer (1990), Bennet & Hulbert (1986), Bennet *et al.* (1991), van Vliet *et al.* (1991), Cartwright and Dewhurst (1998) and Dewhurst *et al.* (1998, 1999a, 1999b), Dewhurst et al. (1999c) published an excellent review of the mechanism of syneresis and how it affects consolidation of clay-rich sediments while describing the polygonal fault system of the North Sea Basin. The following is a detailed discussion of the syneresis process, based on the work of Dewhurst *et al.* (1999c).

5.6.3.1 Gelation of fine-grained sediments

The process of sedimentation of fine-grained clays is dependant on a number of parameters, but the single most important factor crucial to the final structure of a clay is considered to be the electrochemical environment during sedimentation and its effect in establishing flocculating or dispersing conditions. Clay particles in a flocculating environment can link via edge-face (E-F) and edge-edge (E-E) contacts and tend to form voluminous high porosity sediments, which may lead to gelation. Clays which are deposited in a dispersing environment tend to have face-face (F-F) contacts and form more compact, low porosity sediments which do not gelate. These environmental constraints on initial clay fabric result from the delicate balance of electrochemical forces present in clays that result from their size, structure and composition. Different initial fabrics result in different physical and mechanical properties of the clay mass.

The main forces likely to be important during shallow burial and compaction include the clay double layer, electrostatic attractions and Van der Waals forces. The latter two govern the attraction of clay particles to one another, while repulsion is governed by the overlapping of adjacent double layers. The electric double layer in clays results from negatively charged faces and positively charged edges of clay plates. These charges are balanced by the attraction of appropriately charged ions in solution to retain electrical neutrality, forming the double layer. The presence of an electrolyte such as seawater suppresses the double layer, while Van der Waals forces are virtually independent of seawater concentration, so clay particles can flocculate in the water column as attractive forces now dominate. In fact, as electrolyte concentration increases, so the double layer is further suppressed. Gelation is a special form of flocculation where clay particles link together on the sea floor after sedimentation to effectively form a large mass comprising a single floc. This mass has both rigidity and elasticity as a result of the cardhouse type structure formed through E-F and E-E linking of clays. Gelation is influenced by the valence of the flocculating ion in that the lower the valence of the ion, the higher the electrolyte concentration needed to gelate a given amount of sediment.

In smectitic clays, the formation of a rigid gel structure is shown by the development of Bingham yield stress (thus allowing brittle failure) as the viscosity increases. Illite and kaolinite show little reaction to electrolyte concentration except at the lowest concentrations, but for example, in a 3.2% solution of sodium montmorillonite, yield stress continues to increase above the lowest levels of electrolyte concentration. It should be noted that a 3.2% solution of sodium montmorillonite is more than sufficient to form a rigid gel structure, and gelation has been noted in North Sea smectitic clays at clay concentrations as low as 1-2%.

5.6.3.2 Syneresis and failure of smectitic gels:

Syneresis as already defined is the spontaneous contraction of a gel network that occurs without evaporation of the pore fluid. The cause of syneresis depends on the composition of the colloid group involved, and as such has been variously attributed to condensation reactions, surface energy, hydrophobic attractions, Van der Waals forces or electrostatic attractions. The most likely cause of syneresis in marine smectitic clays is probably Van der Waals attractive forces that can predominate over repulsive forces under the environmental conditions noted above. Although gelation causes a marked change in bulk scale mechanical properties of the sediment, at the molecular scale it is relatively insignificant. The clay double layers are still suppressed after gelation since the pore fluid is a strong electrolyte, so Van der Waals forces are still operative. These forces continue to bring clay particles into closer contact, forming new bonds, although they are now laterally confined. Thus the gel network becomes even stiffer, continues to evolve and begins to contract long after gelation of the sediment.

Permeability of the sea floor and near surface sediments is likely to be high, resulting in high compressibility, so if stresses generated by gravitational compaction induce 1-D flattening strains that exceed the contractional strains resulting from syneresis, then shrinkage will not occur immediately. However, although gravitational compaction probably inhibits initial failure near the sea floor, contraction of the gel network may start by the breaking down of unstable longer chains of E-E linked clays, thus reducing permeability. This latter process is also likely to be aided by gravitational compaction in the early stages of burial and may result in an increased rate of syneresis.

The rate of syneresis is controlled by a number of properties of a gel, the most important being permeability and viscosity. Particle attraction, and hence syneresis, may also increase with changing pore fluid composition during contraction. Further suppression of clay double-layer repulsion may occur during compaction if the pore fluid becomes more concentrated, or the valency of the dominant ion in the pore fluid increases, e.g. from Na^+ to Ca^{2+}. As clays are well known as semi-permeable membranes, these processes of ion concentration or valency change may occur during dewatering of the sediments.

Syneresis strain rate α can be implicitly expressed by

$$\alpha = \sqrt{\frac{(1+\nu)\eta_{fl} \; D^2}{3(1-\nu)K\eta_g}}$$

where, ν is Poisson's ratio, η_{fl} is pore fluid viscosity, D is half the distance to a drainage boundary, K is the absolute permeability (m^2) and η_g is the viscosity of the gel network. The larger the value of α, the greater will be the variation in strain rate throughout the sediment body. This condition occurs while both permeability and viscosity of the gel is low. Differential syneresis strain rates and lateral confinement of weak sediments result in the tensile failure of such sediments by brittle faulting, and continuing shrinkage of the sediments provides the mechanism for further fault slip after failure has occurred (Figure 5.18).

It can be seen from Eq. (1) that when the rate of increase of viscosity exceeds the rate of permeability decrease, α will begin to decrease, eventually leading to the cessation of syneresis. This is consistent with the development of tiers of faults, in that the structural data indicate that the faulted tiers appear to have formed sequentially, i.e. that syneresis in the first tier of faults had already resulted in significant extensional strain before the commencement of syneresis in the second tier. It further accounts for the lack of faults above the late Miocene, where probably smectitic clays decrease in abundance by either coarse grained silt and sand together with illite, kaolinite and chlorite which generally do not form such impermeable sedimentary gels.

Eq. (1) states that the rate of contraction depends on the size, shape and thickness of the sediment mass. Analysis of laboratory-based syneresis studies have shown that the contraction rate is much slower in larger gel bodies and that there is a time lag between gel formation and the commencement of syneresis. High permeability and compressibility at the seafloor probably inhibit shrinkage near the sediment-water interface (Figure 5.18). In laboratory tests, as gel permeability begins to decrease, syneresis is inhibited by the development of local abnormal pressures within the gel due to pore fluid resistance to contraction.

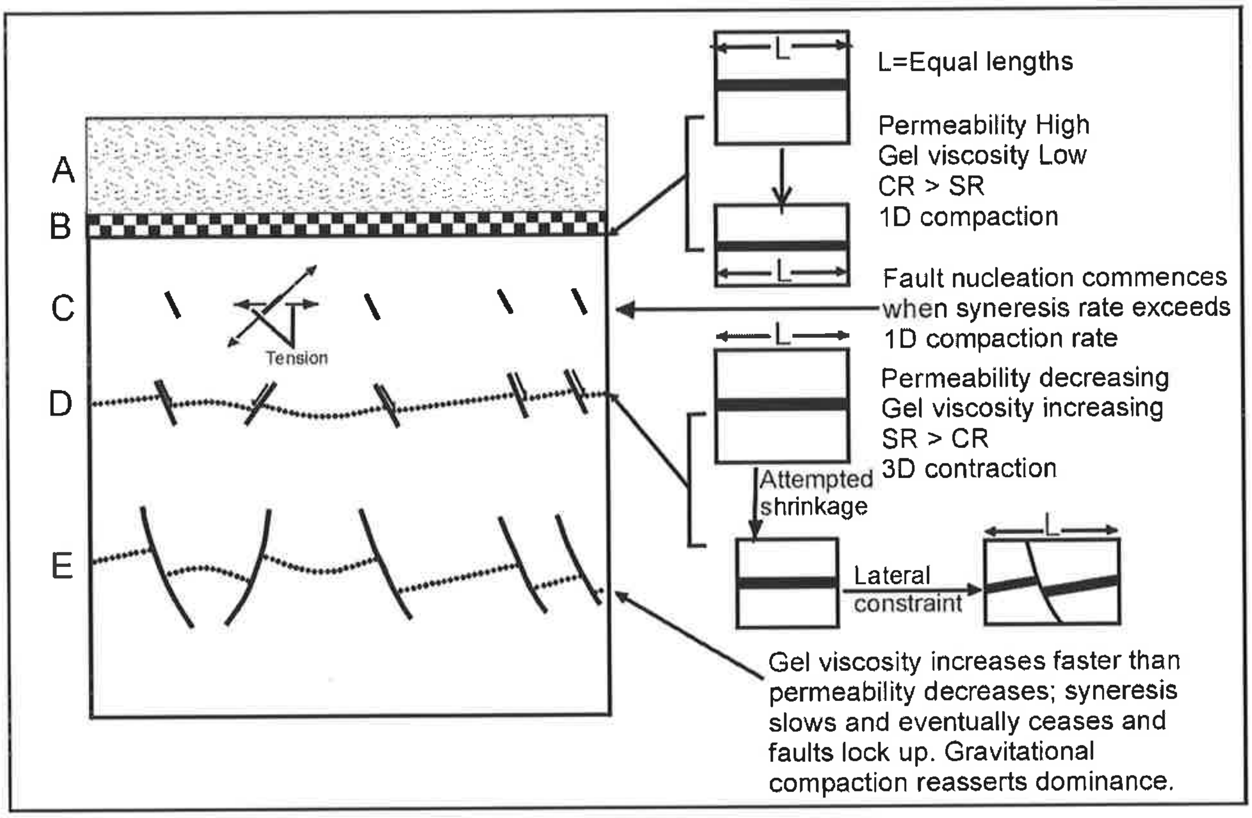


Figure 5.18. Schematic diagram illustrating the process of syneresis. In the water column (A), clay particles flocculate and sediment out. On the sea bed (B), attractive forces still predominate and flocs of clay particles form larger agglomerates and thus gelate, resulting in a voluminous high porosity sediment body. In the first few metres below the sea floor, gel permeability and compressibility are high, gel viscosity is low, and compaction rate (CR) is greater than syneresis rate (SR), so the sediment body compacts one- dimensionally. Compaction rate decreases as gel permeability falls, so strains caused by syneresis begin to dominate, resulting in fracture nucleation at shallow depth (<100m; layer C). The gel attempts to contract three-dimensionally, but is laterally confined, so the stresses set up by contraction are resolved by normal faulting. Continued shrinkage provides a mechanism to drive slip along fault planes, propagating the failure surface both upward and downward (C/D). In layer E, the rate of increase of gel viscosity exceeds the rate of permeability decrease, which eventually causes the cessation of syneresis. Faults cease to propagate, but may be rotated further by continuing gravitational compaction (Modified after Dewhurst et al., 1999c).

The main factor inhibiting syneresis after it has started is the increasing viscosity of the gel network and high viscosity will eventually shut down the process of syneresis but not until relatively high strains have been attained. Conversely, the eventual interlocking of the mechanically stronger and coarser non-clay particles may inhibit syneresis in tiers with lower smectite content and higher abundance of coarser grained detritus, resulting in lower overall strains. This is probably the case in the lower-most unit, which exhibits no or very little polygonal faulting in the eastern and southeastern parts of the 3-D area close to Yolla-1.

5.6.3.3 Mechanism of failure and slip

Figure 5.19 shows a schematic Mohr circle interpretation of the stresses affecting compacting and contracting mudrocks. The failure envelope is assumed to have a shallow gradient (low friction angle) typical of smectite-rich sediments.

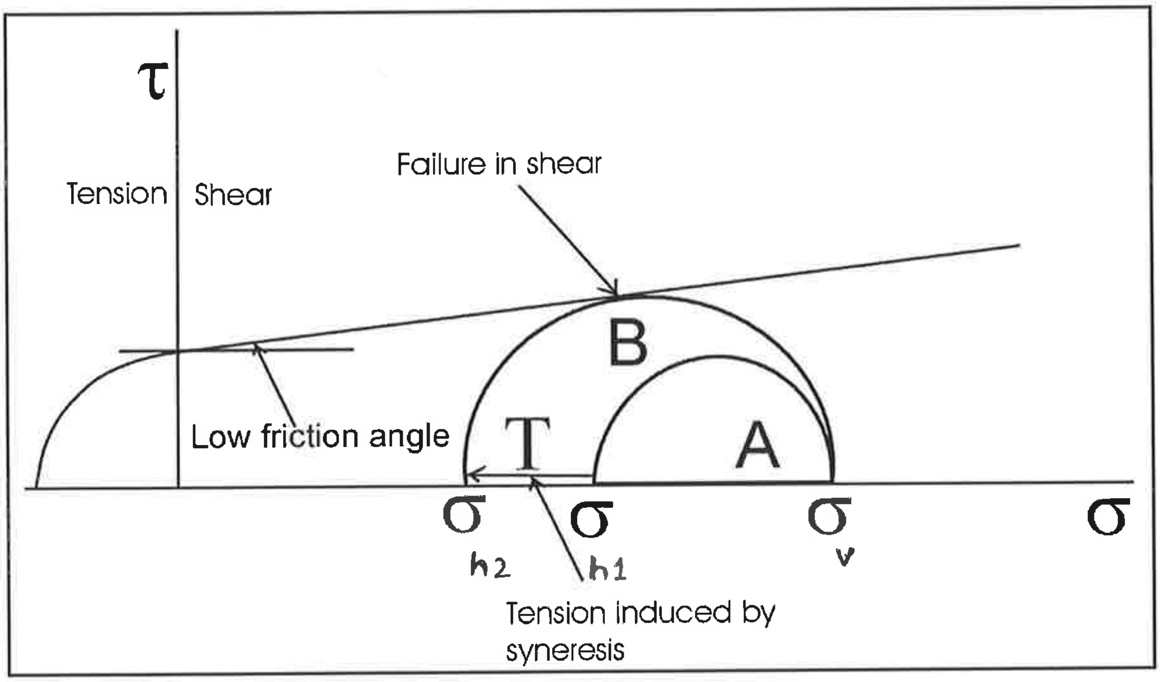


Figure 5.19 Schematic Mohr circle representation of failure in muddy sediments undergoing syneresis. Circle A represents a sediment body compacting one-dimensionally in a passive subsiding basin under gravitational forces, with vertical stress σ_v resulting in a passive horizontal stress, σ_{h1}. When contraction induced by syneresis becomes the dominant dewatering mechanism, overburden load (σ_v) remains constant, but the resultant tensile force (T) reduces horizontal stress, increasing the circle radius (circle B) , until failure occurs in shear at σ_{h2}. This results in the formation of inclined shear fractures rather than the vertical tensile fractures that are commonly caused by overpressuring (Modified after Dewhurst et al., 1999c).

Circle A shows a stress state compatible with one-dimensional compaction of a sediment residing below the failure envelope. However, syneresis induces a component of tensile stress to the sediment mass bringing a reduction in the value of σ_h. Overburden stress remains constant at this point, or may increase dependant on time and sedimentation rate. This results in the expansion of the radius of the Mohr circle to intersect the failure envelope (circle B) in the shear domain, thus resulting in the propagation of non-vertical faults. When differential stresses caused by contraction reach high enough levels to induce failure, faulting initiates at shallow depth rather than at the sediment-water interface. Continued shrinkage drives fault propagation both upward and downward and also leads to progressive reduction in bed length. The incremental reduction in bed length through contraction is balanced by an incremental increase in amount of displacement distribution on the polygonal faults.

5.6.4 Density inversion mechanism

Watterson *et al.* (2000) studied the Lake Hope 3-D survey in the Eromanga Basin and found that polygonal cell boundaries coincide approximately with the downward termination and near convergence of conjugate pairs of normal faults. The faults there have a systematic geometric relationship with folds, with anticlines in the mutual hangingwalls of fault pairs and broader footwall synclines that define the shallow dish forms of the polygons. Polygon boundaries coincide with anticlinal ridges on the interface between the faulted sequence and an underlying low velocity, low density, overpressured layer. From an analysis of the geometric and kinematic aspects of the structures, they concluded that the folding and faulting were interdependent and coeval. To suggest a genetic mechanism for the formation of the Lake Hope polygonal system, they proposed a density inversion process beneath the faulted sequence wherein a diapiric process was initiated but ceased while still at an embroynic stage, possibly due to depletion of the low density layer. While replying to a discussion raised by James *et al.* (2000), the same authors reported that the new but alternative model for polygonal fault system needs further refinement. Hence, till their model reaches a matured stage, the current debate on the formation of these typical structures will remain alive.

5.7 Implications for reservoir geometry in petroleum exploration

Lonergan and Cartwright (1999) published a classic case study of the intricate relationship between the reservoir sand development and polygonal fault structure in the Lower Tertiary Alba Field in the North Sea. Figure 5.20 is the location map of the Alba Field and figure 5.21 is a seismic section from the 3-D seismic dataset over the field showing the position of the sandstone body hosting the oil accumulation. They noted large-scale modification of reservoir geometry through sand remobilization and sand withdrawal during early burial by analysing 3-D seismic time slices and horizon maps over the field. On a mapped marker horizon in the mudrocks 80-120m above the reservoir, there is a marked decrease in polygonal fault density compared to areas away from the reservoir (Figure 5.22). On a horizon in mudrocks within 5-

50m of the base of the reservoir, there is an increase in horizon disruption by small faults directly below the sand body.

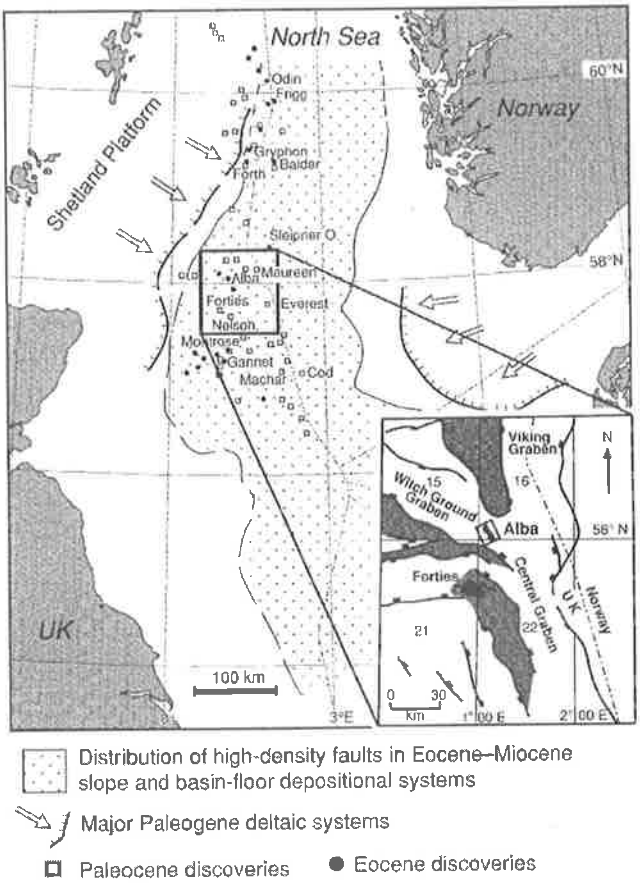


Figure 5.20 Location map of the Alba Field in North Sea (After Lonergan and Cartwright, 1999).

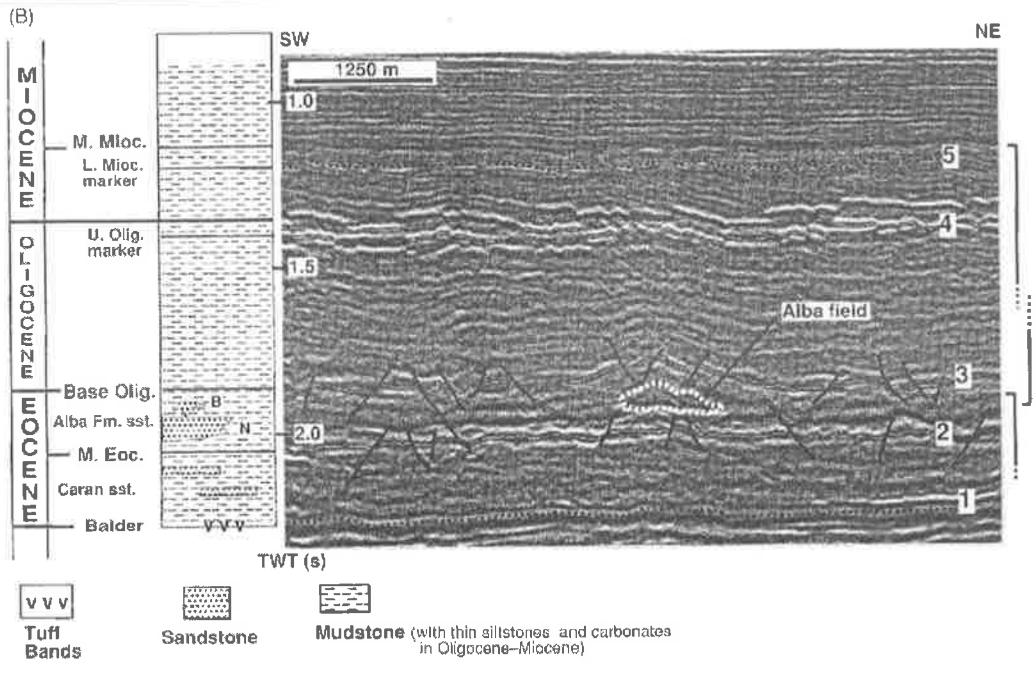


Figure 5.21 3-D seismic line over the Alba Field (After Lonergan and Cartwright, 1999).

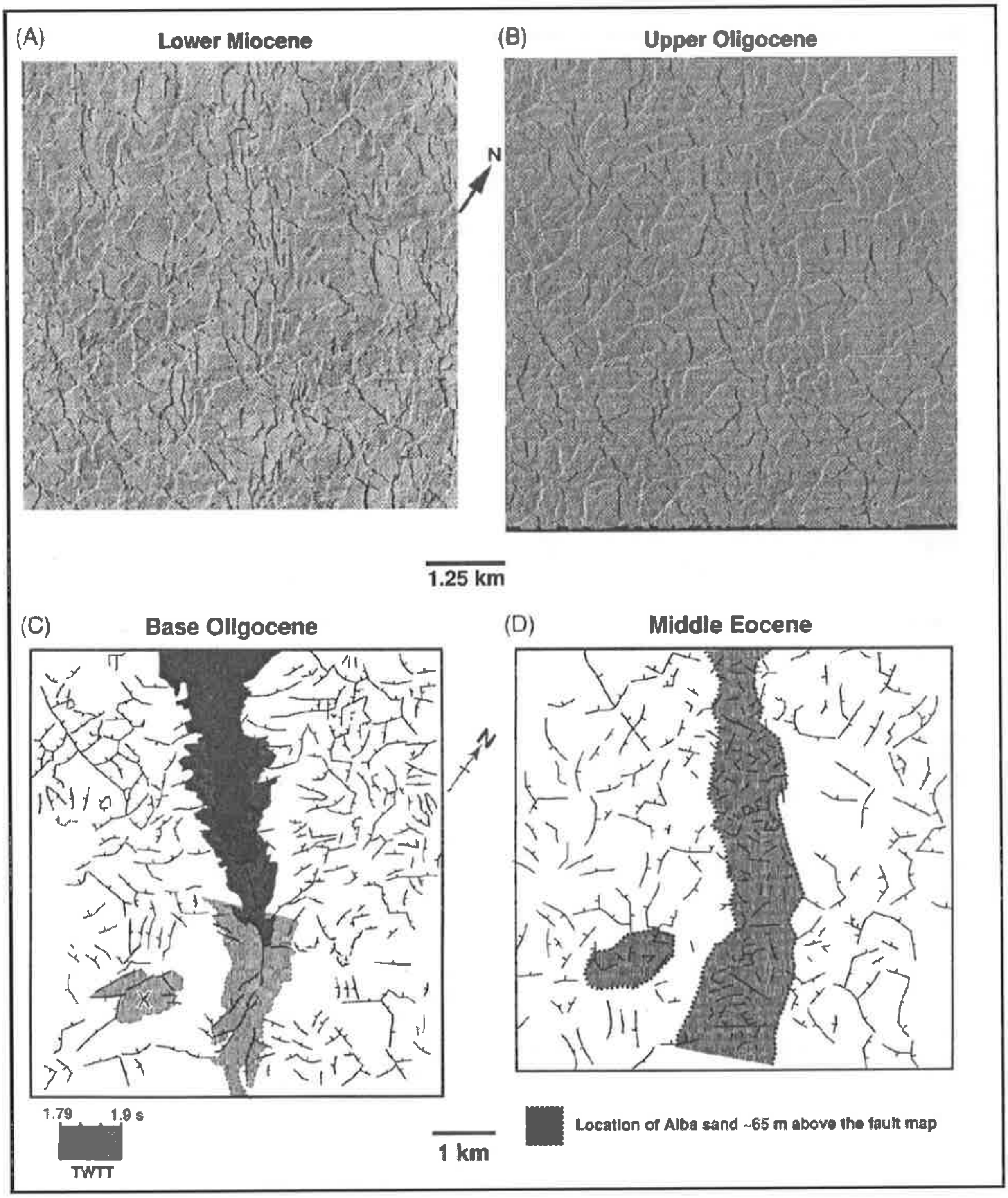


Figure 5.22 Fault maps of four horizons across the Alba field of the North Sea (Modified after Lonergan and Cartwright, 1999). (A and B) show time dip attribute maps showing the fault features by white or dark shading. Illumination being from the northwest, faults dipping in a north-northwest direction are illuminated (pale) and those dipping to the east and south are in shadow (dark). The Alba reservoir occurs between the layers shown by maps in (C) and (D). The time contours on (C) show he closure above the reservoir, and the stripple on (D) indicates the position of the reservoir approximately 65m above the mapped horizon. The polygonal fault density of C is very low whereas that of D is very high in the area corresponding to the reservoir.

Thus changes in polygonal fault density and pattern in the hemipelagic mudstones may indicate the presence of a sandstone reservoir and may prove a useful exploration tool in the search for subtle sand bodies. Though exploration in the Bass Basin currently targets the Late Cretaceous to Early Tertiary Eastern View Coal Measures, it is suggested that future exploration strategies involving Late Tertiary plays, if any, exploration for subtle sand bodies capped by fine-grained mudrocks in the Late Tertiary of the Bass Basin will require similar attention to the details of development of the polygonal fault system.

5.8 Recommendations for future work

The evidence for development of a layer-bound polygonal fault system through analysis of a 3-D seismic dataset and the possible extent of this structural style in the Bass Basin from circumstantial evidence based on 2-D seismic data has been clearly established. With the availability of other 3-D seismic datasets, it would be possible to map the polygonal fault pattern in other areas of the basin and it would be possible to categorise the geometry of the fault pattern akin to the work done in the widely-studied North Sea Basin.

Significant work remains to be done on the exact nature of variation in the intensity of faulting and its style in response to different lithofacies variation, clay mineralogy, depositional setting and grain-size variation from detailed XRD and SEM work on either core or cutting samples from this deformed interval in the basin from several wells. The Yolla-2 well drilled about 3 km to the south of Yolla-1 is well within the 3-D seismic area but the data are not yet open file. Once the well data become open file, it would be possible to test the hypothesis that grain size and smectite content of the muddy rocks constitute the critical factors in development of the polygonal fault system. The lowermost polygonally-faulted horizon mapped in this study shows that, close to Yolla-1, polygonal faulting does not exist whereas to its south near Yolla-2, it is evident. This difference in structural style between two wells is an ideal situation to study the influence of clay mineralogy and grain-size variation on polygonal fault development. Comparing the results from detailed XRD and SEM analysis of rock samples in two wells, in conjunction with closer analysis of the structural styles of the polygonal fault system from 3-D seismic map patterns in the vicinity of the two wells, the relationship between fault development and associated clay mineralogy and grain-size, if any, can be established.

No study has so far been made anywhere on the influence of palaeo-climate and salinity / composition of seawater on the development of polygonal fault system. It is suspected from the close similarity between the barred intracratonic sag basin setting of the Bass Basin and the North Sea Basin and their geological evolution through time, that these two factors (palaeo-climate and salinity / composition of seawater) during deposition of the deformed interval could influence polygonal fault development. Any future study into these aspects will reveal the influence of another broader geochemical environment of seawater and its interaction with compacting sediments into formation mechanism of this polygonal fault system.

At present, there still is a hot debate about the mechanism that is responsible for formation of polygonal fault system (Goulty, 2001). Considering all the factors that suggest a volumetric contraction model (Cartwright and Lonergan, 1996), it was believed all through the present study that the syneresis of colloidal sediments pose the most plausible model for formation of polygonal fault system.

CHAPTER 6 HYDROCARBON POTENTIAL

6.1 Introduction

Exploration for hydrocarbons in the Bass Basin since 1963 has so far yielded reasonable success with the major Yolla oil/gas discovery and sub-commercial Pelican gas/condensate field discovered. Though commercial oil / gas production is yet to begin in the basin, it is still under the active watch of the oil industry to upgrade the commercial viability of the basin in general. Yolla-2, drilled in 1998 (Lennon *et. al.*, 1999) confirmed the probable commercial viability of the Yolla oil/gas field. The estimated gas resource is about 450-600 BCF OGIP. White Ibis-1, drilled in 1998, located in adjacent acreage, encountered a gas accumulation with an estimated 85 BCF OGIP resource. Several oil and gas/condensate recoveries have been made from some other wells viz. Cormorant-1, Bass-3, Aroo-1, Poonboon-1, Toolka-1A and King-1 in addition to the large accumulation of gas/condensate in the Pelican field.

Lennon *et al.*, (1999) have given a synopsis of the source, maturation and charge history of the basin. Geochemical typing and maturity modelling suggest the richest mature source potential occurs within the EVCM particularly in the M. *diversus* to L. *balmei* interval in the Cormorant, Yolla and Pelican troughs (for location see Figure 6.1). Total organic carbon content typically ranges from 2-6% (good to excellent organic richness) and source rocks are primarily of Type II and Type III, that is largely gas prone but with significant potential for hydrocarbon liquids generation (Miyazaki, 1995). The drilling information is restricted to Upper Cretaceous in most of the wells in the basin, particularly in the Bass area. Although good source and reservoir potential may exist in the Lower Cretaceous and the Otway Group of rocks and even in the Upper Cretaceous, it has not been proven (Davidson and Morrison, 1986). In spite of several oil and gas discoveries in the basin, the exact source to oil correlation has not yet been established in any of the oil/gas found so far (Donaldson *et al.*, 1987), which means that the full extent of the number and type of petroleum systems (Magoon and Dow, 1994) operating in the basin have not yet been properly deciphered. The oil/gas found in the Bass-3 /White Ibis –1 and from the Yolla Field could have been sourced from two different petroleum systems but no work has been done yet to characterise the possible different petroleum systems in the basin. (R. G. Lennon, personal communication, 1999; Diane Edwards, personal communication, 1999)

6.2 Flat spot anomaly exploration-a case study

Detection of direct reflection from a flat fluid contact unconformable with the surrounding rock reflections to indicate gas filled reservoirs is now well established in the geophysical literature. Backus and Chen (1975) suggested that, under favourable conditions, a direct reflection from the flat fluid contact unconformable with the surrounding rock reflections could be recognised in

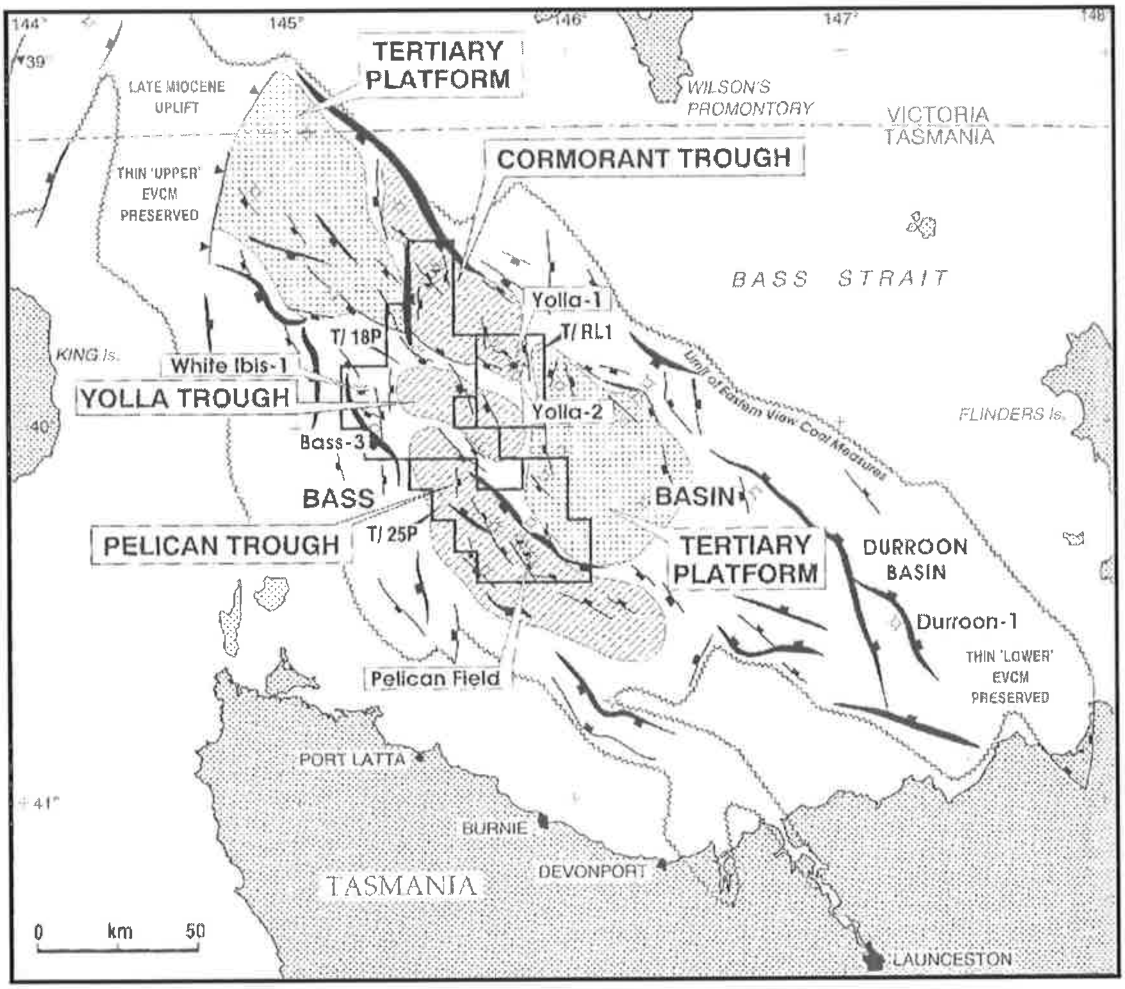


Figure 6.1 Location of Cormorant, Yolla and Pelican Troughs (After Lennon et al., 1999).

gas-filled reservoirs. Because of the uniquely low density and velocity of gas-bearing sandstone, it often shows anomalous seismic amplitudes. Also, the flatness of the fluid contact is the unique property of the reservoir that can be used to reject a multitude of amplitude anomalies caused by lithologic changes and other phenomena. These concepts provide the 'flat spot exploration' technique with enough robustness to predict the occurrence of gas in certain situations.

Blackburn (1986) has observed that the crude oils found in Bass and Gippsland basins have high API gravity (light crude) and although under-saturated, have high gas/oil ratios and therefore have a greater likelihood of detectability through the 'direct detection of hydrocarbon' method. He has also cited a good seismic example of a flat seismic event with variable amplitude, discernible from a seismic line over the Pelican-1 well. Lennon *et al.* (1999) have noted that Yolla-1, drilled in 1985 to test a distinct seismic amplitude anomaly coincident with the structure at Top Eastern View Coal Measures (EVCM) level, confirmed the predicted 'direct hydrocarbon indicator'.

The above reasons make it important to give special attention to the identification of any such 'flat spot anomaly' while carrying out the seismic interpretation work. During the course of interpretation work, two such 'flat spot anomalies' were identified from the seismic data, one in the Durroon area and the other in the Bass area The 'flat spot anomaly' in the Bass area has been investigated in detail with regard to the prospectivity of the anomaly paying all attention to the nearby well data and a close grid of seismic data over the anomaly (Das and Lemon, 2000b). This case study is presented below with fuller description. However, the area close to the 'flat spot anomaly' in the Durroon area suffers from the broad spacing of the seismic grid. Hence, only one seismic line that shows the anomaly in the Durroon area has been presented here. It is however, recommended that the close grid seismic data near the Durroon area anomaly should be investigated in a similar manner to evaluate the prospectivity.

6.2.1 Location of the area

Figure 6.2 shows the study area corresponding to the Bass 'flat spot anomaly'. The locations of seismic lines and two wells (Koorkah-1 and Seal-1) used in this case study in the Permit T99 -1 are indicated in Figure 6.5.

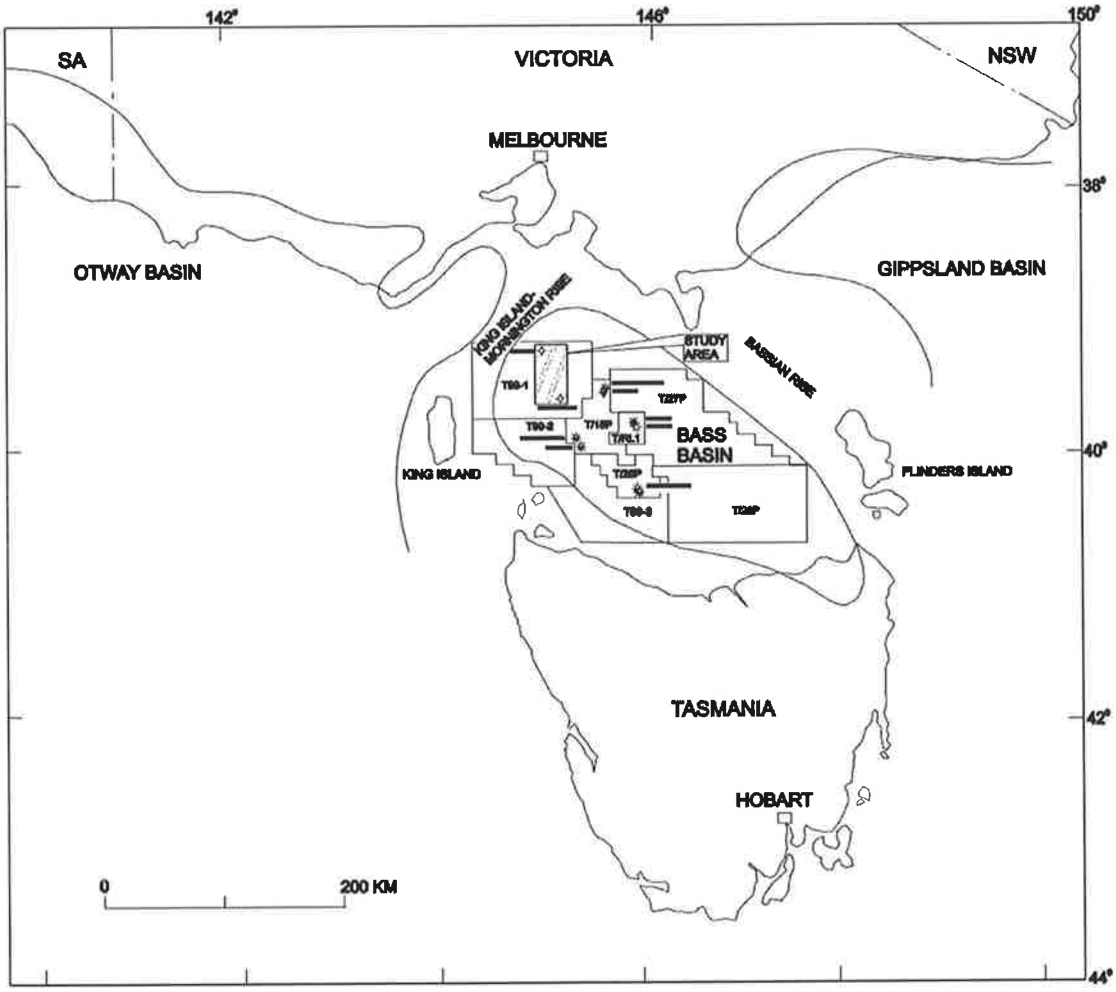


Figure 6.2 Location map of the study area covering the 'flat spot anomaly' in the case study.

The 'flat spot anomaly' identified in the seismic data is at the *M.diversus* palynological zone and stratigraphically, the level of the anomaly corresponds to sandstone of non-marine to marginal-marine facies of the Upper Eastern View Coal Measures Group (Refer to figure 2.5).

6.2.2 Seismic data

A 'flat spot anomaly' is evident on two almost perpendicular seismic lines in the NW Bass Basin, near the wells Koorkah-1 and Seal-1. The two lines are migrated stack seismic sections, one along strike and the other along dip of a small, faulted, anticlinal closure at the middle *M.diversus* palynological zone. Figure 6.3 shows part of the seismic line TNK4-40 in the variable area display mode. This line illustrates the bright seismic amplitudes at the mapped middle *M.diversus* level and also a flat horizontal reflector at TWT 1670 milliseconds (between shot points 1410 - 1595), just beneath the anticlinal feature. The flat horizontal reflector is clearly seen to be unconformable with the reflector corresponding to the middle *M.diversus* zone. The seismic line TNK4-01, running almost perpendicular to the previous line and along the strike of the structural closure, also shows the bright seismic amplitudes and a similar flat horizontal reflector (between shot points 120 - 365)(Figure 6.4). This flat horizontal reflector beneath the anticlinal feature is clearly seen in the other two seismic lines that are placed over the anomaly in the area.

In all, 24 seismic lines (about 550 line-km) covering approximately an area of 400 km^2, have been used in the mapping of the reflector that corresponds to the top middle *M.diversus* Zone (Table 6.1). The seismic data at shot point 434 on seismic line TNK4-11 (the nearest projection of the well Koorkah-1) have been tied with the well data of Koorkah-1 by correlating the depth, palynology/age and seismic two-way-time using a synthetic seismogram. There is, however, a non-uniform distribution of seismic lines and in the western half of the mapped area, data are relatively sparse.

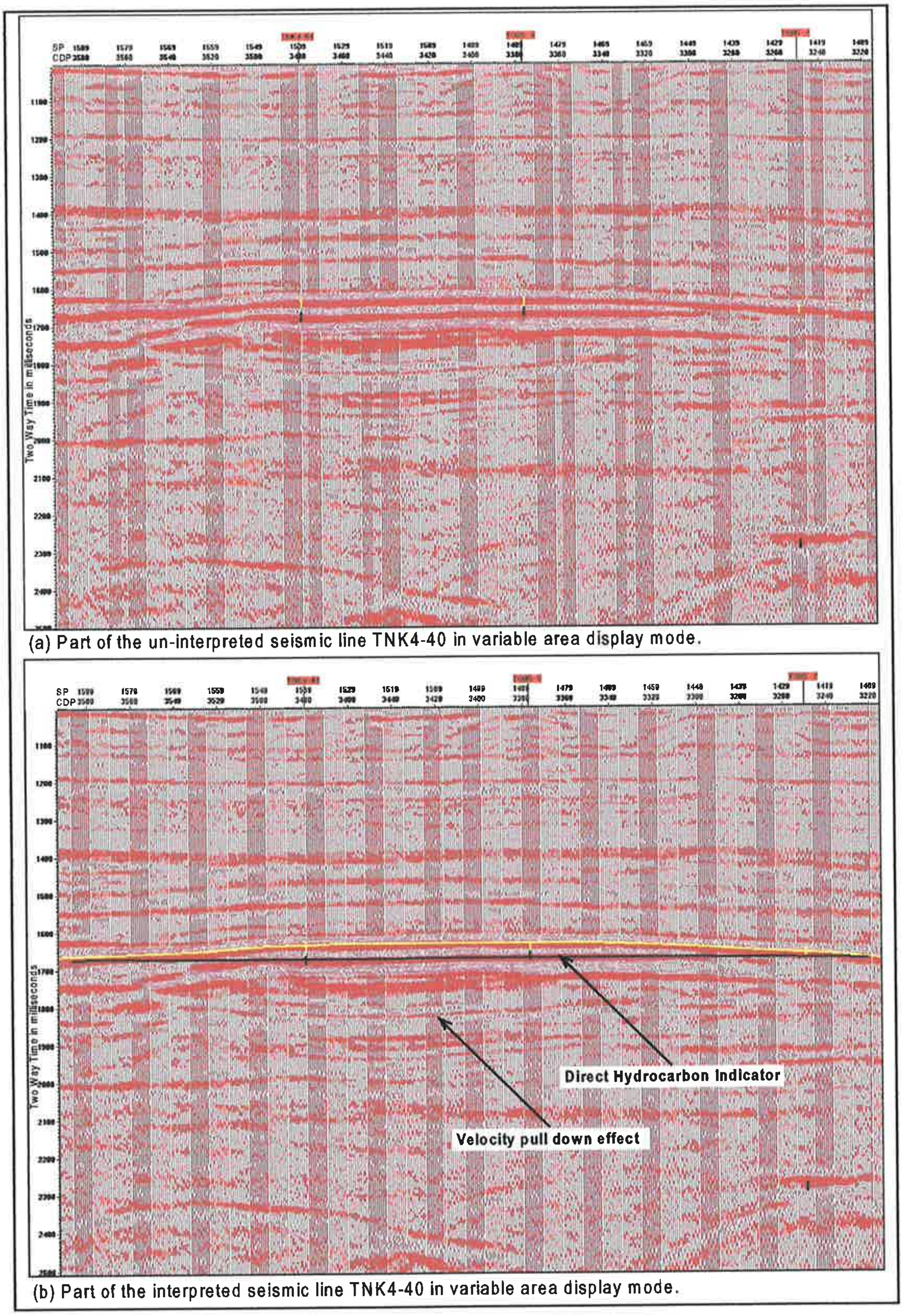
Table- 6.1

List of seismic lines used to map the 'flat spot anomaly' in the Bass area.

TQH5-01	TQH5-03	TQH5-05	TQH5-07	TQH5-09	TQH5-11
TQH5-13	TQH5-15	TQH5-56	TQH5-58	TQH5-62	TQH5-94
TQH5-96	TNK4-01	TNK4-03	TNK4-05	TNK4-07	TNK4-09
TNK4-11	TNK4-13	TNK4-20	TNK4-22	TNK4-34	TNK4-40

A two-way time structure contour map has been presented in Figure 6.5. It shows the broad outline of the anticlinal feature with a general NE-SW strike trend. Both the northern and southern margins of the anticlinal feature are faulted with the downthrown side towards south.

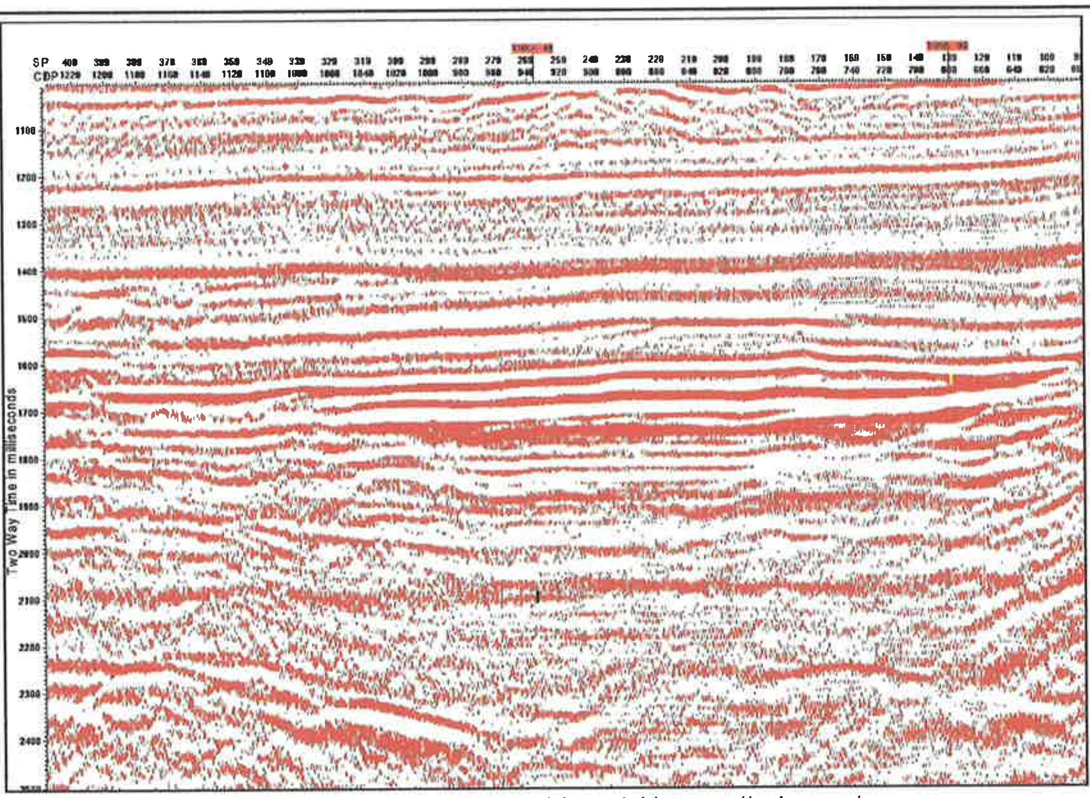
The size of the structural closure is about 40 km^2 in area with a strike length of 7.6 km, a width of 5.2 km and the maximum relief of the structure is about 40 milliseconds two-way time (Figures 6.3 and 6.4).



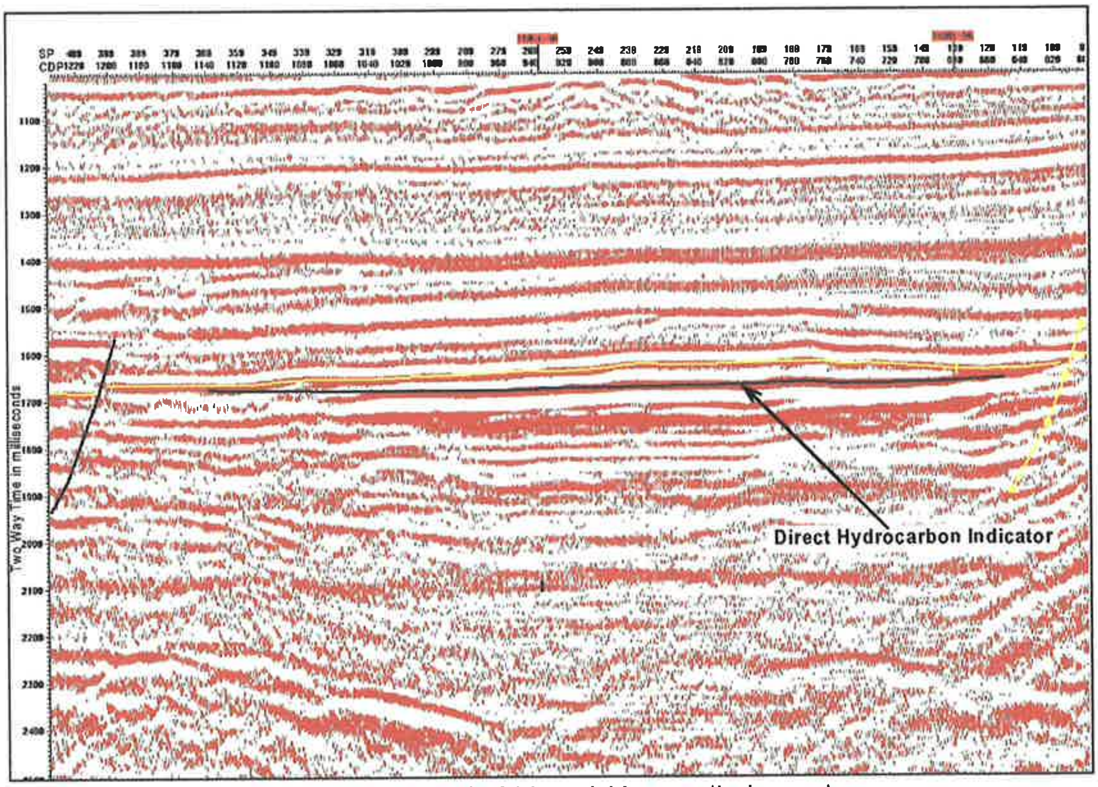
(a) Part of the un-interpreted seismic line TNK4-40 in variable area display mode.

(b) Part of the interpreted seismic line TNK4-40 in variable area display mode.

Figure 6.3 Dip line TNK4-40 showing the anticlinal structure with the 'flat spot anomaly'. The velocity pull down effect is also clearly observed.



(a)Part of the un-interpreted seismic line TNK4-01 in variable area display mode.



(b) Part of the interpreted seismic line TNK4-01 in variable area display mode.

Figure 6.4 Strike line TNK-01 showing the faulted anticlinal closure and the 'flat spot anomaly'.

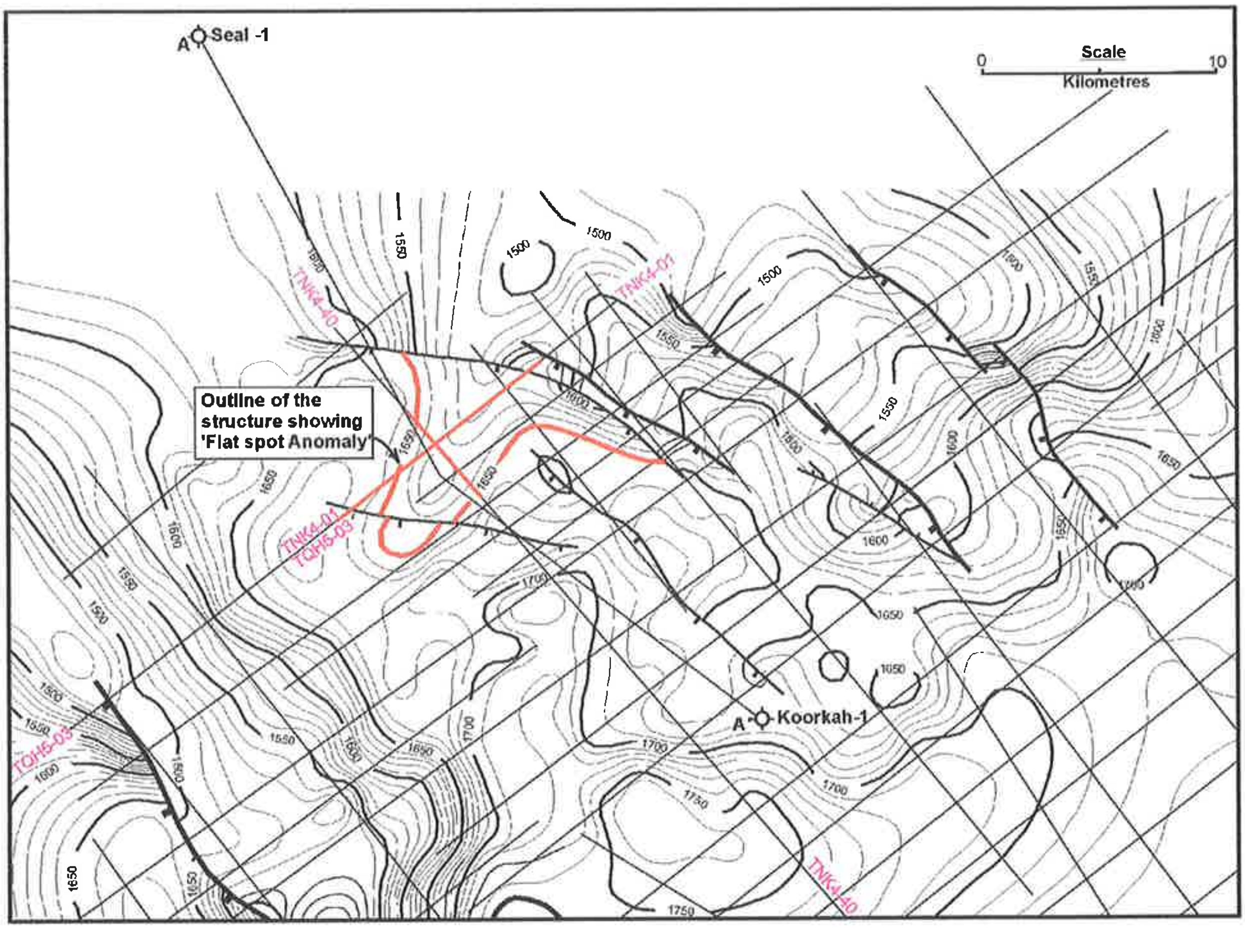


Figure 6.5 Two-Way Time structure map on top of middle M.diversus horizon. Seismic lines and wells used in the case study also indicated. The faulted anticlinal closure corresponding to the 'Flat Spot Anomaly' is shown with the red contour and the portions of seismic lines presented in Figures 6.3 and 6.4 are indicated in red.

6.2.3 Well data

Seal-1

Seal-1 (see Figure 6.5 for location), encountered the *M.diversus* unconformity at 1261.2m sub-sea (Furr, 1986) overlying a 474.5m thick lower Eocene sequence of claystone, sandstone and siltstone with minor amounts of coal. It also encountered 112.5m of igneous intrusive probably of Miocene age consisting of altered dolerite at a sub-sea depth of 1452.7m. The top of the lower EVCM belonging to Paleocene age (*L. Balmei* palynological zone) was encountered at 1565.2m and is characterised by silty claystones with fine-grained sandstone. The sediments were deposited in a non-marine to marginally marine environment.

Organic geochemical analyses of 8 sidewall cores and 2 cuttings samples between the subsea depths 1181.2m and 1604.7m (Cadman, 1986), to characterise reservoir hydrocarbons thought to be present at depths between 1327.5m to 1331.7m, suggest that the sediments are thermally mature and oil generative (1315.7m to 1348.7m). The sediments are moderately good to excellent source rocks. Though hydrogen indices (HI) suggest that the organic matter is predominantly gas-prone, a minor but significant oil-prone component may also be present (sediments at 1314.7m have HI=240).

Koorkah-1

Koorkah-1, drilled in 1985 to test the early Eocene to Late Cretaceous sandstone of a NW-trending faulted anticlinal closure, encountered the top of the EVCM at a sub-sea depth of 1582.7m. The entire 1543.9m thick EVCM sequence is dominated by interbedded sandstone and claystone with minor coal, siltstone and cherty limestone / dolomite (Amoco, 1986). The environments of deposition within the EVCM range from marginal marine (1582.7m-2218.7m) to non-marine (2218.7m-3126.6m) and the sediments were possibly deposited in upper delta plain to meandering river environments with associated flood plains (Morgan, 1986). Though the well was abandoned, there were indications of hydrocarbon fluorescence in sidewall cores. Geochemical analyses of some samples in the upper part of the EVCM showed that the sediments to a depth of 2927.7m were thermally immature to marginally mature for the generation of liquid hydrocarbons. The dispersed organic matter suggests that claystone and siltstone are relatively more vitrinite and inertinite-rich than resinite-rich and hence are more likely to be gas-prone. An igneous body encountered at a depth of 2072.7m-2106.7m may have enhanced the maturity level of the sediments (Amoco, 1986).

Stratigraphic section along Koorkah-1 – Seal-1

Figure 6.6 shows a stratigraphic section between the wells Koorkah-1 and Seal-1 constructed on the basis of the information obtained from the relevant well completion reports. The entire EVCM section has been correlated along with the Demons Bluff Formation and the Torquay Group. The M. *diversus* unconformity was identified in Seal-1 whereas it is not certain in the Koorkah-1. Similarly, the position of the Paleocene top is not certain in Koorkah-1.

The relative structural position of the 'flat spot anomaly' area is clearly seen in the figure with respect to the two wells. Though the position of the anomaly is structurally lower than Seal-1, it is actually much higher than the Koorkah-1 which encountered a very large thickness of source-rich sediments.

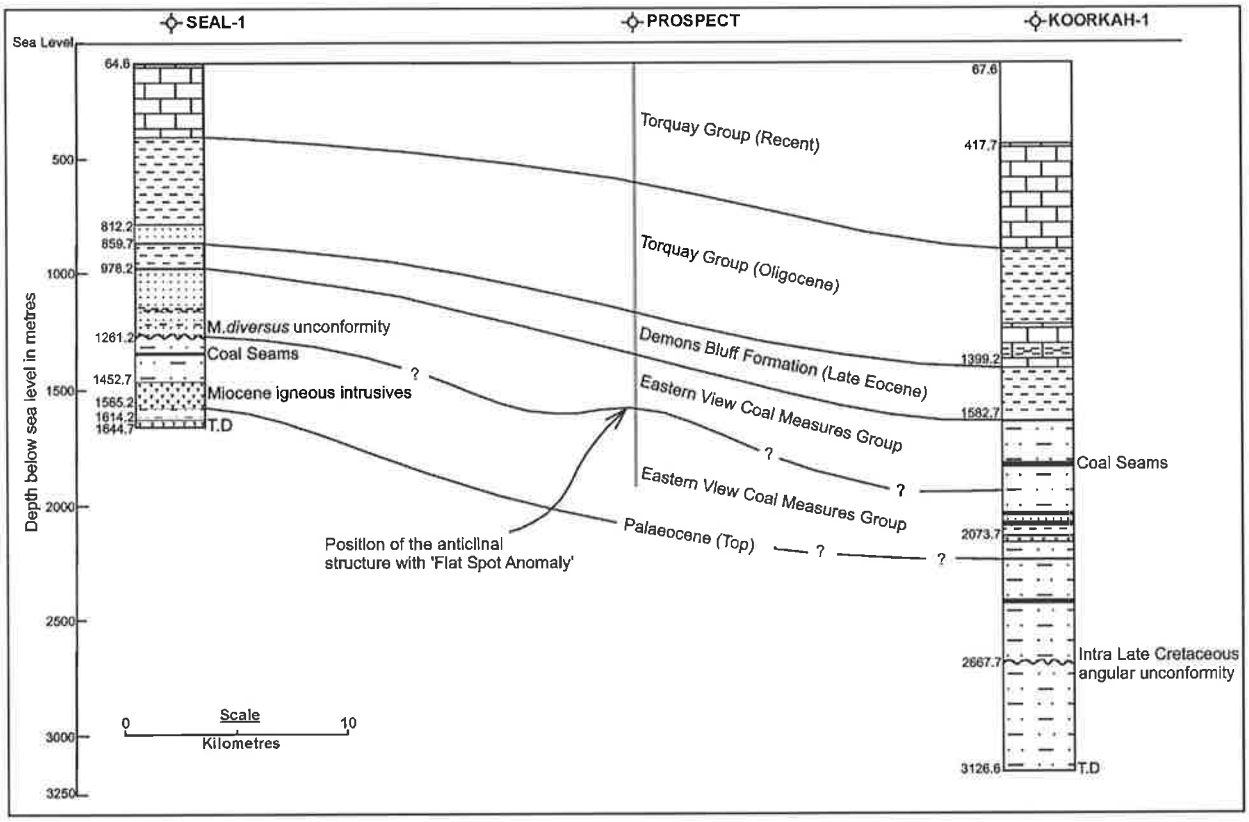


Figure 6.6 Stratigraphic section along AA' from Seal-1 to Koorkah-1 based on information from the well completion reports. The 'flat spot anomaly' detailed in the case study is also projected.

6.2.4 Significance of the T 99/1 'flat spot anomaly'

As seen from Figures 6.3 and 6.4, two important criteria of the 'flat spot reflection' technique of Backus and Chen (1975) are met in the present case:

1. The seismic amplitude of the horizon corresponding to the top of the structure (top middle *M.diversus* zone) is relatively high compared to those corresponding to the surrounding reflections.

2. The reflector beneath the structure is flat and unconformable with the rock reflections around it. This effect is also seen in other two lines (not presented in this paper) that are situated over the anomaly.

The possibility that the bright seismic amplitudes may be from igneous intrusives can be ruled out by the following observations:

1. The anticlinal feature is very clear and has maximum relief of about 40 milliseconds two-way time. The structure developed during a regional tectonic event at lower Eocene age (middle *M.diversus* unconformity) and is probably contemporaneous to similar early Eocene structures

viz. Pelican, Poonboon and Dondu described by Robinson (1974). Davidson *et al.* (1984), however, have suggested that wrenching and later compressional reactivation of some of the pre-Middle Eocene normal faults have produced anticlinal closures viz. those structures at Bass-1, Bass-2, Bass-3 and Dondu-1.

2. An igneous body would not create such a flat reflector right beneath the positive relief of the structure and it is evident that a horizontal interface in an area of dipping strata can arise from an interface between two fluids (Backus and Chen, 1975; Dobrin and Savitt, 1988).

3. There are clear indications of a velocity pull down effect on most of the major reflectors (see Figure 6.3) immediately beneath the 'flat reflector' within 300 milliseconds (up to approximately 2000 milliseconds TWT data) which is most likely caused by the presence of low velocity gas-bearing sandstone. A basaltic layer is expected to be of higher velocity than porous sediments and to generate velocity pull-up than the pull down observed.

The structure is perhaps filled with gas to its spill point, which can be interpreted from the seismic data. The 'flat spot' intersects the lowest closing contour.

Ostrander (1984) has shown that the amplitude-versus-offset (AVO) method can be a reliable technique to distinguish between gas-related seismic amplitude anomalies from other types of anomalies, viz. basalt. The AVO response in the pre-stack common midpoint gather (CMP) data shows an increase in the amplitude with increasing offset in the case of gas-bearing sandstone whereas in the case of basaltic material, it decreases. Rutherford and Williams (1989), however, have further classified gas-bearing sandstone into three families: high-impedance sandstone, near-zero impedance contrast sandstone and low-impedance sandstone depending on different geological situations and the lithofacies of the reservoir sandstone and the encasing sealing rock types. They have demonstrated that the AVO response in each case is quite different from the other. Recently, an AVO study offshore Central West Greenland (Skaarup *et al.*, 2000) indicated the possible presence of untapped hydrocarbon accumulations from the AVO and bright spot anomalies following up from crude oil seeps in Paleocene basalts and Cretaceous sediments. A lack of access to pre-stack CMP gather data has meant AVO analysis has not been done as part of this case study. Such a study should corroborate or refute the prediction of gas-bearing sandstone in the reservoir.

6.2.5 Example from the Pelican Field

Blackburn (1986) has published an example of 'direct hydrocarbon detection' anomaly from the Pelican Field along a seismic section passing through the well Pelican–1(Figure 6.7). The Pelican Field is about 110km southeast from the T-99/1 'flat spot anomaly' presented in the case study above. The 'flat spot anomaly' in the Pelican Field example at about 1.8 seconds immediately southeast of Pelican-1 (Figure 6.7) is a clear indication of the gas-bearing sandstone at that level which has already been proven for the Pelican Field.

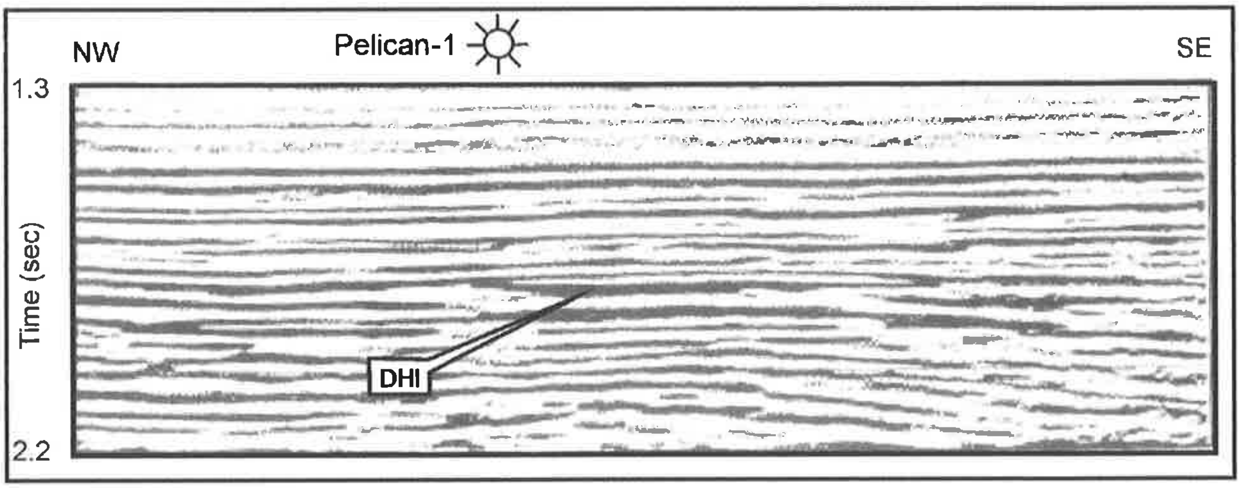


Figure 6.7 Seismic line through Pelican-1 showing a 'direct hydrocarbon indicator' (From Blackburn, 1986)

6.2.6 'Flat spot' anomaly in the Durroon area

Figure 6.8 shows the location of the 'flat spot anomaly' identified on the BMR88-309 line in the Durroon area. The reflector corresponding to the intra-Durroon (C.*triplex palynological zone*) which is chronostratigraphically older than the syn-rift sequence-2 and younger than the syn-rift sequence-1 (refer to section 4.1.4) shows a prominent faulted anticlinal structure with a 1.5 km long 'flat spot anomaly' right beneath the crestal portion of the structure (Figure 6.9).

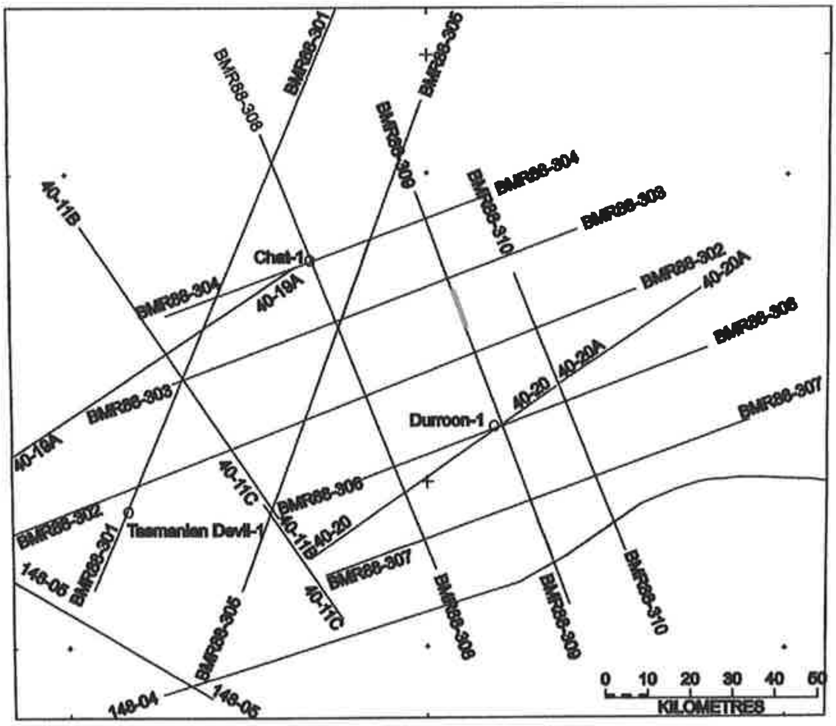
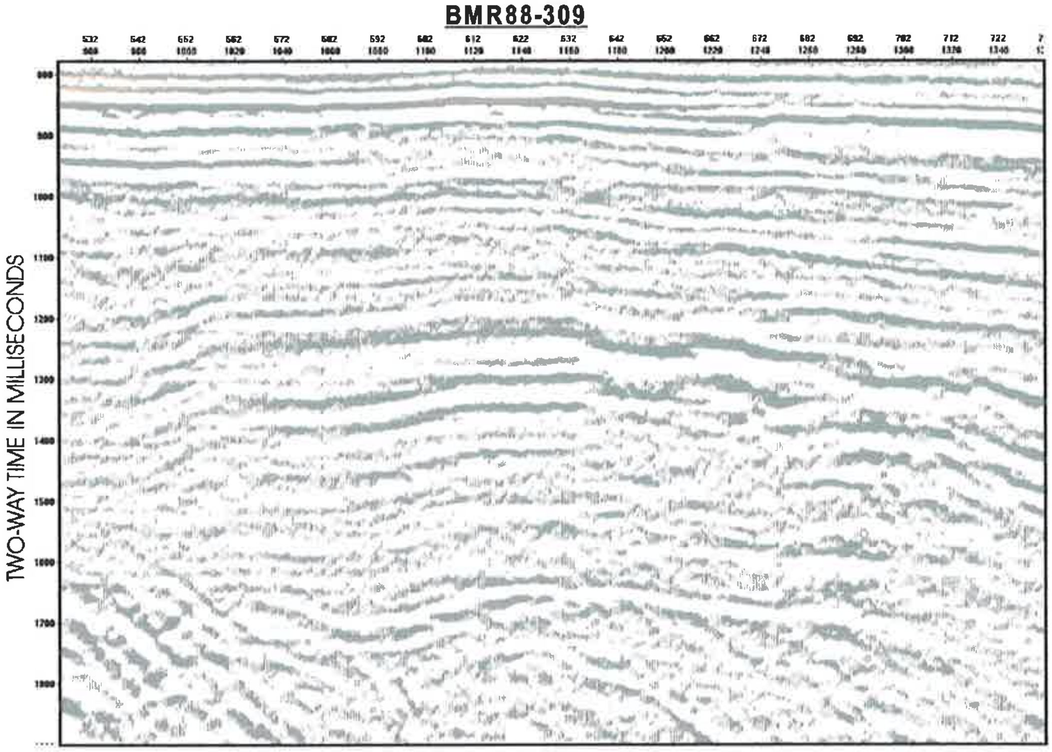
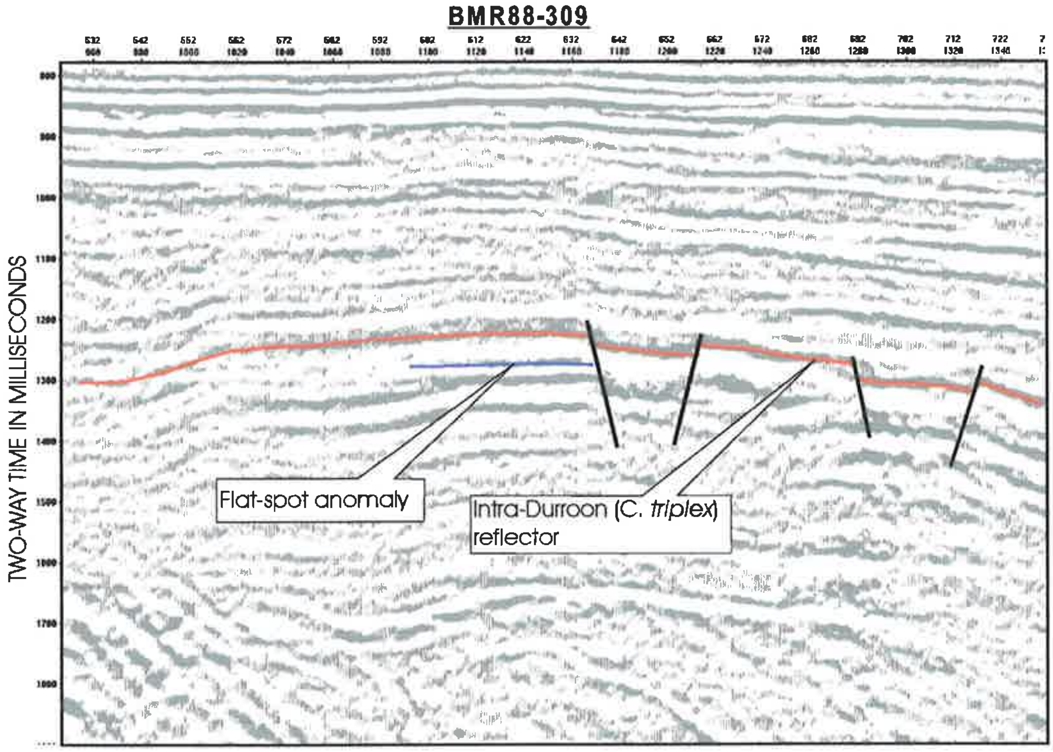


Figure 6.8 Location of the part of the line BMR88-309 showing 'flat spot anomaly' (shown in red).



(a) Part of the uninterpreted seismic section BMR88-309.



(b) Part of the interpreted seismic section BMR88-309.

Figure 6.9 'Flat spot anomaly' in the Durroon area at C.triplex level.

The 'flat spot anomaly' is evident on the line between shot points 600-635 and is at about 1370 milliseconds TWT. The data near the structure showing the 'flat spot anomaly' is very sparse and hence this anomaly has not been investigated in further details. However, it is recommended here that the details of this feature could be further investigated to ascertain its prospectivity.

6.3 Stratigraphic trap exploration

During the last almost four decades of hydrocarbon exploration activity in the basin, existence of suitable source rocks, adequate maturity level, good quality reservoir rocks and seal potential has been proven beyond doubt. However, all attention to date has been directed towards testing structural prospects, either independent closures or fault-related closures. There has been practically no attempt to look at the stratigraphic trap potential in the basin.

There is however, excellent scope for stratigraphic trap development in the syn-rift sequences on the tilted basement fault blocks of the Durroon area. A simplified model is shown in the Figure 6.10 taking into account the Durroon-1 well information and the seismic facies analysis (section 4.1.4) to diagrammatically represent the regional distribution of the various lithofacies in the Anderson fault block. The reflector positions corresponding to the eight sequence boundaries (section 4.1.3) are numbered 1 to 8 for basement to DemonsBluff Top respectively. This model is valid for any other fault block. Sandstone-rich reservoir facies can be expected near the hangingwall slope and on the footwall slope in a fan delta system. The deeper water lacustrine facies in the depocentre corresponding to each fault block would be source-rich. Interdigitation of the source-rich lacustrine facies with sandstone-rich facies in the hangingwall and footwall-derived fan deltas would present suitable stratigraphic trapping mechanism. The Durroon-1 well drilled on the footwall crest of the Boobyalla block might have missed the hydrocarbon bearing sandstone units lying relatively deeper on the slope. Although the two wells (Durroon-1 and Chat-1) were drilled on suitably located structural prospects, the huge Durroon area should not be disregarded as regards its stratigraphic trap potential.

Play types involving stratigraphic pinchouts in the sandy units as reservoirs can be good targets in any future exploration initiative in the Bass Basin in general but Durroon area in particular (Figure 6.10). These reservoirs might contain hydrocarbons migrated laterally migrated from the source-rich deeper water sediments and be sealed by lake highstand shale layers. Good quality seismic resolution of the classic rift-related sequences developed in the Durroon area make it possible to predict these play types as discussed above. Apart from the syn-rift sequences, the pre-rift Otway package also could hold good prospective targets in the tilted reservoir strata capped by the unconformities. However, the timing of hydrocarbon generation, migration and entrapment in such a situation is important with regard to time of tilting of the fault blocks and formation of the unconformities.

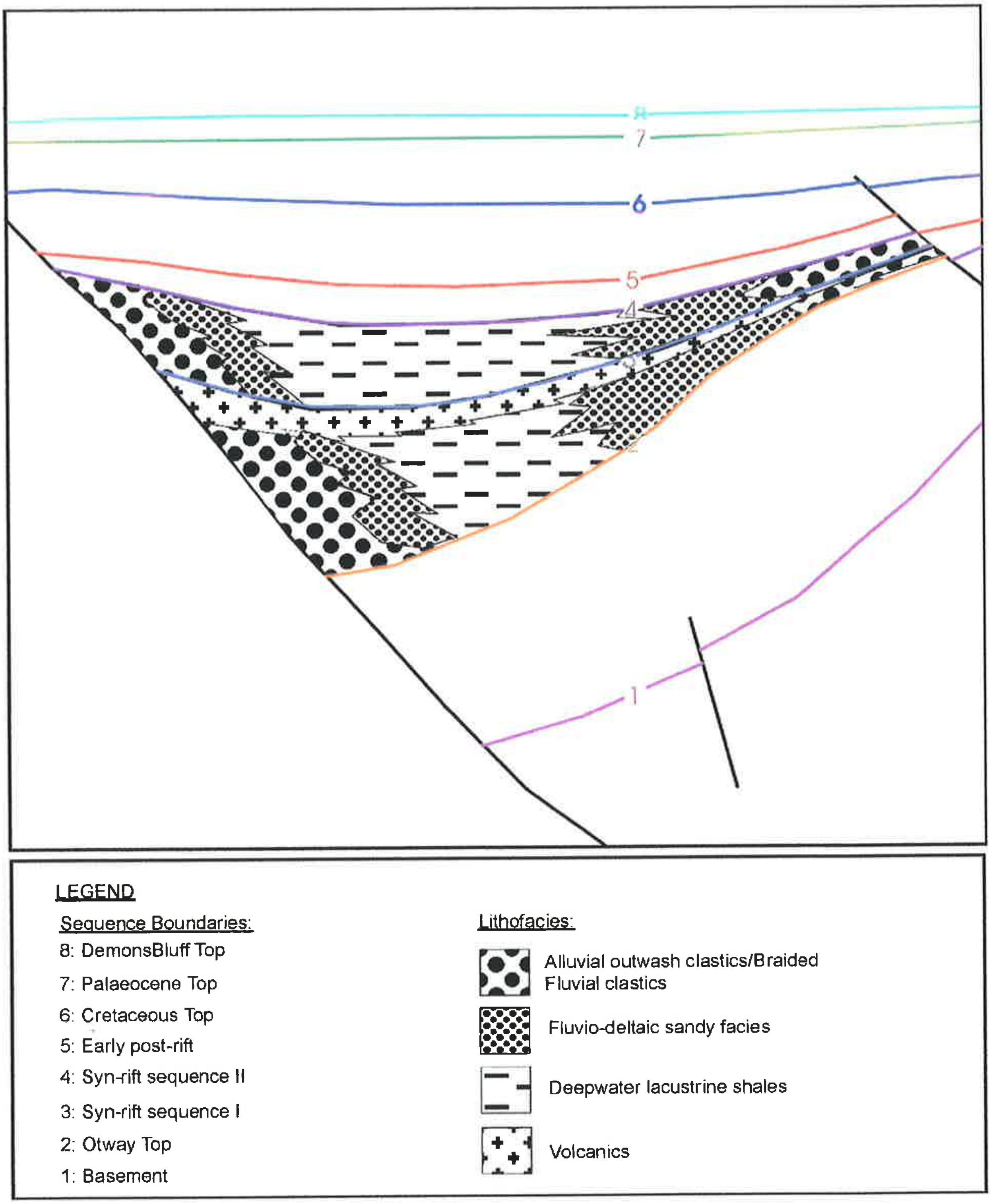


Figure 6.10 Schematic diagram showing possible lithofacies distribution in the syn-rift sequences in a tilted basement fault block (Based on data from Durroon-1 well information and seismic reflection characteristics within the Anderson block).

Morgan (1991) has reported that the Durroon Mudstone, which was deposited during C.*triplex* / P.*mawsoni* palynology zone is the age equivalent of the Golden Beach Group in the Gippsland Basin as discussed by Lowry (1987). The extensive occurrence of algal cysts in the Durroon Mudstone suggests that they could be excellent oil source material. He also noted that the algal cysts comprise 50-60% of palynomorphs in two distinctive horizons, and are likely to indicate times of maximum algal productivity in large lake systems formed during highstands. He also suggested that burial depth of Durroon Mudstone below 2000m off structure should bring it within the oil window. From seismic data, it is quite obvious that conditions for such burial in the depocentres within different fault blocks exist. All these factors enhance the hydrocarbon potential of the relatively unexplored Durroon Sub-basin to a great extent.

The fact that the Burnie Sub-basin has not been studied at all in terms of its deeper stratigraphy older than the T.*lilliei* palynological zone (Campanian) does not rule out the possibility that the age equivalent to the Durroon Mudstone, which is effectively of C.*triplex* / P.*mawsoni* palynological zone, could be widespread. The seismic sequence stratigraphic interpretation discussed in 4.2 clearly indicates that this sequence deposited in C.*triplex* zone could hold very good hydrocarbon potential for effective play types as outlined in the above paragraphs.

Morley *et al.* (1990) have emphasised the role of transfer zones in hydrocarbon accumulations in many different basins in the world viz. North Sea, Gulf Suez. Similarly, several major oil/gas fields are located along transfer zones in the Reconcavo Basin of Brazil (Figueiredo, 1994). Setiawan (2000) suggested that the Pelican-Squid transfer zone could be a suitable locale for exploration of sandstone-rich deltaic facies containing both source/trap combinations. The Pelican Field is located along this transfer zone. Future exploration work should be focussed to locate suitable hydrocarbon traps along this transfer zone.

CHAPTER 7 CONCLUSIONS

7.1 Otway rifting Vs Tasman rifting

- The pre-rift package in the Durroon Sub-basin has been correlated with the Otway Group deposited during the Otway rifting episode. The major erosional unconformity seen on seismic data near the Durroon-1 well within the Early Cenomanian succession has been correlated with the **'rift-onset unconformity'**. This unconformity marks the onset of the Tasman Sea rifting phase (dated 96 Ma, *A. distocarinatus* zone). This major tectonic event may have been synchronous with the initiation of sea-floor spreading in the Southern Ocean.

- The subdivision of the sedimentary succession into rift-related depositional units has clearly identified the separate influences of the Tasman Sea rift and the Otway rift in the Bass Basin. This study shows that the Tasman Sea rift exerted far more control over the Bass area that the Otway rift, both in structural style and stratigraphic development. This is in line with the rapid extension on the former and the slow movement of the latter.

- Four recent publications to date have categorically described the separate identities of the Bass Basin and the Durroon Basin. (Baillie and Pickering, 1991; Hill *et al.*, 1995; Gunn *et al.*, 1997a, 1997b). The definition of the Durroon Basin on the basis of a different structural history to that of the Bass Basin, by Baillie and Pickering (1991) suffers from the fact that they have not presented any material from the Bass Basin proper, to support their conclusions. The present study investigated data from the entire basin (both the Bass and Durroon area) and it is observed that all the previous publications have some element of truth in them although the exact details of how the two basins are different has not yet been properly understood.

- The interpretation that the Durroon Formation (Smith, 1986) is confined to the Durroon area and has been deposited under the influence of the Tasman Rift and a later interpretation that the Durroon Formation is equivalent to the Golden Beach by Hill *et al.*, (1995, Figure 3) suggest that the extensional stresses due to Tasman Sea rifting event did not extend beyond the Durroon area. The depositional style of the Durroon Formation, i.e., sedimentation during active rifting on tilted fault blocks, is certainly analogous to the Golden Beach Group in the Gippsland Basin. On the basis of the present study, however, it is believed that although the extensional stresses of Tasman Sea rift had greater impact on the Durroon area in terms of massive displacement along normal extensional faults, leading to greater amount of fault related extension, the seismic data in the Bass area suggest that the impact of Tasman Sea rift is expressed quite differently. Chronostratigraphically, an equivalent unit to that of the Durroon Formation could be correlated in the Bass area based on seismic reflection characteristics but remains subjective in the absence of actual drilling information in this area. However, it is safe to predict that even though the two units might have been deposited during the same time interval (Early Cenomanian-Early Campanian), they would be lithostratigraphically quite different.

7.2 Rift geometry

- Transfer zones in rift basins show a wide range of deformation including either discrete faults affected by normal slip, oblique slip or strike-slip or wide complex zones of pure normal faulting, transtension or broad warping. Transfer zones can occur at a variety of scales depending upon the size of the faults, and transfer systems of different scales may be nested one within another (Morley, 1990). Again, they can be represented by a single fault to a broad area. A single fault acting as a transfer zone may link two normal faults, basins or sub-basins with different amounts of extension or different polarity or areas of different block rotation (e.g. areas with planar faults vs areas with listric faults)(Moustafa, 1997). On the other hand, transfer zones covering a broad area exist between half-grabens with opposite tilt directions or even between extended parts of the crust characterised by different structural styles (e.g. areas with predominantly horst and grabens vs areas with titled fault blocks. These zones help transfer the throw from one half-graben to the next as the rift propagates.

- It is possible that during the Otway rifting episode in the Late Jurassic-Early Cretaceous, the Arthur Magnetic complex acted as a transfer zone between the Bass area and the Durroon area in so far as dividing the areas into two separate zones wherein two different faulting styles are observed. In the Bass area, horst and graben structures persist whereas in the Durroon area the style is one of large tilted basement fault blocks.

- Setiawan (2000) inferred a transfer zone along the Pelican-Squid alignment that also is similar to one of the two the transfer zones interpreted in the early part of this study (Das and Lemon, 1999). Gunn *et al.*, (1997b) commented that 'offshore continuation of the Arthur Lineament appears to have acted as a pre-existing line of weakness that accommodated basin extension in the northeasterly direction'.

- Integrating all the above observations, it can be concluded that perhaps the zone along the Arthur Magnetic Lineament acted as a transfer zone in the early stage of Otway extension which separates the two areas, the Bass area and the Durroon area, in terms of the style of extensional faulting.

- The same zone perhaps acted as a buttress to limit the tensional stresses during the Tasman rifting event (Das and Lemon, 2000a). A preliminary estimate of the amount of extension by measuring the pre-faulted basement length and the post-faulted basement length estimates extension of 1.36 in the Durroon area compared to 1.22 in the Bass area. Also, the fact that the Otway Top horizon is displaced by faulting similar to the basement in the Durroon area by an order of about 5 km compared to 2-3 km in the Bass area suggests that the effect of Tasman Sea rifting is more pronounced in the Durroon area compared to the Bass area.

- Although most of the previous workers in the basin have concluded that the amount of extension in the Durroon area is less than the Bass area (Etheridge *et al.*, 1985; Setiawan, 2000), the results of the present study do not support that view. However, a relatively exact estimate of extension can be worked out by considering the effect of compaction and sediment loading by the method of 'backstripping' each layer of sediments onward from the time when the first rift sedimentation started. In a recent work in progress, Cummings *et al.* (2001) have considered all the above factors while attempting to measure the amount of extension along two regional seismic lines, one across the Bass area and another across the Durroon area. The preliminary results of their work suggest an extension amount 1.12 for the Bass area and 1.26 for the Durroon area. It was expected that the reconstruction of seismic sections by depth conversion, decompaction and backstripping each layer would give proportionately a lower absolute value for extension amount because of the additional effect of sediment load on the fault displacement as observed at present day.

7.3 Different styles of extensional subsidence: Comparison between Durroon Sub-basin and Burnie Sub-basin

- The structural style in the Durroon area is characterised by large, domino-style, tilted basement fault blocks. Fault throws are in the order of 4 - 5 Km and occur along rotational normal faults. Three major fault blocks, the Bark, Anderson and Boobyalla, have undergone extremely large tectonic subsidence related to extensional normal faulting. This is in complete contrast to the structural style in the adjacent Bass area to the northwest. Although two opposing major half-graben bounding-faults have accommodated initial extension with a large fault displacement (2-3 km) in the Bass area, the predominance of sedimentary loading and syn-sedimentary faulting marks the structural style there with a large number of both planar and listric faults. The probable granitic nature of the basement in the Durroon area as opposed to basement interpreted to be composed of metasediments in the Bass area is thought to have played an important role in controlling the way in which the two areas behaved in response to crustal extension.

- The Bass area was subjected to an increased rate of post-rift sedimentation compared to the Durroon area. The maximum TWT thickness of the sediments in the Bass area is about 5 seconds (along line 40-11A) with about 4 seconds in the Durroon area (along line BMR88-307). An estimate of the approximate thickness of sediments in the Durroon area by utilising average velocity of 3.5 km/sec up to total depth gives the maximum thickness to be around 7000m. A similar estimate for the Bass area by considering the average velocity 4 km/sec, the maximum thickness of sediments is about 10 km. This variation in the thickness of the sediments deposited in the Bass area actually is the result of rapid accommodation development brought about by sediment loading and crustal flexure in the Bass area during thermal cooling stage. The huge accommodation space generated during the thermal cooling phase is perhaps linked to the fact that probably the Bass area was hotter than the Durroon area during the two rifting events. Setiawan (2000) has rightly shown a NW-SE-oriented cross section model showing the Pelican-Squid transfer zone acting as a basement high area during Otway rifting separating the Burnie and Durroon Sub-basins as unconnected depocentres. This was followed by progressive burial (his figure 14) with a much greater thickness of sediments deposited during Tasman Rifting (named as Bass rifting by Setiawan following the terminology of Hill *et al.*, 1995) and thermal subsidence phase. However, Setiawan does not give a reason for the large thickness variation. Perhaps, as mentioned above, it was caused by hotter crust beneath the Burnie Sub-basin than beneath the Durroon Sub-basin during the rifting phase, leading to increased thermal relaxation and regional downwarping in the Burnie Sub-basin compared to the Durroon Sub-basin. Future study in the basin might help in determining volume of sediments in each stratigraphic interval and would better clarify the hypothesis further.

- Although Setiawan concluded greater extension in the Burnie Sub-basin than the Durroon Sub-basin, this study, as well the study by Cummings *et al.* (2001), proved otherwise and the sediment thickness variation between the two sub-basins therefore is not directly related to the magnitude of extension alone.

- Gunn *et al.* (1997a) concluded that a major magnetic intrusion (Mesozoic igneous intrusion) underlies the main basin depocentre underneath the Bass area resulting from magma generation caused by mantle depressurisation associated with basin extension and crustal thinning beneath the basin. This could be the reason why the area beneath the Bass area was hotter and hence the thermal decay stage of the whole area caused a greater thickness of post-rift sediments to be deposited in the Bass area compared to the Durroon area.

- Volcanic activity as late as Oligocene and Miocene suggests, however, that the temperature of the Bass area remained high during the time of thick post-rift sedimentation. It suggests therefore, that perhaps apart from cooling, the effect of sediment loading and isostatic subsidence of a weaker crust in the Bass area played a very important role in accumulation of thicker sediments.

- The contrasting structural styles in the Durroon and Bass areas and the different amounts of extension undergone by the two areas, suggest that the Durroon and Bass areas be referred to as two separate sub-basins of the Bass Basin instead of current proposal in recent literature for two separate basins. It is newly proposed to use Durroon Sub-basin and Burnie Sub-basin respectively for the two areas. Setiawan (2000) had proposed Northern Bass Sub-basin to the Bass area and Durroon Sub-basin to the Durroon area. In order to keep the distinctiveness for the Bass area and the Durroon area from the generic name of the Bass Basin, the new name Burnie Sub-basin (after the important town in the northern coast of Tasmania) is considered to be more appropriate for the Bass area and Durroon Sub-basin for the Durroon area.

- The structural style in the Durroon Sub-basin has been shaped through three major extensional phases that are reflected in different fault trends. The initial response to extensional forces was the development of discrete fault segments across the basin. The present day fault pattern subsequently developed through progressive propagation and linkages of the initial fault segments. The obliquity of Tasman rifting to the earlier trend had a profound influence on the structural development. The offshore extension of the NE-trending Arthur Lineament probably acted as a buttress to limit the tensional stresses, resulting in the Late Cretaceous history in the Burnie and Durroon Sub-basins being very different. However, from the Early Tertiary, the two sub-basins became linked. The development in common of the latter stages of their history further strengthens the argument for the sub-basin nomenclature.

- All the studies, including this one, suffer from very poor data coverage in the intervening area between the two sub-basins corresponding to the proposed Pelican-Squid transfer zone (Setiawan, 2000; Das and Lemon, 1999) along the Arthur Lineament. There is not a good network of long regional seismic lines over this intervening area that could help in deciphering the structural picture in a more realistic manner. The existing seismic lines are short, prospect level, survey lines and are inherently not well suited for very deep imaging. The data beyond 2.5-3 seconds in the entire basin and particularly in the transition zone between the two sub-basins is of very poor quality, which therefore makes seismic interpretation very subjective.

- A simple shear-pure shear model of extension has been invoked to explain the major westerly-dipping normal faults seen in the Moore Basin west of the Lord Howe Rise (Stagg *et al.*, 1999) and the ENE-dipping basin-bounding faults in the Durroon Sub-basin. These two areas perhaps formed initially as conjugate pair in the southeastern Australian Continent during the rift valley development prior to the opening of the Tasman Sea. Otway rifting had already created major weakness in the crust in the area and superimposition of later Tasman Sea rift tectonics resulted in an apparently clean, sharp split in the continental crust seen in the vicinity of the shelf and slope area of the SE Australian margin.

- The oldest rocks encountered in the Burnie Sub-basin, are seen in the Bass-3 well where the basement was encountered at a depth of 2350m. Unlike in the Durroon Sub-basin, the clear-cut identification of different syn-rift sequences is extremely difficult in the Burnie Sub-basin, masked by the large burial depth and poor quality of seismic data. There is evidence of erosional truncation of the Otway Group strata in the crestal portion of the footwall slope within the Yolla Trough. The Otway Group strata show parallel to sub-parallel internal reflection geometry which are parallel to the basement, suggesting that the Otway strata predate the Tasman rifting episode. This is well shown in the Durroon Sub-basin.

- The reflection surface corresponding to the Cretaceous Top shows a very clear truncation of tilted strata at this level which proves there is missing section at this boundary. The available biostratigraphic information however, is confusing, as no such break has been noted in the relevant reports. Revision of palynology might therefore clarify this problem.

- By the T. *lilliei* palynozone, the Durroon area reached the early post-rift stage with relative quiescence from major tectonic activity. In the Bass area, however, it is clear that there is tilting and erosion of some strata indicating thereby that the tectonic activity continued to the Maastrichtian. It is proposed that rapid sedimentation along initial extensional fault-bounded depocentres was superseded by sediment load-induced subsidence. Renewed fault activity tilted strata and possibly brought about subaerial erosion. The other possibility is that the by end of Late Cretaceous / Early Paleocene (Maastrichtian age), the whole region including these basin depocentres (around Aroo-1 well) were peneplaned following uplift and denudation. There is evidence of regional Palaeocene / Maastrichtian peneplanation (Hill *et al.,* 1995; Wilcox *et al.*, (1992) and Duff *et al.*, 1991) of the Late Cretaceous sequences overlain by a basal Tertiary sequence of coastal plain to nearshore barrier systems (Bodard *et al.*, 1985). These authors have suggested that the regional peneplanation may have been associated with Late Cretaceous-Palaeocene compressional deformation recorded in the Otway Basin, west of the Sorrel Fault (Hall, 1994).

- There is indication of thickening of the strata against the main bounding fault within the Otway and syn-rift sequences in the Cormorant Trough. Although the angular relationship between the Otway sequence and syn-rift sequence strata is not discernible, there is however, clearly some marked difference in the frequency content and reflector continuity in the two sequences within the Cormorant Trough. Setiawan (2000) has interpreted the absence of Otway Group in the Cormorant Trough while Williamson *et al.*, (1987) have interpreted the presence of Otway Group in the same area. It is believed that, although the sedimentation style and fault movement during the Otway period might have been different for the Yolla and Cormorant Troughs, time equivalent stratigraphic intervals may be present. Actual well data, when available, would confirm these observations.

- The most distinguishable difference between the syn-rift sequence strata in the Yolla and Cormorant troughs is that while the Yolla Trough shows some indication of either delta front build up or shore line movement, there is no such evidence in the Cormorant Trough. The delta front progrades towards the east with the bottomset downlapping or terminating against the bounding fault of the Yolla Trough. This difference in the reflection characteristics is indicative of the important role the Aroo horst played in controlling the sedimentary architecture within the two adjacent troughs. The initial extensional history in the two troughs has been different. The half-graben of the Yolla Trough shows substantial tilting of the Otway strata indicating rotational motion along the bounding fault with large displacement of 2-3 km creating an extensive depression in which fluvio-deltaic sedimentation occurred. Sediment source was from the west. The major extension occurred in a very short time within the Tasman Sea rifting episode. The Otway Group does not show any major thickening towards the bounding fault although there is a small increase in thickness downdip. Either the bounding fault existed during Otway period and was reactivated with large displacement or it was initiated only during the Tasman Sea rifting episode.

- The different seismic facies observed within the Yolla and Cormorant troughs suggest that sediment source/composition and sediment transport direction might be different in the two depocentres apart from the fact the timing or style of fault movement has also been distinctly different. It is suggested from the reflection geometry in the two troughs, however, that rapid sedimentation into the troughs created during initial extension loaded the crust sufficiently to allow isostatic subsidence to become dominant force in creating further accommodation space along with syn-sedimentary fault growth. That means the predominant force for accommodation space development in the Bass area was sediment load and syn-sedimentary fault growth unlike in the Durroon area where passive infill of the holes created by large fault displacement in the tilted basement fault blocks built the sedimentary architecture in that area.

- The major fault close to the western side of the Bass-3 horst did not record any movement during the Otway and Tasman rifting episodes except perhaps towards end of the Late Cretaceous period. This means the movement along this F0 fault is not equivalent to the age of movement along the two other major extensional normal faults, Faults F2 and F3, which control sedimentation in the Yolla and Cormorant Troughs. This is in contrast to what happened in the Durroon Sub-basin where all the major block-bounding faults probably moved in unision.

- The faulting style during the early post-rift history and later thermal cooling stage is marked by large number of faults with relatively less displacement caused by syn-sedimentary fault growth and the effects of burial and compaction. The major block-bounding faults (F2 and F3) are seen to have reactivated at quite a late stage, i.e., up to Late Eocene. This reactivation is again interpreted to be due in part to sediment loading and or compaction. The sediments in the depocentres separated by the Aroo horst are likely to have different lithofacies in equivalent time interval. The rift onset unconformity is the Otway top unconformity and corresponds to the beginning of Tasman rifting episode similar to the Durroon Sub-basin. The fact that Smith (1986) interpreted Durroon Formation to be restricted to the Durroon Sub-basin only makes some sense only when considered in terms of the sediment character rather than the age equivalence. It is believed that Tasman rifting caused enormous sedimentation in these two troughs but exact age correlation can only be established when deeper drilling information is available.

- From the description of the seismic stratigraphy, it is safe to conclude that the hydrocarbon play types in the various depocentres in the Bass Basin would be controlled by separate petroleum systems because the source-rich sediments in the deeper water lacustrine deposits would be distinct from each other. However, this can be corroborated only with further study.

- By Late Eocene time, the basin had entered the passive subsidence phase with the reflectors showing regional extent and covering the entire breadth of the basin. Similar effect is seen in the Durroon Sub-basin.

7.4 Inversion in the Bass Basin

- The Cormorant structure in the Burnie Sub-basin is a classical inverted feature. The large anticlinal Cormorant structure was formed during Late Oligocene-Early Miocene period. Till the Late Eocene beginning from Upper Cretaceous, the sedimentary thickness increases towards the Cormorant area, from south to north. It was during the Late Oligocene-Early Miocene period, when the Australian Plate collided with the Indonesian archipelago to its north, the resulting compressional stresses propagated across the continent to have a profound effect in the Burnie Sub-basin inverting the pre-Oligocene Cormorant Trough into its present day anticlinal structure. There were other continental collisions and realignments at this time, which were also close and could have a great influence on this area. E-W compression has been observed in New Zealand, the Gippsland Basin, the Eastern Otway Basin and the St Vincent Gulf Basin around Adelaide. It could be that near E-W compression caused some wrenching along the southern edges of Australia to build this structure (Young *et al.*, 1991; Davidson, 1980). The effect of this inversion is also seen in the faults which acted as extensional normal faults during deposition of Cretaceous-Late Eocene sediments but clearly show a reverse sense of throw towards the upper tips. There is clear evidence from the reflection seismic data of thinning of strata and possible erosion at the Late-Oligocene – Miocene level.

- Although inversion structures have earlier been identified from seismic data in the case of Cormorant structure (Robinson, 1974; Etheridge *et al.* 1985), this study has identified for the first time seismic evidence of inversion structures in the Durroon Sub-basin along some seismic lines both to the extreme east (BMR88-306) and in the southwestern part of the Durroon Sub-basin (BMR88-307). Some of the pre-existing normal faults have been subjected to compressional reactivation during the late Paleocene.

- Basin inversion has previously been inferred from thermochronology data and erosion of strata on the crests of tilted fault blocks in the Durroon area (Hill *et al.*, 1995). It is suggested that the inversion inferred from erosion of strata on uplifted footwall blocks, as seen near Durroon-1 well, should be concluded with caution because it can be explained by simple extensional model of hangingwall subsidence and footwall uplift related to the thermo-mechanical behaviour of crust and subsequent erosion of the emergent crests of the fault blocks.

- In the Bass area, the Cretaceous-Paleocene boundary (Maastrichtian unconformity) shows the effect of peneplanation of Cretaceous strata which could be the result of regional scale uplift and denudation. Evidence of such inversion in the Otway and Gippsalnd Basins has been reported in literature.

7.5 Volumetric contraction leading to extensional faulting

- A new type of layer-bound fault system was identified on 2D seismic data in most parts of the basin. Details of this system have been drawn from a 3D seismic data set over the Yolla Field. This fault system in map view is composed of almost randomly oriented, high density, minor extensional faults organised in a polygonal network. The component faults are typically 500m-1000m long and have throws ranging from 10m to 40m. The average fault spacing ranges from 60m to 500m. In seismic sections, there are at least three units or stratigraphic tiers observed within the deformed interval showing variations in the reflection characteristics and fault pattern development. The polygonal network of the fault system shows a near equal distribution of fault strike orientations, suggesting an isotropic stress regime during deformation of each unit within the deformed interval.

- The polygonal fault system deforms the very fine-grained Late Tertiary calcareous clay and marl-dominated succession of the basal Torquay Group throughout almost the entire offshore Bass Basin. The seismic expression of this pervasively-deformed unit suggests that there is no displacement transfer to the basement structures and the stratigraphy overlying and underlying this sequence is undisturbed, characterised by continuous reflection patterns.

- The development of polygonal fault systems by three-dimensional volumetric contraction of muddy sediments during early burial was first reported in literature from the central North Sea Basin (Cartwright, 1996). Layer-parallel volumetric contraction measured in seismic sections from the North Sea has recently been attributed to *syneresis* of colloidal smectitic gels during the early compaction of sediments (Dewhurst *et al.*, 1999). Syneresis results in the spontaneous contraction of a sedimentary gel without evaporation of constituent pore fluid. This process is driven by inter-particle attractive forces in marine clays and is governed by the change of gel permeability and viscosity with progressive compaction. A similar process is attributed to the seismically-observed complex polygonal fault system in the Bass Basin.

7.6 Hydrocarbon potential

- Detection of direct reflection from flat fluid contacts unconformable with the surrounding rock reflections to indicate gas-filled reservoirs is now well established in the geophysical literature. Two new 'flat spot anomalies', one in the Durroon Sub-basin and the other in the Burnie Sub-basin have been identified. The 'flat spot anomaly' in the Burnie Sub-basin, near Koorkah-1 and Seal-1 corresponds to a possible gas-liquid contact within the reservoir sands of the M. *diversus* zone of the early Tertiary Eastern View Coal Measures. The structure has been studied in good detail and shows an elongated anticline oriented NE-SW, with faulted margins to both the north and south. Velocity pull down, clearly seen on the dip line over the structure suggests a lowering of seismic velocity by a gas-charged reservoir.

- The exploration wells, Koorkah-1 (about 18 km to the southeast) and Seal-1 (about 20 km to the northwest of the mapped anomaly) both contain good reservoir sands in the *M. diversus* palynological zone, the same stratigraphic level as the anomaly. Fair quality source rocks in the shales, siltstones, claystones and coals of the Eastern View Coal Measures have attained adequate thermal maturity to charge the reservoir. Intraformational shales within the Eastern View Coal Measures have proven to be effective seals in other hydrocarbon accumulations in the basin. Although the two nearest wells were ultimately abandoned as dry wells, traces of gas and insitu hydrocarbons were found in the sands at the *M. diversus* level in Seal-1 and there were traces of free oil in the Palaeocene section in Koorkah-1. These factors together make the prospectivity of the structure good. The other 'flat spot anomaly' identified in the Durroon Sub-basin has not been detailed due to paucity of both seismic and nearby well data.

- The Bass Basin holds very good potential for stratigraphic hydrocarbon play development. There are several possibilities in which the interdigitation of syn-rift sandstone-rich fluvio-deltaic reservoirs with source-rich deep water lacustrine facies in the footwall and hangingwall slopes of tilted basement fault blocks might hold good hydrocarbon traps.

7.7 Recommendation for future work

- The exact definition of the 'Tasman Sea rift onset unconformity' has been established from seismic data in conjunction with well data at only one location near the Durroon-1 well. There has been no actual subsurface information from drilling data in any other part of the basin to support the findings from this study, particularly with regard to the deeper stratigraphy. There is a gross lack of factual subsurface information in the basin from the Lower Cretaceous Mesozoic section. This serious shortcoming can be overcome only by drilling a very deep well at the deepest part of the basin to collect all requisite subsurface information and enhance confidence in the lithostratigraphic, chronostratigraphic and biostratigraphic interpretations in the basin carried out so far.

- Since the data quality in general is very poor below 2.5-3 seconds TWT, reprocessing of the available 2D seismic data for enhancement of reflector continuity and better imaging of the subsurface from the deeper part of the stratigraphy could refine some of the observations in terms of exact correlation and mapping. The poor imaging of seismic data from deeper parts has forced subjective and interpretative correlations at many places, which could be improved through better processing routines.

- It is proposed here that any future study should then attempt construction of regional basin scale paleogeographic maps to study the sediment distribution patterns when the horizons can be confidently picked at deeper levels in the Burnie Sub-basin, particularly in the Pelican Trough. This should resolve the major discrepancy between previous publications (Etheridge *et al.*, 1985; Willliamson *et al.*, 1987) and some unpublished work (Setiawan, 2000). This discrepancy, made more apparent by the use of much of the available seismic data in this study, is thought to have arisen because the previous workers tried either to force interpretation of some of the poor quality data or used large spacing of seismic lines.

- The revision of palynology at just one well Durroon-1 (Morgan, 1991) has made an enormous impact on the understanding of the tectonostratigraphic evolution of the basin and it is believed that rigorous palynological revision in the available samples in several key wells and new drilled data can help in refinement of many of the conclusions so far.

- A few close grid 2D lines should be shot above the Pelican-Squid transfer zone, the transitional area between the Durroon and Burnie Sub-basins, to better clarify the role of this important zone in the rifting history of the basin. The lines should be long regional lines with an aim to extract very deep information.

- Although a new type of layer-bound polygonal fault system has been identified in the Lower Tertiary Torquay Group and studied in some detail from seismic data, significant work remains to be done on the exact nature of variation in the intensity of faulting and its style. The proposed work should concentrate on the response to different lithofacies variation, clay mineralogy, depositional setting and grainsize variation from detailed XRD and SEM work on either core or cutting samples from the deformed interval in the basin.

- The basin holds good hydrocarbon potential and the two 'flat spot anomalies' identified in this study need further investigation through AVO and subsequent drilling to test the prospectivity of the DHIs. There is good scope of untested stratigraphic play types in the basin and need to be followed up through drilling to better realise the untapped hydrocarbon potential in the basin.

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APPENDICES

1. Transfer zones in extensional basins and their control on structural style and stratigraphy-implications for hydrocarbon exploration, P. K .Das and N. Lemon, APPEA Journal, 39, p727-728, Poster Paper presented at the 1999 APPEA Conference, Perth (Abstract included).

2. A new model for the tectonic evolution of the southeastern Bass Basin, Australia, APPEA Journal, 40, p775-776, Poster Paper presented at the 2000 APPEA Conference, Brisbane (Abstract included).

3. Flat spot anomaly: Possible gas sands in the NW Bass Basin ?, Pradipta Kumar Das and Nick Lemon, PESA Journal, 28,2000, p51-57 (Reprint included).

*4. Seismic expression of layer-bound polygonal fault system in the Late Tertiary of the Bass Basin: indications of syneresis of colloidal sediments ?, P. K. Das, A. Cummings and N. Lemon, Paper presented at the SGTSG Conference at Ulverstone, Tasmania, 2001, Geological Society of Australia, Abstracts, Number 64, p32-33. (***Joint winner of the Best Student Paper Award)** (Abstract included and the copy of the letter from the conference organisers included)

5. Seismic expression of layer-bound polygonal fault system in the Late Tertiary of the Bass Basin: A Case Study, P. K. Das, A. Cummings and N. M. Lemon, March 2001, Paper submitted to ASEG Journal for publication (Abstract included).

P.K. Das and N. Lemon (1999) Transfer zones in extensional basins and their control on structural style and stratigraphy-implications for hydrocarbon exploration.
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P.K. Das and N. Lemon (2000) A new model for the tectonic evolution of the southeastern Bass Basin, Australia.
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P.K. Das A. Cummings and N. Lemon (2001) Seismic expression of layer-bound fault system in the Late Tertiary of the Bass Basin: indications of syneresis of colloidal sediments.
Geological Society of Australia, Abstracts, no. 64, pp. 32-33, 2001

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Dear Mr Das,

Congratulations! You were selected as the joint-winner of the oral presentation student prize at the recent SGTSG conference in Ulverstone. The cash prize was split with the other winner, Nathan Daszko of New Zealand. The committee of Dave Gray, Garry Davidson and Peter Cawood were impressed with your enthusiasm, scientific merit, and excellence in presentation. Good luck in your future endeavours.

Regards

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ASEG Journal, submitted, March 2001

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