The structural history of Tasmania: a review for petroleum explorers

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Abstract

The Mesoproterozoic basement of Tasmania is overlain by a Neoproterozoic passive margin sequence. Deposition was interrupted by a major Cambrian arc-continent collision. Ordovician to early Devonian, shallow-water sedimentation dominated across the Western Tasmanian Terrane, while a turbidite succession built rapidly across eastern Tasmania. Most sequences were metamorphosed to an overmature state (greater than or equal to 300ºC) during a Devonian orogeny.

In the latest Carboniferous, deposition began in the Tasmania Basin. The Late Carboniferous-Permian Lower Parmeener Supergroup consists of glaciomarine sediments, and is unconformably overlain by Triassic non-marine sandstones and coal measures of the Upper Parmeener Supergroup. A large volume of tholeiitic dolerite intruded the Tasmania basin during the Middle Jurassic. The main body of the Tasmania Basin reached maturity in the Mesozoic and there is evidence for widespread hydrocarbon generation. Locally, evidence for hydrocarbon migration has been detected but no hydrocarbon accumulations have been found.

The Sorell, Bass and Durroon basins were initiated in the latest Jurassic–Early Cretaceous by extension related to rifting between Australia and Antarctica. By the Late Cretaceous active spreading had begun in the Tasman Sea. The Bass Basin continued to propagate southwards, extending onshore with the opening of the Tamar Graben in the latest Cretaceous, the Devonport-Port Sorell Sub-basin in the Early Paleocene, and the Longford Sub-basin in the Late Paleocene. On the west coast, the Sorell Basin extended onshore with the development of the Macquarie Harbour Graben in the Late Paleocene. During the Late Paleocene–Early Eocene, Tasmania was moving north along a left-lateral transform against Northern Victoria Land. In the Eocene, Australian-Antarctic motion became more divergent along this margin and the most active extension migrated to the southern rift basins. Only the Bass Basin has yielded economic accumulations of gas. Minor hydrocarbon shows have been encountered in the Sorell Basin. The younger southern basins have not accumulated sufficient sediment to reach maturity.

Keywords: Tasmania, structure, tectonics, basin

Introduction

The aim of this paper is to provide a broad overview of the structural history of Tasmania. The history is complex (Fig. 1) and to meet this aim in a single paper requires that the direct supporting data cannot be included. The reader is referred to (Burrett & Martin 1989) for a pre-1989 view of the geological history. Major changes in the interpretation of the structural history began with the 'Contentious Issues' mini conference in Hobart (Cooke & Kitto 1994). More recent influential papers report modern radiometric dating (especially Black et al. 1997 and Turner et al. 1998), stratigraphic re-interpretations (e.g. Calver & Walter 2000), and structural reappraisal of Proterozoic and Cambrian history (Meffre et al. 2000; Holm & Berry 2002). Most advances have been in the emerging view of the Cambrian and Neoproterozoic history of Tasmania and its place in Gondwana and Rodinia. The most recent influential research into the post-Devonian history are the new seismic surveys offshore (Drummond et al. 2000) and onshore. The full impact of this work has yet to be realised.

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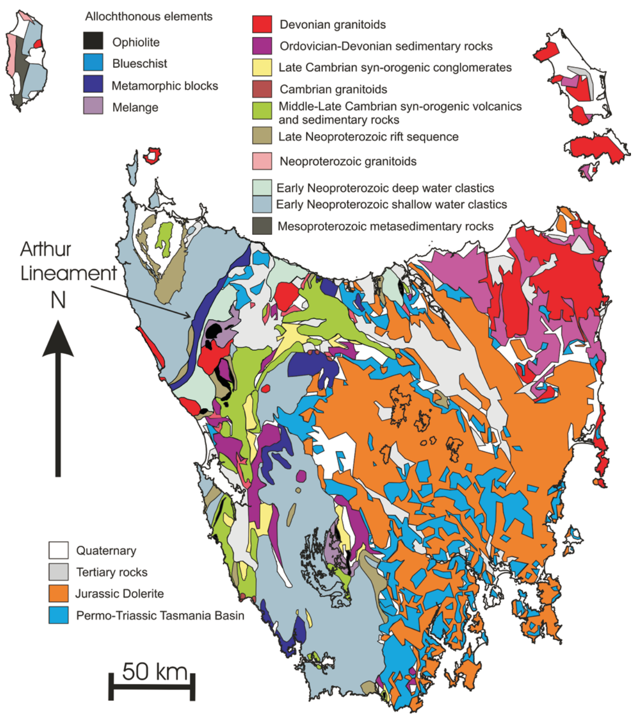

Figure 1. Geological map of Tasmania showing the distribution of major rock types. Simplified and modified from 1:250 000 digital geological map of Tasmania (Brown et al. 1995).

Proterozoic history

New dating on western King Island indicates a metamorphic event at 1,270 Ma (Berry & Burrett 2002) in siliciclastic rocks less than 1,350 Ma old (Black et al. 2004a). This event is very restricted in distribution and the significance is still debatable. The "Western Coast metasediments" of King Island may be a shallow part of the basement. The metamorphism on King Island is similar in age to the early stages of the Grenville Orogeny in the eastern Albany Fraser Province and the Musgrave Ranges (White et al. 1999). Berry and Burrett (2002) argued this age provides evidence that Tasmania was a microcontinental fragment rifted from north of Broken Hill during the breakup of Rodinia. However, zircon inheritance in granites demonstrates that the basement of the Western Tasmania Terrane is 1,600 Ma old (Black et al. 2004b). A possible scenario is shown in Figure 2a with the NW Tasmanian crust having a thin 1,270 Ma upper crust over a Palaeoproterozoic lower crust. The Eastern Tasmanian Terrane did not exist at this stage with Black et al. (2004a) identifying the lower crust as less than 1,000 Ma. Alternatively, the western King Island section may be an allochthonous element with no relationship to the larger scale history of Tasmania.

The early Neoproterozoic stratigraphy of northwest Tasmania (Fig. 2a and Fig. 3) is dominated by shallow water siliciclastics (Rocky Cape Group in the north west, Clark Group in the south) and turbidites (Burnie and Oonah Formations, Badger Head Block in the north) (Seymour & Calver 1995). The age of these units is poorly constrained. All the units have similar detrital zircon patterns (Black et al. 1997) and the youngest zircon detected is 1,000 Ma. Calver and Walter (2000) argued that parts of the unconformably overlying succession are 750 Ma. The Burnie Formation contains dolerite sills (Cooee Dolerite) that have a 725 ±35 Ma minimum age (Crook 1979). This sequence is probably older than the Wickham Orogeny (Fig. 3).

The internal stratigraphy of the early Neoproterozoic siliciclastic sequences varies dramatically. In the NW there are three basic facies varying from deep shelf though storm dominated shelf to massive quartzites deposited above fair weather wave base. The Clark Group includes similar shallow marine orthoquartzite units but has more carbonate units (Seymour & Calver 1995). The deep-water facies (Burnie and Oonah Formation) shallow up from classical turbidites to storm-affected siliciclastics and carbonates. The stratigraphic relationship between these units remains contentious but a possible scenario is shown in Figure 2a where the deep-water facies are to the NE. Very little work has been done on the petroleum potential of these sequences. The shallow water units are low in carbon. Mudstones in the turbidite units are more graphitic. All the sections have low porosity. The rocks are generally referred to as low greenschist facies and the very meagre data available supports this statement but the regional variation in metamorphic grade is not known. No evidence of petroleum migration has been reported.

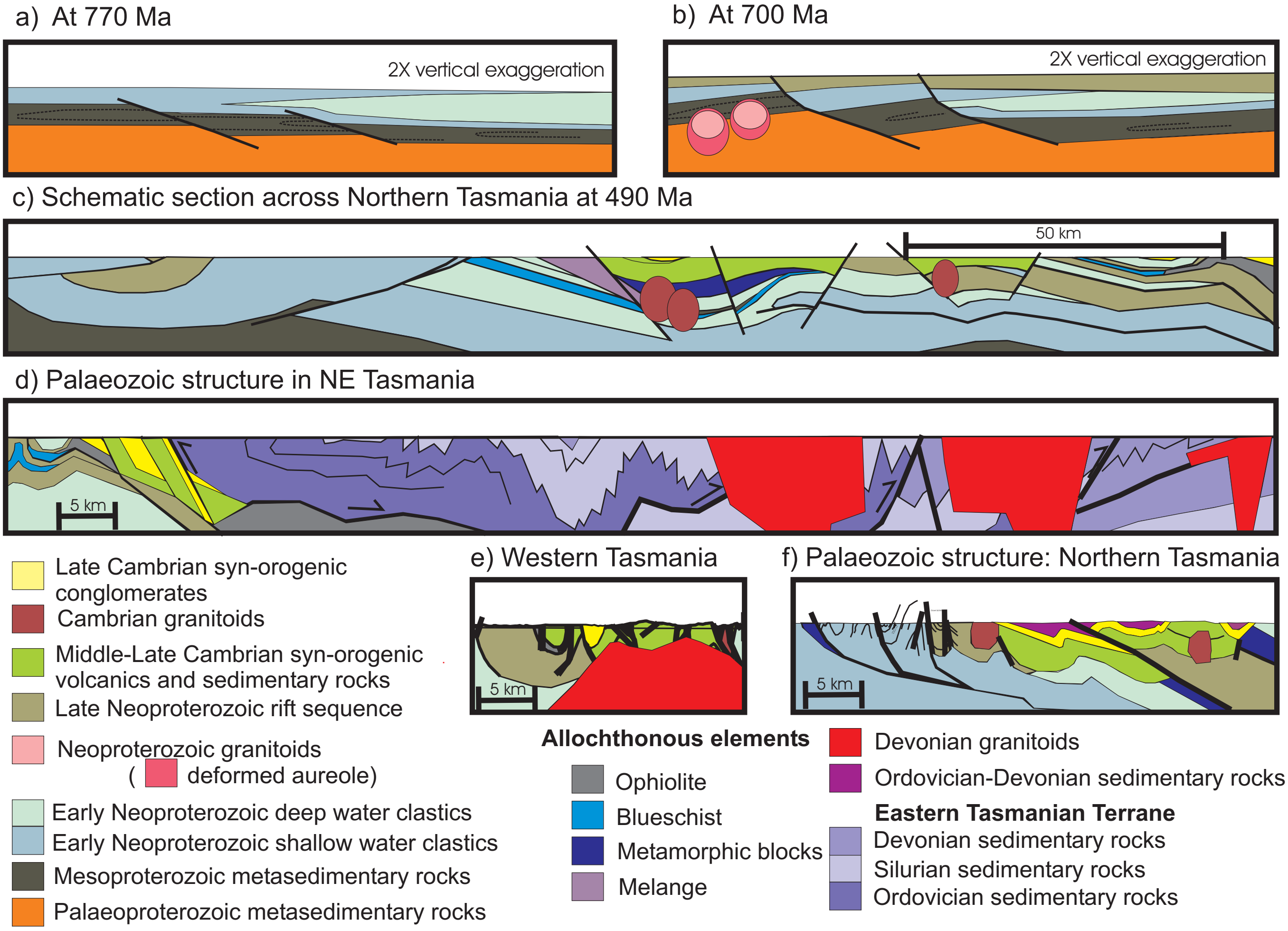

Figure 2. Schematic cross-sections showing the style of early structure in Tasmania: **a)** Early Neoproterozoic basin development over a Palaeoproterozoic basement. **b)** Late Neoproterozoic rifting (approximately 740 Ma). **c)** A Late Cambrian section across Tasmania with the affects of Devonian deformation removed. **d)** Eastern Tasmanian Terrane. (Section modified from Patison et al 2001). **e)** Dundas Trough. (Section modified from Berry 1999). **f)** The northern margin of the Tyennan block (section partly after Woodward et al 1993, McClenaghan et al 2001 and Berry & Bull 2004). Section locations are shown on Fig. 5.

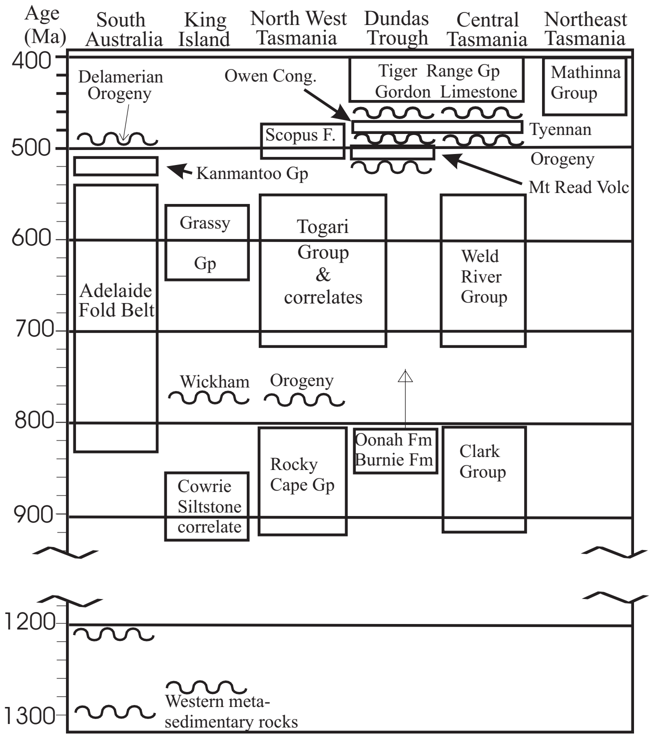

Figure 3. Space-time diagram for the Western Tasmanian Terrane up to the Middle Devonian (modified from Calver & Walter 2000)

A second extensional phase followed in the late Neoproterozoic, and possibly started as early as 740 Ma (Adams et al. 1985; Calver & Walter 2000). The base of the late Neoproterozoic section is a regional-scale low angle (20˚) unconformity (Calver 1998). A more intense deformation (Wickham Orogeny) is known from King Island where there was polyphase deformation and extensive granitoid intrusion at about 760 Ma (Cox 1989; Turner et al. 1998). The granites have intensely deformed aureoles (Fig. 2b), but are not linked to any larger scale deformation features. Holm et al. (2003) argued the King Island granitoids are related to rifting and the breakup of Rodinia. The Wickham Orogeny may correlate with the low angle unconformity observed on mainland Tasmania.

The late Neoproterozoic rift sequence (e.g. Togari Group, Fig. 3) has an initial siliciclastic shallow water section, followed by a lower dolomite and diamictite. Subsequent widespread intrusion of tholeiitic dolerite dykes (Rocky Cape dyke swarm), extrusion of tholeiitic basalts and associated volcanogenic sediments were followed by a second dolomite section and finally marine siliciclastics (Calver & Walter 2000; Brown 1989; Turner 1989; Crawford & Berry 1992). The most complete example of this section is in NW Tasmania and elsewhere only parts of the section are known (Seymour & Calver 1995; Calver 1998). Calver and Walter (2000) correlated the early dolomite with the Sturtian glaciation and the later dolomite with the Marinoan glaciation. The late Neoproterozoic sequence is interpreted as a rift-drift facies associated with the breakup of Rodinia (Crawford & Berry 1992).

The Neoproterozoic sequence in Tasmania has largely been considered as “unmetamorphosed” but detailed petrographic descriptions of the basalts commonly include pumpellyite and to a lesser extent prehnite. These units are best regarded as very low-grade metamorphic rocks. The age of this metamorphism is problematic. No structural model developed so far explains the inferred depth of burial (approximately 10 km) of these rocks during either the Cambrian or the Devonian orogeny. Bacon et al. (2000) reported that the Neoproterozoic mudstones have very low organic C contents. No evidence of petroleum migration has been reported.

Palaeozoic events

The Tyennan Orogeny in Tasmania was a complex event with rapidly changing stress patterns. The locus and style of deformation changed very rapidly. At 520 Ma, Tasmania collided with an oceanic arc, and major slices of fore-arc lithologies were thrust over the Proterozoic sedimentary rocks (Fig. 4). The only allochthonous element specifically identified by Berry and Crawford (1988) was the mafic/ultramafic complexes. Other possible allochthonous elements identified since are the Forth, Badger Head, Port Davey, Franklin, Mersey River and the Arthur Metamorphic Complexes (Meffre et al. 2000), and the Wings Sandstone (Black et al. 2004a). These blocks are scattered across Tasmania lying structurally above the late Neoproterozoic rift facies and are unconformably overlain by late Middle Cambrian and younger sedimentary rocks. The Arthur Lineament (Fig. 1) was formed during the Tyennan Orogeny (Turner et al. 1998), but its tectonic significance remains in doubt (Turner 1989; Berry 1994). This lineament forms the western limit to allochthonous blocks in Tasmania and appears to mark the maximum extent of the thrust complex. The early part of the thrust emplacement is recorded in high temperature mylonites, and indicates thrusting towards the west (Berry 1989b, Fig. 4a). A major phase of thrusting to the south (Fig. 4a), recorded in the amphibolite facies metamorphic rocks and widespread cataclasites in western Tasmania (Findlay 1993; Findlay & Brown 1992; Berry et al. 1990; Holm & Berry 2002), is interpreted as the second stage of the obduction process.

The second stage of the Tyennan Orogeny was a Middle Cambrian extensional event that produced rapid subsidence, active syn-orogenic deposition and major post-collisional felsic-dominated volcanism (Mt Read Volcanics, Fig. 3). East–west extension is recorded in the hydrothermal vein geometry and the Henty dyke swarm. The extensional phase was closely followed, or may be coincident with, a N–S compressional event that produced E-W trending folds to the east of the Mt Read Volcanics (Fig. 4b).

In the Late Cambrian, the last phase of the Tyennan Orogeny inverted earlier extensional faults (e.g. Henty Fault). Major reverse faults and upright open N-trending folds were formed in western Tasmania. This phase also caused uplift of the Tyennan block with syn-orogenic sediments (Owen Conglomerate) accumulating in synclinal cores and other structural depressions.

Most of the folding in the Neoproterozoic stratigraphy in Tasmania is correlated with the Tyennan Orogeny. Two generations of early recumbent folding are recognised in allochthonous blocks and high strain underlying elements. These are overprinted by upright NS (e.g. NW Tasmania) or EW (e.g. Tyennan block) folding and related high angle reverse faults. The structure across northern Tasmania at the end of the Cambrian is shown in Figure 2c. West of the Arthur Lineament upright folding and thrusting dominated the Proterozoic. The allochthonous units are scattered across the rest of Tasmanian and overly a variably thrust-influenced Neoproterozoic section. Two middle Cambrian grabens are visible on this section. They are strongly influenced by volcanism and intruded by high level Cambrian granitoids. The basins were inverted in the Late Cambrian and syn-orogenic conglomerates accumulated in structural depressions.

Following the Late Cambrian basin inversion western Tasmania was peneplained and a new cycle of deposition began in the Middle Ordovician. This cycle has a thin basal sheet sand followed by widespread intertidal to shallow marine tropical carbonate (Gordon Group) that is overlain by a Silurian to Early Devonian shallow marine siliciclastic sequence (Tiger Range Group). At the same time, the Mathinna Group was deposited in NE Tasmania. The earliest unit, Stoney Head Sandstone, is a proximal turbidite and is overlain by Middle Ordovician deep-water mudstone unit. The Silurian and Early Devonian section is a highly variable turbidite section with rapid facies variations and containing evidence of arc volcanics in the source region.

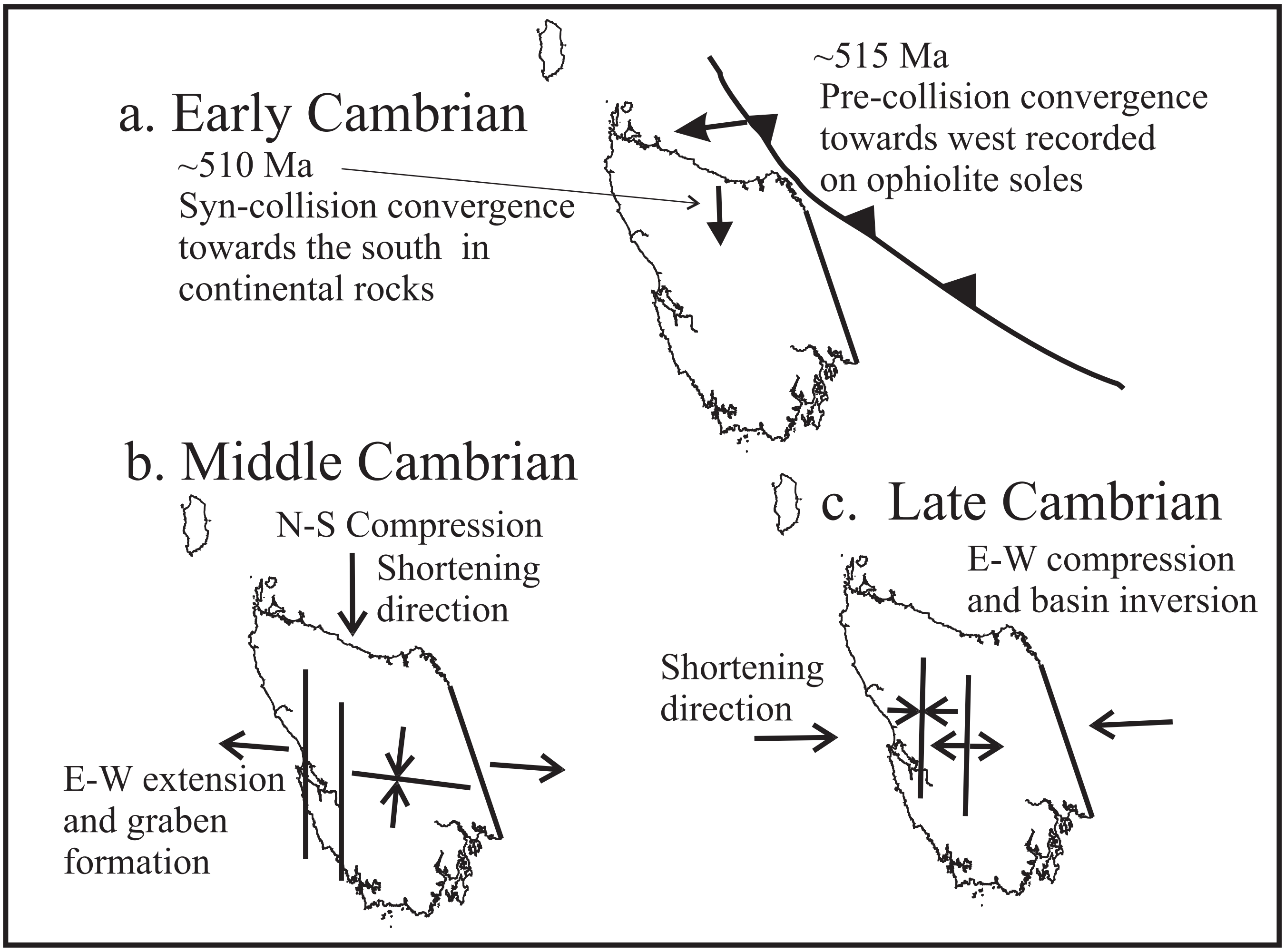

Figure 4. Pattern of Cambrian deformation, Tyennan Orogeny (after Holm & Berry 2001).

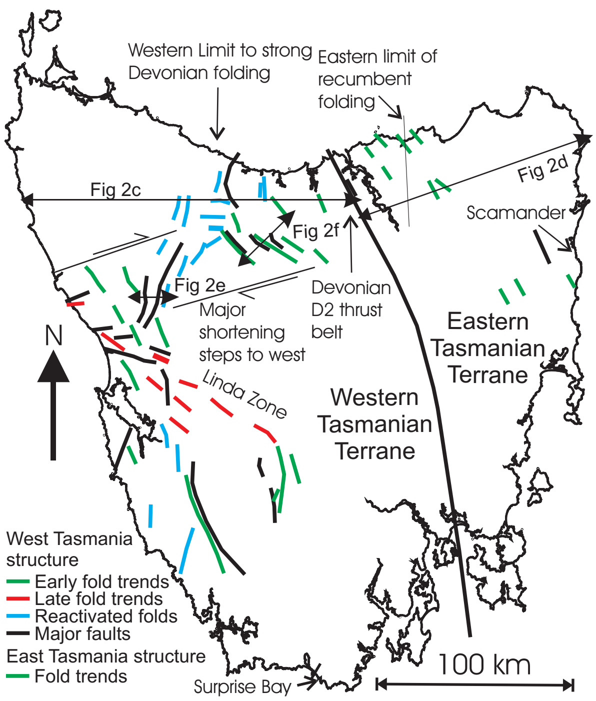

Figure 5. Devonian structural trends in Tasmania (modified from Williams 1989).

Middle Devonian deformation (390 Ma: Black et al. 2004b) occurred throughout Tasmania and is characterised by the complexity of fold orientations (Fig. 5), explained, in large part, by reactivation of older structures. In many areas, the fold geometry is controlled by the Cambrian fold trends, which were tightened during the Devonian. This led to Devonian cleavage orientations that are not parallel to the axial plane of the folds with which they are associated. In the north, E-W Cambrian folds were tightened. The Devonian structural style in the north is shown in Figure 2f. In the west, N-trending Cambrian folds are tightened with an associated NNW-striking Devonian cleavage. NNE-trending folds, north of Tullah, are controlled by the reactivation of the Henty Fault. This region acted as a transfer zone in the Devonian and the NNE fold trend is probably a rotated N trending Cambrian trend (Fig. 5). A cross-section of the Dundas Trough (Fig. 2e) shows the strong fault control on deformation and the influence of granitoids.

The subsequent N- to NNE-trending compression produced WNW-trending folds and thrusts in the south, and NW-striking thrusts and associated folds in the north (Fig. 5). On the west coast, there is an associated phase of brittle strike slip faults dominated by NNE-striking sinistral movement on the Henty Fault (Berry 1989a).

In NE Tasmania, the early phase of deformation thrust the passive margin east across the deep-water section with recumbent folds in the Georgetown area, and this was followed by back thrusting, especially strong in the Beaconsfield zone (Fig. 2d). A late stage of N trending compression produced strike slip movement on some faults and large-scale kinks (Goscombe et al. 1994).

The regional metamorphic grade associated with the Devonian orogeny is prehnite-pumpellyite with local zones of greenschist facies in the vicinity of late syn-to post-orogenic granites. Areas of low grade metamorphism occur in the Western Terrane at Surprise Bay (Burrett 1992) and in the Mathinna Group on the far east coast at Scamander (Patison et al. 2001) (Fig. 5). Only these areas are still within the oil and gas window.

Large-scale granitoid intrusion in NE Tasmania started before the Devonian deformation and continued after (400–380 Ma). The more restricted granite intrusions scattered throughout western Tasmania and King Island, post-date peak deformation (375–350 Ma: Black et al. 2004b).

Latest Carboniferous to Juriassic

Tasmania Basin

Following the Devonian deformation there was large-scale erosion. Deposition restarted in the late Carboniferous with 1.5 km of generally flat lying sedimentary rocks of Late Carboniferous to Late Triassic age deposited in the Tasmania Basin. The basin is unconformable on Late Devonian Granites and older folded rocks, and is subdivided into two broad lithological and environmental associations, the Lower and Upper Parmeener Supergroups. The Lower Parmeener Supergroup consists of glacial and glaciomarine sedimentary rocks, while the Upper Parmeener Supergroup comprises non-marine sedimentary rocks. Both units contain subordinate coal measures (Forsyth et al. 1974). The basin, covering an area greater than 30,000 km2 (Fig. 6), is best preserved in northern and southeastern Tasmania. However, the present basin limits are erosional, rather than depositional, and the basin was once considerably larger (Bacon et al. 2000).

The Tasmania Basin has been variously labelled as a sag basin (Veevers 1984), or a foreland basin (Collinson et al. 1987). The accumulation of 1.5 km of sediment in 100 million years is an order of magnitude slower than classic foreland basins (Schwab 1986). While a Mesozoic foreland basin developed to the east, the effect of continental margin loading was not significant this far inboard. There is evidence for minor normal faulting early in the basin history, but the limited nature of such faulting and the relatively slow sedimentary accumulation rate is not typical of rift basins. The early development of a marine basin and its subsequent infilling may relate to a eustatic sea-level rise rather than any tectonic events, or thermal subsidence. The history of sedimentary accumulation is more typical of a continental margin ("pericratonic”) basin.

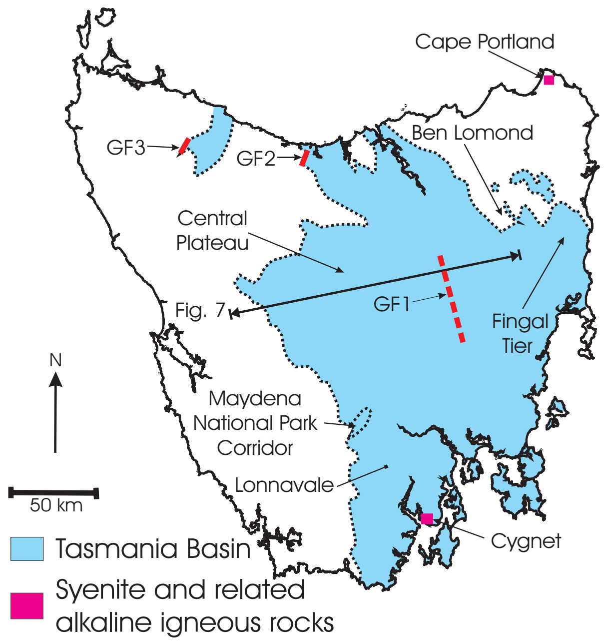

Figure 6. Current extent of the Tasmania Basin, locations of alkaline igneous intrusions and growth faults associated with early basin structure.

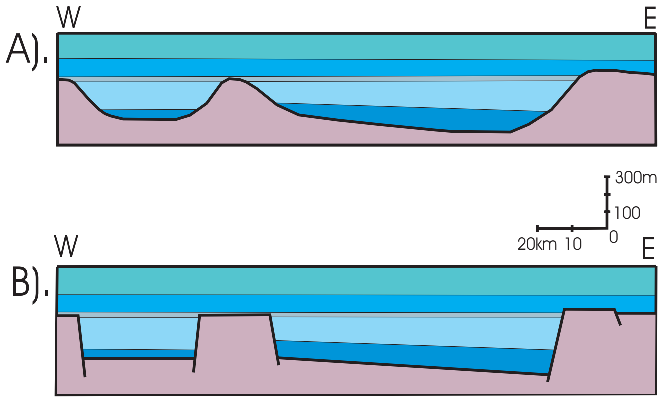

Figure 7. Two models for the initial development of the Tasmania Basin: A). Infilling of glacial valleys, B). Growth on a west-dipping fault (location of section shown on Fig. 6).

The variable development of lower glacigene beds, especially evident in the lowermost units (Wynyard Tillite and the Woody Island Siltstone), has generally been interpreted as the infilling of glacial valleys (Fig. 7a). The Wynyard Tillite and Woody Island Siltstone occupy an elongate depocentre through central and southern Tasmania, which rapidly thins to the east (Reid et el. 2003) (Fig. 6, GF1). The present data set could also be interpreted as the result of growth along a west dipping normal fault (Fig. 7b). Further evidence for structural control on the basins evolution can be seen in north and northwestern Tasmania. The Devonport Sub-basin is now bounded in the east by an east-dipping normal fault against the Forth Metamorphics (Everard et al. 1996) (Fig. 6, GF2), although this fault could be younger than the basin. In north-western Tasmania approximately 300 m of growth in the Wynyard Tillite can be seen on an east-dipping fault adjacent to the Arthur Lineament (Burns 1963) (Fig. 6, GF3), suggesting reactivation of the Arthur Lineament during the Late Carboniferous to Early Permian. No other growth faults have been observed in the Tasmania Basin either from outcrop mapping, drilling or from recent seismic data acquired across the Central Plateau and the Longford Sub-basin. Faulting was minor during the deposition of the Tasmania Basin and limited to the earliest stages of basin development.

Jurassic faults

There is evidence of faults in Tasmanian Basin rocks that were active before the mid-Jurassic intrusion of dolerite. By far the best example is the strong transpressional zone along the Maydena-National Park corridor (Fig. 6). This zone has the appearance of a dextral positive flower structure and is truncated at the eastern end by a major dolerite dyke (Dunster 1981). Leaman (1976) recognised a Jurassic phase of rifting, pre-to syn-dolerite, in the Hobart area but the evidence supporting this interpretation is scarce. In areas with less detailed work, no evidence has yet been found for mid-Jurassic rifting. The nature of late Triassic to mid-Jurassic deformation remains contentious.

Jurassic dolerite

By the early Jurassic the Parmeener Supergroup formed a shallow syncline, plunging towards the SSE, with possibly some gentle folding in an otherwise sub-horizontal succession (Hergt et al. 1989). Large volumes of tholeiitic dolerite intruded into the Tasmanian crust during the Middle Jurassic are probably related to a major thermal anomaly occurring along the eastern margin of Gondwana. The dolerite formed mainly as sills in the Tasmania Basin. The dolerite is exposed over an area of 30,000 km^2 and has an estimated volume of 15,000 km^3 (Hergt et al. 1989).

The dolerite has been a major deterrent to petroleum exploration in Tasmania, with nearly every part of the basin being intruded by at least one dolerite sill. Recently acquired seismic data across the Central Plateau and the Longford Sub-basin, demonstrates the variations in data quality associated with acquiring seismic through the dolerite. When at, or near, the surface, dolerite is generally highly diffusive resulting in poor resolution of underlying events (Fig. 8a). Whether this results from the effects of weathering, or the occurrence of boulders or remnant boulder fields in the soil profile is unclear. At depth, the dolerite is characterised by a strong positive event at its top and base and by weak and scattered events in between. Seismic events beneath the dolerite at depth are in general, better resolved (Fig. 8b).

Most dolerite intrusions have the form of a flattened cone connected to a source or sources at the deepest point, the limbs are concordant (sills) or approximately concordant with abrupt transgressions when rising to higher levels (Leaman 1976). The metamorphic effects resulting from dolerite intrusion are usually confined to within a few metres of the intrusion margin, the effect being more severe at the roof of the intrusions. In the Hobart area, two or three sills are commonly present. The sills are usually either less than 1 m, or 300 m to 400 m thick. The thick sills in middle or lower Permian rocks are typically 30 km^2 in area, while sills in Triassic rocks are more extensive (Leaman 1975). In contrast, only a single sill, intruding the Upper Parmeener Supergroup, has been recognised in the northern part of the basin (Central Plateau, Ben Lomond and the Fingal Tier, Fig. 6) (Bacon et al. 2000). A single, 650 m thick, sill was intersected near the Upper–Lower Parmeener Supergroup boundary in the Hunterston-1 DDH (Reid et al. 2003). From the interpreted seismic this sill appears to cover many hundreds of square kilometres.

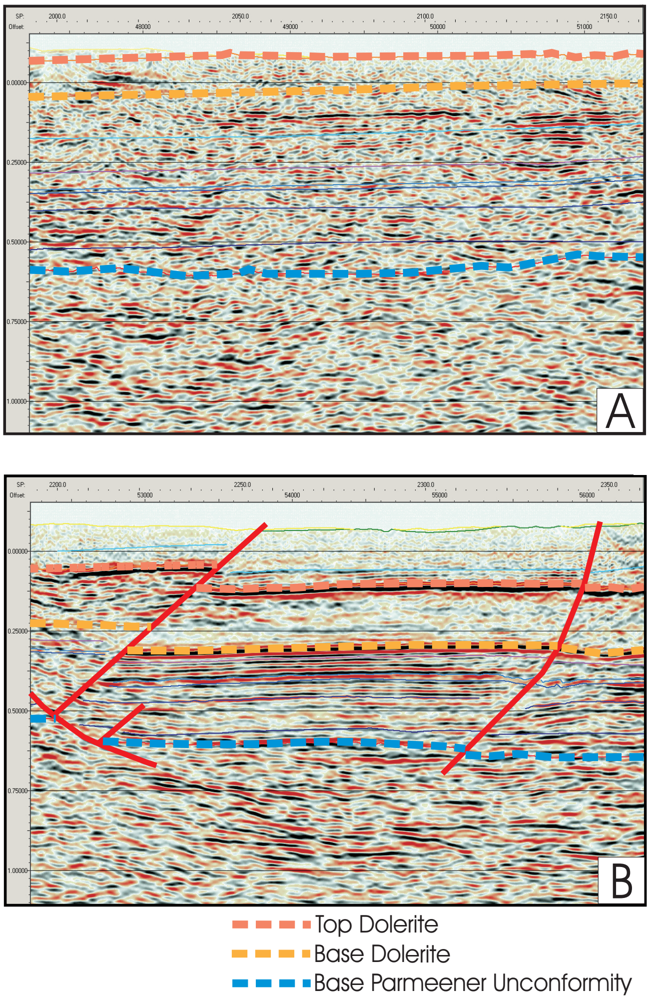

Figure 8. The effect of Jurassic Dolerite on seismic resolution: A) at/near surface (SP 2,000–2,150) and, B) at depth (SP 2,200–2,350).

Hydrocarbon potential of the Tasmania Basin

The Tasmania Basin has yet to be fully explored using modern exploration methods. Potential source, reservoir and seal rocks of the Gondwana Petroleum System are contained within the Lower Parmeener Supergroup and recent work indicates the main body of the Tasmania Basin is mature (Reid et al. 2003). Maturation and initial migration probably resulted from an elevated geothermal gradient in the Middle to Late Cretaceous, however subsequent restructuring has probably modified the trap geometries and complicated the recognition of migration pathways.

There have been many reports of oil and gas seeps in Tasmania; almost all have proved to be erroneous. In the Tasmania Basin, a single bitumen seep has been confirmed at Lonnavale, approximately 42 km WSW of Hobart (Fig. 6). The seep, located in Jurassic Dolerite close to a faulted contact with fossiliferous Permian mudstone, was derived from a *Tasmanites*-bearing oil shale (Revill 1996). While oil shales aren't known in southern Tasmania, disseminated *Tasmanites* have been found in Woody Island Siltstone at Maydena (Bacon et al. 2000). One theory argues that the heat from intruding dolerite was sufficient to generate hydrocarbons from an oil shale source, while migration was facilitated by fractures in the dolerite (Bacon et al. 2000).

Late Jurassic to Cretaceous events

Extension related to the rifting between Australia and Antarctica initiated the Bass, Durroon and Sorell Basins in the latest Jurassic to Early Cretaceous (Fig. 9). The direction of extension in Bass Strait is contentious, with seismic data acquired north of Tasmania revealing extensional features with NW–SE, NE–SW and N–S orientations (Muller et al. 2000).

The exploration results from these basins have been disappointing when compared to the neighbouring Gippsland and Otway Basins. Only two wells (Yolla-1 and Pelican-5) have flowed hydrocarbons in the Bass Basin (Morrison & Davidson 1989). While in the Sorell Basin four exploration wells have been drilled, with only Cape Sorell-1, in the Strahan Sub-basin, recording minor hydrocarbon shows (Baillie & Hudspeth 1989).

Extension in the Bass Basin resulted in the development of a series of NW–SE trending half-grabens with faults dipping to the SW, these faults are generally planar though listric and ramped faults have been found (Young et al. 1991; Etheridge et al. 1985; Hill et al. 1995). Etheridge et al. (1985) observed that these normal faults can terminate abruptly, and are displaced across transverse zones which span the basin, he argued that the transverse zones were not simple strike-slip faults but rather transfer faults similar to oceanic transforms, as they are accommodation structures that allow for variations in the geometry of extension. However, Williamson et al. (1985) interpreted these transfer zones as more complex structures. Gunn et al. (1997), using magnetic data, recognised three main compartments corresponding to the Etheridge et al. (1985) interpretation.

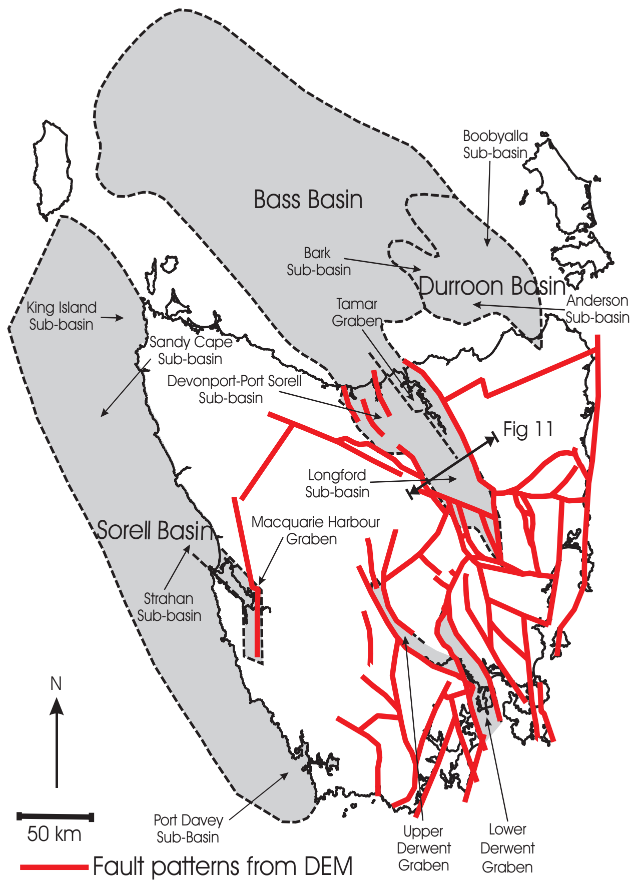

Figure 9. Late Jurassic to Middle Tertiary basins and fault patterns interpreted from the high resolution DEM.

The Durroon Basin comprises three sub-basins (Baillie & Pickering 1991), the Bark, Anderson and Boobyalla sub-basins (Fig. 9). The basin adjoins the southeast Bass Basin and its earliest development is related to the same extensional forces, but has a different structural history (Baillie & Pickering 1991). Deposition of Otway Group correlates probably began in the Early Cretaceous in a broad linear depression with fault bounded margins (Baillie & Pickering 1991).

Sediments of the Sorrell Basin cover the western continental margin of Tasmania (Willcox et al. 1989). The basin contains four depocentres, from north to south the King Island, Sandy Cape, Strahan and Port Davey Sub-basins (Fig. 9), each with over 4 km sediment thickness. These sub-basins were probably developed in the latest Jurassic to earliest Cretaceous (Hill et al. 1997), and have been interpreted as "relieving bends" on a major left-lateral, strike-slip fault, associated with extension along Australia's southern margin (Moore et al. 1992). The sub-basins are bounded along their eastern edges by strike-slip faults, except for the King Island Sub-basin that lies inboard of the major fault. They are separated by basement highs. Both the King Island and Sandy Cape Sub-basins were initiated in or before the Early Cretaceous, and contain Early Cretaceous Otway Group correlates (Moore et al. 1992). The King Island Sub-basin is a perched half-graben, apparently formed by extension (Moore et al. 1992), that deepens eastward to a high angle NW dipping normal fault (Willcox et al. 1989). The Strahan Sub-basin is a NNW–SSE trending slot comprising two EW trending half-graben, which shallow to the southwest from a northern depocentre (Moore et al. 1992). Moore et al. (1992) considered the basin boundary faults and the extensional faults bounding the tilt-blocks to be a linked fault system responsible for basin formation. Opinion remains divided on the timing of the earliest deposition in the Strahan Sub-basin, in spite of its close seismic coverage, and a single exploration well, Cape Sorell-1, that only just penetrates the upper Cretaceous. In Moore et al. (1992), the junior authors argue the basin was formed along a strike-slip zone during the Late Jurassic and was contemporaneous with the formation of the Eyre Sub-Basin and the Great Australian Bight Basin. In contrast, Moore (1991) argues that the basin is unlikely to contain any Lower Cretaceous section and with the Port Davey Sub-basin is the product of strike-slip and transform plate interaction in the Late Cretaceous to Palaeocene. The latter case is consistent with the age of basin formation decreasing southwards, and links their relative age of formation to the northward movement of Tasmania with respect to the Antarctic Plate (Moore et al. 1992).

There is little evidence to suggest that the Tasmania Basin or any onshore Tasmanian rocks have been affected by a Late Jurassic-Early Cretaceous rift phase. The distinction of this rifting phase from the Late Cretaceous and Tertiary extension remains a challenge for the future.

Middle to Late Cretaceous events

By the mid-Cretaceous, rifting along the southern margin of Australia had given way to spreading. In the Bass Basin the event is recorded as an angular unconformity at the top of the Otway Group, extension continued during this period with reduced movement on normal faults (Young et al. 1991). While in the Sorell Basin (at least in the King Island and Sandy Cape Sub-basins see Moore et al. 1992) rifting was followed by low-energy sag fill or late rift deposition which was succeeded in the late-Cenomanian by uplift and erosion (Hill et al. 1997). Extension coupled with lithospheric cooling in the Durroon Basin split the basin into northwest trending graben and half-graben separated by major listric faults (Baillie & Pickering 1991). Graben development in the basin is asymmetrical with fault blocks tilted southwest (Seymour & Calver 1995). The Durroon Basin experienced increased geothermal gradients (up to 55ºC/km) from 100 Ma to 90 Ma, followed by Cenomanian uplift and erosion (Duddy 1992). A 110 Ma to 90 Ma cooling event is recorded throughout eastern Tasmania. This is partly in response to km scale erosion of a thick overlying succession, probably including Jurassic to Early Cretaceous stratigraphy (O'Sullivan & Kohn 1997; Kohn et al. 2002; O'Sullivan et al. 2000), but also influenced by a Cretaceous magmatic pulse.

Onshore, intrusions of alkaline igneous rocks occur in the southeast (Cygnet-Oyster Cove area) and in the far northeast (Cape Portland) (Fig. 6). The Cygnet Syenites occur as a series of sheet-like bodies and sills a few tens of metres thick, and as numerous widely dispersed smaller sills, dykes and pipes (Farmer 1985). They mainly intrude the lowermost units of the Lower Parmeener Supergroup. At Cygnet, the Lower Parmeener Supergroup is domed above a 750 m thick laccolith lying close to the surface (Leaman 1990). The alkaline rocks near Cape Portland occur in a series of small flows, dykes and irregular plug-like intrusions (Jennings & Sutherland 1969).

Tasmania experienced an elevated geothermal gradient in the Middle to Late Cretaceous. K-Ar ages indicate the age of emplacement of the Cygnet intrusion is 100.5 ± 3 Ma (Evernden & Richards 1962; McDougall & Leggo 1965), and the rocks at Cape Portland are dated at 101.3–102.3 ± 2.6 Ma (McDougall & Green 1982). Organic matter taken from the Parmeener Supergroup in the

Cygnet area has been totally carbonised by regional heating associated with the intrusions (Farmer 1985) while the Gordon Group at Ida Bay has been remagnetised by a Late Cretaceous heating event that persisted for about 10 million years (Sharples & Klootwijk 1981). This Cretaceous thermal event is widely recognised across the adjacent parts of Antarctica and New Zealand (Tessensohn 1994; Veevers 2000).

By the Late Cretaceous active spreading in the Tasman Sea had begun. In the Bass Basin, relatively high rates of subsidence associated with extension continued (Young et al. 1991). In the Durroon Basin there was significant growth on NE trending extensional faults in the early Late Cretaceous, resulting in thick (up to 2.5 km) wedges of Durroon Formation, unconformably overlain by a Campanian to Maastrichtian sag sequence (Hill et al. 1995). Baillie and Pickering (1991) report that by the latest Cretaceous subsidence in the basin had diminished with regional subsidence migrating towards the present day depocentre of the Bass Basin.

In the latest Cretaceous, a series of extensional structures began to develop across NE Tasmania. The onshore extension of the Bass Basin, the Tamar Graben was the first of these structures, beginning to develop in the Late Cretaceous (Fig. 9). The Tamar Graben is a NW striking structure, defined by a series of graben and half-graben dipping to the west, which contain several hundred metres of latest Cretaceous to Tertiary non-marine sediments (Forsyth 1989).

A phase of strike slip faulting is prominent in western and southern Tasmania. These faults are widespread but largely minor in nature. No major faults of this age have been recognized on the Tasmanian mainland. Representative orientations of these minor faults and fractures are shown in (Fig. 10). The faults affect rocks as young as 100 Ma and these fault movements predate the normal faulting in the Derwent Graben (Berry & Banks 1985). The faults increase in intensity towards the SW and no faults or fractures of this style are known in NE Tasmania. Cretaceous reconstructions for Gondwana place Tasmania very close to Northern Victoria Land. Rossetti et al. (2003) records a major dextral transpressional zone along the Rennick Glacier, which matches up with western Tasmania. This belt was interpreted as a Late Cretaceous deformation zone by Fleming et al. (1993). Although Rossetti et al. (2003) argued that the dextral faulting was active at 50 Ma. Further south again, a dextral transtensional zone in Marie Byrd Land is well dated at 95 Ma (Siddoway et al. in press; Siddoway et al. 2004) and we consider this the most likely age for the dextral faulting in the Rennick Glacier and in western Tasmania. This short period of dextral transtension during breakup contrasts with the long-term sinistral movement along this margin from the Late Jurassic to Eocene. The deformation associated with this event may correlate with Cenomanian inversion in the Sorell Basin.

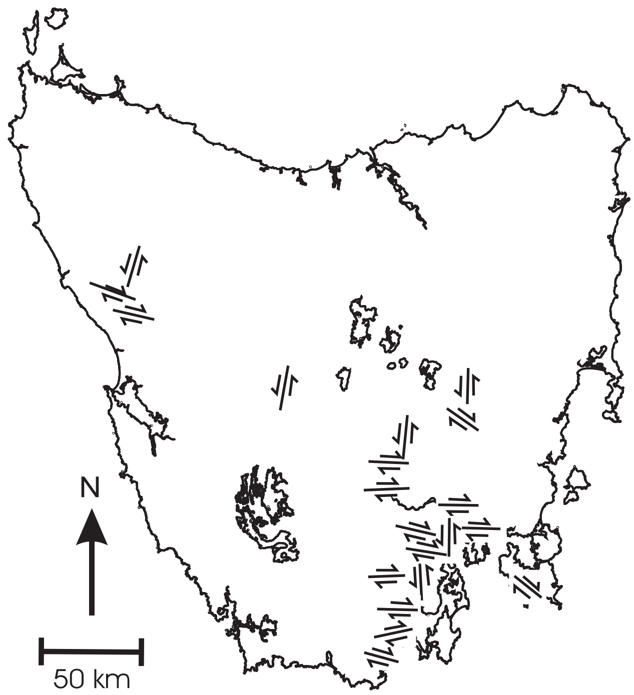

Figure 10. Pattern of Late Cretaceous strike-slip faulting in Tasmania.

Early Tertiary

During the Late Palaeocene to Early Eocene Tasmania was moving NNE along the left-lateral transform, which ran along the eastern margin of Antarctica. At about this time the western domain of the South Tasman Rise detached from Antarctica and joined the Australian Plate, sea-floor spreading propagated south of the South Tasman Rise and changed to a more southerly direction (Royer & Rollet 1997). Spreading in the Tasman Sea ceased in the Early Eocene (55–50 Ma) (Royer & Rollet 1997). Major structuring related to these events ceased by the end of the Eocene. The vertical displacement of the Early Tertiary faults can still be recognised in the modern topography. A new high-resolution digital elevation model (DEM) for Tasmania (DPIWE 2002) has greatly simplified the recognition of these structures (Fig. 9).

The Devonport-Port Sorell Sub-basin contains interbedded Paleocene to Late Oligocene non-marine sediments and basalt flows. The sub-basin is structurally related to the Bass Basin (Cromer 1989). The dominant structures are NNW–SSE trending, sub-vertical, en echelon normal faults. Gravity data suggest that the basin was land-locked, until at least the end of the Eocene (Cromer 1989). Analysis of the DEM shows three NNW striking faults including the east and west boundary of the basin. In the south, the basin is bounded by the uplifted Central Plateau (Fig. 9).

The Longford Sub-basin has received considerable attention, its structure being variously described as a graben with a central horst comprising Mt Arnon and Hummocky Hills (Carey 1947), and as a south-westerly dipping surface fractured by normal faulting downthrowing to the east (Longman 1966; Longman & Leaman 1971). Direen and Leaman (1997) considered the basin to be an asymmetrical depression developed on multiple, fluvially incised blocks, most of which exhibit half-graben rotation. Interpretation of recent regional seismic data acquired across the basin by Lane (2002), found the basin to consist of a NW trending western graben and eastern half-graben separated by a central horst (Fig. 11). These structures developed sub-parallel to the basin bounding Tiers and Castle Carey Faults. Lane (2002) interpreted the western graben to have formed at the onset of rifting, with a series of NE trending transfer faults active during the initial extension. Reactivation of the transfer structures initiating the development of the eastern half-graben (Lane 2002). Many of these faults were continuously active during deposition. Lane (2002) interpreted several hundred metres of growth on the western sides of the western graben and eastern half-graben (Fig. 11).

Basin fill was deposited under lacustrine/fluviatile conditions (Mathews 1989; Lane 2002) and was largely derived from the erosion of basement rocks (Mathews 1989). The sediments in the basin are largely Eocene in age, but extend back to the Paleocene in the deepest part of the basin near Hagley (Mathews 1983).

The large-scale topographic depressions produced by the Tamar Graben, the Devonport-Port Sorell and the Longford Sub-basins, are easily observed on the DEM, as are the main boundary faults of these structures (Fig. 9). The most obvious is the Tiers

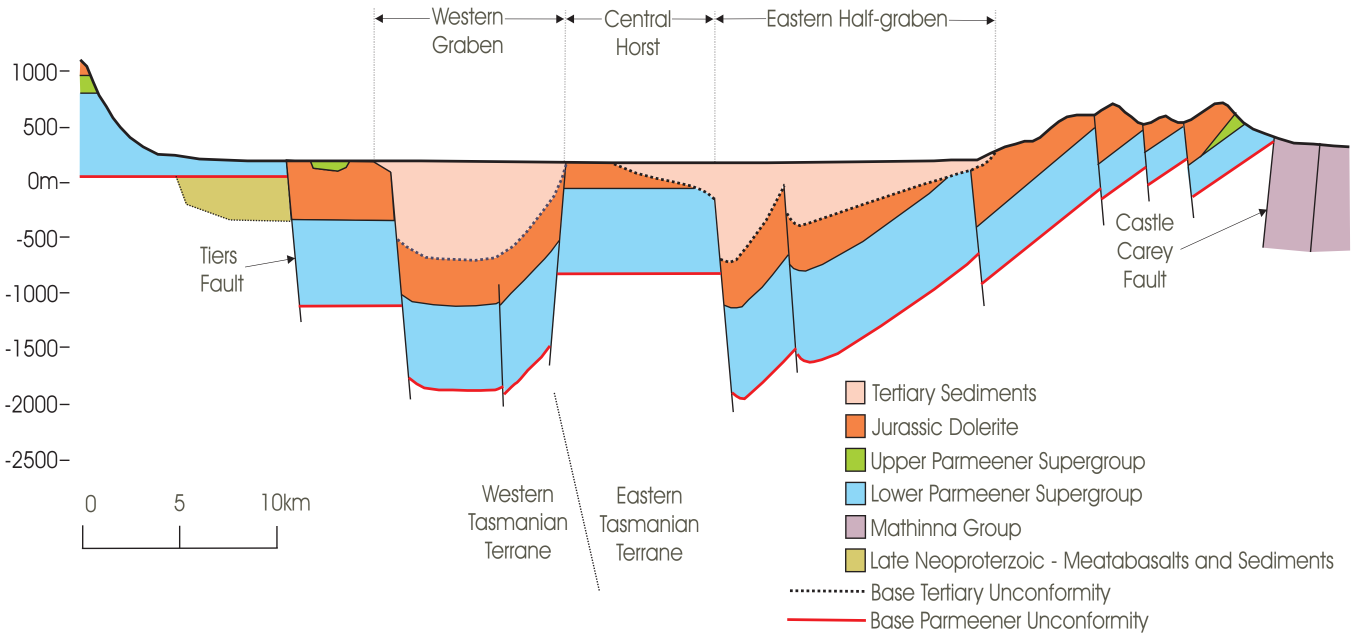

Figure 10. Cross-section across the Longford Sub-basin showing architecture and Tertiary growth faults (modified from Lane 2002) (location of section shown on Fig. 9).

Fault, a major structure with several hundreds of metres of normal movement in the Early Tertiary. Erosion was greatest in the NE with only remnants of Permian strata remaining on dolerite-capped hills, indicating that uplift and subsequent erosion was probably at its greatest in this area.

In the Strahan Sub-basin, Paleocene to mid-Oligocene strike-slip-related structures have been interpreted (Willcox et al. 1989; Moore et al. 1992). An en echelon, onshore extension of the Strahan Sub-basin, the Macquarie Harbour Graben (Fig. 9) contains about 500 m of Eocene marginal marine sediments (Baillie & Hudspeth 1989).

The Derwent Graben consists of two linked structures, the lower-and upper-Derwent Graben. The Lower Derwent Graben is a narrow NW trending structure, bounded on the west by the Cascades Fault system and by the Meehan Ranges in the east. The structure contains only a few hundred metres of sediment, the oldest being of Paleocene age, indicating faulting initiated in the early Tertiary (Colhoun 1989). Analysis of the DEM indicates the structure is paralleled by the Derwent Valley as far north as Bridgewater where it continues NW followed by the course of the Jordan River, and terminating near Melton Mowbray. The throw on the Cascades system decreases to the north from Hobart.

The Upper Derwent Graben is bounded on its southwestern side by a shallowly concave normal fault, down to the NE that approaches the Cascades Fault near Hobart. The fault follows the upper Derwent Valley before dying out in a horsetail splay on the Central Plateau. The intersection between the two faults is a lateral ramp, with no direct intersection between the two structures. Curved, non-linked faults as identified here are indicative of a regime of low extension, while the overall complexity of the observed pattern may indicate oblique extension (McClay et al. 2002), although an element of inheritance cannot be discounted. A complex transfer zone dominated by north-and east-striking faults links the Longford and Derwent Grabens (Fig. 9).

Miocene to Pleistocene compressional tectonics

Mio-Pliocene compressional tectonics is widely recognized in Victoria and measurements of the modern stress field indicate that Tasmania is also affected by a NNW compressional stress. However there is very little evidence for neotectonic structures in Tasmania. The only active fault scarp is the Lake Edgar Fault. Tasmania is actively rising at present (Murray-Wallace & Goede 1995) but this appears to be a regional response to high heat flow rather than a structural response to regional stress.

Summary

The oldest Neoproterozoic sedimentary rocks in Tasmania formed on an open shelf. A deeper shelf sequence to the east may be younger or a lateral equivalent. A Late Neoproterozoic (740–550 Ma) rift phase developed restricted half grabens with kilometre thick sequences containing tholeiitic basalts. This rift phase is correlated with the breakup of Rodinia. Tasmania was deformed during the Tyennan Orogeny 515–490 Ma and all units were apparently overmature by the end of the Cambrian. Evidence of petroleum generation is rare and restricted to the least deformed parts of the sequence, in NW Tasmania.

Shallow water deposition recommenced in the Ordovician. By the Late Ordovician all of western Tasmania was covered by a shallow-water tropical limestone. In the Silurian-Early Devonian siliciclastic sedimentation dominated on a shallow shelf across the Western Tasmanian Terrane. A turbidite succession built rapidly across eastern Tasmania. Deposition ceased with the onset of a Middle Devonian orogenic event throughout Tasmania. These sequences were metamorphosed to an overmature state (approximately 300ºC) except in the far south at Surprise Bay, and east at Scamander.

In the latest Carboniferous-Early Permian deposition began in a pericratonic basin. The Tasmania Basin comprises 1.5 km of flat-lying sedimentary rocks, known as the Parmeener Supergroup. The Late Carboniferous to Permian Lower Parmeener Supergroup consists of glaciomarine sediments and non-marine coal measures, and is unconformably overlain by Triassic non-marine sandstones and coal measures of the Upper Parmeener Supergroup. Recent work indicates the main body of the Tasmania Basin is mature but questions remain regarding distribution of reservoirs, migration paths and traps.

The large volumes of tholeiitic dolerite intruded into the Tasmanian crust during the Middle Jurassic are probably related to a major thermal anomaly occurring along the eastern margin of Gondwana. Dolerite mainly occurs as sills in the Tasmania Basin.

The Sorell, Bass and Durroon rift basins were initiated in the latest Jurassic-Early Cretaceous by a NE–SW oriented, extension related to rifting between Australia and Antarctica. Australia was drifting slowly north by the mid-Cretaceous. The Bass Basin has yielded only sub economic accumulations of gas. Minor hydrocarbon shows have been encountered in the Sorell Basin.

By the Late Cretaceous, active spreading had begun in the Tasman Sea. The Bass Basin continued to propagate southwards, extending onshore with the opening of the Tamar Graben in the latest Cretaceous, the Devonport-Port Sorell Sub-basin in the Early Paleocene, and the Longford Sub-basin in the Late Paleocene. On the west coast, the Sorell Basin extended onshore with the development of the Macquarie Harbour Graben in the Late Paleocene.

During the Late Paleocene-Early Eocene Australia was moving north along a left-lateral transform against the Antarctic plate. At about the time the Australian-Antarctic motion became more divergent along this margin (Eocene) the major southern rift basins were opened (Derwent Graben, Coal River Graben). The pattern of this extension is still preserved in the modern topography. These basins have not received sufficient sediment to reach maturity.

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