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Seismic Transects of Australia's Frontier Continental Margins

M.B. Alcock, H.M.J. Stagg, J.B. Colwell, I. Borissova, P.A. Symonds & G. Bernardel

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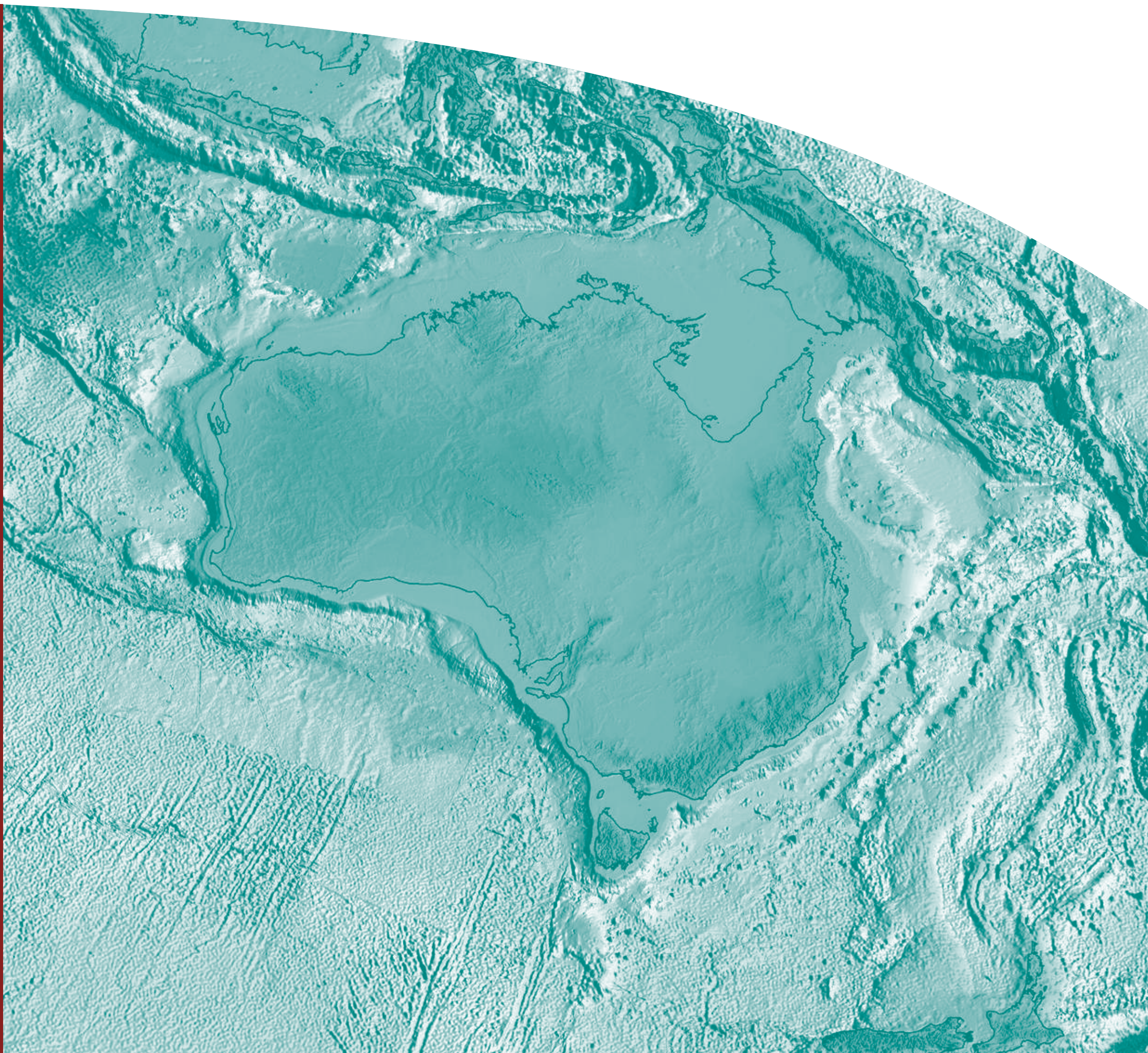

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DEPARTMENT OF INDUSTRY, TOURISM & RESOURCES

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SEISMIC TRANSECTS OF AUSTRALIA'S FRONTIER CONTINENTAL MARGINS

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INTRODUCTION

In October 2004, the Australian government made its submission to the United Nations Commission on the Limits of the Continental Shelf (CLCS) in which the outer limit of the extended Continental Shelf (ECS) was defined in accordance with article 76 of the United Nations Convention on the Law of the Sea (UNCLOS). The Australian submission covered areas of ECS in ten discrete areas (Fig. 1), *viz.*:

- Argo Abyssal Plain
- Australian Antarctic Territory
- Great Australian Bight
- Kerguelen Plateau
- Lord Howe Rise
- Macquarie Ridge
- Naturaliste Plateau
- South Tasman Rise
- Three Kings Ridge
- Wallaby and west Exmouth Plateaus

As part of the Australian submission, the Law of the Sea Project at Geoscience Australia prepared seismic transects from each of the submission areas, with the exception of Macquarie Ridge for which no high-quality seismic data were available. The rationale of these transects was to provide a summary of the geology of what are generally poorly-known remote areas of Australia's continental margins. The purpose of this Record is to publish a selection of these transects, together with some supporting text and figures on the geomorphology and geology. The text is only intended to provide basic background information; far more detailed information is contained in the references.

It should be noted that the level of detail in the interpretations is not consistent between areas, reflecting the levels of previous exploration and of available sample data. In the case of Three Kings Ridge, the transects are shown in uninterpreted form as the sediment cover is generally thin and undated. Regional stratigraphic diagrams have been included as figures or plates for those areas where such diagrams are available.

All seismic profiles have been reproduced at a horizontal scale of 1:200 000 and a vertical scale of 3 cm/sec, with the exception of some profiles from the Lord Howe Rise and Three Kings Ridge region, which were reproduced at a horizontal scale of 1:500 000 because of the lengths of the lines.

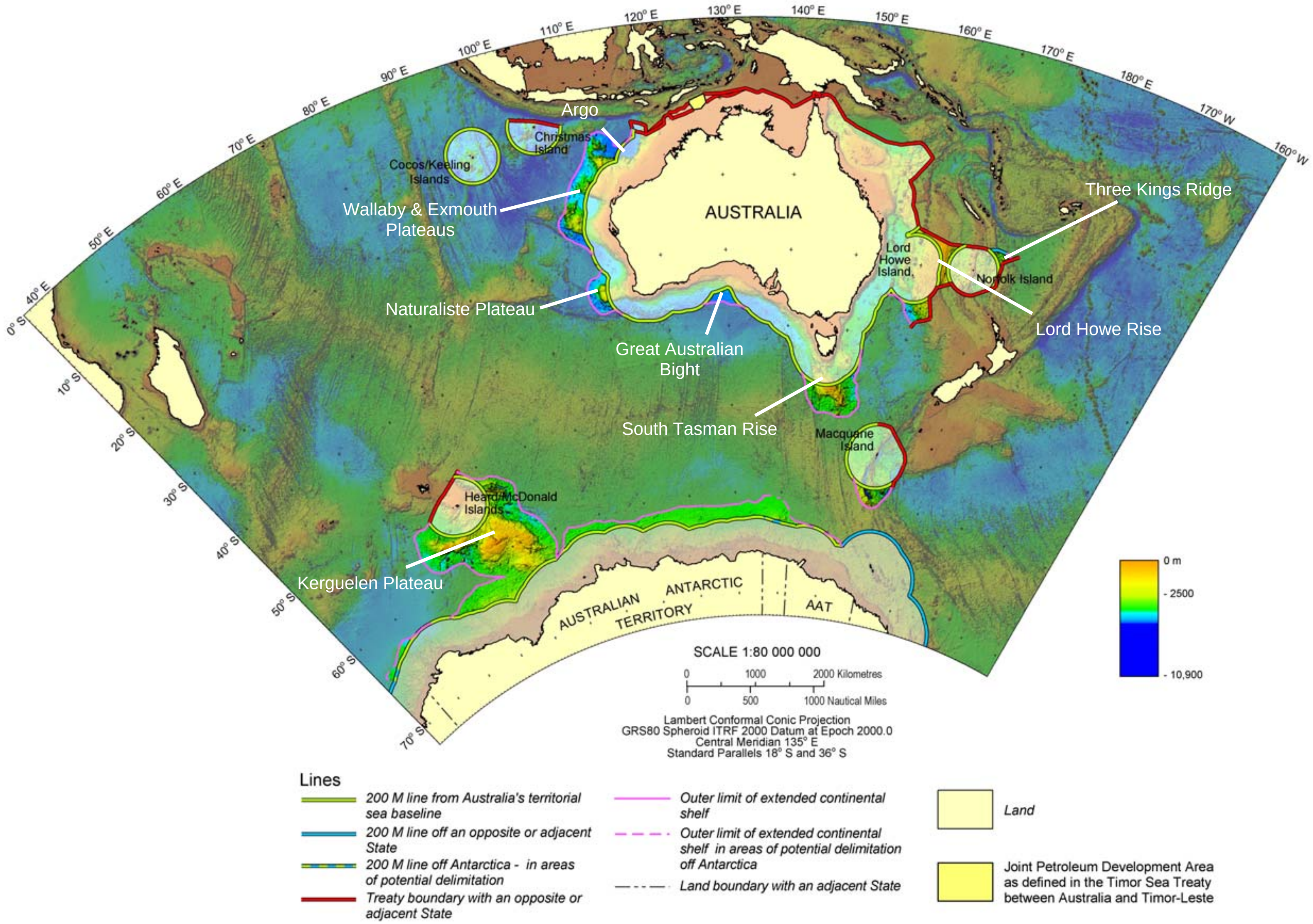

Figure 1: *Location of areas of extended continental shelf submitted to the United Nations Commission on the Limits of the Continental Shelf in November 2004.*

ARGO ABYSSAL PLAIN

INTRODUCTION

The Argo region continental margin lies adjacent to the central sector of the North West Shelf (Fig. 2) and includes the major morphological features of the Argo Abyssal Plain, Scott Plateau and Rowley Terrace. (Fig. 3). The Scott Plateau and Rowley Terrace are elements of the Browse and Roebuck Basins (Fig. 4), which are major components of the North West Shelf.

The Argo Abyssal Plain developed by seafloor spreading following the Middle to Late Jurassic (Callovian–Oxfordian) breakup of Australia's northwestern continental margin. It is an area of oceanic crust overlain by post-rift flat-lying sediments of variable thickness. The sediments range from claystones to carbonate turbidites.

The region has been explored by a number of geophysical surveys since the early 1970s. The extensive single and multichannel seismic data sets acquired on these surveys have been supplemented by several key cores and dredges and by a number of Deep Sea Drilling Program (DSDP) and Ocean Drilling Program (ODP) sites.

GEOMORPHOLOGY

The Argo region (covering part of the Argo Abyssal Plain; AAP) has an almost flat seafloor topography at water depths of 5600–5700 m with a slight regional southeasterly dip (Rao et al, 1994). It is bounded to the east by the Scott Plateau, and to the southeast by the Rowley Terrace (Figs 2 and 3). More distant, the region is flanked by the Sunda (Java) Trench to the north, the Joey and Roo Rises to the west, and the Exmouth Plateau to the south. The margins of continental Australia bordering the AAP to the south and east vary from relatively low gradient (outer part of the Scott Plateau) to steep and heavily incised by canyons (lower slopes of the Rowley Terrace and northern Exmouth Plateau).

The Scott Plateau lies at water depths of 2200–5000 m. It has a dome-shaped central part (including three bathymetric culminations at less than 2000 m depth), and broad canyons on its western flank (Jongsma & Johnston, 1993a; Stagg & Exon, 1981). The gradients on the upper slope and across the plateau crest are typically less than 2º, whereas gradients of the mid-slope are up to 5º. Gradients of the lower slope are typically less than 1º due to the presence of lava flows and other volcanic emplacements which are part of the regionally-extensive break-up-related volcanic province along this part of Australia's northwestern continental margin. In places, spurs of likely volcanic origin extend from the continental slope adding to the morphologic complexity of the continental margin.

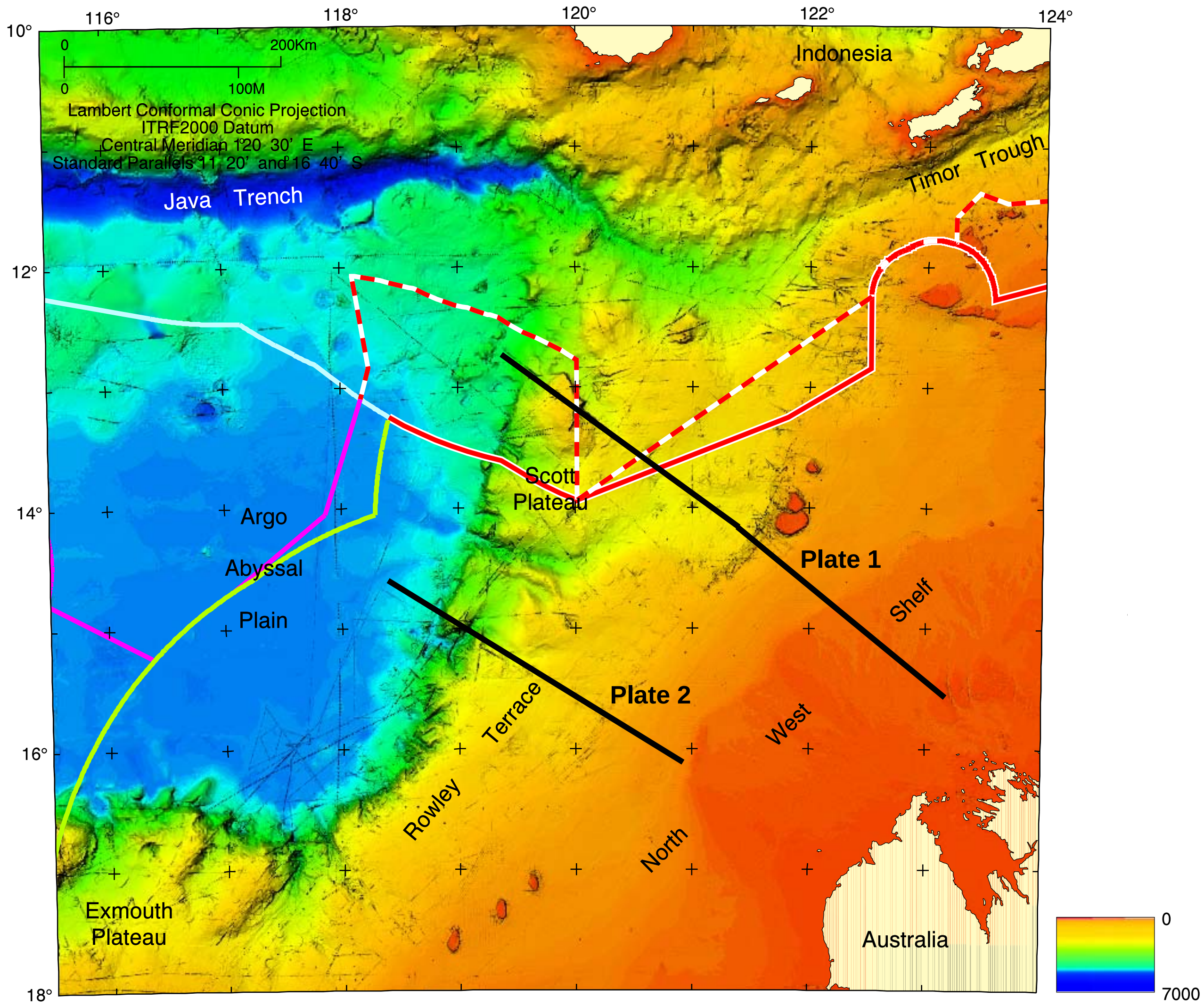

Figure 2: *Bathymetric image of the Argo region showing the location of the illustrated seismic transects. Major morphological features are labelled. Line colours: green – 200 M; magenta – outer limit of extended continental shelf (as submitted to UN CLCS, November 2004); light blue – 200 M (Indonesia); solid red – 200 M boundary with Indonesia; red& white dash – seabed treaty boundary with Indonesia.*

The Rowley Terrace lies to the south-southwest of the Scott Plateau at waters depths of 200–5000 m. Its northwestern margin abuts the deep ocean floor of the AAP. The lower continental slope is steep and incised by submarine canyons. As with the flanks of the Scott Plateau, the Rowley Terrace margin is sediment-starved and on most profiles there is no evidence of a continental rise.

The Exmouth Plateau margin to the south of the AAP has a complex morphology. It includes a series of spurs and small plateaus separated by large canyons (Fig. 2). Much of this margin has gradients of greater than 15º (Rao et al., 1994).

The geomorphology of the Argo region is primarily the result of the breakup of Australia's northwestern continental margin. This breakup was accompanied by extensive volcanism, particularly on the outer part of the Scott Plateau and the northern margin of the Exmouth Plateau.

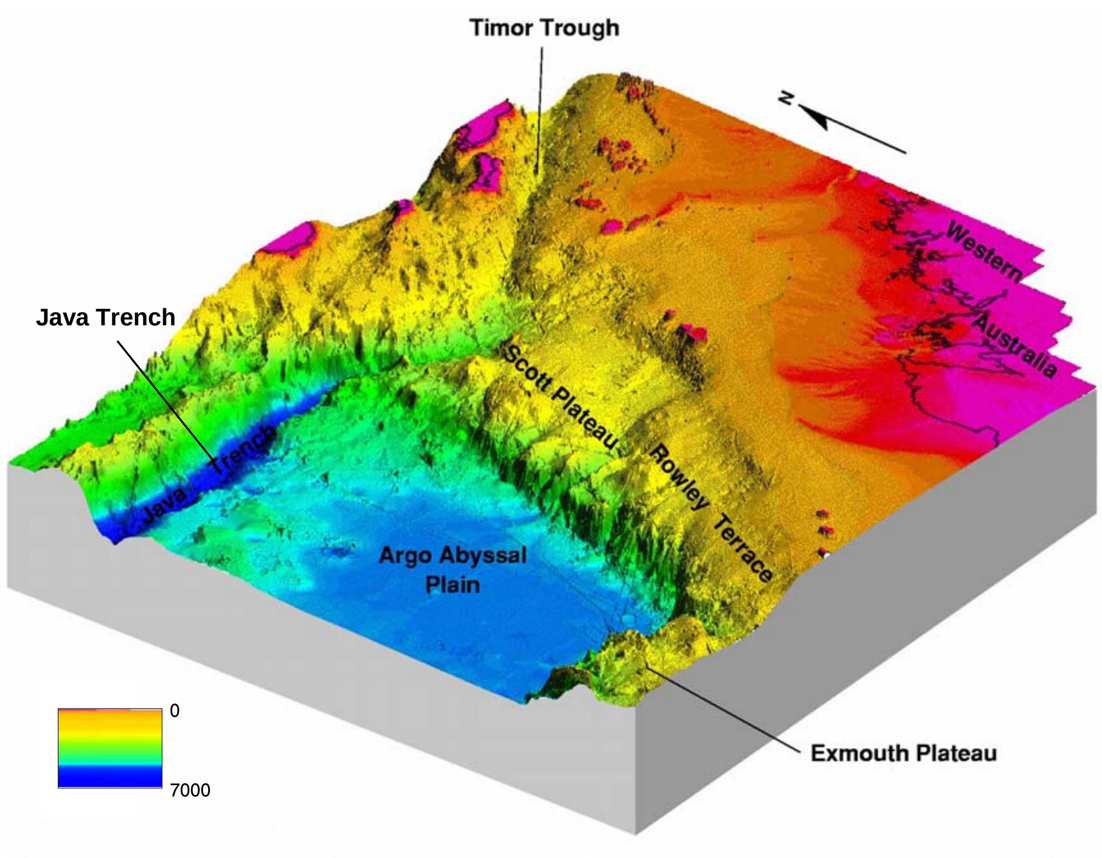

Figure 3: *3-D bathymetric image of the Argo region viewed from the southwest.*

GEOLOGY

Plate Tectonic Setting

The Argo Abyssal Plain (AAP) is the result of Gondwana breakup in the Middle / Late Jurassic to Earliest Cretaceous. The oceanic crust generated immediately after breakup along the Argo margin is the oldest in the Indian Ocean. Drilling at ODP Site 765 in the southeastern (oldest) part of the AAP adjacent to the Rowley Terrace sampled oceanic crust radiometrically dated at ~155 Ma (Ludden, 1992).

Mesozoic M26 – M10 seafloor spreading magnetic anomalies with an east-northeast strike were identified in the AAP by Powell & Luyendyk (1982) and Fullerton et al. (1989). This orientation was significantly different to that interpreted for anomalies in the Gascoyne, Cuvier and Perth Abyssal Plains, which were interpreted with a northeast orientation. This difference led to the interpretation that the Argo spreading was a quite separate event to that which produced the other ocean basins off the western Australian margin.

Mihut & Müller (1998a) revised the interpretation of AAP magnetic anomalies and derived similar trends to the anomalies identified previously in the Gascoyne and Cuvier Abyssal Plains. This revision suggested that the formation of the Argo Abyssal Plain occurred as the first stage of a systematic and progressive rupture between Australia and Greater India by means of a southward-propagating rift system.

Several different models have been proposed for this breakup. According to the most recent interpretation, based on the combined analysis of magnetic and other geophysical and geological data (Heine et al., 2002, 2004), the seafloor spreading started in isolated compartments of the southern Argo and Gascoyne Abyssal Plains simultaneously at M25 time on the same azimuth. The spreading ridge is suggested to have continued to the west along the northern margin of Greater India, joining up with the spreading system in the early Somali Basin, separating Madagascar / Greater India and Africa. A southward ridge jump occurred in the Argo Abyssal Plain at about M14 time, and the final breakup along Australia's western margin took place in the Valanginian, documented by a 10–15° counter-clockwise rotation of the spreading direction in the northern Argo Abyssal Plain. Despite this recent interpretation, there remains considerable doubt about the nature and age of assumed early seafloor spreading magnetic anomalies adjacent to the Gascoyne Abyssal Plain.

Geology of the region

The Argo margin (Fig. 5) is part of volcanic province which is up to 400 km in width (Symonds et al., 1998). Breakup-related volcanism significantly modified the margin, particularly in the Scott Plateau and outer Exmouth Plateau (Gascoyne margin) areas (Veevers & Cotterill, 1978; Symonds et al., 1998).

A variety of magmatic features have been recognised along the Argo margin in samples from wells and dredges. These include flood basalts on the Rowley Terrace (Ramsay & Exon, 1994) and other volcanics dredged on the Exmouth and Scott Plateaus and along the Argo margin (Colwell et al., 1994a; Crawford & von Rad, 1994).

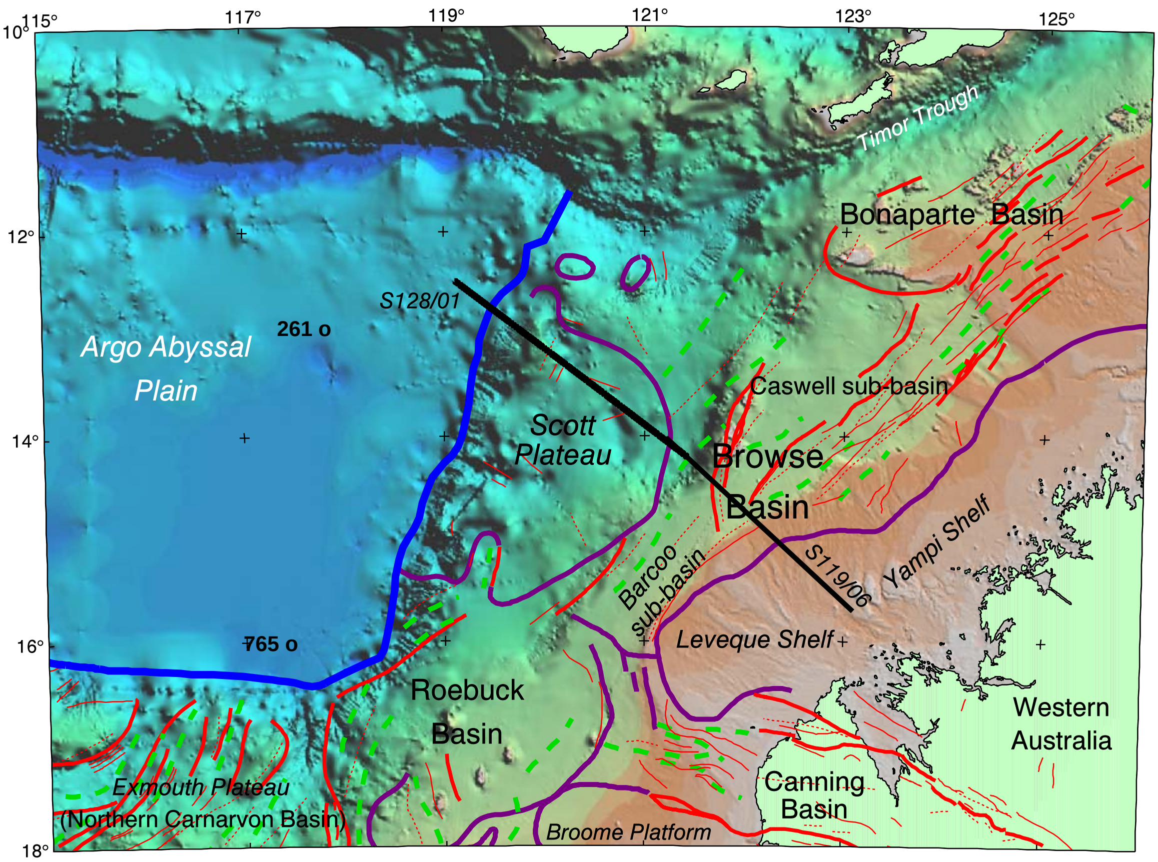

Figure 4: *Major geological and structural features of the Argo region. Also shown are outlines of the major sedimentary basins which comprise this part of the North West Shelf and the locations of DSDP Site 261 and ODP Site 765. The location of the profile shown in Figure 5 is indicated by the black line.*

Three main volcanic provinces and facies types have been identified in seismic data on the Argo margin, mainly associated with the Scott Plateau (outer part of the Browse basin) and the northern Rowley Terrace (outer Roebuck Basin; Symonds et al., 1998). These are:

- landward flows related to subaerial flood basalts in the south and throughout the Browse Basin;
- a rifted volcanic zone consisting of extensive flows, buildups and intrusives over much of the inner and central Scott Plateau;
- a transition zone on the outer edge of the plateau consisting of volcanics and faulted continental blocks.

Voluminous magmatism has modified and covered much of the Scott Plateau, and spread into the Browse Basin (Blevin et al., 1998). The various volcanic facies and features produced extend throughout a 400 km wide zone, and appear to have been emplaced at about the time of breakup in the Callovian–Oxfordian. Flexural isostatic modelling indicates that this magmatic episode was associated with significant lower crustal/upper mantle extension (Struckmeyer et al., 1998) beneath the Scott Plateau and Browse Basin, producing regional uplift along the locus of future Argo breakup, as well as over much of the Scott Plateau.

The AAP is underlain by a generally flat-lying Late Jurassic and younger sedimentary section up to 1000 m thick, that overlies Late Jurassic to Early Cretaceous oceanic crust. The sediments in the AAP were first sampled during DSDP Leg 27 at Site 261 (Fig. 4), where they consist of a thin section of Late Cainozoic ooze and clay unconformably overlying a thicker section of Late Jurassic to Early Cretaceous claystone. Site 261 reached total depth in the Late Jurassic oceanic crust (Buffler, 1994). In 1988, ODP Leg 123 drilled Site 765 (Fig. 4) and sampled a more complete sedimentary section as well as drilling into the underlying oceanic crust (Gradstein et al., 1992). The site was drilled at the approximate location of magnetic anomaly M-25A (162 Ma, Oxfordian). This age is considerably older than the oldest sediments recovered above basement (150 Ma, Tithonian), suggesting a long period of sediment starvation in the newly formed deep basin. Ludden (1992) radiometrically dated basement at this site at approximately 155 Ma. Relatively rapid sediment deposition (especially in structural depressions in the south and west) started in the Early Cretaceous. A mixture of pelagic clay and carbonate turbidites and debris flows were deposited on the steadily subsiding abyssal plain, with the turbidites and debris flows producing the highest sedimentation rates. Sedimentation rates from the Middle Miocene onwards were relatively high, particularly on the central part of the AAP.

ILLUSTRATED TRANSECTS

Two transects are included to illustrate the geology of the Argo region continental margin (locations shown in Fig. 2):

Plate 1: (lines GA-128/01 & GA-119/06) extends northwest from the Kimberley Block, across the central Browse Basin and the Scott Plateau, to the northern Argo Abyssal Plain.

Plate 2: (line GA-128/03) extends northwest from the southern Browse Basin, across the southern Scott Plateau, to the southern Argo Abyssal Plain.

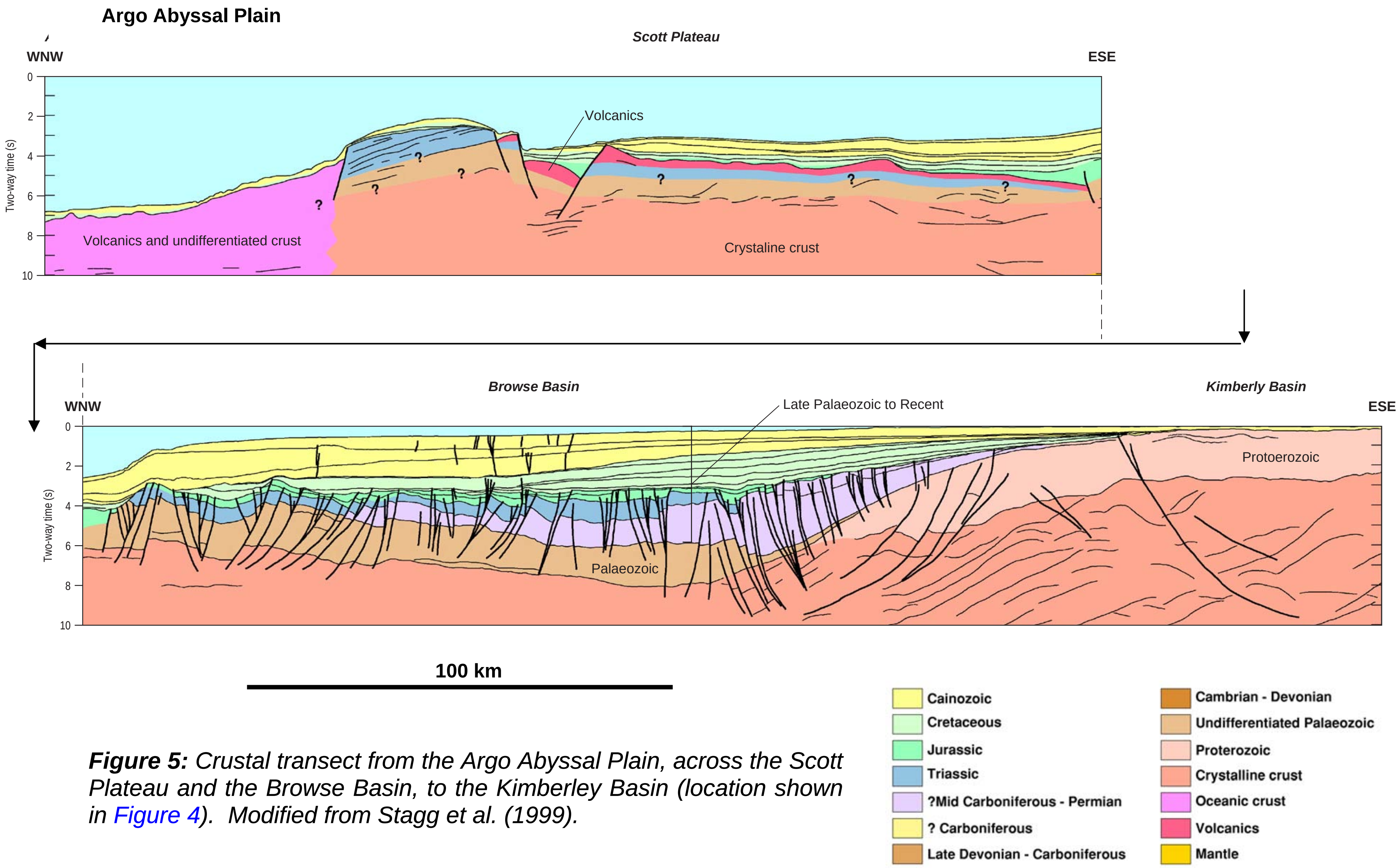

Figure 5: *Crustal transect from the Argo Abyssal Plain, across the Scott Plateau and the Browse Basin, to the Kimberley Basin (location shown in Figure 4). Modified from Stagg et al. (1999).*

AUSTRALIAN ANTARCTIC TERRITORY

INTRODUCTION

The Australian Antarctic Territory (AAT) comprises the part of the Antarctic continent and offlying islands lying between 45–136ºE and between 142–160ºE. Onshore, it is bounded to the west by Dronning Maud Land (Norway) and to the east by the Ross Dependency (New Zealand), while the sector from 136–142ºE comprises Terre Adélie (France). The continental margin of the AAT encompasses approximately 60% of the margin of East Antarctica. The western sector of the continental margin was formed by the rifting and breakup of Greater India from Antarctica-Australia in the Early Cretaceous, while the eastern sector of the continental margin was formed by the rifting and breakup of Australia and Antarctica in the mid- to Late Cretaceous.

Since the early 1980s, the continental margin of the AAT has been investigated by a number of widely-spaced surveys carried out by Australia, France, Japan, Russia/USSR and the USA. Sample data are scarce, being limited to a few sites drilled by the Deep Sea Drilling Program (DSDP) and the Ocean Drilling Program (ODP) and a small number of dredge and core samples. The extended continental shelf submission for the AAT is primarily based on the more than 20 000 km of high-quality deep-seismic data acquired by the Australian Antarctic and Southern Ocean Profiling Project in 2001 and 2002.

GEOMORPHOLOGY

The development of the margin of the AAT has been dominated for the past several tens of millions of years by glacial processes that have produced a morphology very different to ‘normal’ passive continental margins. In particular, the morphology of the shelf and slope are affected by the ice loading on the continent and by erosion and depositional processes arising from the interaction of the outlet ice streams and the Circum-Antarctic Current. The sediments deposited by these processes cover the margin with thick sequences that form a continuous blanket from the morphologic continental shelf, across the continental slope, to the deep ocean floor.

The nearshore margin of the Antarctic continental shelf is often more than 600 m deep, and occasionally attains depths of more than 1000 m in narrow troughs near the coast. The shelf break typically lies at depths of 300–400 m.

Beyond the shelf, the margin rarely exhibits the classic morphology of a continental slope and rise. Kennett (1982) states that the slope gradient on a non-glaciated margin averages about 4º and is greater than 1:40 (~1.4º); in contrast, the continental rise is marked by a gentle seaward gradient of 1:100 to 1:700 (~0.57–0.1º). However, the margin of the AAT typically shows a steep upper slope with gradients of 1º to >6º, and a lower slope with gradients of <0.4º to about 1º.

The 5500 km length of the AAT margin can be divided into segments that have characteristic morphologies. The following sections briefly describe the morphology in each sector.

Enderby Land (38–58ºE)

The continental margin in this area is characterised by a narrow continental shelf (30–70 km) and a continental slope that is dominated by spur and canyon topography for 150–250 km oceanwards from the shelf edge (Fig. 6). The upper slope, in particular, is characterised by closely-spaced, small-scale spurs and canyons. Down-slope, these features coalesce into broad (70–90 km width) spurs and canyons with up to 1000 m of relief that merge with the northwest-deepening Enderby Basin at more than 4500 m depth.

Gradients on the continental slope are highly variable, due to the local topography. The upper slope has an average gradient of about 1.5–2º. At approximately 3000 m depth, the gradient decreases to generally less than 1º, although local gradients can be more than 10º on the walls of canyons.

Offshore Prydz Bay (58–80ºE)

The continental shelf in this sector is highly variable in width, ranging from <100 km to the west of Prydz Bay (the Mac.Robertson Shelf), to more than 280 km in central Prydz Bay, and about 180 km east of Prydz Bay (Fig. 6). Prydz Bay is the outlet for the Lambert Glacier, the largest ice stream in East Antarctica; sediments derived from the Lambert Glacier are the major controlling factor for the margin morphology.

West and east of Prydz Bay, the upper continental slope is steep, with gradients typically more than 5º down to about 3000 m depth. The surface of the upper slope is extensively incised by small-scale canyons. Below the upper slope, the margin is dominated by large-scale fans and troughs (Wild Drift; Wild Canyon; Prydz Channel trough mouth fan). These features extend for more than 250 km from the upper slope and are characterised by their broad, low relief. Bathymetric profiles from the shelf edge to the eastern Enderby Basin typically have a catenary form with no significant breaks in gradient.

Princess Elizabeth Trough (80–87ºE)

The morphology of this part of the continental margin is quite simple. The shelf is broad, with a width of 100–200 km, and is partially covered by permanent ice. The upper slope is steep, with gradients of as much as 10-20º. At 2000–2500 m depth the morphology changes to a lower slope with gradients of 1–2º down to about 3000–3500 m. Below this depth, the seabed slopes more gently down to the axis of the Princess Elizabeth Trough at about 3800 m.

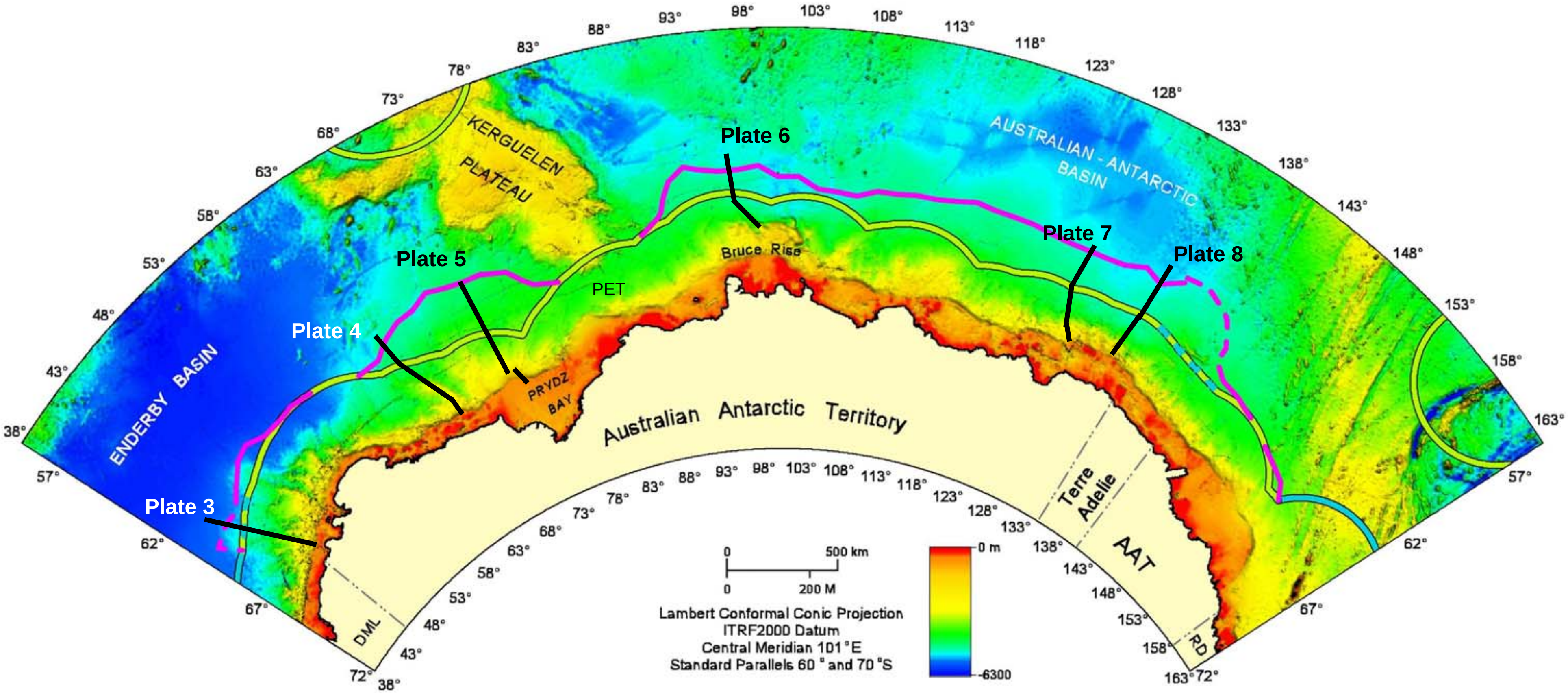

Figure 6: *Colour-gridded bathymetry image of the region of the Australian Antarctic Territory, showing the location of the illustrated seismic transects. PET – Princess Elizabeth Trough; DML – Dronning Maud Land; RD – Ross Dependency. Line colours: green – 200 M off AAT; magenta – outer limit of extended continental shelf (as submitted to UN CLCS, November 2004); blue – other 200 M lines (dashed in areas of potential delimitation).*

Wilhelm II – Queen Mary Lands (87–108ºE)

This area is one of the most morphologically complex parts of the margin of the AAT. The shelf is variable in width, ranging from ~100–150 km wide to the east and west of Bruce Rise, to about 250 km wide in the centre of the sector, inboard of Bruce Rise. Approximately 30% of the shelf in the central sector is covered by the permanent ice of the Shackleton Ice Shelf.

West of Bruce Rise, the shelf edge is strongly linear with an ENE-WSW trend, probably reflecting the Early Cretaceous India-Antarctica breakup phase. Canyon development on the upper slope is subdued in this area. The gradient of the upper slope is about 7º down to about 2000 m depth. Below this depth, there are consistent gradient segments of about 1–1.5º and 0.6º and less. There is some canyon development in this lower slope province.

East of Bruce Rise, the shelf edge generally trends ESE–WNW. Canyon development on the upper slope is widespread and gradients are more variable than to the west. While the overall gradient of the slope is low (generally less than 1º), local gradients can also be more than 10º. Despite the low gradients, the area is strongly incised by canyons down to depths of 3500–4000 m and it is therefore classified as a lower slope province.

This segment of the margin is dominated by Bruce Rise, the only significant marginal plateau off the AAT. It comprises two distinct parts, separated by a northwest-trending canyon (Bruce Canyon) that incises more than 1000 m into the seabed and debouches on to the lower slope at more than 3500 m depth. The eastern half of Bruce Rise is the shallower, at less than 1500 m depth. It is bounded by the upper slope to the south, Bruce Canyon to the west, an E–W trending rugged lower slope with average gradients of 3–4º to the north, and the NW–SE trending Vincennes Fracture Zone (Tikku & Cande, 1999) to the east. The western half of Bruce Rise is deeper than the eastern half, at slightly less than 2000 m, and its flanks are less well-defined.

North of Bruce Rise, the deep ocean basin lies at depths of close to 4500 m.

Wilkes Land – Terre Adélie (108–140ºE)

The generally E–W striking Wilkes Land and Terre Adélie margins formed during the rifting of Australia and Antarctica and have a relatively consistent morphology along their 1400 km length (Fig. 6). The shelf is broad (100–200 km width) and contains some characteristic inner shelf basins with water depths greater than 1000 m (cf. Mac.Robertson Shelf above).

As with much of the AAT margin, there is an upper slope with gradients of 2–4º that extends down to about 2000 m. There is then a distinct gradient change to a lower slope down to ~3500 m depth with gradients between 0.7–1.4º. Beyond this lower slope province, gradients are variable between 0.2–0.6º. The slope in this sector is extensively incised by canyons or valleys that extend from the shelf edge out to the edge of the abyssal plain at about 4000 m depth.

George V Land (140–160ºE)

The morphology of the George V Land sector is strongly influenced by the strike-slip motion between East Antarctica and southeast Australia, prior to clearance of the plates in the Oligocene. The shelf is highly variable in width, ranging from about 120 km in the west of the sector, to almost 300 km at about 150ºE (Fig. 6). Much of this shelf is semi-permanently covered by ice, and little is known of its morphology.

The morphology of the slope is similarly variable. In the west of the sector, where the plate tectonic setting is dominated by normal extensional processes, the upper slope has gradients that average 3º but are locally from 10–20º. At 2000–3000 m depth, the gradient abruptly decreases to less than 1º, with this gradient being fairly consistent down to the deep ocean basin at about 3500 m depth. The slope in this sector is extensively incised by canyons that generally trend oblique to the slope.

In the easternmost part of this sector, where the major fracture zones between western Tasmania and Antarctica converge with the Antarctic margin, the slope has a quite simple morphology. Here the margin is dominated by strike-slip movement. The upper slope is steep (5–20º) and extends down to 2000–2500 m depth. Beyond this province, gradients are low (~0.5º) and depths increase gradually to about 3000 m. The deep ocean basin here shows strong NW–SE trending relief, reflecting both the fracture zone traces and the thin sediment cover.

GEOLOGY

Plate tectonic setting

The continental margin of the Australian Antarctic Territory was formed by the extension and breakup of eastern Gondwana that culminated in the emplacement of oceanic crust between Greater India and East Antarctica in the Early Cretaceous (e.g. Brown et al., 2003a) and between southern Australia and East Antarctica in the Late Cretaceous (e.g. Cande & Mutter, 1982; Veevers, 1986; Tikku & Cande, 1999; Sayers et al., 2001). These discrete episodes of seafloor spreading have produced geologically distinct margin segments offshore from the western (India-Antarctica sector; 38–84ºE) and eastern (Australia-Antarctica sector; 107–160ºE) sectors of the Australian Antarctic Territory. The central sector of the offshore AAT (Triple junction sector; 84–107ºE) was formed near the major triple junction between southwest Australia, India and Antarctica and structuring in this sector consequently reflects both of the major spreading episodes.

Geology of the AAT margin

For clarity, the following sections separately summarise the geology of the western, central and eastern sectors of the continental margin of the AAT. The interpretation contained herein is principally based on the deep-seismic data acquired by the Australian Antarctic and Southern Ocean Profiling Project in 2001 and 2002 (Stagg et al., 2005b) and is illustrated here by a map of the interpreted tectonic elements off the entire AAT continental margin

(Fig. 7) and by detailed tectonic elements maps and representative seismic profiles from the continental margin off Enderby Land, Bruce Rise and Wilkes Land (Figs 8–13), as well as Plates 3–8.

India–Antarctica sector (38–84^{o}E)

The following section largely follows the interpretation presented by Stagg et al. (2005a, b) and is illustrated by reference to the tectonic elements map in Figure 8 and the seismic profile in Figure 9.

This part of the Antarctic continental margin formed during the breakup of the eastern margin of India and East Antarctica, which culminated with the onset of seafloor spreading in the Valanginian (e.g. Brown et al., 2003a).

The shelf edge and upper slope are interpreted to be underlain by a major basin-bounding fault system, beyond which the crystalline basement that underlies much of the shelf and immediate hinterland is down-faulted oceanwards by at least 6 km. The continental margin is underlain by a rift basin that ranges from about 100 km wide off western Enderby Land to more than 300 km wide off eastern Enderby and Mac.Robertson Lands. The thickness of rift-phase sediments in this basin is uncertain, due to the loss of seismic energy in the very thick post-rift sedimentary section.

The sedimentary section appears to be primarily of post-rift age. This section is at least 8 km thick north of Prydz Bay and at least 6 km thick off western Enderby Land. The section thins gradually oceanwards, but is still more than 2 km thick at 500 km from the shelf edge.

The margin is divided into distinct western and eastern sectors by a strong, north-south crustal boundary at about 58ºE (Fig. 8). Structuring in the western sector appears to be strongly influenced by the mixed rift–transform setting. In contrast, the eastern sector was formed in a normal rifted margin setting, albeit with complexities caused by the major N-S trending crustal structure of the Lambert Graben and the overlying Prydz Bay Basin.

As with the marginal rift basins, the continent-ocean boundary (COB), defined here as the inboard edge of unequivocal oceanic crust, shows a marked change in character across the crustal boundary at 58^{o}E. In the western sector, the location of the COB is defined mainly on the basis of a change in the reflection seismic character, supplemented by potential field modelling; here, its location appears to coincide with the zone of deepest basement on the margin (Stagg et al., 2005a). However, in the eastern sector, the COB is a prominent and sharp boundary in the reflection seismic data which correlates with a marked change in the crustal velocity profile as shown by the interpretation of sonobuoy records. Potential field modelling confirms this interpretation. As also noted by Gandyukhin et al. (2002), this boundary often correlates with an oceanward step-up in the basement level of up to 1 km (Fig. 9).

The character of the oceanic crust is also highly distinctive, again with a major west-to-east character change at 58^{o}E. In the western sector, the basement surface is of variable character, from rugged with a relief of more than 1 km over distances of 10–20 km, to rugose with low-amplitude relief on longwavelength undulations. The crustal velocity structure is unusual, with velocities of 7.6–7.95 km.s^{-1} being recorded at several stations at a depth that gives a thickness of overlying crust of only 4 km. It is possible that these velocities are from mantle, in which case the thin crust may be due to the presence of fracture zones. Alternatively, the velocities may be coming from a lower crust that has been heavily altered by the intrusion of mantle rocks. Oceanic crust in the eastern sector has a more typical oceanic velocity structure and is particularly characterised by its internal reflection fabric, which comprises: a smooth upper surface underlain by short, seaward-dipping reflectors; a transparent upper crustal layer, probably correlating with a sheeted dyke complex; a lower crust dominated by dipping high-amplitude reflections that probably reflect intruded or altered shears; a strong reflection Moho, confirmed by refraction modelling; and prominent landward-dipping upper mantle reflections on several adjacent lines. Current models for reflective oceanic crust cannot be readily applied to this crust.

Potential field modelling indicates that the gross margin structure is relatively simple, and that both the continental and oceanic crusts have behaved as a semi-rigid plate that has been depressed landwards by the thick post-rift sediment loading.

Illustrated Transects

Three transects are included to illustrate the geology of the India–Australia sector of the continental margin of the AAT (locations shown in Fig. 6):

Plate 3: (line GA-229/35) extends across the margin off western Enderby Land.

Plate 4: (lines GA-33/62 & GA-228/06) extends across the margin off Mac.Robertson Land.

Plate 5: (lines GA-33/33 & GA-229/30) extends across the margin off Prydz Bay.

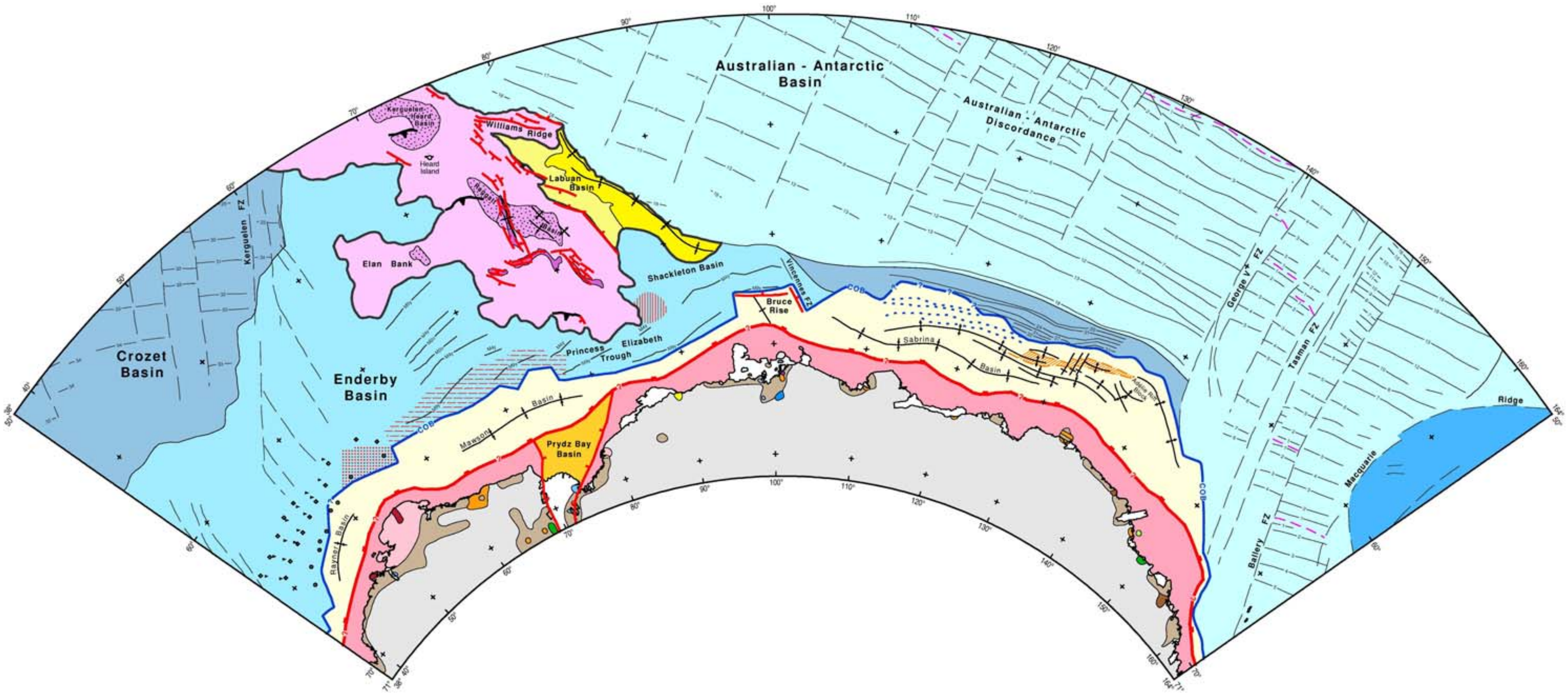

Figure 7: *Caption and legend on next page.*

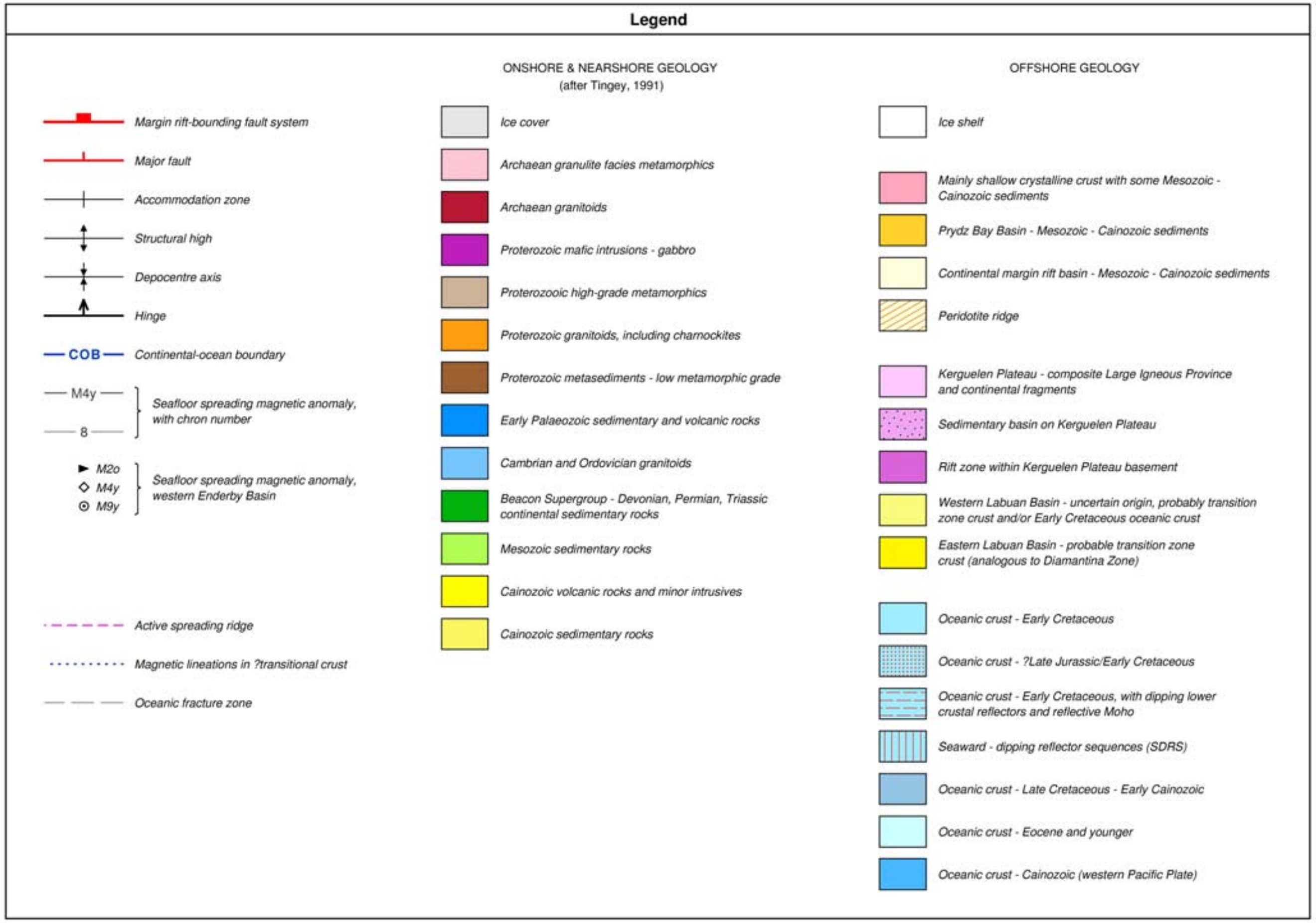

Figure 7: *Tectonic elements of the continental margin and adjacent ocean basins in the region of the Australian Antarctic Territory (After Stagg et al., 2005b).*

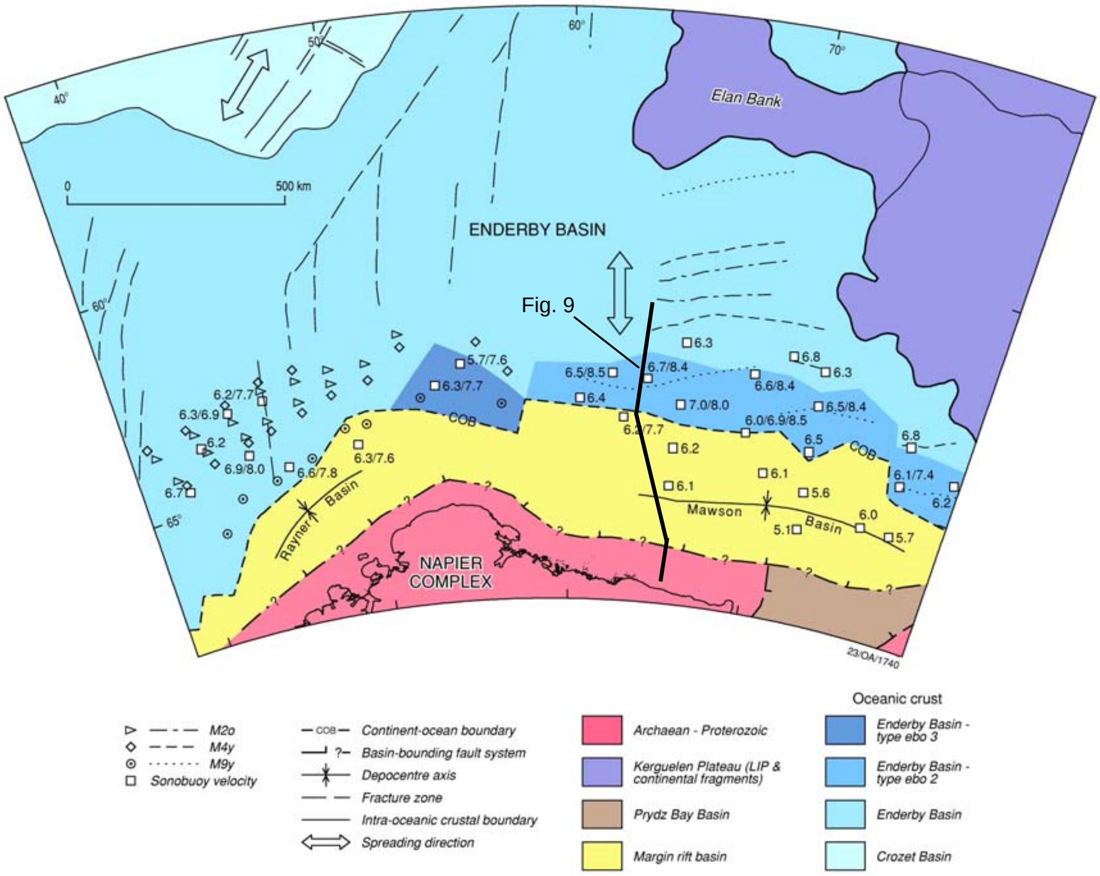

Figure 8: *Tectonic elements of the Enderby and Mac.Robertson Land margin and Enderby Basin (after Stagg et al., 2005a), showing the location of the profile shown in Figure 9.*

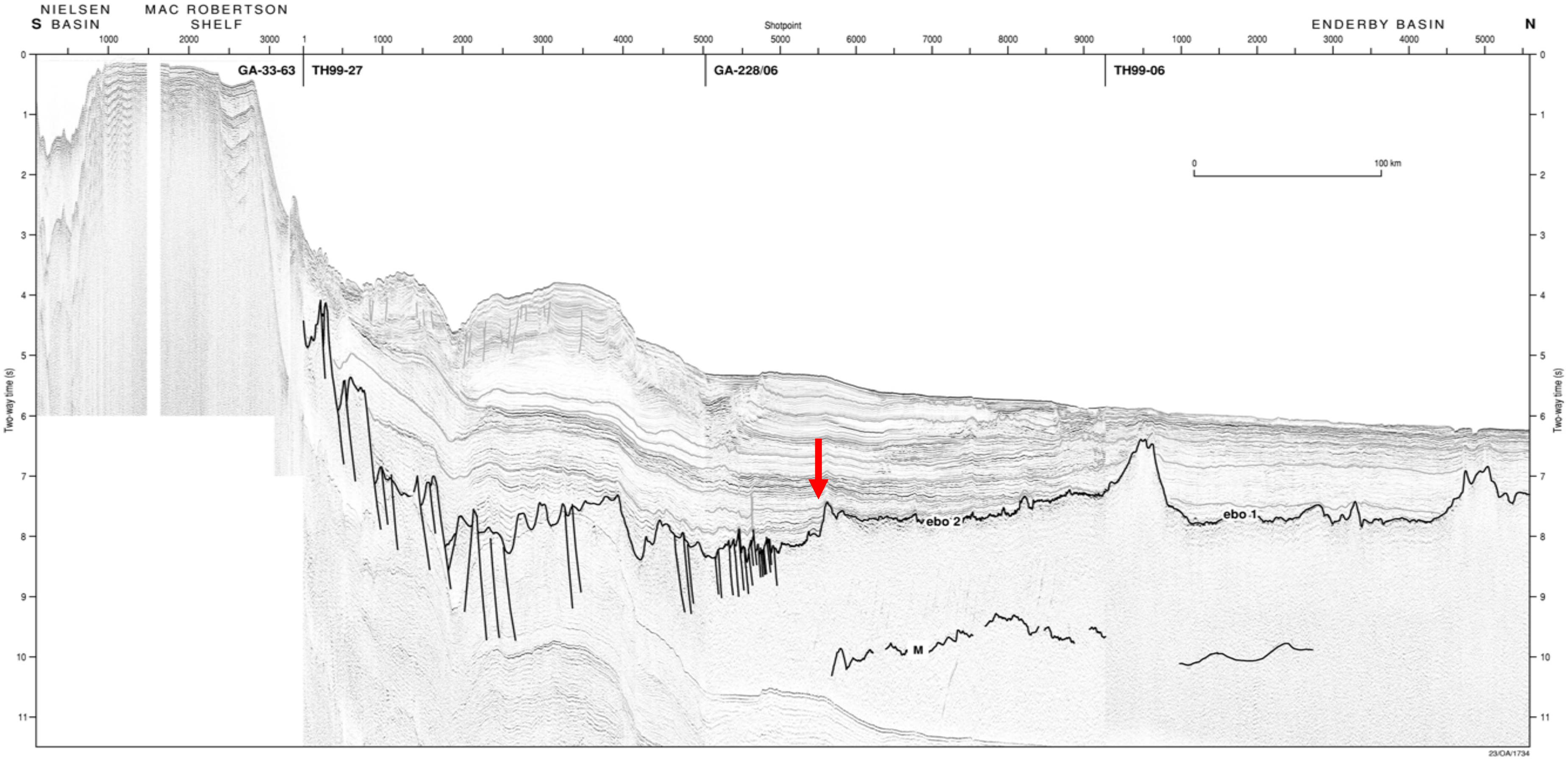

Figure 9: *Composite seismic profile from the Mac.Robertson Shelf to the Enderby Basin (after Stagg et al., 2005a). Location shown in Figure 8. Line numbers and shot-point numbers are annotated along the top of the section. Lines GA-33/63 and TH99-27, and GA-228/06 and TH99-06 do not directly tie. Arrow shows inboard edge of oceanic crust. M is reflection Moho; ebo1 and ebo2 are top of oceanic crust.*

India–Antarctica–Australia sector (84–105ºE)

This sector of the Antarctic continental margin was formed by the overprinting of the Early Cretaceous separation of Greater India from Antarctica-Australia by the Late Cretaceous breakup of Australia and Antarctica and is illustrated by reference to the tectonic elements map in Figure 10 and the seismic profile in Figure 11. Margin structuring is complex and it is difficult to discriminate between primary and overprinted structures.

Because of the generally extensive ice cover in this area that is mainly derived from the Shackleton Ice Shelf, there is no seismic coverage of the continental shelf and only very sparse lines on the upper continental slope. As there are also no seabed samples from the shelf and uppermost slope we have no knowledge of the geology of this area. However, as with the sectors to the east and west, it is likely that the shelf edge is underlain by a major fault system with crystalline basement being down-faulted oceanwards to depths of several kilometres.

As noted previously, this sector is dominated by the presence of a mid-slope marginal plateau, the Bruce Rise. Seismic coverage of this structure is extremely limited, except on the plateau margins. The eastern margin is crossed by several Japanese seismic lines (TH-94 survey; Ishihara et al., 1996) and one survey GA-229 line, while the northern margin is crossed by five survey GA-228, GA-229 and TH-94 lines. The TH-94 survey also recovered three dredge samples from the eastern margin of the Bruce Rise. These dredges contained granite, schist and sandstone and confirm the continental affinity of this feature (Ishihara et al., 1996).

Seismic data show that the Bruce Rise is very similar to its conjugate feature on the Australian margin, the Naturaliste Plateau; it is underlain by shallow acoustic basement, with faulted pockets of interpreted Cretaceous sediments (Fig. 11). Both the Bruce Rise and the Naturaliste Plateau are covered with a veneer of sediments a few hundred metres thick. The eastern and northern margins of the Bruce Rise are sharply defined: the extremely steep, northwest-trending eastern margin is underpinned by the Vincennes Fracture Zone, while the E-W trending northern margin is down-faulted by at least 3 km across several major, high-angle faults. The western margin of the Bruce Rise is poorly-defined, with no seismic lines extending landward of the lower slope.

The deep ocean basin adjacent to the Bruce Rise is highly variable in structure and seismic character and its origins are enigmatic. Between the western flank of the Bruce Rise and the eastern end of Princess Elizabeth Trough, the crust appears to be oceanic. Potential field modelling (Stagg et al., 2005b), the identification of Early Cretaceous seafloor spreading magnetic anomalies (Stagg et al., 2005b) and similarities of seismic character with oceanic crust in the Enderby Basin, suggest that this crust was generated during the separation of India from Australia-Antarctica. In contrast, the basement between the northern flank of the Bruce Rise and the fast-spreading crust of the Australian-Antarctic Basin to the north has seismic characteristics suggesting that, despite the presence of interpreted seafloor spreading magnetic anomalies (Stagg et al., 2005b, in press), there may be fragments of continental crust that have been left stranded in an overall setting of oceanic crust. Alternatively, these fragments may comprise oceanic crust that has subsequently been faulted during the period of very slow spreading between Australia and Antarctica in the Late Cretaceous. The small, triangular-shaped crustal fragment east of the Vincennes Fracture Zone is also enigmatic. Satellite gravity data suggest that it was also formed during the separation of India from Antarctica-Australia.

Illustrated Transects

One transect is included to illustrate the geology of the India–Antarctica–Australia sector of the AAT continental margin (location shown in Fig. 6):

Plate 6: (line GA-228/13) extends from the outer flank of the Bruce Rise, across the Shackleton Basin to the fast-spreading crust of the Australian–Antarctic Basin.

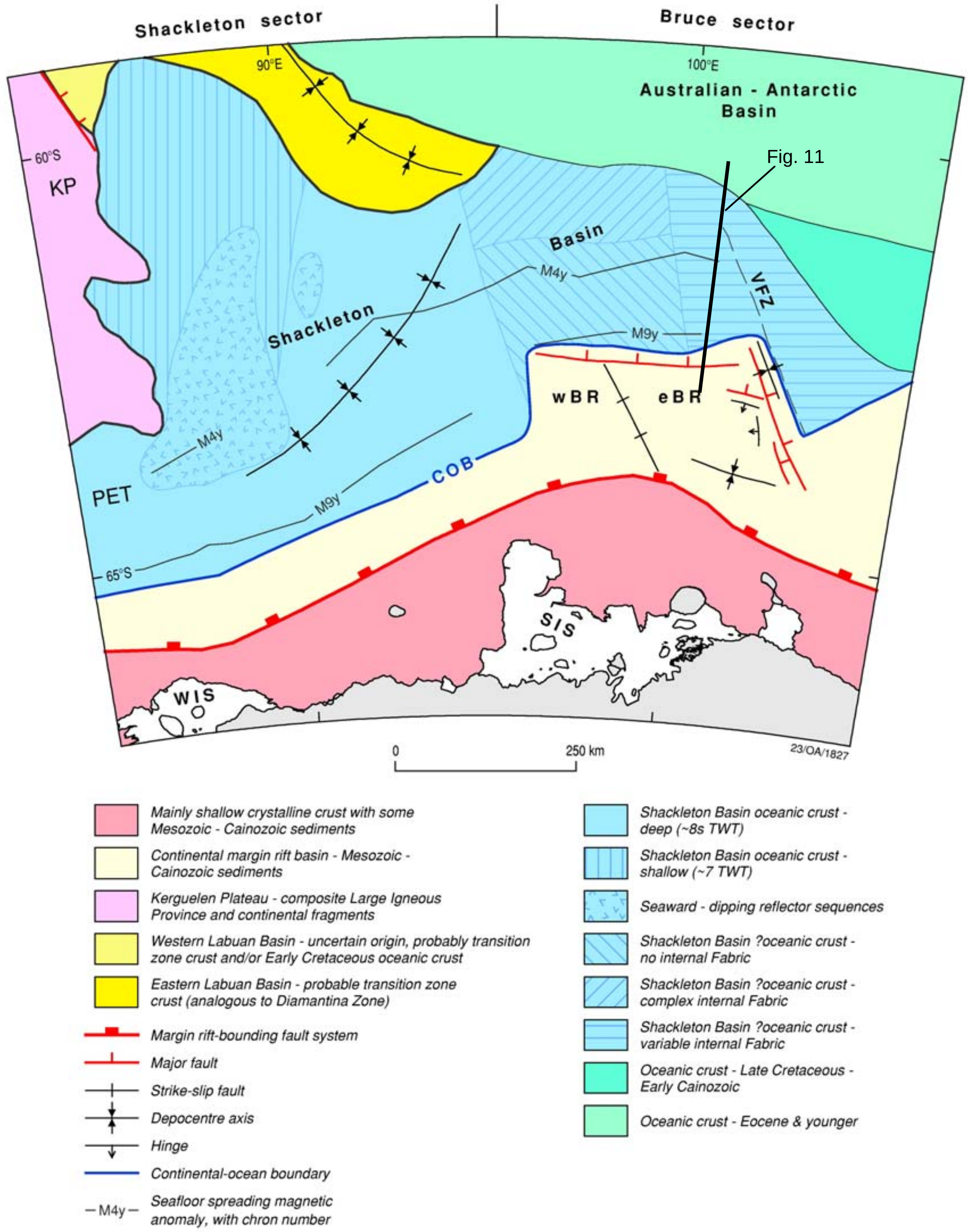

Figure 10: Tectonic elements of the continental margin in the area of the Bruce Rise and Shackleton Basin (after Stagg et al., in press). Shows the location of the seismic profile illustrated in Figure 11. VFZ – Vincennes Fracture Zone; wBR – west Bruce Rise; eBR – east Bruce Rise; KP – Kerguelen Plateau; PET – Princess Elizabeth Trough; COB – continent-ocean boundary; WIS – West Ice Shelf; SIS – Shackleton Ice Shelf.

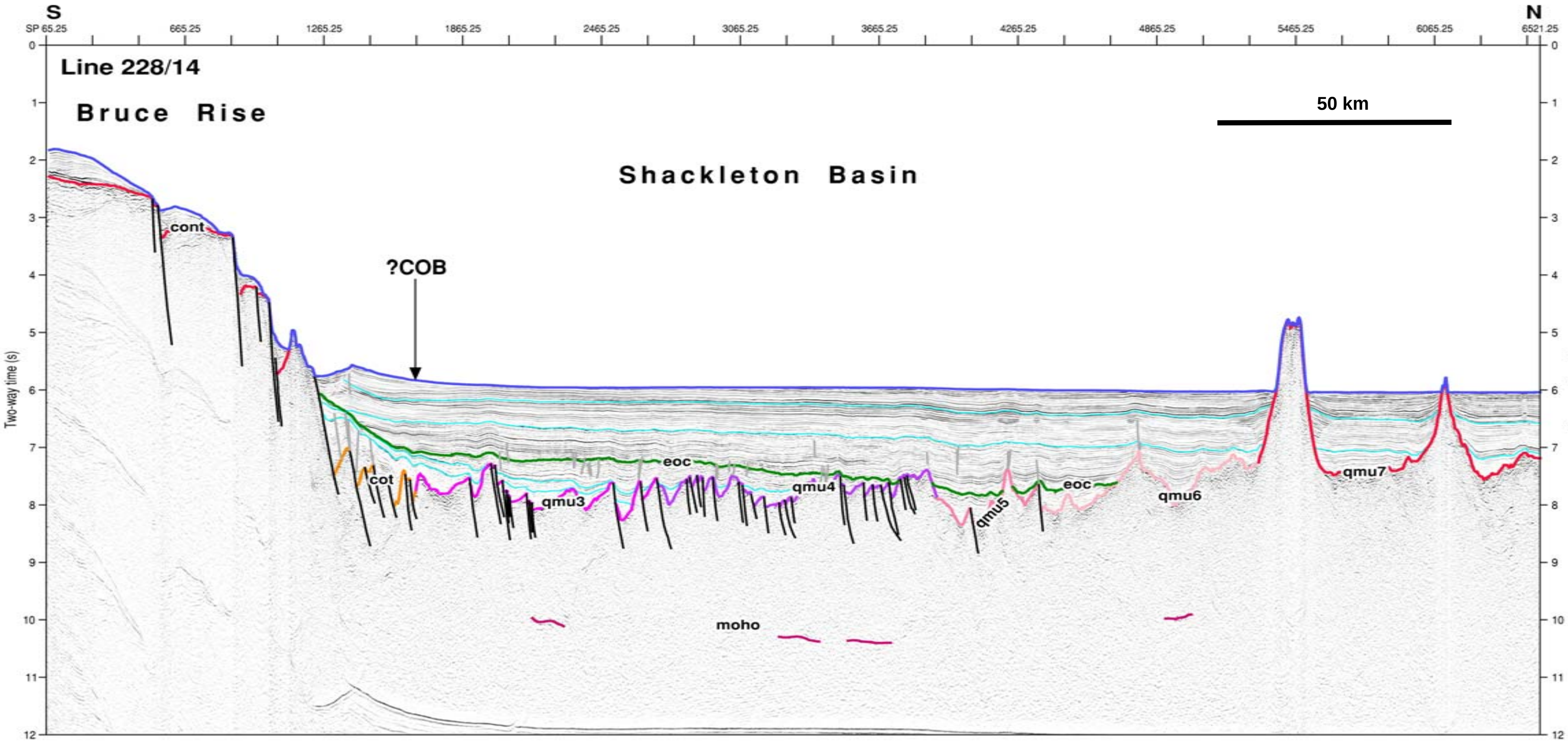

Figure 11: *Seismic profile from the Bruce Rise and adjacent Shackleton Basin. Location shown in Figure 10. Line number and shot-point numbers are annotated along the top of the section. COB indicates the location of the interpreted continent-ocean boundary.*

Australia-Antarctica sector (105–154^{o}E)

The interpretation presented in this section largely follows that of Colwell et al. (2006). The geology of this part of the margin is illustrated by the tectonic elements map (Fig. 12) and a representative seismic profile (Fig. 13).

This part of the Antarctic continental margin formed during the breakup of the southern margin of Australia and East Antarctica, which culminated with the onset of seafloor spreading. The timing of final continental separation has been extensively debated since the first identification of seafloor spreading magnetic anomalies by Weissel & Hayes (1972). While it is generally accepted that seafloor spreading was initiated in the Late Cretaceous, current estimates of the timing of this event range from 83 to 95 Ma (e.g. Cande & Mutter, 1982; Veevers, 1986; Tikku & Cande, 1999; Sayers et al., 2001).

West of 140^{o}E, the margin was formed mainly by orthogonal rifting between Antarctica and Australia and these rift structures are largely oriented parallel to the margin. East of 140^{o}E, margin formation was strongly influenced by the oblique and strike-slip separation of Antarctica from southeast Australia and the South Tasman Rise.

As with the India–Antarctica sector of the AAT continental margin, the shelf edge and upper slope are probably underlain by a major basin-bounding fault system, beyond which the crystalline basement underlying much of the shelf and immediate hinterland are down-faulted oceanwards by several kilometres. The broad continental slope is underlain by a major rift basin that extends for at least 1500 km along the margin offshore from Wilkes Land and Terre Adélie. Beneath this basin, the crust thins oceanwards through extensive faulting of the rift and pre-rift sedimentary section and by mainly ductile deformation of the crystalline crust. The total thickness of pre-rift, rift and post-rift sedimentary rocks in this basin is probably at least 7 km.

Outboard of the rift basin, a 90 to 180 km-wide continent-ocean transition zone is interpreted to consist primarily of continental crust with magmatic components that can account for the lineated magnetic anomalies that have been interpreted in this zone. The thick sedimentary section in the COT zone is floored by dense lower crustal or mantle rocks indicating massive (>10 km) thinning of the lower and middle crust in this zone. As with the conjugate Australian margin, the first unequivocal seafloor spreading magnetic anomalies place the start of oceanic spreading at about chron 33o time.

The boundary between the margin rift basin and the COT is marked by a basement ridge which potential field modelling indicates is probably composed of altered/serpentinised peridotite. This ridge is similar in form and interpreted composition to a basement ridge located in a similar structural position at the inboard edge of the COT on the conjugate margin of the Great Australian Bight (Sayers et al., 2001). Both ridges are probably in part the product of mantle up-welling and partial melting focussed at the point of maximum change/necking of crustal thickness.

Integrated deep-seismic and potential field interpretations and dredged rock samples of continental origin off Terre Adélie (Tanahashi et al., 1997; Yuasa et al., 1997) point very strongly to the boundary between unequivocal oceanic crust and largely continental crust of the continent-ocean transition as lying in very deep water, and considerably seaward of earlier interpretations (often based on inadequate seismic data or magnetic data only). The continent-ocean boundary is considered to be well-constrained from 124–131^{o}E and unequivocal from 131–140^{o}E, but more debatable in the sector from 110–124^{o}E.

There is strong symmetry between the formerly conjugate margins of southern Australia and East Antarctica east of about 120^{o}E. This symmetry is evident both in the crystalline crust and in the overlying rift section. In contrast, the post-rift sections are markedly different in thickness, reflecting the different depositional histories during much of the Cainozoic. In addition to the symmetry between the margins, there is also a strong correlation in the seismic characters, which gives us some confidence in dating the major unconformities as being of base Turonian, Maastrichtian and early Middle Eocene age.

Illustrated Transects

Two transects are included to illustrate the geology of the Australia–Antarctica sector of the AAT continental margin (locations shown in Fig. 6):

Plate 7: (lines GA-227/2301 & GA-228/24) extends across the margin off central Wilkes Land out to the Australian–Antarctic Basin.

Plate 8: (line GA-227/2601 & GA-228/28) extends across the margin off eastern Wilkes Land out to the Australian–Antarctic Basin.

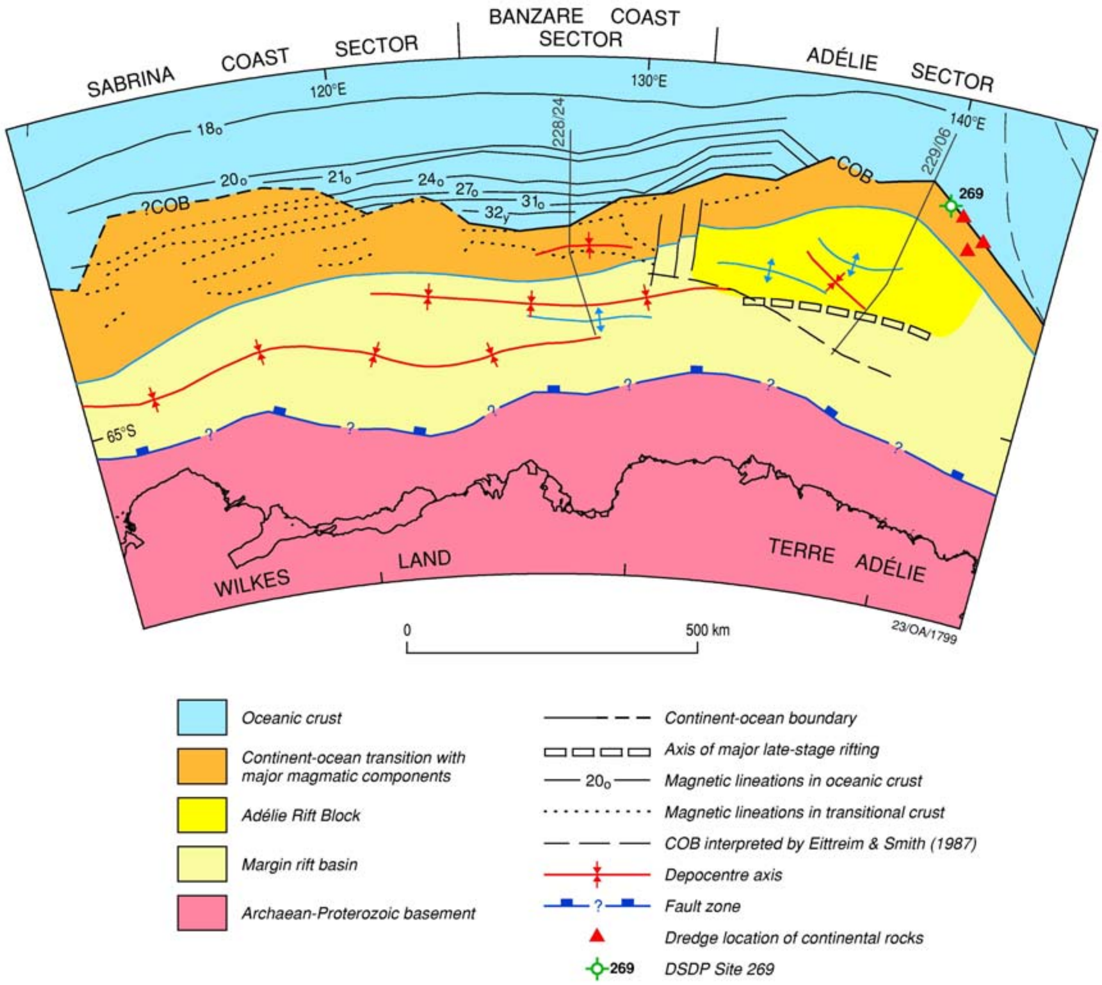

Figure 12; *Tectonic elements of the continental margin of Wilkes Land and Terre Adélie (after Colwell et al., 2006).*

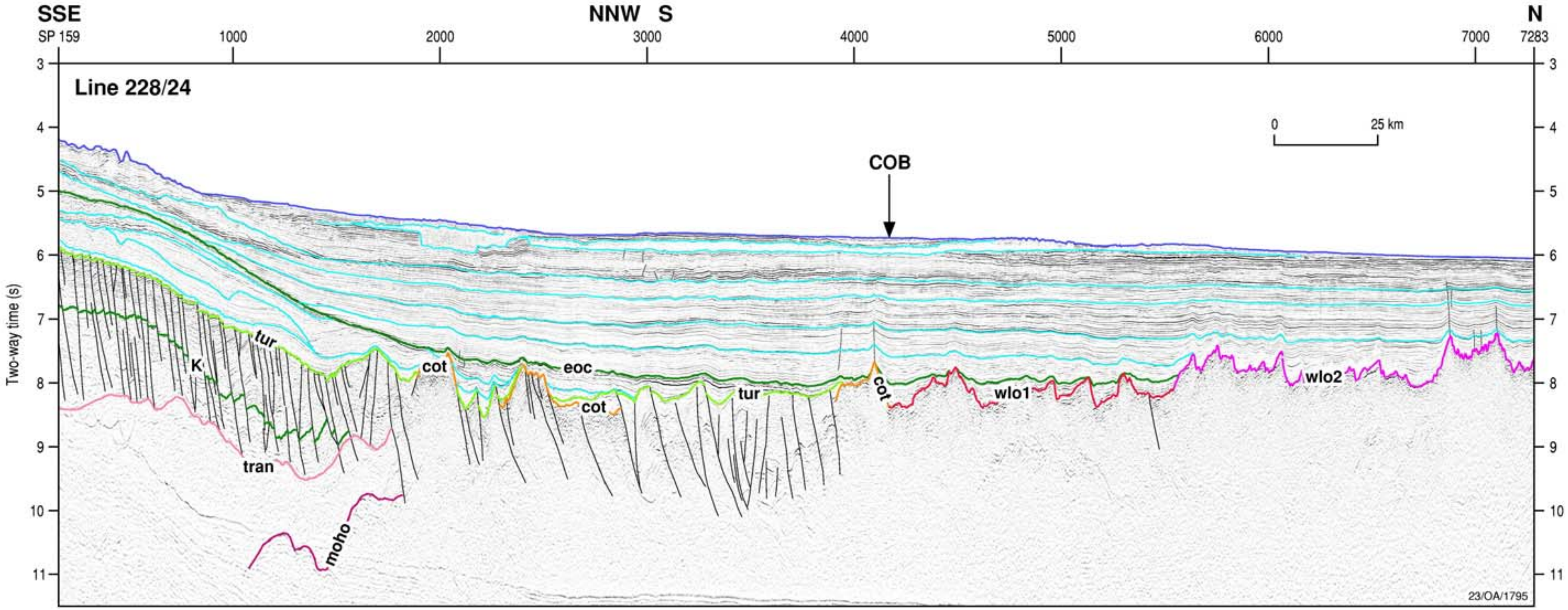

Figure 13: *Seismic profile GA-228/24 from the continental margin off Wilkes Land (after Colwell et al., 2006). Location shown in Figure 12. Line number and shot-point numbers are annotated along the top of the section. COB indicates the location of the interpreted continent–ocean boundary.*

GREAT AUSTRALIAN BIGHT

INTRODUCTION

The Australian continental margin in the Great Australian Bight (GAB) region is a rifted non-volcanic margin formed by continental extension and breakup between Australia and Antarctica. The margin is clearly part of the submerged prolongation of the landmass of Australia. Extension between the continents commenced in the Middle Jurassic (~165 Ma) with final breakup occurring in the Late Cretaceous (late Santonian – early Campanian, ~83 Ma; Sayers et al., 2001). Seafloor spreading was at very slow rates until about 40 Ma (Middle Eocene) at which time the spreading rate increased markedly, creating the deep-water Australian–Antarctic Basin that now separates Australia and Antarctica.

The GAB region has been explored by a number of petroleum exploration and scientific seismic surveys dating back to the late 1960s. Peaks of exploration activity occurred in the early 1970s, early 1980s, and since the late 1990s (e.g. Bein & Taylor, 1981; Longley, 2003). Less than 20 petroleum exploration and Ocean Drilling Program (ODP) wells have been drilled on the margin, with the most recent being drilled in 2003. The region has been of considerable scientific interest, particularly in terms of the rifting, breakup and spreading history between Australia and Antarctica (e.g. Talwani et al., 1978; Cande & Mutter, 1982; Willcox & Stagg, 1990; Stagg & Willcox, 1992; Tikku & Cande, 1999; Norvick & Smith, 2001).

GEOMORPHOLOGY

The continental margin of the GAB region is broadly concave in outline (Fig. 14). It is dominated by a wide continental shelf (the Eucla Shelf), two major mid- to upper-slope terraces (the Eyre Terrace in the west and the Ceduna Terrace in the east), and a broad lower slope province in the west (the Recherche Lower Slope). It is bounded to the south by the deep ocean floor of the Southern Australian Abyssal Plain (Figs 14 and 15).

West of about 129ºE, the Eucla Shelf is up to 120 km wide. Here the margin descends beyond the shelf break, at a depth of about 200 m, through the Eyre Slope and Recherche Lower Slope, to the Southern Australian Abyssal Plain. The Eyre Slope consists of a 10–15 km-wide upper slope province; the up to 70 km-wide Eyre Terrace, in water depths of 500–1500 m; and a lower province with average gradients of 1–4º that extends to a depth of about 3500 m where it merges with the lower-gradient Recherche Lower Slope. The Recherche Lower Slope is up to 180 km wide with gradients typically varying from 0.4–0.9º. It merges with the deep ocean floor at a depth of about 5500 m. As is discussed in the geology section below, seismic data clearly show that the Recherche Lower Slope is not a continental rise.

East of about 129ºE, the Eucla Shelf broadens to a width of about 200 km. It descends through the shelf break (lying at about 200 m) to the upper part of the Ceduna Slope province. The upper slope is narrow (<20 km wide) and merges with a major marginal plateau, the Ceduna Terrace, at a depth of about 1000 m. The Ceduna Terrace is up to 200 km wide and mostly lies at water depths of 1000–2500 m. Below the terrace, the lower part of the Ceduna Slope is relatively steep (typical gradients of ~1º to 3º). This part of the Ceduna Slope province abuts the deep ocean floor of the Southern Australian Abyssal Plain at a depth of about 5600 m.

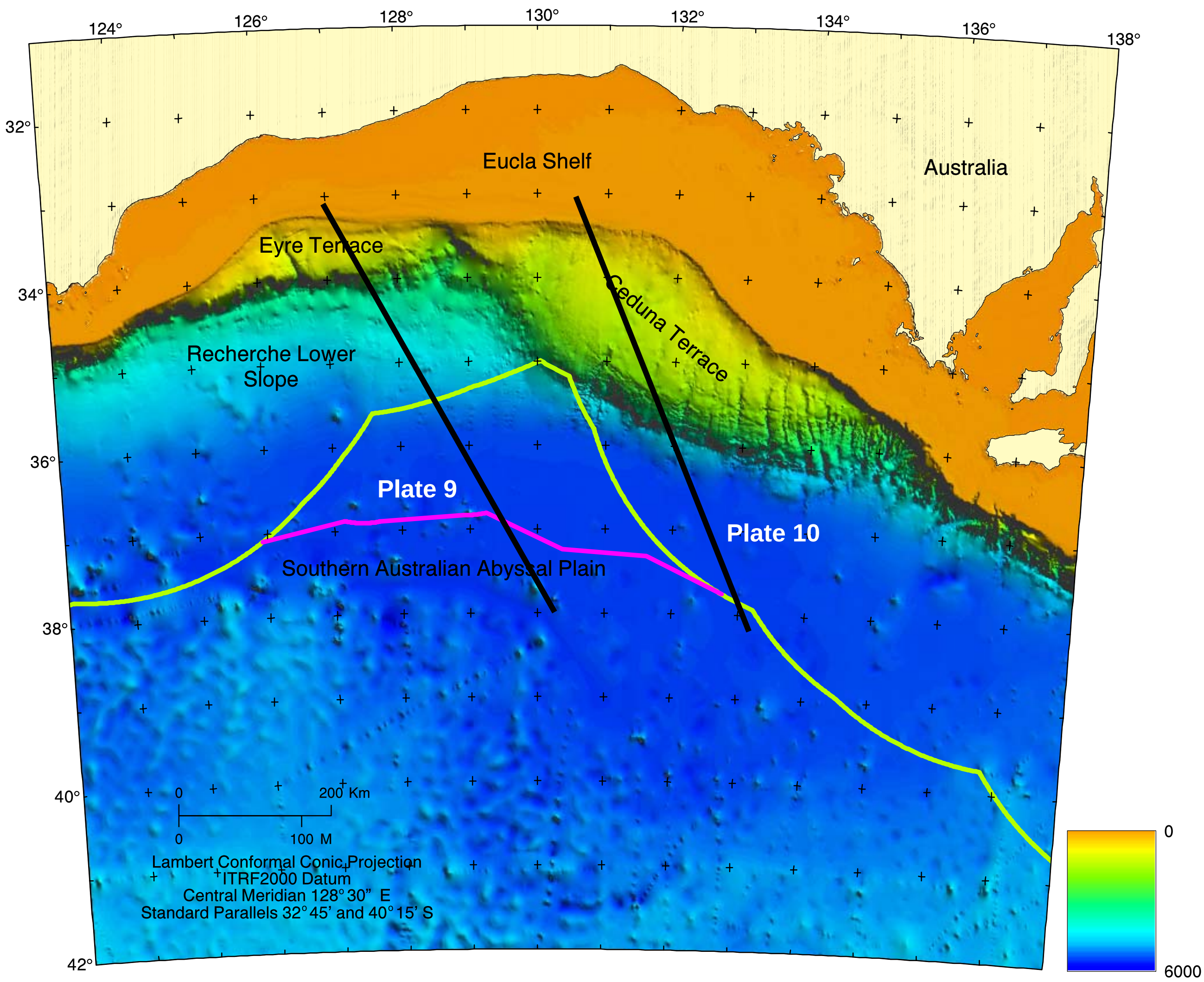

Figure 14: *Bathymetric image of the Great Australian Bight region, showing the location of the illustrated seismic transects. The major morphological features are labelled. Line colours: green – 200 M (Australia); magenta – outer limit of extended continental shelf (as submitted to UN CLCS, November 2004).*

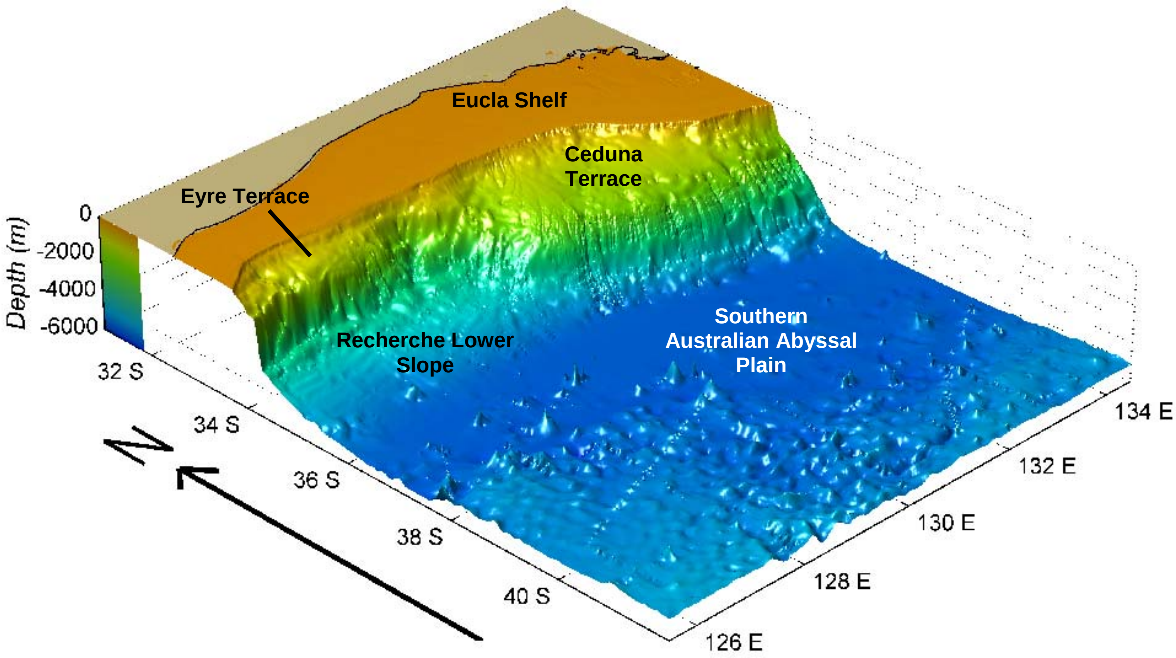

Figure 15: *3-D bathymetric image of the Great Australian Bight region viewed from the southwest.*

GEOLOGY

Plate tectonic setting

The continental margin of Australia in the Great Australian Bight region is the product of extension, breakup and seafloor spreading between Australia and Antarctica. Extension occurred along this part of margin from the Middle Jurassic (~165 Ma) to the Late Cretaceous (~83 Ma). This rifting, and the consequent thermal subsidence, produced the major depocentres of the Bight Basin: the Eyre, Ceduna, Duntroon, and Recherche Sub-basins (Sayers et al., 2003; Bradshaw et al., 2003; Figs 16–18).

A number of workers have identified magnetic anomalies in the Southern Ocean adjacent to Australia. These were first identified by Weissel & Hayes (1972) as indicating breakup in the Eocene, and were subsequently re-interpreted by Cande & Mutter (1982) who estimated breakup between Australia and Antarctica took place between 110 and 90 Ma; they included an initial period of very slow spreading. Veevers (1986) and Veevers & Eittreim (1988) further refined the estimate of the age of breakup along the southern margin of Australia to 95+/-5 Ma (Cenomanian – Turonian).

Tikku & Cande (1999) re-examined all magnetic data between Australia and Antarctica. They confirmed the presence of magnetic anomaly 34y and modelled breakup at 95 Ma. However, in an integrated interpretation of deep-seismic and potential field data, Sayers et al. (2001) showed that anomaly 34y was located on the crust of the continent-ocean transition zone rather than on unambiguous oceanic crust. They interpreted the first magnetic anomaly over oceanic crust to be 33o, indicating that Australian-Antarctic breakup in the GAB region occurred in the early Campanian (~80 Ma).

After breakup, ultra-slow seafloor spreading took place between Australia and Antarctica. Slow spreading continued until the Early Eocene when there was a marked increase in the spreading rate, coincident with a change in the spreading azimuth from broadly NW–SE to N–S and with the major global plate reorganisation (chron C18, Veevers et al., 1990; Tikku & Cande, 1999; Norvick & Smith, 2001).

Geology of the margin and adjacent areas

As shown in Figure 16, the continental margin in the GAB region is characterised by several major depocentres of the large (>800 000 km^2 area) Bight Basin. The structural and stratigraphic framework of these depocentres is detailed in Totterdell et al. (2000), Sayers et al. (2003), Bradshaw et al. (2003) and Totterdell & Bradshaw (2004). The Eyre, Ceduna and Recherche Sub-basins of the Bight Basin are located in the central GAB region (Fig. 16).

Eyre Sub-basin

The Eyre Sub-basin is a perched, Middle Jurassic and younger rift basin that mainly underlies the physiographic Eyre Terrace. It consists of a series of half graben that contain up to 3500 m of Mesozoic sedimentary rocks overlain by up to 1000 m of Cainozoic sediments. Although it is bounded by shallow basement to the north, south and west, the sub-basin fill is continuous with the Ceduna and Recherche Sub-basins to the east and south.

Ceduna Sub-basin

The east-southeast-trending Ceduna Sub-basin extends over an area of >120 000 km^2 and is the major Middle Jurassic to Late Cretaceous depocentre of the Bight Basin. It contains at least 15 000 m of syn-rift and post-rift Mesozoic strata and underlies the physiographic Ceduna Terrace.

Recherche Sub-basin

The Recherche Sub-basin lies in deep water beneath the lower slope of the GAB continental margin. It lies outboard of the Ceduna and Eyre Sub-basins and its strata are on-lapped by the Eocene and younger sediments of the Australian–Antarctic Basin. It has a maximum width of approximately 300 km immediately west of the Ceduna Sub-basin. The sub-basin contains up to 10 000 m of mainly flat-lying, ?Middle Jurassic to Early Cretaceous strata.

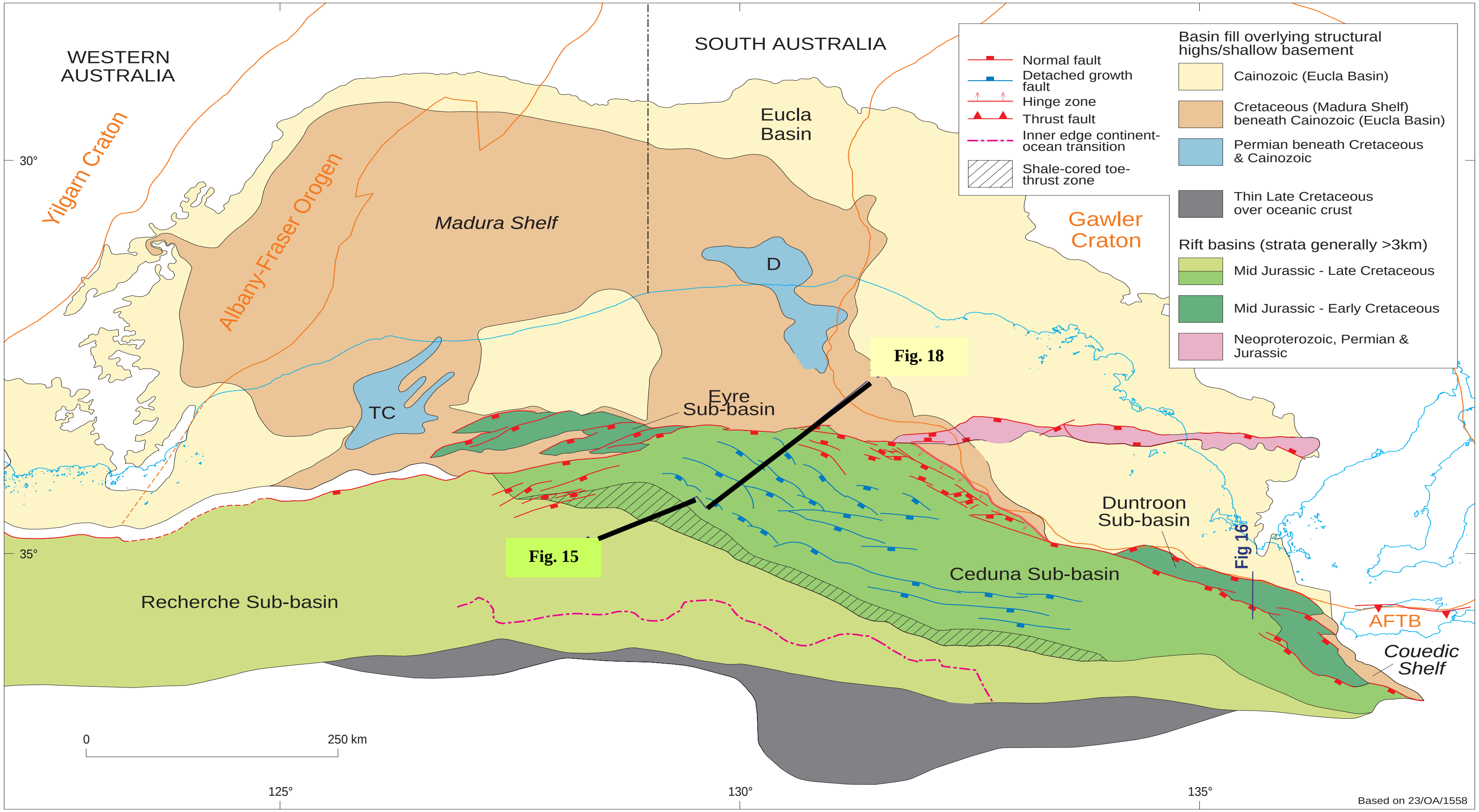

Figure 16: *Major structural elements of the Great Australian Bight region (after Bradshaw et al, 2003).*

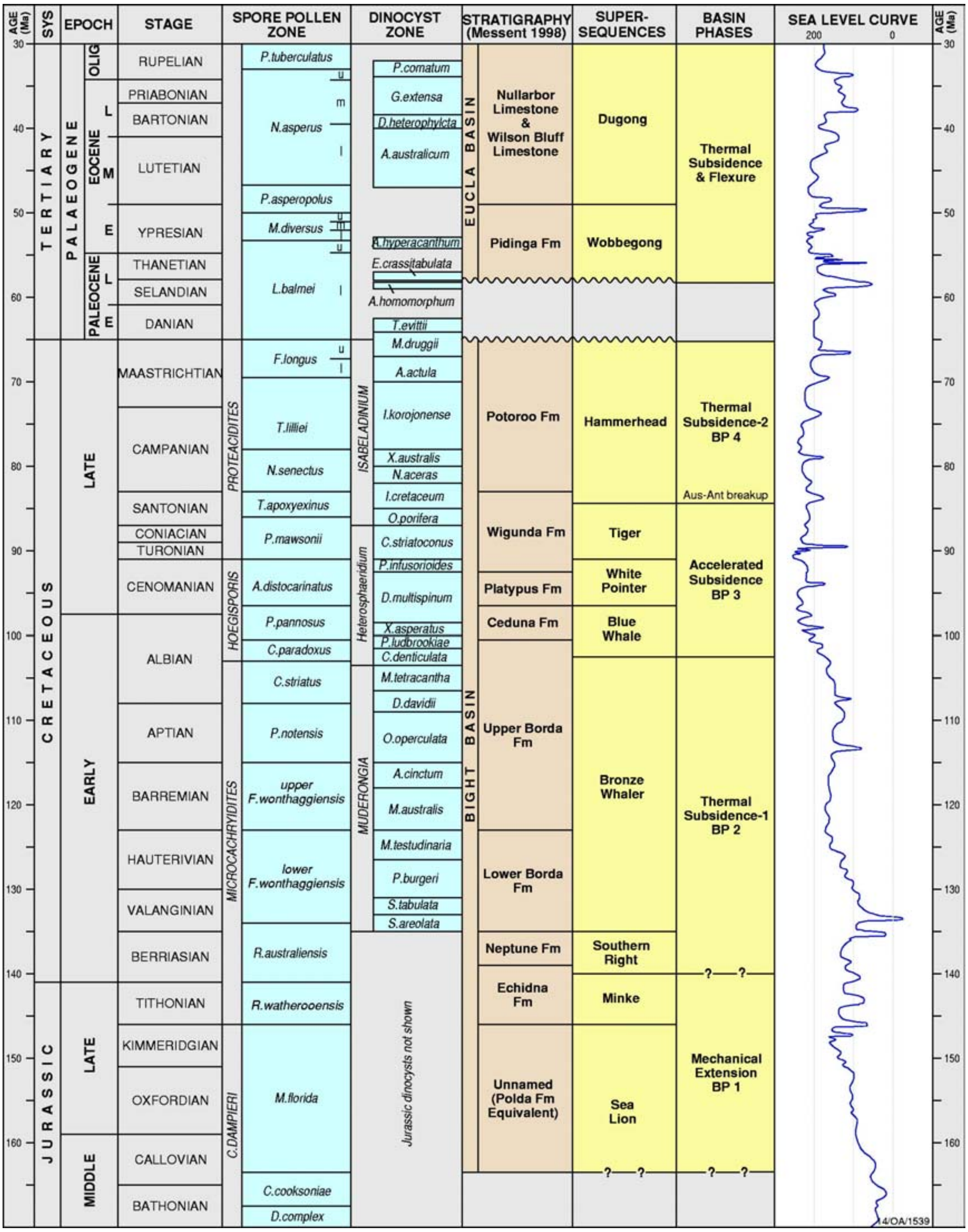

***Figure 17:** Stratigraphic correlation chart for the Bight Basin (after Totterdell & Krassay, 2003).*

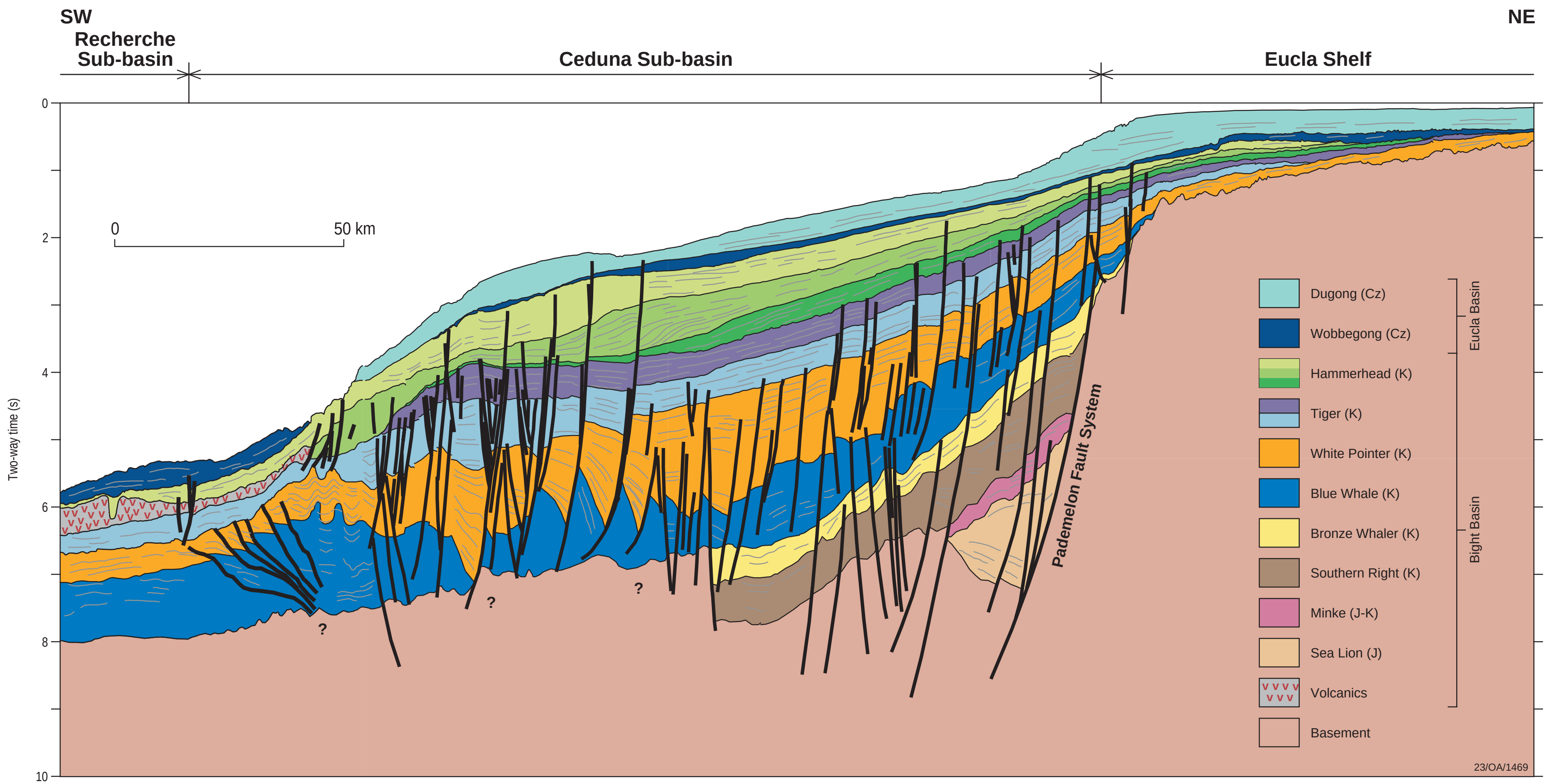

Figure 18: *NE–SW cross-section from the Eucla Shelf to the lower continental slope through the Ceduna and Recherche Sub-basins (after Bradshaw et al., 2003). Location of the section is shown in Figure 16.*

Deep-water continent–ocean transition zone

The geology of the deep-water continent-ocean transition zone (COT) in the central Great Australian Bight region has been interpreted by Sayers et al. (2001) by integrating the interpretation of deep-seismic reflection data with seismic refraction data and potential-field modelling (Fig. 19). They identified a 50–120 km wide zone between the outer edge of highly-attenuated continental crust and the inner edge of unequivocal oceanic crust as the continent-ocean transition (COT). This zone is characterised by a thin apron of post-breakup sediments overlying complexly deformed sediments and intruded crust bounded on its landward edge by a basement ridge complex and oceanward by the continent-ocean boundary (COB), beyond which is rugged oceanic crust. Figure 19 shows the modelled transect across the margin.

The COT is interpreted by Sayers et al. (2001) to be underlain by extended continental lithosphere. As noted above, in this interpretation, the continent-ocean boundary (COB, the outer limit of the COT) appears to be associated with magnetic anomaly 33o, indicating that breakup and seafloor spreading commenced at about ~80 Ma (early Campanian). This timing is more or less consistent with structural, stratigraphic and biostratigraphic evidence from the Ceduna Sub-basin, which suggests that breakup occurred in the latest Santonian (Totterdell & Bradshaw, 2004).

Based on its gravity and magnetic signature, the large basement ridge complex which characterises the inner flank of the COT is probably composed of a mixture of serpentinised peridotites and mafic intrusions and extrusions derived by mantle upwelling and limited partial melting. A similar basement ridge complex is seen on parts of the conjugate Antarctic margin (Colwell et al., 2006). Many of the features and relationships observed on the outer margin of the GAB can be explained in terms of extension within a lithosphere-scale “pure-shear” environment involving four layers: brittle upper crust and upper mantle, and ductile lower crust and lower lithospheric mantle (Sayers et al., 2001).

The widespread identification of the so-called magnetic anomaly 34 within the COT of the GAB may be due to the cooling of magmatic products injected into the COT zone as a result of the mantle upwelling and partial melting. This anomaly lies well inboard of the COB (Fig. 19).

Australian–Antarctic Basin

The Australian-Antarctic Basin sedimentary succession and onlaps oceanic crust and the rocks of the Recherche Sub-basin. This succession was deposited during a mostly continuous period of open-marine sedimentation after the onset of fast seafloor spreading in the Middle Eocene (Tottderdell et al., 2000; Sayers et al., 2003).

Geological history

Based on the identification of ten supersequences and three major phases of basin development, Totterdell et al. (2000; Fig. 17) described the evolution of the Bight Basin from a series of small, isolated intracratonic rifts to a major, complex passive margin basin. This history can be summarised as follows:

- In the Late Jurassic, upper crustal extension resulted in fluvial–lacustrine deposition in a series of isolated half graben.
- Subsequently a period of thermal subsidence led to the deposition of fluvio-lacustrine to restricted marine sediments in the Early Cretaceous.
- Accelerated subsidence in the Albian–Santonian led to the deposition of widespread marine facies across the region.
- In the Cenomanian, extensive growth faulting occurred; this was linked to decollements within over-pressured shales.
- The commencement of seafloor spreading in the latest Santonian – Early Campanian led to a second phase of thermal subsidence during which a thick Santonian–Maastrichtian succession was deposited in a vast progradational deltaic shelf margin system, particularly in the Ceduna Sub-basin.
- In the early Tertiary, a dramatic decrease in the sediment supply to the region resulted in the establishment of a sediment-starved shelf.
- An increase in the rate of seafloor spreading in the Middle Eocene led to accelerated margin subsidence and the development of an extensive carbonate shelf.

ILLUSTRATED TRANSECTS

Two transects are included to illustrate the geology of the continental margin in the Great Australian Bight (locations shown in Fig. 14):

Plate 9: (line GA-199/09) extends southeast from the shallow basement of the Eucla Shelf, across the Eyre and Recherche Sub-basins, to the oceanic crust of the Australian–Antarctic Basin.

Plate 10: (lines GA-199/08 and GA-199/11) extends southeast from the Eucla Shelf, across the Ceduna and Recherche Sub-basins, to the Australian–Antarctic Basin.

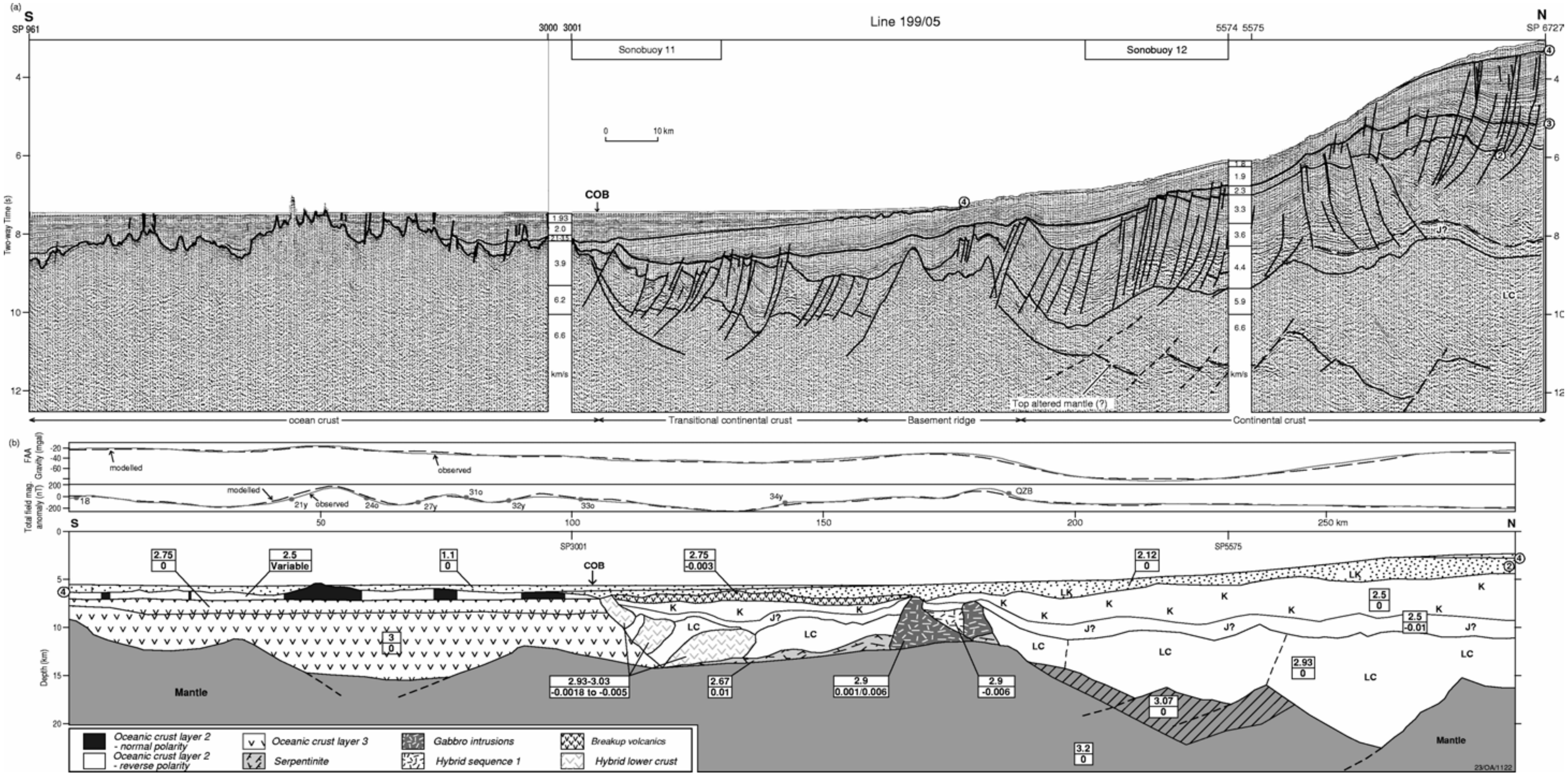

Figure 19: *Caption on next page.*

Figure 19: *(after Sayers et al., 2001)*

(a) *Interpreted seismic profile GA-199/05, showing sedimentary sequences and crust from the outer Ceduna Sub-basin, across the COT, COB and onto oceanic crust. Interpretation is based on seismic reflection data and refraction modelling. 1–4, Megasequences 1 (Upper Jurassic to uppermost Cenomanian); 2 (Turonian to Santonian); 3 (lower Campanian to Middle Eocene); and 4 (Middle Eocene to Recent). J?, possible Jurassic; LC, lower crust. Dashed subhorizontal line represents a possible sequence boundary roughly equivalent to the basal part of Megasequence 1 as modelled from potential field data. Sonobuoy refraction model velocities shown as vertical strips, positioned at one end of the shot range for display purposes.*

(b) *Potential field model for GA-199/05. Magnetic anomaly picks are based on Tikku & Cande (1999). Upper and lower values in the boxes are density ($t.m^{-3}$) and magnetic susceptibility (c.g.s.), respectively, for the associated layer. The normal polarity bodies used to model magnetic anomalies over oceanic crust have been placed where the data demand a change in magnetic properties. The bodies integrate a number of magnetic polarity changes associated with the slow-spreading crust. 'LC' between dashes represents possible modified lower crust. Megasequences 1–4 are shown; LK, Lower Cretaceous; K, Cretaceous; J?, possible Jurassic; LC, lower crust. Hachured area represents an area of probable altered mantle interpreted from modelled refraction velocities of 7.2–7.6 $km.s^{-1}$.*

KERGUELEN PLATEAU

INTRODUCTION

The Kerguelen Plateau region (Fig. 20) is dominated by the Kerguelen Plateau, a large, composite, mid-ocean, submarine plateau that developed during the extensional and magmatic breakup of Greater India, Australia and Antarctica. It forms the continental margin to Heard Island and McDonald Islands (Australia), and Îles Kerguelen (France). The plateau is located in the Southern Ocean and extends for approximately 2300 km from north-northwest to south-southeast, and is an average of 600 km in width. To the west, north and east, the plateau is abutted by deep ocean basins, while in the south it is separated from the Antarctic continent by the Princess Elizabeth Trough. The geology is dominated by major episodes of volcanism, although drilling and other evidence indicate that parts of it contain continental components, rather than being solely the product of magmatism.

The plateau and its surrounds have been explored by a number of French, Australian and US geophysical surveys since the 1970s. The single and multichannel seismic data sets acquired on these surveys have been supplemented by several key cores and dredges and by a number of Ocean Drilling Program (ODP) sites.

GEOMORPHOLOGY

The Kerguelen Plateau is a broad and elongate mid-ocean plateau that stands 2000-4000 m above the adjacent deep ocean basins (Figs 20, 21). To the west, northwest, northeast and southeast it is abutted by the Enderby, Crozet, Australian-Antarctic and Labuan Basins, while to the south it is separated from the Antarctic continent by the east-west trending, 3500 m-deep Princess Elizabeth Trough. The plateau consists of a number of morphologically diverse provinces that include the Northern, Central and Southern Kerguelen Plateau, Elan Bank and Williams Ridge.

The Northern Kerguelen Plateau (NKP) is the shallowest and morphologically most prominent part of the plateau, encompassing the northern quarter of the feature. It also contains the major land mass on the plateau – the Îles Kerguelen (France). Water depths over most of this sector range from less than 500 m to more than 1000 m, with the transition to the continental slope occurring at about 1500 m depth. The margins of the NKP are extensively incised by canyons and are disrupted by the westwards-protruding, 1000–2500 m-deep Western Banks area in the northwest and by minor east- and northeast-extending ridges in the northeast.

The Central Kerguelen Plateau (CKP) is the central province of the Kerguelen Plateau; it is bounded to the north by the NKP and to the south by the Southern Kerguelen Plateau (SKP); to the southwest, where it abuts the SKP, it is bounded by the Elan Bank, which is a westward salient of the CKP/SKP; and to the east it is bounded by the Williams Ridge, which is a continuation of a basement high that extends into the CKP. Both the Elan Bank and the

Williams Ridge are components of the Kerguelen Plateau. The CKP is flanked by the deep ocean floor of the Australia-Antarctic Basin to the northeast and the Enderby Basin to the southwest.

The CKP is surmounted by the Heard and McDonald Islands, both of which are land territories of Australia. The crest of the plateau lies generally at a few hundred metres depth, although Big Ben, the highest point on the volcanically active Heard Island, rises to over 2500 m above sea level. The CKP is geologically complex, with large-scale faulting, at least one significant sedimentary basin (the Kerguelen-Heard Basin), and evidence of sub-aerial erosion and basalt flows.

The Southern Kerguelen Plateau (SKP) encompasses the southern one-third of the Kerguelen Plateau. It is considerably deeper than the NKP and CKP, with crestal water depths slightly less than 1000 m (Fig. 20). The eastern margin of the SKP is steeper than its southern and western margins. Canyon development occurs on all the margins. The SKP is separated from the central Kerguelen Plateau by a broad, east–west trending saddle with minimum axial water depths of approximately 2500 m.

The Elan Bank is a submarine elevation that extends for approximately 400 km westwards from the CKP. Crestal water depths range from approximately 1000 m in the east to slightly deeper than 500 m to the west. It is separated from the main part of the Kerguelen Plateau by a saddle with axial water depths of slightly less than 3000 m. The southern flank of the bank is quite steep, whereas the northern and western flanks have more gentle gradients. The almost orthogonal orientations of Elan Bank and the Kerguelen Plateau suggest that the two features have different origins.

There has been substantial sediment accumulation in the basins that abut the SKP, especially to the south and east in the Princess Elizabeth Trough and Labuan Basin, respectively. In some areas this has resulted in the development of a continental rise, and especially on the eastern margin this rise has been modified by bottom currents forming a bounding moat and levee system on its inner edge.

The Kerguelen Plateau is flanked by the Enderby Basin to the west and by the Australian-Antarctic Basin to the northeast. These basins lie at water depths of 3500–5000 m and were formed by seafloor spreading between India and Antarctica and between Broken Ridge and the Kerguelen Plateau. The generally flat seafloor in each of these basins is frequently disrupted by seamounts and ridges that rise from a few hundred to more than 1500 m above the adjacent seafloor. While the Labuan Basin to the southeast generally lies at water depths greater than 4500 m, its origins are less clear.

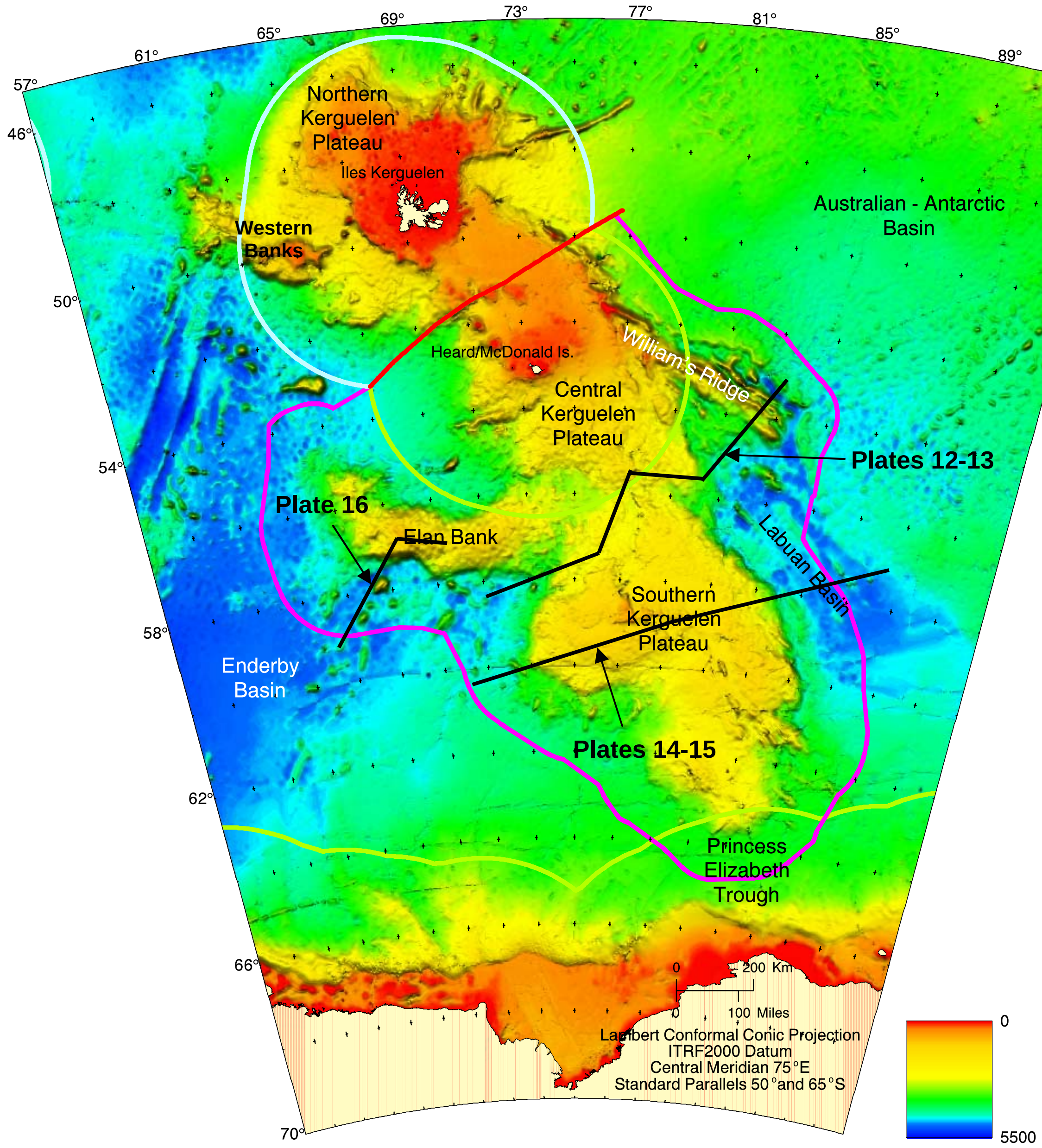

Figure 20: *Bathymetric image of the Kerguelen Plateau region, showing the location of the illustrated seismic transects. Major morphological features are labelled. Line colours: green – 200 M (Australia); light blue – 200 M off Kerguelen (France); magenta – outer limit of extended continental shelf (as submitted to UN CLCS, November 2004)*

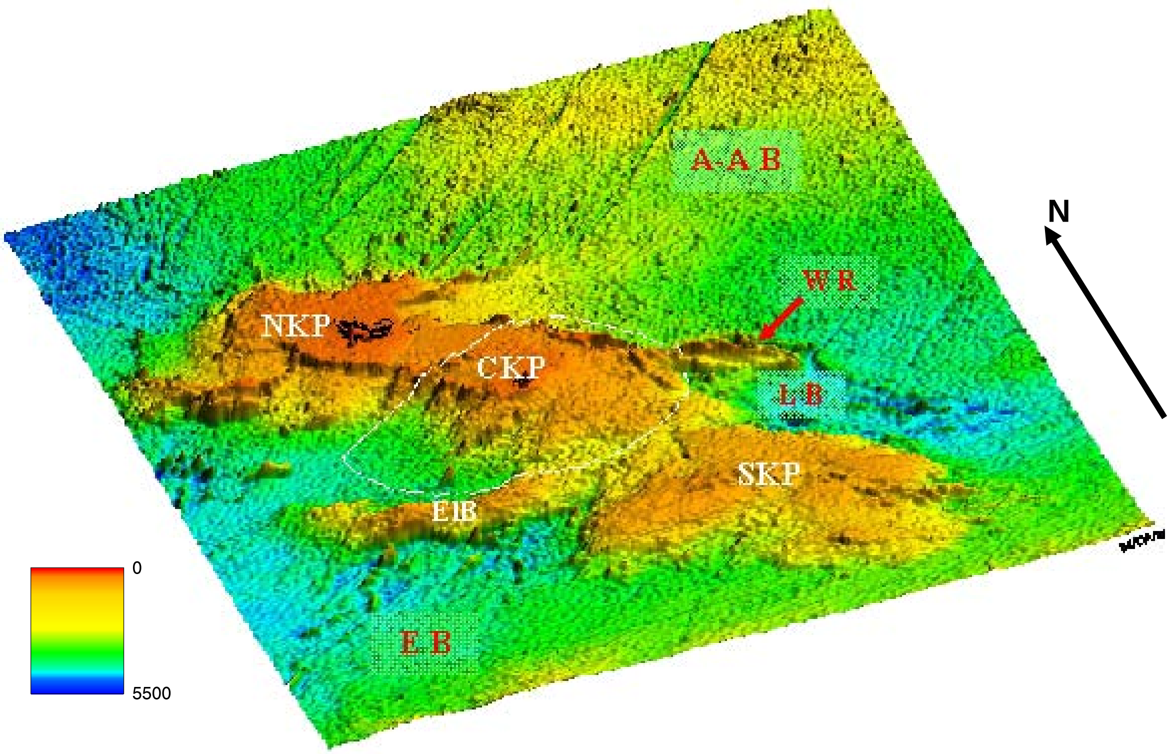

Figure 21: *3-D bathymetric image of the Kerguelen Plateau region viewed from the southwest. NKP – Northern Kerguelen Plateau; CKP – Central Kerguelen Plateau; SKP – Southern Kerguelen Plateau; ElB – Elan Bank; LB – Labuan Basin; WR – Williams Ridge; EB – Enderby Basin; A-A B – Australian-Antarctic Basin. White line is 200 M line.*

GEOLOGY

Plate tectonic setting

The tectonic history of the Kerguelen Plateau is closely linked to the breakup of India, Australia and Antarctica. Much of the plateau formed during the Barremian–Albian (Coffin, Frey, Wallace et al., 2000) as a large igneous province (LIP) over the Kerguelen Plume.

The plateau is located on the Antarctic Plate, south of the junction between the Southwest Indian and Southeast Indian spreading ridges, and is flanked by mainly oceanic crust of variable ages. The oldest magnetic lineations are in the Enderby Basin, southwest of Elan Bank. Here, Brown et al. (2003a) have identified anomalies M10 to M2 (~130 Ma to ~124 Ma; Valanginian–Hauterivian) that date the first separation of Greater India (and Elan Bank) from East Antarctica. Ishihara et al. (1999) have also identified the slightly older anomalies M12 to M10 in the Princess Elizabeth Trough between the southern Kerguelen Plateau and Antarctica. However, this is a geologically complex area with the development of thick seaward-dipping reflector sequences (SDRS; Stagg et al., in press), and it is possible that this identification is in error. In the Crozet Basin, northwest of the Kerguelen Plateau, anomalies A34 to A23 (Royer & Sandwell, 1989) have been identified, which date this crust as mid- to Late Cretaceous, comparable in age to the earliest spreading between Australia and Antarctica. The northeast flank of the Northern Kerguelen Plateau and Williams Ridge lie adjacent to crust that ranges in age from Late Eocene (A18) to Oligocene (A11) (Cande & Mutter, 1982). This crust was generated during the fast-spreading phase between the Australian and Antarctic Plates. No magnetic lineations have been identified in the Labuan Basin in the southeast, and the origins of this basin remain enigmatic.

Geology of the Kerguelen Plateau

The geology of the Kerguelen Plateau and the Labuan Basin are illustrated by a map of the tectonic provinces (Fig. 22), by transects of the central and southern parts of the plateau, from the Enderby Basin in the west to the Australian–Antarctic Basin in the east (Figs 23, 24), and by a chart of the stratigraphy of the region (Plate 11)..

Crustal structure

Because of its remote location and the low number of surveys that have been undertaken, the crustal structure of the Kerguelen Plateau remains poorly understood. Ocean Drilling Program (ODP) drilling revealed that the plateau is underlain mostly by magmatic crust generated in the Barremian-Cenomanian (119–95 Ma) by excessive volcanism attributed to a large hotspot (Coffin, Frey, Wallace et al., 2000). The plateau has been described as a Large Igneous Province or LIP (Coffin & Eldholm, 1994). However, an entirely magmatic origin for the plateau has been disproved by ODP drilling results on the Elan Bank (Site 1137, Leg 183; Coffin, Frey, Wallace et al., 2000). Gneissic metamorphic and felsic igneous clasts recovered in a volcanoclastic conglomerate deposited in a fluvial environment provided unambiguous evidence of the presence of some continental crust. Together with geochemical evidence that points to the presence of continental lithosphere beneath the southern part of the plateau (Storey et al., 1992; Mahoney et al., 1995), this sampling indicates a more significant involvement of continental crust in the foundations of the plateau than was previously assumed. It seems likely that the plateau is floored by a combination of different crustal blocks, including continental fragments and magmatic crust.

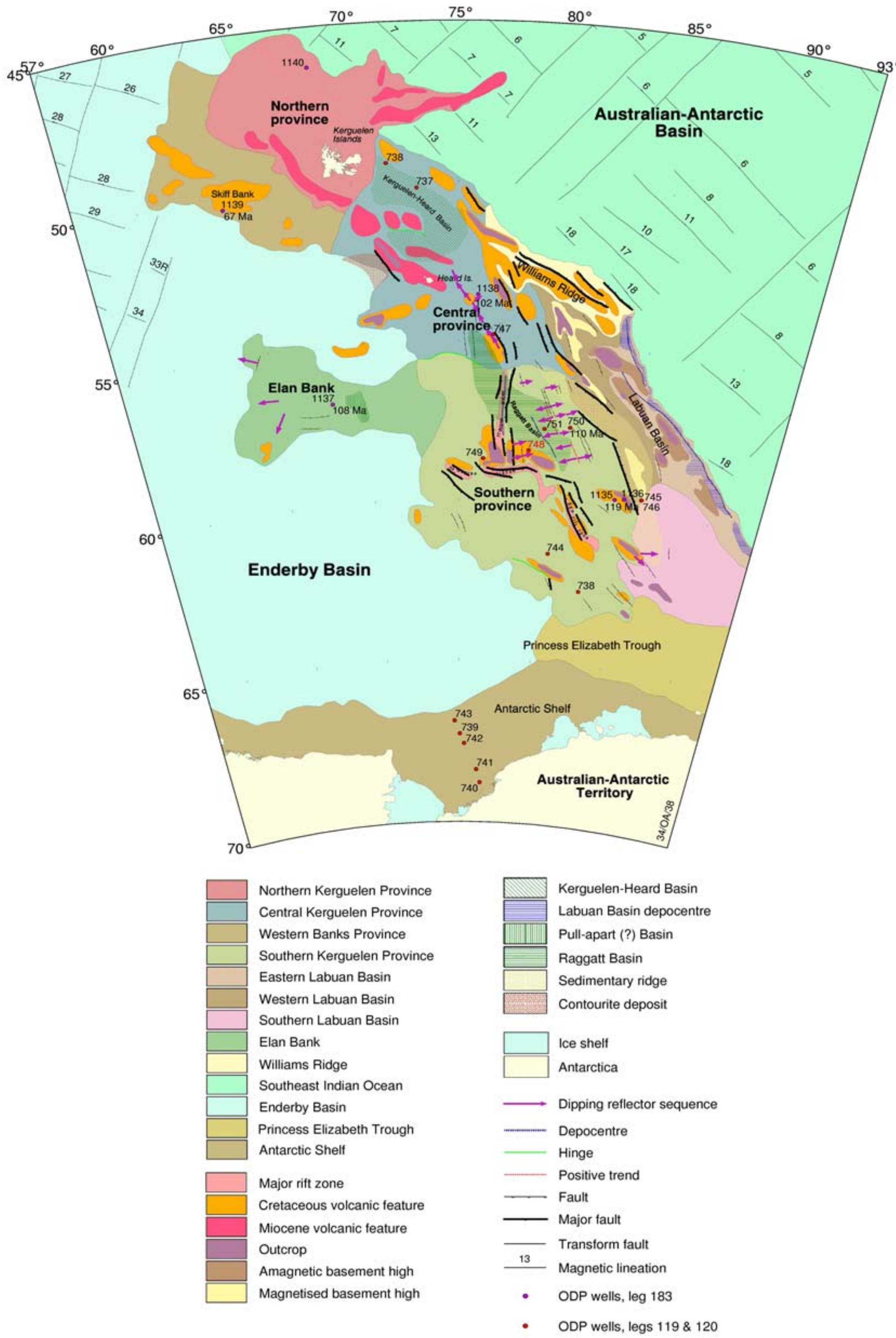

***Figure 22:** Tectonic provinces of the Kerguelen Plateau and adjacent areas (after Borissova et al., 2002).*

Possible continental origins for substantial parts of the plateau are also supported by integrated potential field modeling and deep-seismic interpretation (Direen, 2004b) and by seismic refraction models of the southern Kerguelen Plateau (Operto & Charvis, 1995, 1996).

Northern Kerguelen / Western Banks Province

Most of the Northern Kerguelen Province (45–50ºS; Fig. 22) was formed in the Oligocene-Miocene and it includes the volcanically active Kerguelen Archipelago. ODP drilling on the Skiff Bank (ODP Site 1139; Coffin, Frey, Wallace et al., 2000) recovered Late Cretaceous trachytes, which may indicate a wider presence of Cretaceous basement beneath the northern part of the plateau.

Central Kerguelen / Williams Ridge Province

The Central Kerguelen Province (CKP; 50–55ºS; Figs 22, 23) includes the volcanically active Heard and McDonald Islands and contains a major sedimentary basin, the Kerguelen-Heard Basin.

Basement rocks at ODP Site 747, in the southern part of the CKP, comprise basalts erupted at about 85–88 Ma (Cenomanian; Munschy et al., 1992), while Site 1138, located about 150 km southeast of Heard Island, reached basement in dacitic lavas and pyroclastic flow deposits of trachyte (Coffin, Frey, Wallace et al, 2000).

The Kerguelen–Heard Basin is a sag basin in the northern half of the CKP, more than 40 000 km^2 in area and containing more than 2 000 m of sediments that range in age from Albian to Pleistocene (Borissova et al., 2002). ODP drilling indicates that the basal sediments contain wood fragments, leaves and fern fronds indicative of regolith formation (Coffin, Frey, Wallace et al., 2000). Shallow neritic conditions were established by the Cenomanian/Turonian (~99–89 Ma) and the Late Cretaceous and Cainozoic section was deposited in an open marine environment. A prominent seismic reflector within the basin appears to be caused partly by a diagenetic 'front' (Borissova et al., 2002) associated with a Late Paleogene to Early Neogene unconformity. Deformation of the section above this unconformity is the result of syn-depositional bottom current activity during the Neogene.

Williams Ridge is the southeast extension of a prominent basement ridge mapped beneath the eastern part of the CKP. Williams Ridge is underlain by 12–15 km thick crust (Gladczenko & Coffin, 2001), which is similar in thickness to that beneath the Kerguelen Plateau. A basement high on the northward structural continuation of Williams Ridge has been sampled in several locations. MD48 dredge 8 recovered Miocene saprolitic basalt, and MD35 cores recovered Late Cretaceous sediments (Fröhlich, 1983).

Southern Kerguelen Province

Volcanic basement beneath the Southern Kerguelen Province (SKP; south of 55ºS) is of Aptian–Albian age (119–110 Ma; Coffin et al., 2002). Three major rift systems have been mapped in this province. The 77º graben extends N–S for more than 400 km; the 59º graben trends approximately E–W, to the south of the 77º Graben; and the Southern Kerguelen Plateau rift zone trends NW–

SE, to the south of the 59º Graben. The age and origin of these rift systems remain unclear. It is estimated that 77º Graben was formed in the latest Cretaceous (~75 Ma); however, faults adjacent to the rift have been reactivated at different times with the latest movements taking place as late as the Miocene (Borissova et al., 2002).

The Raggatt Basin is a major 'sag' basin in the northern part of the SKP (Colwell et al., 1988; Fig. 24). The basin occupies an area of approximately 58 000 km^2, and contains at least 2000 m of Cainozoic sediments. Seismic data show that the eastern flank of the basin is underlain by an igneous ridge. This ridge is characterised by sequences of dipping intra-basement reflections, which can be traced for up to 120 km westwards beneath the Raggatt Basin. This dipping reflector sequence is at least 2 s TWT (ca 4–5 km) thick and the reflections probably represent sub-aerial lava flows.

In the southern Raggatt Basin, prominent mound features can be seen in the Late Cretaceous–Paleocene section. ODP Site 748 intersected upper Campanian and Maastrichtian biogenic carbonates at the same stratigraphic level, which suggests that the mounds are likely to have a biogenic rather than a volcanic origin (Borissova et al., 2002). Seismic interpretation indicates that these mounds continued growing until the Paleocene.

Elan Bank

Seismic, gravity and bathymetry data indicate that Elan Bank consists of two large basement highs displaced laterally along a NW-SE trending structure in the central part of the bank (Borissova et al., 2003). ODP Site 1137, drilled on the central part of the bank, reached basement and recovered 153 m of mainly brecciated and massive basalts, dated at 109.3 Ma (Coffin et al., 2002). Volcaniclastic conglomerates in basement contain clasts of garnet-biotite gneiss (Nicolaysen et al., 2000) indicating that Elan Bank contains continental crustal components. Basement is overlain by Campanian packstone and Late Eocene to Pleistocene pelagic oozes.

Volcanic sequences on Elan Bank may have formed either during the Valanginian breakup of India/Elan Bank and Antarctica, or during the later Albian breakup of India and Elan Bank, when the bank was transferred from the Indian to the Antarctic plate via a ridge jump (Gaina et al., 2003). In either case, massive Albian volcanism has overprinted and radically altered the continental sliver forming the core of the Elan Bank.

Labuan Basin

The Labuan Basin is located along the eastern margin of the Kerguelen Plateau and generally contains 2.5 to 4 km of sediment above rugged faulted basement. In the north it is separated from the Australian–Antarctic Basin by Williams Ridge (Fig. 22). The eastern boundary is delineated by a deep (9 s TWT), sediment-filled basement trough containing up to 5 km of sediments which separates the Labuan Basin from the oceanic crust of the Australian-Antarctic Basin.

Analysis of seismic data shows significant differences in basement character between the eastern, western and southern parts of the basin. The western part of the Labuan Basin is 130–150 km wide and is extensively faulted. The dominant fault-style is extensional, with planar normal faults typically dipping to the SW and resulting in a series of tilted fault blocks. The eastern Labuan Basin is about 120 km wide and is characterised by large NNW-trending basement highs. The appearance of these blocks is different from the faulted blocks of the western part of the basin in that they are usually larger and dome-shaped. Analysis of magnetic anomalies over these basement blocks shows that most of them do not have a magnetic signature typical of intrusions. One of these blocks in the northern part of the basin was dredged in 1991 (Montigny et al., 1993) and yielded 1.5 tonnes of metamorphic and granitic rocks interpreted as ice rafted debris. However, if these rocks are representative of Labuan Basin basement, rather than being ice-rafted, then the absence of a magnetic signature over the blocks is readily explained.

Direen (2004a) has produced potential field models for the Labuan Basin and eastern flank of the Kerguelen Plateau that are integrated with interpretation of the deep seismic data. This modelling suggests:

- There is a possibility of some oceanic crust in both the eastern and western Labuan Basins, although the exact nature of the crust underlying much of the basin remains problematic.
- The eastern and western Labuan Basins are separated by a dense, magnetised peridotite ridge.

The area shown in Figure 22 as the southern Labuan Basin is probably floored by oceanic crust comparable in age to that in the Enderby Basin (Stagg et al., 2005b, in press). This area was renamed the Shackleton Basin by Stagg et al. (2005b) to remove any implication that it is genetically related to the Labuan Basin located to the north.

Only two wells have been drilled in the Labuan Basin, both of which sampled the uppermost part of the sedimentary section; the stratigraphy of the Mesozoic and Early Cainozoic section remains unknown. ODP Sites 745 and 746 were drilled on the western margin of the Labuan Basin (Barron, Larson, et al. 1989) near the crest of the East Kerguelen Sedimentary Ridge.

ILLUSTRATED TRANSECTS

Three transects are included to illustrate the geology of the Kerguelen Plateau region (locations shown in Fig. 20):

Plates 12–13: (line GA-179/07) extends northeast from the northeast corner of the Enderby Basin, across the Southern and Central Kerguelen Plateau, the northern end of the Labuan Basin, and Williams Ridge, to the oceanic crust of the Australian–Antarctic Basin.

Plates 14–15: (lines GA-47/29 & GA-47/33) extends northeast from the Enderby Basin, across the Southern Kerguelen Plateau, the Labuan Basin and Williams Ridge, to the oceanic crust of the Australian–Antarctic Basin.

Plate 16: (lines GA-179/05 & GA-179/06) extends northeast from the northern Enderby Basin on to the crest of the central part of the Elan Bank and then runs eastwards along the crest of the bank.

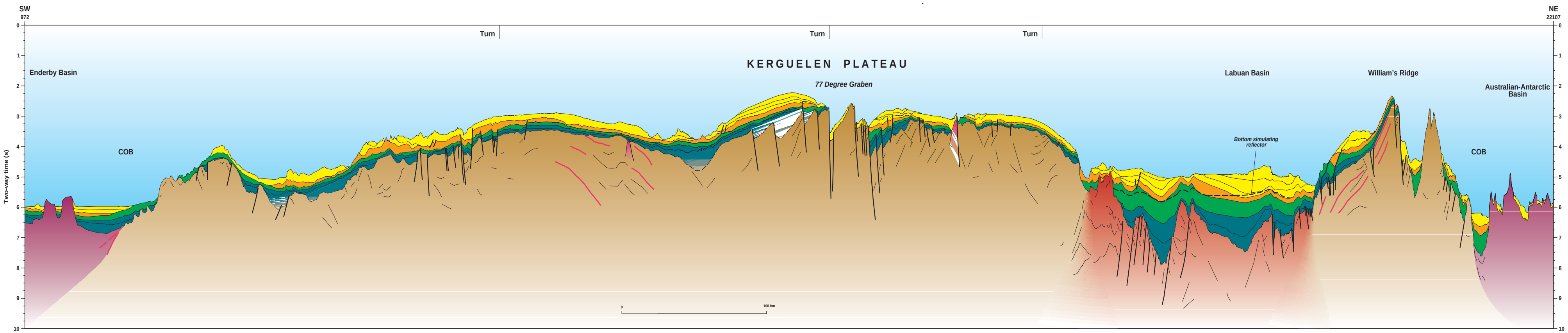

Figure 23: *Central Kerguelen Plateau transect, line GA-179-07 (after Borissova et al., 2002).*

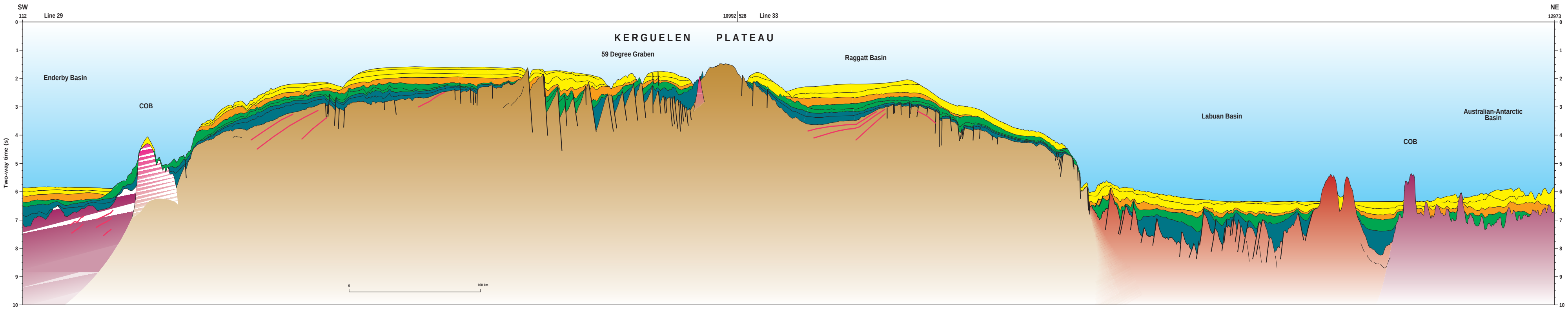

Figure 24: *Southern Kerguelen Plateau transect, lines GA-47-29+GA47-33 (after Borissova et al., 2002).*

LORD HOWE RISE

INTRODUCTION

The Lord Howe Rise region is dominated by the Lord Howe Rise, a large submerged continental plateau. This feature extends for more than 1600 km northwards from the Challenger Plateau, northwest of New Zealand, to southwest of New Caledonia and east of the Marion Plateau, in water depths generally greater than 1000 m. It is the western-most element in a broad zone of north- and northwest-trending continental plateaus, ridges (including the Norfolk Ridge and Three Kings Ridge), and narrow depressions, located between the deep ocean floors of the Tasman Basin in the west and the South Fiji Basin in the east. Lord Howe Rise is underlain by continental crust which, in the south, completely detached from eastern Australia during the continental margin break-up that led to the formation of the Tasman Basin from about 84–52 Ma (Gaina et al., 1998).

Since the early 1970s, the Lord Howe Rise has been explored by a number of regional seismic surveys carried out by Australia, France, New Zealand and the USA. Sample data are scarce, being limited to some sites drilled by the Deep Sea Drilling Program (DSDP) and a small number of dredge and core samples.

GEOMORPHOLOGY

The Lord Howe Rise (Fig. 25) extends from the eastern Coral Sea in the north, to the Bellona Trough in the south. The Bellona Trough separates the Lord Howe Rise from the Challenger Plateau which abuts the South Island of New Zealand. At about 2500 km in length and with a width of 450–650 km, the combined Lord Howe Rise – Challenger Plateau province has a total area of about 1 500 000 km^2, comparable to the area of Australia's North West Shelf and adjacent continental margin.

The NW-trending northern segment of Lord Howe Rise merges with a region of complex topography, which includes the Chesterfield Plateau and the NW-trending Fairway Ridge (Dubois et al., 1974; Fig. 25). The central segment of the rise trends north–south for some 1100 km from about 24–34ºS. The southernmost segment of the rise again trends NW-SE, and is separated from the Challenger Plateau adjacent to the New Zealand continental margin by the 3000 m deep, NNE-trending Bellona Trough at about 39ºS. In the north, there is morphological connection between the Kenn/Chesterfield/Mellish plateau complex and the northeast Australian continental margin across the 2000–3000 m deep Cato zone – an area of terraces and troughs that includes the 3000 m deep Cato Trough. The Cato zone forms an 800 km long complex saddle that lies 1500 m shallower than the deep ocean floors of the >4500 m deep Coral Sea and Tasman Basins, to the north and south respectively.

Excluding the islands and banks of the N-S trending Lord Howe seamount chain on the western flank of the rise, Lord Howe Rise is shallowest in the east where crestal water depths generally range from 1000–1500 m. The western flank of the rise is morphologically complex. Along the N–S trending, central segment of the rise, the Dampier Ridge (crestal water depths of 2000–2500 m) is partly separated from the main part of the Lord Howe Rise by the 3000–4000 m deep Lord Howe and Middleton Basins. In the south, the western flank of the rise is formed by the NW-trending Monowai Sea Valley, at about 3000 m depth, and the parallel and slightly shallower Monowai Spur to the southwest.

Lord Howe Rise is separated from the eastern Australian margin by the Tasman Basin (Fig. 25). The Tasman Basin is triangular in outline, narrowing from about 1100 km in the south, between Tasmania and New Zealand, to about 150 km in the north, southeast of the Marion Plateau. The basin floor lies at water depths of 4500–5100 m, and the topography varies from flat and smooth to extremely rugged, depending on the presence of fracture zones. The Tasman Sea is notable for the presence of the linear N–S trending Tasmantid seamount chain, which extends from the northern Tasman Sea to about 37ºS, where it appears to change trend to the SSW. These seamounts have relief of up to 2500 m above the seabed and some of them have planated summits (guyots), indicating that they have been subaerially exposed in the past.

To the east of Lord Howe Rise, lies the sub-parallel New Caledonia Basin. This basin extends for about 2000 km from the continental margin of the North Island of New Zealand to west of New Caledonia. As with the Lord Howe Rise, there are NW–SE trending segments in the north and south, separated by a N–S trending central segment. The basin has strong linearity, with an average width of about 150 km. The floor of the basin is generally very flat-lying, at a depth of about 3000 m.

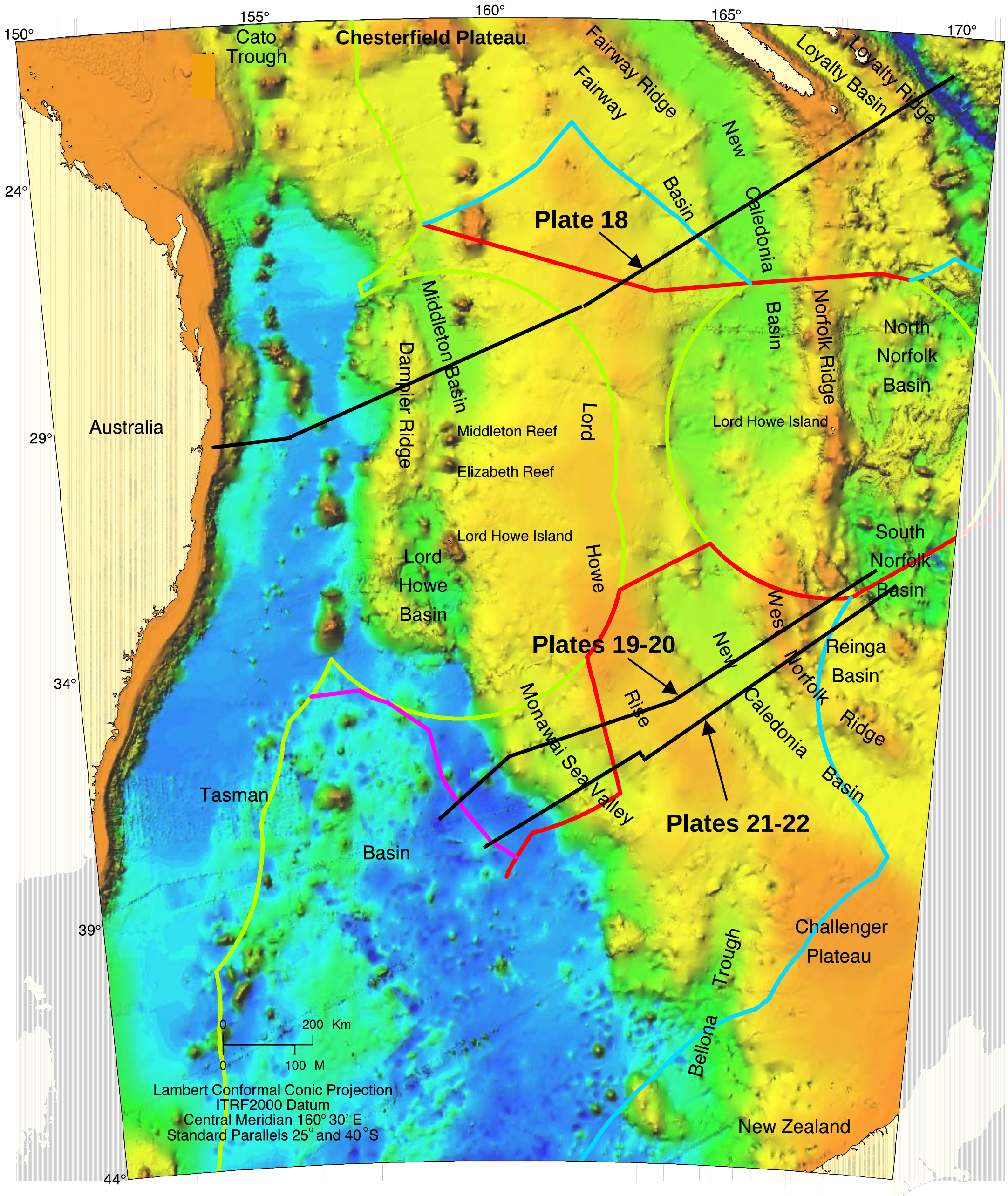

Figure 25: *Bathymetric image of the Lord Howe Rise region, showing the location of the illustrated seismic transects. The main geomorphological features are labelled. Line colours: green – 200 M (Australia); magenta – outer limit of extended continental shelf (as submitted to UN CLCS, November 2004); light blue – 200 M (New Zealand and France); red – treaty boundaries.*

GEOLOGY

Plate tectonic setting

Separation of the Lord Howe Rise 'continental ribbon' from mainland eastern Australia took place in the Late Cretaceous (Early Campanian, chron 33) by seafloor spreading in the Tasman Sea (Hayes & Ringis, 1973; Weissel & Hayes, 1977). Gaina et al. (1998) have revised plate reconstructions for the opening of the Tasman Basin based on a compilation of all the available magnetic and satellite gravity data. Their interpretation is considerably more complex than previous work, and is modelled using thirteen newly identified microplates that were active during breakup and spreading. Movement of microplates was accompanied by changes in spreading azimuth, and involved the failure of several rifts. In addition, Norvick et al. (2001) suggest that seafloor spreading in the Tasman Sea was preceded by the propagation of a strike-slip fault that gradually worked its way northwards, approximately parallel to Australia's eastern margin.

Geology of the Lord Howe Rise & adjacent areas

Introduction

Lord Howe Rise (LHR) extends for more than 1600 km from the Bellona Trough in the south to the Chesterfield Plateau in the north. The regional geology is illustrated here by reference to tectonic elements maps of the southwest Pacific (Fig. 26) and by a geological cross-section of Lord Howe Rise (Fig. 27). The stratigraphy of the region is illustrated in Plate 17.

Lord Howe Rise is from 400–500 km in width and comprises four provinces that are sub-parallel to the trend of the LHR and that extend for most of its length (Fig. 26). From east to west, these provinces comprise:

1. Shallow, planated, probable Palaeozoic basement of the Lord Howe Platform, overlain by a few hundred metres of mainly Cainozoic siliceous and carbonate oozes. To the east, the basement of the New Caledonia Basin is about 4 km deeper and of uncertain crustal affinity, while the western boundary of the platform is defined by a sharp Cretaceous hinge.
2. The Central Rift province, adjacent to the Lord Howe Platform and characterised by a series of poorly-defined basement blocks, normally down-faulted to the west, with 2–4 km of Upper Cretaceous and Cainozoic syn- and post-rift section. This province includes the Moore Basin in the south and the Faust Basin in the north.
3. The Western Rift province, which is separated from the central rift by a broad fault zone across which basement is down-faulted to the west. Basement and water depths are considerably deeper than in the central rift, and the syn- and post-rift sediments are considerably thicker. This province includes the Monowai Basin in the south and the Capel Basin in the north. In the vicinity of Lord Howe Island, the central and western rifts

cannot be separated, and the combined rift has been referred to as the Gower Basin.

4. A western bounding complex ridge system of known continental origin. In the north, the Dampier Ridge is separated from the western rift province by the Lord Howe and Middleton Basins, which may be underlain by highly extended lower continental crust. Further south, where there has been less extension, the Monowai Ridge forms an intact outer margin to the Monowai Basin.

Volcanic rocks of Cretaceous and Tertiary age form a significant component of the geology of Lord Howe Rise, the most conspicuous of these being the large seamounts and guyots of the Lord Howe seamount chain along the western part of the feature. These volcanic rocks reflect important geological processes that have continually modified and added to the heterogeneous continental crust of Lord Howe Rise throughout much of its development as a continental margin.

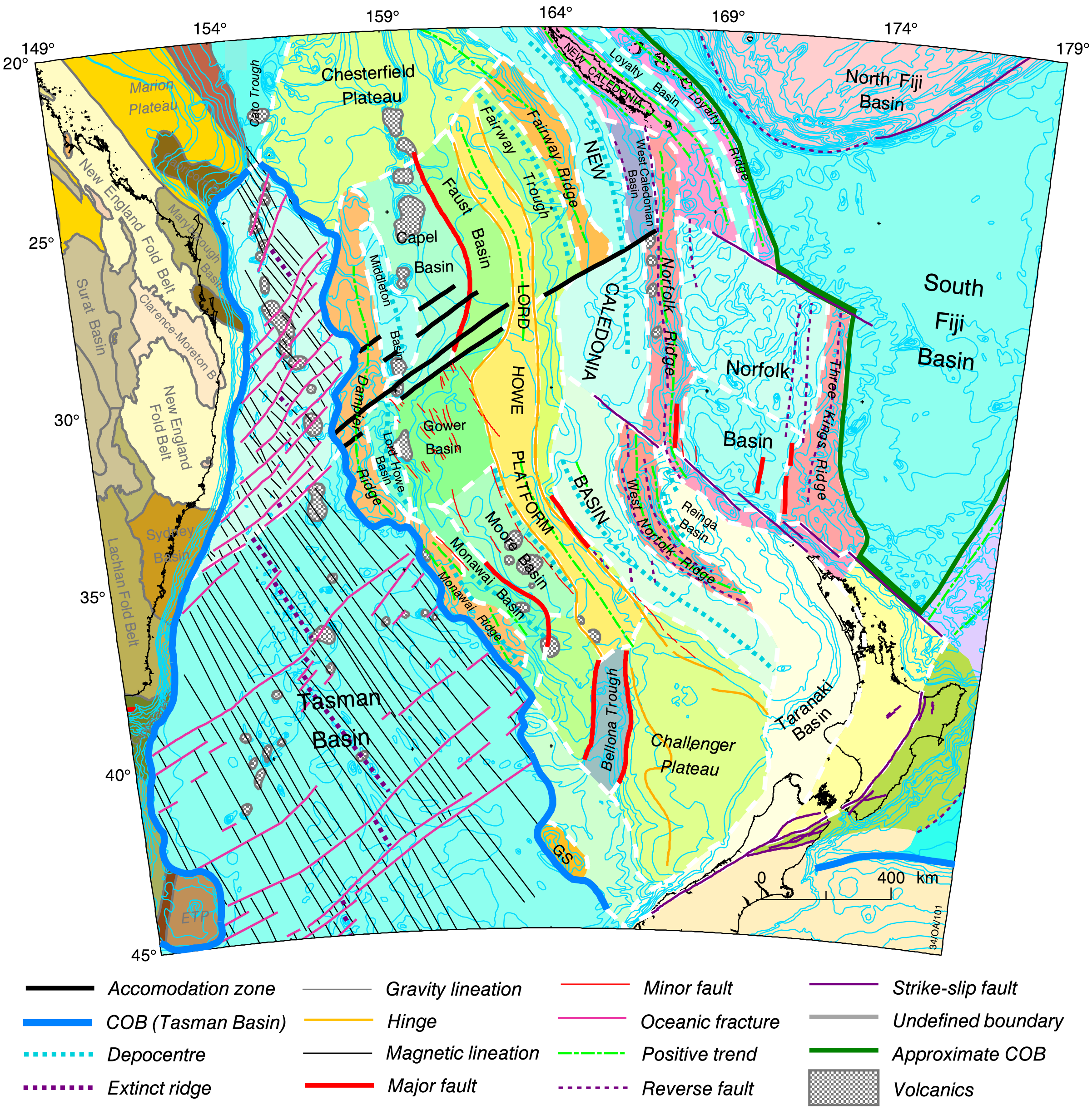

Figure 26: *Map showing the distribution of tectonic provinces and features in the southwest Pacific (after Stagg et al., 2002).*

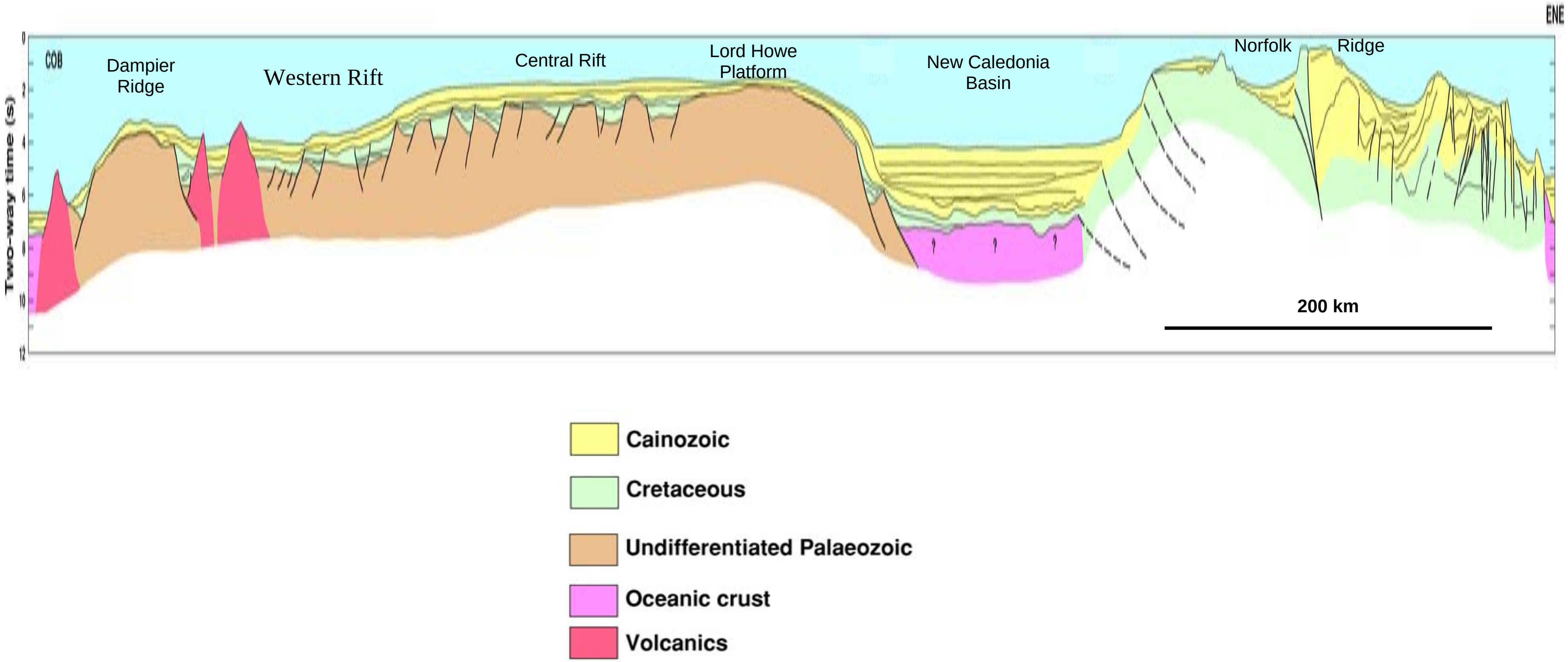

***Figure 27:** Geological cross-section across the southern Lord Howe Rise (after Stagg et al., 1999).*

Lord Howe Platform

The Lord Howe Platform generally underlies the shallowest part of Lord Howe Rise in water depths from 1000–1300 m, and extends from the boundary with the Challenger Plateau at the Bellona Trough at about 37ºS, northwards to at least 23ºS, a distance of at least 1600 km. The average width of the platform is less than 100 km along most of its length, with a maximum width of about 200 km east-northeast of Lord Howe Island.

The internal seismic character of the basement suggests that it comprises in part Palaeozoic fold belt sediments and meta-sediments, correlating with either New England or Lachlan Fold Belt rocks of eastern Australia. DSDP Site 207, at the southern end of the province, recovered Cenomanian rhyolitic lapilli tuffs and vitrophyric rhyolite flows from within the acoustic basement (Burns, Andrews et al., 1973; McDougall & van der Lingen, 1974). These volcanics cannot be distinguished from the interpreted Palaeozoic basement and their areal extent and thickness is uncertain. Refraction velocities in the range of 5.26–5.95 km.s^{-1} are consistent with a Palaeozoic and/or volcanic composition.

The sedimentary section overlying basement at Site 207 comprises 38 m of glauconitic silty claystone (sandstone at the base), overlain by 167 m of Paleocene to Middle Eocene and 142 m of Middle Miocene and younger carbonate oozes (Burns, Andrews et al., 1973). A major hiatus in the Oligocene can be traced in DSDP well sites and seismic data throughout the Lord Howe Rise region. The Cainozoic section is ubiquitous across the Lord Howe Rise, and thickens westwards into the central rift province and eastwards into the New Caledonia Basin.

Gravity modelling by Zhu & Symonds (1994) and refraction stations reported by Shor et al. (1971) indicate crustal thicknesses in the range 29–34 km for the Lord Howe Platform, suggesting that the original crust has only undergone minimal thinning during pre-breakup rifting.

The margins of the platform are structurally distinctive. The eastern flank is marked by a 40–50 km-wide crustal zone across which basement is down-faulted into the New Caledonia Basin. Throws on the faults are up to 4 km and, while most of the faulting is downthrown into the basin, in some places there is pronounced reverse faulting indicating localised compression. Willcox et al. (1980) interpreted the eastern flank of the Lord Howe Rise to be a pre-Maastrichtian palaeo-continental margin.

The western flank of the platform is characterised by a basement hinge and a flanking half-graben, which is faulted on its western margin. This half-graben is up to 20 km wide, and contains about 2 km of interpreted Cretaceous sediments that are generally folded and eroded, indicating compression in the latest Cretaceous or earliest Tertiary. This structure can be traced for at least 400 km along the flank of the platform beneath the southern Lord Howe Rise and probably marks a major underlying crustal boundary. In some areas, this structure also coincides with a distinct change in the reflection character of the underlying crust.

Central and Western Rift Provinces

With the improved quality data that are now available, the ‘rift zone’ originally described by Mutter & Jongsma (1978) can structurally be divided into two longitudinal provinces – the *Central Rift* and the *Western Rift*. The boundary between these provinces is not clear cut; however, in small-scale seismic sections in which gross structures are more evident, there is an obvious distinction between them. This division into two parallel rift provinces, which are sub-parallel to the Lord Howe Platform, appears to be a valid distinction along much of the length of the Lord Howe Rise.

The Central Rift underlies the western half of the bathymetric culmination of the Lord Howe Rise and is approximately 150 km wide. It is separated from the Lord Howe Platform by a long and narrow half graben, and from the Western Rift by a broad fault zone across which basement is downfaulted to the west by about 2 s TWT (*ca* 3 km).

The Central Rift is characterised by a series of basement blocks, normally down-faulted to the west. The average basement depth is about 3 s TWT (*ca* 3 km), which is about 1 s (1–1.5 km) deeper than the basement beneath the Lord Howe Platform, indicating greater crustal thinning beneath the Central Rift. This is supported by the gravity modelling of Zhu & Symonds (1994), who show a crustal thickness of about 20 km in this zone, and refraction station N16 of Shor et al. (1971) which gave a crustal thickness of 18 km beneath the northern Lord Howe Rise. This zone was named the Moore Basin by Stagg et al. (1999).

Within the Central Rift, the tops of the basement blocks are typically ill-defined, even where the overlying sedimentary cover is relatively thin. Volcanics, presumably of Cretaceous age, are therefore interpreted to comprise a significant component of the rift fill and are possibly also included in the tops of the rift blocks (i.e. as pre-rift volcanics).

The Western Rift is principally identified beneath the Monowai Sea Valley and Monowai Spur, which together form a subsided western flank to the Lord Howe Rise; this depocentre was named the Monowai Basin by Stagg et al. (1999). Strongly faulted and intruded basement lies at depths of 3.5–5.5 s TWT (*ca* 3–5 km), up to 2 km deeper than in the Central Rift, suggesting greater crustal thinning than beneath the Central Rift and Lord Howe Platform. This is again supported by the gravity modelling of Zhu & Symonds (1994) who interpret a crustal thickness of slightly less than 20 km beneath the Monowai Basin.

The western flank of the Monowai Basin is a structural high, underpinning the bathymetric Monowai Ridge. Basement beneath this ridge is typically 1–1.5 s TWT (*ca* 1–2 km) shallower than that beneath the basin. It is likely that the Monowai Basin and the Monowai Ridge are correlatives of the Lord Howe and Middleton Basins and their flanking western ridge, the Dampier Ridge. In this interpretation, extension in the Lord Howe and Middleton Basins proceeded much further than it did in the Monowai Basin, with the possibility (though unlikely) of incipient breakup occurring in parts of the northern basins. If this correlation is valid, then it follows that pre-breakup and perhaps early post-breakup extension, were concentrated between the Dampier Ridge and Lord Howe Rise in the north and west of the Monowai Ridge in the south. The location of this jump in the locus of extension corresponds to the major boundary between the central and northern Lord Howe Rise blocks interpreted in the reconstruction of Gaina et al. (1998).

As with the Central Rift, the basement blocks underlying the Monowai Basin are often indistinct; again, this is probably due to the presence of extensive volcanics within the sedimentary fill and possibly incorporated into the blocks prior to the onset of rifting.

Tasman Basin

Basement in the Tasman Basin is generally rugged and lies at depths of 6.6–8 s TWT (*ca* 5–7 km). The rugged basement surface appears to be due to four causes: the original oceanic crust emplacement process; the presence of many rugged fracture zones; subsequent magmatic intrusions and extrusions; and Cainozoic reactivation, including volcanism, which has produced both normal and reverse faulting with throws of more than 1 km in places.

The internal character of the Tasman oceanic crust is generally reflection free, except for some isolated locations adjacent to the Lord Howe Rise, where the lower crust contains strong, cross-cutting reflections.

The boundary between Tasman Basin oceanic crust and the thinned continental crust of the western flank of Lord Howe Rise is often not sharply defined, and it is possible that this is related to a transition zone of mixed oceanic (or magmatic) and continental crust, similar to some intermediate volcanic margins.

The sedimentary fill of the eastern Tasman Basin is interpreted to comprise thick sequences of Paleocene to Early Eocene and Middle to Late Eocene pelagic sediments, overlain by a thin and irregularly distributed sequence of Miocene and younger sediments. Internal reflector configurations indicate that strong bottom currents have been an important factor in controlling sediment distribution throughout the development of the Tasman Basin.

ILLUSTRATED TRANSECTS

Three transects are included to illustrate the geology of the Lord Howe Rise region (locations shown in Fig. 25):

Plate 18: (line GA-206/04) extends east-northeast from the eastern Australian margin, across the Tasman Basin, Lord Howe Rise, New Caledonia Basin, Norfolk Ridge, Loyalty Ridge and New Hebrides Trench, to the North Fiji Basin.

Plates 19–20: (lines GA-LHRNR-F, GA-114/02) extends northeast from the eastern Tasman Basin, across the southern Lord Howe Rise, New Caledonia Basin and Norfolk Ridge complex to the South Norfolk Basin.

Plates 21–22: (lines GA-LHRNR-G, GA-114/09 & GA-114/04) extends northeast from the eastern Tasman Basin, across the southern Lord Howe Rise, New Caledonia Basin, Norfolk Ridge complex and Reinga Basin, to the South Norfolk Basin.

NATURALISTE PLATEAU

INTRODUCTION

The Naturaliste Plateau region lies off the southwestern tip of Australia. It is dominated by the Naturaliste Plateau, a large, rectilinear (400 km x 250 km) continental margin plateau that protrudes westward from the Australian continental landmass (Figs 28, 29). The plateau is flanked to the north by the Perth Abyssal Plain, to the west by the Naturaliste Fracture Zone and Gonneville Triangle, to the east by the Naturaliste Trough, and to the south by the Diamantina Zone. It is one of several plateaus that remain attached to the western Australian continental margin after the rifting and breakup of Gondwana during the Mesozoic. The plateau is largely underlain by continental crust, a fact confirmed by the dredging of Cambrian gneiss from the southern flank of the plateau (Royer & Beslier, 1998).

GEOMORPHOLOGY

The Naturaliste Plateau lies in water depths of approximately 2000–5000 m. It forms a prominent protrusion of the western Australian continent margin and has an approximately rectilinear outline. This outline is largely the result of different episodes and directions of rifting and seafloor spreading affecting its northern, western and southern flanks.

The plateau has a relatively flat upper surface which is outlined by the 2500 m isobath (Fig. 30). The southern and western flanks are steep (gradients up to 7º) whereas gradients on the northern flank are generally more gentle (typically 2–4º). The physiography of the western flank is complicated by the presence of a large seamount on the lower slope at about 33.5ºS and by ridges of the Naturaliste Fracture Zone that are embedded in the lower slope from about 34.5–35.5ºS (Figs 28, 29). On the southern margin of the plateau, the steep upper to middle slope changes to a complex and extensively-canyoned lower slope at 4000–4500 m depth which descends to the abyssal depths (typically 4500–5500m and deeper) of the Diamantina Zone and adjacent seafloor.

North of the Naturaliste Plateau, the seafloor of the Perth Abyssal Plain is typically flat, whereas to the west and south it is disrupted by a series of basement ridges and troughs. The seafloor is particularly rugged to the south, where it consists of a series of alternating, broadly ESE–trending basement ridges and troughs of the Diamantina Zone. This zone is up to 200 km wide and extends from the eastern-most part of Broken Ridge (~100ºE) to about 120ºE. Most of the ridges comprise exposed basement. In the intervening troughs, the sedimentary cover is patchy and thin (typically no thicker than a few hundred metres). The boundary between the Diamantina Zone and the Australia-Antarctica Basin to the south is marked by a rapid shallowing of the sea floor (6000–6500 m to 4500–5000 m) and a change to a much smoother topography.

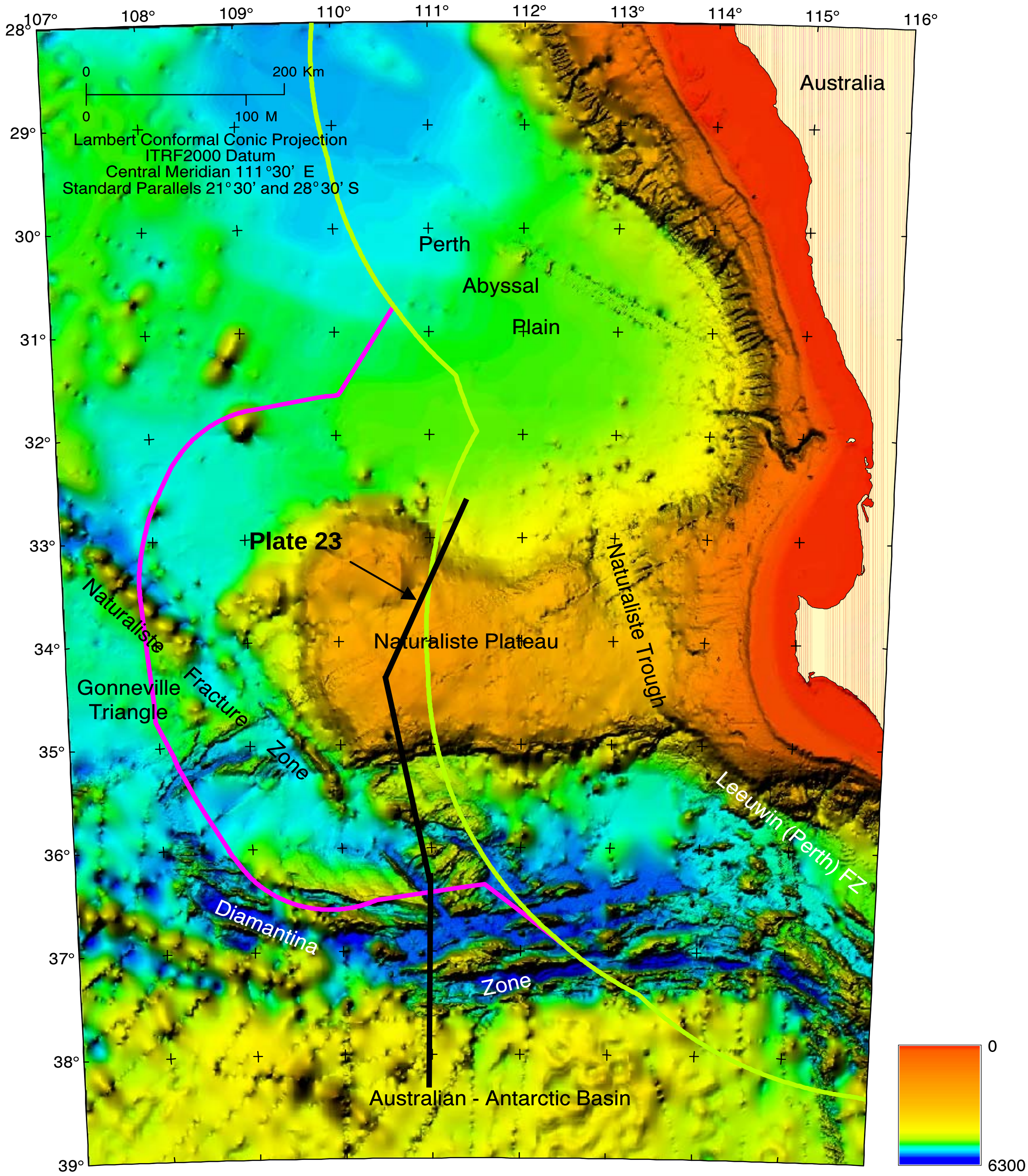

Figure 28: *Bathymetric image of the Naturaliste Plateau region, showing the location of the illustrated seismic transect. Major geomorphological features are labelled. Line colours: green – 200 M (Australia); magenta – outer limit of extended continental shelf (as submitted to UN CLCS, November 2004).*

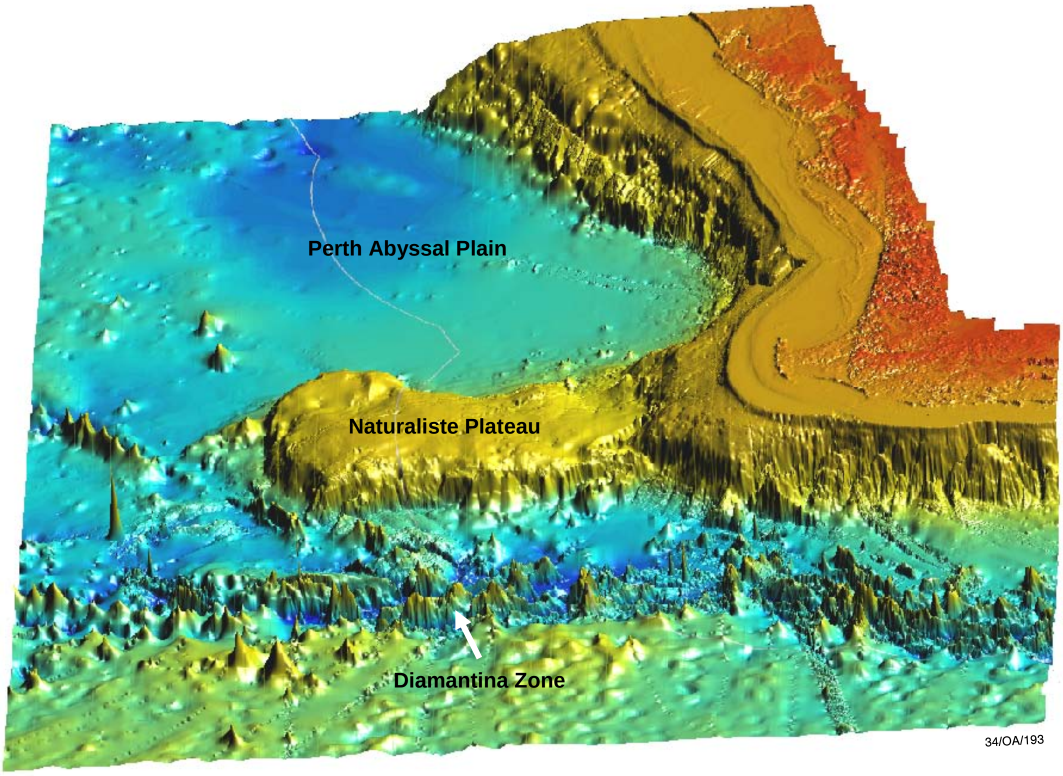

Figure 29: *3-D bathymetric image of the Naturaliste Plateau region viewed from the south.*

A continental rise is not present along most of the plateau margins due to strong current action and low sedimentation rates. As is the case with most parts of the Australian continental margin, the Naturaliste Plateau region is sediment-starved.

To the east of the Naturaliste Plateau, the Naturaliste Trough is a northerly-trending bathymetric saddle between the plateau and the upper continental slope and shelf of mainland Australia. The trough and its flanks are underlain by the Mentelle Basin (Fig. 31).

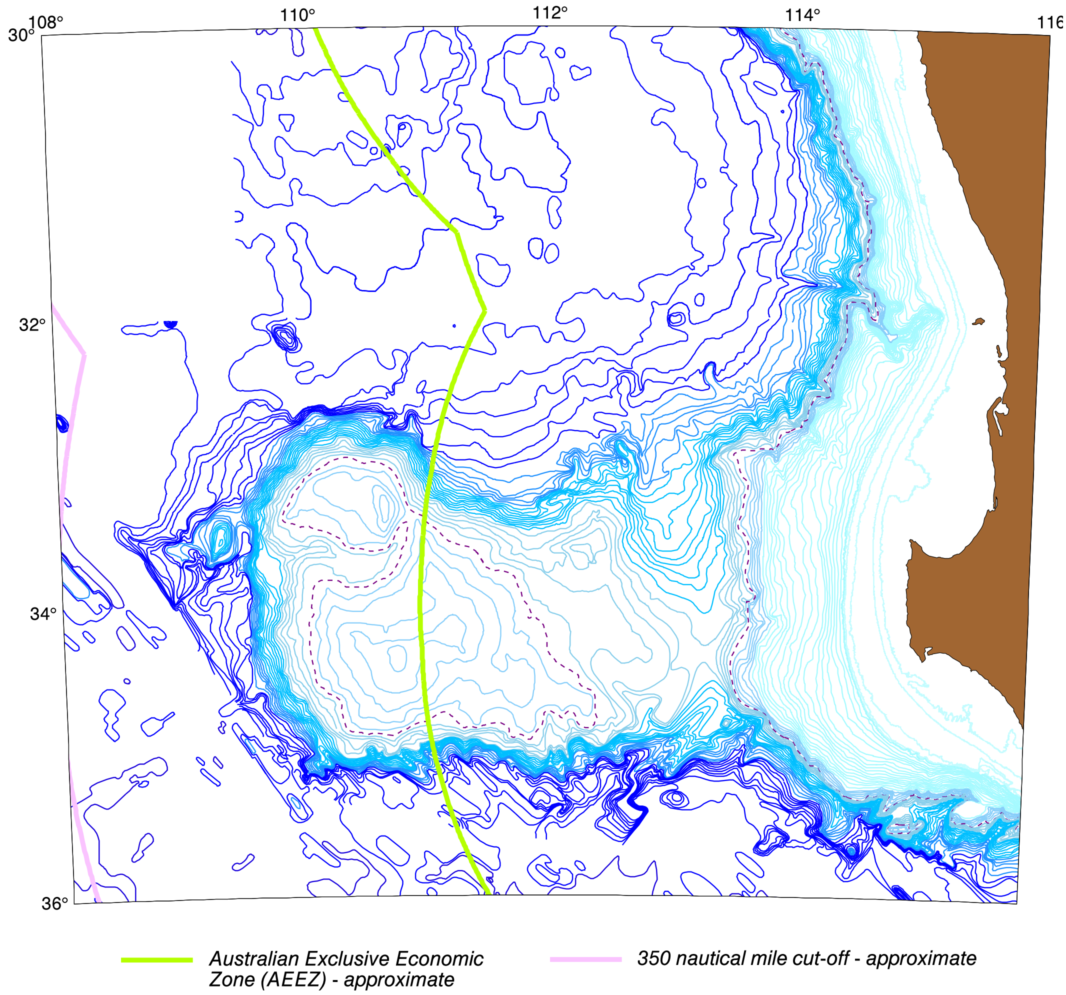

Figure 30: *Bathymetric contour map of the Naturaliste Plateau region from Geoscience Australia's ORMS bathymetric map series (Jongsma & Johnston, 1993b). The 2500 m isobath is shown as a dashed line. Green line is the approximate 200 M line, and pink line the approximate 350 M constraint.*

GEOLOGY

Plate tectonic setting

During the breakup of the super-continent Gondwana, the Naturaliste Plateau was located near the junction of three major plates: Australia, Antarctica and Greater India. Magnetic anomalies in the Perth Abyssal Plain, north of the Naturaliste Plateau, contain evidence of the Early Cretaceous breakup between Greater India and Australia/Antarctica. Breakup commenced in the Argo Abyssal Plain off northwestern Australia in the Late Jurassic and gradually propagated to the southwards (see for example, Veevers, 2000). The oldest mapped anomaly in the Perth Abyssal Plain is chron M10 – Early Hauterivian (Cande et al., 1989). However, seismic data indicate that most of the anomalies currently mapped as M10 and M9 are located on continental/transitional crust and are therefore not seafloor spreading anomalies (Borissova, 2002). M8 and M7 are the oldest reliably-mapped anomalies in the Perth Abyssal Plain.

The Gonneville Triangle, which is separated from the Naturaliste Plateau by the Naturaliste Fracture Zone (Fig. 28), does not contain any identified magnetic anomalies. Based on variety of geophysical data, Munschy (1998) suggested that it is floored by oceanic crust 120–90 Ma in age. This period corresponds to the Cretaceous Magnetic Quiet Period.

No magnetic anomalies have been identified in the Diamantina Zone south of the Naturaliste Plateau and its origin is speculative, with interpretations ranging from trapped Early Cretaceous oceanic crust (Munschy, 1998) to extended continental crust (Chatin et al., 1998). Tikku & Cande (2000) re-examined existing magnetic data and suggested that the southern part of the Diamantina Zone was formed during an ultra-slow spreading phase between Australia and Antarctica between chron 34 and chron 20.

Geology of the plateau and adjacent areas

Deep Sea Drilling Project (DSDP) holes 258 and 264 were drilled on the Naturaliste Plateau (Fig. 31) and recovered a succession of Cretaceous to Miocene sedimentary rocks (Davies et al., 1974; Hayes, Frakes, Barrett et al., 1975). While neither site reached basement, volcaniclastic conglomerates of Cenomanian or older age were recovered at Site 264 a short distance above acoustic basement (Ford, 1975).

The first sample of the plateau basement was dredged by the *Eltanin* (ELT55-12) from the northern margin of the plateau (Heezen & Tharp, 1973; Fig. 31). Plagioclase-rich rocks from this dredge were originally interpreted as being of continental origin. However, a more detailed petrographic analysis (Coleman et al., 1982; Storey et al., 1992) showed that these rock clasts are in fact altered tholeiitic basalts similar to the Bunbury Tholeiitic Suite (132–122 Ma) on the Australian mainland.

Four plagioclase-rich *Eltanin* dredge rocks (ELT55-12) and a cobble from volcaniclastic conglomerate at DSDP Site 264 were analysed by Mahoney et al. (1995), who confirmed their mostly basaltic composition. However, they also noted that Naturaliste Plateau samples are chemically and isotopically different from typical hotspot basalts and that their composition suggests contamination of the magmas by continental crust.

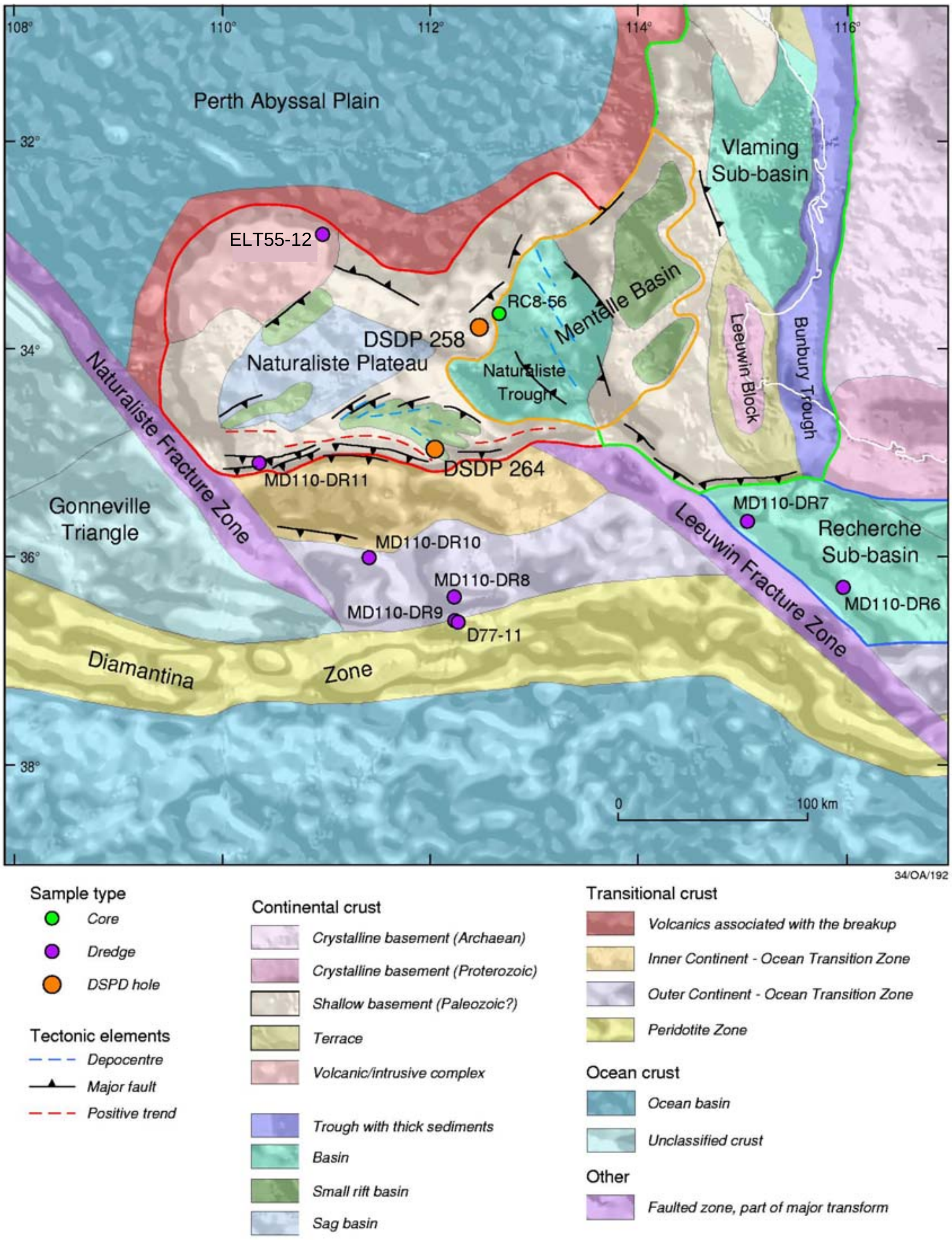

Figure 31: *Geological features and provinces of the Naturaliste Plateau region (after Borissova, 2002).*

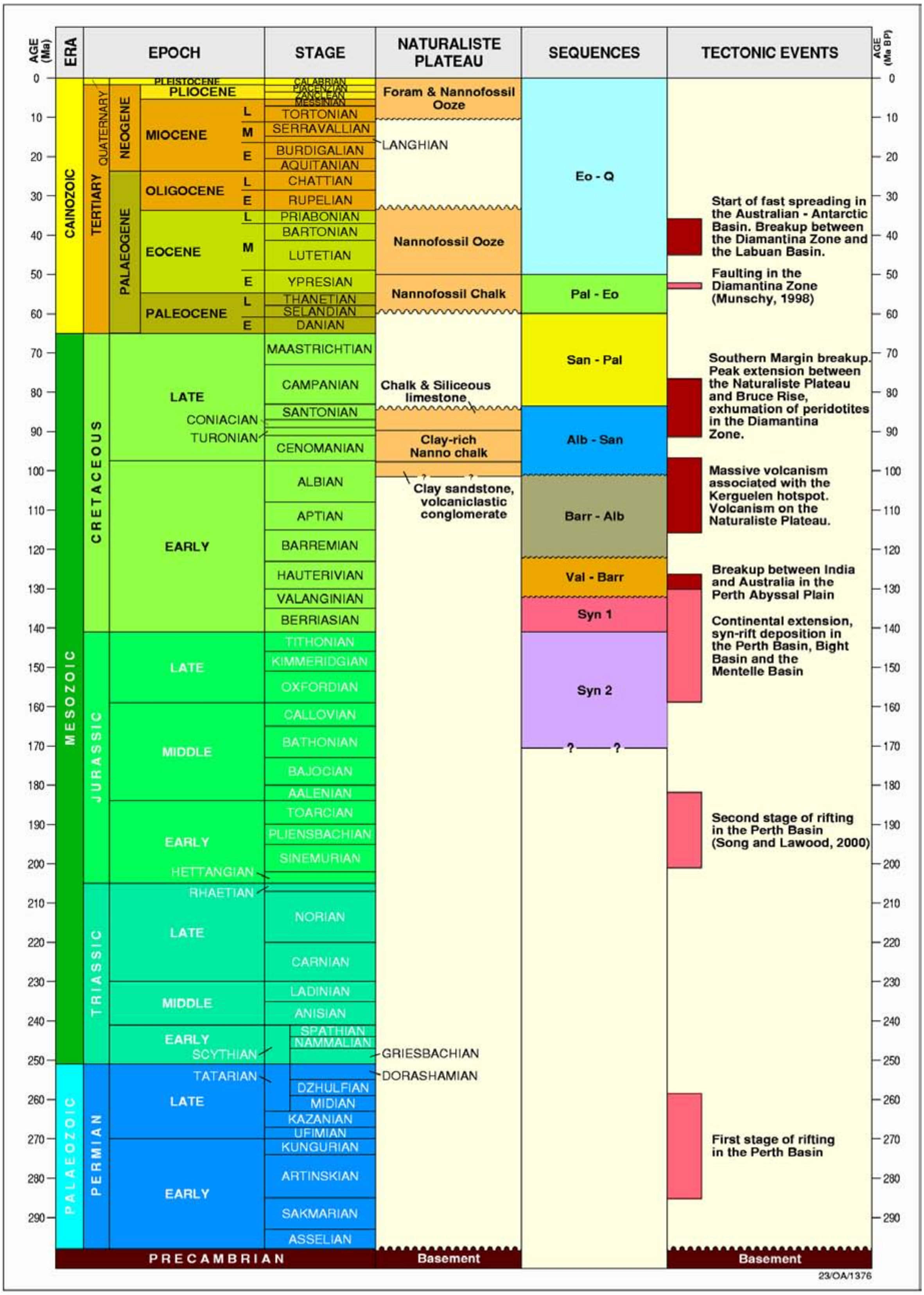

Figure 32: *Summary stratigraphic diagram for the Naturaliste Plateau region (after Borissova, 2002). Bars in Tectonic Events column indicate approximate duration of each event.*

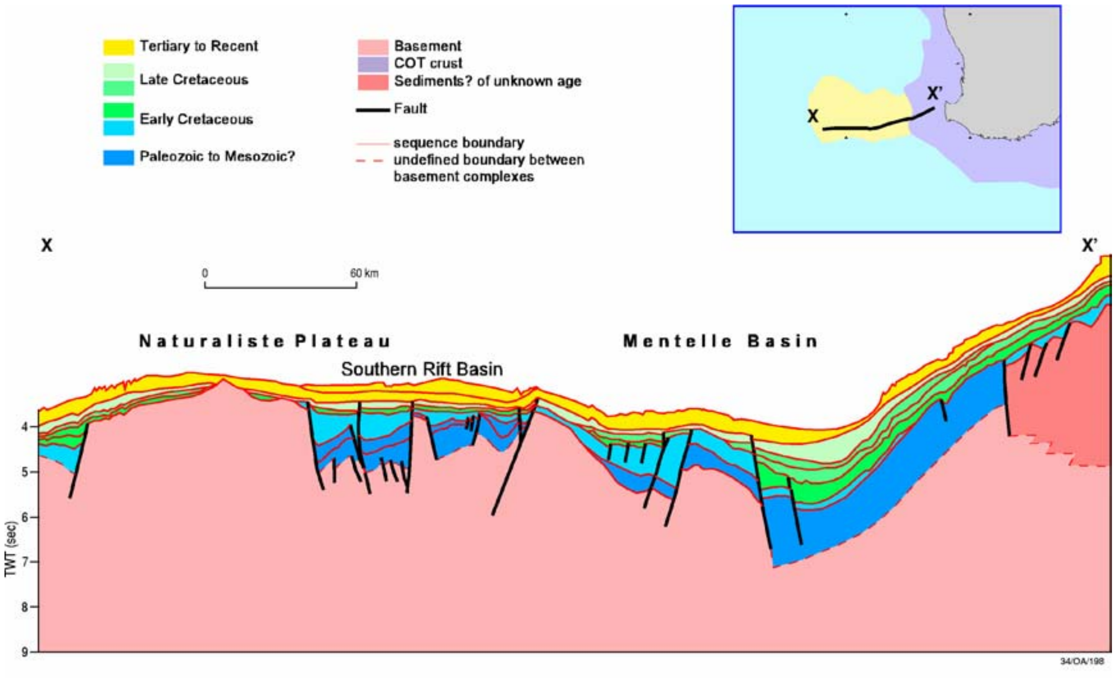

Figure 33: *W–E cross-section across the southern part of the Naturaliste Plateau, the Mentelle Basin (Naturaliste Trough) and the upper continental slope (after Borissova, 2002).*

A continental origin for the Naturaliste Plateau is supported by a dredge sample recovered by the *Marion Dufresne* in 1998 on the faulted southern flank of the plateau (MD110-DR11; Royer and Beslier, 1998; Fig. 31). Analysis of the recovered material showed the presence of high-grade gneisses similar in composition to those of the Australian craton (Beslier et al., 2001). It is now apparent that the Naturaliste Plateau, as with other Indian Ocean plateaus such as the Wallaby Plateau and Kerguelen Plateau, is at least partly cored by continental rocks, as originally proposed by Petkovic (1975). Recent potential-field modelling is consistent with basement of the plateau being largely of granitic and/or gneissic composition (Direen, 2004c).

The sediment thickness over much of the Naturaliste Plateau does not exceed 1 s TWT (*ca.* 1 km; Borissova, 2002). Thicker sequences occur in small rift basins, particularly on the southern part of the plateau, and in the Mentelle Basin which underlies the Naturaliste Trough to the east (Bradshaw et al., 2003; Figs 31, 33).

Eight major sequences have been identified on the Naturaliste Plateau and the adjacent Mentelle Basin (Borissova, 2002; Fig. 32). However, only the post-Albian sequences are reasonably well dated, from control provided by DSDP Sites 258 and 264.

The nature of the continent–ocean transition (COT) zone adjacent to the Naturaliste Plateau is complex (Fig. 31). Along most of the northern margin of the plateau, the COT zone is 20–90 km wide and contains features common to volcanic rifted margins. To the west, the COT zone is disrupted by the Naturaliste Fracture Zone that is embedded in the lower continental slope from about 34.5–35.5ºS (Figs 28 & 31).

To the south, the Naturaliste Plateau is abutted by highly-faulted basement alternating with large intrusive-type bodies (Borissova, 2002). The southern area of this province is part of the Diamantina Zone. The crustal origin of the Diamantina Zone is equivocal. It has been variously interpreted as: an oceanic fracture zone (Markl, 1978); re-rifted Early Cretaceous oceanic crust (Munschy, 1998); the product of ultra-slow spreading in the Late Cretaceous to Eocene (Tikku & Cande, 2000); or as highly-extended continental lithosphere (Chatin et al., 1998, Borissova., 2002). Potential field modelling along a transect extending south from the Naturaliste Plateau across the Diamantina Zone to the oceanic crust of the Australia-Antarctic Basin is consistent with the zone being composed of exposed mantle peridotites and serpentinites overlain by syn-breakup volcanics and post-breakup marine sediments (Direen, 2004c).

Geochemical analysis of peridotites dredged in the Diamantina Zone provides evidence for emplacement in a rift setting, allowing rapid exhumation and a consequent limited amount of melting, rather than formation at a slow-spreading oceanic ridge (Nicholls et al., 1981; Chatin et al., 1998). This supports a continental origin for much of the Naturaliste Plateau region.

ILLUSTRATED TRANSECTS

One transect is included to illustrate the geology of the continental margin in the region of the Naturaliste Plateau (location shown in Fig. 28):

Plate 23: (lines GA-187/07 & N323) extends north from the oceanic crust of the Australian–Antarctic Basin in the south, across the Naturaliste Plateau to the Perth Abyssal Plain in the north.

SOUTH TASMAN RISE

INTRODUCTION

The South Tasman Rise region contains two large continental margin plateaus – the South Tasman Rise (STR) and East Tasman Plateau (ETP; Fig. 34). The STR is an elongate plateau that extends approximately 700 km south-southeast of Tasmania. The ETP, approximately one third of the area of the STR, extends for approximately 300 km to the southeast of Tasmania. A variety of geological and geophysical data indicate that both the STR and ETP are underlain by continental crust.

The South Tasman Rise region has been extensively covered by geophysical surveys since the early 1970s. The seismic data are supplemented by extensive swath-mapped bathymetry data acquired by France and Australia. Some sampling information is available, mainly from the Deep Sea Drilling Program (DSDP) and from the Ocean Drilling Program (ODP), but also including some dredge samples.

GEOMORPHOLOGY

The morphology of the seafloor south of Tasmania is dominated by the 200 000 km^2 area of the STR (Figs 34, 35). The STR is separated from Tasmania by the west-northwest-trending, 3000 m-deep South Tasman Saddle. The STR rises from abyssal depths of about 4500 m to the west, south and east to a crestal depth of slightly less than 1000 m. It comprises two distinct terranes: a complexly deformed terrane in the west, and a structurally more simple terrane in the east and south.

The western terrane is characterised by a generally rugged seabed that is delineated in the west by the linear, 450 km-long, composite Tasman Escarpment and Needwonne Ridge. This feature is the northernmost expression of the Tasman Fracture Zone, which is one of the major N-S trending fracture zones along which the main seafloor spreading axis between Australia and Antarctica is progressively offset to the southeast. The scarp is up to 2000 m high, with the seabed on average dropping from about 3000 m to 4500 m. The abyssal region west and southwest of the STR is floored by oceanic basalts, with a well-developed seafloor spreading fabric (Exon & Hill 1999). The rough abyssal basement surface has little or no sediment cover (Hill & Moore, 2001). To the east of the Needwonne Ridge, the seafloor forms a north–south trending trough between the ridge to the west and the main part of the STR to the east. The morphology of this area is controlled by the presence of numerous, small, underlying pull-apart basins. The seabed at the northern end of the trough is quite rugged while to the south it becomes progressively smoother and deeper.

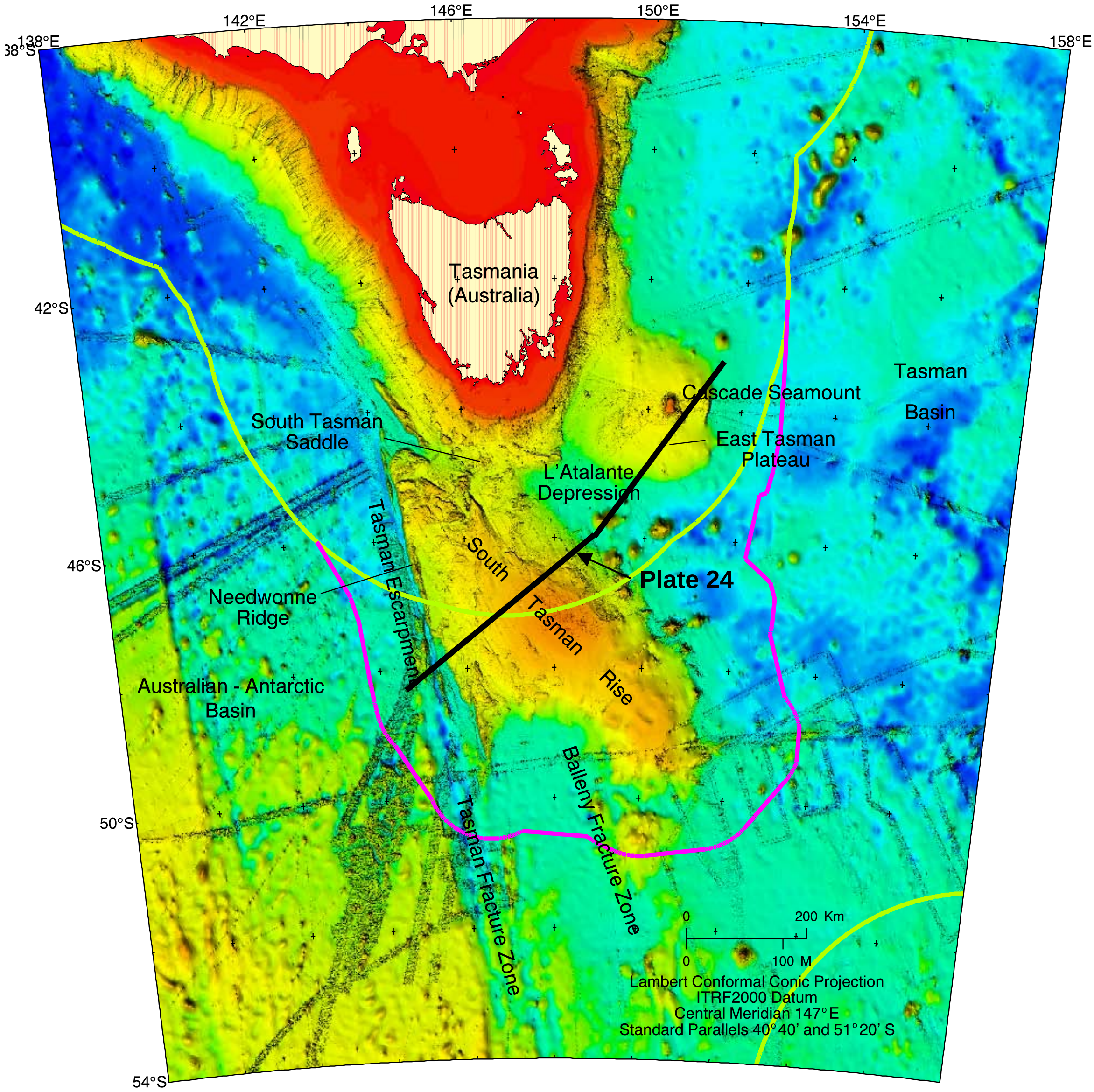

Figure 34: *Bathymetric image of the South Tasman Rise region, showing the location of the illustrated seismic transect. Major morphological features are labelled. Line colours: green – 200 M; magenta – outer limit of extended continental shelf (as submitted to UN CLCS, November 2004)*

The eastern terrane comprises the major part of the STR and is characterised by a relatively simple morphology that is underpinned by shallow continental basement. The crest of the STR has the form of a broad dome, elongate on a south-southeast azimuth, and includes the main culmination of the STR at less than 1000 m depth. To the east, the STR descends at relatively low gradients into the Tasman Basin. Volcanic seamounts are embedded within the deeper parts of this margin. The boundary between the continental crust of the STR and the oceanic crust of the Tasman Basin is not clearly defined in the morphology. The southeastern extent of the STR is marked by an E–W trending saddle at more than 4000 m depth south of which a small, un-named plateau rises to a depth of about 2300 m. This plateau is interpreted as a continental fragment that was rifted from the STR at the time of breakup (Hill & Moore, 2001).

The 50 000 km^2 ETP is a circular plateau located to the southeast of Tasmania (Figs 34, 35), that mainly lies at water depths of 2500–3000 m and is separated from Tasmania by the NNE-trending East Tasman Saddle. The ETP is capped by a large volcanic guyot, the Cascade Seamount, which rises to 650 m below sea level (Exon et al., 1997a). The ETP is flanked to the northeast, east and southeast by abyssal plain of the Tasman Basin at depths of 4000–4500 m, while the northwest-trending L'Atalante Depression separates the ETP from the STR to the southwest. The generally flat seafloor of the L'Atalante Depression and adjacent abyssal areas is interrupted by volcanic seamounts.

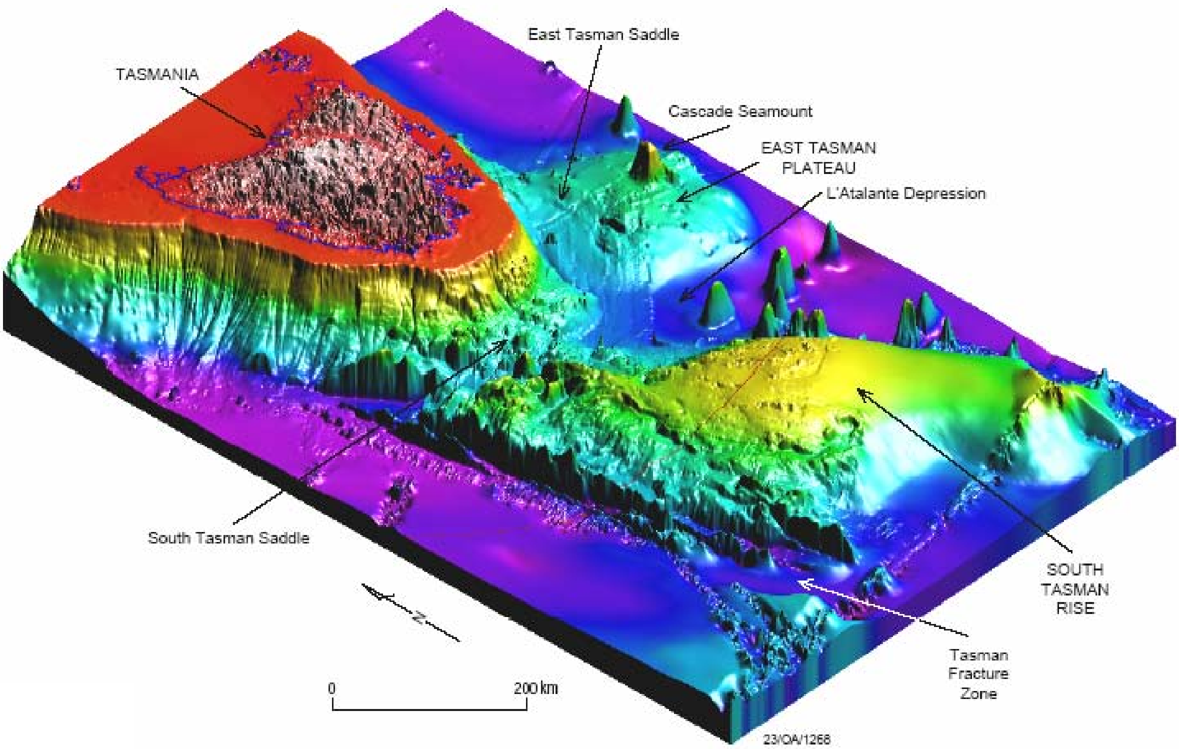

Figure 35: *3-D perspective image of the bathymetry of the South Tasman Rise region viewed from the southwest. Key physiographic features are labelled. Red line is 200 M line.*

GEOLOGY

Plate tectonic setting

The Tasmanian margin (including both the STR and ETP) was part of Eastern Gondwana and was formed by complex interactions during the separation of southern Australia and Antarctica, leading to the formation of the Australian–Antarctic Basin, and the separation of eastern Australia and Lord Howe Rise, leading to the formation of the Tasman Basin. Plate tectonic events in the region of the South Tasman Rise have been summarised by Royer & Rollet (1997).

The breakup of Australia and Lord Howe Rise in the Late Cretaceous–Palaeogene separated the ETP from southeast Tasmania and the eastern STR (Exon et al., 1997a). The L'Atalante Depression, interpreted to be floored by oceanic crust, was formed by seafloor spreading on a WSW–ENE azimuth between the ETP and the eastern STR in the Campanian-Maastrichtian (Fig. 36). A ridge jump to the northeast, which occurred a few million years later, initiated the main phase of spreading in the Tasman Basin.

The opening of the Australian–Antarctic Basin commenced along the southern Australian margin with very slow spreading in the Late Cetaceous (Tikku & Cande, 1999). Based on satellite gravity data and plate tectonic reconstructions, Royer & Rollet (1997) concluded that the present-day STR formed from the two separate terranes (Fig. 36). In this scenario, the western terrane lying between the Tasman Fracture Zone and a N–S oriented boundary at 146º30'E was initially part of the continental shelf of North Victoria Land, Antarctica, while the eastern terrane was part of the Tasmanian margin.

By the Late Paleocene, seafloor spreading between Australia and Antarctica had propagated far enough to the southeast to cause the detachment of the western STR from the Antarctic Plate; this fragment then became 'welded' to the eastern STR terrane by the Early Oligocene (Royer & Rollet, 1997). During margin breakup, the western terrane underwent intense wrenching and left-lateral shearing between the Antarctic shelf break and its western boundary. Deformation continued after breakup along the transform margin until the Early Miocene. Clearance of the STR and Antarctica at the southwestern tip of the STR and the establishment of an open seaway occurred in the Early Oligocene (~33 Ma; Exon et al., 2001).

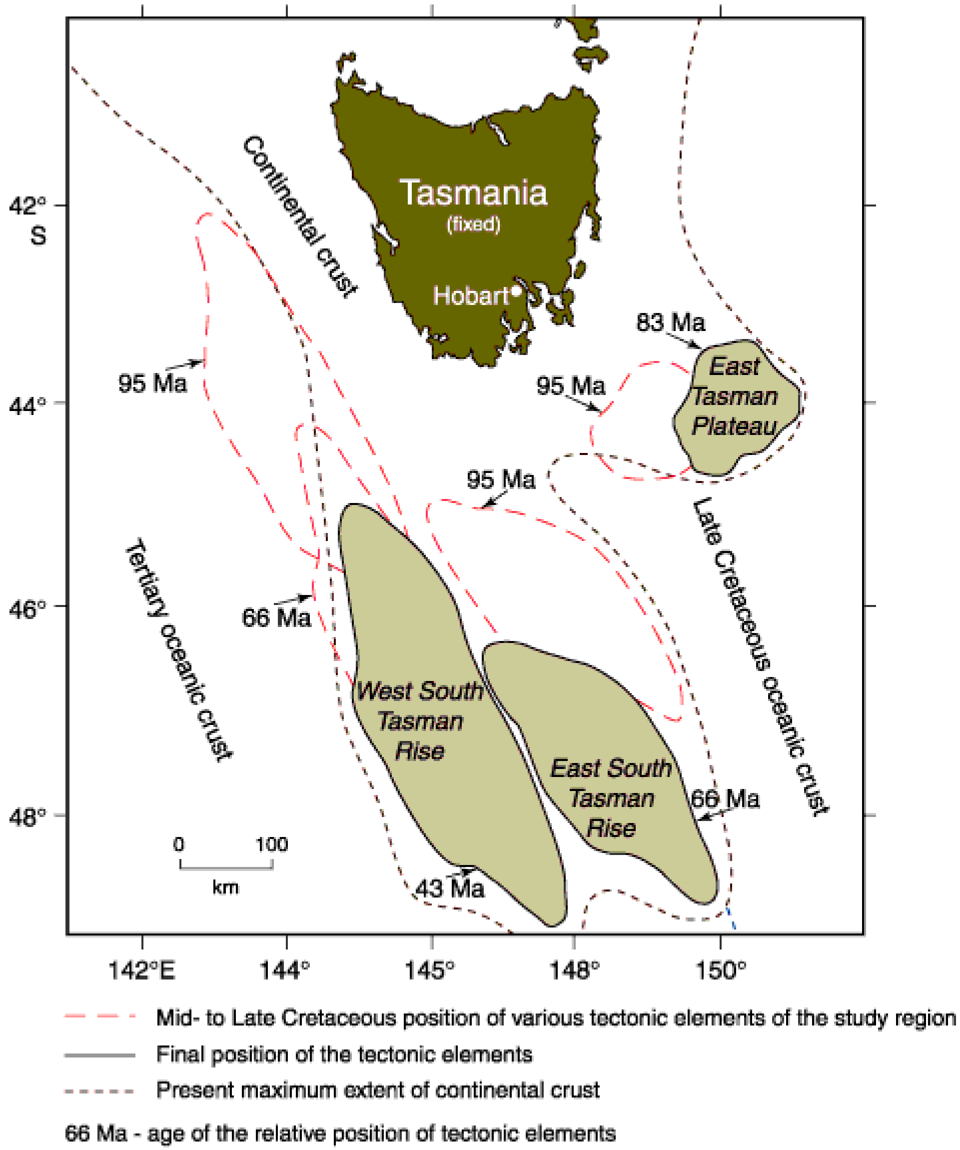

Figure 36: *Sketch map of Cretaceous and Paleogene movements, relative to a fixed Tasmania, of the western South Tasman Rise block, eastern South Tasman Rise block, and East Tasman Plateau (modified from Exon et al., 1997b, based on Royer & Rollet, 1997).*

Geology of the South Tasman Rise & East Tasman Plateau

The STR and ETP are geologically complex continental blocks (Fig. 37), with basement that comprises gneisses, granites and metasediments which show similarities to Palaeozoic basement rocks on Tasmania and Antarctica (Berry et al., 1997). Bedrock is exposed in some parts of the region, including the western edge of the STR (adjacent to the Tasman Fracture Zone), the Lowreenne Massif (at the northwestern corner of the STR), the summit area of the central STR (Hill et al., 2001) and the eastern flank of the ETP.

The region has been extensively sampled by scientific surveys since the 1970s (Figs 37, 38). DSDP Site 281, on the southern margin of the STR, drilled into late Carboniferous mica schist basement, providing the first evidence of a continental origin (Kennett, Houtz et al. 1975). Dredging during R/V *Rig Seismic* Survey GA-147 and R/V *Sonne* Cruise 36 recovered metamorphic rocks from the STR and ETP region, which further supported a continental origin (Berry et al., 1997; Royer & Rollet, 1997). ODP Leg 189 drilled five sites south of Tasmania (Fig. 37, Sites 1168-1172; Exon et al., 2001). These sites recovered Late Cretaceous to Late Quaternary marine sedimentary rocks.

Seismic data show that the northern part of the STR is highly structured (Fig. 39). In contrast, the central STR is composed of two distinct structural terranes; a complexly-deformed western terrane and a structurally simpler eastern terrane (Fig. 40). These terranes are also reflected in the morphologic terranes described previously. Metamorphic rock samples indicate that the western terrane correlates with the Wilson Terrane of Antarctica, whereas the eastern terrane closely matches metamorphic rocks of Tasmania (Berry et al., 1997). The boundary between the terranes is delineated by gravity and magnetic anomalies associated with a westward-deepening basement step (Royer & Rollet, 1997). The more complex nature of the western terrane is due to deformation within the Tasmanian-Antarctic Shear Zone during its transfer from the Antarctic to the Australian plate in the Paleocene, and then during N–S wrenching associated with transform movement along the Tasman Fracture Zone (Exon et al., 2001). The eastern STR exhibits mainly NW–SE and NE–SW structural trends consistent with the Tasman rift and spreading directions.

Sedimentary basins on the STR contain at least 4 km of section consisting of Late Cretaceous to Eocene siliciclastics, overlain by Oligocene and younger pelagic carbonates. Where drilled, the siliciclastics comprise deltaic and shallow marine silty claystones, often rich in organic carbon. The western STR contains numerous small wrench basins, whereas the basement high beneath the main part of the STR is flanked by narrow NW–SE trending rift basins, with at least 3 km of sedimentary section and 20–25 km wide.

The East Tasman Plateau is bounded by high-angle faults and its outer margins are underpinned by planated and exposed continental basement (Fig. 41). The central part of the plateau is dominated by the Eocene-age (40–35 Ma) volcanic Cascade Seamount and its volcanic apron (Exon et al., 1997a). Steeply dipping faults cut basement and the overlying volcanic apron

(Fig. 41). The basement of the plateau was sampled at several locations during Geoscience Australia Survey GA-147. Orthogneiss and rhyolite were dredged from the eastern flank of the plateau. This orthogneiss had close affinities with samples taken in three dredges from the southeastern continental slope of Tasmania and the northeastern slope of the South Tasman Rise (Berry et al., 1997).

Rocks dredged from the Cascade Seamount include Late Eocene alkali olivine basalts, volcanic breccia and hyaloclastite (Exon et al., 1997a). The Cascade Seamount is considered to be part of the ocean island chain, which includes the Balleny Islands and extends almost to Antarctica (Lanyon et al. 1993; Sutherland, 1994). The Balleny Islands have been postulated to be formed above a hotspot (the Balleny Hotspot ; Crawford et al., 1997), which in the Late Eocene was located immediately to the southeast of Tasmania. Plate kinematic reconstructions of the region show that the Australian/Pacific plate boundary passed over the Balleny Hotspot at approximately 20–30 Ma. Basalts dredged from seamounts in the L'Atalante Depression, and also from the STR, show strong affinities with the Cascade Seamount basalts (Crawford et al., 1997) suggesting a similar origin and possibly a similar age.

The L'Atalante Depression, located between the STR and ETP, is interpreted to be floored by oceanic crust (Exon et al., 2001). Seismic data suggest that the East and South Tasman Saddles are mostly floored by extended continental crust.

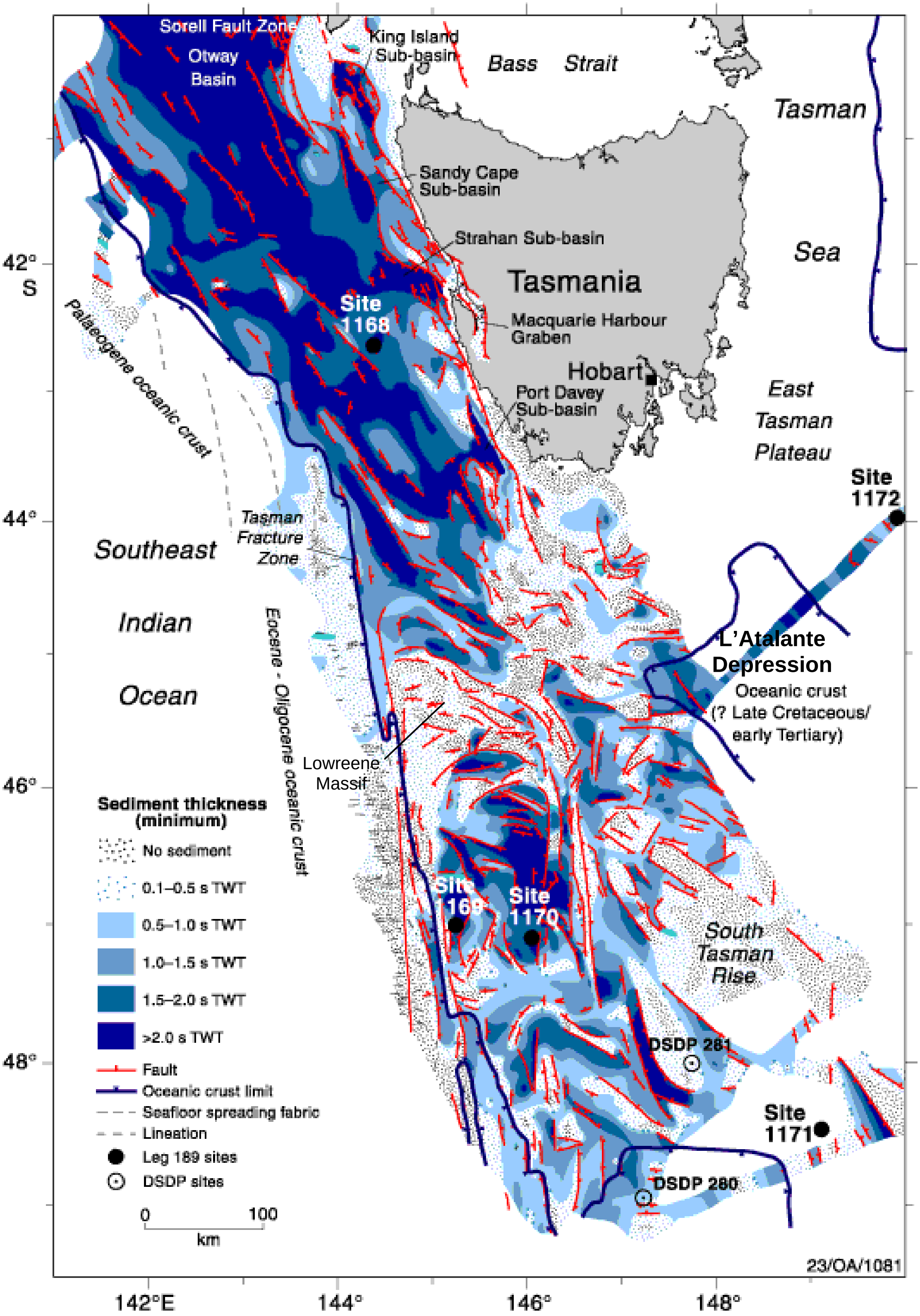

Figure 37: *Map of structures and minimum sediment thickness of the area west and south of Tasmania after Hill et al. (1997), based on multichannel seismic profiles. ODP and DSDP sites are shown.*

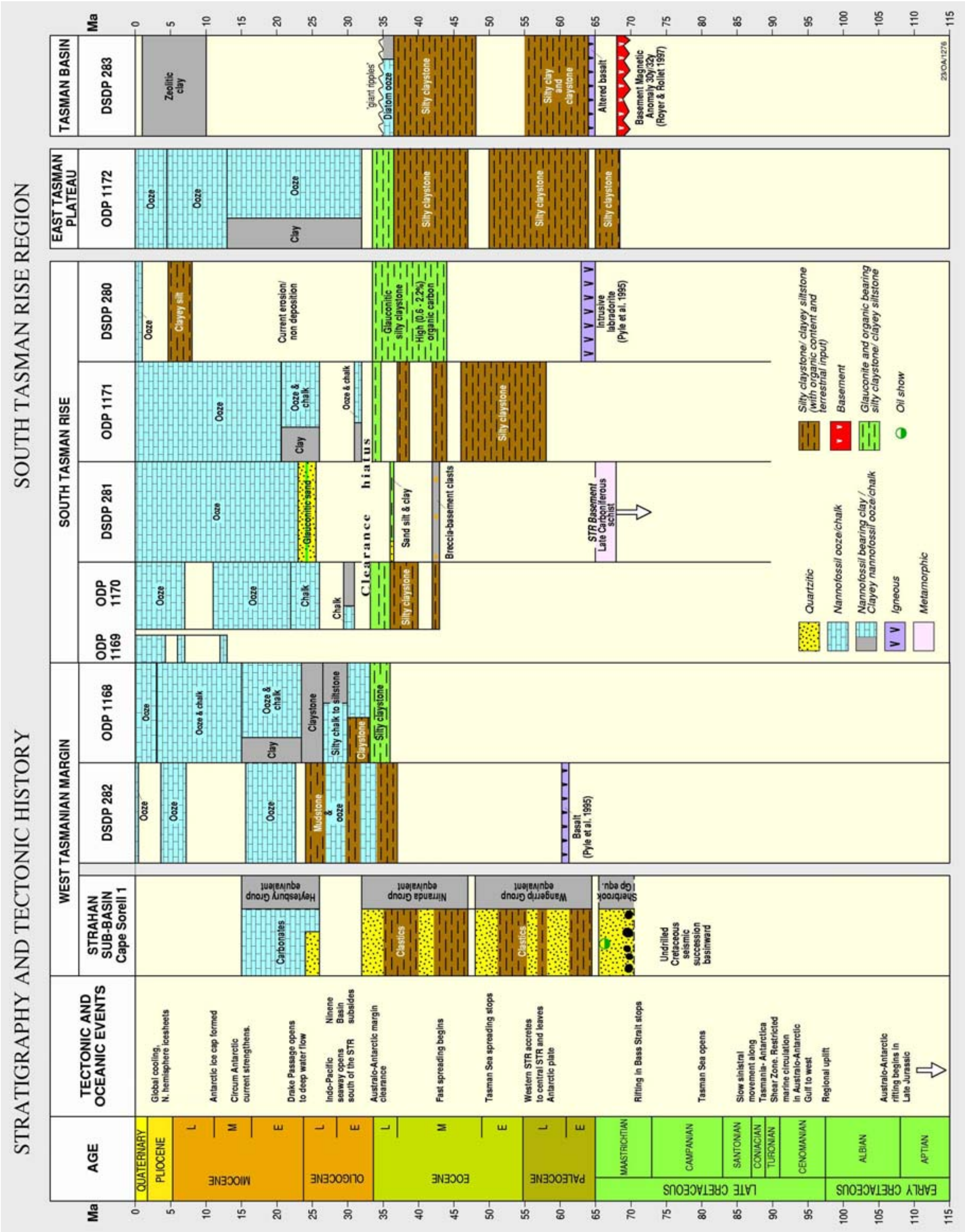

***Figure 38:** Stratigraphic diagram for the South Tasman Rise region, based on Cape Sorell–1 exploration well and DSDP and ODP drill sites (after Hill et al., 2001).*

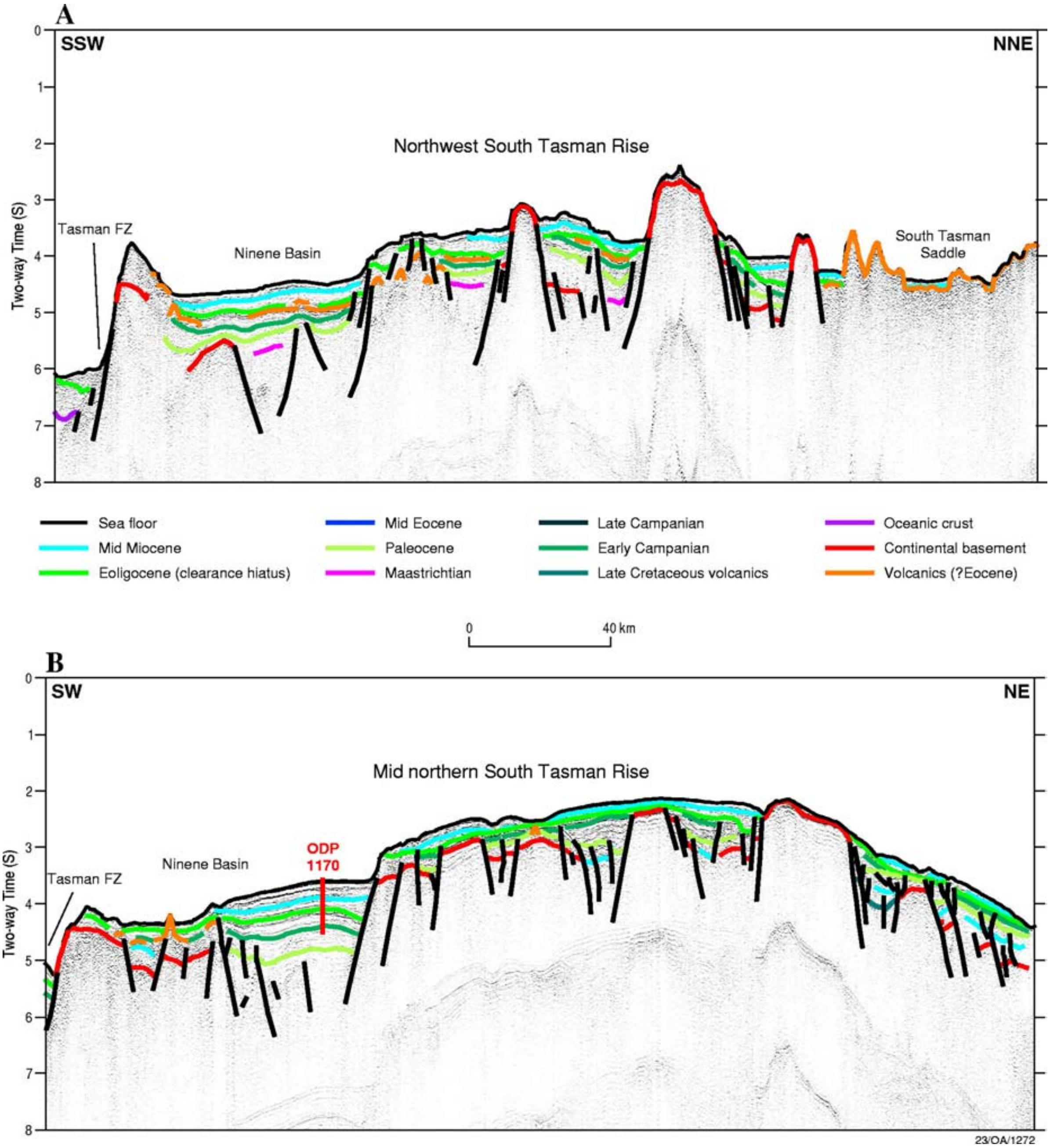

Figure 39: *Interpreted seismic profiles – A) SO36-62 and B) SO36-58 on the South Tasman Rise (after Hill et al., 2001).*

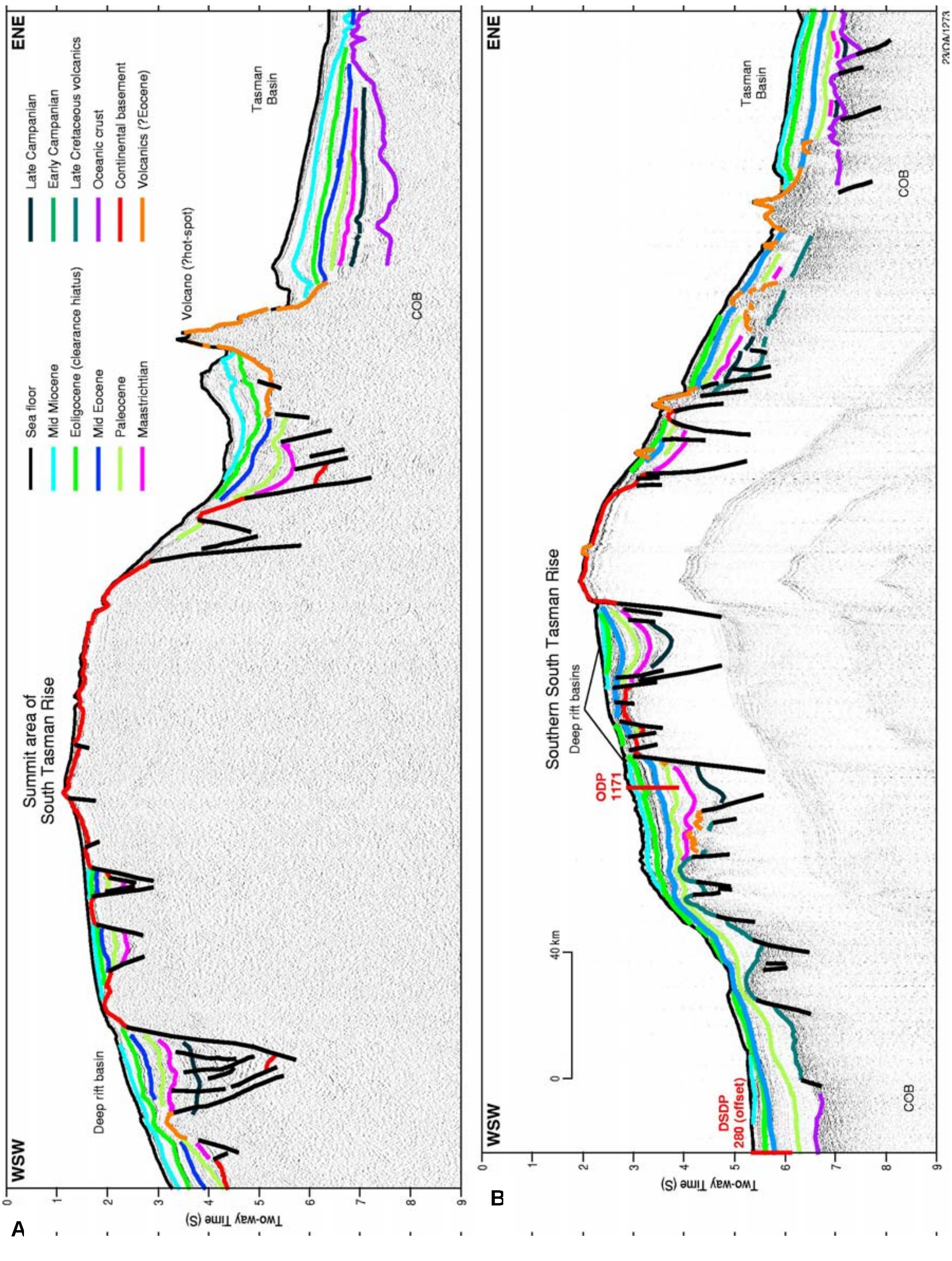

Figure 40: *Interpreted seismic profiles – A) GA-202/2 and B) PD00-5 on the South Tasman Rise (after Hill et al., 2001).*

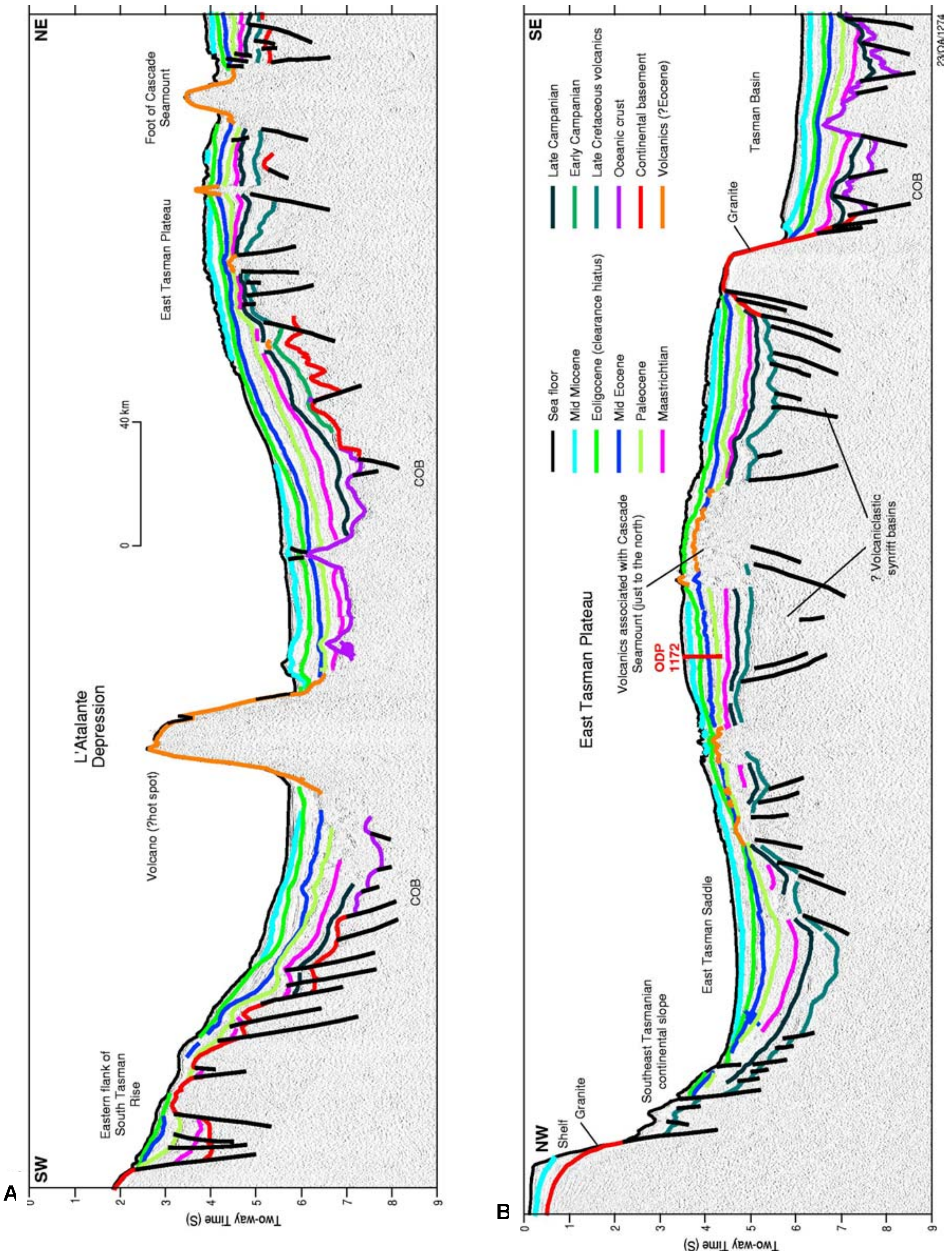

Figure 41: *Interpreted seismic profiles – A) GA-202/12 and B) GA-226/4 – GA-202/1 on the South Tasman Rise and East Tasman Plateau (after Hill et al., 2001).*

Tectono-stratigraphic evolution

The following summary of the tectono-stratigraphic evolution of the region of the South Tasman Rise is largely taken from Hill & Moore (2001).

Basin development south of Tasmania probably dates back to the latest Jurassic to Early Cretaceous when this area is likely to have developed as an arm of the Southern Rift System – the rift that preceded breakup along the continental margin of southern Australia.

The main phase of basin development occurred post-Cenomanian. In the east, this was the direct result of Tasman Basin rifting, while in the west it was due to wrenching as the NW–SE sinistral Tasmanian-Antarctic Shear became active due to extension in the Great Australian Bight (GAB) further to the west. Seafloor spreading in the southern Tasman Basin, adjacent to the eastern STR and ETP, began in the early Campanian, at the same time that spreading was initiated in the GAB. The L'Atalante Depression opened first, separating the STR and ETP, but spreading then failed several million years later as the locus of spreading jumped to the northeast of the ETP, leading to the main phase of WSW–ENE Tasman opening. The East Tasman Saddle was a short-lived transform zone between Tasmania and the ETP during the initial breakup phase. Up to the Paleocene, the structural development of the STR was controlled by two stress regimes, extension to the northeast (Tasman rifting and spreading), and NW–SE strike-slip movements associated with the Tasmanian-Antarctic Shear. The resulting transtension produced the deep, narrow rifts co-aligned with the main NW–SE-oriented axis of the STR. In contrast, the structural style of the eastern margin of the STR and on the ETP is dominated by normal faulting, reflecting the development of the Tasman Basin passive margin.

There was a major change in the tectonics of the region post-Maastrichtian, due to changing azimuths of plate motions. As the relative movement between Australia and Antarctica became more N–S, the western STR, which up to then had been part of the Antarctic Plate, detached from that plate and became attached to the eastern part of the STR. During this change and with subsequent wrenching sub-parallel to the Tasman Fracture Zone, the western STR terrane underwent major deformation with wrenching and rotation of its constituent blocks. Seafloor spreading began south of the STR in the Paleocene, as spreading in the Tasman Basin slowed and then ceased.

With development of fast N–S seafloor spreading between Australia and Antarctica in the Middle Eocene, the region behaved largely as a passive margin, except for continuing deformation along the Tasman transform margin, affecting the western STR. By the end of the Eocene (*ca.* 33 Ma), the Australian and Antarctic continents had cleared, and by the Early Miocene (~23 Ma), the Southeast Indian Ridge had passed the southwestern tip of the STR; this marked the end of all wrench movements in the region.

Volcanism has been widespread in the South Tasman Rise region, and appears to be related to two main phases, (i) ~Campanian, related to margin breakup, and (ii) Eocene, related to motion over the Balleny Hotspot and possibly enhanced by regional tensile stress in the lithosphere associated with the onset of fast seafloor spreading. The Eocene volcanics comprise a number of large isolated basaltic seamounts, 1000–2000 m high, mainly in the L'Atalante Depression, but also on the ETP (Cascade Seamount), in the deep eastern margin of the STR, and in the South Tasman Saddle (Hill et al. 2001). Small volcanic cones, up to several hundred metres high, and associated flows are common in the region. In some places the seafloor is dominated by such volcanic terrain, notably in the South Tasman and East Tasman Saddles, on the summit area of the STR and on the far southeastern tip of the STR.

ILLUSTRATED TRANSECTS

One transect is included to illustrate the geology of the South Tasman Rise region (location shown in Fig. 34):

Plate 24: (line GA-202/12) extends northeast from the Australian–Antarctic Basin, across the South Tasman Rise, L'Atalante Depression and East Tasman Plateau, to the western Tasman Basin.

THREE KINGS RIDGE

INTRODUCTION

The Three Kings Ridge region (Figs 42, 43) is an area of interconnecting plateaus and elongate highs separated by narrow depressions that has a complex evolution related to extension, break-up and probable island arc formation through the Cretaceous and Cainozoic. It is part of a broad zone of north- and northwest-trending continental plateaus, ridges and narrow depressions, that includes the Lord Howe Rise, Norfolk Ridge and Three Kings Ridge, and extends from the Tasman Basin in the west to the South Fiji Basin in the east.

Physiographic features described here include: the Norfolk Ridge, West Norfolk Ridge, South Norfolk Basin, North Norfolk Basin, Nepean Saddle, Kingston Plateau, Bates Plateau, Loyalty Ridge and the Cook Fracture Zone (see Bernardel et al. (2002) for a detailed description). Many of these features have been surveyed in recent years by swath mapping (Mauffret et al., 2001).

GEOMORPHOLOGY

The Three Kings Ridge (TKR) is a broad, north–south trending feature that is asymmetrical in east–west profile, with the ridge crest being offset to the east, producing a broad western flank and a steep and narrow eastern flank. Crestal water depths generally range from less than 500 m to about 1000 m. The deeper part of the eastern flank is dominated by a chain of shallow seamounts that extends as far east as the Julia Fracture Zone (Kroenke & Eade, 1982).

The TKR is flanked to the east by the South Fiji Basin, to the north by the Cook Fracture Zone, to the south by the Vening–Meinesz Fracture Zone, and to the west by an area of complex morphology which connects the TKR with the Norfolk Ridge. The area of complex morphology to the west includes the Bates and Kingston Plateaus and the Nepean Saddle (Bernardel et al., 2002; Fig. 42). In the north, this area is partly separated from the TKR by the 2500–3500 m deep Philip and Cagou Troughs.

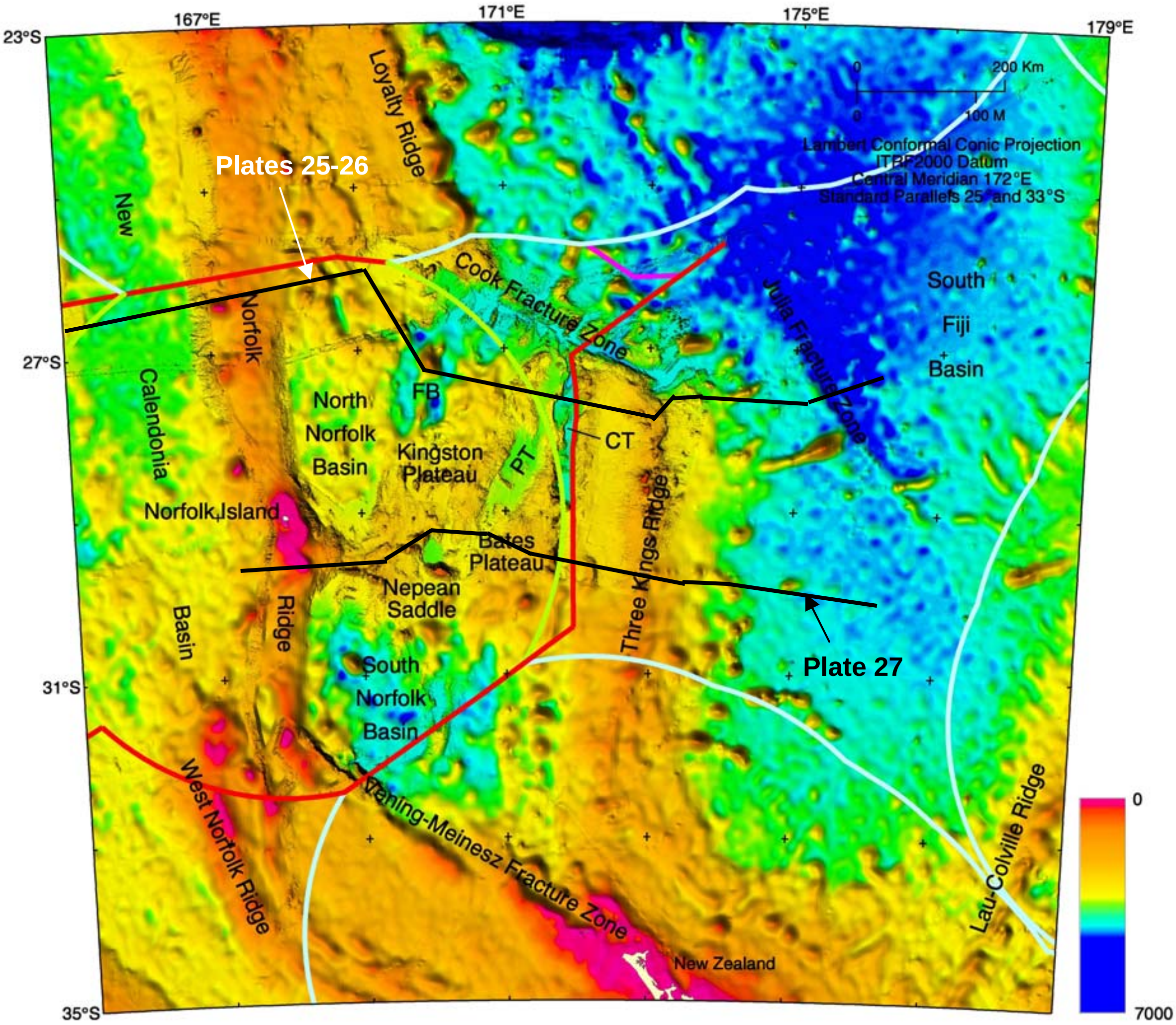

Figure 42: *Bathymetric image of the Three Kings Ridge region, showing the location of the illustrated seismic transects. Major geomorphological features are labelled. FB – Forster Basin; PT – Philip Trough; CT – Cagou Trough. Line colours: green – 200 M (Australia); magenta – outer limit of extended continental shelf (as submitted to UN CLCS, November 2004); light blue – 200 M (other countries); red – treaty boundaries.*

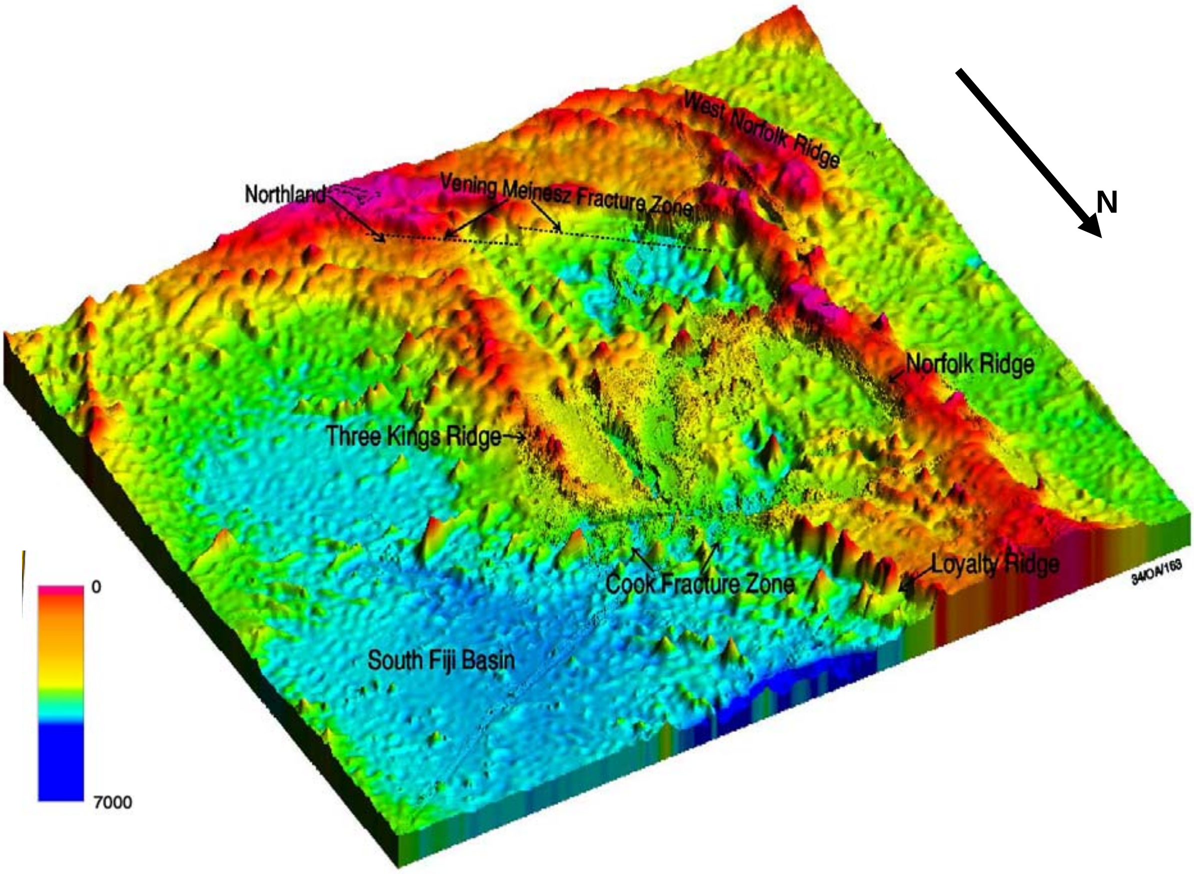

Figure 43: *3-D bathymetric image of the Three Kings Ridge region viewed from the north-northeast. Note the area of complex bathymetry between the Norfolk and Three Kings Ridges.*

Other significant structural and geomorphological features in the region include:

- The Norfolk Ridge, a narrow, north–south trending bathymetric high with summit water depths largely in the range of 500–1500 m. Norfolk Island is located midway along its length.
- The Norfolk Seamount Chain that is embedded in the western flank of the Norfolk Ridge.
- The West Norfolk Ridge, a broad, moderately-shallow and curvilinear system of ridges and basins connecting the southern part of the Norfolk Ridge to the North Island of New Zealand. Water depths are typically in the range of 500–2000 m.

- The South Norfolk Basin, an approximately rhomboidal-shaped, 3500–4000 m deep basin enclosed by the Norfolk Ridge to the west, the Vening-Meinesz Fracture Zone to the south, the Three Kings Ridge complex to the east; and the broad zone of complex morphology to the north that connects the Norfolk and Three Kings Ridges. The basin has an average water depth of 4000 m.
- The North Norfolk Basin, an area of moderately-deep (3000–3500 m) seafloor separated from the South Norfolk Basin by the Nepean Saddle, an area of undulating morphology that includes large and small seamounts.
- The Kingston Plateau, a NNE–SSW trending area of elevated seafloor comprising small seamounts, subdued ridges and scarps and interspersed elevated blocks and broad flat-lying depressions. The plateau lies at average water depths of 2000–3000 m.
- The Bates Plateau, a triangle-shaped morphologic high that extends northward from the South Norfolk Basin and connects to the Three Kings Ridge in the east. Water depths are typically 2100–2600 m.
- The Loyalty Ridge, a northern feature that is similar in form to the TKR but which trends north-northwest. It is offset from the TKR by the Cook Fracture Zone.

GEOLOGY

Plate tectonic setting

The ribbon-like development of the entire Tasman Basin/Dampier Ridge/Lord Howe Rise/New Caledonia Basin/Norfolk Ridge system (Fig. 26) is the result of a progressive rifting, breakup and seafloor spreading regime established along the eastern margin of Gondwana from approximately the Late Jurassic to the Early Eocene (see, for example, Yan & Kroenke, 1993; Symonds et al., 1999; Sdrolias et al., 2001; Crawford et al., 2003).

Geology of the TKR and adjacent provinces

A comprehensive study of the geology, history and crustal structure of the entire TKR region, synthesizing the interpretations of bathymetry, seismic reflection and refraction, dredge analysis and satellite gravity, has not yet been undertaken. The following is a summary of the state of knowledge as currently published.

The Three Kings Ridge is generally considered to be a product of arc volcanism (Kroenke & Dupont, 1982; Davey, 1982; Herzer & Mascle, 1996); however, its age and polarity have remained controversial. Its arc-related origin has been supported by the dredging of subduction-related basalts and shoshonites (Mortimer et al., 1998) and peridotites and boninites (Bernardel et al., 2002). Detailed swath bathymetry shows a seamount-dominated chain west of an elevated forearc-like front, which is consistent with an island-arc origin (Bernardel et al., 2002). Furthermore, the Loyalty Ridge to the north, which is the product of arc volcanism, is thought to be a continuation of the

Three Kings Ridge, separated from it by sinistral offset along the Cook Fracture Zone (Lapouille, 1977). An alternate view is that the Three Kings Ridge is a continuation of the Northland region of New Zealand (Mortimer & Herzer, 2000).

The **Norfolk Ridge** is probably a rifted ribbon of eastern Gondwana continental crust (Eade, 1988; Eissen et al., 1998; Mortimer & Herzer, 2000). This is partly confirmed by the dredging of granitoid-sourced, Late Cretaceous carbonaceous rocks on the adjacent West Norfolk Ridge (Herzer et al., 1999). The only exposed parts of the central Norfolk Ridge are Norfolk Island and the adjacent Philip Island. These islands were formed by volcanic activity between 3.1 and 2.3 Ma (Jones & McDougall, 1973; Aziz-ur-Rahman & McDougall, 1973).

The origins of the **North and South Norfolk Basins** are largely unknown. They have been interpreted as back-arc basins developed by westwards-directed subduction (Launay et al., 1982; Herzer & Mascle, 1996; Mortimer et al., 1998) or as remnant pieces of the Cretaceous Pacific Plate trapped between the rifted Norfolk Ridge and the Three Kings Ridge arc (Eade, 1988; Bernardel et al., 2002). However, recent studies suggest that this region of seafloor is in part a complex terrane resulting from a forearc collision front initially related to eastward subduction, that was fragmented by several episodes of extension and seafloor spreading to the east (Bernardel et al., 2002; Crawford et al., 2003).

The **South Fiji Basin** is a back-arc basin that developed in response to the subduction of Pacific Plate crust along the Tonga–Kermadec subduction system. It is believed to have formed entirely in the Oligocene by spreading at two triple junctions: one to the northeast of the northern tip of the TKR and one to the east (Sdrolias, 2000; Sdrolias et al., 2001).

The complexity of the geological evolution of the TKR region as part of the southwest Pacific is summarised by Crawford et al. (2003), who summarise the following regional tectonic evolution:

- Regional extension along the eastern margin of Gondwana after about 120 Ma (Barremian) led to the breakup of several microcontinent 'ribbons' which were separated by marginal ocean basins.
- At about 55 Ma (Paleocene–Eocene boundary) a major reorganisation of plate boundaries led to the initiation of subduction along former spreading centres in the easternmost of the marginal basins and the construction of an arc behind a boninitic forearc terrane.
- At 45 Ma (Middle Eocene), a major change in plate kinematics resulted in the initiation of west-dipping subduction.
- From 35–30 Ma (Late Eocene to Early Oligocene), the South Fiji Basin developed and the Vitiaz Arc extended southward.
- At about 25 Ma (Late Oligocene), a western spreading centre limb in the South Fiji Basin propagated into the Norfolk Ridge, detaching the Three

Kings Ridge and terminating the east-dipping subduction. Spreading in the South Fiji Basin ceased at about the same time.

This regional tectonic evolution has significant implications for the Three Kings Ridge region. It suggests east-dipping subduction of 70–55 Ma-aged oceanic crust beneath the Three Kings Ridge prior to 45 Ma, followed by arc reversal and the initiation of the modern west-dipping Tonga-Kermadec subduction system and formation of the backarc South Fiji Basin from about 35–25 Ma. This change in tectonism caused significant extension of the old forearc to the west of Three Kings Ridge, eventually initiating seafloor spreading in the Forster Basin at about 25 Ma, and movement along the Cook Fracture Zone. This history implies that the zone of relatively shallow seafloor connecting the Norfolk and Three Kings Ridge probably consists of extended forearc (peridotite and boninite), pieces of old Cretaceous down-going oceanic or transitional crust, and continental fragments rifted from the Norfolk Ridge in the Cretaceous.

The diversity of the tectonic evolution outlined above is largely responsible for the geological and geomorphological complexity of the entire region between the Tasman and South Fiji Basins, including the TKR and adjacent areas.

ILLUSTRATED TRANSECT

Two uninterpreted transects are included to illustrate the geology of the Three Kings Ridge region (locations shown in Fig. 42):

Plate 25–26: (lines GA-206/04, GA-177/LHRNR-BA, GA-177/LHRNR-B, GA-177/NZ-E & GA-177/NZ-F) extends from the east Australian margin, across the Tasman basin, Lord Howe Rise, New Caledonia Basin and Norfolk Ridge complex to the Three Kings Ridge and the South Fiji Basin.

Plate 27: (lines GA-LHRNR-C, GA-177/NZ-HA, & GA-177/NZ-H) extends east from the New Caledonia Basin to the South Fiji Basin, via the Norfolk Ridge, Nepean saddle, Kingston Plateau, Philip Trough, Bates Plateau, Cagou Trough and Three Kings Ridge.

WALLABY AND WEST EXMOUTH PLATEAUS

INTRODUCTION

The Wallaby and Exmouth Plateaus (Figs 44, 45) are large continental margin plateaus that formed on the northwest margin of Australia during the break-up of Gondwana in the Middle/Late Jurassic to Early Cretaceous. Except in the east where they abut the Australian landmass, both the Wallaby and Exmouth Plateaus are surrounded by oceanic crust.

Parts of the Wallaby and Exmouth Plateaus region have been explored to varying degrees since the late 1960s. A number of scientific surveys have acquired geophysical and geological data in the region which is now considered a type example of an intermediate volcanic rifted margin. A large quantity of petroleum exploration data has been recorded over parts of the Exmouth Plateau since the 1970s and a number of petroleum exploration wells have been drilled. These wells have made a number of significant hydrocarbon discoveries. Seven Ocean Drilling Program holes have been drilled on the Exmouth Plateau as well as two Deep Sea Drilling Program holes in the region, one on the Cuvier Abyssal Plain between the Wallaby and Exmouth Plateaus, and one on the Gascoyne Abyssal Plain to the west of the Exmouth Plateau. In addition, extensive sampling of the margins of both plateaus has been undertaken by vessels from both Australian and foreign institutions.

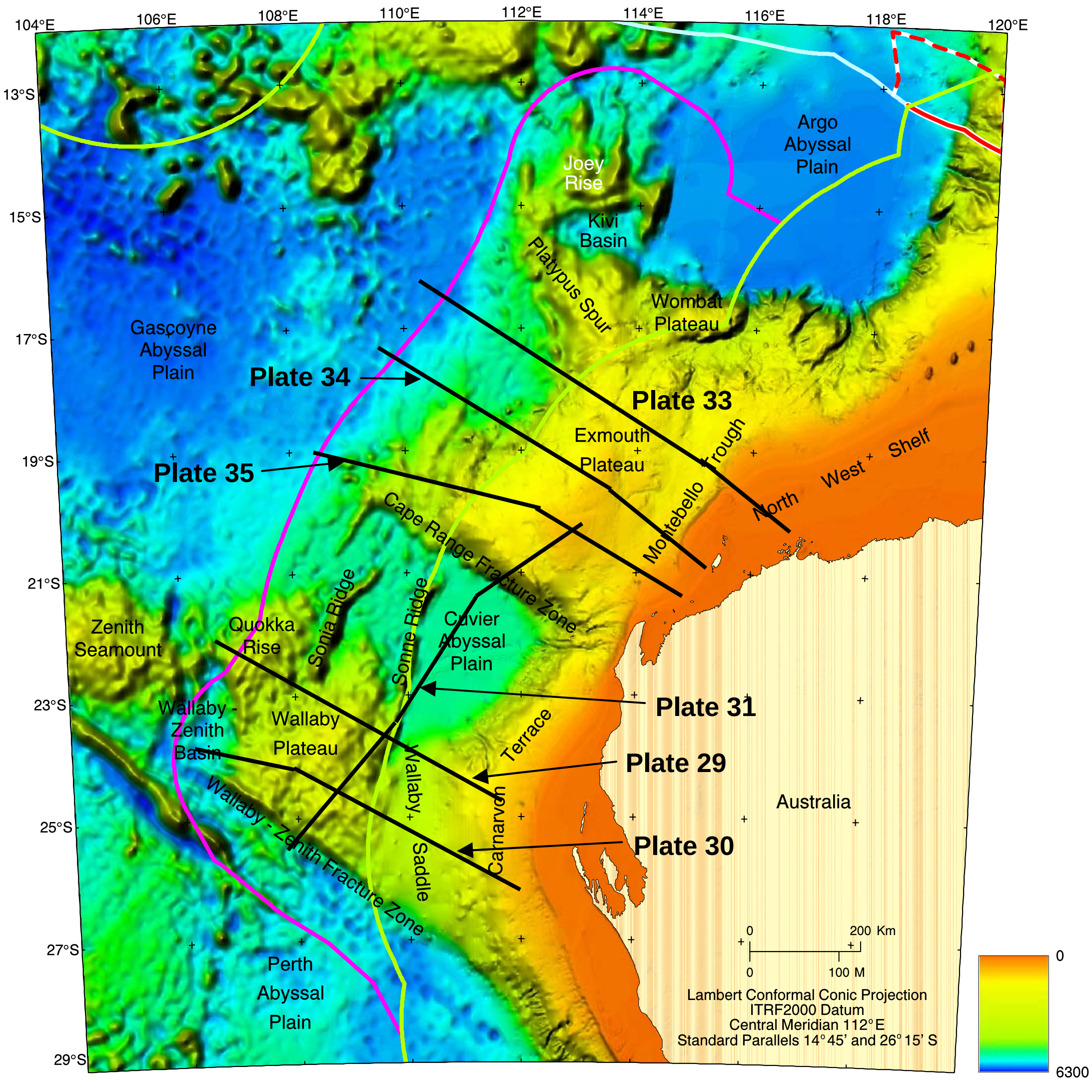

Figure 44: *Bathymetric image of the Wallaby and Exmouth Plateaus region, showing the location of the illustrated seismic transects. Major geomorphological features are labelled. Line colours: green – 200 M (Australia); magenta – outer limit of extended continental shelf (as submitted to UN CLCS, November 2004); light blue – 200 M (Indonesia); solid red and red & white dash – treaty boundaries with Indonesia.*

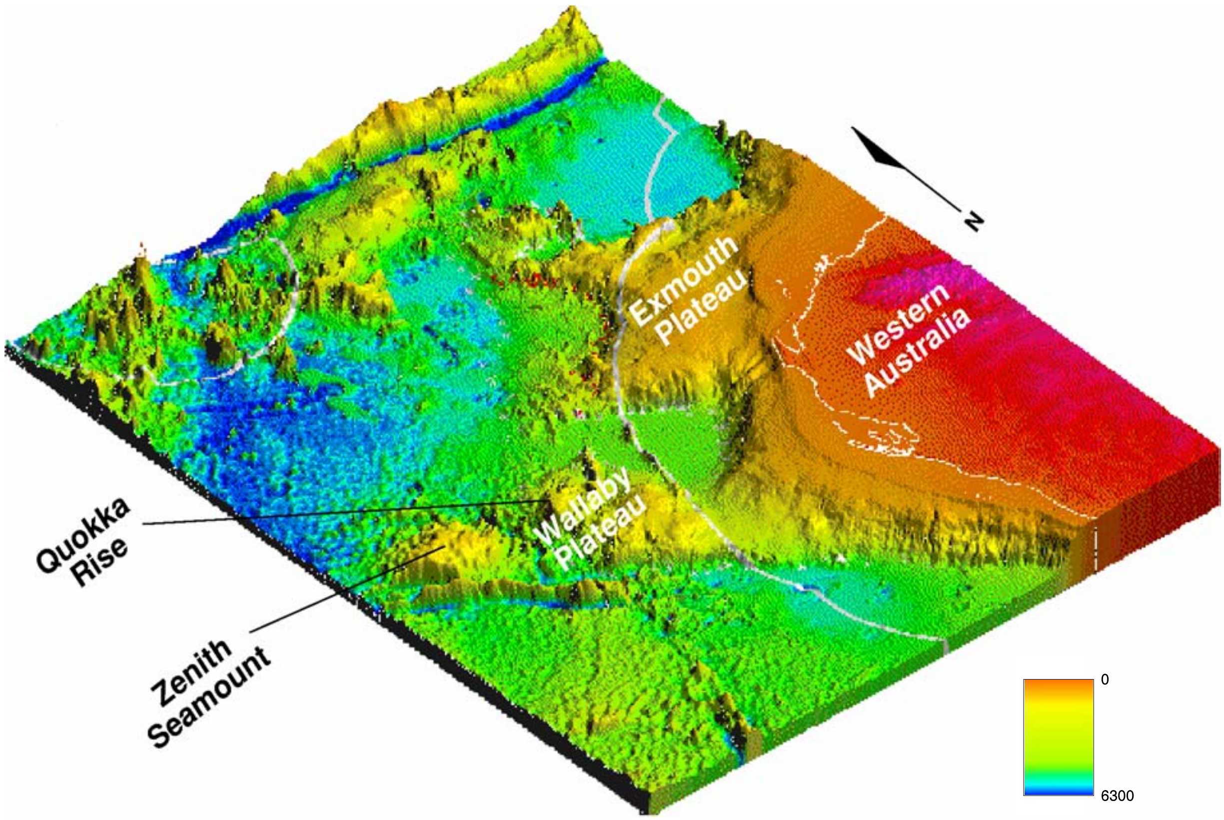

Figure 45: *3-D bathymetric image of the Wallaby and Exmouth Plateaus region viewed from the southwest. White line is 200 M line.*

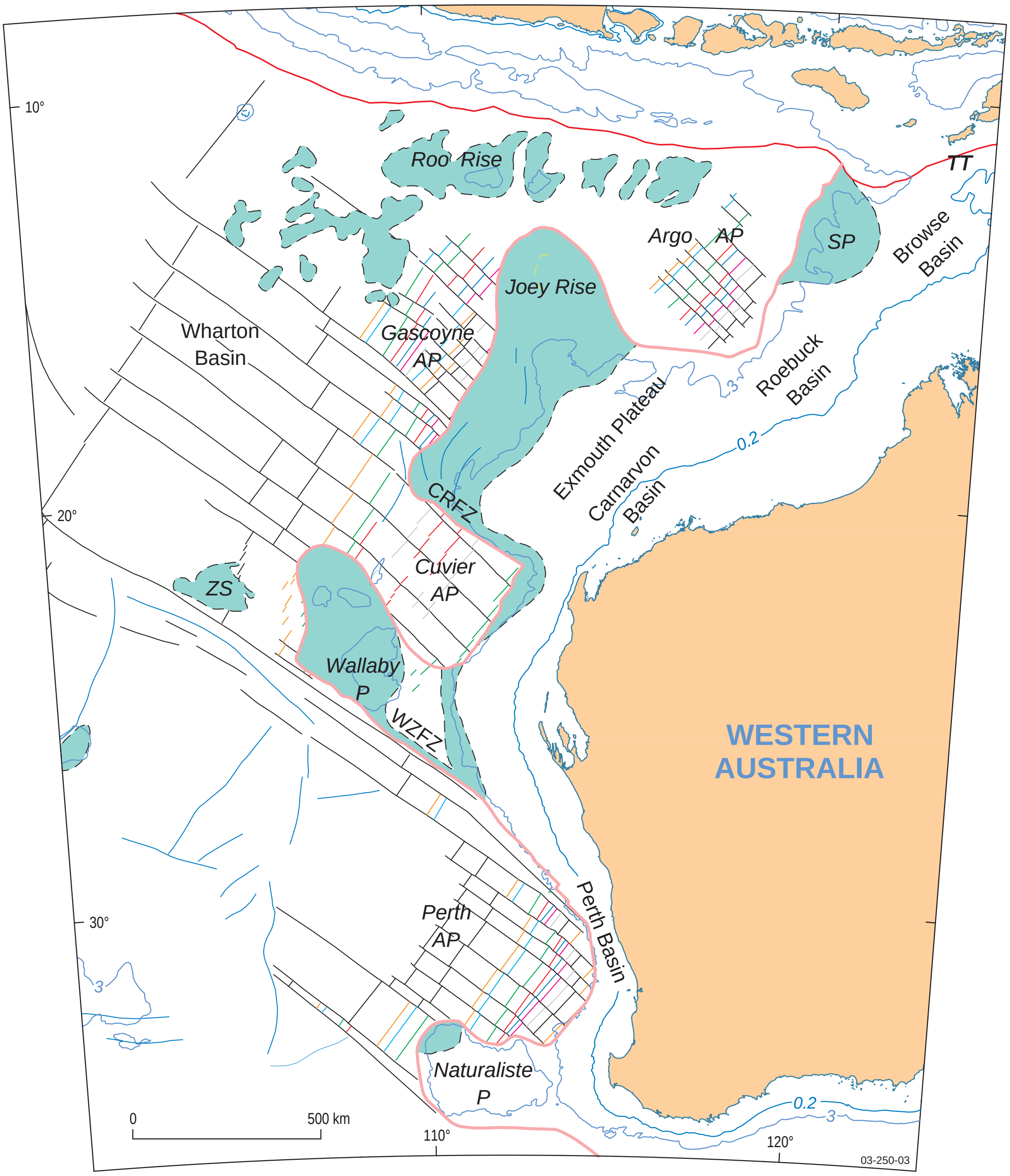

Figure 46: *Regional setting of the Wallaby and Exmouth Plateaus (after Stagg et al., 2004). Solid pink line is the approximate continent/volcanic margin – ocean boundary based on an integrated geophysical interpretation. Seafloor spreading magnetic anomalies are modified from Müller et al., (1998) and Mihut & Müller (1998a). Green areas are volcanic-dominated provinces on continental, volcanic margin, or oceanic crust. ZS – Zenith Seamount; SP – Scott Plateau; AP – abyssal plain; TT – Timor Trough; CRFZ – Cape Range Fracture Zone; WZFZ – Wallaby – Zenith Fracture Zone. Solid red line in the north is the approximate convergent plate boundary.*

WALLABY PLATEAU

Geomorphology

The 70 000 km^2 Wallaby Plateau is a large continental margin plateau situated between the Cuvier and Perth Abyssal Plains southwest of the Exmouth Plateau (Figs 44–46). It is connected to the Carnarvon Terrace/Bernier Platform on the inboard part of the western Australian margin by the 3500 m deep Wallaby Saddle. To the northwest lies another large bathymetric feature, the Zenith Seamount, separated from the Wallaby Plateau by the Wallaby–Zenith Basin, a 5000 m-deep northerly-trending trough with unusually rough topography.

The Wallaby Plateau is subdivided into two main parts: a northern high called the Quokka Rise, and a larger southern high that makes up much of the Wallaby Plateau proper (Falvey & Veevers, 1974; Veevers et al., 1985;). The southwest margin of the Wallaby Plateau is formed by the steep, 2000 m-high, northwest-trending scarp produced by the transform margin associated with the Wallaby–Zenith Fracture Zone. In contrast, the plateau's northeastern and northwestern slopes are more gentle (typically about 2º), merging with the Cuvier Abyssal Plain in the northeast and the Wallaby–Zenith Basin in the northwest. Water depths vary from less than 2100 m on the crest of the plateau to about 4500–5000 m at the base of the bounding slopes. The Quokka Rise has a minimum water depth of just over 2500 m.

Two ridges extend from the northern margin of the Wallaby Plateau part way across the Cuvier Abyssal Plain: the Sonne Ridge in the east and the Sonja Ridge in the west. The Sonne Ridge is composed of several culminations where minimum water depths range from 3500 m in the north to 3000 m in the south where it merges with the eastern flank of the Wallaby Plateau. It is narrow and asymmetric in the north and broader and more symmetric in the south. The Sonja Ridge is shorter and more continuous, with crestal water depths ranging from 2700 m in the south to approximately 4000 m in the north.

Geology

A combination of initial continental extension, ongoing volcanic margin extension, extrusion and intrusion, and seafloor spreading, rift propagation and ridge jumping in a zone adjacent to a developing transform margin segment explains many of the features found in the Wallaby Plateau region (Sayers et al., 2002). The region remains one of the more poorly-understood segments of the western Australian margin, mainly because of its complex history.

Plate Tectonic Setting

The Wallaby Plateau is part of the western Australian continental margin (Fig. 46). It is surrounded to the north by oceanic crust of the Cuvier Abyssal Plain, to the south by oceanic crust of the Perth Abyssal plain, immediately to the northwest by a complex volcanic-margin province (including the Zenith Seamount and Wallaby–Zenith Basin), and further to the northwest by oceanic crust of the Gascoyne Abyssal Plain/Wharton Basin. It is located immediately to the northeast of the transform Wallaby–Zenith Fracture Zone.

As shown in Figure 46, seafloor spreading magnetic anomalies in the region are interpreted to have a consistent northeast-southwest orientation. The oldest seafloor anomalies identified on oceanic crust adjacent to the Wallaby Plateau are of Early Cretaceous (Valanginian) age in the Cuvier Abyssal Plain (Müller et al., 1998).

Geology and evolution of the Wallaby Plateau

The major volcano-tectonic provinces and features of the Wallaby Plateau region are shown in Figure 47. A stratigraphic chart for the Wallaby Plateau region is shown in Plate 28.

The current understanding of the geology and evolution of the Wallaby Plateau region is based largely on seismic data and seabed sampling (Sayers et al., 2002). Hopper et al. (1992) and Colwell et al. (1994b) described seaward dipping reflectors at the base of the continental slope east of the Wallaby Plateau, and interpreted them as volcanic wedges emplaced during break-up. Frey (1998), Frey et al. (1998) and Symonds et al. (1998) analysed volcanic features of the region and identified several seismic volcanic facies that had also been recognised on other typical volcanic margins. Among them were seaward-dipping reflector sequences (SDRSs), hyaloclastic build-ups, sills and landward flows, indicating voluminous magmatism at about the time of, or immediately following continental break-up. A cross–section based on seismic reflection data across the Wallaby Plateau region is shown in Figure 48.

Dredging of the margins of the Wallaby Plateau and of the Sonne Ridge has so far only recovered volcanic rocks (von Stackelberg et al., 1980; Colwell, Graham et al., 1990; Exon, 1994a). A re-interpretation of seismic data in the region (Colwell et al., 1994b), and in particular the presence of well-layered crust beneath the central part of the plateau, suggested that the basement structure under the plateau was of composite origin, similar to that of other large plateaus (e.g. the Kerguelen Plateau).

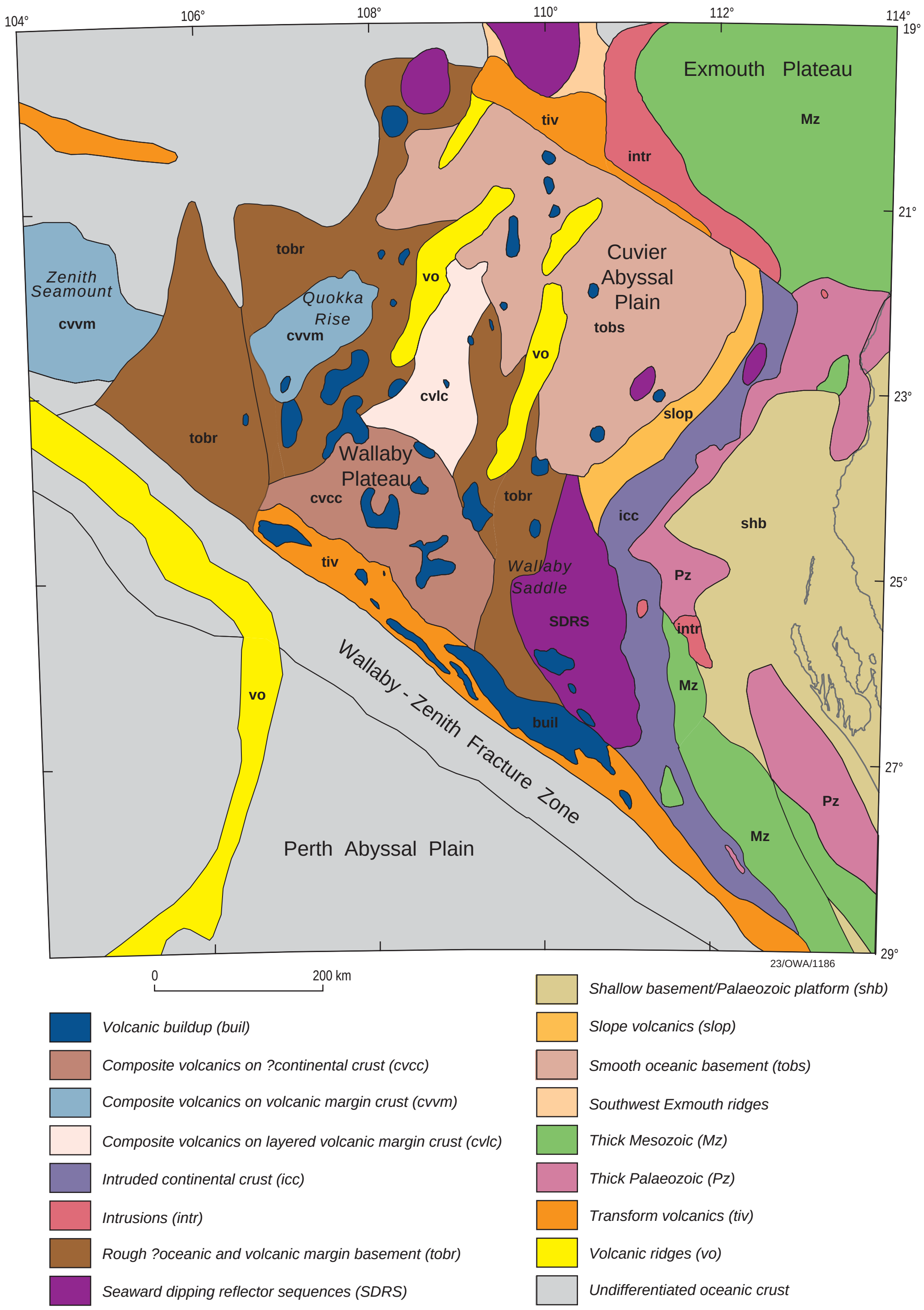

Figure 47: *Distribution of volcano-tectonic provinces and features in the Wallaby Plateau region (after Sayers et al., 2002).*

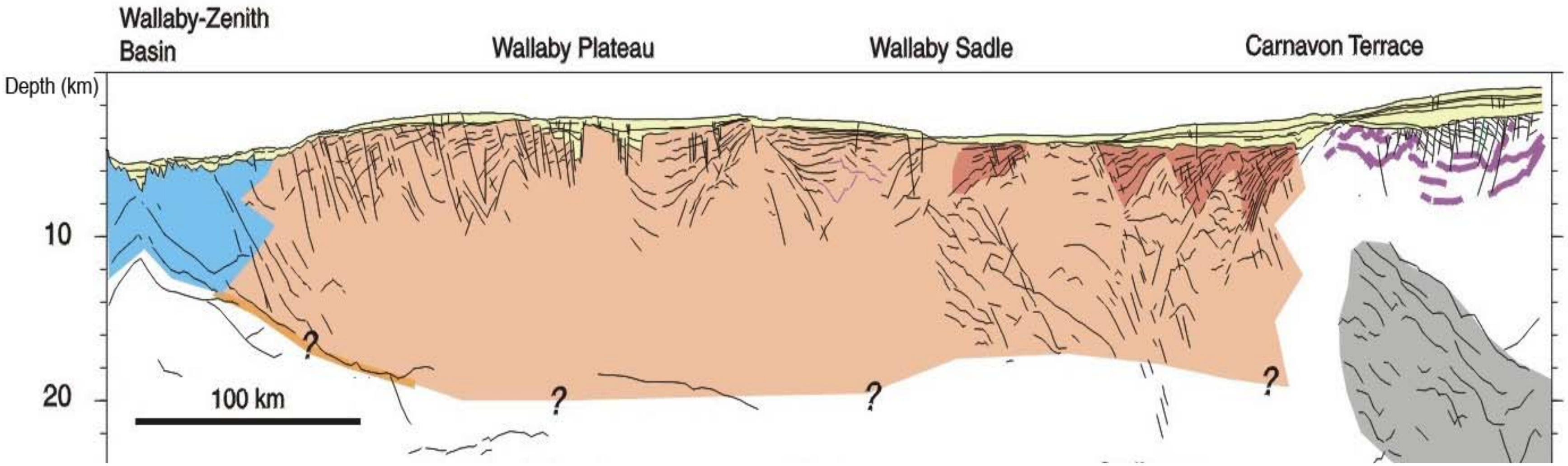

Figure 48: *Cross-section across the Wallaby Plateau region based on Geoscience Australia deep-seismic data (after Symonds et al., 1998). SDRSs are shown in dark brown; highly-reflective lower crust in grey; sills as purple lines; mixed volcanic-margin and continental crust is shown in light brown; ?oceanic crust in blue; and the overlying sediments in yellow.*

Planke et al. (1996) compared the tectonic evolution of the western Australian margin to the well-studied volcanic margins of the Atlantic, and proposed an alternative model for the formation of the Wallaby Plateau. According to their model, the plateau has a core of extended continental crust mixed with magmatic material, and is blanketed by basaltic extrusive bodies. Symonds et al. (1998) suggested that the Wallaby Plateau complex, with its variable seismic character, trace element indications of sub-continental lithosphere (Colwell et al., 1994b), and rift-transform corner setting, is likely to have a complex development history. Like Planke et al. (1996, 1997), Symonds et al. (1998) interpreted the Wallaby Plateau complex to be a composite feature cored by some extended continental blocks that have been modified and blanketed by voluminous magmatism associated with breakup and transform development along its southwestern margin. The latest model for the evolution of the region is shown in Figure 49. In some places, the volcanic margin crust may be newly created by magmatism that occurred during continental margin development; however, in other places, such as beneath parts of the Wallaby Plateau and Saddle, the margin-forming magmatic processes modified existing extended continental crust (Sayers et al., 2002).

Illustrated Transects

Three transects are included to illustrate the geology of the Wallaby Plateau region (locations shown in Fig. 44):

Plate 29: (line GA-135/05) extends east-southeast from the Quokka Rise, across the northern Wallaby Plateau and Wallaby Saddle, to the Carnarvon Terrace.

Plate 30: (line GA-135/08) extends southeast across the main part of the Wallaby Plateau and the Wallaby Saddle to the Carnarvon Terrace.

Plate 31: (line GA-135/11) is a margin strike line that extends northeast from the Perth Abyssal Plain, across the Wallaby Plateau and the Cuvier Abyssal Plain to the Exmouth Plateau.

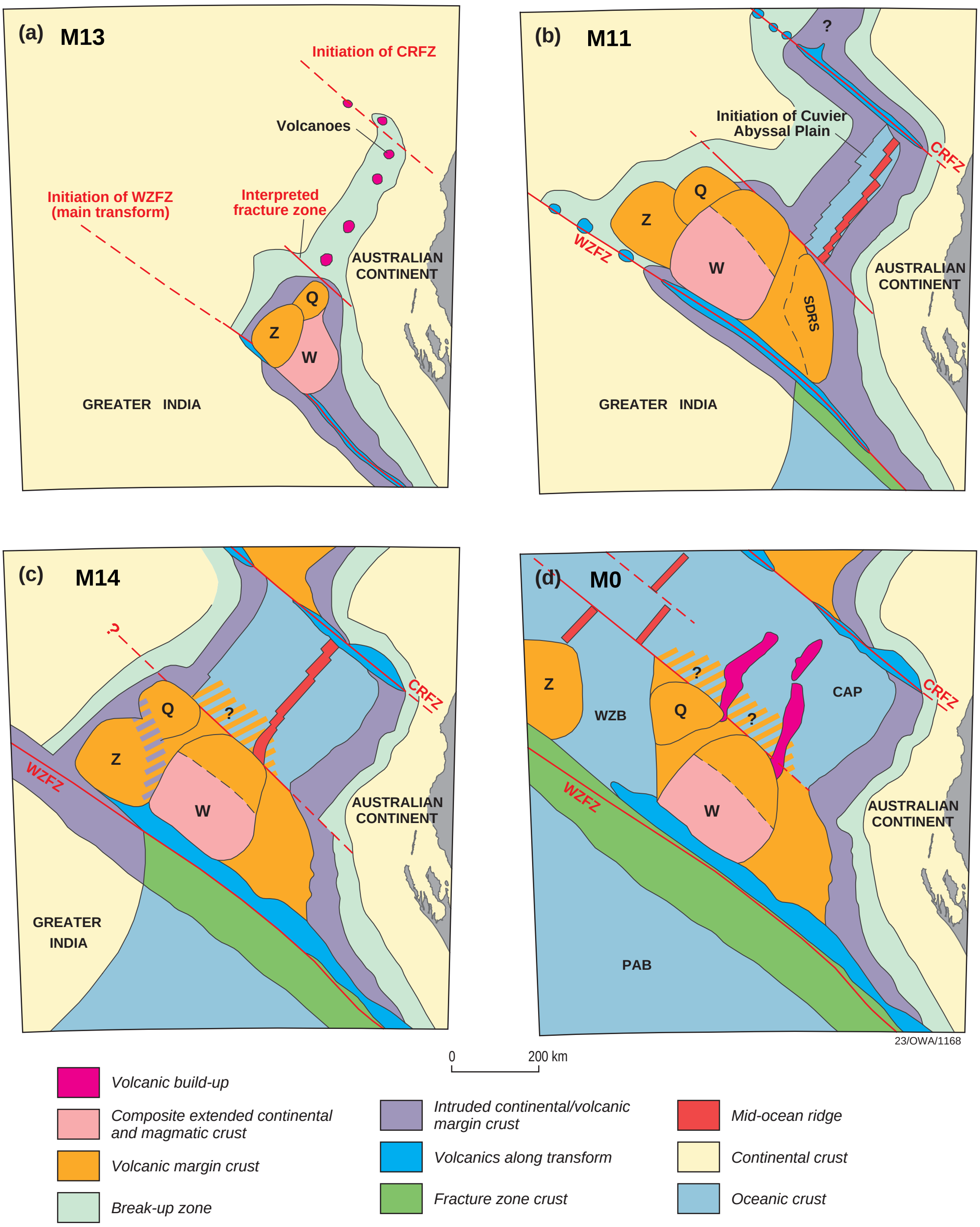

Figure 49: *Conceptual evolution of the Wallaby Plateau region during the Early Cretaceous (after Sayers et al., 2002). (a) chron M13 ; (b) chron M11; (c) chron M4; (d) chron M0. WZFZ – Wallaby-Zenith Fracture Zone, CRFZ – Cape Range Fracture Zone, W – central part of the Wallaby Plateau, Q – Quokka Rise, Z – Zenith Seamount, CAP – Cuvier Abyssal Plain, PAB – Perth Abyssal Plain, WZB – Wallaby – Zenith Basin.*

EXMOUTH PLATEAU

Geomorphology

The Exmouth Plateau is a large mid-slope plateau on the continental margin of northwest Australia (Figs 44–46). The plateau is flanked by the North West Shelf to the southeast, the Argo Abyssal Plain to the north, the Gascoyne Abyssal Plain to the northwest and the Cuvier Abyssal Plain to the southwest. These abyssal plains typically lie at depths of 5000–5700 m and are the product of seafloor spreading in the north between the Exmouth Plateau and Argo Land starting in the Mid/Late Jurassic, and between the Exmouth Plateau and Greater India in the northwest and southwest starting in the Early Cretaceous (Fig. 46). The shallowest area of the plateau is the NNE-trending and centrally located Exmouth Plateau Arch, which culminates at a depth of about 780 m; the deepest parts of the plateau's margins extend down to approximately 5000 m. The Montebello Trough, marking the inboard flank of the plateau, has axial water depths of about 1200 m and debouches northwards through the Montebello Canyon into the Argo Abyssal Plain.

The broadly east–west trending, commonly steep (up to 17º in places) northern margin of the Exmouth Plateau is morphologically complex. It is dominated by a range of plateaus, canyons and spurs, that include the Wombat Plateau, the Platypus Spur and the Joey Rise (Fig. 44). The plateaus and spurs commonly lie at water depths of 2000–2500 m, while the canyons have frequently cut down into the seafloor by 1000–2000 m. The Joey Rise is surmounted by three north-south trending, elongated highs separated by depressions. The southwestern part of the Joey Rise is connected to the Exmouth Plateau proper by the Platypus Spur and lies at water depths of 2500–3000 m. The Joey Rise shallows to slightly less than 2500 m. The Platypus Spur, the Joey Rise, and a saddle area between the Joey Rise and a complex seamount region to the north, jointly separate the Argo and Gascoyne Abyssal Plains (Fig. 44).

The northwestern (Gascoyne) margin of the Exmouth Plateau is characterised by a generally rugged, low-gradient (commonly less than 2º) lower slope zone reflecting its origin as part of a volcanic rifted margin in which volcanic flows, seaward-dipping reflector sequences (SDRS) and volcanic intrusions dominate the geology (see Geology section). Canyon development is limited. The lower slope along this part of the margin is about 60 km wide in the north and widens to about 250 km in the central part of the margin. The southwestern corner of the Exmouth Plateau is characterised by a series of north- to northeasterly-trending ridges of probable volcanic origin which splay from the plateau.

The southwestern (Cuvier) margin of the Exmouth Plateau is strongly linear and steep, reflecting its origin as a transform margin which is underlain by the Cape Range Fracture Zone (CRFZ). Gradients on the lower continental slope of the CRFZ are typically steep (6–15º). Southwest of the CRFZ, the Cuvier Abyssal Plain separates the Exmouth Plateau from the Wallaby Plateau.

Geology

The Exmouth Plateau is a major component of the Northern Carnarvon Basin of the North West Shelf geological province. This basin is the southernmost element of the late Palaeozoic to Cainozoic Westralian Super-basin that underlies the continental margin of Australia from North West Cape to the Timor Sea (Yeates et al., 1987). The Northern Carnarvon Basin is currently Australia's prime hydrocarbon province, with production derived from giant gas-condensate fields and a number of small to medium-size oil fields. On the Exmouth Plateau, hydrocarbon discoveries include the yet-to-be-developed giant Scarborough Gas Field on the southern-central part of the plateau.

The current understanding of the geology and evolution of the plateau comes from a number of sources. These include:

- petroleum exploration and Geoscience Australia surveys that provide high-quality seismic data sets tied to petroleum exploration wells (see, for example, Barber, 1988; Stagg & Colwell, 1994; AGSO North West Shelf Study Group, 1994; Longley et al., 2002);
- papers dealing with the breakup history of the margin, based largely on marine magnetic lineations (for example, Fullerton et al., 1989; Müller et al., 1998);
- papers dealing with the continental margins and adjacent abyssal plains, based largely on reflection seismic data and seabed sampling (for example Exon & Willcox, 1980; von Stackelberg et al., 1980; Exon et al., 1982; Mutter et al., 1989; papers in Exon, 1994b, including Exon & Colwell, 1994; Brown et al., 2003b);
- reports dealing with the drilling results of the Deep Sea Drilling Project and the continuously cored Ocean Drilling Program (for example, Veevers, Heirtzler et al., 1974; Haq, von Rad, O'Connell et al., 1990; von Rad et al., 1992b; von Rad, Haq et al., 1992; Gradstein, Ludden et al., 1992); and
- a major study/synthesis of the geology and evolution of the Exmouth Plateau and adjacent areas by undertaken by AGSO in the late 1990s (Stagg et al., 2004).

The plateau and surrounds can be divided into a number of tectonic elements (Figure 50). A stratigraphic chart for the Exmouth Plateau region is shown in Plate 32 and a cross-section of the plateau is illustrated in Figure 51.

Most of the Exmouth Plateau is underlain by 10–15 km of generally flat-lying, faulted sedimentary section that was mainly deposited during the extension that preceded breakup of Argo Land and Australia in the Middle Jurassic, and Greater India and Australia in the Early Cretaceous. This includes a very thick Triassic section. Around the time of breakup, the Exmouth Plateau was strongly modified by the effects of magmatism. This phase of development caused intrusion of the central part of the plateau, and modified and expanded its developing margins, forming broad zones of volcanic margin crust, spurs, and numerous other typical volcanic margin features. Since breakup, the Exmouth Plateau and its environs have been largely sediment-starved, with the mid-Cretaceous to Cainozoic marine section rarely being more than a few hundred metres thick. Since breakup, the Exmouth Plateau has slowly subsided to its present water depths, mainly in response to thermal cooling rather than sediment loading.

The margins of the Exmouth Plateau are defined by a number of different features that developed in response to continental breakup. The northern (Argo) margin is structurally the most complex, reflecting a mixed rift/transform setting that has overprinted underlying early Palaeozoic structures. The major manifestation of volcanic activity on this margin is the Joey Rise, which is connected to the main part of the plateau by the Platypus Spur.

The Wombat Plateau is one of the most striking features on the northern margin of the Exmouth Plateau. It was drilled during Leg 122 of the Ocean Drilling Program, revealing a geological history that includes the development of a series of shallowing-upward cycles of fluvio-deltaic siliciclastics in the Late Triassic, carbonate platform development in the latest Triassic, volcanism and major block faulting in the Jurassic prior to breakup along the margin, and subsequent rapid subsidence (von Rad et al., 1992a).

The northwestern (Gascoyne) margin of the Exmouth Plateau is characterised by a rugged, commonly low-gradient slope and a very thin cover of sediment. The margin is a typical example of an intermediate volcanic rifted margin (Symonds et al., 2000). Igneous extrusions and intrusions dominate the margin, strongly influencing the physiography and masking the underlying sedimentary section (Symonds et al, 1998; Frey et al., 1998). A number of volcanic provinces can be recognised, including landward flows, slope volcanics, seaward-dipping reflector sequences (SDRS), and buildups of varying styles and origins (Fig. 50). Multiple SDRS wedges are separated by blocks of interpreted continental origin and volcanic build-ups.

A recent study of the western part of the Exmouth Plateau and adjacent areas using a combination of seismic reflection data and potential-field (gravity and magnetics) modelling has provided further insights to the nature of the Gascoyne margin (Direen, 2004a; Direen et al., 2004). Lineated magnetic anomalies in this area clearly overlie SDRSs and not, as previously interpreted, normal oceanic crust. This has ramifications for the previously published Valanginian and older ages of breakup along this margin.

The strongly linear, southwest (Cuvier) margin of the Exmouth Plateau reflects its origin as a transform margin. Since breakup, this margin has undergone major uplift (more than 2 km) with consequent erosion of the Triassic-Cretaceous sedimentary section (Stagg et al., 2004). This uplift was partly due to a thermal anomaly associated with the close proximity of the Cuvier spreading centre to the margin and partly to a significant intrusion beneath the uplifted rim of the Exmouth Plateau, and transform-related volcanics adjacent to the Gascoyne margin located further to the west.

Illustrated Transects

Three transects are included to illustrate the geology of the continental margin of the west Exmouth Plateau ((locations shown in Fig. 44):

Plate 33: (lines GA-162/07, GA-128/08 & GA-101R/09) extends southeast from the Gascoyne Abyssal Plain, across the Exmouth Plateau to the flank of the Pilbara Craton, on the continental shelf.

Plate 34: (lines GA-162/01, GA-110/15 & GA-101R/07) extends southeast from the Gascoyne Abyssal Plain, across the Exmouth Plateau to the Barrow Sub-basin beneath the continental shelf.

Plate 35: (lines GA-162/04, GA-135/01 & GA-110/12) extends southeast from the Gascoyne Abyssal Plain, across the southern Exmouth Plateau to the Exmouth Sub-basin beneath the outer continental shelf.

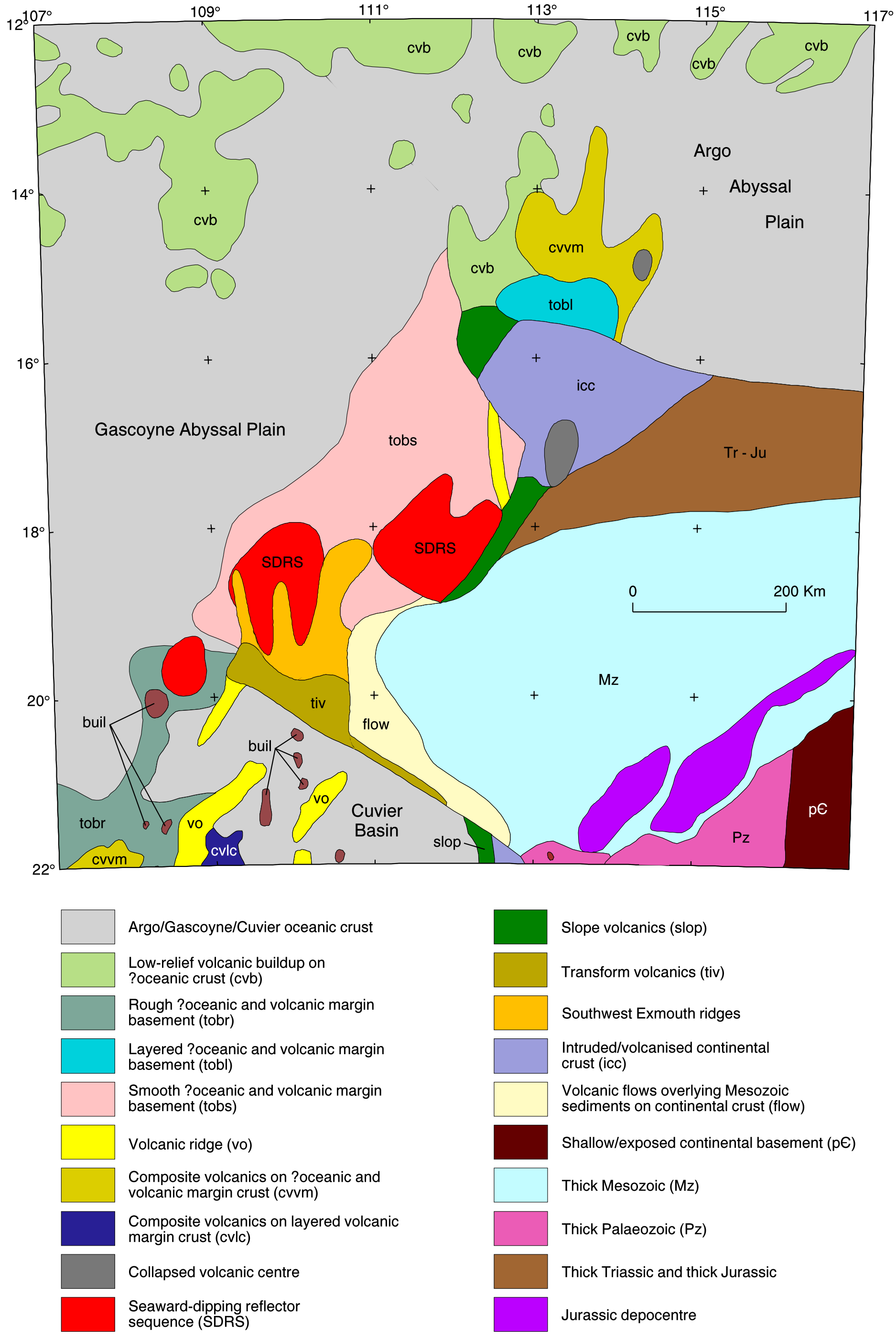

Figure 50: *Tectonic elements of the Exmouth Plateau and its surrounds (after Stagg et al., 2004).*

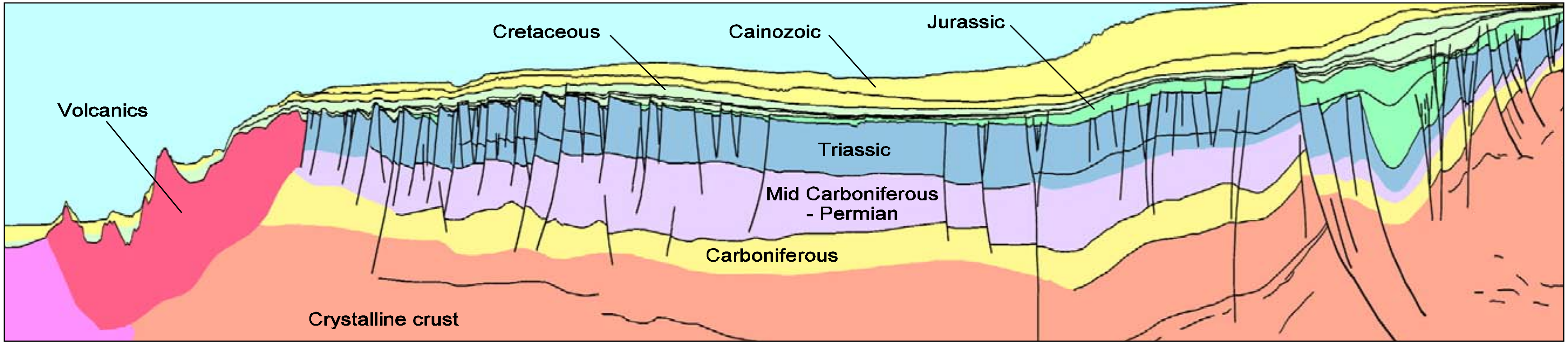

Figure 51: *Cross section (based on a composite seismic reflection profile) extending from the inner margin of the Northern Carnarvon Basin in the east, across the Exmouth Plateau to the Gascoyne margin of the plateau in the west, showing the main structural elements (after AGSO North West Shelf Study Group, 1994).*

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