3. Review and Synthesis of Literature for the East Marine Region

3.1. INTRODUCTION

The tectonic history, oceanography, late Quaternary evolution and surficial sedimentology of the East Marine Region (EMR) have been the focus of extensive research by various authors and government agencies at different temporal and spatial scales. The eastern margin of Australia covers a vast area and extends from tropical to temperate latitudes and although over 350 references were reviewed for this report there are still large gaps in our basic knowledge. The EMR includes notable geomorphic features such as reefs, seamounts and canyons and covers an extensive area of shelf, slope and abyssal plain/deep ocean floor along with a small area of abyssal rise. Geoscience Australia has contributed extensively to the study of the region and has published records and bulletins on the southeast margin offshore of New South Wales and southern Queensland (Marshall, 1977; Davies, 1979; Marshall, 1980; Colwell & Roy, 1983; Colwell & Coffin, 1987; O'Brien & Heggie, 1990; Heggie et al., 1992; Hill, 1994; Tsuji et al., 1997; Glenn et al., 2007, O'Brien et al., 1994; Maung et al., 1997; Stephenson & Burch, 2004), the Lord Howe Rise (Willcox et al., 1981; Dickens et al., 2001; Stagg et al., 2002; Willcox & Sayers, 2002; Van de Beuque et al., 2003; Alcock et al., 2006), northeast Queensland margin offshore of the Great Barrier Reef (Mutter, 1977; Symonds et al., 1992; Struckmeyer et al., 1994; Wellman et al., 1997; Isern et al., 1998; Earl et al., 2002; Exon et al., 2005; Exon et al., 2006a) and the Norfolk Island region (Bernardel et al., 2002; Exon et al., 2004c).

Several expeditions of the international Deep Sea Drilling Project/Ocean Drilling Program (DSDP/ODP Legs 21, 29, 30, 90, 133 and 194) have made borehole and regional studies in and adjacent to the EMR (Burns et al., 1973; Kennett et al., 1974; Andrews et al., 1975; Kennett et al., 1986; Davies et al., 1991a; McKenzie et al., 1993; Isern et al., 2002; Anselmetti et al., 2006). Drill Sites 206, 207, 208, 209, 210, 283, 287, 587, 588, 589, 811-826 and 1192-1199 provide useful data. The location and water depth of these sites is listed in Appendix C Table 3.1. Key geomorphic features and provinces of the EMR have been mapped using a consistent bathymetric grid of Australia's EEZ (National Bathymetric Map Series, 1976; Kroenke et al., 1983; Webster & Petkovic, 2005; Heap and Harris, in press-a) and past scientific literature (Fig. 1.2 & 3.1).

The EMR has been divided into the four geomorphic provinces as defined in Table 1.1 (Fig. 4.1): the shelf; middle shelf; slope; rise; and abyssal plain/deep ocean floor. These divisions are made on the basis of water depth and the geomorphic provinces described in a recent study on the geomorphology of the Australian margin (Heap and Harris, in press-a). The geomorphic features defined in Table 2.1 are described individually within these provinces.

3.1.1 Tectonic History

The principal geomorphic features in the East Marine Region (EMR) of the Tasman and Coral Seas were formed during rifting and thinning of the continental crust of eastern Australia in the late Cretaceous between ~110 and 80 million years ago (Ma) followed by a period of seafloor spreading with the formation of new basaltic oceanic crust. The sea floor spreading continued until early Eocene (~52 Ma) and created the ocean basins, failed-rift troughs, ridges and plateaus. Since the cessation of seafloor spreading periodic volcanism and subsidence have determined the basic present-day seafloor geomorphology.

3.1.1.1. Tasman Sea

The main geomorphic features in the Tasman Sea are basins, plateaus, ridges and seamounts (Fig. 3.1). Tectonic features related to the geomorphology are the boundaries between oceanic and continental crust, faults and fracture zones, oceanic ridges and younger volcanism delineated by Stagg et al., (1999a) and shown in Figure 3.2. Prior to seafloor spreading a rifting phase started in the middle to late Cretaceous (ca. 110 Ma) with stretching and thinning of the continental crust and the development of topographic highs and basins due to the associated normal and strike-slip faulting (Fig. 3.3; Gaina et al., 1998b, Norvick et al., 2001). The newly formed basins filled, or partially filled, with sediment in a fluvial or lacustrine environment (Stagg et al., 1999b). Volcanism accompanied this rifting but data as to how widespread it was are lacking. On the eastern margin of the Tasman Basin DSDP Site 207 on the southern Lord Howe Rise sampled rhyolitic tuffs and flows dated at 94 Ma (McDougall and van der Lingen, 1974). They have been interpreted as having been deposited in a subaerial or shallow marine environment (Burns et al., 1973). North of the Tasman Basin the rift-related Whitsunday-Proserpine volcanics on the Queensland coast are dated at 120 - 100 Ma (Ewart et al., 1992). On the southeastern Australian margin emplacement of igneous intrusions occurred along the NSW south coast near Montague Island which is composed of lavas from nearby Mt Dromedary and a quartz syenite from this lava is dated as 97 Ma. Offshore of Montague Island, on the mid-continental slope, a quartz monzodiorite (101 Ma) was also emplaced during the rifting phase (Hubble et al., 1992). Rocks of similar age and provenance occur south of the EMR in the Gippsland Basin (O'Halloran and Johnstone, 2001).

Magnetic anomalies in the basaltic oceanic crust beneath the Tasman Sea have been used to determine the age, direction and rate of seafloor spreading (Ringis, 1972, Hayes and Ringis, 1973, Weissel and Hayes, 1977, Shaw, 1979, Gaina et al., 1998b). Rifting was not uniform and resulted in the continental crust being broken up into 13 micro-continental blocks (Fig. 3.3; Gaina et al., 1998b). Seafloor spreading (breakup) in a SW-NE direction started south of 38^0S at ~84 Ma (Campanian, Chron 33) where what is now called the Monawai Ridge met its conjugate margin at Bass Strait. Initially, the spreading rate was very slow (4 mm yr^{-1}) forming a narrow ocean propagating to the north.

Spreading reached the southern end of the EMR at 80 Ma when the southern block of the Dampier Ridge separated from what is now the base of the NSW continental slope. At 79 Ma the spreading rate increased to 22 mm yr^{-1} (Van de Beuque et al., 2003). Rifting between the Dampier Ridge and the Lord Howe Rise occurred from 79 to 71 Ma before separation from the NSW margin. This continental crust attenuation formed the Middleton Basin in the north and the Lord Howe Basin in the south. There is no magnetic evidence for oceanic crust in these basins (van de Beuque et al., 2003). Seafloor spreading started west of the Dampier Ridge at 73 Ma and continued to propagate northward until it ceased at 52 Ma (early Eocene) when the direction of spreading was SSW-NNE (Fig. 3.11). Magnetic anomalies date the age of the oceanic crust adjacent to the continental crust at the southern boundary of the EMR as 80 Ma, at Coffs Harbour (32^0S) as 67 Ma, and offshore of Fraser Island as 60 Ma (Fig. 3.4; Gaina et al., 1998a). These ages indicate the maximum age for marine sediments resulting from the transgression as a narrow sea flooded the subsiding oceanic crust as it propagated northwards.
The geology on the conjugate margins is related in the following way:

- Kenn Plateau (far northern Lord Howe Rise) with the Maryborough and Nambour Basins of southern Queensland;
- Northern Lord Howe Rise (northern Dampier Ridge; Middleton Basin; Faust and Capel Basins) with the New England Fold Belt and the northern margin of the Sydney Basin;
- Central Lord Howe Rise (southern Dampier Ridge, Lord Howe Basin and Gower Basin) with the Sydney Basin and the Lachlan Fold Belt.

Regional structural lineaments related to tectonic events still shape the present seabed morphology by forming the margins of the major crustal blocks and scarps within those blocks. In particular the lineaments have two trends: northeast-southwest parallel to the Tasman Sea spreading; and northwest-southeast related to the ridge and basin complex in the Norfolk Island region (Stagg et al., 1999b). One of these lineaments, known as the Barcoo-Elizabeth-Fairway Lineament, extends from the slope offshore of Jervis Bay for about 1,800 km northeast as far as the Norfolk Ridge. It dissects the Dampier Ridge, separates the Middleton Basin from the Lord Howe Basin and continues across the Lord Howe Rise (LHR) as the mostly sediment buried Elizabeth-Fairway Lineament. Another major lineament with topographic expression is the Vening-Meinsez Fracture Zone which trends NW from the southern EMR around Norfolk Island and crosses the Lord Howe Rise.

The fabric of the ocean crust in the Tasman Basin is clearly defined in satellite-derived sea surface gravity anomalies and shows major fracture zones (Smith and Sandwell, 1997). These fracture zones were a controlling factor in the shape of the continental crust breakup and their imprint still exists in the bathymetry of the base of the Australian stet slope and its conjugate margin on the Dampier Ridge and Kenn Plateau. The fracture zones also form basaltic ridges in the oceanic crust where they are mostly, but not entirely, buried by subsequent sedimentation.

The change in orientation of the Australian continental margin to NNW in southern Queensland is due to a change in direction of spreading and strike-slip movement (70 - 64 Ma) on the southern margin of the Marion Plateau and opening of the Capricorn Trough as a failed rift (Gaina et al., 1998b; Muller et al., 2000). The 150 km-long NNE trending Cato Fracture Zone forms the northern boundary of the oceanic crust in the Tasman Basin (Exon et al., 2006a) at this location. Further north in the Tasman Sea the Recorder Fracture Zone meets the base of slope where it is offset off Noosa (Queensland) and was formed when the Kenn Plateau moved away from southern Queensland (Hill, 1992). A strike-fault at this time (70 Ma) also formed the steep scarp at the base of the continental slope where it widens northwards from Coffs Harbour to Tweed Heads. Similarly the rectilinear pattern of the western margin of the four blocks that make up the Dampier Ridge and the steep scarps that occur at the base of the continental slope in southern NSW were formed by initial break-up along transform (strike-slip) faults.

The basement rocks of the conjugate margins of eastern Australia and Dampier Ridge/LHR form an asymmetric pair in the subsurface profile. Seismic lines show that the extensional basins formed during the rifting stage are now largely confined to the western half of the LHR with only minor rift basins on the slope of eastern Australia (Fig. 3.5; Willcox and Sayers, 2001). The main sedimentary basins in the sub-seafloor of this 'Central Rift Zone' are the Capel and Faust Basins east of Capel and Gifford Seamounts, the Gower Basin east of Lord Howe Rise and the Monawai and Moore Basins at the southern margin of the EMR (Zhu and Symonds, 1994). These basins contain over 4 km of sediment along with igneous intrusions, some of which reach the seabed as seamounts. In contrast the basement under the eastern half of the LHR is relatively flat ('Lord Howe Platform' of Willcox, et al., 2001).

To account for both the differing crustal thickness of the conjugate margins, and hence current water depth, and the lack of major rift basins on the Australian margin Jongsma and Mutter (1978) concluded the final breakup was asymmetric. Their model had the emplacement of oceanic crust occurring along the western boundary of a wide rift system of basins and ridges. Etheridge et al., (1989) developed this model further with a low angle detachment fault creating the Australian margin as the upper plate margin along with the uplift of the eastern Australian Highlands taking place prior to breakup while the Dampier Ridge/Lord Howe Rise formed from the thinner crust of the lower plate margin.

Since their formation the passive margins of the Tasman Basin have been subsiding to their current depths. The rates and timing of subsidence are not well known. Mudstone samples of late Cretaceous Campanian age (84 to 71 Ma) from the NSW continental slope (Heggie et al., 1992) indicate widespread marginal marine conditions and DSDP drilling suggests fully marine conditions on the northern LHR at this time. By this time the basins formed during rifting have mostly filled with sediment and are no longer topographic basins. Today they underlie the Lord Howe Rise and some of the Australian continental slope. However, topographic basins still exist for the Middleton and Lord Howe Basins. The Cretaceous rocks in these basins have not been sampled but are interpreted from seismic evidence to be fluvial sands and gravels and lacustrine and marginal marine muds. These are overlain by post-breakup pelagic limestone and calcareous oozes.

Since seafloor spreading ceased the region has been moving north at ca. 7 cm yr-1 as part of the Australian plate. During this time three hot spots have resulted in north-south volcanic mountain chains. The most westerly affected the east coast of Australia (Wellman and McDougall, 1974, Sutherland, 1998), the central hotspot formed the Tasmantid Seamount Chain (McDougall and Duncan, 1988), and the eastern hotspot produced the Lord Howe Seamounts on the western flank of the Lord Howe Rise (Vogt and Connolly, 1971, Slater and Goodwin, 1973, McDougall et al., 1981). Other post-breakup volcanics have been sampled and imaged by seismic at many locations on the Lord Howe Rise, Dampier Ridge and on the Australian continental slope (Hill, 1992; Heggie et al., 1992; van de Beuque et al., 2003; Glenn et al., 2007).

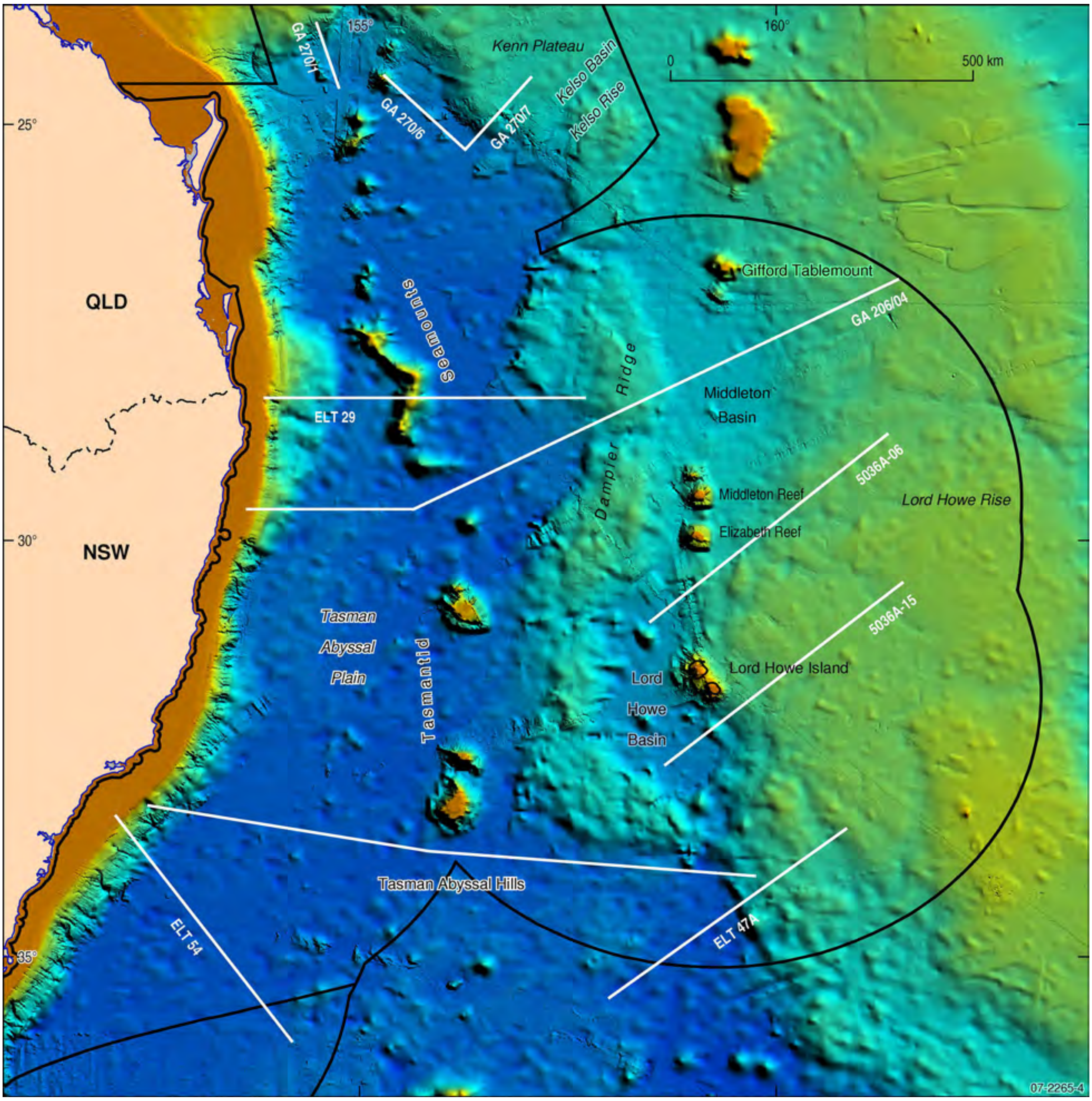


Figure 3.1. False-colour image of the Tasman Sea showing the geomorphology and bathymetry. The main geomorphic features are labelled. Location of features and seismic lines displayed in other figures is marked. Black line is the EMR boundary.

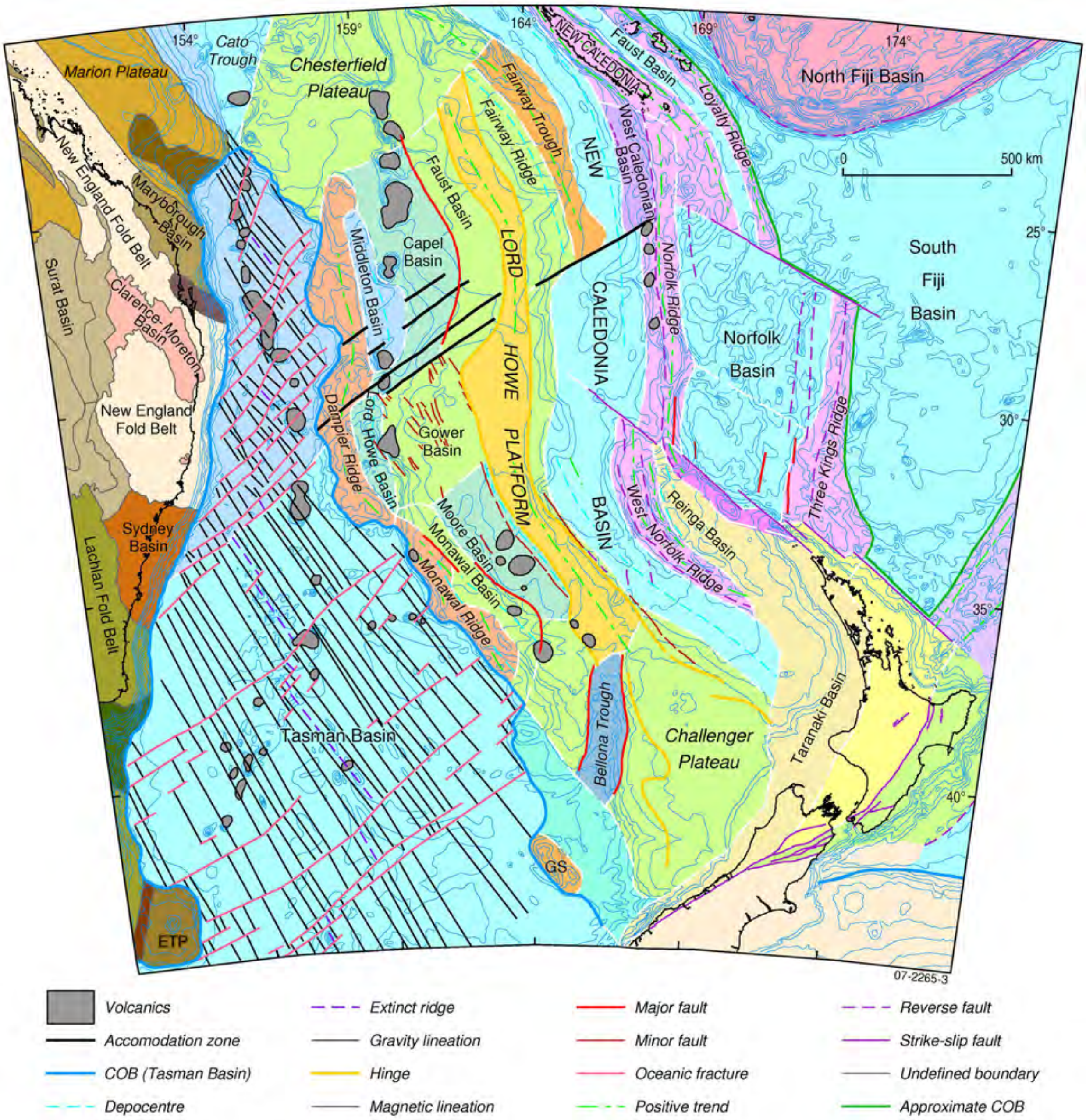


Figure 3.2. Map showing the tectonic features of the Tasman Sea and the geology of the conjugate margins. Stagg *et al.,* (1999a); figure modified by Alcock *et al*., (2006).

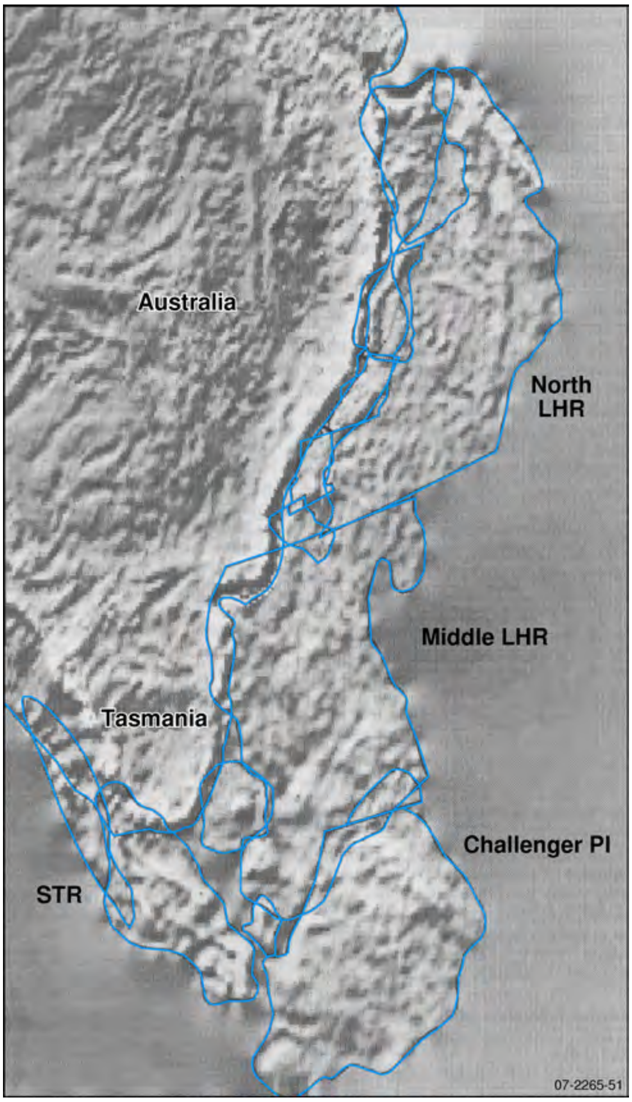


Figure 3.3**.** Mosaic of continental block pre-breakup time (90 Ma). Blue lines are modern boundaries of the continental crust with oceanic crust. Overlap is due to synrift crustal extension. Gaina *et al*., 1998b.

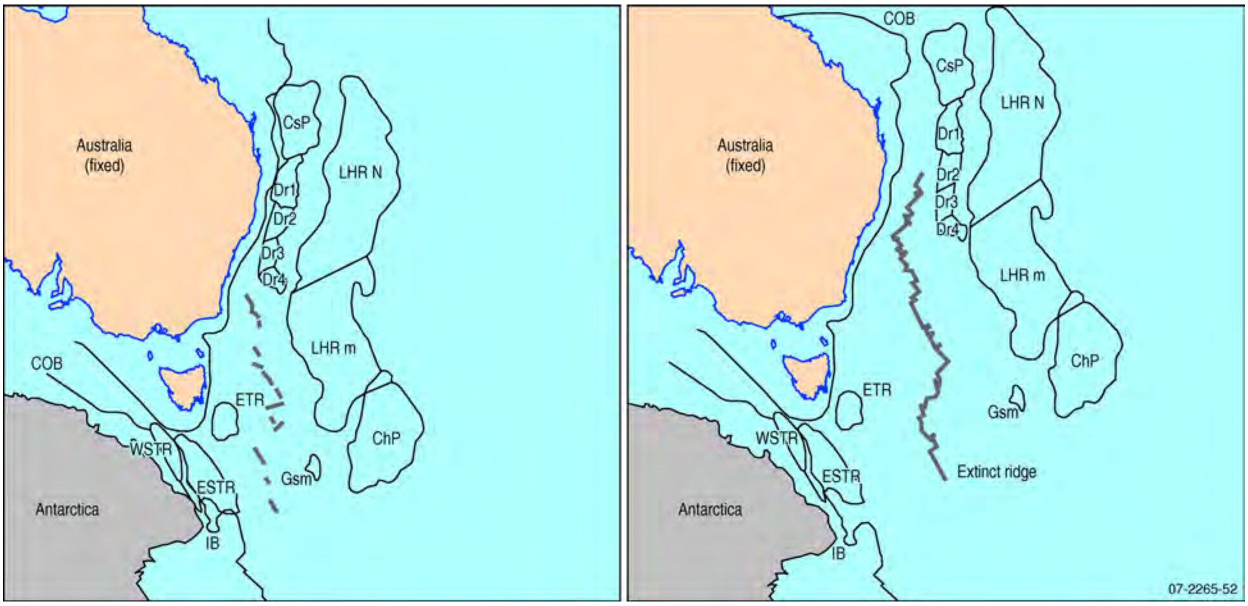


Figure 3.4. Reconstruction of the Tasman Sea opening at 67.7 Ma and at 52 Ma when spreading ceased. Gaina *et al*., 1998b.

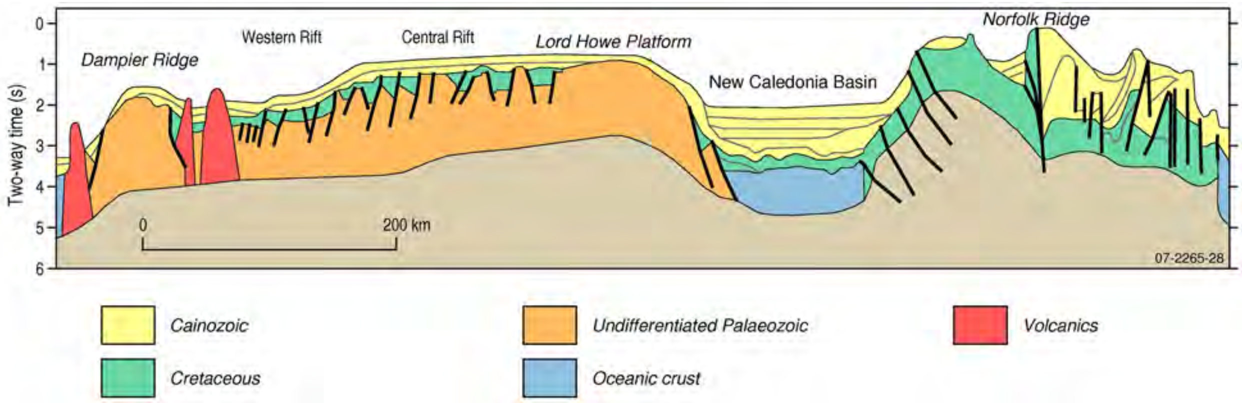


Figure 3.5. Interpreted seismic profile from the Tasman Basin across the Dampier Ridge, Lord Howe Rise, New Caledonia Basin and Norfolk Ridge showing how the tectonic elements of the rifted crust, oceanic crust and volcanics determine the geomorphic features. Note the relatively steep lower slopes, the thick sediment caps on highs and sediment fill in basins. (Alcock et al., 2006).

3.1.1.2. Western Pacific Ocean: Norfolk Island

The shape of the seabed in the EMR around Norfolk Island is controlled by the tectonic fabric that trends north-south as a result of the breakup of the eastern margin of Gondwana and its relationship to the plate boundary along this margin. The EMR extends from the eastern slope of the Lord Howe Rise in the west to the western edge of the Three Kings Ridge in the east. The Norfolk Ridge is the central feature and is part continental crust with relatively thick sedimentary basins, and part island arc volcanics with younger late Cainozoic volcanism adding to the complexity (Eade, 1988; Mortimer et al., 1998).

The Fairway Basin and the New Caledonia Basin lie between the Lord Howe Rise and Norfolk Ridge and were formed in the Early Cretaceous as a result of extension and rifting commencing around 130-120 Ma (Eade, 1988, Muller et al., 2000) and continued as a result of extensive rollback of the Pacific Plate (Crawford et al., 2004). The volcanic basement ridge between the Fairway Basin and the New Caledonia Basin is believed to have formed in the early Cainozoic (Auzende et al., 2000). According to Sdrolias et al. (2002) these basins are underlain by continental crust but others interpret oceanic crust beneath the New Caledonia Basin (Exon et al., 2004c). Continued rifting from 120 and 95 Ma is supported by large amounts of felsic volcanics of this age along the most easterly of these continental fragments, the Norfolk Ridge (Bryan et al., 1997). It must have been at a convergent plate boundary (Crawford et al., 2004). Sdrolias et al. (2001) propose that the Norfolk Ridge was a volcanic arc above a west-dipping subduction zone of the Pacific Plate prior to 100 Ma and that the New Caledonia Basin is a back-arc rift basin with trapped continental crust or Pacific Plate Cretaceous oceanic crust in the Norfolk Basin (a view also proposed by Eade, 1988). They also provided geophysical evidence for a reversal to east-directed subduction from 90 to 45 Ma with the trench lying on the western margin of the Three Kings Ridge but no arc type volcanics of this age have been found. A reversal in subduction at ~45 Ma resulted in the further opening of the Norfolk Basin from 45 to 35 Ma.

Regions of elevated topography in the Norfolk Basin are interpreted as thickened Cretaceous oceanic crust while the topographic basins are underlain by Miocene oceanic crust. Widespread volcanic activity occurred on the plateaus and along their margins in the Miocene before the basins were formed (Herzer et al., 1997; Mortimer et al., 1998, Sdrolias et al., 2004). Mortimer et al. (1998) provide evidence from petrology and isotopic dating of dredge samples to conclude that the Norfolk Basin is a back-arc basin formed by 18-20 Ma. Another view is given by Crawford et al. (2004) who support sea floor spreading east of the Norfolk Ridge from 70 to 55 Ma forming the Norfolk Basin as a back arc basin between the Norfolk Ridge and the Three Kings Ridge, a volcanic arc with a trench to its east. According to these authors subduction had reversed by 50 Ma to become eastward dipping beneath the Three Kings Ridge, and most of the back arc Norfolk Basin was subducted, before subduction ceased at 30 Ma. The plate boundary then jumped eastward and the modern day westward dipping subduction of the Pacific Plate began.

While it is generally agreed that the Norfolk Basin was formed by some kind of back-arc extension there is little agreement on the timing, process and hence nature of the crust. DiCaprio et al. (in press) have synthesized all of the available data, and concluded that the Norfolk Basin is composed of continental fragments along with Cretaceous and Miocene aged oceanic crust. They identify a thrust fault on the Kingston Plateau and metamorphic rocks from the Bates Plateau dated at 38 Ma to indicate compression during the late Eocene. Extension and new seabed formed during the early Miocene, but was limited to the South Norfolk Basin and the Forster Basin.

The Cainozoic volcanic history of the Norfolk Ridge is poorly known because it is covered with a thick sedimentary sequence. Norfolk Island consists of alkaline basalt lavas of Late Pliocene age dated as 3.1-2.3 Ma (Jones and McDougall, 1973). Along the western edge of the Norfolk Ridge is a seamount chain part of which is a large caldera and many small cones (Exon et al., 2004b). A dredge sample of basalt from this chain gave a late Oligocene age of 26.3 Ma (Mortimer et al., 1998). Another chain 100 km to the east occurs as seamounts on oceanic crust. Mortimer et al. (1998) also dated basalt from seamounts in the South Norfolk Basin as being of Miocene age (16.7-21.4 Ma). The West Norfolk Ridge is considered to be an igneous/metamorphic basement block with a generally flat eroded top with varying amounts of sedimentary cover, mostly in half-grabens (Herzer et al., 1997).

3.1.1.3. Coral Sea

The seabed features in the Coral Sea, like the Tasman Sea, are the result of the nature of the breakup of the continental crust along the eastern margin of Gondwana and the location of new oceanic crust. The general pattern of breakup of the northeastern margin of Australia was established by Taylor and Falvey (1977), Mutter (1977), Weissel and Watts (1979) and Shaw (1979). More recently Struckmeyer and Symonds (1997) and Gaina and Muller (1999) have re-evaluated all the geophysical data and show how the breakup and seafloor spreading of northeastern Australia has formed the major tectonic blocks that exist today (Fig. 3.6).

Prior to 62 Ma rifting and extension of continental crust occurred north of the Tasman Sea Basin oceanic crust to form the Townsville and Queensland sedimentary basins which underlie the present-day bathymetric features of the Townsville and Queensland Troughs (Fig. 3.7). The first oceanic crust formed east of the Queensland and Marion Plateaus at Chron 27 (61.2 Ma) when a triple junction formed where the Louisiade Plateau and Mellish Rise continental fragments met the Queensland Plateau (Fig. 3.6). This new seabed propagated northwest, northeast and south forming the Coral Sea/Osprey Embayment, Louisiade Trough and Cato Trough, respectively. Between 57.9 and 55.8 Ma the triple junction migrated south and the transform fault in the Louisiade Trough became extinct and a new strike-slip fault became active between the Mellish Rise and the Kenn Plateau. Spreading became extinct at 52 Ma. An east-west cross section over the Marion Plateau-Cato Trough-Kenn Plateau shows the underlying geology beneath a sediment drape (Fig. 3.8).

Gaina and Muller (1999) identify 190 km of oceanic crust in the Louisiade Trough that formed between 61.2-57.9 Ma, and 150 km of oceanic crust in the Cato Trough and 180 km of oceanic crust in the Osprey Embayment between 57.9 and 52 Ma. They also propose a transform fault along the northwestern part of the Kenn Plateau, which might account for later emplacement of the seamounts that now occur along this trend. The structural trends underlying the complex geomorphology in the Cato Trough area are shown in Figure 3.9. The plateaus and ridges are bounded by major faults, and major volcanic extrusions have utilized these crustal weaknesses.

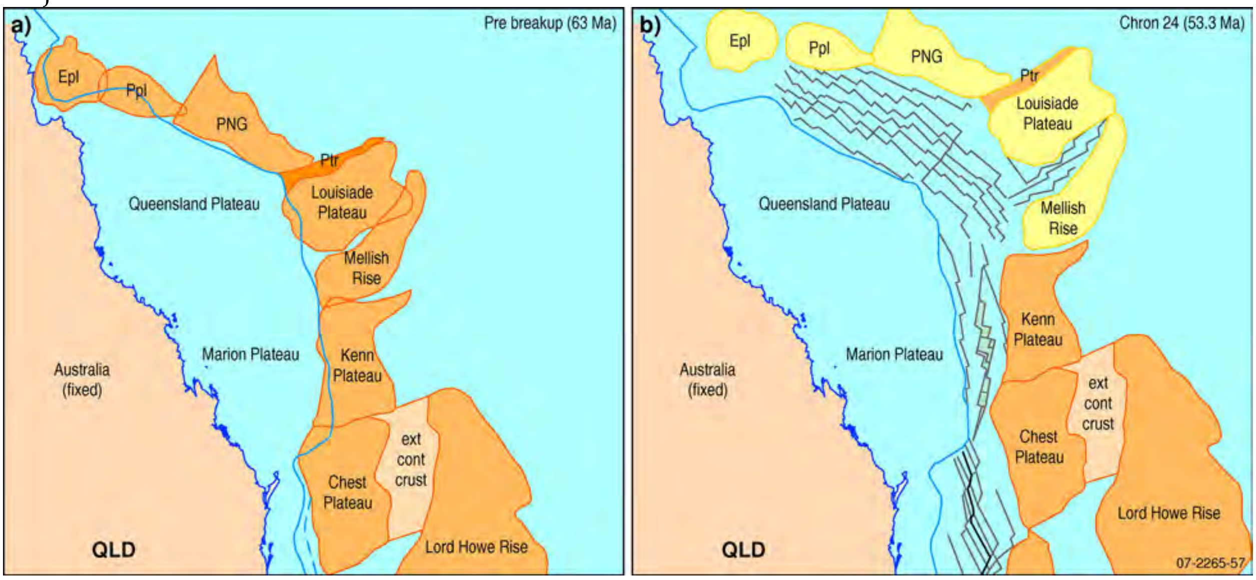


Figure 3.6. Tectonic block reconstructions for the evolution of the plateaus, troughs and basins in the Coral Sea offshore of Queensland. Gaina *et al*., (1999); Exon *et al*., (2005).

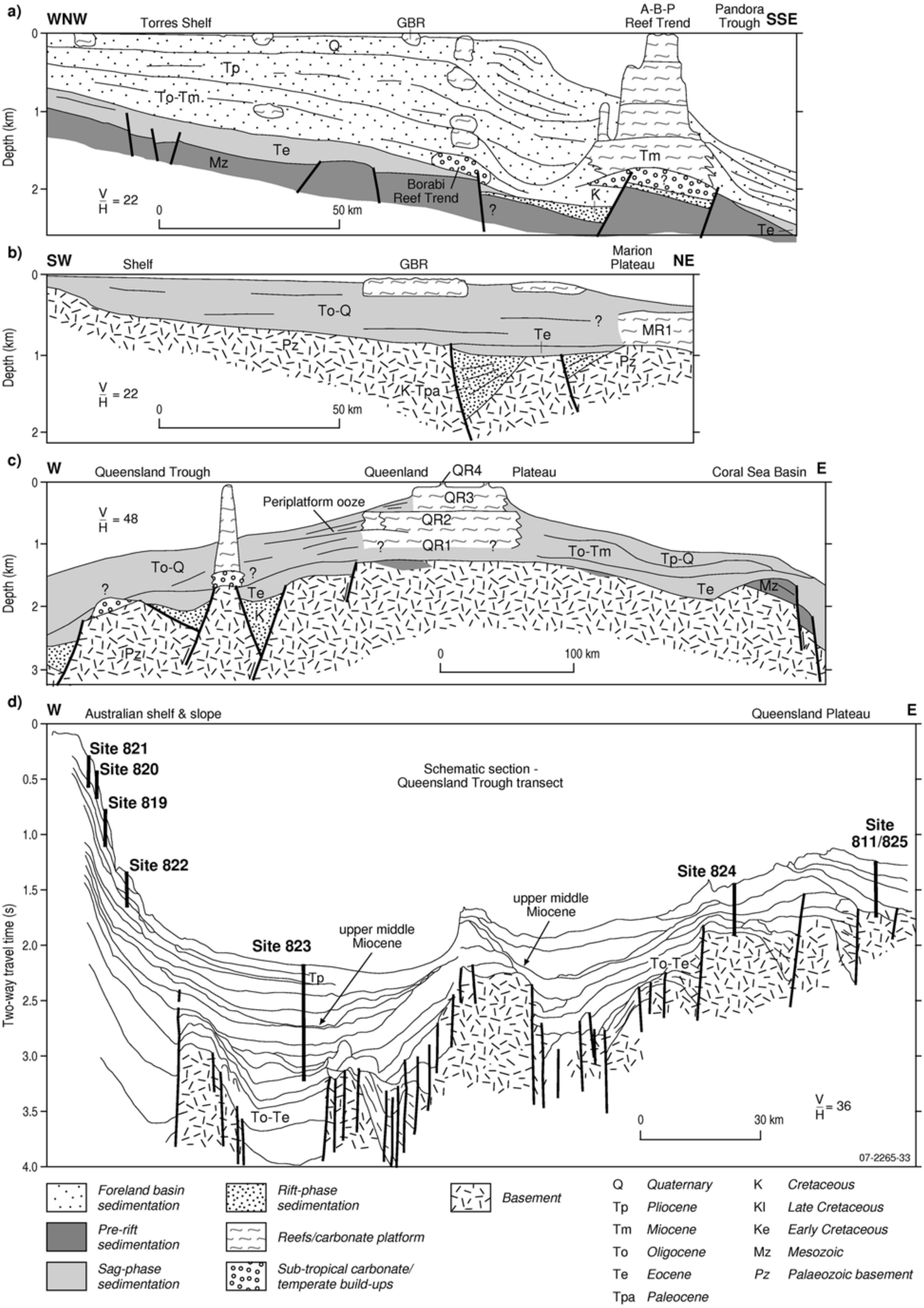


Figure 3.7. Profiles showing generalized basement structure and sedimentary sequences offshore of Queensland. Wellman et al., (1997).

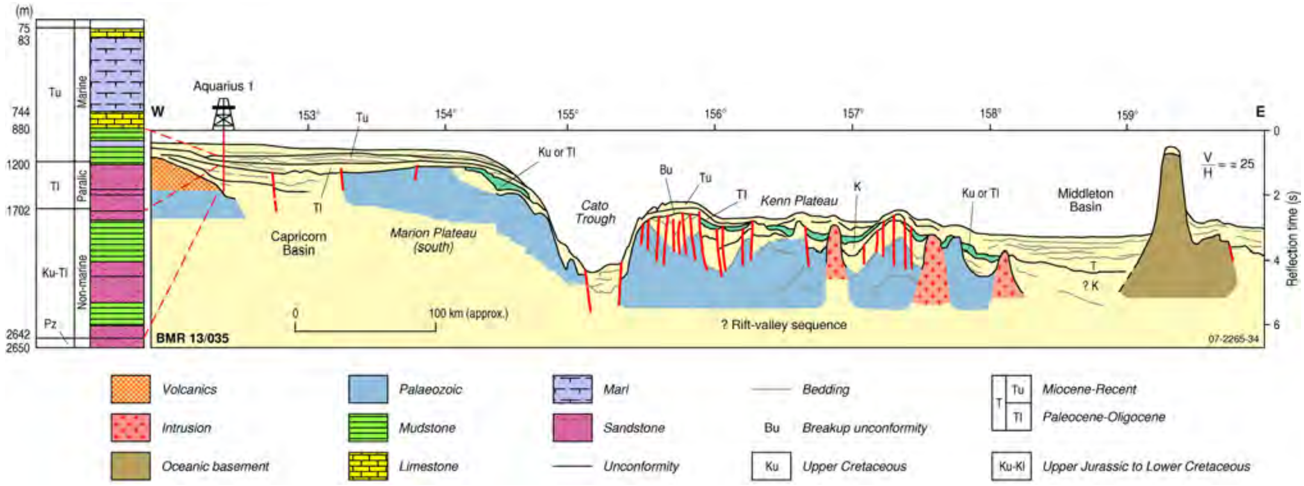


Figure 3.8. Geological and tectonic interpretation of an east-west seismic profile from the Marion Plateau to the Middleton Basin. Line BMR 13/035. Location in Figure 3.44. Willcox, (1981).

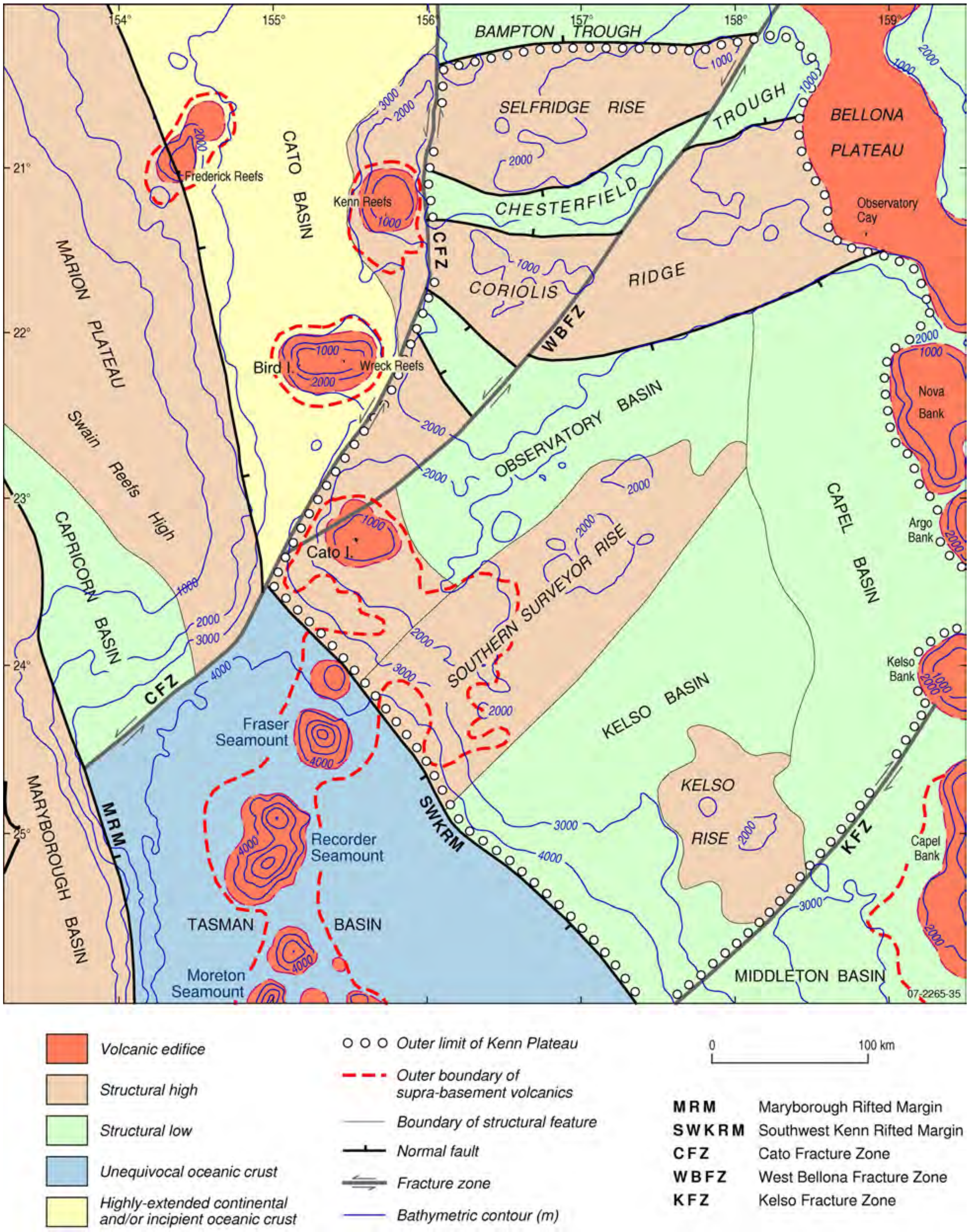


Figure 3.9. Tectonic elements that form the geomorphic features in the Cato Trough – Kenn Plateau area. Exon *et al.,* (2006b).

3.1.2 Oceanography

The surface water circulation in the EMR is dominated by the East Australian Current (EAC) which forms on the Queensland Plateau at -15°S from the South Equatorial Current (SEC) (Pickard et al., 1977; Godfrey et al., 1980, Creswell et al., 1983; Burrage et al., 1996). Part of the SEC forms a clockwise circulation northwards into the Gulf of Papua (Wolanski et al., 1984; Wolanski et al., 1995; Church and Craig, 1998: Keen et al., 2006) while the rest flows over the Queensland Plateau and south over the Queensland Trough seaward of the Great Barrier Reef (GBR). It then flows over the Marion Plateau and along the shelf and upper slope of eastern Australia (Fig. 3.10). It is strongest between 25° and 30°S. The flow is directed off shelf at major headlands such as Fraser Island, Smoky Cape and Sugarloaf Point, and is a significant process in sweeping sediments off the shelf edge (Boland and Hamon, 1970).

At 30°S the EAC is on the outer shelf and upper slope and surface currents of 1-2 ms^{-1} (2-4 knots) have been measured (Godfrey et al., 1980; Church, 1987). South of Sugarloaf Point (32°S) the strength of the current decreases rapidly and the current breaks up into anticyclonic gyres or eddies (Marchesiello and Middleton, 2000). These eddies may impinge on the shelf in the southern part of the EMR and a hydrographic study by Mulhern (1983) infers the influence of EAC eddies down to 2,000 m. Scours and lee-side mounds around bedrock on the upper slope are evidence that the current can modify the sediment down to water depths of 1,000 m (Glenn et al., 2007). The prevailing wind pattern means that upwelling of deeper nutrient-rich water is not a regular phenomenon along this margin (Gibbs et al., 2000).

Part of the EAC is deflected offshore at around 30°S along the Tasman Front (Subtropical Divergence) forming a warm subtropical anticyclonic gyre (Boland and Church, 1981). The Tasman Front forms the interface between the warm waters of the Coral Sea and the cooler waters of the Tasman Sea and it moves north and south with the seasons from 30°S in winter to south of Lord Howe Island and along 34°S in summer (Martinez, 1994a).

A cooler subtropical cyclonic gyre originates in the southern part of the west Tasman Sea and meets subantarctic waters along the subtropical convergence (45°S). Below these water masses at about 1,000 m depth the Antarctic Intermediate Water (AAIW) flows from south to north and beneath it is the Antarctic Bottom Water (AABW) which also flows north in the western part of the Tasman Sea (Tomczak and Godfrey, 1994). Jenkins (1984) mapped erosion on the abyssal floor to infer the clockwise circulation of deep-water currents in the Tasman Basin (Fig. 3.10). Modelling of currents for the GBR suggests a southerly flow of bottom water in the Queensland Trough (Luick et al., 2007).

Storms, longshore currents (littoral drift) and internal waves all affect the nature of the sediment on the seabed. Moderate to high energy southerly swell and much larger storm and cyclone generated north-east and easterly waves affect the NSW and southern Queensland shelf in the EMR (Short and Wright, 1981). Re-suspension, winnowing and transport of sediments by waves occur down to water depth of at least 100 m along this high energy margin (Gordon and Hoffman, 1986; Short and Trenaman, 1992; Roy et al., 1994a; Middleton et al., 1997). On the inner shelf there is northward littoral transport of sediment associated with a northward flowing current (Gordon et al., 1979; Cowell and Nielson, 1984, Huyer et al., 1988). The shelf in the EMR is classified as microtidal so tide generated currents have limited impact except at the entrance to large bays and around islands (Boyd et al., 2008).

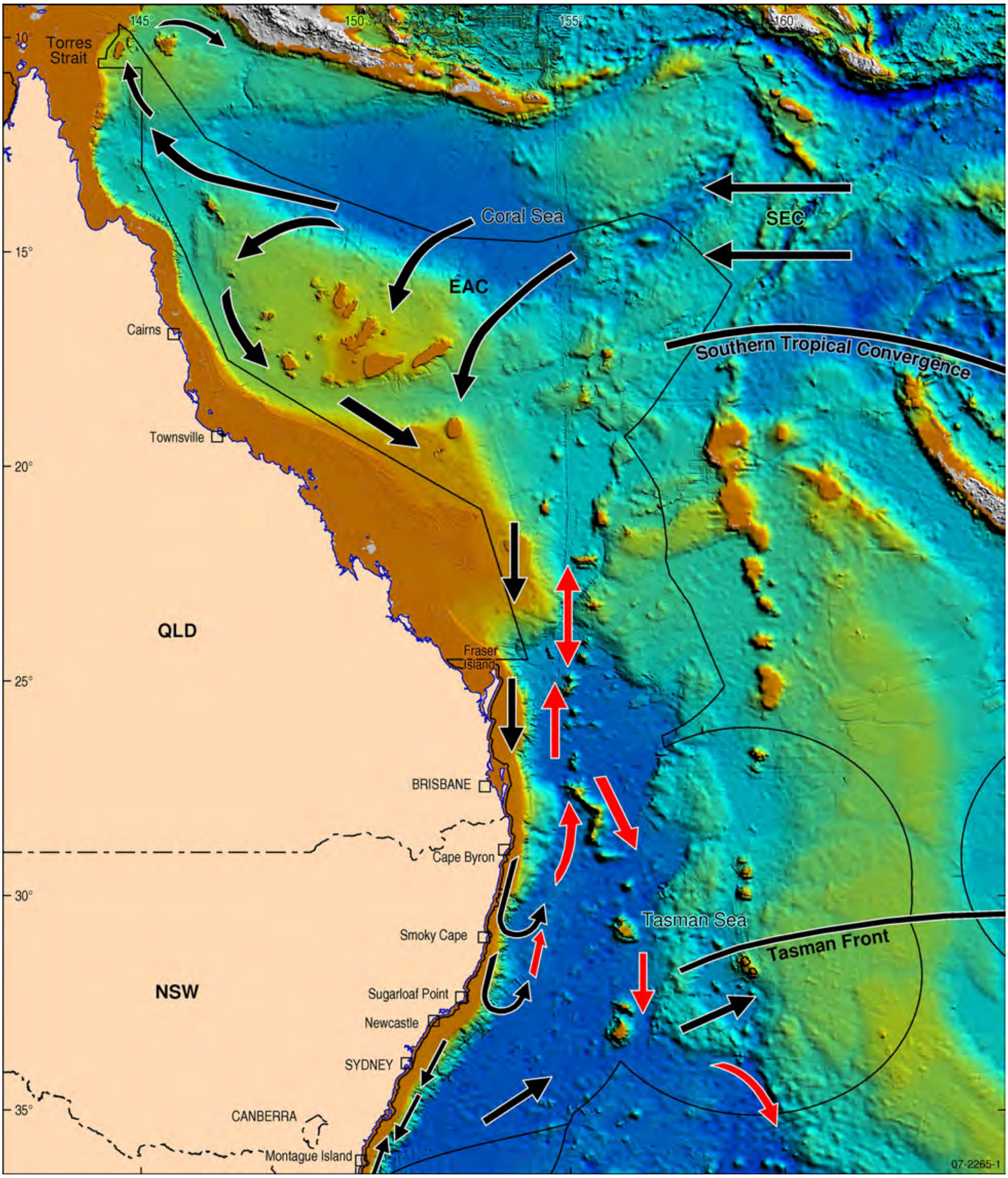


Figure 3.10. Physical oceanography of the EMR, showing the main water masses influencing the region: the East Australian Current (EAC), South Equatorial Current (SEC), West Wind Drift (WWD) and inferred abyssal currents. (CSIRO; Jenkins, 1984; Kawagata, 2001).

3.1.3 Quaternary Evolution

The Quaternary (<2.0 Ma) evolution of the EMR is characterised by a fluctuating climate, oceanography and sea level, which is reflected in seabed sediments and geomorphic features of the region. For half of the past 300,000 years sea level has been 70 to 120 m below its present level (Lea et al., 2002). In half of the last 100,000 years sea level has been at depths between 40 and 80 m. When sea level was lower most of today's shelf was exposed, with a shallow shelf in the Wollongong-Sydney-Newcastle area and very little or no shelf elsewhere along the NSW and Queensland coast. Surprisingly, little terrigenous sediment reached this shelf off Sydney during the last glacial lowstand (Ferland and Roy, 1997). Sediments of this age on the slope off Sydney show an increase in terrigenous mud (Troedson and Davies, 2001). Off Fraser Island the same study found slowly accumulating terrrigenous muds during the lowstand and maximum accumulation rates due to increases in carbonate production during the transgression as the shelf was flooded.

The carbonate platforms, atolls and banks of the Queensland Plateau and Coral Sea have been islands for more than half of the past 300,000 years. The area of these features would not have changed significantly due to their steep sides (>500 m water depth). These features, along with the then-exposed GBR, would have been karstified. In the Cato Trough area, the shallow water connection between the Coral Sea and the Tasman Sea would have been greatly reduced. The Marion Plateau would have been only partly submerged and formed a major promontory along the coastline with a shallow carbonate platform connection to the island formed by Saumarez Reef and Marion Reef at the shelf edge despite sea level rise (Pigram, 1993). The area of the Marion Plateau remaining below sea level was significantly shallower than it is today and swept by strong currents (Liu et al., 1998). East of the Cato Trough the modern day reefs of Mellish, Frederick, Kenn, Wreck and Cato would be slightly enlarged and exposed as islands. East of Kenn Reef, Selfridge Bank would be exposed as an island. South-east of Kenn Reef a bank would be at or near sea level on the Coroilis Ridge.

In the east of the EMR, Norfolk and Philip Islands would become one, much larger island whenever sea level was more than 50 m below present. Areas on the north of Norfolk Ridge currently less than 500 m deep would become banks during low stands. South of Norfolk Island on the boundary of the EMR Wanganella Bank would have been a relatively large island surrounded by a shallow bank for more than half of the past 300,000 years.

The subaerial exposure of what is known today as living reefs probably led to a dramatic reduction in productivity of the shallow water benthic community and a corresponding reduction in carbonate particles from this source. Dunbar et al. (2000) analysed cores from the Queensland Trough and concluded that the glacial lowstand was a period of low terrigenous and carbonate input to the basin. The maximum input of terrigenous muds occurred during the transgression following the lowstand. In the north of the EMR the shoreline was at the shelf edge during the lowstand. This led to a significant input of terrigenous sediment to the slope and basin from the rivers of New Guinea (Francis et al., in press).

The EMR is considered as tectonically stable during the Quaternary with no significant uplift, subsidence or faulting taking place. The larger seamounts continued to subside at a slow rate. Younger seamounts at the southern end of the chains, including Lord Howe Island are currently subsiding faster, but at a rate of centimeters per thousand years. Hydro-isostatic adjustment occurred on the shelf during the Quaternary with rising sea level loading the shelf and falling sea level unloading. Loading during the most recent ocean transgression tilted the shelf seaward by ~1 m.

There is no evidence for volcanic activity during the Quaternary, but it cannot be ruled out as the youngest volcanic flows on Norfolk Island have been dated at approximately 2 Ma and evidence of an undersea eruption 180 km north east of the island was reported in 1981 (Royal Australian Navy

Hydrographic Service, Chart Aus 4602). Other areas with possible Quaternary volcanism occur on the Lord Howe Rise and the Mellish Plateau.

3.2. SOUTHEAST AUSTRALIAN SHELF

3.2.1. Geomorphology

The shelf in the EMR of the Tasman Sea (Montague Island to Fraser Island) is 1, 435 km in length and varies in width from the minimum values of 10 to 20 km off headlands and islands to the maximum of 50 to 75 km off embayments (Fig. 3.11). It is widest between Sydney and Sugarloaf Point (53 km wide off Newcastle) and between Moreton Island and Fraser Island (75 km wide off Noosa) and narrows off Montague Island (17 km), Jervis Bay (16 km), Smoky Cape (12 km), Cape Byron (14 km), Cape Moreton (<10 km) and Sandy Cape (<17 km). The shelf ends at the shelf break, defined as a change in gradient from the shelf with a slope of < 0.50 to > 10 as the seabed deepens on the upper continental slope. The variability in shelf width and depth is illustrated by the 12 profiles superimposed in Figure 3.12.

The shelf is deeper off southern NSW and shallower in northern NSW and southern Queensland. Boyd et al. (2004b) have plotted the shelf break from southern NSW to Fraser Island and show a general trend of shallowing north of Sugarloaf Point (320 30'S) (Fig. 3.13). South of Sugarloaf Point the shelf break is between 145 and 160 m with the deepest shelf break at the margin of the wide shelf between Sydney and Newcastle (172 m) where there is a gradual roll-over on the outer edge of the sediment wedge. Figure 3.14 shows a typical profile of the shelf and upper slope from 40 m to 600 m water depth offshore of Broken Bay where the shelf is wide and the shelf break is not distinct. South of Wollongong the shelf break is relatively sharp. North of Coffs Harbour to Moreton Island the shelf break is between ~80 and 100 m but the break is not as distinct as it is further south and there is a transition to the upper slope. Offshore of Fraser Island there is a distinct terrace and nick-point at a depth of 105 m below a 20 m cliff which forms the shelf break (Fig. 3.15, Marshall et al., 1998). At Breaksea Spit at the north end of Fraser Island the shoreline and inner shelf is effectively at the shelf edge (55 m water depth) as the sand island extends itself across the entire shelf.

The depth and morphology of the shelf is controlled by the underlying basement geology, the prograding depositional sediment wedge, carbonate reef/mound growth (north of Sugarloaf Point), differential subsidence and erosion (Jones et al., 1975; Davies, 1975, 1979). Most of the shelf in the EMR can be divided on the basis of morphology into an inner shelf (<60 m water depth), middle shelf (60-120 m) and an outer shelf (120 m to shelf break). In general, the inner shelf is relatively steep down to 60 m water depth, the middle shelf has a more gently slope seaward and the outer shelf is a flat, near-horizontal plain. Figure 3.14 shows a typical profile off Sydney where the mid-shelf in this area has a gradient of 0.30-0.50 and the outer shelf plain 0.10 (Ferland and Roy, 1997). Davies (1979), Marshall (1979) and Roy and Thom (1981) also present figures that display the variety of shelf profiles from south to north.

At the few places where detailed surveys by side-scan sonar, multibeam or bottom photography have been made significant geomorphic variability is revealed on the shelf (Jones and Kudrass, 1982;

Gordon and Hoffman, 1986; Marshall et al., 1998; Boyd et al., 2004b; Roberts and Boyd, 2004; Payenberg et al., 2006). Using detailed bathymetry of two areas of the NSW shelf Boyd et al. (2004a) recognise 6 shore-parallel zones: shoreface, inner plain, inner mid-slope, outer mid-slope, flat outer plain and hummocky outer plain. They also recognized considerable along-shelf variability in morphology enabling further subdivision into mounds, lobes, tongues, depressions, deltas and bedrock. Mounds, reefs and banks are significant features near the shelf edge off Sugarloaf Point and Fraser Island. The variability of the seafloor at scales of one to hundreds of meters is visible on side-scan sonar images of the inner shelf plain in water depths between 15 and 60 m off northern NSW (Roberts and Boyd, 2004). They distinguished gravel patches, bedrock reefs, sand ripple fields and featureless seafloor.

The inner shelf is generally concave-up and steeper off headlands (1-50) where the bedrock outcrops on the seabed. Shelf sand bodies are often present on the inner to mid-shelf adjacent to prominent bedrock headlands in water depths of 20-120 m (Ferland, 1991). These are linear, shore-parallel features with lengths of 5-35 km and widths of 1-5 km. Terraces and nick-points are common down to 160 m and were formed by erosion during lower sea levels (Jones et al., 1975). Relict drowned beach-barrier systems also occur and underlie the inner shelf plain where it occurs (Browne, 1994, Boyd et al., 2004a).

Topographic depressions on the shelf are rare. Boyd et al., (2004a) document a series of linear depressions oriented along the mid-shelf 2.5-5 km wide over distances of 20 km and relief of up to 10 m. They interpret them as erosional and occurring offshore of the submarine extension of headlands, such as Sugarloaf Point, where the East Australian Current is accelerated due to the constriction. Evidence for strong currents on the mid-shelf also comes from side-scan sonar and seabed photographs that show dunes and ripples in sand in water depths of 50-70 m off northern NSW (Jones and Kudras, 1982).

In the south of the EMR the outer shelf plain is the main geomorphic feature and it is particularly extensive from Montague Island (360 10°S) north to Sugarloaf Point (32° 30′S). South of Montague Island the shelf is oriented N-S, is narrow (18 km), and has a relatively smooth and steady gradient from the shoreline to a distinct shelf break at 142 m. At Montague Island the orientation of the shelf break changes to NNE. On the landward side of Montague Island currents have scoured a channel to a depth of 46 m, some 20 m deeper than the surrounding seabed. East of the island the seafloor is concave-up with thin sediment cover and rock outcrop with seismic profiles indicating erosion by currents (Roy and Thom, 1991). Sand waves on the outer shelf (water depth 70-100 m) with amplitude of 4-9 m and wavelengths of 300-500 m were also reported by Roy and Hudson (1987). The eroded sand has been redeposited to form a convex-up shelf sand body extending 30 km south from the island in 50-110 m water depth (Ferland, 1991). The 115 and 120 m isobaths on the outer shelf immediately north of Montague Island define a 10 km long broad (5 km) ridge oriented south and extending towards the shelf edge. This may be a sand lobe deposited by the southerly flowing current.

Between Montague Island and Jervis Bay the outer shelf plain is 15-18 km wide with three broad basins 15 x 5 km, ~5 m deep and oriented along shelf. At Jervis Bay (35°S) the shelf narrows and there are broad N-S ridges on the outer shelf with 5 m of relief defined by the 140 m isobath. The heads of three canyons between 35° 22′S and 35° 07′S have incised the shelf break by 1-2 km. North of Jervis Bay a ridge with up to 20 m of relief, known as Sir John Young Banks, extends 12 km NE across the inner and mid-shelf to 90 m water depth where it becomes buried by sediments. This has been interpreted as a basement high by Phipps (1966, 1967) and Davies (1975) that is a continuation of a fault at the coastline. North of this ridge the outer shelf plain widens and the orientation of the shelf break becomes ENE.

Between Jervis Bay and Sugarloaf Point Phipps (1966) and Davies (1979) define terrace nick points at 40-60 m, 80-100 m and 120-140 m water depth and suggest they were cut by erosion during still-stands of sea level during the Pleistocene. Offshore of The Entrance (33° 24'S) there is a 10 km long N-S ridge on the mid-shelf with up to 25 m of relief that is probably rock outcrop. The relatively steep inner shelf adjacent to headlands along this section of coastline is associated with elongate, shore-parallel sand bodies. These have been mapped south of Port Jackson (Field and Roy, 1984; Ferland, 1991; Albani and Rickwood, 2000). These sand bodies generally lie in water depths of 40-80 m, 1-5 km offshore and are 10-30 m thick. They are underlain by bedrock and often terminate abruptly against bedrock outcrop. They form a convex-up seabed with the seaward face the steepest at 30-70.

At Sugarloaf Point the shelf again narrows and there are N-S ridges of sand on the outer shelf along with mounds near the shelf break described in detail by Boyd et al. (2004a). The mounds are 2-10 km long and 0.6-3 km wide with local relief of 5 to 15 m. North of Sugarloaf Point the shelf narrows and trends more northerly. At 31° 40'S a broad depression (canyon head?) has incised 5 km into the shelf edge to the 112 m isobath. North of this the shelf break is less distinct as it merges seaward with a gentle upper slope. This continues to Coffs Harbour (30° 20'S) where the upper slope again becomes steeper. From 30° 20'S to 29° 50'S there are numerous small islands that rise steeply from water depths of 40 m on the inner shelf (<60 m) which itself extends halfway across the shelf, 12 km from shore. These islands are part of structural trends in the basement rock. North of here the shelf widens significantly to ~30 km which is maintained until Stradbroke Island. In this area Roberts and Boyd (2004) identify an inner shelf plain (slope: 0.16-0.310) extending from the base of the shoreface to 60 m water depth (8-12 km), a mid-shelf slope (0.31-0.630), and an outer shelf plain with a width of 6-12 km and a slope of 0.16-0.470. The only feature on this flat shelf seaward of the 40 m isobath is Windarra Bank (28° 28'S, south of Tweed Heads), a small pinnacle of bedrock that rises steeply from depths of 50-60 m to 31 m.

The northern ends of both Stadbroke and Moreton Islands extend eastward across the shelf, reducing its width to less than 10 km with shoals extending to the shelf edge offshore of Cape Moreton. North of Moreton Island the shoreline steps westward 30 km and the shelf widens. Here, there are several banks of notable size on the outer shelf with a broad (several kilometers wide), shallow (4-5 m deep), linear depression for 75 km along shelf behind them. Barwon Bank (260 30'S) extends for 15 km along the shelf edge offshore of Noosa. It is <2 km wide and has relief of 10-15 m and a minimum depth of about 22 m. North of Barwon Bank the shelf edge and slope step eastward 20 km to create the widest shelf on this margin (75 km).

Offshore of Fraser Island the shelf edge starts to trend NNW and there are significant banks. North Gardner Bank and Gardner Bank extend some 30 km parallel to the 85 m isobath (Marshall et al., 1998). These banks rise from depths of 50-60 m on the outer shelf to within 22-30 m of the sea surface (Marshall, 1977; Harris et al., 1991). They have generally rugged topography and are cemented hardgrounds believed to have developed on drowned reefs and aeolian sand ridges (Marshall, 1977; Jones and Kudras, 1982; Marshall et al., 1998). These carbonate platforms form the shelf edge and because they are lithified create a particularly steep (140) steps on the upper slope (Payenberg et al., 2006) (Fig. 3.15).

Breaksea Spit extends for 30 km north from Fraser Island as a submerged sand ridge to reach the shelf edge which is only 20 m deep at this location (Boyd et al., 2008). Immediately north of Breaksea Spit there is an incised, sinuous valley on the shelf, up to 50 m deep, 600 m wide and at least 5 km long, cut into a lithified carbonate platform (Payenberg et al., 2006). This valley was the course of the Mary and Burnett Rivers during the last sea level lowstand. Multibeam mapping of the seabed in this area by Boyd et al. (2004b) has revealed large subaqueous dunes with amplitudes of 1-2 m and wavelengths of 100 m that are migrating from the south. These dunes cover the carbonate platform underlying the shelf in this area and sand transported by the dunes has partly filled the shelf valley. Scours in the valley indicate erosion by tidal currents (Payenberg et al., 2006). Between Breaksea Spit and the shelf valley is Stingray Shoal, a roughly circular carbonate reef approximately 1 km across at the top. It rises steeply from the sand dunes which surround its base in 35-40 m of water to come within 17 m of the surface (Boyd et al., 2004b).

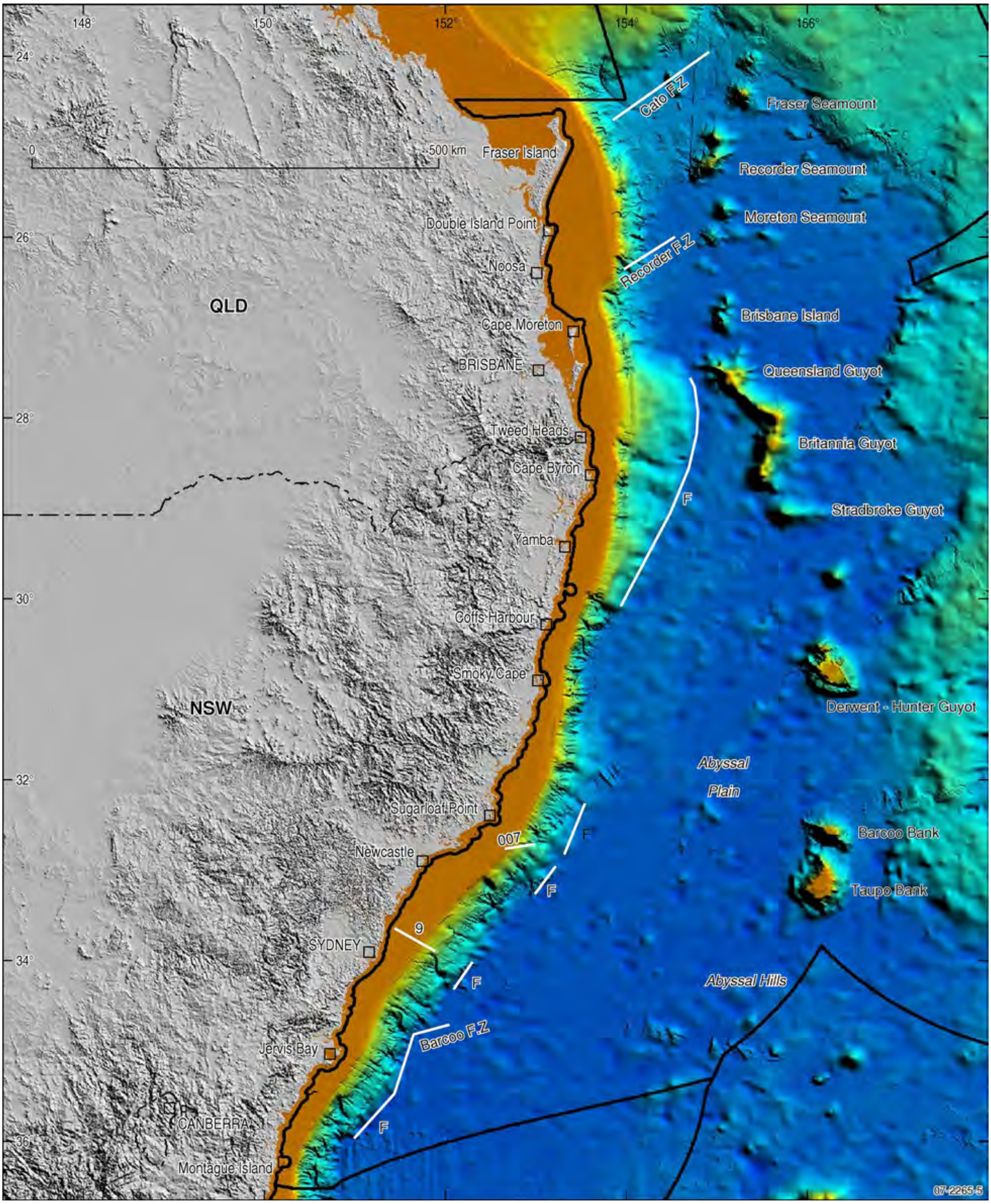


Figure 3.11. False colour map showing the geomorphology and bathymetry of the east Australian margin. Note the varying width of the continental shelf and continental slope and the relationship between the location of major fracture zones in the oceanic crust and shape of the continental slope. The location of seismic profiles 007 and 9 are shown. F = major fault scarps.

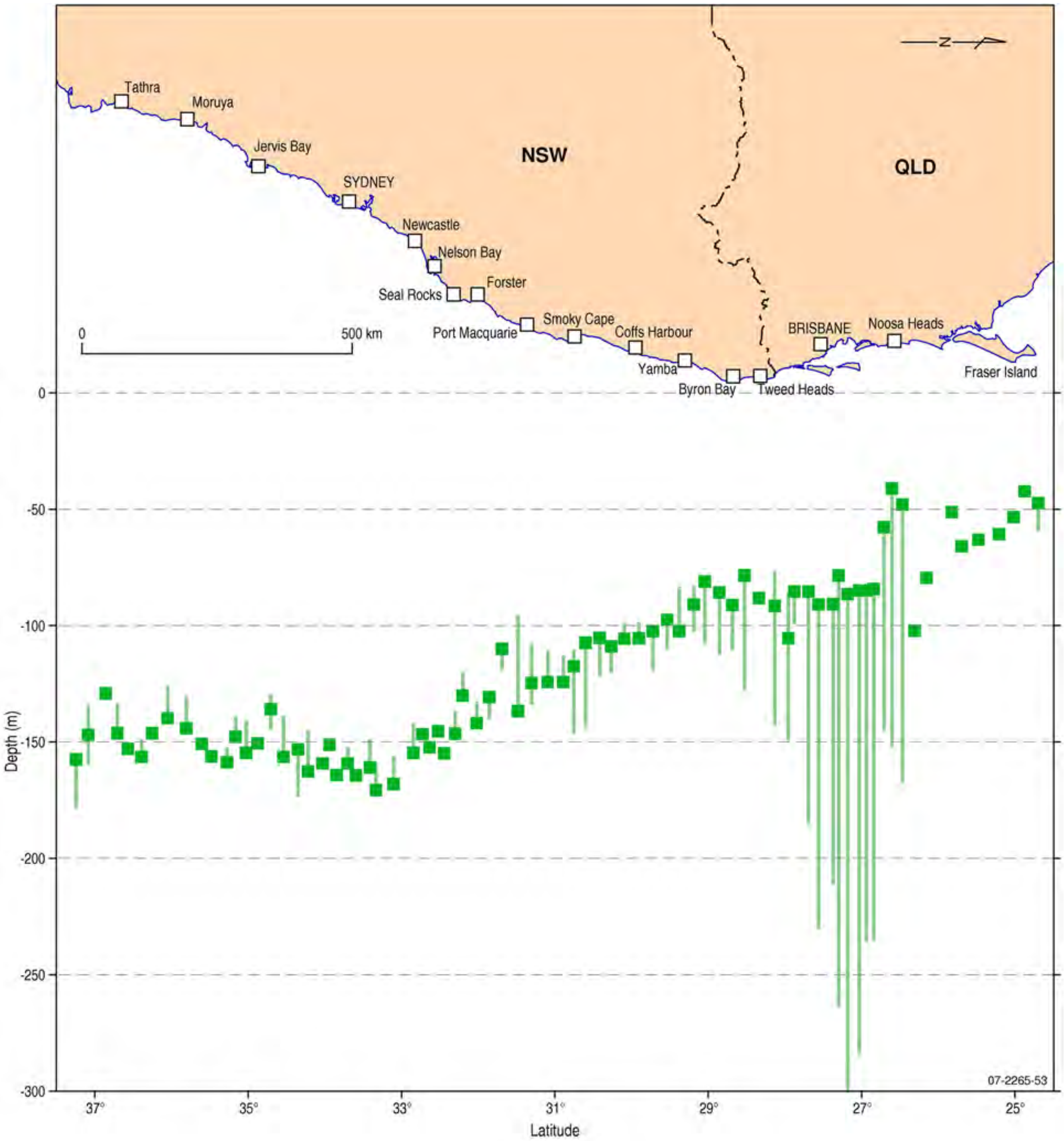


Figure 3.12. Plot of the depth of the shelf break on the eastern Australian margin. Green bars indicate a transitional zone from shelf to slope. (Boyd et al., 2004b).

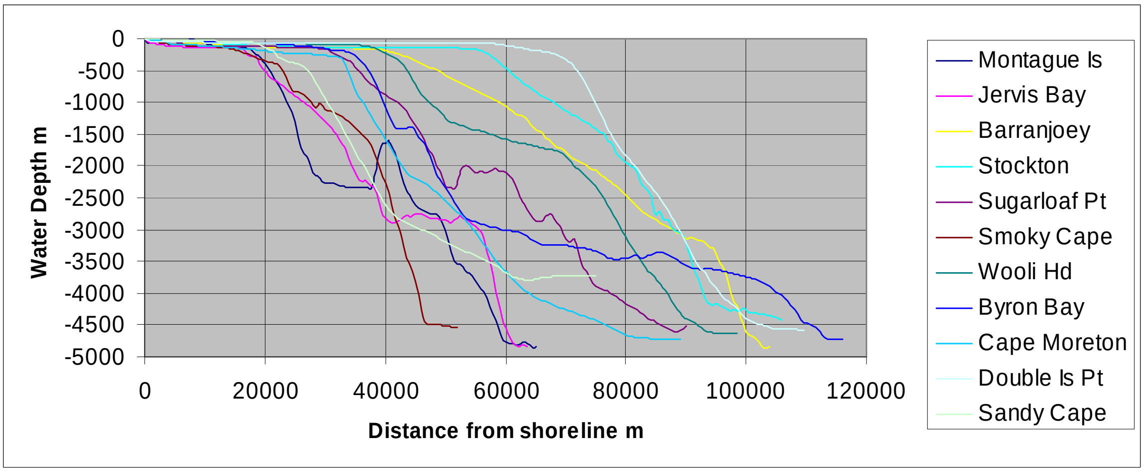


Figure 3.13. Profiles of the continental shelf in the EMR normal to the shelf break from the shoreline to the shelf break. It shows the variety of depths and widths of this shelf. Approximately equally spaced from Montague Island to Fraser Island.

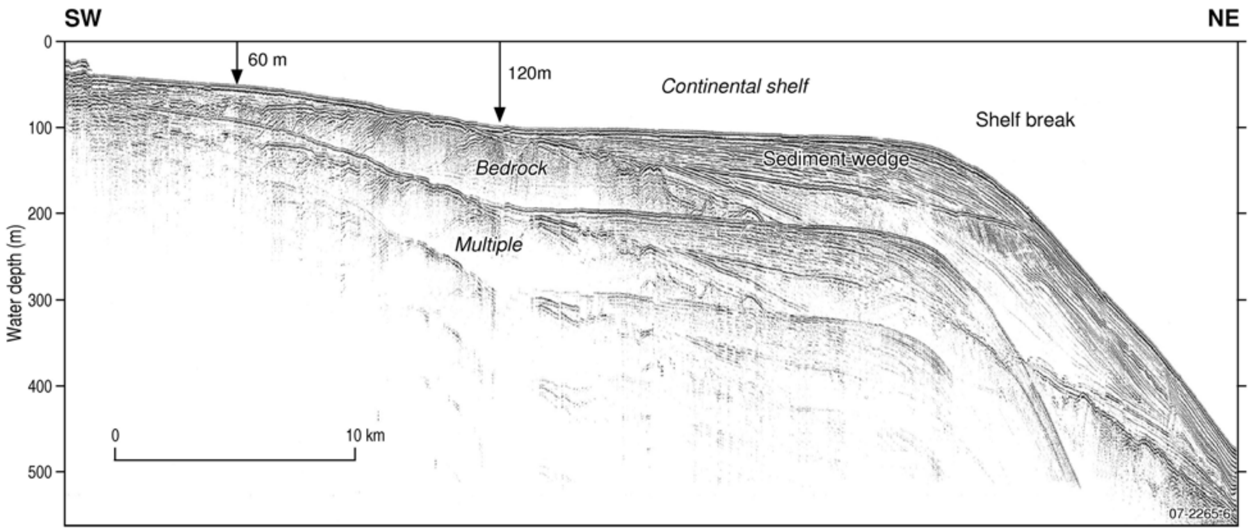


Figure 3.14. Seismic profile of the continental shelf and upper slope offshore of Barranjoey Head, Broken Bay, NSW, showing the relatively steep inner to mid shelf with rock outcrop and the outer shelf plain forming the top of the prograding sediment wedge. The location of the 60 m and 120 m isobaths are marked to show the location of the midshelf mud deposit. The shelf-break is gradual along this part of the coast at ~150 m. Location on Figure 3.11. Line 9A, (Heggie et al., 1992).

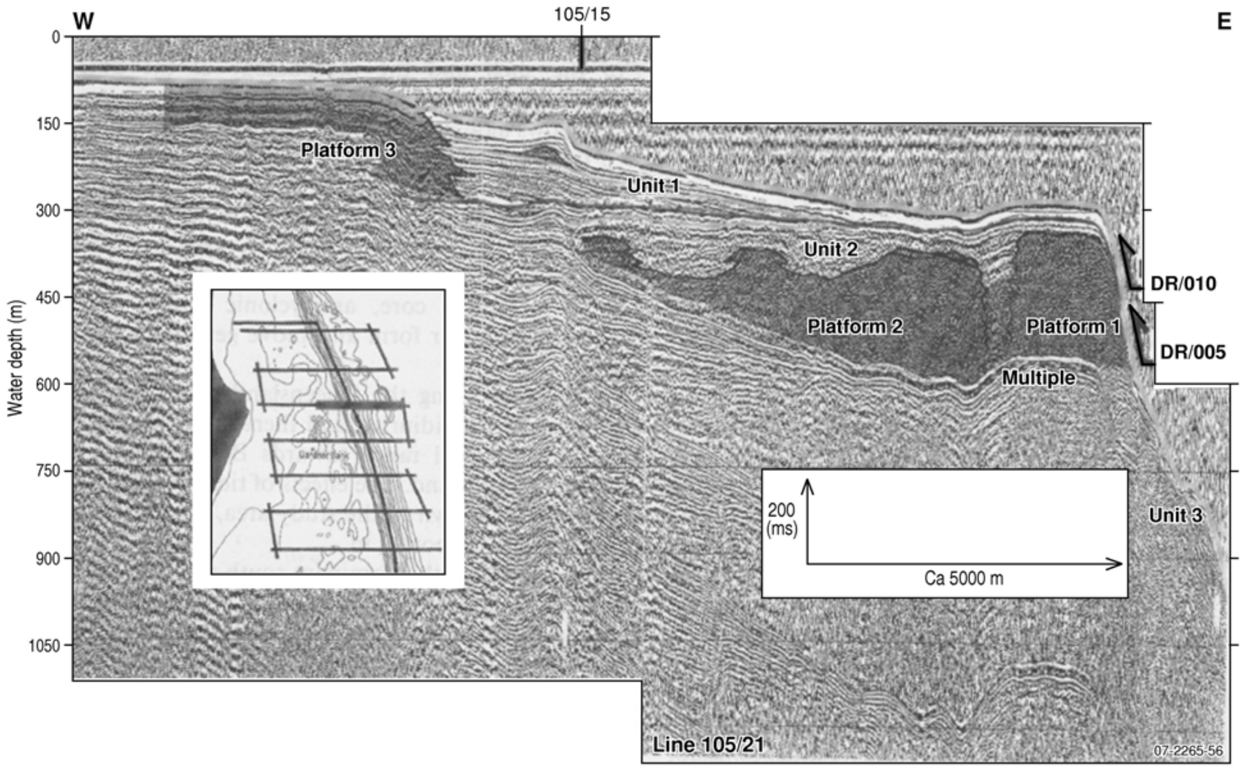


Figure 3.15. Seismic profile showing the carbonate platforms and sedimentary units on the outer shelf and upper slope offshore of Fraser Island (Marshall et al., 1998).

3.2.2. Surface Sediments and Rocks

There is a relatively simple first-order division of surface sediments along the east Australian continental shelf with terrigenous quartzose sediments on the inner shelf and carbonate dominated-sediments on the outer shelf (Fig. 3.16; Davies, 1979: Marshall, 1980; Lane and Heggie, 1993). This boundary is generally at about 100-120 m water depth for the NSW shelf and shallower (~50 m isobath) on the southern Queensland shelf (Marshall et al., 1998). Because of the high wave energy and EAC along this shelf there is little mud accumulation. The grain size trend is medium to coarse sand on the inner shelf, fine sand (with mud in limited areas) on the mid shelf and coarse sand and gravel on the outer shelf (Figs. 3.17 and 3.18). Overall this shelf is one that has had a low sediment supply throughout the late Cainozoic. Most of the sediment deposited along this margin has accumulated over the shelf edge on the upper slope, forming a wedge of sediment that has prograded with time. This wedge of sediment is usually at its maximum thickness below the shelf break in the Sydney-Newcastle area where it is up to 700 m thick (Rule et al., 2007). Thus the shelf break is a depositional feature on the top of this sediment wedge. A consequence of low sediment supply is the occurrence of bedrock outcrop as patches at all depths on the shelf. Relict features such as terraces and shorelines from lower sea levels also occur on the present-day seabed (von Stackelberg, 1982a). Details about the cores and bottom photographs from the shelf are given in Appendix C Tables 3.2 and 3.3.

Bedrock outcrop is common on the inner shelf along with concave-up sand bodies. Rock outcrop has been mapped out to the 50 m water depth 6 km offshore just south of Yamba and for a similar distance offshore off Sydney to the limit of the mapping in 80 m of water (Albani et al., 1988; Gordon and Hoffman, 1989). The shore-parallel sand bodies are composed of well-sorted, fine- to medium-grained quartzose sand with <10% biogenic carbonate and are either relict shoreline sand ridges or regressive deposits formed as sea level rose to its present level (Ferland, 1991; Roy, 1998). The 'inner plain' (20-75 m) of Boyd et al., (2004a) is more common north of Sugarloaf Point and is interpreted as the top of drowned beach ridges that now fill the shelf valleys and form a thin veneer of sand on bedrock (Fig. 3.19).

South of Sugarloaf Point the NSW shelf sediments can be divided into inner shelf, mid-shelf and outer shelf zones based on texture (Shirley, 1964; Davies, 1979). The inner shelf extends down to depths of ~60 m is mantled in medium- to coarse-grained, orange-yellowish quartzose sand with a carbonate content of less than 10% (Gordon and Hoffman, 1989). From Jervis Bay to Sugarloaf Point on the mid-shelf between 60 and 120 m there are muddy sands and sandy muds that form a discontinuous strip of carbonate-rich, olive-green sediment containing gravel sized calcareous shells and worm tubes. This mud-rich belt is 10-20 km wide of muddy sands with mud content between 10 and 30% (Matthai and Birch, 2000a and b). Highest mud content in this zone occur offshore of the major rivers, namely: Shoalhaven, Hawkesbury and Hunter, and the lowest mud contents occur off Sydney (Boyd et al., 2004b). In general, the highest mud contents correspond to where the shelf is widest. Birch and Davey (1995) and Matthai and Birch (2000b) found that the concentrations of heavy metals in the mud fraction off the major urban areas of Wollongong, Sydney and Newcastle were above the regional background values due to dumping of harbour sediments and sewage outfalls.

South of Jervis Bay the mid-shelf sediments continue to be finer grained than those of the inner and outer shelf only with less mud than those sediments further north. North of Sugarloaf Point mud is lacking on the shelf. Only offshore of the Manning and Clarence Rivers is there a mid-shelf zone of fine sand and slightly muddy fine sand, respectively (Marshall, 1980; Boyd et al., 2004a; Roberts and Boyd, 2004).

South of Sugarloaf Point the outer shelf extends from 120 m to the shelf edge and is mantled in coarse-grained, brownish-grey, calcareous, gravelly-sands. The carbonate content is 50-80% (Shirley, 1964; Roy and Thom, 1981; Ferland and Roy, 1997). The outer shelf is particularly coarse grained off the Shoalhaven River and south of Montague Island (Fig. 3.17; Davies, 1979). The biogenic components of the outer shelf sediments are a mix of modern and relict grains consisting of molluscan skeletal fragments, bryozoa, foraminifera and calcareous worm tubes. Pelagic foraminifers become the dominant species towards the shelf break. The non-carbonate components are authigenic glauconite, quartz and cemented ferruginous pebbles. Some of the shells are corroded and colonized by boring sponges, worm tubes and other encrusting organisms and are clearly relict, deposited during a previous low sea level (Roy and Thom, 1981). Beach-rock from 128 m water depth off Sydney has been dated as 15,925 BCE. (Jones et al., 1975). Offshore of Newcastle there are mounds on the outer shelf described by Boyd et al. (2004a) as cemented carbonate consisting of bivalves, bryozoans and other shelly debris. These mounds are interpreted as temperate carbonate reefs formed at the shelf edge during lower sea levels.

North of Sugarloaf Point the texture of the sediment on the outer shelf continues as a coarse sand with some gravel and the carbonate content of the sediment remains high (>60%). Significant amounts of calcareous red algae occur on the outer-shelf north of 30°S (Marshall and Davies, 1978). The carbonate content is reduced in some areas due to the presence of siliceous sponge spicules and glauconite formation. The outer shelf offshore of the Clarence River has less carbonate (Boyd et al., 2008). Offshore of northern NSW carbonate sands and hardgrounds occur on the outer shelf seaward of the 75 m isobath (Roberts and Boyd, 2004). The hardgrounds consist of carbonate sands cemented by calcite, encrusting bryozoans, and by calcareous algae. Seismic profiles in this area indicate mounds at the shelf edge which are interpreted by Roberts and Boyd (2004) as temperate-water reefs.

Off southern Queensland quartz sand continues to form a narrow inner shelf belt with carbonate dominating the mid- and outer-shelf (Fig. 3.20). Carbonate forms >90% of the sediment on the modern biohermal accumulations known as Gardner and Barwon Banks where red algae is a common component (Marshall and Davies, 1978; Marshall, 1980; Tsuji et al., 1997; Marshall et al., 1998; Davies and Peerdeman, 1998). The coralline algae are encrusting and have bound other skeletal fragments to form banks and hardgrounds (Marshall, 1980). The mid- to outer- shelf sediments offshore from Fraser Island are sands, gravels and crusts consisting of coralline algae, hermatypic corals, large benthic foraminifers, bryozoans, Halimeda, molluscs and algal rhodoliths up to pebble-sized (Marshall et al., 1998). Seafloor photographs in Marshall et al., (1998) show living corals on Gardner Bank and rhodolith gravel blanketing the sediment surface on the outer shelf off Fraser Island. The East Australian Current sweeps the seabed here and the rhodoliths are moved by southerly flowing bottom currents with speeds of up to 1.34 ms^{-1} in water depths of 40-140 m (Harris et al., 1996). A limestone platform underlies the present day sediments on Gardner Bank and outcrops where it forms the shelf edge. Barwon Bank hosts few corals and lacks Halimeda but has abundant bryozoans, barnacles, benthic foraminifers, serpulid encrustations and pelecypods to form a temperate reef (Tsuji et al., 1997).

North of Fraser Island, in Hervey Bay, the shelf is underlain by a lithified (limestone) tropical carbonate platform covered by a thin veneer of migrating quartz sand dunes (Marshall, 1977, Boyd et al., 2004b, Payenberg et al., 2006). A multibeam image of Stingray Shoal shows it to be roughly circular, approximately 1000 m across at the top (17 m water depth) and rising steeply from the sand dunes which surround its base in 35-40 m of water (Boyd et al., 2004b). It is composed of coral and coralline algae, some living. Shoreward (west) of these outer-shelf dunes are poorly-sorted fine to coarse sands with some gravel and shell particles, which form the seabed in Hervey Bay (20 m water depth). Their composition of feldspar and rock fragments show them to from the local rivers and reworked Pleistocene shore ridges (Boyd et al., 2004b).

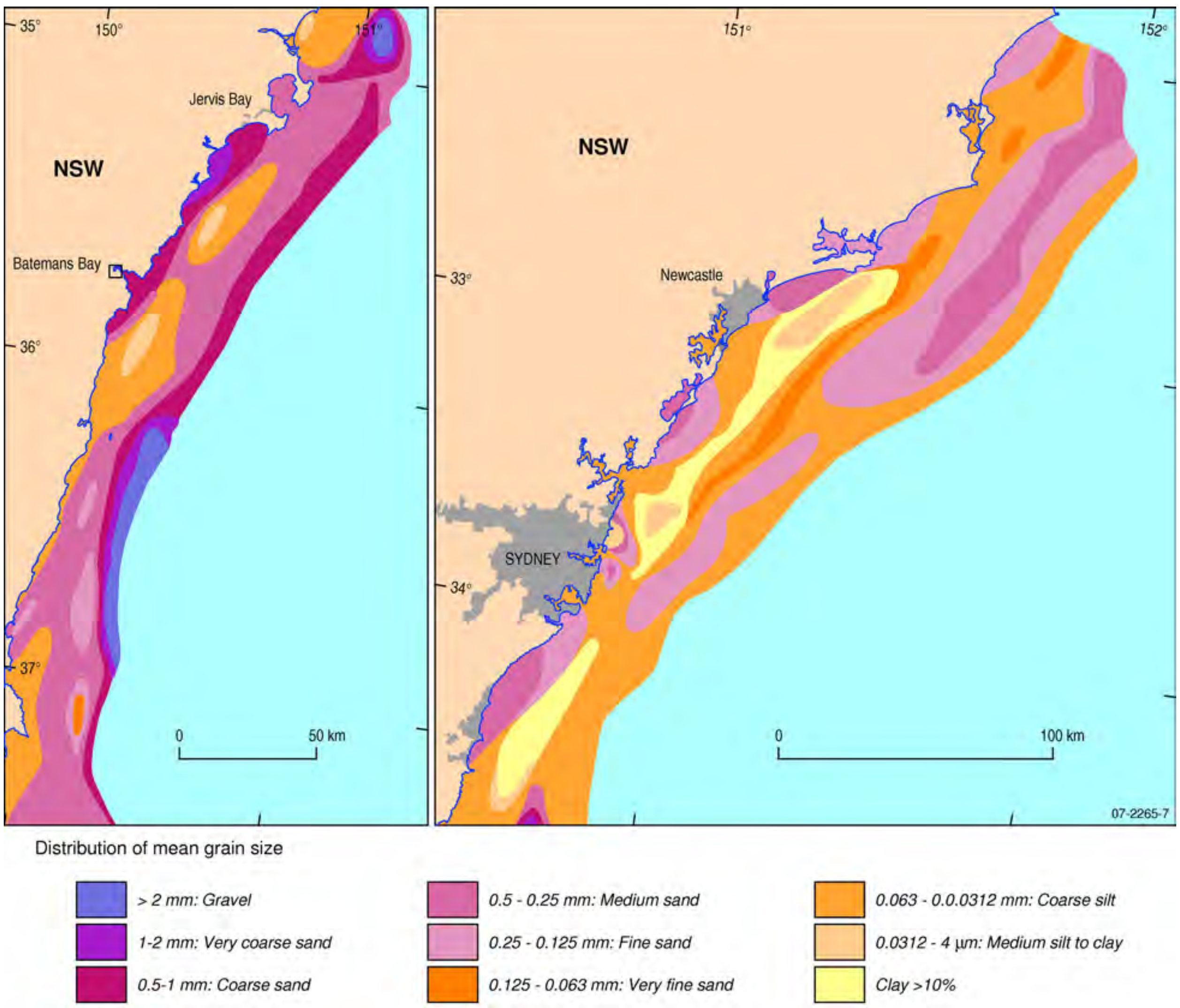


Figure 3.16. Maps showing the distribution of grain size of surface sediment on the continental shelf and upper slope south of 32^0S. (Davies, 1979).

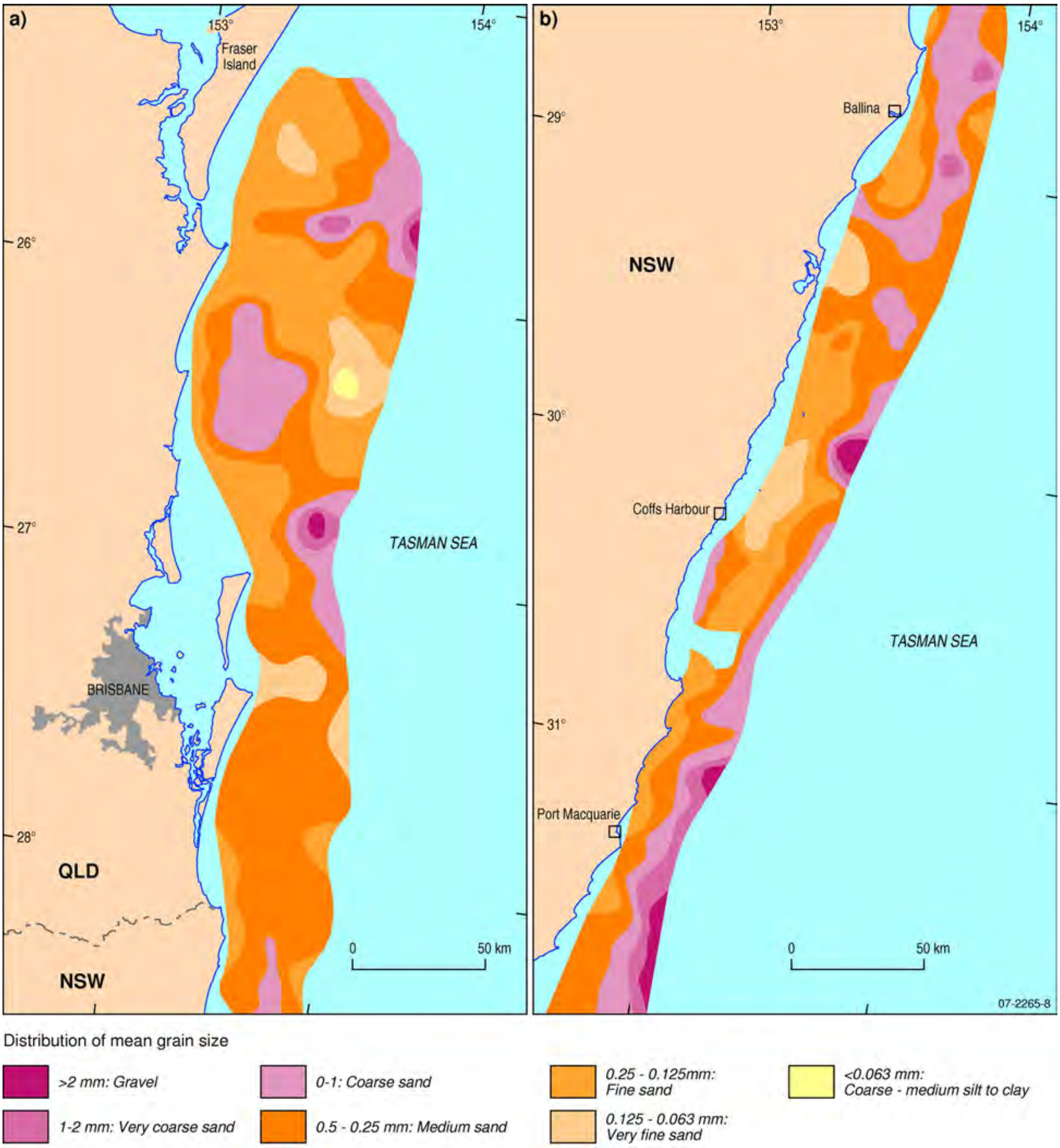


Figure 3.17. Maps showing the distribution of grain size of surface sediments on the continental shelf and upper slope from Fraser Island to 32^0S. (Marshall, 1980).

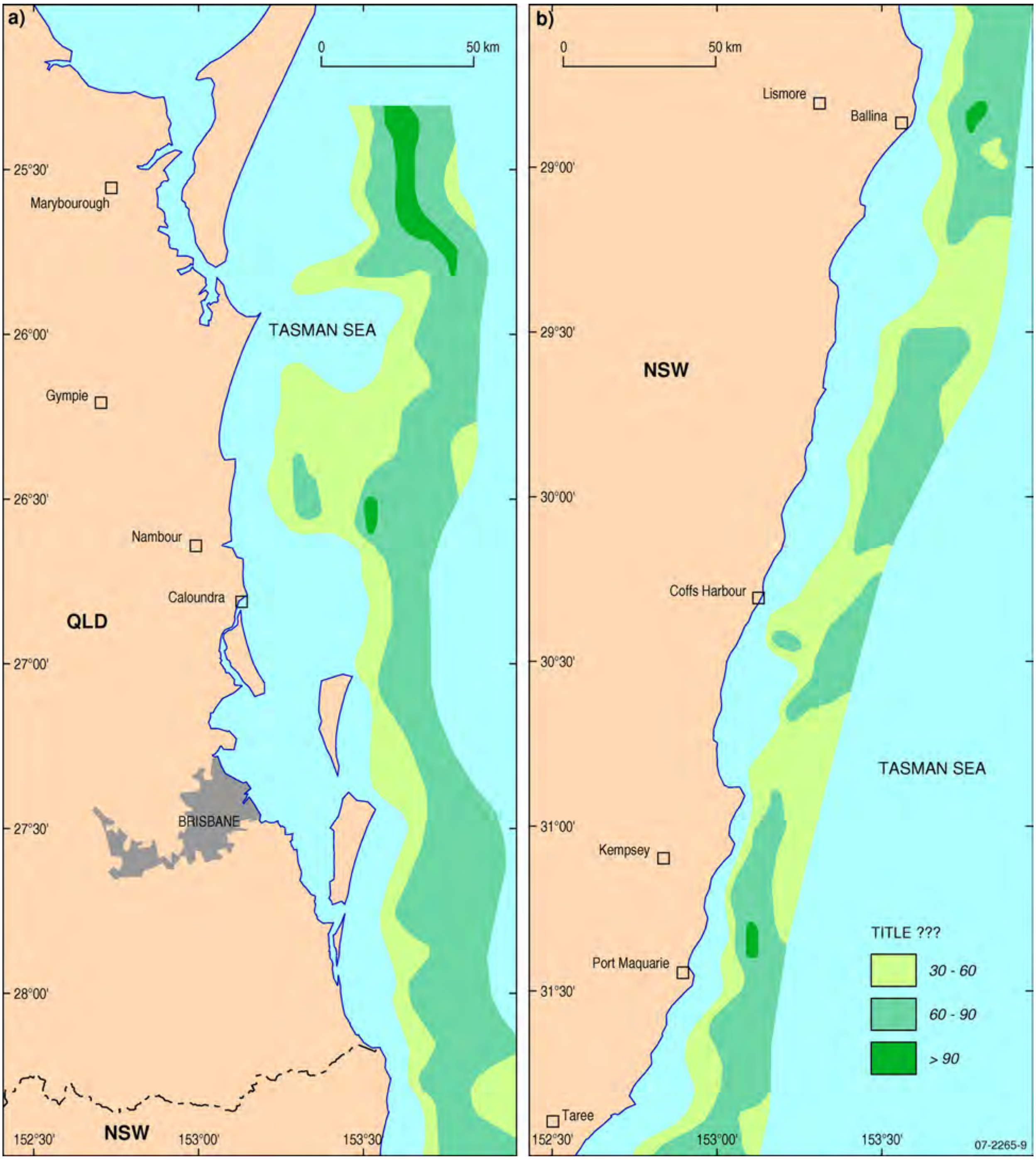


Figure 3.18. Maps showing the distribution of calcium carbonate (weight %) in surface sediments on the continental shelf and upper slope from Fraser Island to 32^{0}S. (Marshall, 1980).

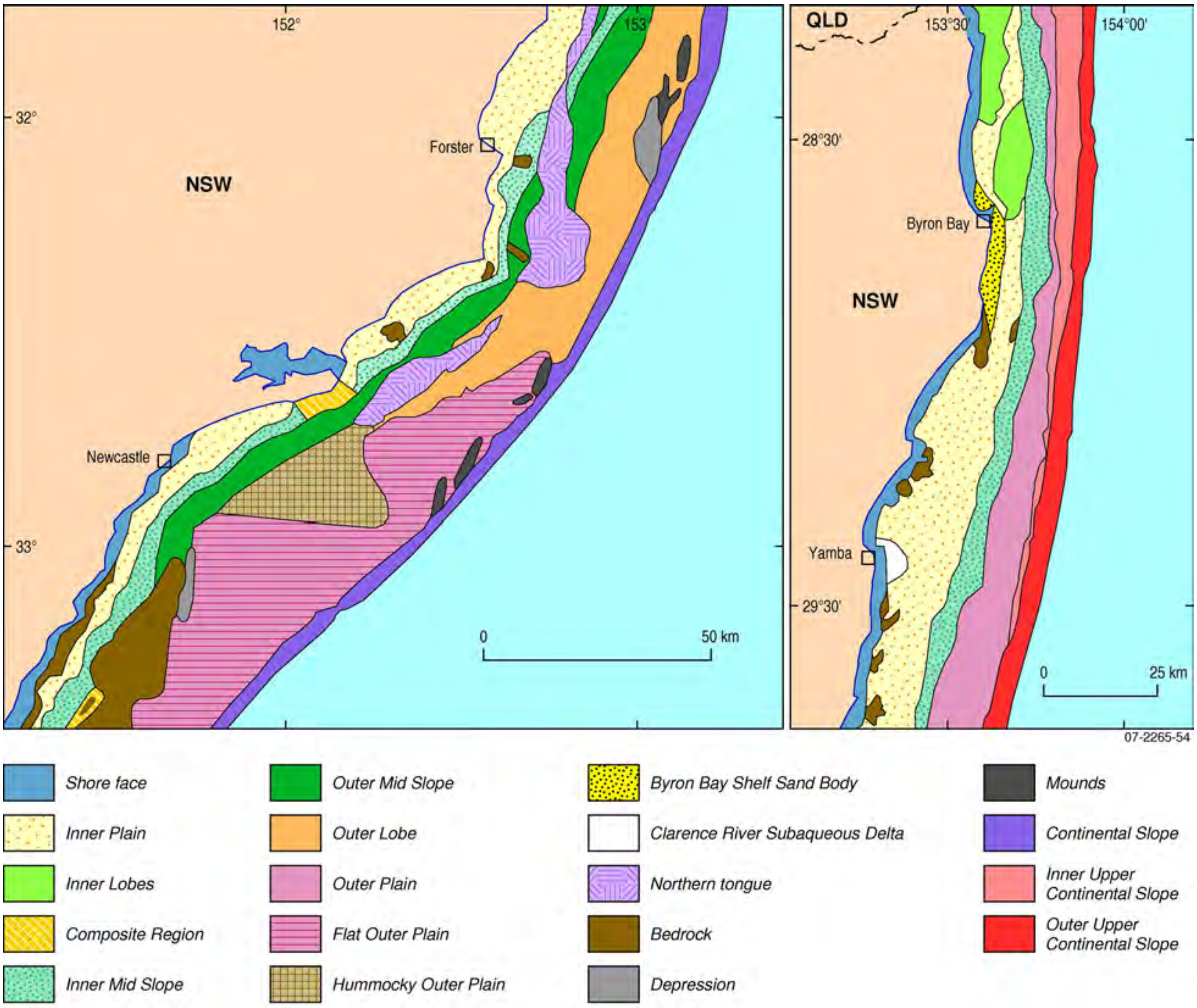


Figure 3.19. a) Distribution of morphological units and location of profiles, central NSW. Boyd et al*.,* 2004. b) Distribution of morphological units and location of profiles, northern NSW. (Boyd et al., 2004).

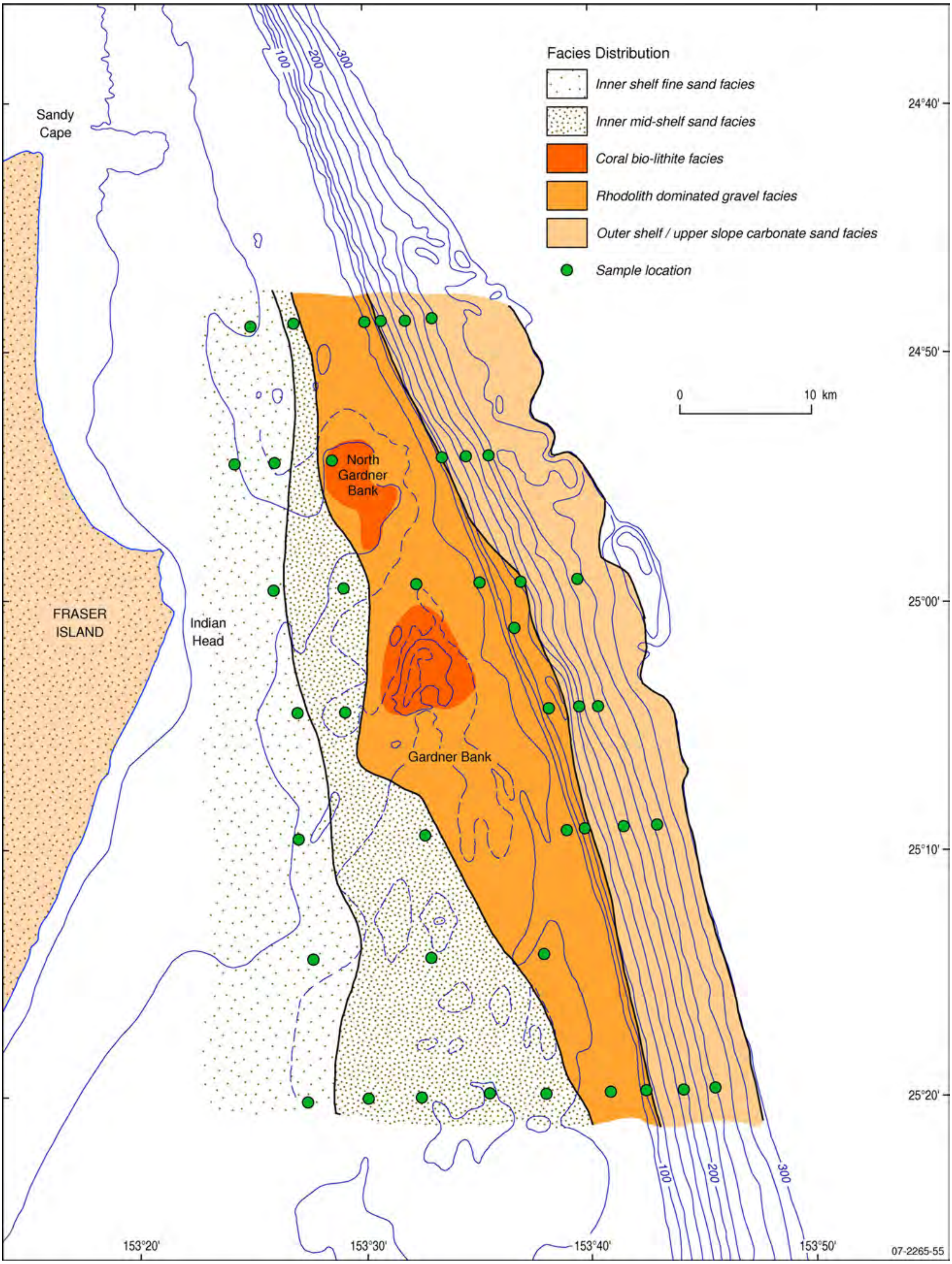


Figure 3.20. Bathymetric map offshore of Fraser Island showing the distribution of sedimentary facies on the shelf and upper slope around North Gardner Bank and Gardner Bank. (Marshall et al.,1998).

3.3. NSW AND SOUTHERN QUEENSLAND SLOPE

3.3.1. Geomorphology

The bathymetric data used for this review of the geomorphic features on the continental slope consists of two single-beam navy surveys collated by Packham (1983), a multi-beam survey of the central NSW slope by Glenn et al., (2007) and multi-beam surveys of the slope off Fraser Island by Boyd et al., (2004b, 2008). Ships of opportunity have also contributed their multi-beam bathymetry to the Geoscience Australia 250 m grid digital dataset but considering the area of the continental slope the multi-beam coverage is low (Petkovic and Buchanan, 2002). Seismic profiles of the slope have also been used to identify geomorphic features and subsurface geology by Jenkins (1984); Colwell et al. (1987); Heggie et al. (1992); Jenkins and Keene (1992); Colwell et al. (1993) and Glenn et al. (2007).

A characteristic of the NSW and southern Queensland continental slope is that its overall shape and orientation has been inherited from the initial rifting of the continental crust (Fig. 3.11). Canyons and mass wasting have modified this initial slope. Only on the upper slope has sedimentation determined the morphology. Most of the geomorphic features on the mid and lower slope are controlled by basement structures, particularly ridges and faults (Colwell et al., 1993; Glenn et al., 2007) (Fig. 3.21). Canyons exploit structural trends such as faults and along with mass wasting have shaped the slope, particularly below 1,500 m water depth. The location and pattern of canyon development is partly determined by tectonic features and underlying lithologies. Most canyons have linear segments where they are diverted by basement ridges or follow structural weaknesses. Many also have steps in their floors with scoured depressions as they descend over resistant strata.

The width of the slope from the shelf break to the abyssal plain/deep ocean floor varies by a factor of six. It is relatively narrow (40-50 km) in the south of the EMR, wider (50-60 km) off Sydney-Newcastle, wider still (60-70 km) off Sugarloaf Point-Port Macquarie, then narrower (20-35 km) off Port Macquarie-Coffs Harbour, gradually becomes wider (90-120 km) off Cape Byron-Moreton Island and then steps westward and gradually narrows to be 25-30 km wide off Fraser Island (Fig. 3.11) (Heap and Harris, in press). This variable width results in a range of slopes from 3° to 10°. Profiles of the slope vary from location to location but on a regional scale they can be used to divide the slope into: a) a generally convex upper slope above ~1,500 m that is relatively steep (4°-10°) particularly below 500 m; b) a rugged generally concave mid-slope down to ~3,000 m with locally steep scarps along slope, box canyons and flat terraces; and c) a very steep (10°-30°) and rugged lower slope below 3,000 m. The variety of slope profiles and widths is shown in Figure 3.22.

The two sections where the slope is particularly narrow are off Montague Island (south of 36°S) and between Port Macquarie and Coffs Harbour (310 30'S to 30°S). In these areas the slope is concave with many small canyons eroding the upper slope and joining to form large canyons down slope that cut through exhumed basement to make their way to the abyssal plain. Only remnants of a mid-slope terrace are preserved between the canyons or upslope from resistant ridges. It appears that these slopes have undergone considerable retreat due to erosion.

Off Montague Island the upper slope is steep (average 8°-10°) with many small slumps and canyons that amalgamate into larger canyons on the mid-slope. There are remnants of a mid-slope terrace at

2,200-2,400 m separated from the lower slope by a distinct along-slope basement ridge rising to 1,500 m that has dammed sediments on its up-slope side (Fig. 3.21; Colwell et al., 1993). The large canyons on the mid-slope are diverted along slope by this resistant basement ridge. At the northern end of this ridge is a large canyon with a dome-shaped seamount on the north side formed by an early Cretaceous igneous intrusion (Hubble et al., 1992). The lower slope is also steep and dissected by large canyons. The base of the slope is an abrupt contact with the abyssal plain/deep ocean floor at 4,850-4,870 m water depth.

Off Port Macquarie and Coffs Harbour there are no basement ridges but the large canyons have up to 2,000 m of relief on their sides on the lower slope. At both locations the large canyons have many small tributaries with their heads on the upper slope at 200-500 m water depth (Packham, 1983). At 30°S, offshore of Coffs Harbour, a particularly large canyon has incised to the 120 m isobath. A list of the canyons that have their heads at the 200-300 m isobath are given in Table 3.1. Such canyons may be younger than those with their heads in deeper water. They also provide access for deeper water to reach the shelf via upwelling.

Table 3.1. Location of canyon heads that reach the 200-300 m isobath

Canyon Head Offshore of:	Latitude
Montague Island	36^0 18’S
Jervis Bay	35^0 22’S
Jervis Bay	35^0 12’S
Jervis Bay	35^0 07’S
Newcastle	33^0 22’S
Newcastle	33^0 14’S
Port Macquarie	31^0 15’S
Port Macquarie	31^0 40’S
Coffs Harbour	30^0 00’S
Cape Byron	28^0 45’S
Tweed Heads	28^0 15’S
Noosa	26^0 16’S
Sandy Cape	24^0 40’S

The slope between Montague Island and Port Macquarie (36°S to 31° 30’S) is distinguished by a sediment wedge forming the upper slope and large canyons occurring approximately every 50 to 100 km on the mid and lower slope. The canyons are separated by a slope where the terrace between 2,000 and 3,500 m is largely preserved. The canyons are of two general types: those that have developed wide ‘box’ heads in the mid slope at ~1,500 m (e.g, Newcastle Canyon) and those like Sydney Canyon that have linear segments, few tributaries and have their heads in the upper slope.

The Newcastle Canyon and box canyons off Jervis Bay also have small, narrow tributary canyons that have incised the upper slope to the shelf break. There are many slumps on the mid-slope along this section of margin, plus some in the toe of the upper slope sediment wedge.

Between Coffs Harbour (30°S) and Moreton Island (27°S) the mid-slope terrace widens to form what can be defined as a small marginal plateau 300 km long and up to 120 km wide (Figs. 3.23 & 3.33). Two large, long canyons cut this plateau but little detail is known because this section of the slope is mostly unsurveyed.

North of Moreton Island the southern Queensland slope lacks a mid-slope terrace and also lacks large canyons, but is cut by many small to moderate size canyons. At 26° 35'S the slope steps east by 20 km. This step in the slope coincides with a possible change in basement lithology (Palaeozoic basement/Maryborough sedimentary basin) and the Recorder Fracture Zone in the adjacent oceanic crust (Hill, 1994; Exon et al., 2006a and b). A canyon has formed at this location and heads in 220 m of water (Marshall, 1978). The slope continues north with a width of 50-60 km until it turns NNW at 25° 20'S, again at a fracture zone, and again there is a relatively large canyon on the slope. Here the trend of the slope is probably controlled by the Maryborough rifted margin fault (Exon et al., 2006a) as it gradually curves westward until 24° 40'S, offshore of Sandy Cape and at the Cato Fracture Zone which is the northern boundary of oceanic crust. At this location the slope trends NW and becomes more gentle. The break in slope at its base is less distinct and shallower at approximately 3,700-3,500 m. This section of slope was formed later in the history of the opening of the Tasman Basin which, along with the lack of a large drainage basin in the hinterland may explain the absence of large canyons. Substrate geology controls the shape and course of canyons.

The base of the slope and its orientation is defined in several locations by linear fault scarps formed in the late Cretaceous-Paleogene as the result of rifting of the crust (Fig. 3.24). South of 36°S the base of slope trend north. At 36°S the orientation of the base of slope changes abruptly to trend more easterly and becomes a distinct linear scarp, probably the expression of a strike-slip fault formed during rifting. This ridge rises to 2,800 m and descends on its eastern side by a spectacular 1,000-2,000 m high scarp (average gradient 25°) to the abyssal plain/deep ocean floor at 4,800 m. It may have been formed immediately prior to spreading by the strike-slip fault proposed by Norvick et al., (2001). This trend continues to 34° 50'S where the base of slope steps seaward. This step is where the Barcoo Fracture Zone in the oceanic crust meets the continental crust of the continental slope. North of the Barcoo Fracture Zone the base of slope continues to the NE with a sharp contact with the abyssal plain/deep ocean floor at depths of 4,810-4,870 m and the slope becomes wider (47 to 65 km) to the north. At 33° 10'S (east of Newcastle) the base of slope again steps eastward to become a 10 km long, 800 m high scarp (4,000 to 4,860 m water depth) trending NE. This corresponds with an increase in the width of the lower slope. North of this scarp the base of slope trends more northerly at depths of 4,400 – 4,550 m. This continues to 31° 30'S off Port Macquarie.

The contact at the base of slope with the abyssal plain is shallower where small sediment fans occur off the Hunter River (base of slope 4,100 m) and off Port Macquarie - Coffs Harbour where three small fans give the base of slope an undulating contact ranging from 4,500 to 4,000 m (Packham, 1983; Jenkins, 1984). It is possible there is also a fan off the Clarence River but data are lacking. South of Moreton Island the base of the slope is very steep with a 1,200 to 1,400 m high scarp between 28° 03' and 28° 17'S and a sharp contact with the abyssal plain at water depths between 4,800 m and 5,000 m (Figs. 3.23 and 3.33). North of Moreton Island the base of the slope shoals to 4,500 m off Noosa, but remains a steep lower slope of faulted basement with an abrupt contact with the abyssal plain/deep ocean floor (Hill, 1992). The base of slope occurs at depths of ~3,800 m east of Fraser Island and shallows further to 3,500 m east of Breaksea Spit (Boyd et al., 2004b) and 4,000 m along the southern margin of the Marion Plateau (Fig. 3.35). This shoaling of the abyssal plain northward, and hence base of slope, is due to increased sedimentation on the abyssal plain in the north. Basement outcrop is common on the lower slope because sediment supply to the slope has been low and erosion has occurred.

In contrast to the rugged lower slope the upper slope, (above 1,500 m) is generally smooth because it is formed by deposition on the distal face of the prograding sediment wedge. Some geomorphic features do occur on the upper slope as bedrock outcrop, or due to currents scouring depressions and depositing ridges as well as mass wasting forming canyons and slump scars. Incipient slumps form arcuate ridges on the upper slope and are common below 1,000 m water depth. South of Jervis Bay the upper slope is less convex due to the steeper slope above 500 m compared with that depth interval further north.

The only canyons to erode into the shelf break are offshore of Jervis Bay. Three canyons have their heads in 150 m of water and are known as the Shoalhaven Canyons. Off Newcastle three of the Hunter Canyons have incised their heads to water depths of 200 m and the Manning Canyon off Taree starts in 370 m of water.

Significant basement outcrops occur on the upper slope off Wollongong, off Sydney and off Port Macquarie. On the upper slope off Wollongong and extending for 11.4 km along slope is a large outcrop of basement that has a relief of at least 111 m, rising to a depth of 650 m from water depths of 790-680 m and continuing down slope as rock outcrop to 990 m on the eastern side (Fig. 3.25). It has an erosional moat along the western margin, presumably due to the south flowing EAC. A depositional lee ridge has accumulated sediment in the southern or downdrift area behind the basement block. Offshore of Sydney the Mt Woolnough seamount rises 130 m from the surrounding seabed (~550 m) to three peaks, the shallowest being 383 m (Fig. 3.26). It is a rugged feature (1 km wide at base) and is the eroded remains of a basalt volcanic cone. Associated smaller rock outcrops extend 4 km to the NE with up to 50 m of relief. To the SE, in 780 m of water, are five en echelon ridges of rock 1.3 km long with up to 25 m of relief. The effect of the EAC encountering the Mt Woolnough seamount and nearby rock obstacles can be clearly seen in the sediment bedforms. The current eddies in the lee (south side) have eroded a 40 m deep moat at the base of the seamount with a broad (800 m wide) 10 m high depositional levee forming a 2 km long tail downstream. There is a similar sediment ridge extending 2 km on the north side. The rock outcrops at 800 m water depth also have current scours in the surrounding sediment indicating that the EAC erodes sediment to at least this depth.

Another basement outcrop occurs offshore of Port Macquarie between the 200 and 300 m isobaths. It is 8 x 2 km forming a ridge along the upper slope rising to within 175 m of the surface with a scoured channel to depths of 254 m on its upslope side and outcropping to 280 m on its down-slope side. Boyd et al. (2004a) describe mounds on the upper slope (140-165 m) north of Port Stephens (Fig. 3.19a). They interpret them as temperate carbonate reefs or possibly gravity slumps.

The width of the slope offshore of Jervis Bay is 47 km. The Jervis Bay region has a very steep upper slope of 80 to 90 between the 400 and 600 m isobath and 70 to 80 down to approximately 1,500 m.

Prominent features on the upper slope are six narrow canyons and many smaller down slope gullies (Fig. 3.24). In their upper reaches these canyons are 200 m deep, meander and are v-shaped. Below 1,100 m water depths the canyons erode into the lower part of the sediment wedge and are flat-bottomed, 200-400 m wide and only 100 m deep. The slope of their thalweg becomes abruptly steeper below 1,500 m (from 4.30 to 90) where they join a large box canyon on the mid slope. On the mid-slope between 1,500 and 2,500 m there are scarps up to 180 and there are numerous retrogressive slides. Below 2,500 m is a 15 km wide terrace, largely un-surveyed, with a prominent linear ridge forming its outer margin.

The narrow and relatively straight canyons extending up the slope in the Jervis Bay region are related to the Shoalhaven River drainage and its deposition of sediment on the shelf and slope during the Cainozoic. They were incised into the sediment wedge above the 1,500 m isobath by turbidity currents carrying sediment down slope. Meanders in the canyons are caused by deflection by more resistant strata in the sediment wedge as are steps in their otherwise flat floors. The particularly large step down in the floor of the canyon at 1,500 m is caused by a change to a more resistant basement lithology.

Large mass wasting features occur off Wollongong. Figure 3.25 images a straight-sided slide in the upper slope sediment wedge that is 5-7 km wide and extends down slope for 15 km between depths of 400 m and 1,300 m. It starts as two broad slumps at the 400 m isobath. The floor of the slide scar is asymmetrically deeper to the south, with the southern scarp up to 200 m high and the northern scarp stepped, and a maximum of 120 m high. The floor of the slide has grooves ranging from 2-40 m deep, from 90-500 m wide, and up to 5.5 km long. To the north of this slide is the Bulli Slide (Fig. 3.25) located immediately down slope from the large outcrop of basement described earlier. It is the largest gravity failure feature discovered so far on the margin. It has an arcuate plan shape with a maximum width in the upper erosional slide area of 11 km. The slide feature descends from 950 m to over 3,300 m water depth over a distance of 22 km. The floor of the slide has a slope of 5.90. The margins of the slide are up to 200 m high.

The head of Sydney Canyon is relatively subdued, consisting of slump scars in 880 m of water that open down slope into a broad amphitheatre 4 km wide with 140 m relief at 1,000 m. The relief on the slumps is subdued and slope-parallel, perhaps because they are draped by younger sediment or have been modified by the EAC. From here the canyon runs down slope to the southeast for 11 km to 1,800 m. The southern side is linear and steeper than the northern side. Subdued slump scars are common on the continental slope on either side of the canyon from around its head to 1,800 m where it turns south and is linear and oblique to the slope for further 12 km to 2,800 m and joined by four tributaries. Relief in the main canyon is greatest (750 to 600 m) where the canyon floor is between 1,800 and 2,500 m, the steepest side is initially the western side but below a floor depth of 2,100 m the steepest side is on the east (seaward) as the floor continues to descend to the south. A linear ridge forms its eastern side like a rampart along this length. It then turns southeast where the floor descends from 2,300 to 3,100 m over a series of steps with 'plunge pools' and continues for 20 km to 3,850 m before turning south. The floor of the canyon at 3,850 m steps down 300 m into a 'plunge pool' that is 100 m deep (water depth 4,200 m) in the 3.5 km wide floor. At this point the canyon sides have relief of 600 m up to the adjacent slope depths of 3,600 m. The canyon continues for a further 12 km to exit on the abyssal plain at 4,896 m as a broad, 6 km wide channel 400 m deep. The total length of the canyon is 55 km.

Seaward of the Hunter River mouth the most obvious features on the slope are a series of seven canyons on the upper slope that begin in around 200 m water depth, with incision of around 350 m and canyon widths of 1-2 km and continue down slope for 15-20 km to 1,500-2,000 m depth where they join a large box-canyon (Fig. 3.27). The canyons are sinuous and the v-shaped (20-25° slopes) down to the 800 m isobath. In deeper water they become linear, flat bottomed and terraced. Parts of the toe of the sediment wedge between these small canyons have been removed by gravity slides. On one slip surface are circular depressions up to 600 m across and 70 m deep (Fig. 3.27), which may be due to water and/or gas seepage.

The seven small canyons on the upper slope feed into a large box-canyon feature beginning in ~1,200 m water depth at the top of several failure scars have eroded the mid-slope. These failure scars coalesce down slope forming a single narrow canyon 7 km wide with over 800 m of relief. This is known as the Newcastle Canyon and is mainly developed in bedrock with around 400 m of relief at its head (Fig. 3.28). The canyon continues for 22 km before ending in around 4,300 m water depth where it is incised into a fan forming the rise. Irregular topographic features on the order of 400 m high and 2 km wide occur in the canyons and on the slope below 1,500 m off Newcastle. These narrow ridges and pinnacles occur as resistant bedrock within the canyons and on the edge of scarps and are most likely eroded volcanic intrusions (Fig. 3.28).

Offshore of Sugarloaf Point several significant slides occur on the seaward edge of the sediment wedge at water depths of 800-1,500 m (Fig. 3.29). One is from 3-5 km wide, is a minimum of 6 km long and has steep sides (20°) up to 140 m high. The slide mass has fed into a canyon down slope. The floor of the slide has a slope of around 3°. Another slide begins in a relatively shallow 800 m water depth and its erosional scar is up to 3.6 km wide and 3 km long and extends to 970 m water depth. This slide is unusual in that it has formed a debris flow that has been preserved further down slope with more than 80 individual blocks up to 40 m high strewn across a debris apron 5 km long and 4 km wide. The blocks are probably semi-consolidated clays and chalks. Boyd et al. (2008) used multi-beam bathymetry to describe the slope in the area off Fraser Island in detail. This is the only area where significant amounts of sediment are moving across the shelf and reaching the heads of canyons on the upper slope. These canyons can be considered active whereas the canyons further south are not receiving sediments, because they have been cut off from the rivers that supplied them by the rise in sea level. Off Fraser Island the slope is relatively narrow (20-30 km), highly erosional and steep with average gradients of 6.5-100 (Boyd et al., 2008). The upper slope is steep (14°) and consists of two or three steps formed by lithified Quaternary carbonate platforms (Payenberg et al., 2006). Below this at a depth of 150 m many gullies have their heads. They are up to 330 m wide and 40 m deep and they join to form larger canyons in the middle-lower slope, which in turn feed into the Capricorn Sea Valley at the base of the slope. The abyssal plain abuts a steep lower slope at 3,500 m. To the south of this area, off Gardner Banks, the top of the slope is at 85 m water depth where there is a 'nick-point' at the base of a 20 m high cliff (Marshall et al., 1998). Seaward of this reef there is a gently sloping (1-30) terrace for 6-10 km with a mound on its outer margin at 300 m water depth, seaward of which the slope increases. Marshall et al., (1998) choose this outer change in slope due to a submerged carbonate platform as their shelf break resulting in their shelf breaks in southern Queensland being particularly deep at between 210 and 450 m. The lower slope is steep and basement outcrops along a fault which meets the abyssal plain/deep ocean floor at 4,500 m water depth (Hill, 1991).

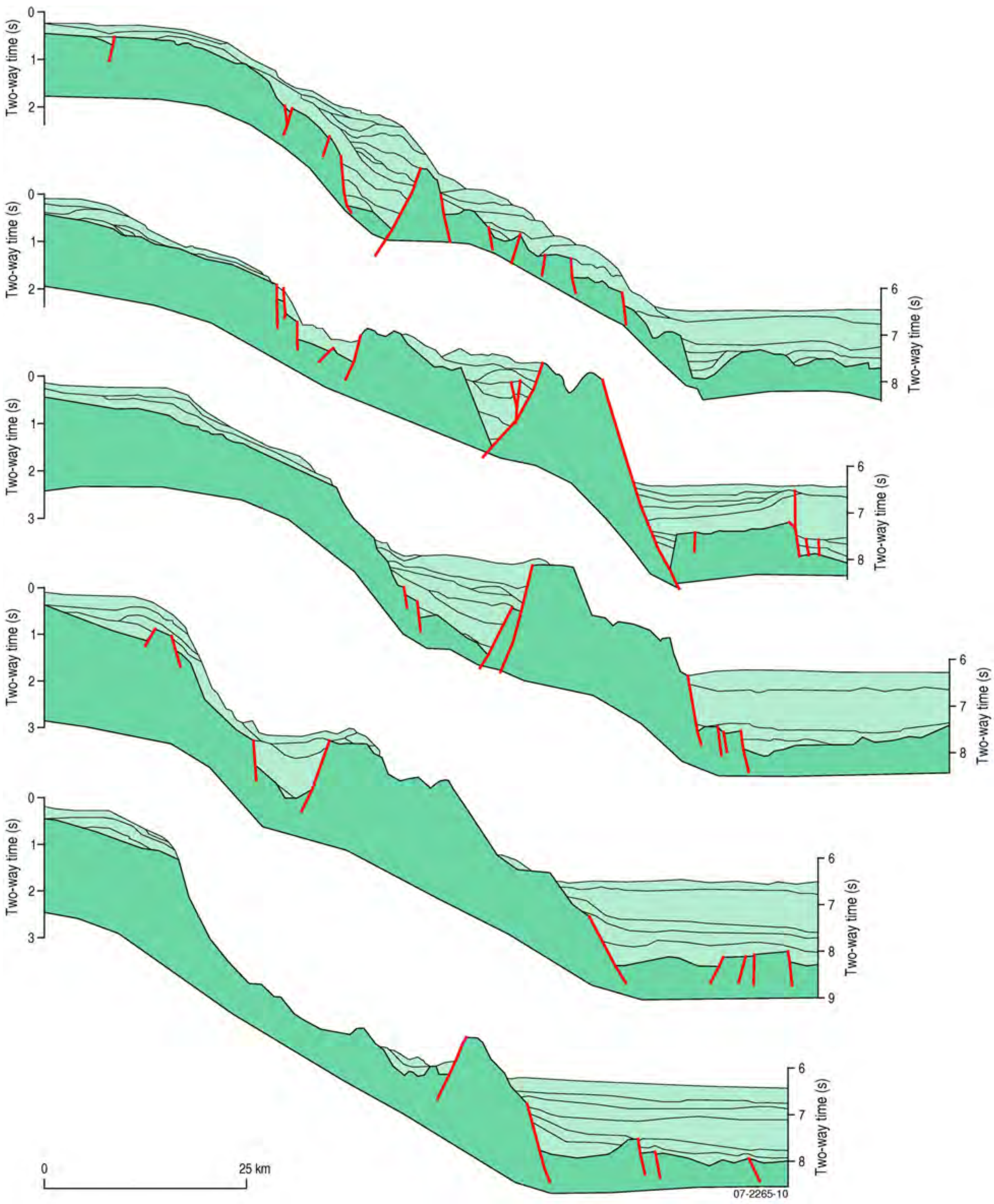


Figure 3.21. Line drawings of interpreted seismic profiles from the southern NSW slope and adjacent abyssal plain showing the relationship of sediments to basement. Note the convex upper slope with sediment wedge, concave mid-slope with erosion or sediments behind an outer ridge and a rugged and steep lower slope. BMR Survey 68, (Colwell et al., 1993).

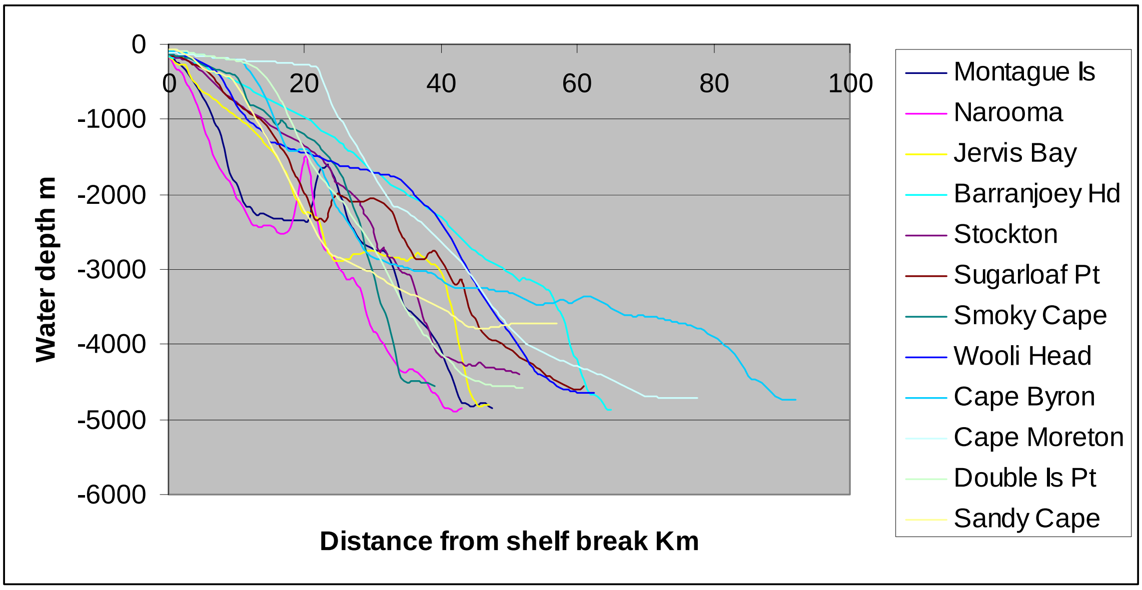


Figure 3.22. Profiles of the slope from the shelf break to the abyssal plain. A range of widths and slopes occur on this margin. Profiles are approximately equally spaced from Montague Island to Fraser Island.

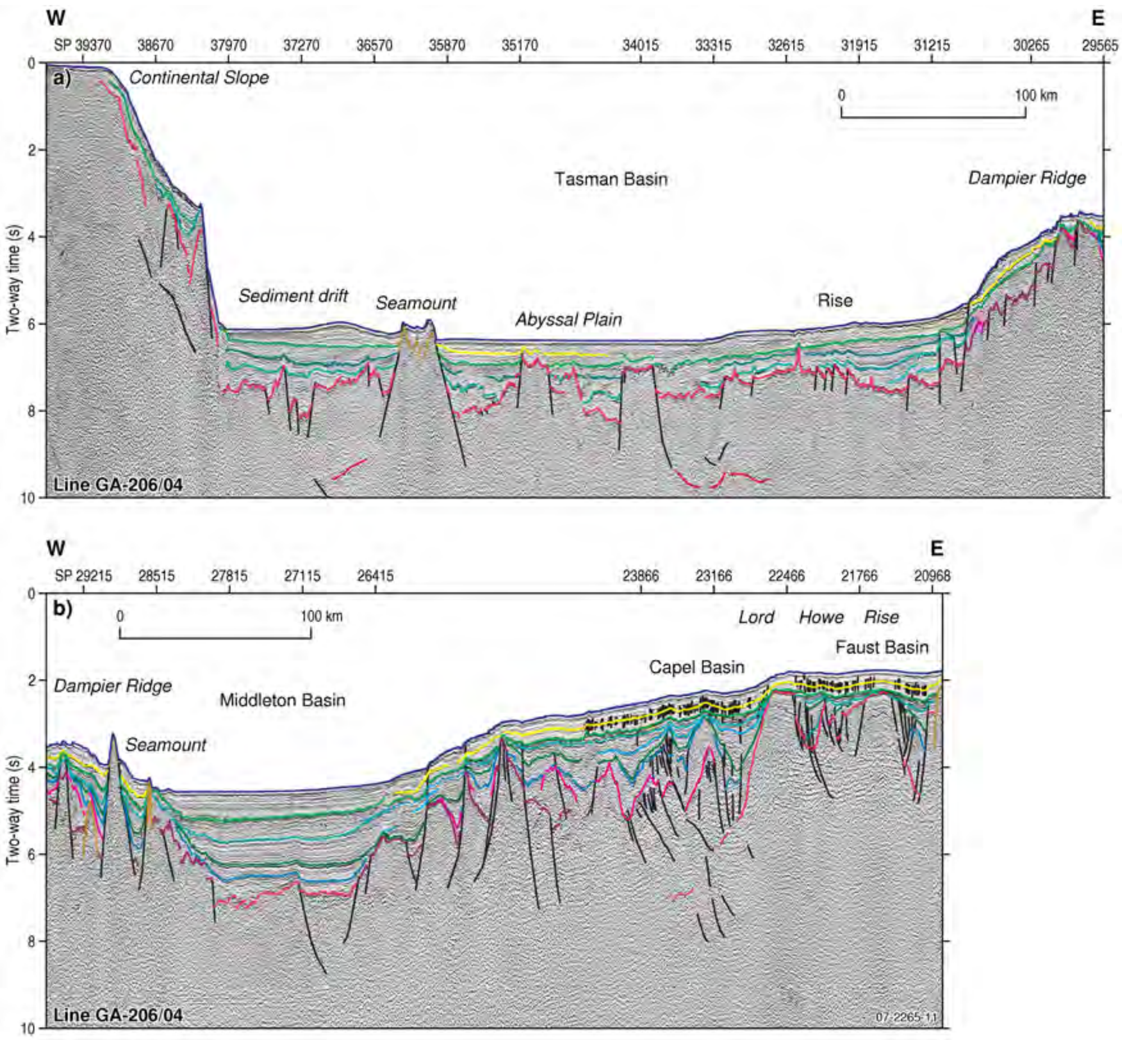


Figure 3.23. Seismic profile from Yamba, NSW, across the Tasman Basin to the Dampier Ridge. It shows the wide slope, lack of rise, abyssal plain and the rise at the foot of the Dampier Ridge. Note the drift deposit on the abyssal plain/deep ocean floor and the basement seamount. Location on Figure 3.1. (Van der Beuque et al., 2003).

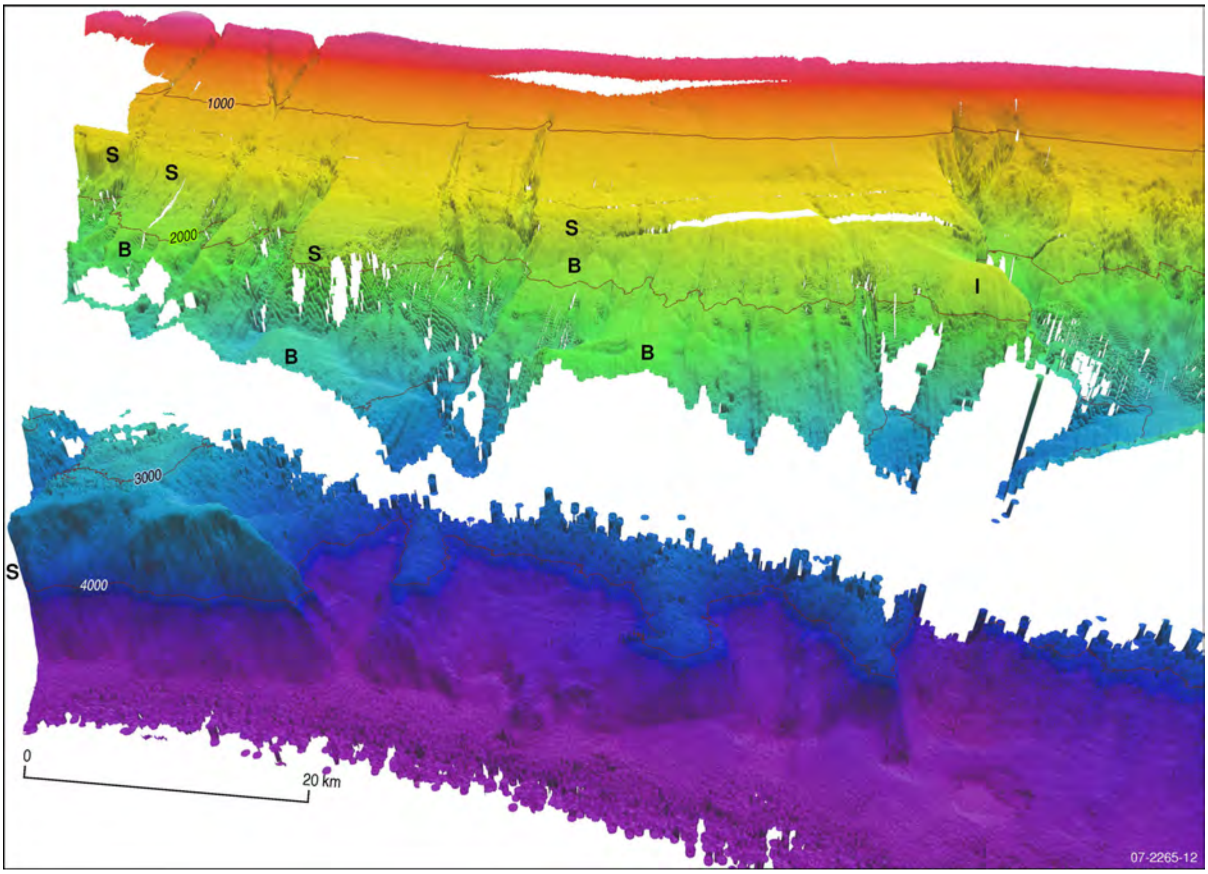


Figure 3.24. Oblique view of the slope off Jervis Bay looking up slope to the west. Note the canyons in upper slope, mid slope terrace and large canyons. Other features are domed hills of resistant igneous rocks (I), scarps (S) and slide blocks (B). The scarp at base of slope is a major feature. V.E. 6x. Location on Figure 3.11.

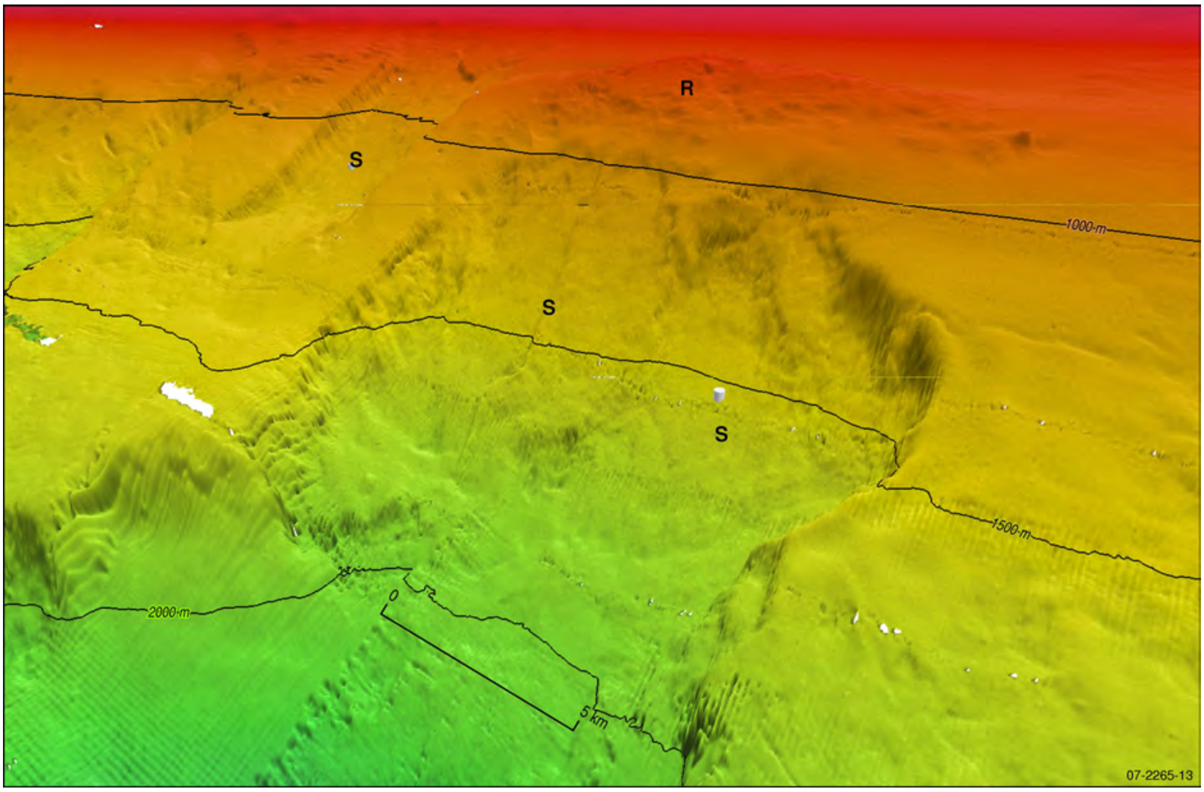


Figure 3.25. Oblique view looking up slope off Wollongong to the west. Note the rock outcrop (R) on the upper slope, slide scars (S). The slide on the left is in the sediment wedge whereas the larger slide is on the seaward face of the basement outcrop. V.E. 6x. Location on Figure 3.11.

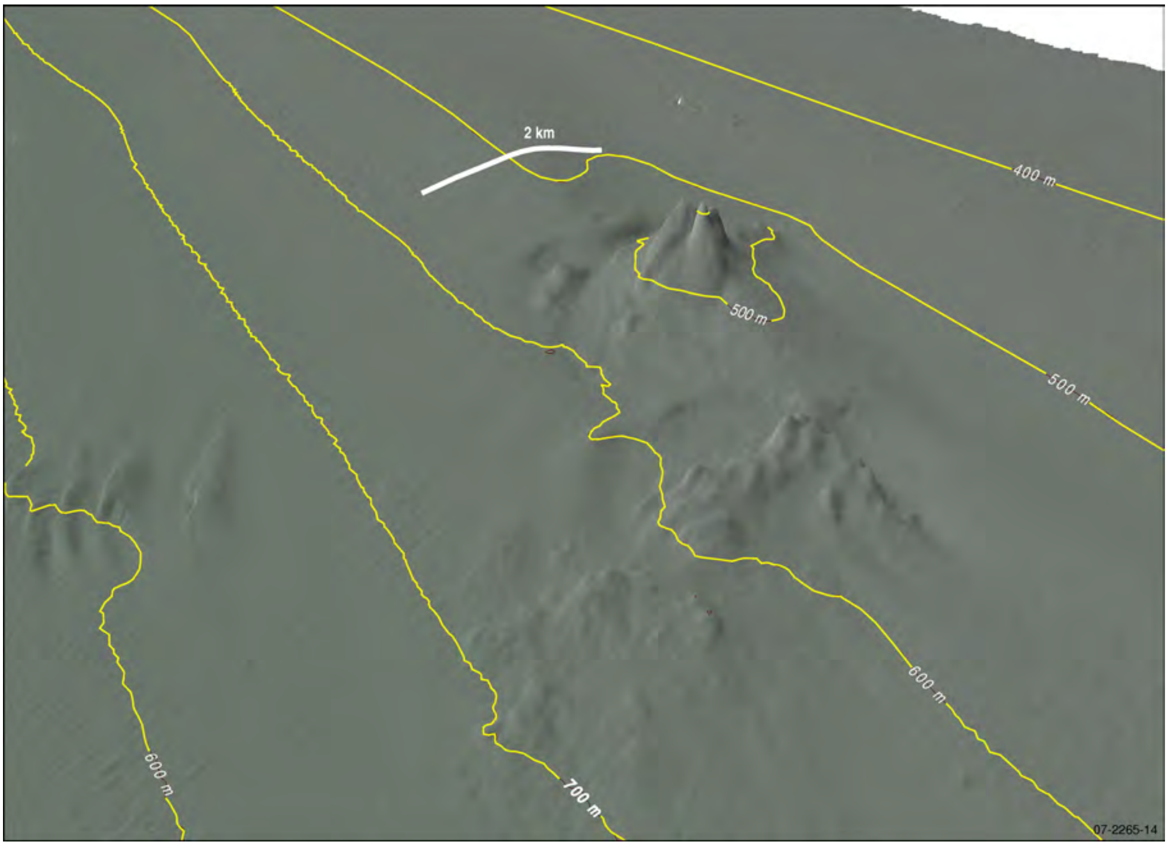


Figure 3.26. Oblique view of Mt Woolnough on the upper slope offshore of Sydney looking southwest. Note the erosion moat and depositional lobes down to 800 m water depth around the bedrock due to the EAC. V.E. 6x. Location on Figure 3.11.

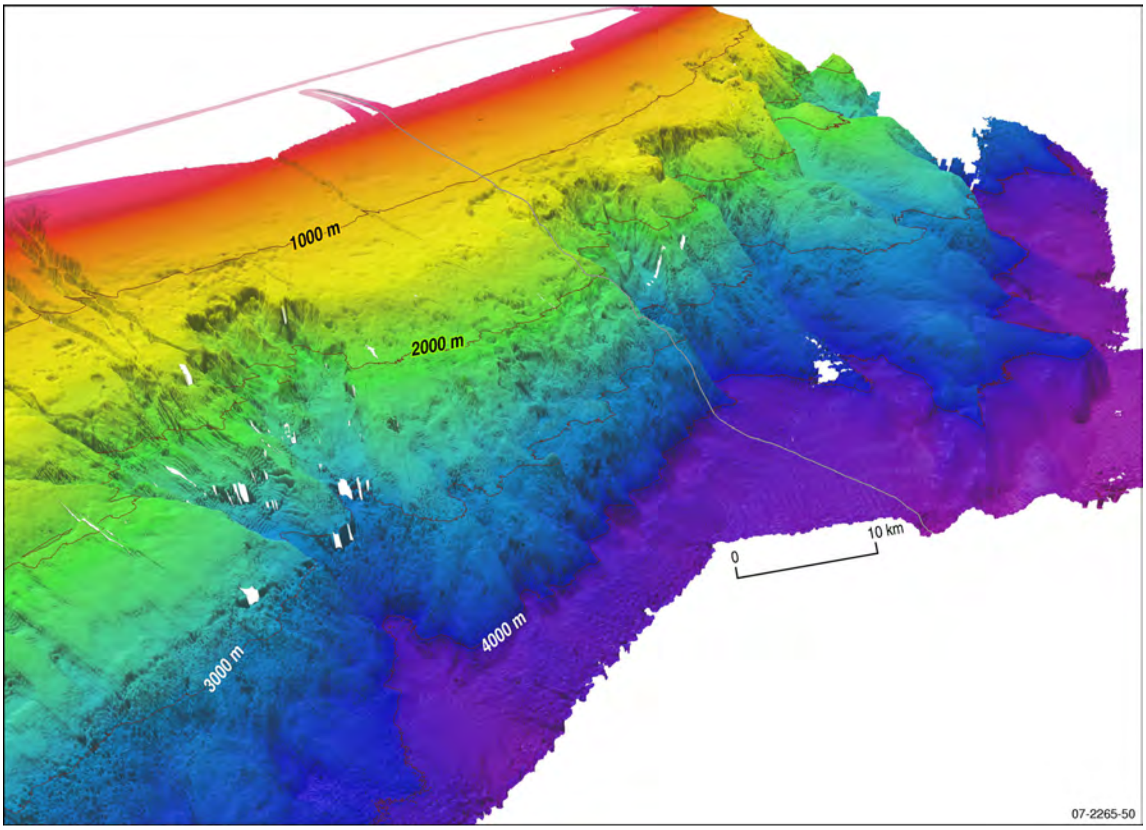


Figure 3.27. Oblique view looking north up slope off Newcastle to the northwest. Note the small canyons (Hunter Canyons) and slumps in the upper slope sediment wedge. They feed into wide box-canyons on the mid-slope. The lower slope is

rugged with linear (fault?) scarps. The canyon on the left (Newcastle Canyon) has incised into a fan at the base of slope. Note the offset of the lower slope to the east at the top right of image. V.E. 6x. Location on Figure 3.11.

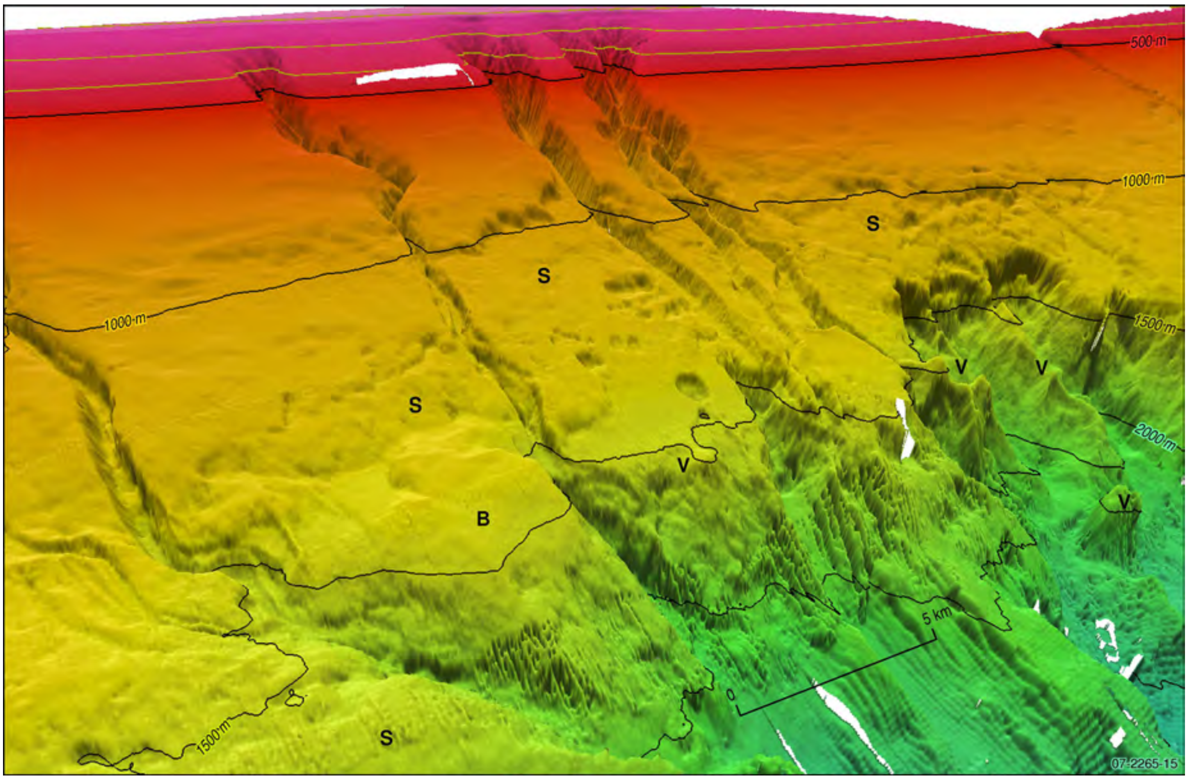


Figure 3.28. Oblique view enlargement of Figure 3.27 showing the Hunter Canyons incised in the upper slope sediment wedge. They feed into the wide box canyon in the mid and lower slope. Isolated round peaks are interpreted as volcanic (V). Slide scars (S) and displaced blocks (B) indicate headward erosion of the box-canyon. V.E. 6x.

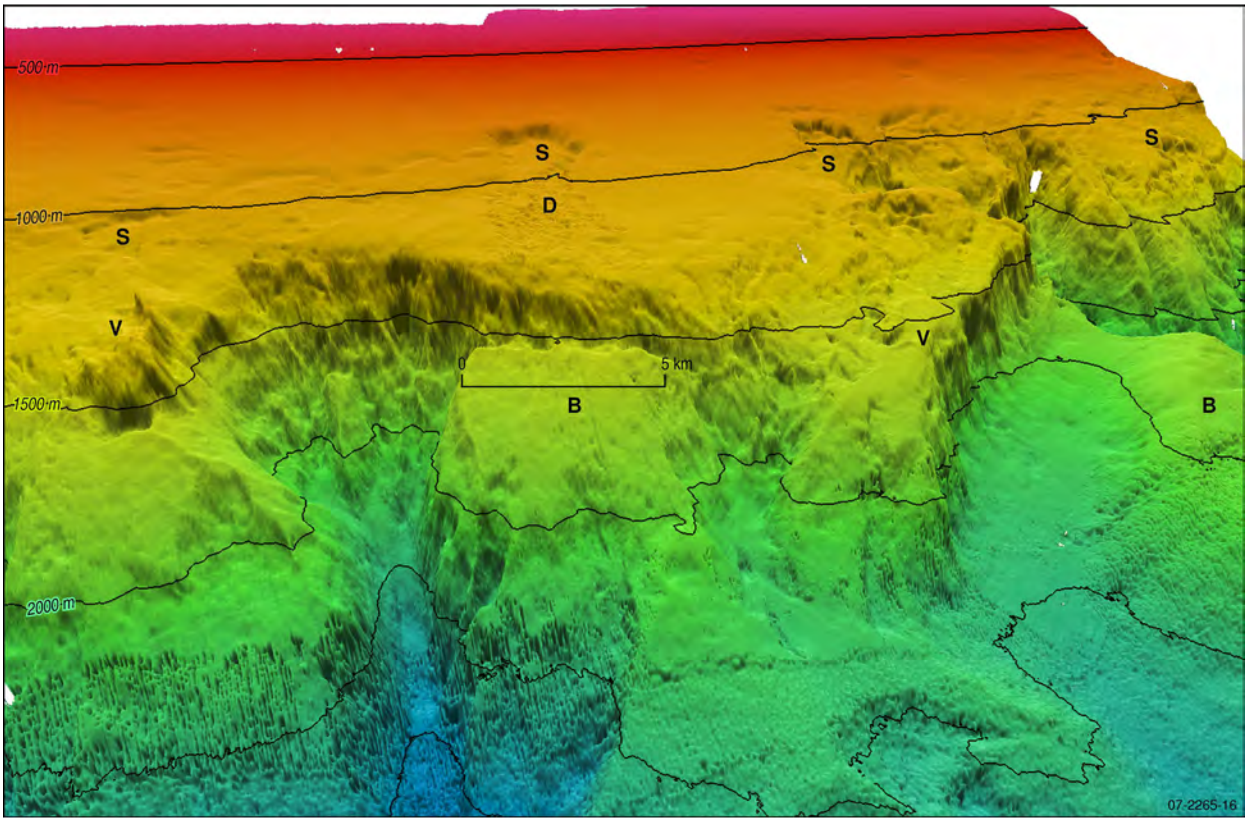


Figure 3.29. Oblique view enlargement of Figure 3.27 showing slope failure in toe of the sediment wedge (S = slide scars), debris flow (D), displaced blocks (B) and volcanic ridges and pinnacles (V). V.E. 6x.

3.3.2. Surface Sediments and Rocks

The surface sediments on the slope of the EMR are typical for this hemi-pelagic environment: a suite of biogenic and terrigenous sediments coming from the shelf mixing with the remains of plankton that live in the water column above the slope and benthos that live on and in the seabed. Packham, (1983), Hubble and Jenkins (1984a and b), Heggie et al., (1992), Howard (1993), Lane and Heggie (1993), Troedson (1997) and Troedson and Davies (2001) have made studies of parts of the slope. A summary of the published data on bottom photographs, cores and dredges from the continental slope is given in the Appendix C Tables 3.4, 3.5 and 3.6.

The main terrigenous components in the slope sediments are clay minerals and silt-size quartz, whereas the main plankton components are foraminifers, coccolithophores and minor pteropods. The principal remains of slope benthos are echinoderms and sponges (both siliceous and calcareous). Green glauconite pellets are a common sand-size authigenic mineral on the upper slope and phosphate nodules and clasts occur off northern NSW (von der Borch, 1970; O'Brien and Heggie, 1990). The texture and composition of the sediment varies with water depth and their colour is various shads of grey and olive with a very thin brownish oxidized layer at the surface. Bioturbation keeps the sediment relatively homogenous.

The upper slope sediment is calcareous muddy sand or a sandy mud, due to the relatively high content of sand-size foraminifers, glauconite, benthic particles and fine-sand size quartz. The texture varies along the margin. O'Brien and Heggie (1990) found sediment off northern NSW between 350-460 m water depth to be sands containing 30-60% carbonate (mostly planktic foraminifers), 20-35% silt to fine-sand size angular quartz, 7-30% glauconite pellets with coccoliths and clay minerals in the minor amount of mud. Boyd et al. (2004a) report significant amount of siliceous sponge spicules in the sediment on the upper slope offshore of Tweed Heads. The sediments on the upper slope offshore of major rivers tend to be less calcareous, presumably due to turbid plumes, and the sediments on the slope off southern Queensland are the most carbonate rich due to particles from the shelf, particularly algae (Troedson, 1997; Glenn et al., 2007).

The sediments on the mid-slope are more mud-rich due to the remains of coccoliths. The carbonate content is generally higher than for those on the upper slope, as there is less dilution by terrigenous mud. The oxygen minimum zone is poorly developed so the seabed is well oxygenated. The organic carbon content of the sediments is <1%. Bioturbation is abundant but beneath a brownish grey oxidizing layer of a few centimeters the sediments are an olive grey colour and are reducing. The lower slope sediments are also muds but have undergone carbonate dissolution because they are below the lysocline. The foraminifers are preferentially dissolved. Siliceous plankton such as diatoms and radiolarians are a rare component of the sediment because of the relatively low productivity in the waters along this margin. Thus there is no impact on the benthos of the oxygen minimum zone as is the case in more productive areas.

Cores from the upper and mid-slope (500-3,000 m) offshore of Sydney show surface sediment with increasing mud with depth: from 25% at 500 m, 40-55% at 1000-1500 m and >80% at 2,000-3,000 m, and with carbonate content varying between 40-50% (Howard, 1993; Troedson and Davies, 2001). Quaternary glacial/interglacial cycles have been reorganized in these sediment cores, through the changing mud and particularly carbonate content. Dating of the cores gives a sedimentation rate of 2-

5 cm ka-1 for the Holocene. A 5 m core from the lower slope off Port Macquarie (water depth 3768 m) consisted of surface sediment of calcareous mud (46% $CaCO_3$), with fine terrigenous sediment mixed with foraminifers and coccoliths (Eade and van der Linden, 1970). Below the reddish oxidized surface the sediment is olive grey to yellowish grey, bioturbated with no bedding present and the carbonate content varies from 30 to 57%. The surface sediments from cores on the slope offshore of Noosa (Queensland) in water depths of 842 and 1,022 m are 70-80% mud and 50-60% carbonate (Troedson and Davies, 2001).

The distribution of sediments on the slope, and hence their texture and composition, is also affected by physical processes such as contour currents, turbidity currents, grain flows and debris flows. Colwell et al. (1993) and Glenn et al. (2007) identified debris flow deposits and current scouring and redeposition of sediment particularly around basement highs. Troedson (1997) recognised sandy turbidites in cores from the slope, and Marshall et al. (1998) reported redeposited carbonate sands to a depth of over 300 m offshore of Fraser Island. Boyd et al. (2008) found quartz sands from the inner shelf in the floor of submarine canyons down the entire slope off Breaksea Spit. This is in contrast to the hemi-pelagic mud and carbonate ooze on the slope adjacent to the canyons, that contain 23-57% mud and 25-43% carbonate. Marshall et al. (1998) found similar values in slope sediments down to 250 m offshore of Fraser Island with the mud content increasing from <5% to >20% and the carbonate content decreasing from >90% to < 50% down slope in less than 5 km.

A feature of the upper slope off northern NSW is the presence of hardgrounds, iron rich nodules and phosphate nodules in water depths <450 m (von der Borch, 1970; O'Brien and Veeh, 1980; Marshall, 1983; O'Brien and Heggie, 1990). These have been recognized on seismic profiles, in bottom photographs and by sampling. Many hardgrounds are present at water depths of 250-380 m north from Yamba (29° 25'S to 29° 08'S) and south of Coffs Harbour to Port Macquarie (O'Brien and Heggie, 1990). Ferruginous phosphorite occurs as both discrete dark-brown nodules 1-15 cm in diameter and as iron cemented hardgrounds. Marshall (1983) found nodules with iron contents up to 36% on the upper slope in shallower water depths of 197-274 m. Von der Borch (1970) distinguishes three types of phosphatic sediments on the seafloor: a) equant light to dark grey nodules 2-3 cm in size encrusted with calcareous organisms, b) nodules that have a thick coating on iron oxide and contain glauconite, and c) ferruginous-glauconitic- phosphorite conglomerate slabs 10 cm or more in size. The phosphate-rich nodules have a late Pleistocene or Holocene age whereas the iron-rich nodules are older than 800,000 years and probably Miocene (O'Brien and Veeh, 1980; Kress and Veeh, 1980; O'Brien and Heggie, 1990). O'Brien et al. (1981) propose that the younger nodules form by cementation with bacterial apatite.

Roberts and Boyd (2004) cored mud-rich biogenic gravel (bryozoa and sponge spiclules) and carbonate cemented clastic hardgrounds in 125 m water depth off northern NSW. Similar mounds with hardground outcrop have been identified on seismic profiles further south between 32° 30'S and 32° 55'S offshore of Sugarloaf Point and Nelson Bay (Boyd et al., 2004a; Glenn et al., 2007). They occur as strong seafloor reflectors on the shelf edge and upper slope between water depths of 150 to 180 m along with an exposed older surface that could be a hardground. This surface also outcrops where erosion has occurred further south (33° 15'S) near the heads of the Newcastle canyons in water depths of 180 m.

The upper slope in northern NSW is the only place where phosphate and silica sponge spicules form a significant part of the surface sediment. Both are indicators of relatively high surface water productivity in this area. The presence of nodules, slabs and hardgrounds on the seabed in this area is probably due to the low sedimentation rate because of winnowing and erosion by the EAC. Local upwelling also occurs where canyon heads have cut into the upper slope. Currents probably flow up and down these canyons on a regular basis and winnow the sediment on the seabed. Thus these canyons may have a more sandy floor than those in deeper water.

Seismic profiling has confirmed the presence of basement and volcanic outcrop as ridges normal to the slope (Fig. 3.30). They have been dissected by canyons and exposed by slumping. Dredging on the slope has revealed a variety of rock types representing basement lithologies (metamorphic rocks, igneous rock, sandstones, shales and limestone), igneous intrusions contemporaneous to rifting (monzodiorite, serpentinite), syn-rift sediments (mudstones and sandstones), post-rift sediments (limestones and chalks) and volcanic rocks (Heggie et al., 1992; Hubble et al., 1992; Quilty et al., 1997; Quilty and Packham, 2006). Rock outcrop on the seafloor also provides habitats for benthic organisms. One dredge offshore of Sydney in 1,600 m of water recovered a benthic community of living corals, sponges, annelids, echinoderms, brachiopods, bivalves and gastropods (Heggie et al., 1992). The location and composition of dredge samples from the EMR continental slope are given in the table in Appendix C Tables 3.6.

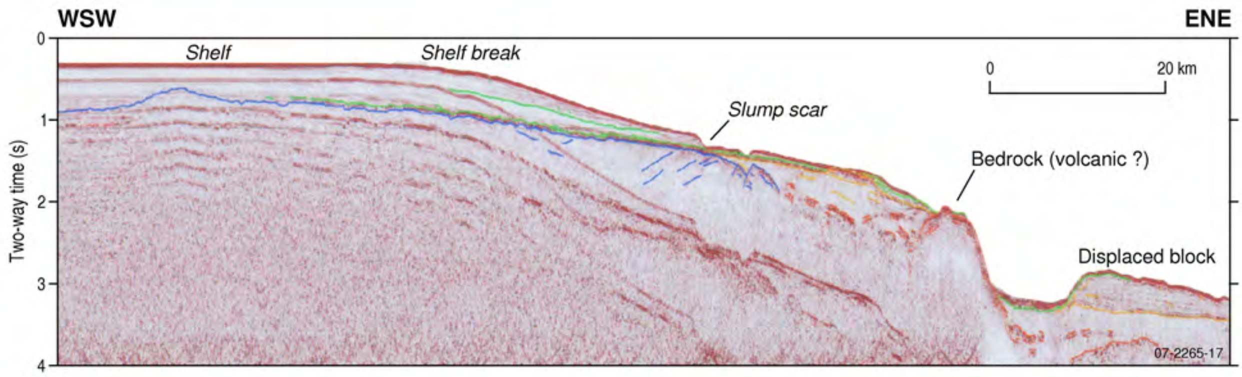


Figure 3.30. Seismic profile across the slope illustrated in Figure 3.29. Note the upper slope sediment wedge, slide scar, older rift sediment basin in the mid slope with a volcanic ridge exposed on the down slope side. Location on Figure 3.11. Line 007, Glenn et al., (2007).

3.4. ABYSSAL PLAIN/DEEP OCEAN FLOOR

3.4.1. Geomorphology

The Tasman Sea Basin is roughly triangular in shape with its apex in the north at the southern flank of the Marion Plateau and Cato Trough and widening to the south. This shape reflects its plate tectonic origin when the faulted margins of continental crust opened from the south and new oceanic crust was emplaced (Gaina et al., 1998a & b). The rectilinear margins of the basin, on a scale of 10 to 200 km persist today as a result of this original structural break-up pattern of the continental crust (Fig. 3.1). Within the EMR the basin is 1,200 km from north to south and 800 km wide at the southern boundary. Most of the seabed in the basin lies at depths of between 4,500 and 4,900 m and it gradually shoals to the north where the seabed is 3,500 m and 3,750 m off the northern tip of Fraser

Island and at the entrance to the Cato Trough respectively (Walker 1992; Exon et al. 2005; Boyd et al., 2008).

The floor of the basin is unusual in that the greatest depths are adjacent to the slope off NSW where the seabed is at 4,900-5,000 m and similar depths are reached at the base of the Monawai Ridge which forms the western margin of the southern Lord Howe Rise (Stagg et al., 2002). The maximum depth detected is over 5,100 m in eroded channels along the eastern margin of the basin and in depressions south of Taupo Seamount.

The basin can be divided into two geomorphic areas: a) the abyssal plain which forms the northern and western two-thirds of the seabed and, b) deep ocean floor that forms the eastern third of the seabed. The boundary between the two topographic styles of seabed is sharp (Fig. 3.31). The base of the slope along the eastern margin of the basin is generally a sharp contact and varies in depth between 3,500 m and 4,500 m. This absence of a rise and submarine fans is unusual for a basin of this age and reflects the lack of sediment input and erosion by bottom currents, particularly in the Neogene.

In the southern part of the EMR the basin is 800 km wide extending from the base of the slope off southern NSW to the base of the Monawai Ridge. North of this the basin narrows to 400 km at 33°S (Newcastle to Dampier Ridge) because both margins step basinward. It narrows again to 250 km at 28°S off Tweed Heads and it then widens to 350 km at 26°S off Noosa. Seamounts of all sizes and shapes rise from the seabed in both the abyssal plain and deep ocean floor areas (Fig. 3.1). Edgeworth David (David, 1932) was the first to draw a profile across the Tasman Sea. His figure is based on the west to east track of SS Britannia off Tweed Heads and identifies Globigerina (foraminifer) ooze adjacent to the Australian margin, then seamounts, red clay and manganese. He described one seamount as rising 4,880 m from the seabed to within 910 m of the surface along with a dredge sample of volcanic rock from its side (28° 42'S; 155° 37'W). He named it Britannia Seamount, which is now part of the Tasmantid Seamount chain (Standard, 1961) which rises from the seabed along a north-south line from the Cato Trough. General descriptions of the main features of the Tasman Sea floor are also given by Standard (1961) and Conolly (1968 and 1969).

The abyssal plain occupies the western part of the basin where it extends from the base of the slope to a sharp boundary with abyssal hills of the deep ocean floor (Figs. 3.31 and 3.32). Features on the abyssal plain are isolated basement outcrops forming seamounts and ridges, minor fans/debris deposits/channels at the base of slope on the western margin, and elongate drift mounds on the plain itself. The plain is readily identifiable on seismic profiles because it is near horizontal, smooth and internally consists of parallel reflectors. This reflects the principal mode of deposition by turbidity currents. Along its southeastern margin the plain is up to 50 m lower than the sediment draped abyssal hill region (Fig. 3.31). In the south the plain extends more than half way across the basin and the surface of the plain rises gradually to the north. It is 200 to 250 km wide in the southern part of the EMR where it extends eastward to Taupo, Barcoo, and Derwent Hunter Seamounts. It maintains this width northwards to occupy most of the basin north of these seamounts (Conolly, 1969; Jenkins, 1984). The turbidites of the abyssal plain have an abrupt contact with the Kenn Plateau in the northeast (Fig. 3.34).

A characteristic of the Australian margin is the lack of depositional fans at the base of the slope. The only submarine fans identified along the western margin of the basin are very small, less than 30 km wide, and located off the Hunter River (Newcastle), Macleay River (Smoky Cape, NSW) the Clarence River (Yamba, NSW) and Breaksea Spit (Sandy Cape, Fraser Island). Off the Maclay River Packham (1983) described three small fans at the base of slope at 4,000 m and extending to 4,500 m water depth. At 26°S offshore of Noosa Hill (1991) used a seismic profile to defined a slump mass forming positive relief on the seabed in 4,500 m of water and extending nearly 20 km out from the base of the slope. This slump mass is not recent as it has been eroded at the base of slope and is onlapped by turbidites of the abyssal plain on its basinward side.

The eastern margin of the basin adjacent to the Dampier Ridge also lacks fans and is of a similar depth to the conjugate western margin. What appears to be a rise 100 km wide occurs where the seabed shoals into an embayment in the Dampier Ridge between 28° 30′ and 29° 30′S (Figs. 3.1 and 3.23). The seismic profile in Figure 3.23 across this rise, along with a multibeam line further south, show it to have irregular topography on the order of 10's of meters interspersed with flat lying sediments. Small rises occur elsewhere along the base of the Dampier Ridge, and are formed by the accumulation of sediment debris redeposited from the slopes possibly by less subsidence in the underlying crust (Fig. 3.33). The basin is deep (~4,800 m) along the base of the Dampier Ridge where the seamounts are close. There is evidence for erosion by bottom currents in these areas (Baker et al., 1988a). Distinct erosion channels occur at the western foot of the Dampier Ridge where the abyssal seabed between the Dampier Ridge and the Taupo and Barcoo Seamounts is only 50 km wide (Jenkins, 1984). Moats 6-30 km wide and 18-332 m deep also occur at the foot of Recorder Seamount (west side), Britannia Seamount (east side), the Monawai Ridge and an un-named seamount (Appendix C Table 3.9; Fig. 3.33; Jenkins, 1984).

Jenkins (1984, 1992b) has described the depositional drift deposits and current erosion on the floor of the Tasman Basin. A large drift exists 150 km offshore of northern NSW and extends from 33°S for 600 km to the north until it abuts the Britannia Seamount (28°S) with a massive piling-up of sediments (Fig. 3.33). The drift is between 100 and 300 m high and 30 to 50 km wide. At its edges it merges with the abyssal plain. Jenkins (1984) named it the Kennedy Drift and uses internal bedding and erosion scours around small seamounts to conclude that the drift formed between a north flowing Western Boundary Undercurrent and a south-flowing current further east. Jenkins (1984) also defines a smaller sediment drift extending south of Stradbroke Seamount (29.5°S; 155.5°W).

A distinct erosional moat occurs in the abyssal seabed at the base of the slope offshore of Tweed Heads-Stradbroke Island, where the gap between the base of slope and the Queensland and Britannia Seamounts is only 40-80 km (Fig. 3.33). This north flowing current may also explain why the greatest depths on the abyssal plain are adjacent to the base of slope further south. The troughs along the base of slope in the south are smaller, about 20-50 m deep and a few kilometers wide. The shape of the subsurface reflectors suggests non-deposition due to current winnowing formed these troughs (Conolly, 1969; Jenkins, 1984; Colwell et al., 1993). The presence of relatively strong currents on the Tasman Basin seabed has been confirmed by current indicators in seabed photographs of the abyssal plain near the western margin (Jenkins et al., 1986).

At the northern end of the Tasman Basin the abyssal plain shallows to 3,500 m as a depositional rise at the base of slope, but it is not fan shaped (Fig. 3.35). The Capricorn Sea Valley is incised into this deposit (Boyd et al., 2008). It is a prominent feature on the seabed as it curves eastward from the base of the Marion Plateau slope, and runs south-south-east for over 120 km into water depths of 4,700 m. In contrast, at the southern end of the basin the Sydney Canyon exits onto the abyssal plain at the base of the slope (4,896 m) as a broad, 6 km wide channel some 400 m deep and with no evidence of a fan deposit.

Abyssal hills characterise the deep ocean floor in the south eastern part of the basin and are generally shallower (3,500 – 4,500 m) than the abyssal plain but with some depressions down to 5,100 m. The features forming the abyssal hills are broad ridges, hills and swales draped with sediment and with low relief (< 100 m), and hills and ridges of exposed basement with relatively steep relief of 100 to 1,000 m and often with eroded channels at their base.

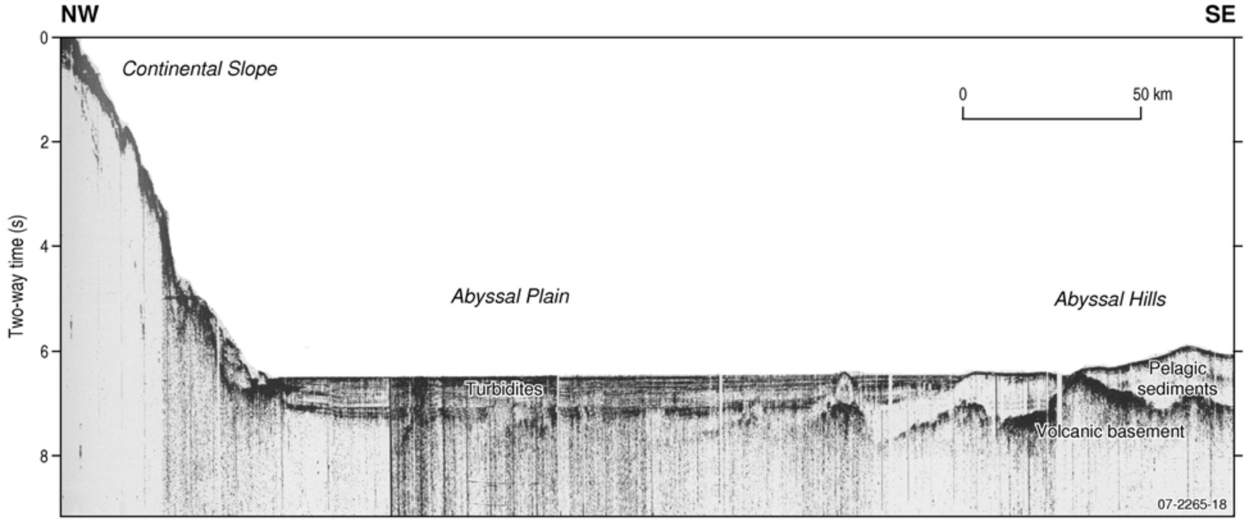


Figure 3.31. Seismic profile to the south-east of Newcastle showing the flat, near horizontal abyssal plain formed by turbidites and the sharp contact with the abyssal hills draped with pelagic sediments of the deep ocean floor region. Basaltic basement outcrops as small seamounts or where sediment is eroded. Location is on Figure 3.1. Eltanin Line 54.

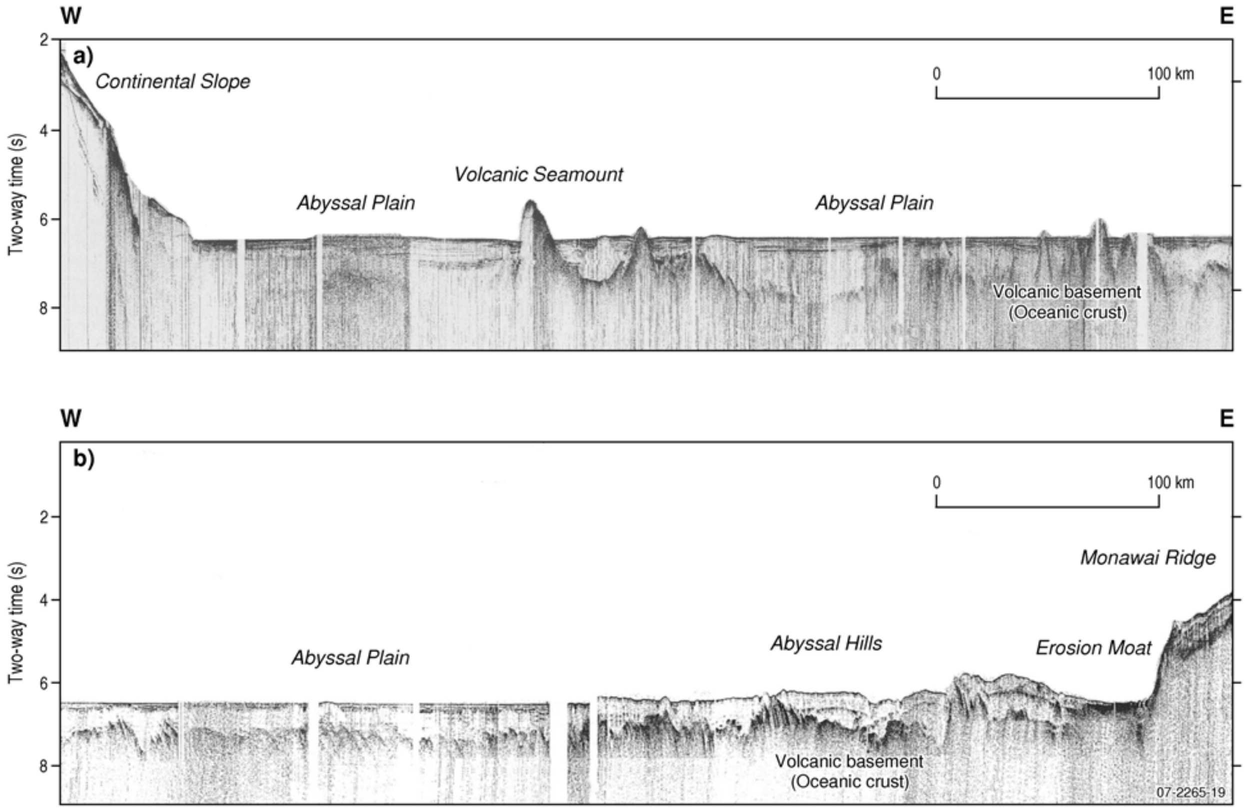


Figure 3.32. Seismic profile from the shelf off Newcastle across the Tasman Basin to the southern Lord Howe Rise. a) The seismic profile to the east of Newcastle shows the flat slightly convex-up abyssal plain. Note the broad moats at the base of the small seamounts of basement rock and b) Pelagic sediments in the eastern part of the basin show draping on underlying

basement topography and erosion adjacent to basement ridges. Note the scarp and the erosion moat at the base of the Monawai Ridge which forms the western margin of the Lord Howe Rise at this location. The Tasman Basin seabed along this transect is mostly in a narrow depth range from 4,800 to 4,900 m. Location is on Figure 3.1. Eltanin Line 47A.

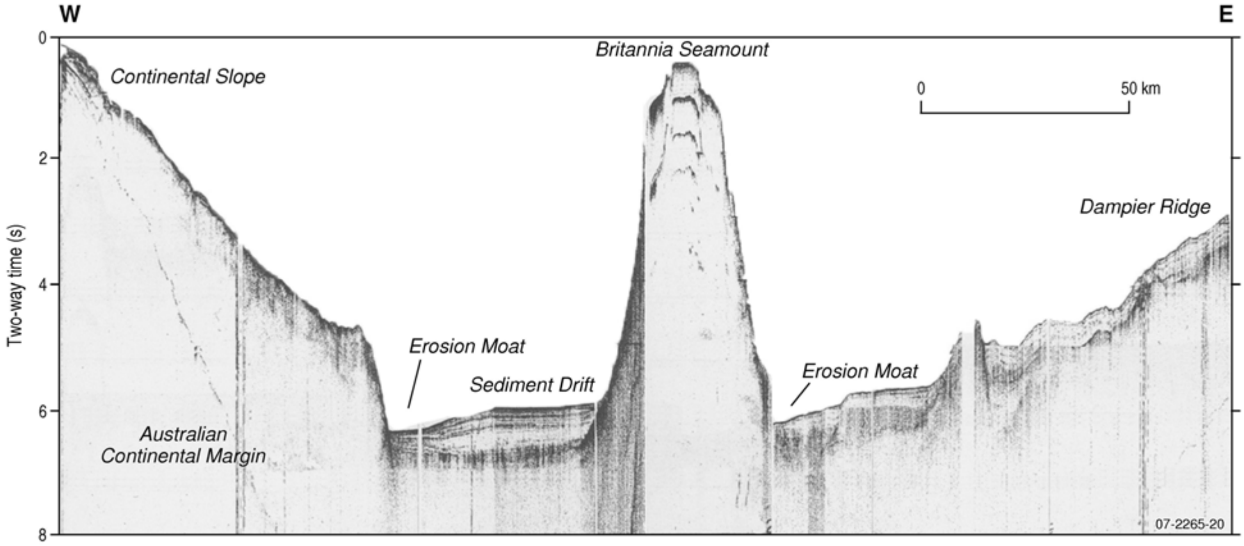


Figure 3.33. Seismic profile in the north Tasman Basin from east of Tweed Heads to the Dampier Ridge passing over Britannia Seamount. Note the wide mid-slope and steep lower slope with erosion moat at its base in the abyssal plain. Deposition by bottom currents has formed a sediment drift and apron at the base of seamount. There is an erosion moat on the east side of Britannia Seamount. Location is on Figure 3.1. Eltanin Line 29.

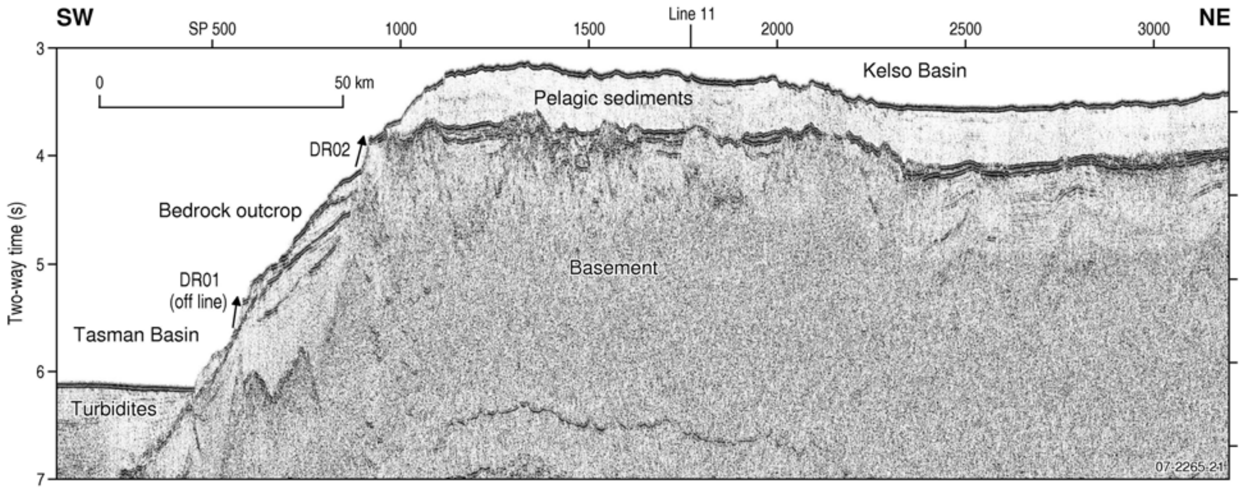


Figure 3.34. Seismic profile showing abyssal plain turbidites in sharp contact with the steep margin of the Kenn Plateau north Tasman Basin. Note the sediment blanket on the plateau. Location is on Figure 3.1. Line 270-7, Exon et al., (2005).

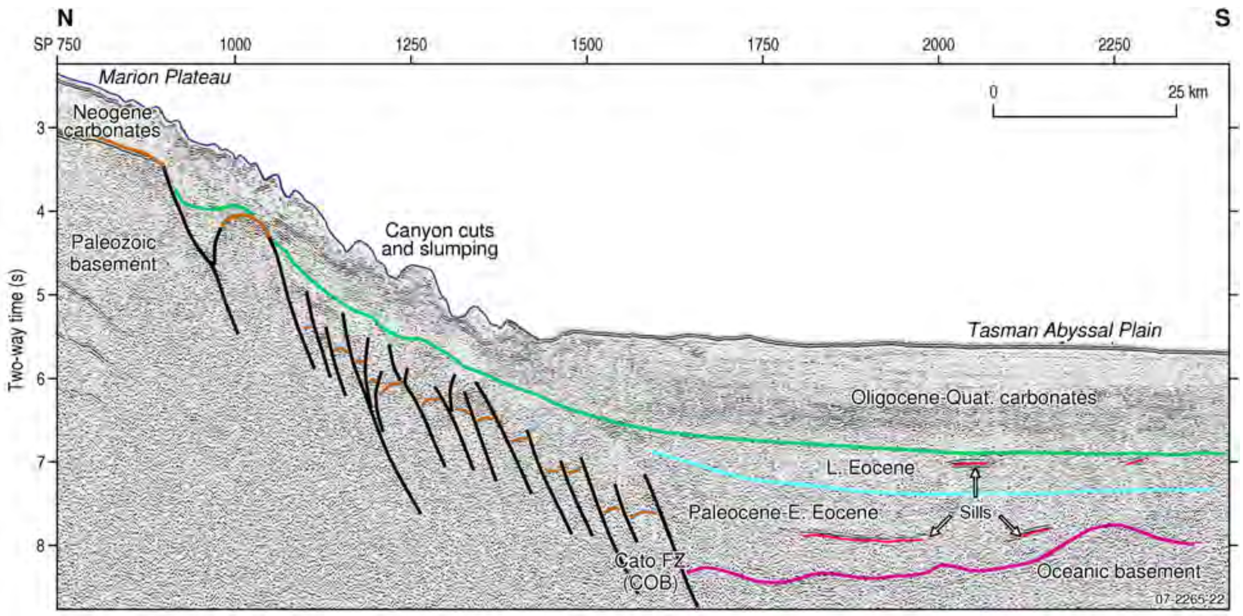


Figure 3.35. Seismic profile showing sediment slumping from the Marion Plateau and relatively shallow turbidites forming the abyssal plain, north Tasman Basin. Note the location of the Cato Fracture Zone separating oceanic crust from continental crust. Location is on Figure 3.1. Line 1. Exon et al., (2005).

3.4.2. Surface Sediments and Rocks

Very few cores or surface samples have been described from the Tasman Basin within the EMR. A list of the bottom photographs and samples described in the literature are given in Appendix C Tables 3.7 and 3.8. In general, calcareous ooze is present where the seabed is above 4,500 m and pelagic brown calcareous 'red' clay at greater depths (Conolly, 1969). The most detailed core analyses are in Eade and van der Linden (1970) and Hubble et al. (1987). They analysed 6 cores along a transect from the lower slope (one core) off Port Macquarie across the abyssal plain/deep ocean floor (four cores) to the flank of the Dampier Ridge (one core).

The composition and texture of the sediments deposited in the Tasman Basin is controlled by their sources, the water depth and the process of deposition. Terrigenous sands, silts and clays are supplied from the Australian continent by wind and currents and similar sized particles come from the ocean, mostly as biogenic debris from planktonic but with some benthonic particles. Their relative proportions depend on their supply/productivity and on water depth. Martinez (1994b) determined the present day lysocline for the Tasman Sea to be at ~3,600 m in the Tasman abyssal plain and ~3,100 m in the New Caledonia Basin to the east. Below the lysocline the calcium carbonate particles, particularly of foraminifers, start dissolving extensively until the calcite compensation depth (CCD) is reached. The CCD is where the rate of carbonate sedimentation is equal to the rate of solution and hence no pelagic carbonate survives on the seafloor below the CCD. Most of the Tasman Sea basin is just above the CCD as the cores analysed contain pelagic carbonate. The processes of deposition in this environment are pelagic settling, turbidity currents, debris flows and reworking by bottom currents. The near-surface strata beneath the plain are younger than the pelagic sediments draping the abyssal hills because they clearly onlap the hills on seismic profiles (Fig. 3.31).

The five cores from the Tasman Basin analysed by Eade and van der Linden (1970) come from water depths of 4,565 and 4,654 m on the abyssal plain, 4,689 and 4,838 m in the abyssal hills and 4,283 m on the flank of the Dampier Ridge. They range in length from 1.6 to 5.4 m. All have a carbonate content in the surface sediment of between 40 and 55%. The sediment type is dominantly light olive grey calcareous mud containing foraminifers in various states of preservation and coccoliths. The core at 4,565 m is only 60 km from the slope and has distinct layers in the upper 55 cm containing shallow water benthic foraminifers. These layers are interpreted as deposits from turbidity currents. Carbonate cycles are present down-core in four of the cores and represent Pleistocene glacial/interglacial cycles. The exception is the deepest core which is 60 km SW of the Derwent-Hunter Seamount and may be a turbidite from that source. A core in 4,747 m of water between Brisbane Seamount and the northern Dampier Ridge had a 15 cm thick graded bed of foraminiferal sand at the seabed with other graded beds below (Jenkins et al., 1986). Photos of the seabed between Derwent-Hunter Seamount and the Dampier Ridge (water depth 4,600 m) show rock debris and ripples as clear evidence of bottom water flow (Baker et al., 1988a and b).

The presence of debris flow deposits at the base of the NSW slope are confirmed by bottom photographs of pebbles and blocks of rock in bioturbated muds in 4,820 m of water adjacent to the slope (Jenkins et al., 1986). These are relatively recent as they are at the surface. Other photos in Harris et al. (1987) show evidence of currents on the abyssal plain near the base of slope off Wollongong (water depth 4,860 m), off Sydney (water depth 4,800 m) and off Port Macquarie (water depth 4,550 m).

The only evidence for present day activity of turbidity currents in the basin is provided by Boyd et al. (2008) who sampled quartz sand along with estuarine and shelf carbonate detritus in the floor of the Capricorn Sea Valley in 3,920 m water depth off Fraser Island. Luminescence dating of the quartz returned modern ages indicating this valley is an active conduit for gravity driven transport processes such as grain flows or turbidity currents.

Debris aprons occur around some of the seamounts and are likely to consist of fine to coarse volcanic sediment and lithified carbonate from their summits. Seabedphotographs in the eroded moat at the base of Gascoyne Seamount showed it to contain gravel and larger rocks as a lag deposits (Appendix C Table 3.10; Jenkins et al., 1986).

A field of manganese nodules has been sampled just south of the EMR by Glasby et al. (1986). Their photographs show abundant manganese nodules covering the seafloor at water depths of 4,408 to 4,717 m in the abyssal hills adjacent to the abyssal plain (Appendix C Table 7). The nodules are coated with a light dusting of calcareous brown clay. Most of the nodules sampled were discoidal in shape and 6 to 10 cm in diameter. Chemically the nodules have a relatively high Mn/Fe ratio of 2.5 along with moderate concentrations of nickel (av. 0.83%) and copper (av. 0.40%). Exon et al. (1980) also found nodules associated with greenish-grey calcareous clay further south and in shallower water (4,300 m). They contained more iron and less manganese, nickel and copper and 0.06% cobalt. Presumably the sedimentation rates are low in this region because the seabed is below the lysocline and this allows the growth of manganese nodules. This nodule field probably extends north into the EMR but its extent is unknown. Cochran and Osmond (1976) calculated a relatively high sedimentation rate of 2.1 cm ka-1 for a core containing 45% carbonate near this area. They concluded that the sedimentation rate in the area varies and is strongly affected by winnowing and deposition by bottom currents and that low accumulation rates occur on the crest of ridges and rises and faster rates occur on the flanks.

3.5. SEAMOUNT CHAINS OF THE TASMAN SEA

Two parallel seamount chains form prominent N-S features on the seafloor in the EMR. They were constructed as the Australian plate moved north over two hot-spots in the underlying mantle. The youngest seamounts are in the south. The western chain extends south from Cato Trough and follows the abyssal seabed of Tasman Basin. The other is along the western margin of the LHR. The chains are 300-400 km apart and the volcanic edifices that come to the surface or within a few hundred meters of the surface are named in Figures 3.1 and 3.11. The major seamounts are listed in Table 3.2 along with the depth of their summits below sea level and the depth of the surrounding seabed. Many smaller seamounts also occur along these two chains. Once extinct these volcanoes subside and if they were above sea level their tops are eroded flat by waves. Reefs have grown on some as they subsided to form limestone caps. Submerged seamounts that are flat-topped are called guyots. Appendix C Tables 10 and 11 list the bottom photographs and samples in the literature from the two seamount chains.

Table 3.2. Seamounts of the Tasmantid and Lord Howe Chains within the EMR. Data from Slater and Goodwin, (1973), McDougal and Duncan (1988) and Royal Australian Hydrographic Office.

Seamount	**Depth to summit m**	**Depth of base m**
Unnamed 30 km SW of Taupo	2,549	5,000
Taupo	120	~4,800
Barcoo	251	~4,800
Derwent-Hunter (two platforms)	280	~4,800
Unnamed 100 km north of D-H	2,375	~4,800
Stradbroke	800	~4,800
Unnamed 25 km S of Britannia	3,370	~4,800
Britannia (two platforms)	411 and 397	~4,800
Queensland (two platforms)	410 and 390	~4,800
Brisbane	1,458	~4,800
Moreton	753	4,400-4,600
Unnamed 50 km NE of Moreton	3,509	4,400-4,600
Unnamed 40 km N of Moreton	2,827	4,400-4,600
Recorder (two platforms)	410 and 1,170	4,400-4,600
Fraser	361	4,000-4,100

Unnamed 40 km N of Fraser	2,445	4,000-4,100
Cato Island	0	1,500
Bird Island/Wreck Reef	0	3,000
Kenn Reef	0	3,000
Frederick Reef	0	3,000
Balls Pyramid Is.	+552	~2,000
Lord Howe Is.	+864	~2,000
Elizabeth Reef	0	~3,000
Middleton Reef	0	~3,000
Unnamed 25 km N of Middleton	303	~3,000
Unnamed 30 km S of Gifford	330	~3,000
Gifford Tablemount	261	~3,000

3.5.1. Geomorphology

3.5.1.1. Tasmantid Seamount Chain

The Tasmantid Seamount chain extends northward onto the eastern margin of the Kenn Plateau and into the Cato Trough (Exon et al., 2005). Cato Island, Bird Island/Wreck Reef, Kenn Reef and Frederick Reef are all volcanic seamounts capped by limestone reefs of varying thickness (Fig. 3.38). Cato Island and Kenn Reef rise from depths of less than 2,000 m because they are on the continental crust of Kenn Plateau (Exon et al., 2005, 2006b). The four oldest seamounts in the north of the chain have living reefs on the limestone that caps the underlying volcano (Fig. 3.9). They have grown to maintain themselves at sea level. Further south Moreton, Brisbane and Stradbroke seamounts remain as volcanic peaks as they have never reached sea level whereas the other named seamounts have all subsided below sea level and been planated. A detailed multibeam survey around Cato Island by Exon et al. (2005) revealed numerous small volcanic cones and larger submerged limestone platforms (Figs. 3.36 and 3.38).

Taupo Seamount is the largest in the EMR and is 60 km in diameter at its base rising from depths of 4,800 m to a flat top only ~120 m below sea level. This shallow platform with relief of less than 10 m is approximately 40 km north to south and up to 15 km wide (McDougall and Duncan, 1988). Barcoo also rises from over 4,800 m to within 300 m of the surface. Its platform is 20 km x 6 km. Derwent-Hunter has an oval-shaped platform 30 by 20 km at about 300 m water depth. Britannia and Queensland both have remarkably flat summits at a depth of approximately 400 m and Recorder has an uneven top at about 450 m. Britannia and Recorder both have bases 43 km across. Fraser rises from a base 20 km across in 4,000 m of water to a small circular platform 4 km across at a depth of 361 m (Exon et al., 2005). Some of the seamounts, for example Fraser (Fig. 3.36), are near circular in plan while others are a complex of several volcanoes elongate parallel to the trend of the volcanic chain.

Fraser Seamount has distinct terraces at 1,030 m and 1,350 m which represent former sea levels. There are many smaller un-named and unsurveyed seamounts along this chain as well as subsidiary cones on the flanks of larger edifices.

The slopes on the side of the seamounts are commonly in the range of 10-20^0 and locally can be much steeper or form a flat terrace cut at sea level (Fig. 3.36). These slopes consist of rugged rock outcrop and boulders and blocks with only a relatively thin drape of sediment cover. The seamounts shed sediment to the adjacent seabed to form an apron at their base (Fig. 3.37). In some cases this apron is removed by bottom currents to form a moat.

McDougall and Duncan (1988) have sampled basalt from Taupo Bank to Queensland Guyot and obtained ages of 6.4 to 24 Ma (late Miocene to early Miocene). The ages of the volcanoes get progressively older to the north, confirming that they were formed as the crust moved over a hot-spot in the mantle. The northern seamounts were probably formed in the Oligocene (Exon et al., 2005). Assuming the hot spot is fixed then the average rate of northward movement of the Australian plate is 6.7 cm year^{-1} over that time interval.

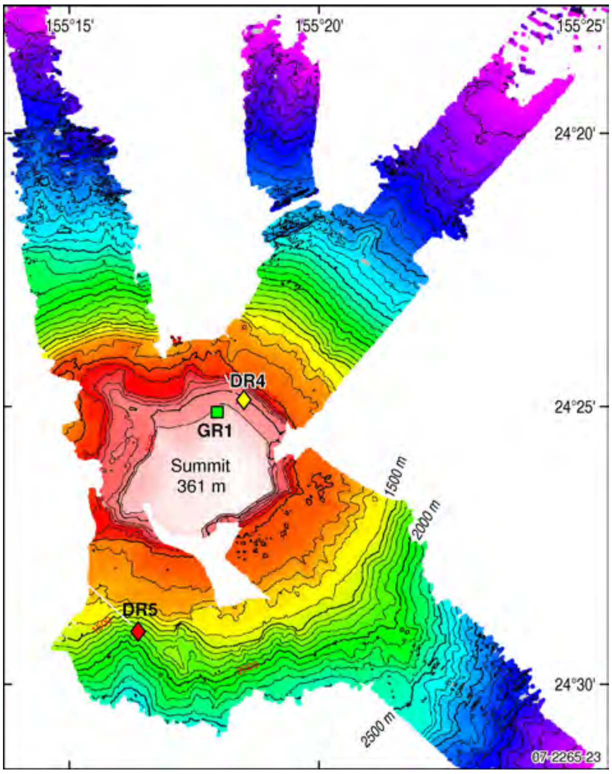


Figure 3.36. Multibeam bathymetric map of Fraser Seamount showing it to be a guyot with a flat summit and steep sides with two prominent benches at 1,000-1,100 and 1,300-1,400m. Location on Figure 3.1. Exon et al., (2005).

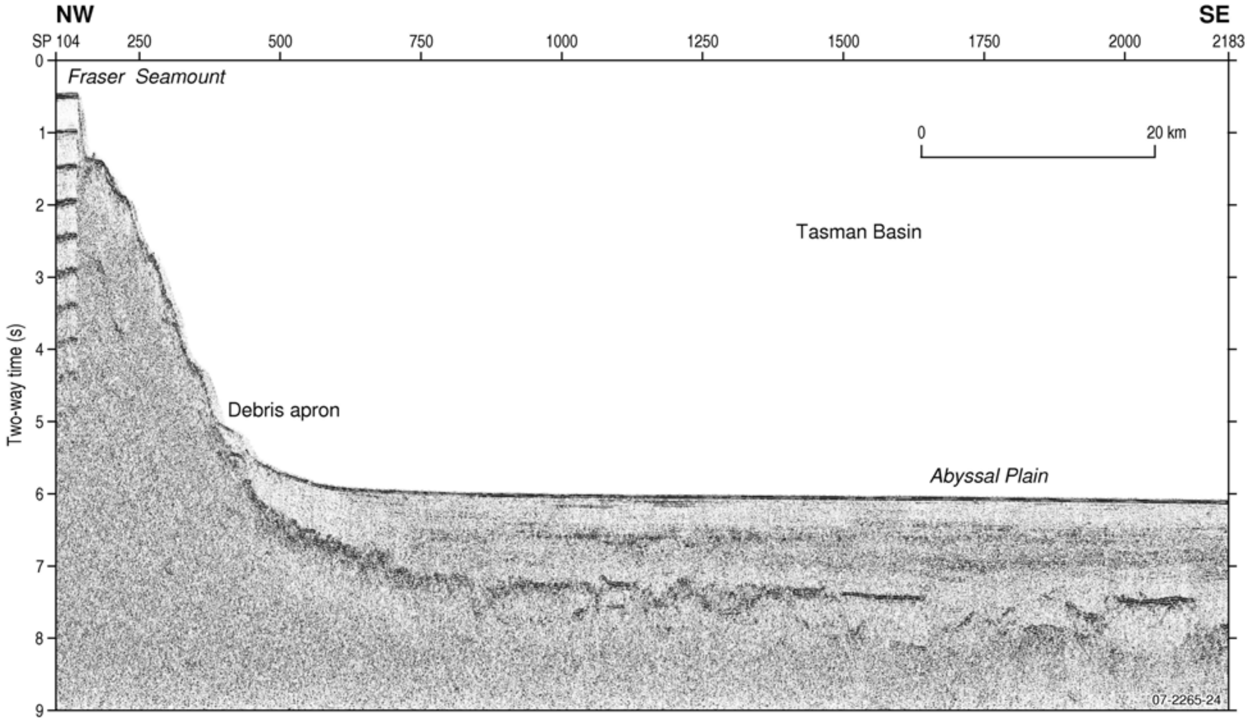


Figure 3.37. Seismic profile of Fraser Seamount showing an apron of slump material at its base merging with the Tasman abyssal plain. Location on Figure 3.1. Exon *et al.*, (2005).

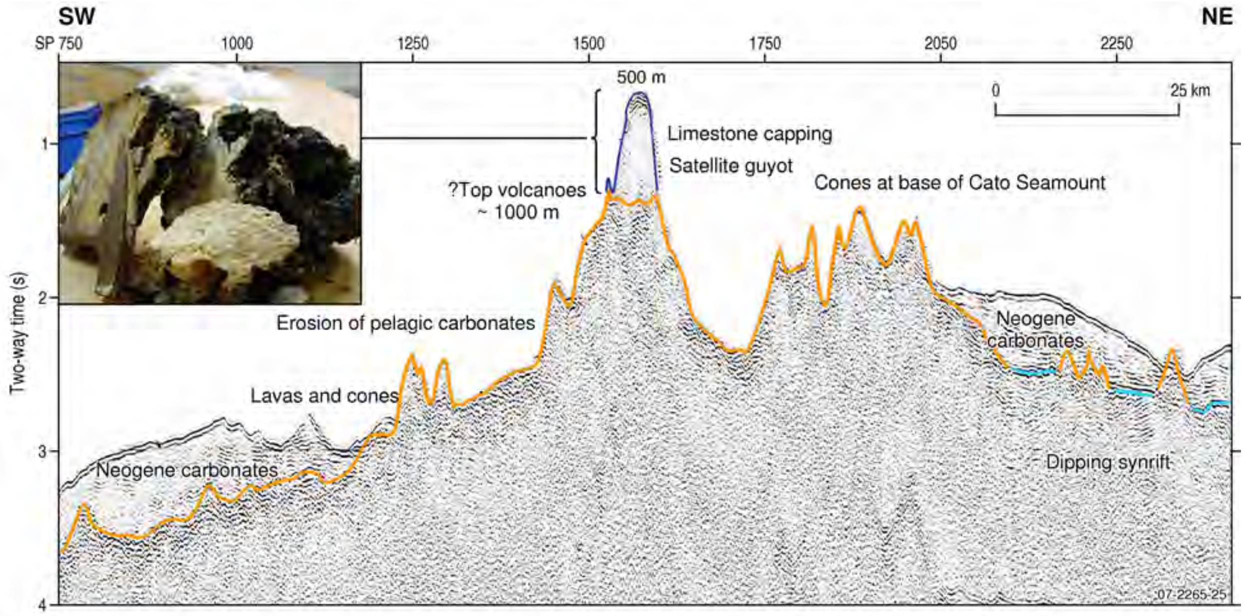


Figure 3.38. Seismic profile across Cato Seamount showing erosion, exposure of bedrock, sediment drifts and carbonate reef as a cap on the volcanic peak. Location on Figure 3.1. Line 3, Exon *et al*., (2005).

3.5.1.2. Lord Howe Seamount Chain (Balls Pyramid, Lord Howe Island, Elizabeth Reef, Middleton Reef, Gifford Tablemount.)

The Lord Howe Seamounts form a linear N-S chain along the western flank of the LHR from Balls Pyramid and Lord Howe Island in the south to Elizabeth Reef, Middleton Reef and Gifford Guyot in the north of the EMR (Fig. 3.39). This chain is parallel to the Tasmantid Chain some 300 km to the west. It is believed that this chain is also formed by the northward movement of the Australian Plate over a 'hot spot' in the mantle. The older seamounts to the north have subsided below sea level allowing reefs to develop limestone platforms on their summits. Gifford Guyot near the northern boundary of the EMR is a 2,000 m high guyot that comes to within ~300 m of the surface and is capped by a drowned limestone platform (Slater and Goodwin, 1973). Other smaller and un-named seamounts occur both along this chain and adjacent to it in the Lord Howe Basin. South of Balls Pyramid seamounts continue along the margin of the LHR, the largest being Flinders Seamount on the Monawai Ridge.

Within the EMR the largest of the seamounts in this chain is the Lord Howe Island/Balls Pyramid volcanic edifice. It is the only one whose age has been determined. It is a basaltic volcano built between 6.9 and 6.4 Ma ago (McDougall et al., 1981). Its base is elongate NW-SE some 40 km wide by 80 km long. It rises steeply from water depths of over 3,000 m on its western and southern sides. Its eastern side merges with the Lord Howe Rise at water depths of less than 2,000 m. Game (1970) describes at least three major eruptive episodes probably starting in the mid-Tertiary and dated the youngest at 7.7 Ma.

There is a broad shelf around both Lord Howe Island and Balls Pyramid due to marine planation (Woodroffe et al., 2005). They present a bathymetric chart around Lord Howe showing that the shelf is nearly square in shape (20 x 20 km) and oriented NW-SE with the shelf break between 40 and 60 m water depth. Most of the coral reef is on the mid-shelf. Balls Pyramid shelf has no reef and is 16 x 8 km and 4 km southeast of the Lord Howe shelf and linked to it by a 500 m deep sill.

Middleton Reef and Elizabeth Reef are located on the saddle joining the Dampier Ridge with the Lord Howe Rise, and separating the Middleton Basin in the north from the Lord Howe Basin in the south. They have roughly circular bases with diameters of ~40 km and rise steeply from 2,500 m water depth. They have also undergone wave erosion as they subsided below sea level resulting in a planated surface for coral growth to form oval shaped rim reefs enclosing a lagoon. Elizabeth Reef is an oval atoll oriented NW-SE and is 10.7 km long by 6.2 km wide (Kennedy and Woodroffe, 2004). It is slightly larger than Middleton Reef which is 9.3 x 5.7 km and oriented NE-SW. The lagoons have a maximum depth of 30 m, but are mostly infilled with sediment and on average are less than 5 m deep (Kennedy and Woodroffe, 2004). On the seaward side of the reef there is a steep drop-off at 30-40 m water depth.

The Lord Howe chain was formed in a similar way to the Tasmantids but only Lord Howe Island has been dated. Exon et al. (2004b) have used sediment associated with basalt from seamounts on LHR north of the EMR to indicate a Miocene age of the eruption. Similar small seamounts are quite common on the seabed of the LHR and the Dampier Ridge and may be the result of more widespread volcanism during the Miocene than that dated from the two dominant chains.

3.5.2. Surface Sediments and Rocks

Slater and Goodwin (1973) dredged four of the seamounts and recovered unconsolidated shelly sands along with limestone, phosphorite and basalt cobbles. Limestone of late Pliocene to early Pleistocene age along with basalt pebbles was recovered from Taupo, Barcoo and Derwent-Hunter. Phosphorite and limestone was recovered from Gifford and phosphorite was also found on Derwent-Hunter. Quilty (1993) identified Miocene shallow-marine limestone from Derwent-Hunter and Barcoo seamounts. McDougall and Duncan (1988) sampled basalt, including well-rounded cobbles, from the flanks of Taupo, Derwent-Hunter, Britannia and Queensland Seamounts in water depths of 500 to 1900 m.

The shelf around Lord Howe Island has well established coral reef. This is the southernmost occurrence of coral reefs in the Tasman Sea. There is a submerged fossil reef on the mid-shelf around Lord Howe Island but Woodroffe et al., (2005) could find no evidence of subsidence since the last interglacial (c.a. 120 ka). Coralline algae are the dominant sediment component with coral and algal rhodoltihs also common (Kennedy et al., 2002). Gravelly muds occur within the lagoon.

Kennedy and Woodroffe (2004) collected sediments from the seabed from around Middleton Reef down to a water depth of 368 m and from within the lagoon of Middleton and Elizabeth Reefs. Gravelly sands and sands dominate in the lagoons with some mud in the deepest parts. The sediment is composed of coral and coralline algae with lesser amounts of Halimeda, molluscs and foraminifera.

3.6. PLATEAUS OF THE TASMAN SEA

3.6.1. Geomorphology

3.6.1.1. Southern Kenn Plateau, Dampier Ridge and Lord Howe Rise

The eastern margin of the Tasman Sea Basin within the EMR is formed by the submerged margins of three tectonic blocks of continental crust clearly outlined by the 3000 m isobath. From south to north they are: the Monawai Ridge (southern LHR), Dampier Ridge and southern Kenn Plateau. The EMR extends over the Dampier Ridge, northern Monawai Ridge and onto the Kenn and Lord Howe Plateaus (Figs. 3.1 and 3.39).

In the south the Monawai Ridge is a linear NW trending feature 100 km wide on the eastern flank of the Lord Howe Rise extending into the planning area for 160 km. On its western side it rises from the abyssal floor depths of ~5,000 m of the Tasman Basin to depths of 2,000-3,000 m (Fig. 3.32). This steep scarp is up to 1,500 m high. Flinders Seamount is a volcanic intrusion along this scarp, and rises from water depths of ~5,000 m in the Tasman basin to water depths of 1,740 m at its summit giving a relief of over 3,000 m (Fig. 3.40c, Stagg et al., 2002). The seabed on its eastern side is at depths of 2,300 m. Between the Monawai Ridge and the Lord Howe Rise proper is the broad 100 km wide Monawai Sea Valley (Alcock et al., 2006).

The Dampier Ridge extends south from the southern Kenn Plateau for 900 km and is approximately 80 km wide (Fig. 3.39). In detail it can be divided into four plateaus in 2,000-3,400 m water depth representing the four continental fragments of Gaina et al., (1998b). They are joined by narrow saddles (20-40 km wide) in 3,500-3,600 m water depth. The southern plateau extends 200 km NW from the northern end of the Monawai Ridge. Its western margin is steep and is only 40-60 km from

Taupo and Barcoo Seamounts. To its north is the smallest of the plateaus 100 km N-S by 80 km. Further north are two more plateaus each 200 km long, and 100 km and 150 km wide respectively, with irregular tops. The western margin of the Dampier Ridge is generally steep, with scarps of 1,000 to 1,500 m common reflecting its plate tectonic origin as transform faults (Gaina et al., 1998b). Gaina et al., (1998b) also identified intrusions along this margin forming seamounts. Wherever the margin is surveyed it shows small canyons eroding the slope and scarps along slope caused by slumping or faulting.

Between the Dampier Ridge and the LHR lies a depression that is divided into two basins, the Lord Howe Basin in the south and the Middleton Basin in the north (Figs. 3.39 and 3.40). The Lord Howe Basin is between 4,000-4,200 m deep, 300 km N-S and 100 km wide with Lord Howe Island and Balls Pyramid on its eastern margin. At the northern end of the basin the depression is occupied by three large seamounts: Elizabeth Reef, Middleton Reef and an un-named seamount that does not reach the surface. North of these seamounts the depression deepens to 3,500 m to form the Middleton Basin (200 km N-S and 100 km E-W). Gifford Guyot is at the northern end of this basin. Both basins have a relatively flat seabed, and small canyons occur on the slopes around the margins of the basins.

The EMR extends approximately two-thirds (400 km) of the way across the LHR and extends N-S for over 800 km. In the southern area the plateau is less than 1,000 m water depth while to the north it is mostly 1,200-1,500 m. In general the seabed on the plateau is smooth because it is draped by thick sediment (Fig. 3.40, Willcox and Sayers, 2002) and it gradually descends to the west into the basins. Exceptions to this do occur, particularly in the south where the margin with the Lord Howe Basin can be steep and rugged (Fig. 3.40b, Willcox and Sayers, 2002). Along this slope are small seamounts protruding though the sediment cover, and small scale roughness on the seabed indicates erosion and mass movement of sediment. In the northeast of the EMR the LHR is cut by the NW trending Vening-Meinsez Fracture Zone resulting in linear scarps, rough topography and seamounts (Fig. 3.41). Erosion by bottom currents has created moats in the sediments at the base of these ridges (Fig. 3.42).

The relationship of the seamounts on the LHR and Dampier Ridge to the Lord Howe seamount chain is unknown. Willcox and Sayers (2002) present seismic images showing numerous igneous intrusions beneath the eastern LHR that dome the seabed sediments and in some cases reach the seabed to form volcanic seamounts, presumably composed of basalt. A relatively large seamount protrudes from the LHR where it is less than 1,000 m deep in the south of the EMR. More recently, a seismic survey on the northern LHR within the EMR discovered numerous intrusions that have domed the seafloor sediments to form broad hills 4-10 km across with relief of 50-100 m (Kroh et al., 2007). Some of these have also reached the seafloor to form steep sided volcanic pinnacles, often with erosional moats 1-2 km wide and ~50 m deep in the sediments at their base. Kroh et al. (2007) suggest that some of this volcanism is very recent because it intrudes young sediment and disrupted sediment above other subsurface intrusions may create pathways for hydrothermal fluids to reach the seafloor.

The northern Tasman Basin is flanked on its east by the southern part of the Kenn Plateau which extends from the Dampier Ridge north to the Bampton Trough. Immediately north of the Dampier Ridge on the Kenn Plateau is the Kelso Rise, a relatively small 100 x 100 km area of seabed at 1,000-1,200 m water depth.

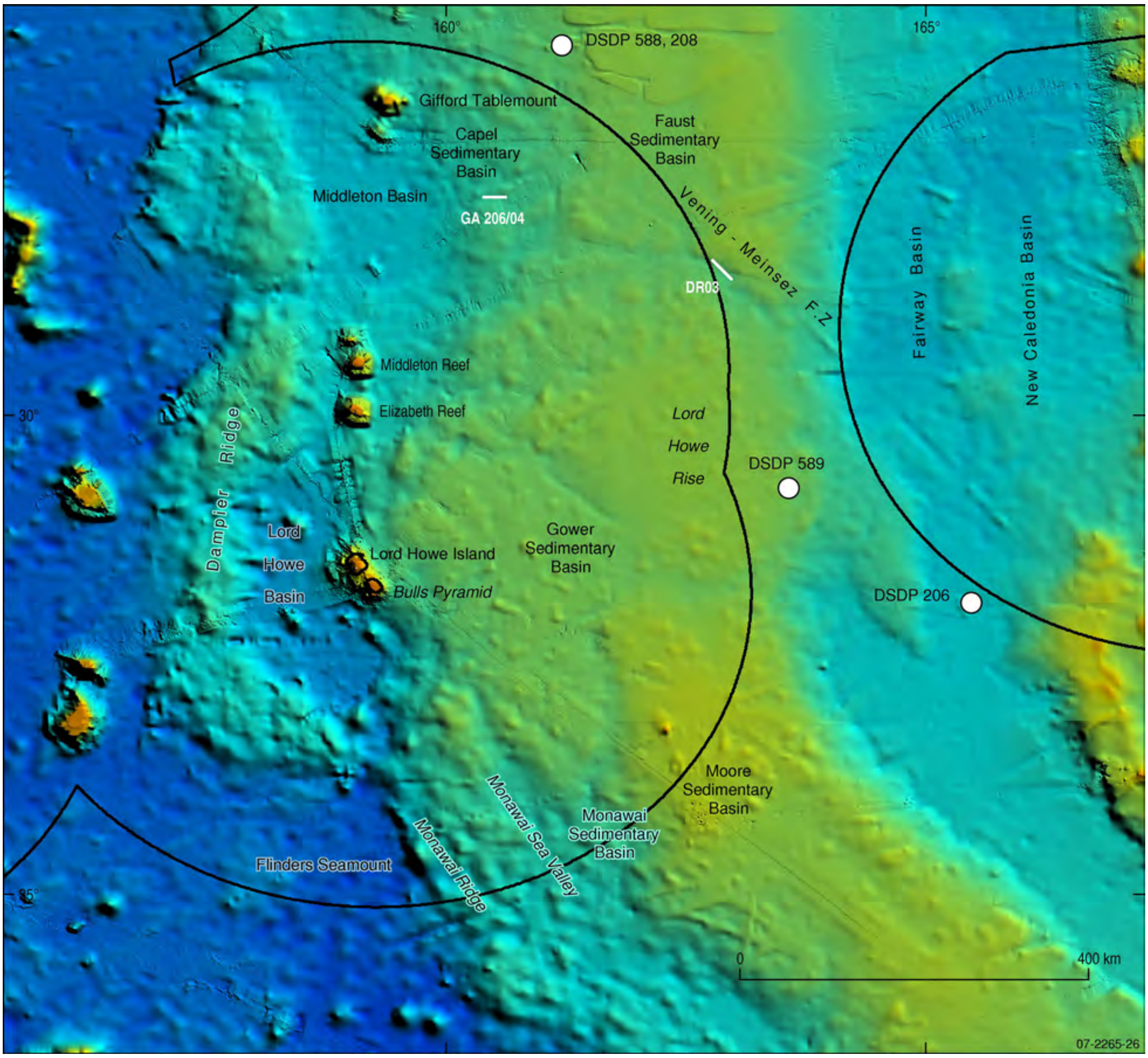


Figure 3.39. False colour image showing bathymetric features including Dampier Ridge, Lord Howe Rise, Middleton Basin, Lord Howe Basin,Lord Howe seamount chain, Monawai Ridge and Flinders Seamount. Location of other figures is marked by a white line. Black line is the EMR boundary. DSDP/ODP drill sites are shown.

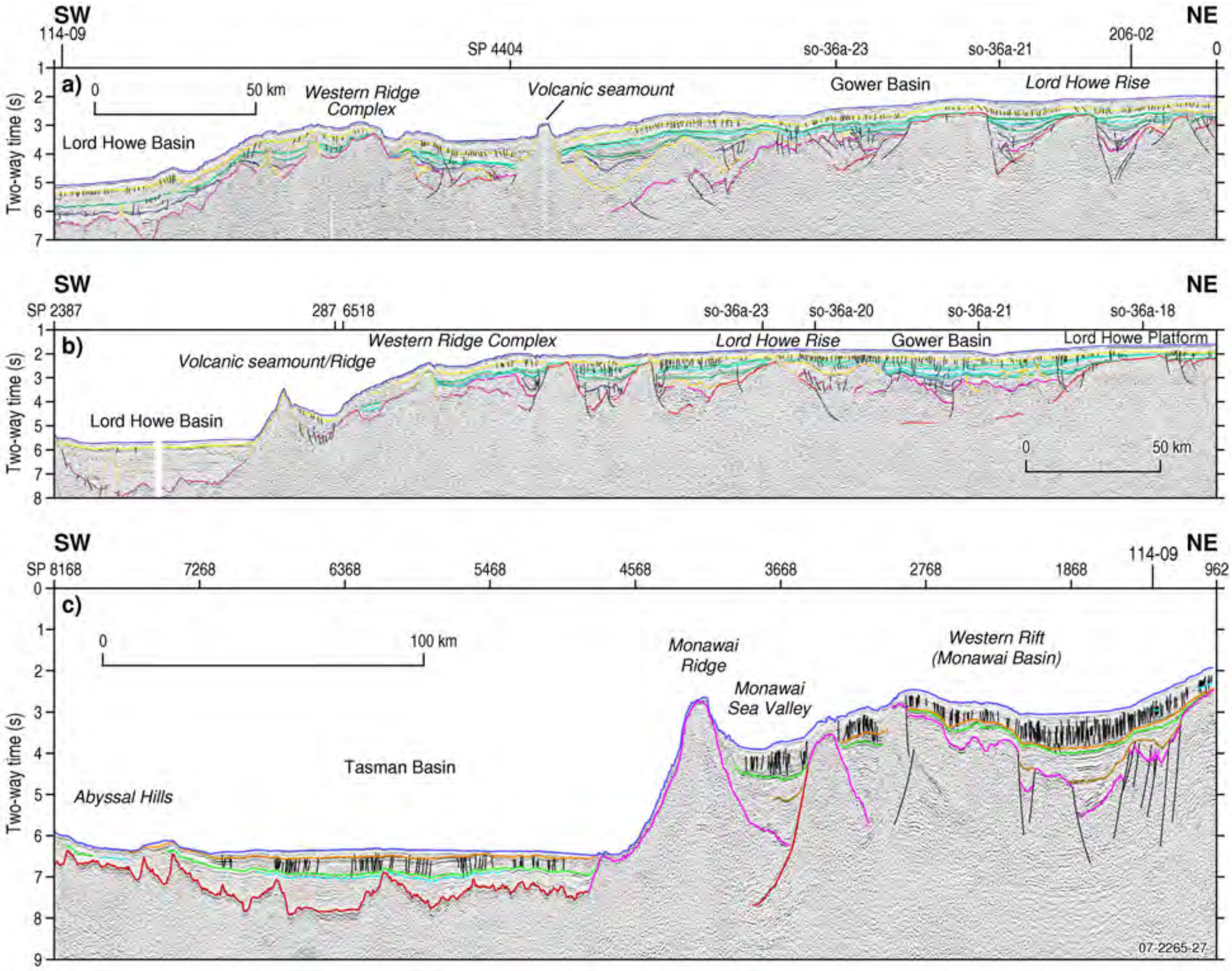


Figure 3.40. a) Seismic profile across the northern Lord Howe Rise showing more topographic relief and volcanic seamounts along the western margin. Line SO30A-06/06AWillcox and Sayers (2002), b) Seismic profile across Lord Howe Basin and central Lord Howe Rise showing abyssal plain in the basin and steep volcanic western flank of the rise. Line 15/15A. Willcox and Sayers (2002) and c) Seismic line from the Tasman Basin across the Monawai Ridge and southern Lord Howe Rise. Note the steep volcanic ridge forming the margin of the plateau and the topographic relief on the rise mimicking the underlying basement. Line LHRNR-E. Stagg et al., (2002).

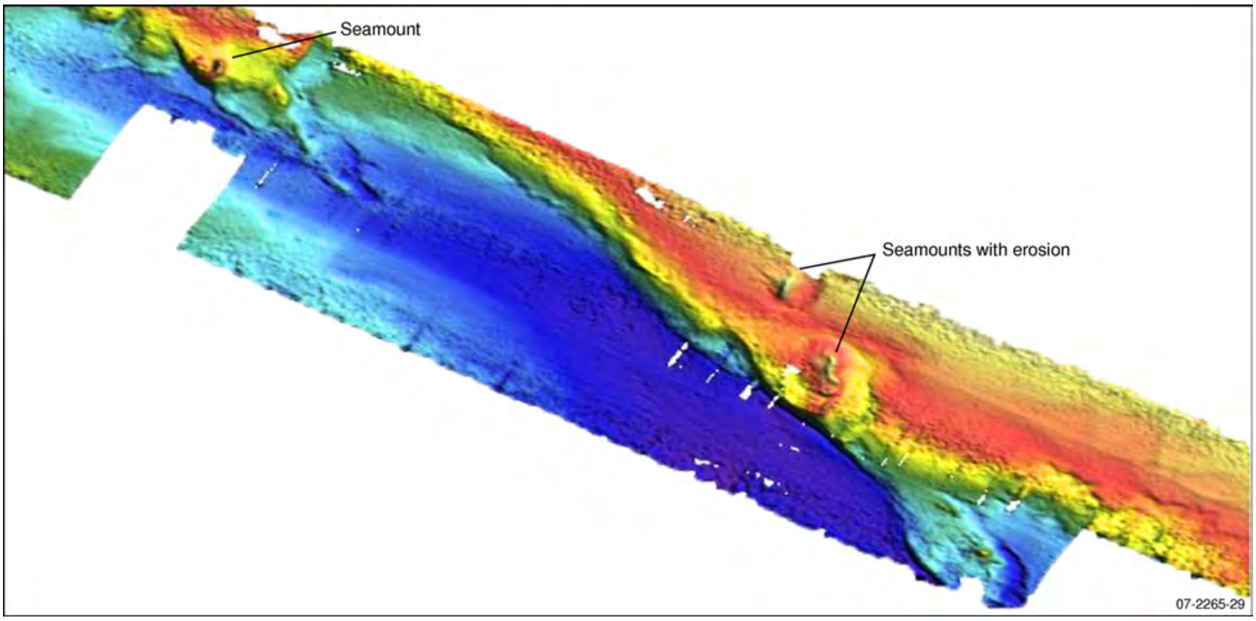


Figure 3.41. Multibeam bathymetry showing a series of fault scarps of the Vening-Meinesz Fracture Zone where it crosses the central part of the Lord Howe Rise. Located on the margin of the EMR but continues to the NW into the EMR. Volcanic seamounts also outcrop. Location DR08 in Figure 3.39. Colwell et al., (2006).

3.6.2. Surface Sediments and Rocks

Our direct knowledge of the composition and stratigraphy of the Lord Howe Rise and Dampier Ridge is from limited sampling. Appendix C Tables 3.12 and 3.13 list the location of samples. The most relevant are:

- Deep Sea Drilling Project (DSDP) and Ocean Drilling Project (ODP) Sites 207, 208, 588, 589, 590 on LHR (Burns et al., 1973, Kennett et al.,1974 and 1986)
- Dredge samples (Launay et al., 1976; Willcox et al., 1981; Roeser et al., 1985; Herzer et al., 1997; Colwell et al., 2006)
- Continental rocks (metamorphosed granite) of Permian age was dredged from the Dampier Ridge (McDougall et al., 1994).

Based on seismic evidence Willcox et al., (2001) agree that the southern Dampier Ridge is continental crust but suggest that the more rugged northern part is volcanic, perhaps of the same age or even younger than the Tasmantid Seamounts. A dredge sample from a scarp on the west flank of the Dampier Ridge in water depths of 2,400-2,700 m (BMR seismic 15/058) contained 20 cm thick Fe/Mn crusts on granite, gabbro and feldspathic sandstone (Roeser et al., 1985, Bolton et al., 1990). Analyses of the manganese crusts showed relatively low Mn/Fe ratios of 0.48-0.91 and low Ni, Cu and Co. Dredges samples from the Vening-Meinesz Fracture Zone just outside the Planning Area have included a variety of continental rocks including sandstone, conglomerate and limestone along with basaltic volcanic rock (Fig. 3.41; Roeser et al., 1985, Colwell et al., 2006). The rocks were covered with a thick manganese crust and a community of benthic organisms used this outcrop as substrate (Fig. 3.42). The biological samples included: Gorgonians, black coral, soft coral, crinoids, bryozoans, bivalves, gastropods, silica sponges, brittle stars, ascidians, tunicates and polychaete worms.

The LHR is draped with pelagic calcareous sediments known as oozes. These oozes at the surface on the seabed are very pale brown to white in colour and consist almost entirely of the calcite remains of foraminifers and coccolithophorids, planktonic protozoans and algae. DSDP Sites on the south and north LHR along with other cores have sampled this rather uniform sediment type (Burns et al., 1973; Kennett et al., 1986; Colwell et al., 2006). The sediments sometimes contain minor amounts of calcareous spicules, radiolarians, silicoflagellates, diatoms and volcanic glass.

Eade and van der Linden (1970) analysed two cores from the LHR in the southern part of the EMR. One was a 3.5 m length core on the western flank in 3,609 m of water, and had carbonate values varying from 72 to 87% with a foraminiferal sand forming the upper 250 cm consisting of 85% carbonate. The second core, 4.2 m, on the rise itself in 1,450 m of water, had more carbonate, ranging from 80 to 97% with 93% at the surface. Colour cycles are also recognizable subsurface in both cores with the sediment alternating between light olive grey and yellowish grey. Bioturbation is also abundant.

Kawagata (2001) use benthic and planktic foraminifers and oxygen isotope stratigraphy from three cores from the crest of the LHR to determine past changes in sea surface temperatures and productivity, and from this determined the position of the Tasman Front over the past 250 ka. One of the cores is from within the planning area in 1,160 m of water while the others are just outside the northern and southern boundary. Nees (1997) also used benthic foraminifers and oxygen isotope stratigraphy to determine that ocean productivity above the LHR was greater during the last glacial period. The samples came from two cores at 1,340 and 1,450 m water depth on the LHR within the EMR.

Kawahata (2002) analysed three cores from the LHR within the EMR to determine the rate of mineral dust accumulation. The sediment in these cores was a light yellow to light grey foraminiferal coccolith ooze. The organic carbon content of these cores was also measured and had a maximum of 0.38% by weight at the surface and generally less than 0.30% subsurface. Variations in dust content in the cores reflect changing wind patterns during the late Pleistocene.

There is evidence in the subsurface of the LHR that diapirs have had an affect on the seabed by causing uplift and possibly faulting (Fig. 3.44). Above some of the diapirs on the seismic profiles there are faults that appear to reach the surface (Exon et al., 2007). There is a characteristic hummocky seabed above these structures that could be the result of fluid escape. These areas also have a bottom simulating reflector at about 500 m sub-seafloor. Exon et al., (1998) interpreted this as the presence of gas hydrates in the sediments.

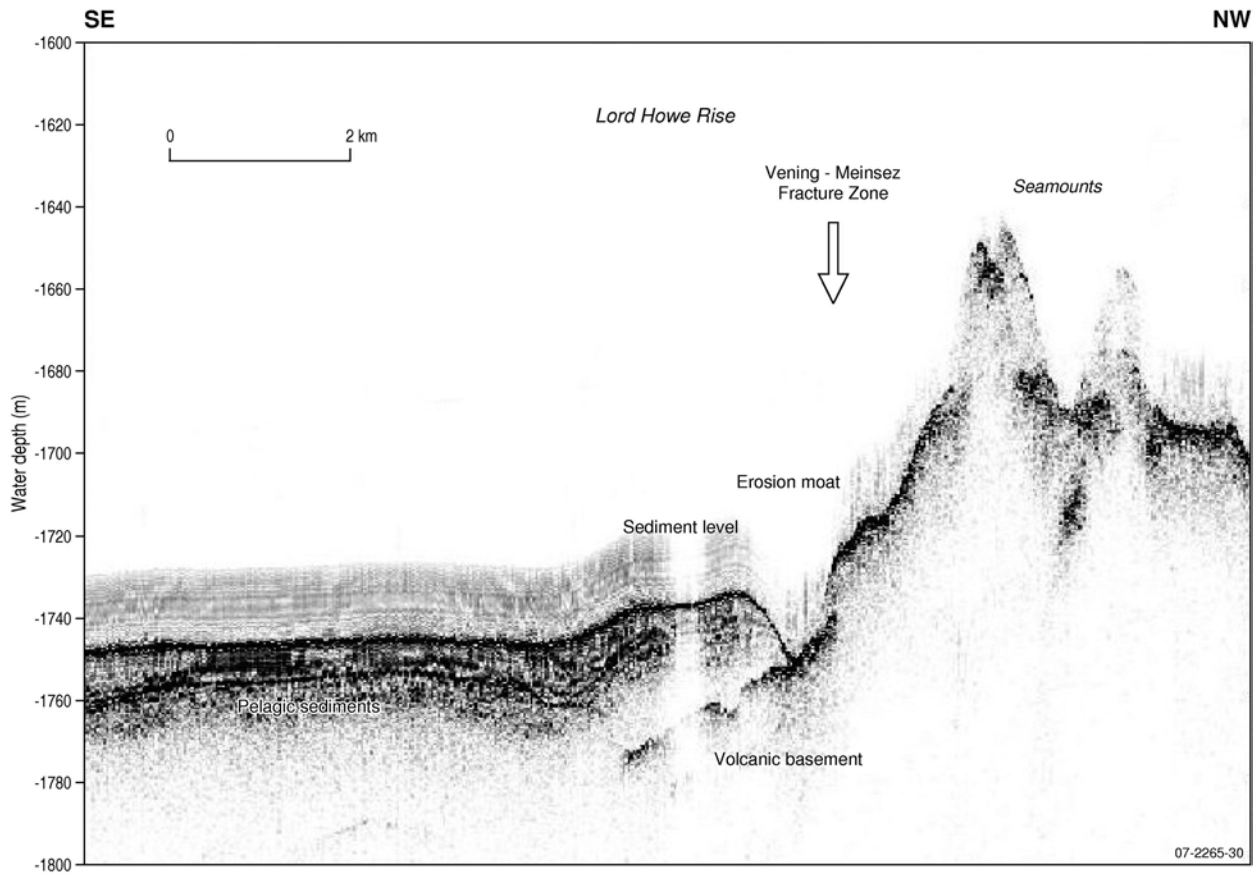


Figure 3.42. Seismic profile across the Vening-Meinesz Fracture Zone on the Lord Howe Rise showing erosion by bottom currents at the base of volcanic ridges. Location on Figure 3.39. Colwell et al., (2006).

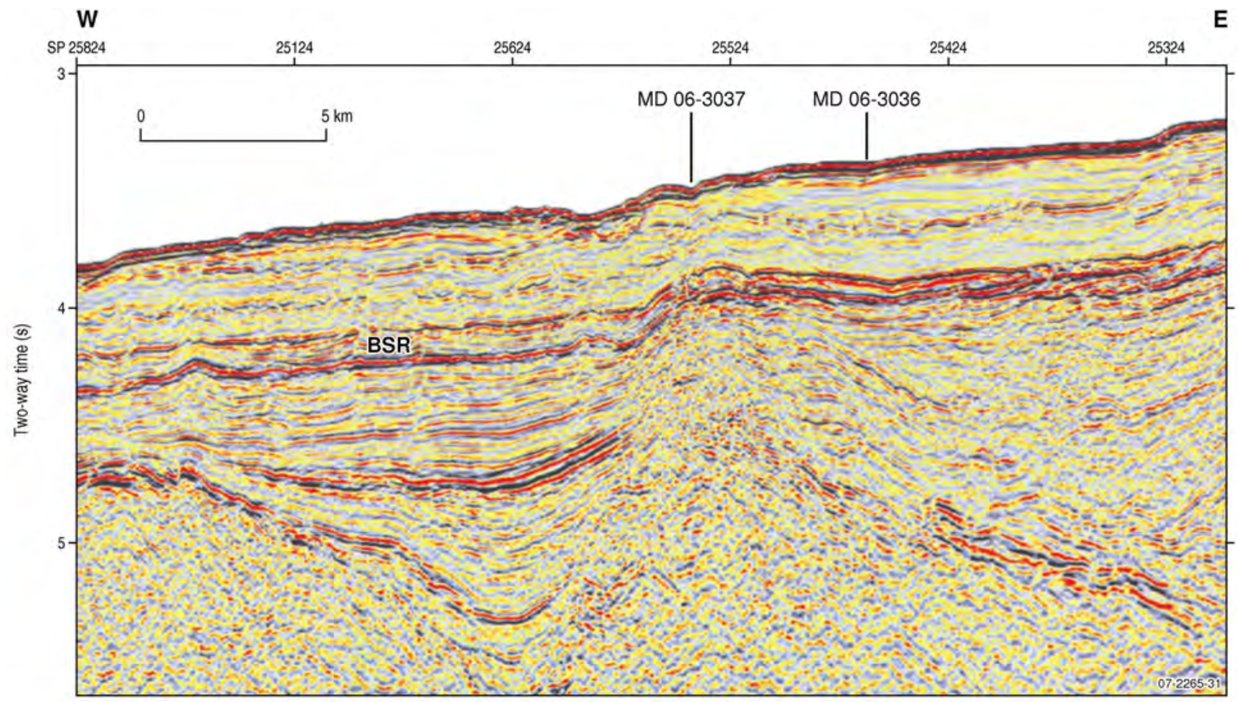


Figure 3.43. Seismic profile showing bottom simulating reflector (BSR) in Capel Basin, western flank of Lord Howe Rise and location of cores. Note the seafloor is irregular above subsurface domes. Location on Figure 3.39. GA line 206/04, Colwell et al., (2006).

3.7. NORTHEAST AUSTRALIAN CONTINENTAL MARGIN CORAL SEA:

3.7.1. Geomorphology

The major features on the seabed in the EMR off northeastern Australia were determined by the breakup pattern of the continental crust and subsequent seafloor spreading (Davies et al., 1989; Pigram, 1993). Subsidence, eustatic sea level and paleoceanographic changes have also controlled the pattern of sediment accretionary features on this margin, particularly for calcium carbonate. Offshore of the Great Barrier Reef are three large marginal plateaus: the Marion Plateau, the Queensland Plateau and the Eastern Plateau (Figs. 3.44 and 3.45). They are separated by deep water embayments known as the Townsville Trough and Queensland Trough/Osprey Embayment. East of the Marion and Queensland Plateaus are a series of smaller troughs and plateaus. To the northeast of the Queensland Plateau is the abyssal plain of the Coral Sea Basin.

Since the tectonic activity ended 52 Ma (early Eocene) the area has undergone differential subsidence and the plateaus were flooded and carbonate sediments began accumulating (Davies et al., 1991b; Isern et al., 2002). Starting in the early Miocene the plateaus have been modified by significant accumulations of carbonate reefs that developed on structural highs (Mutter, 1977; Davies et al., 1989). These reefs formed broad limestone platforms that underpin approximately half of the Marion and Queensland Plateaus (Mutter, 1977; Mutter and Karner, 1980; Isern et al., 2002). These platforms are either exposed as limestone on the seabed or covered by a thin veneer of recent sediment and they form the foundation for extensive present day atoll reefs. The plateau surface surrounding the platforms has a generally flat surface built up by pelagic carbonate ooze and biogenic particles swept off the platforms by currents (Isern et al., 2004). The troughs are also smooth seabed because of the relatively high accumulation rate of redeposited and pelagic sediment. Smaller scale features occur throughout the region and are the result of erosion and deposition by currents, gravity slides and volcanism.

3.7.1.1. Marion Plateau, Capricorn Trough (Channel), Cato Trough, Cato Basin, Kenn (Chesterfield) Plateau, Mellish Rise and Associated Reefs (Saumarez Reef, Marion Reef, Cato Island, Bird Island, Wreck Reefs, Frederick Reef, Kenn Reefs, Selfridge Bank and Mellish Reef)

The Marion Plateau is seaward of the Great Barrier Reef and has a length of about 600 km and an area of 77,000 km^2. It can be divided into two parts because of its varying width and orientation. In the north it extends 200 km NE from the GBR to Marion Reef and in the south it extends SE from the southern end of the GBR for 200 km to the Cato Trough. Isern et al. (2004) describe the two major and two minor carbonate platforms that form about half of the area of the surface of the plateau. They were drowned in the early Pliocene after a period of subaerial exposure and karstification. The southern platform is exposed on the seabed and stands out topographically with moderate relief whereas the northern platform is covered with a veneer of sediment. Two reefs survived the Pliocene drowning: Marion Reef which is built on one of the small platforms and Saumarez Reef which is on the seaward margin of the southern platform. The past and present asymmetric shape of the Marion Plateau platforms is determined by the EAC with the up-current side of the southern platform swept of sediment and forming an erosion moat and deposition occurring on the down-current side (Isern et al., 2001; Isern et al., 2004).

The surface of the northern Marion Plateau has relatively little relief. It is smooth and undulating away from the platform and has a gentle slope to the NE from water depths of 300 m to depths of 600 m (Fig. 3.46; Isern et al., 2004). The southern Marion Plateau slopes east from 300 m to 800 m water depth. Most of the plateau is between 300 m and 400 m water depth.

The southern margin of the plateau runs E-W for 100 km and is relatively steep and rough descending from 800 m to 3,600 m (Fig. 3.35, Exon et al., 2005). A seismic line at 1,500 m water depth crosses volcanic peaks on the seabed several hundred meters high and 3-5 km across (Hill, 1992). The rough topography on this southern slope is also due to erosion of canyons and mass movement of sediment into the northern end of the Tasman Basin. The Capricorn Sea Valley has its origin as a canyon on this slope. The southern margin of the plateau is separated from the Hervey Bay continental shelf by the Capricorn Trough which lies within the Great Barrier Reef marine park.

The eastern margin of the Marion Plateau is relatively straight and trends SSE for 550 km from Marion Reef. Marion Reef and Saumarez Reef lie along this margin. The margin is a relatively gentle convex slope into the Cato Basin where water depths are 3,000 – 3,500 m. The slope is much steeper into the Cato Trough because of a faulted basement ridge forming the lower slope (Mutter and Karner, 1980). Slump deposits are present along the base of this margin (Fig. 3.47; Exon et al., 2005).

The northern margin of the plateau follows the Townsville Trough for 300 km west of Marion Reef. In its eastern section it is relatively steep with a sharp break in slope at about 450 m water depth where a limestone platform crops out and forms the steep upper slope (Fig. 3.46a; Davies et al., 1989). The slope has been modified by currents and drift deposits. The western section is a gradual slope into the Townsville Trough.

The only deep water connection between the Tasman Sea and the Coral Sea is through the Cato Trough. This trough is bounded on the east by the Kenn Plateau and on the west by the southern Marion Plateau and it extends for 550 km from the Tasman Basin to the Mellish Rise. The trough is at its narrowest in the south where for 50 km of its length the flat seabed is only 10 km wide and water depths are 3,400 to 3,500 m (Fig. 3.47). Both sides are steep due to exposure of basement ridges. The trough opens northwards into what is called the Cato Basin by Exon et al. (2005). The Cato Basin is triangular in shape, 200 km E-W and 500 km N-S, with the Mellish Rise along its northern margin. The eastern margin of the Cato Basin trends north-south and rises abruptly from 3,000 m water depth. It is complex in detail as it is formed by steep-sided tectonic blocks of the northern Kenn Plateau. Many small canyons and several larger ones flow into the basin down the eastern margin of the Kenn Plateau (Exon et al., 2005). Cato Island and Kenn Reef are two seamounts that reach the surface along this margin. In contrast the western margin is blanketed in sediment and rises gradually to the Marion Plateau. The basin has a relatively smooth floor 3,100 to 3,200 m deep in the east with occasional seamounts including two that rise to the surface as carbonate banks: Frederick Reef and Wreck Seamount (Wreck Reef and Bird Island).

Most of the Kenn Plateau, as defined by Exon et al. (2005), lies within the EMR (Fig. 3.44). This plateau consists of a series of ENE trending ridges shallower that 2,000 m and troughs (basins) with water depths of 1,800 – 3,000 m (Figs. 3.34, 3.48 and 3.49). The major features from south to north are:

Southern Surveyor Rise (~1,600 m), Observatory Basin (~1,800 m), Coroilis Ridge (~1,000 m), Chesterfield Trough (~2,200 m) and Selfridge Ridge (1,000-2,000 m). The Coroilis Ridge has a 15 km wide platform in 100-300 m of water on its northwestern corner. Selfridge Ridge joins Kenn Reef and banks with Selfridge Rock (water depth 44 m) on the boundary of the EMR.

The margins of the Kenn Plateau are steep with many small canyons and fault scarps and slump scars exposed. Figure 3.50 shows fields of sand dunes occur on the western top of Coriolis Ridge in water depths of ~1,000 m (Exon et al., 2006b). They are 4 to 5 m high, have asymmetric sinuous crests and a wavelength of 250-350 m. These dunes along with scours around pinnacles indicate the currents are flowing to the east or northeast.

To the north of the Kenn Plateau is Bampton Trough, 3,000 m deep and 50 km wide ,which separates the Kenn Plateau from an area of seabed that is generally shallower than 2,500 m, trends NE and is about 700 km long and 250 km wide. It is irregular in shape and with considerable relief but known as Mellish Rise (Exon et al., 2006a). In the southwest Mellish Reef rises as a limestone-capped volcanic seamount from where the plateau is shallower than 2,500m (Terrill, 1975). Other seamounts are common and may represent a northern extension of the Tasmantid Seamount chain (Exon et al., 2006a). The topography is a complex of ridges and valleys possibly caused by faulting and partial breakup of a large plateau (Exon et al., 2006a). Tilted basement blocks are well imaged by multibeam bathymetry in Figure 3.50.

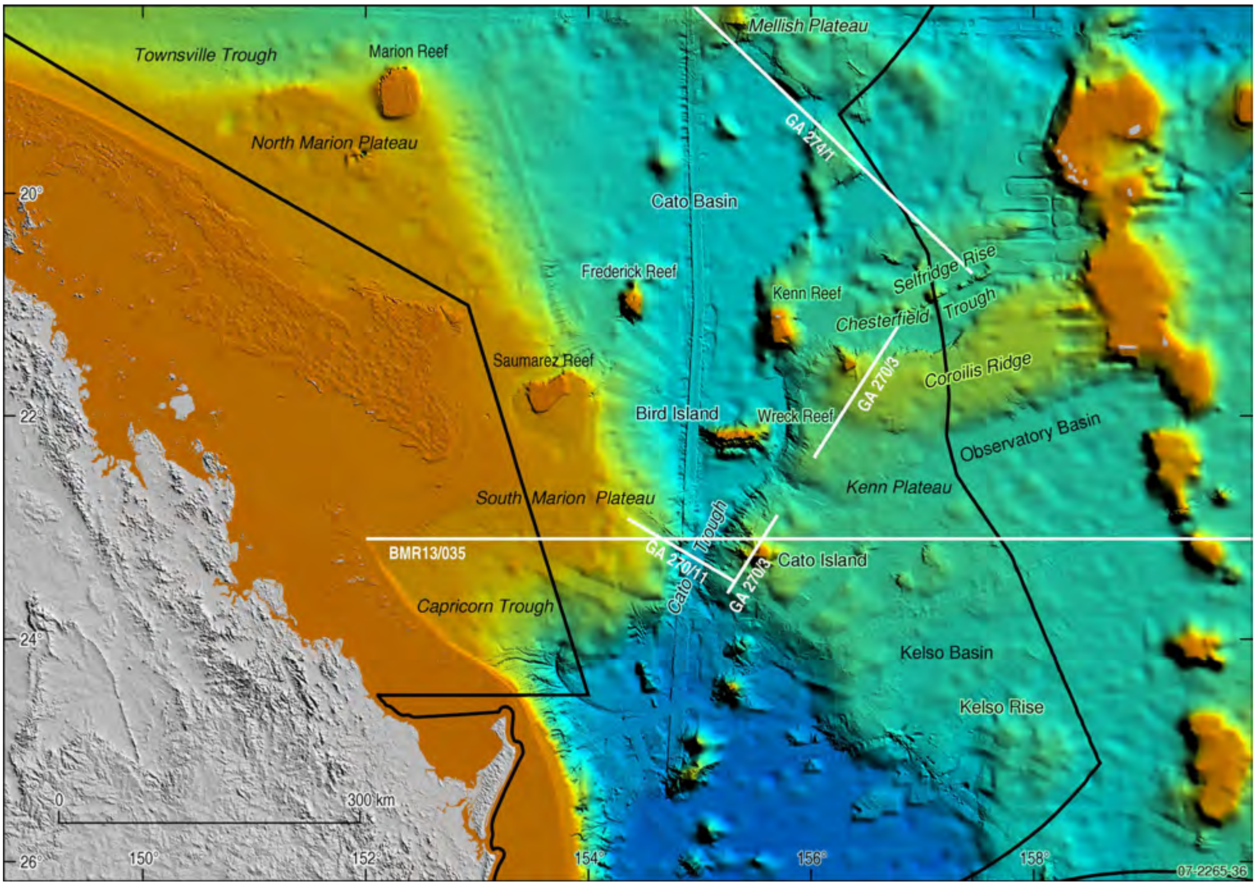


Figure 3.44. False colour image showing geomorphic features of the Cato Trough, Cato Basin, Kenn (Chesterfield) Plateau, Mellish Rise, Marion Plateau and associated reefs. Locations of features and seismic lines displayed in other figures are marked. Black line is the EMR boundary.

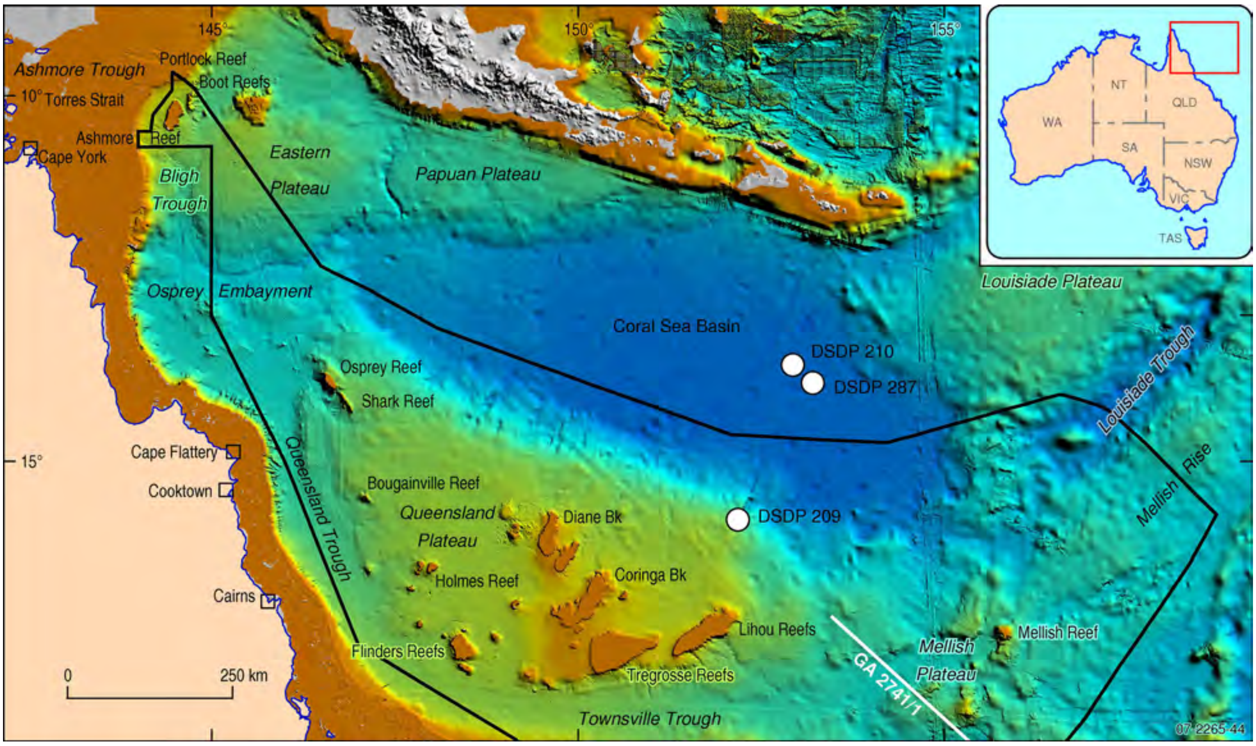


Figure 3.45. False colour image showing geomorphic features of the Queensland Plateau, Townsville Trough, Queensland Trough, Osprey Embayment, Eastern Plateau and associated reefs. Locations of DSDP drill sites are indicated. Black line is the EMR boundary.

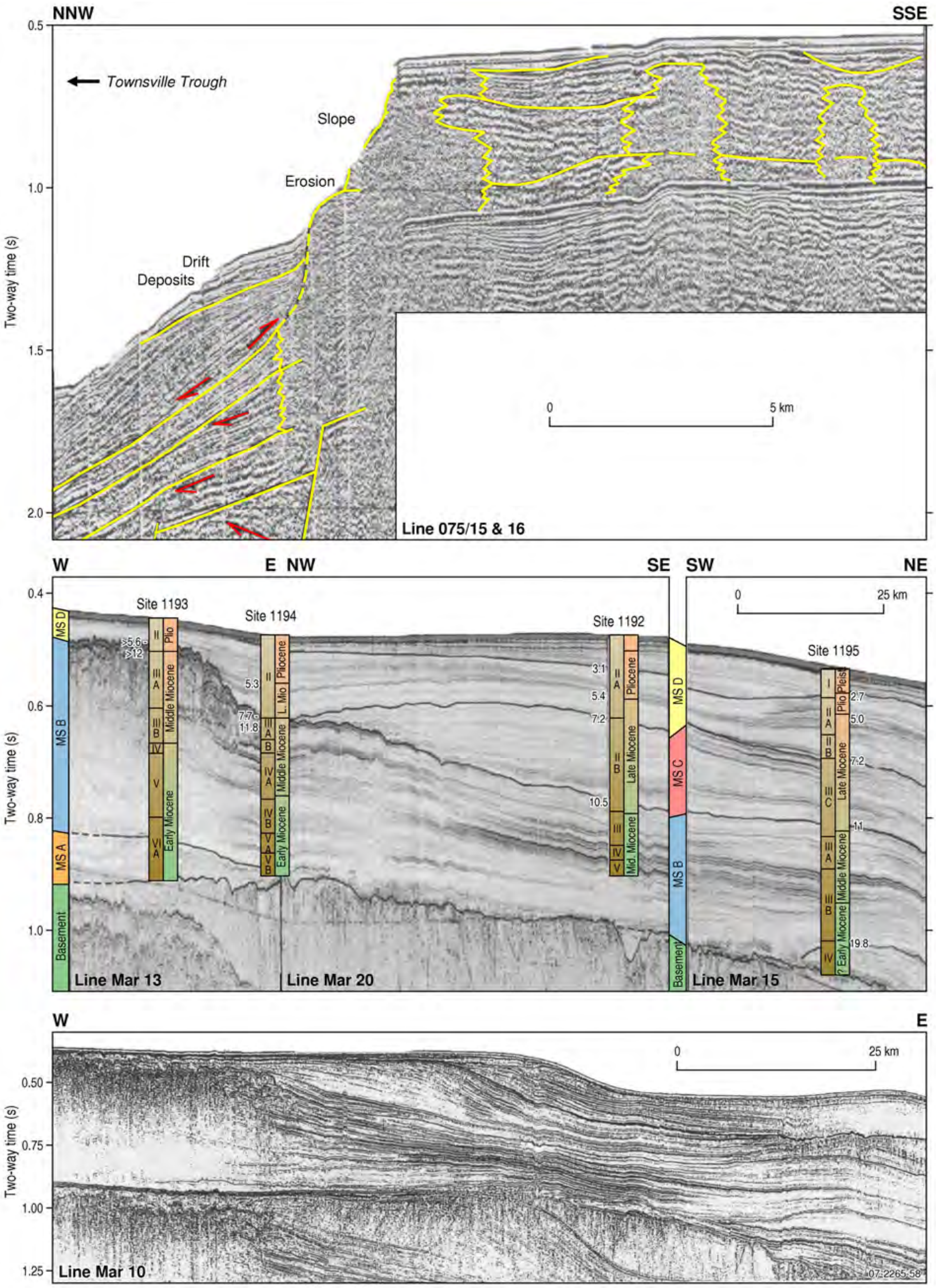


Figure 3.46. a) Seismic profile across the northern Marion Plateau carbonate platform adjacent to the Townsville Trough. Note the sharp break at the platform edge and the erosion and drift deposits on the slope. The sediments beneath the seabed are interpreted as barrier reef (BR), lagoon (L), patch reef (PR) and forereef periplatform (P). (Davies et al., 1989), b) Seismic profile across the southern part of the northern Marion Plateau showing location and results of ODP drill sites. (Isern et al., 2004) and c) Seismic profile across the central part of the northern Marion Plateau from west to east. (Isern *et al.,* 2004).

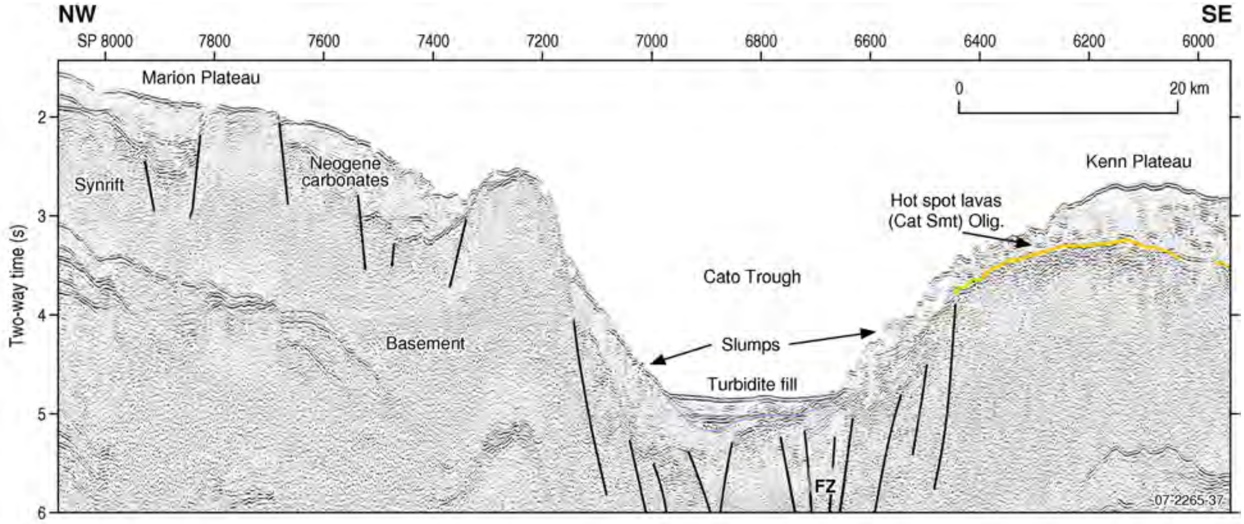


Figure 3.47. Seismic profile across the Cato Trough showing erosion, exposure of bedrock, sediment slumps on the slopes and turbidites in floor of trough. Line 11. Location in Figure 3.44. Exon *et al*., (2005).

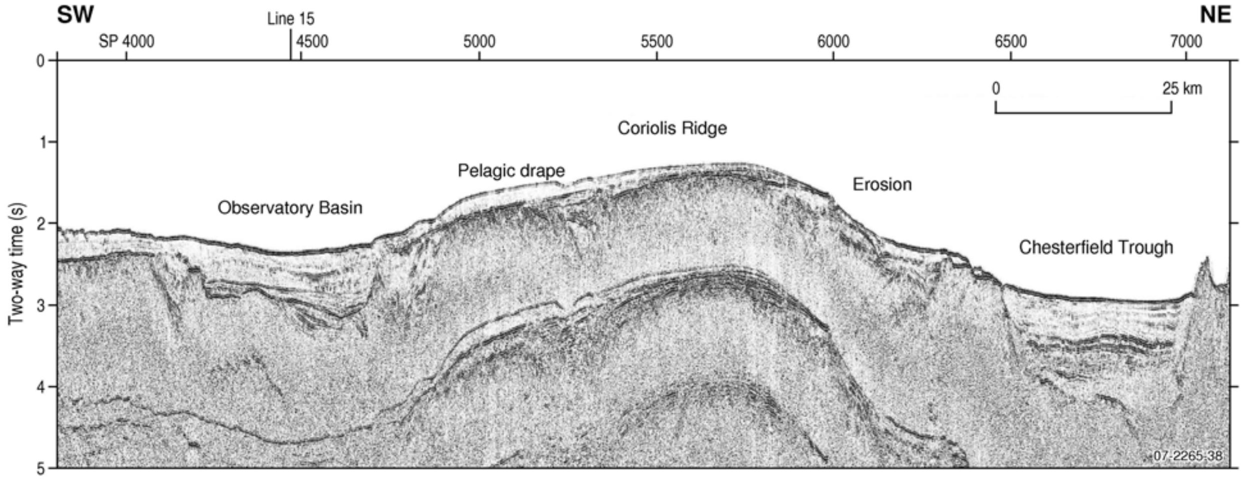


Figure 3.48. Seismic profile across the Observatory Basin, Coriolis Ridge and Chesterfield Trough showing general drape of pelagic sediments. Line 270-3. Location in Figure 3.44. Exon *et al*., (2005).

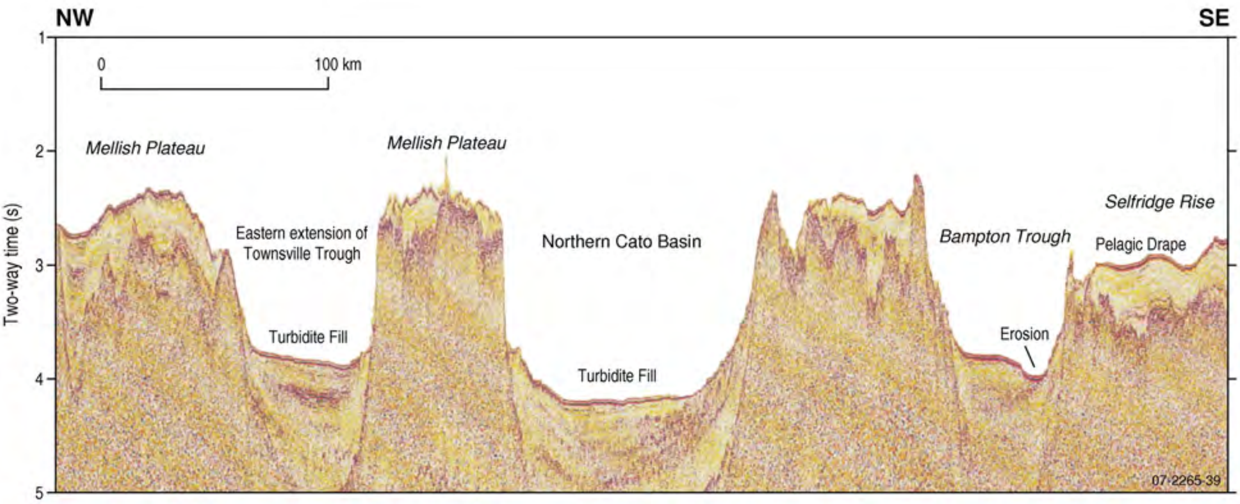


Figure 3.49. Seismic profile from Selfridge Rise to Mellish Rise showing the trough and ridge topography typical in this area. Line GA274/01. Location in Figures 3.44 and 3.45. Exon *et al*., (2006a).

*3.7.1.2. Queensland Plateau, Townsville and Queensland Troughs and Reefs(*Lihou Reefs and Cays, Tregrosse Reefs, Malay Reef, Abington Reef, Flinders Reefs, , Coringa Bank, Magdeline Cays, Willis Islets, Diane Bank, Moore Reefs, Holmes Reefs, Herald Cays, Flora Reef, Bougainville Reef, Shark Reefs and Osprey Reef*)*

The Queensland Plateau is roughly triangular in shape with its northeast side facing the Coral Sea Basin for 800 km, its western side is the Queensland Trough for 550 km and its southern side is the Townsville Trough for 500 km, an area of about 165,000 km^2. Both troughs are rift basins partly filled by sediment (Symonds et al., 1983; Scott, 1993). On the northeast margin there is a change in trend of the contact between the slope and the abyssal plain at 14° 20'S. The slope is 40-60 km wide and occurs between 2,000 and 4,000 m water depth. In the north it has an abrupt contact with the abyssal plain at ~ 4,000 m whereas in the south there is a rise extending out from the base of slope at ~3,500 m. The slope has a convex shape, steeper in the southern section, and is cut by many canyons (Gardner, 1970; Mutter, 1977). East of Lihou Reefs the plateau deepens to form two terraces at depths of 1,400-1,600 m and 2,200 m with a shallow trough between them.

On the basis of water depth the plateau can be divided into half along a line drawn from where the two troughs meet the Coral Sea Basin. The plateau is 350 km wide at this point and the southeast half is mostly shallower than 1,000 m and the north-west half is mostly between 1,000 and 2,000 m water depth with a gentle slope to the north. The seabed on the plateau is generally flat and smooth with numerous steep-sided reefs ranging in size from very small pinnacles to those that reach the sea surface and have formed large platforms. The living reefs, at or near present sea level, form 10-15% of the area of the plateau and are most common in the south and along the western margin (Davies, 1988).

The Miocene was a period of extensive reef growth that formed extensive limestone platforms that are a feature of the Queensland Plateau (McKenzie and Davies, 1993). Modern day reefs have formed smaller platforms on the submerged and largely buried Miocene platforms (Davies et al., 1989; Betzler et al., 1995). The largest of the modern platforms are the Tregrosse Reefs and Lihou Reefs on the southern margin of the plateau. They are each nearly 100 km long from east to west and 50 and 25 km wide, respectively. They rise steeply from water depths of >1,000 m where they are near the edge of the plateau and from ~500 m water depth on the plateau. To their north are the slightly smaller Coringa Bank and Diane-Willis complex of reefs. Osprey Reef on the northern end of the plateau rises steeply from over 2,000 m water depth on its western side to reach sea level. Terraces occur around some of these reefs indicating more extensive growth in the past. A major terrace occurs at 450-500 m and another at 50 m (Davies et al., 1989). Mutter (1977) reports that sediment aprons surround the base of these reefs and extend a considerable distance from them. Slumping, mass transport and shedding of carbonate material from the reefs have created these aprons.

Numerous drowned reefs have been reported on the plateau (Taylor, 1977; Mutter, 1977; Davies et al., 1989). Large isolated pinnacles occur along the western margin of the plateau. Some of these are only 1-2 km across and rise from over 1,000 m water depth to within 10 m of sea level. A pinnacle in the Queensland Trough east of Flinders Reef rises steeply 500 m from a flat seabed to within 675 m of the surface. Davies et al., (1989) sampled reef limestone from this pinnacle. Seismic evidence shows that some of these pinnacles form on the raised corners of fault blocks (Davies et al., 1989). Along the southern margin of the plateau there are numerous small canyons cut in the slope leading down from Tregrosse Reefs and Lihou Reefs into the Townsville Trough.

The Townsville Trough extends from the GBR eastward into the northern end of the Cato Basin and continues as a topographic trough into the Mellish Plateau (Fig. 3.45). The trough is a broad U-shaped feature some 100 km wide at the 2,000 m isobath at its eastern end. It narrows and shoals gradually to the west for 500 km to reach a sill depth of 900 m offshore of Townsville. On the north side of this sill is the Queensland Trough. It extends for 550 km to the north from 18-14°S gradually deepening to 2,900 m north of Osprey Reef where it widens (200 km) into the Osprey Embayment. The floor of the trough is flat with a northerly slope of <10°.

The southern section of the Queensland Trough is broad (90 km) and asymmetric with a much steeper slope on the western margin as it rises to the GBR. North of Bougainville Reef at 15° 10'S it abruptly narrows due to a basement ridge, making the eastern side also steep with 150 to 500 m of relief and a distinct terrace before rising to the plateau. This terrace continues north to Osprey Reef. The GBR side of the trough is mostly outside the EMR but is also steep with many canyons and scars from submarine slides. Johnson (2004) reports failures up to 30 km wide and 50 km long south of Cooktown. North of Cooktown the shelf edge and upper slope are deeply incised by canyons.

3.7.1.3. Osprey Embayment, Eastern Plateau and Reefs (Bligh Trough, Ashmore Reef, Boot Reef, Portlock Reef and Ashmore Trough).

The EMR is narrow here and covers only part of the Osprey Embayment and a wedge-shaped part of the Eastern Plateau (Fig. 3.45). This plateau is between 1,000 and 2,000 m deep with the Bligh Trough along its western margin. North-west of the Eastern Plateau across a sill (1,500 m water depth) separating the northern end of the Bligh Trough from the Pandora Trough are two significant reefs in the EMR. They are Ashmore Reef and Boot Reef which rise from depths on their eastern side of 1,000-1,500 m to form carbonate platforms in less than 100 m of water. Both platforms are elongate N-S. Ashmore is the largest platform at 45 x 20 km and Boot is 20 x 10 km. Boot Reef is only 10 km to the north-east of Ashmore Reef and separated by a narrow, 700 m deep channel. Ashmore is only 50 km east of the GBR from which it is separated by the 700 m deep Ashmore Trough. This trough is a valley that drains south into the Bligh Trough. North of these reefs the shelf edge runs nearly east-west. The shelf edge here is distinguished by a very sharp break and steep upper slope (Harris et al., 1996b). It is interpreted as a drowned barrier reef that was constructed on a beach ridges in the late Pleistocene (Frances et al., in press). At its eastern end is a relict delta from a lower sea level. Further east is Portlock Reef which is a partly buried atoll now on the shelf edge. Boot Reef is an atoll 25 km south of Portlock Reef. The saddle joining Boot Reef with the slope is at 500 m water depth and the saddle at the northern end of Ashmore Reef joining it to the slope is at 650 m water depth. The valley between the two reefs and one north of Boot Reef both drain east into the Pandora Trough. Scalloped morphology on the margins of Ashmore Reef, Boot Reef and Portlock Reef suggest large gravity failures. The possibility of failure is enhanced on atolls because the upper slope (<500 m) is very steep, leading to failure.

The Osprey Embayment extends west for 300 km from the Coral Sea Basin to the GBR. It is 200 km wide from north to south. The seabed has a general slope to the north and consists of small plateaus and ridges with a relief of several hundreds of meters separated by troughs. Its major feature is a broad, sinuous deep sea valley along its northern margin where it is deepest (3,600 m). This channel was named the Bligh Canyon by Winterer (1970) and it meanders east across the EMR for 160 km.

Winterer (1970) mapped a 60 km long section of the canyon in the EMR and showed that it was flat-floored, 10 km wide and entrenched about 100 m below the surrounding seabed. A smaller channel, only a few meters deep, meanders across the flat floor of the canyon. The Queensland and Bligh Troughs act as tributaries to the eastward trending Bligh Canyon.

3.7.1.4. Coral Sea Basin Abyssal Plain, southern Louisiade Plateau and Louisiade Trough

There is a paucity of data for this area. The Louisiade Trough separates the Mellish Rise from the Louisiade Plateau. The southern extensions of both are within the EMR. Like the Mellish Rise the southern margins of the Louisiade Plateau are irregular and the topography, where surveyed, is rough and rises steeply from the surrounding seabed. The geomorphology is inherited from their tectonic origins. Between the Louisiade Plateau and the Mellish Rise lies the Louisiade Trough which runs NE-SW and is deepest in the north. It has a maximum width of about 100 km but its margins are not well defined by surveys. Over 250 km of its length are within the EMR and it is up to 4,500 m deep.

The abyssal plain/deep ocean floor of the Coral Sea Basin extends into the EMR for a distance of up to 140 km over a length of 1,000 km seaward of the Queensland Plateau. It extends from the Osprey Embayment in the north-west, gradually deepening to the southeast where it abuts the southern part of the Louisiade Plateau. The plain lies at a depth of 4,000 to 4,500 m, and in the north it has a gradual contact with the slope of the Queensland Plateau forming a rise.

3.7.2. Surface Sediments and Rocks

The sediments in this area are dominated by pelagic carbonate deposits but with significant contribution of carbonate from the shallow water reefs. The whole area is above the calcite compensation depth so preservation of the carbonate is excellent except in the Coral Sea Basin. The ocean in the area is not highly productive so siliceous plankton are missing from the sediments except in the northeastern margin, where diatoms are found in the sediment beneath the South Equatorial Current. Information about samples and bottom photographs from this area is collated in Appendix C Tables 14 to 20

In general the GBR acts as a barrier to terrrigenous sediments reaching the slope. An exception is the Queensland Trough where up to 40% of the sediment on the seafloor is terrrigenous mud that has been flushed through the GBR from the shelf (Francis et al., 2007). Further north in the Osprey Embayment and in the Coral Sea Basin, there is also significant dilution of the carbonate sediment by terrigenous mud from the rivers of Papua New Guinea. Sediments on the topographically isolated Eastern Plateau have up to 50% terrrigenous muds (Francis et al., in press).

ODP drilling of the limestone and carbonate sediment forming the Marion Plateau showed it to be 450-650 m thick and to have accumulated since the Oligocene (Isern et al., 2002). The limestone lies on basement which was sampled at five ODP sites and consisted of volcanic flows and breccias of altered basalt. They suggest the basalt was emplaced at the time of rifting in the late Cretaceous-Paleocene. This basement outcrops on the faulted margin where the Marion Plateau descends into the

Cato Trough. On the Queensland Plateau the basement was sampled at two sites at depths of 400-450 m below the seabed, and consisted of meta-sedimentary rocks; the overlying limestone deposition started in the middle Eocene (Davies et al., 1991b).

Sedimentation on the Marion Plateau is dominated by the EAC creating erosion around limestone outcrop but elsewhere the current has smoothed the relief and created undulations on the seabed where it has deposited large sediment drifts. Bottom photographs at water depths of 348 m and 374 m show ripple marks in the sand along with cemented slabs encrusted by sponges on the plateau surface east of the northern platform (Isern et al., 2002). Photographs and samples taken on the southern platform show the seabed to be a ferromanganese-stained hardground of limestone, bored and encrusted with serpulids and bryozoans with some unconsolidated carbonate sediment in depressions (Heck et al., 2007). There are also scoured channels around Saumarez Reef on its north and west sides. Away from the platforms the sediment is a sand or muddy sand dominated by planktonic foraminifers with minor skeletal grains of bryozoans, scaphopods, solitary corals, sponge spicules and pteropods (Isern et al., 2002, 2004). Cores analysed by Heck et al., (2004) from east of the southern platform have sedimentation rates of 2.6-3.5 cm ka^{-1} and carbonate content ranges from 79% to 93% reflecting the glacial-interglacial cycles down-core with the higher values from sediment at the surface and at other interglacials intervals. ODP Sites 815 and 816 on the northern margin of the Marion Plateau sampled foraminifer coccolith ooze on the seafloor at water depths of 466 m and 438 m.

Sediments in the Cato Trough and Basin are pelagic oozes (Walker, 1992). Photos of the seabed show ripples in foraminiferal sand where the Trough is narrow and debris blocks from the adjacent slopes (Figs. 3.51 and 3.52). Away from the narrow pass the muddy sand has a rich epifauna and grazing trails and burrows are common (Fig. 3.53). A core from the central Cato Basin in 3,152 m sampled graded beds of foraminifer coccolith ooze with pteropods deposited by turbidity currents (Jenkins et al., 1986). The sediment was >90% carbonate.

Sedimentation on the Queensland Plateau is dominated by very pale orange to greyish orange calcareous pelagic ooze consisting of foraminifers and coccoliths with lesser amounts of pteropods (Gardner, 1970; Burns et al., 1973). The living reefs on the Queensland Plateau are similar to those on the GBR (Done, 1982). These reefs are shedding carbonate particles, particularly coral, algae, foraminifers and molluscs down their steep slopes and onto the surrounding seabed (Orme, 1977). ODP drilled at seven locations on the western and southern Queensland Plateau to sample a variety of depositional environments. Close to reefs on the plateau (Site 811/825, 812, 813, 814) in 462-940 m water depth foraminifer coccolith pelagic oozes were interbedded with redeposited shallow water carbonate sediments and foraminifer-pteropod sands with carbonate values of >95% and sedimentation rates of 1 to 2.4 cm ka-1 (Davies et al., 1991b). Away from the reef platforms, at Sites 817, 818 and 824 in water depths of 749 to 1,017 m on the gently sloping margin of the plateau, the sediments are more mud-rich due to an increased proportion of coccoliths and less foraminifers and pteropods. Interbedded with the pelagic ooze are thin beds of coarser carbonate consisting of molluscs, bryozoans, corals and coralline algae. These along with slump folds in the cores indicate downslope transport of sediment. On the northeast margin of the plateau DSDP Site 209 in water depth of 1,428 m sampled very pale orange to grayish orange foraminifer coccolith ooze (Burns et al., 1973).

The Queensland Trough is filling with hemi-pelagic sediments and redeposited sediments from the GBR and Queensland Plateau. Four large (5-10 km wide and 10s km long) sediment gravity flows have been mapped on the seafloor between 15° and 16° 30'S by Dunbar et al. (2000). They originate as failures on the seaward slope of the GBR and flow down slope and curve north down the trough axis. Cores in these deposits show chaotic sediment structure. Turbidite layers are common in cores in the trough (Watts et al., 1993; Blakeway, 1991; Dunbar et al., 2000). The near surface sediment on the seabed sampled at ODP Site 823 consisted of hemipelagic, clay-rich, foraminifer coccolith mud with 60% carbonate, interbedded with redeposited layers interpreted as turbidites and debris flows (Davies et al., 1991a).

Carbonate content in the surface sediment generally increases from the GBR slope to the trough and plateau and decreases northwards down the axis of the trough (Dunbar, 2000; Dunbar and Dickens, 2003a). It decreases from >90% at the sill with the Townsville Trough to 60% in the north. The carbonate content is >90% on the Queensland Plateau and 100% near isolated reefs. On the GBR slope Francis et al., (2007) have identified a zone between 15^0 and 17^0S with the lowest carbonate content of 40-80% and suggest that inter-reef passages supply the terrigenous mud that dilutes the carbonate in the trough. They further suggest that the pattern of clay mineral concentration indicates transport to the south by surface and bottom currents.

Harris et al. (1990) analysed cores from the western slope of the Queensland Trough offshore of Townsville and found that the surface sediment was sandy mud and muddy sand (20-45% foraminifer-pteropod sand) with 80-90% carbonate in water depths of 690 m to 1,200 m. Below the surface there were cycles with less carbonate due to an increase in terrigenous muds which Harris et al., (1990) interpreted as lowstand deposits. More recent analyses of numerous cores from the Queensland Trough and its slopes have recovered siliciclastics (quartz, feldspar, clay minerals) in surface sediments indicating a considerable flux of terrigenous sediment through the passages in the reef during the present-day sea level highstand when more carbonate was expected to be supplied (Harris et al., 1990; Dunbar et al., 2000; Dunbar and Dickens, 2003a, 2003b; Page et al., 2003; Page and Dickens, 2005; Francis et al., 2007).

Average sedimentation rates for the past 6.5 ka are highest on the slope in front of the GBR (up to 55 cm ka-1), 2-3 cm ka-1 in the floor of the trough and low (<1.5 cm ka-1) on the Queensland Plateau adjacent to the trough (Dunbar et al., 2000). On the southern slope of the Queensland Plateau Cotillon et al. (1994) calculated the late Pleistocene average sedimentation rate to be 5.2 cm ka-1.

Ashmore Reef and Boot Reef are carbonate platforms less than 100 m deep and rise steeply from about 1,000 m. They are separated from each other and the shelf by a 600-700 m deep known as the Ashmore Trough. Portlock Reef forms the outer rim of the continental shelf of eastern Torres Strait. This is the largest modern example of a tropical mixed siliciclastic/carbonate depositional system (Davies et al. 1989). Significant quantities of clay, sand (rock fragments, feldspar and quartz) and other weathering products are discharged from PNG rivers to a shelf, where it is deposited with large amounts of carbonate produced from algae (e.g, Halimeda), foraminifera, mollusks and corals. These shelf sediments along with pure carbonate sediment from platform tops/barrier reef then shed aragonite and high-Mg calcite particles onto the surrounding slopes and basins where they accumulate with low-Mg calcite pelagic carbonates. It appears that large slumps have occurred from the summits of these atolls, and large limestone blocks have been transported to the surrounding seabed.

In the north of the EMR the significant input of sediment, mostly terrigenous muds, from the PNG shelf reaches the slope (Brunskill et al., 1995; Walsh and Nittrouer, 2003). A core taken from the Ashmore Trough at the northern end of the EMR in water depths of 760 m consists of dark grey/pale gray laminated calcareous muds. This hemipelagic sediment consists of 75% carbonate (coccoliths, foraminifers and pteropods) at the surface which decreases down core to 25% during the Last Glacial Maximum (20,000 y.b.p.). The core is useful for paleoclimate studies because it has a very high sedimentation rate of ~25 cm ka-1 and the depositonal laminae have not been bioturbated which allowed de Garidel-Thoron et al. (2004) to use carbon isotopes to suggest possible methane escape from near here that triggered abrupt climate warmings in the Pleistocene. Francis et al. (in press) determined that the sedimentation rate in Ashmore Trough is low for the Holocene. The slope here and in the Osprey Embayment is blanketed in mud drifts deposited from turbid waters from the PNG shelf, probably transported by the clockwise deep water current in the Gulf of Papua (Keen et al., 2006). During the last glacial maximum sea level lowstand beach ridges and deltas formed at the shelf edge and upper slope. During the low stand and regressions leading up to it large volumes of organic-rich sediments were deposited on the toe of slope of the northern Ashmore Trough at 600-800 m of water depth. The laminations in the cores suggest the basin was poorly ventilated. These sediments contain a high resolution archive of Earth climate variability, particularly of the Western Pacific Warm Pool (Beaufort et al., 2005).

The Coral Sea abyssal plain/deep ocean floor has been sampled by shallow cores (Gardner, 1970; Beiersdorf, 1989; Jenkins et al., 1986) and by DSDP Sites 210 and 287 (Burns et al., 1973; Andrews et al., 1975). All the surface samples from the southwest of the basin were similar: a white pelagic foraminifer coccolith ooze at the surface generally less than a meter thick overlying interbedded turbidites and pelagic ooze. In the far east of the basin (15° 15′S; 154° 45′E) in water depths of 4,590 m Jenkins et al., 1986 sampled a diatomaceous foraminifer coccolith ooze overlying olive grey silty clays. This is the only known occurrence of significant quantities of diatoms in sediments in the EMR. The turbidites at both locations are olive grey graded beds of fine sand, silt and clay. The graded beds are terrigenous sediment transported by turbidity currents down canyons to the abyssal plain. Most of the sediment comes from the deltas along the PNG coast, but some turbidites are calcareous and consist of redeposited pelagic and shallow water carbonate from the Queensland Plateau. Gardner (1970) concluded that the pelagic ooze accumulated at a rate of 3.6 cm ka-1.

Coring on the Kenn Plateau recovered pale brown to white calcareous ooze consisting predominantly of foraminifers and coccoliths (Exon et al., 2006a). This is as expected for these water depths and productivity. Cores on Coriolis Ridge near the sand dunes recovered a foraminiferal sand confirming the presence of strong bottom currents.

Dredging on the Mellish Rise by Exon et al. (2006a) recovered basalts and fragmented volcanic debris along with pelagic sediments of late Cretaceous to Cainozoic age. Further south on Selfridge Rise they recovered quartzite indicating continental crust. Rocks from both the Kenn Plateau and Mellish Rise are coated with manganese crust generally <10 cm thick but with some up to 20 cm thick. One dredge on Mellish Rise in 2,630 m water depth recovered manganese nodules along with basalt (Exon et al., 2006a).

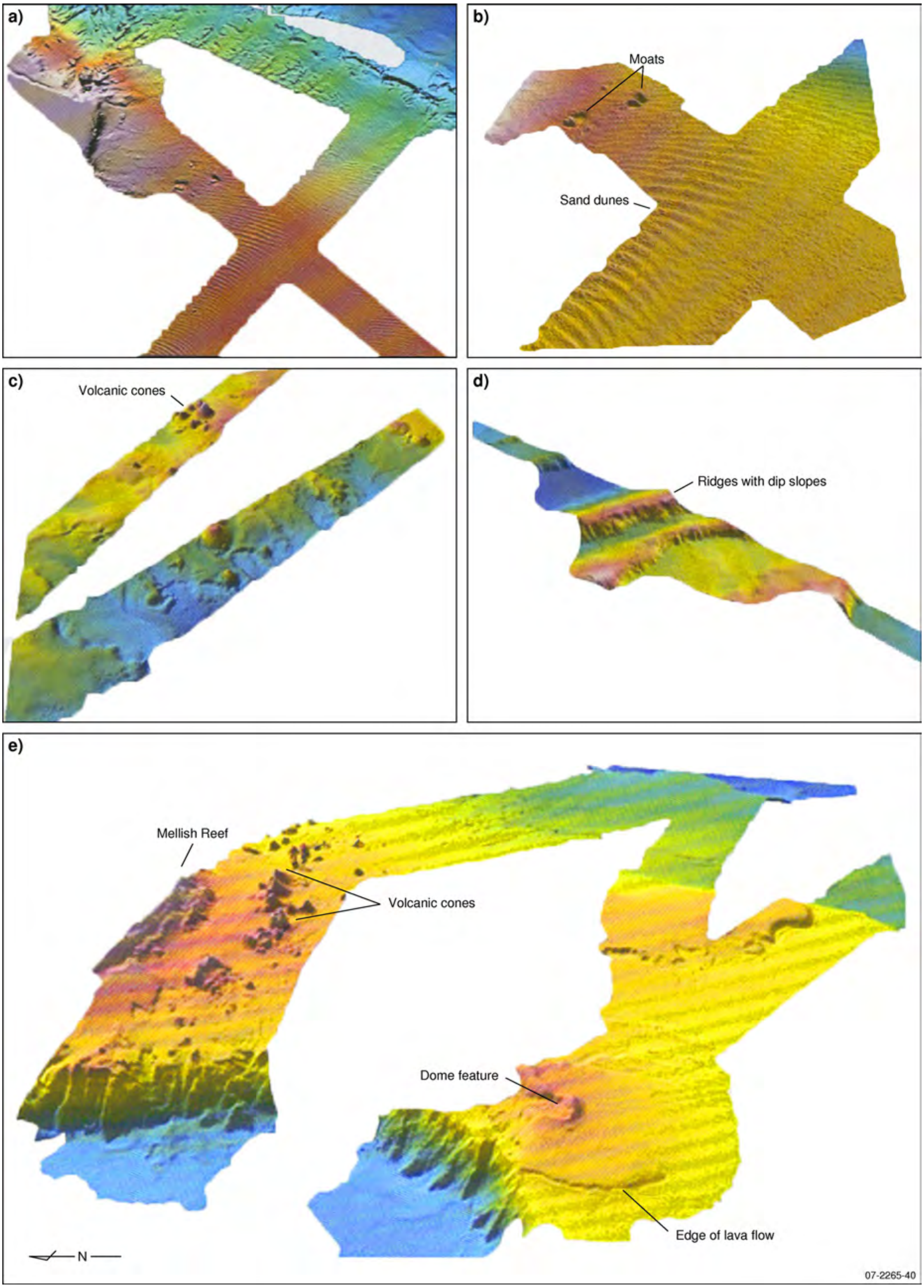


Figure 3.50. Multi-beam 3D imagery of bedforms, erosion and outcrop patterns on the Mellish Rise.a and b) Flat seabed with sand dunes (wavelength of 300 m) and outcrop with erosional moats, c and d) Volcanic cones and tilted basement ridges and e) volcanic cones, domes and lava flows. Locations in Figure 3.44 and 3.45. Exon *et al*., (2006).

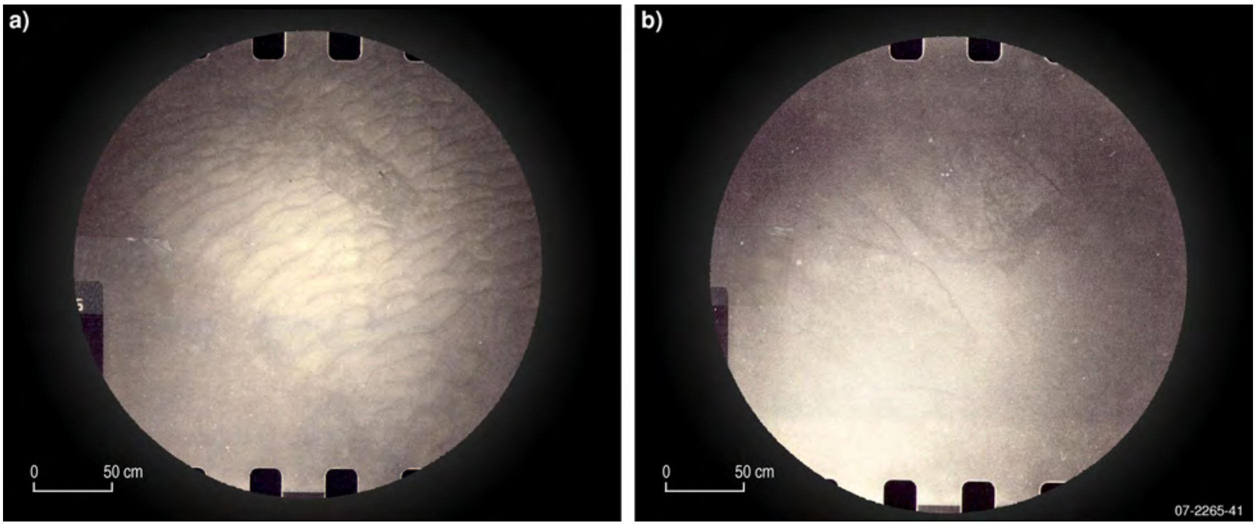


Figure 3.51. Seabed photos from the Cato Pass showing active ripple marks in sand and blocks of rock with organism bending in the current. BC1, Location in Figure 3.44. Walker, (1992).

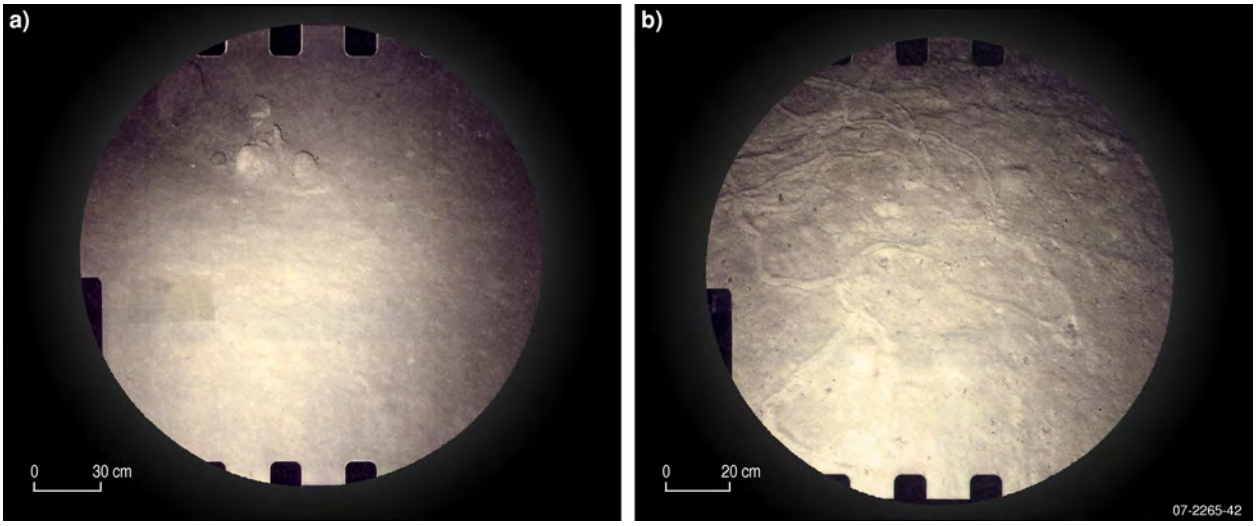


Figure 3.52. Seabed photos from the eastern side of the Cato Trough showing active mounds and tracks in muddy sand and rounded boulders draped with sediment . BC2, Location in Figure 3.44. Walker, (1992).

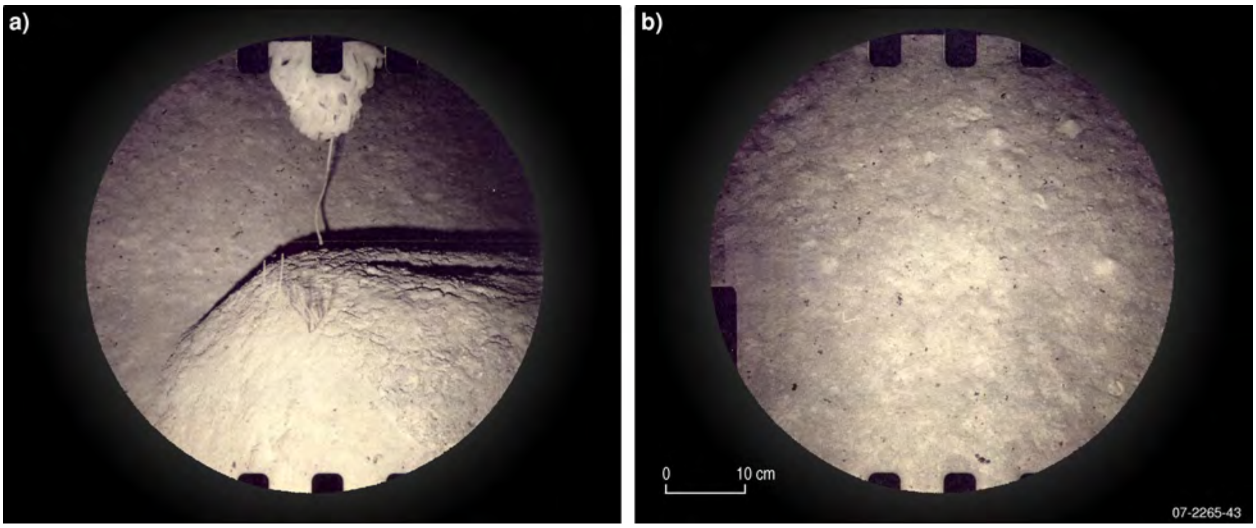


Figure 3.53. Seabed photos from the central Cato Trough showing bioturbated sediment and mounds with sponge and other epifauna. BC3, Location in Figure 3.44. Walker, (1992).

3.8. WESTERN PACIFIC OCEAN: NORFOLK ISLAND

3.8.1. Geomorphology

New Caledonia Basin, Norfolk Ridge, West Norfolk Ridge, North Norfolk Basin, South Norfolk Basin, Vening-Meinesz Fracture Zone, North Norfolk Plateau, Forster Basin, Nepean Saddle, Kingston Plateau, Bates Plateau, Philip Trough.

The geomorphology in the EMR surrounding Norfolk and Philip Islands is dominated by north-south trending rugged volcanic ridges separated by basins partly filled by sediments. The major feature is the Norfolk Ridge which bisects the region (Fig. 3.54). Apart from a slight western offset north of Norfolk Island the ridge runs straight north-south. It is 750 km in length from the northern boundary of the EMR to the southern boundary. Its width varies from 80 to 100 km and is mostly flat-topped at depths shallower than 1,500 m, with steep sides sloping into the New Caledonia Basin on the west and Norfolk Basin on the east (Figs. 3.55 and 3.56). Numerous small canyons cut these slopes. There are three shallower plateaus along its length. One in the north is 150 x 50 km and 500-1,000 m water depth, another around the islands is 200 x 50 km and rises steeply from 500 m water depth to a planated basalt surface at about 75 m, and the third at the southern end is 100 x 50 km, trends NW-SE, rises to 300-600 m water depth and is known as the Reinga Ridge. The orientation of this southern block is north-west as its NE margin is part of the Vening-Meinsez Fracture Zone. A 2,000 m deep narrow (20 km) gap separates this ridge from the Wanganella Bank (water depth 100-1,000 m) on West Norfolk Ridge in the south of the EMR.

To the west of Norfolk Ridge is the N-S trending New Caledonia Basin, an area of very flat seabed 3,500 m deep at the northern margin where it is 150 km wide and shoaling and narrowing to the south where it is 2,500 m deep (Fig. 3.56b). The West Norfolk Ridge forms the southern margin of the basin. This ridge continues NNW across the EMR as a discontinuous bathymetric feature known as the Northern West Norfolk Ridge and separates the Caledonia Basin on its east from the 100 km wide, 3,000 m deep Fairway Basin on its west (Exon et al., 2007). This ridge is probably volcanic and consists of steep sided ridges and seamounts with steep relief of 1,000 m where it rises from the 3,200 m abyssal floor of the New Caledonia Basin (Fig. 3.57).

The Fairway Basin is also N-S trending in the EMR. It is also 150 km wide and slopes to the south from 3,000 m in the north to 3,200 m in the south. The seabed is generally flat in the south of the basin but in the north there are ridges and hills with 100 m of relief and the seabed slopes east (Fig. 3.56a, Exon et al., 2007). These hill and ridges on the seabed appear to have been formed by underlying igneous intrusions and possibly sedimentary diapirs. The EMR extends across the Fairway Basin to take in the lower slopes of the eastern margin of the LHR. The contact of the LHR with the flat abyssal plain is abrupt and the lower slopes are steep, as the sediment cover has draped over the underlying normal-fault scarp formed during rifting (Willcox et al., 1980).

East of the Norfolk Ridge is a zone of complex topography on the seafloor known generally as the Norfolk Basin, but composed of plateaus, seamounts, basins, depressions and fault-bounded troughs (Fig. 3.55). The extent of these geomorphic features has only recently been revealed by multibeam surveys over the central part of the basin (Bernardel et al., 2002, Sdrolias et al., 2004).

Along the northern boundary of the EMR is the North Norfolk Plateau 2,000-3,000 m deep. South of this plateau is the North Norfolk Basin 200 x 100 km and 3,000–3,500 m deep separated from the deeper (4,000–4,250 m) South Norfolk Basin by the Nepean Saddle at 30°S. The basin floors are flat due to sediment fill of several hundreds of meters, thickest at the margins and base of ridges (Eade, 1988). Numerous volcanic ridges, generally oriented NNE-SSW, protrude from the basin floor with varying relief.

East of the North Norfolk Basin and separated from it by a ridge is the deeper Forster Basin (>4,000 m) which continues its NE trend outside the EMR to the Cook Fracture Zone. A large seamount in the Foster Basin has a summit only 800 m below the sea surface. Evidence of a submarine eruption was reported from this area in 1981 (RAN Hydrographic Office chart). South of the Forster Basin is the Kingston Plateau, a large, irregular shaped area approximately 200 x 150 km and between 2,000 and 3,000 m water depth. This plateau has many seamounts, some rising to within 1,000 m of the sea surface. Along its southern margin a ridge rises to water depths of 400-600 m. It is joined to the Norfolk ridge at its south western margin by the 50 km wide Nepean Saddle that is 900-1,400 m deep. Numerous small seamounts on the Nepean Saddle have a linear E-W trend at approximately 290 40'S and may correspond to a fault at this location (Sdrolias et al., 2004).

Southeast of the Kingston Plateau is the Bates Plateau. They are separated by the NNE trending Philip Trough, 3,000 to 3,500 m deep and 20-50 km wide with an irregular boundary. The margins of the plateaus are steep with slopes up to 270 (Sdrolias et al., 2004). Bates Plateau is similar to the Kingston Plateau having a relatively smooth surface at water depths of 2,100–2,600 m with low relief crenulations and the general absence of abyssal hills or ridges. However, they do have many seamounts, some are part of linear volcanic ridges along the margins of the plateaus. The eastern margin of the Bates Plateau has a N-S chain of seamounts in the EMR with three of them rising to depths of 800, 840 m and 570 m respectively (DiCaprio et al., in press). South of Bates Plateau and separated form the South Norfolk Basin by a ridge is an un-named group of small basins deeper than 4,000 m and separated by ridges.

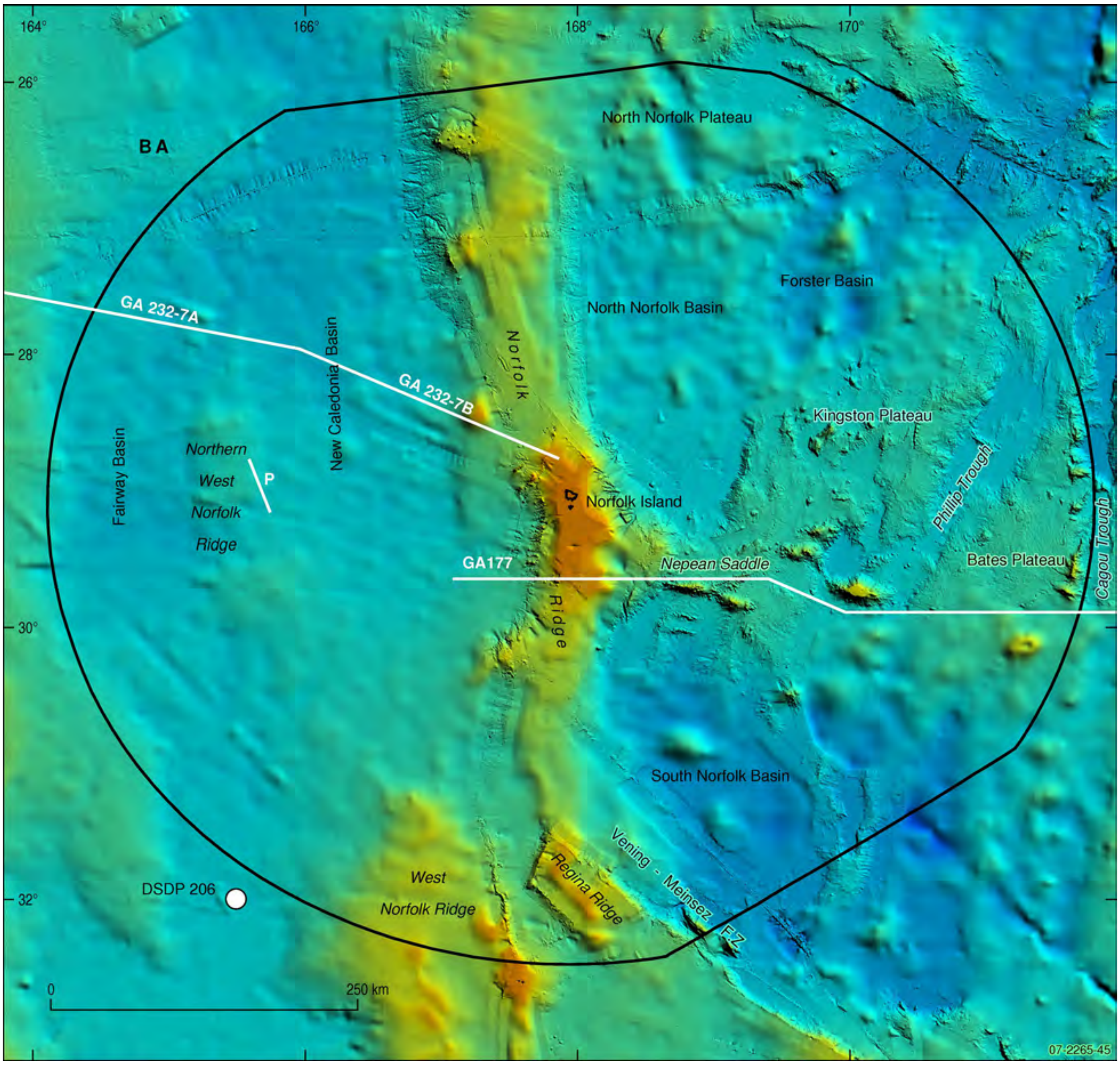


Figure 3.54. False colour image showing the geomorphic features around Norfolk Island. Locations of DSDP drill sites and of features and seismic lines displayed in figures 3.55 and 3.58 are marked. Black line is the EMR boundary.

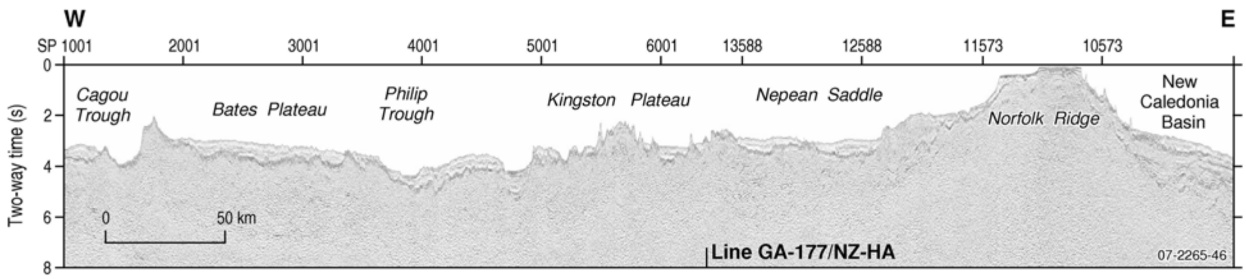


Figure 3.55. Seismic profile from the Norfolk Ridge to the eastern margin of the EMR showing the slope up from the New Caledonia Basin and the rugged topography of the ridges, troughs and plateaus. Location in Figure 3.54. Alcock *et al*., (2006).

3.8.2. Surface Sediments and Rocks

Sediments in this area are dominated by pelagic carbonate consisting of the calcite remains of foraminifers and coccoliths. There is also a minor contribution from siliceous plankton (radiolarians and diatoms). Volcanic ash and pumice are the minor contribution forming non-biogenic particles. Near rock outcrops boulders, blocks and pebbles of rock fragments become common. A list of cores and other samples from the area is given in Appendix C Table 3.21.

The New Caledonia and Fairway Basins are blanketed in a pale brown to light grey foraminifer-rich coccolith pelagic ooze containing minor amounts (<2%) of volcanic glass, radiolarians and diatoms (Dickens et al., 2001; Exon et al., 2004c). The New Caledonia Basin was sampled at DSDP Site 206 just outside the southwestern margin of the EMR (Burns et al., 1973). Sedimentation rates of ~2.5 cm ka^{-1} were calculated by Exon et al. (2004a). Little sediment sampling has been done from the more rugged areas on and to the east of Norfolk Ridge, but it is likely that foraminifer coccolith ooze dominates with redeposited sediments at the base of slopes. Air fall and submarine volcanic ashes and pumice may be mixed with the sediment in some areas.

Sedimentary diapirs occur in the subsurface of the Fairway Basin in the northwest of the EMR and are thought to be shale or possibly even salt (Auzende et al., 2000). Some diapirs have intruded through 3 to 4 km of sedimentary section and have raised the seabed. The diapirs are associated with a Bottom Simulating Reflector (BSR) which occurs at depths of 520-600 m below the seabed (Exon et al., 1998). This has led to speculation that an extensive area of gas hydrates and free methane gas may occur in this area.

Basement ridges occur along both sides of the Norfolk Ridge. They are strongly faulted and often outcrop or are only thinly covered by sediment (Eade, 1988). The ridge on the west is wider than the one on the east and is believed to be volcanic. Between the ridges the sediments are at least 3 km thick (Eade, 1988). The only sediment core from the Norfolk Basin was taken in 4,000 m water depth and contains 66% carbonate and consists of a foraminiferal coccolith ooze with some diatoms and radiolaria (Baker et al., 1988a and b).

DiCaprio et al. (in press) have collated the results of dredging in the region. A wide range of volcanic and metamorphic rocks have been recovered, most encrusted with manganese. Figure 3.56b shows a large volcano that is part of the chain of volcanoes along the western side of Norfolk Ridge. It rises to within 900 m of the sea surface and has a caldera at least 17 km across (Exon et al., 2004b). It has sediment drifts on its top that indicates the possible presence of foraminiferal sand winnowed by currents.

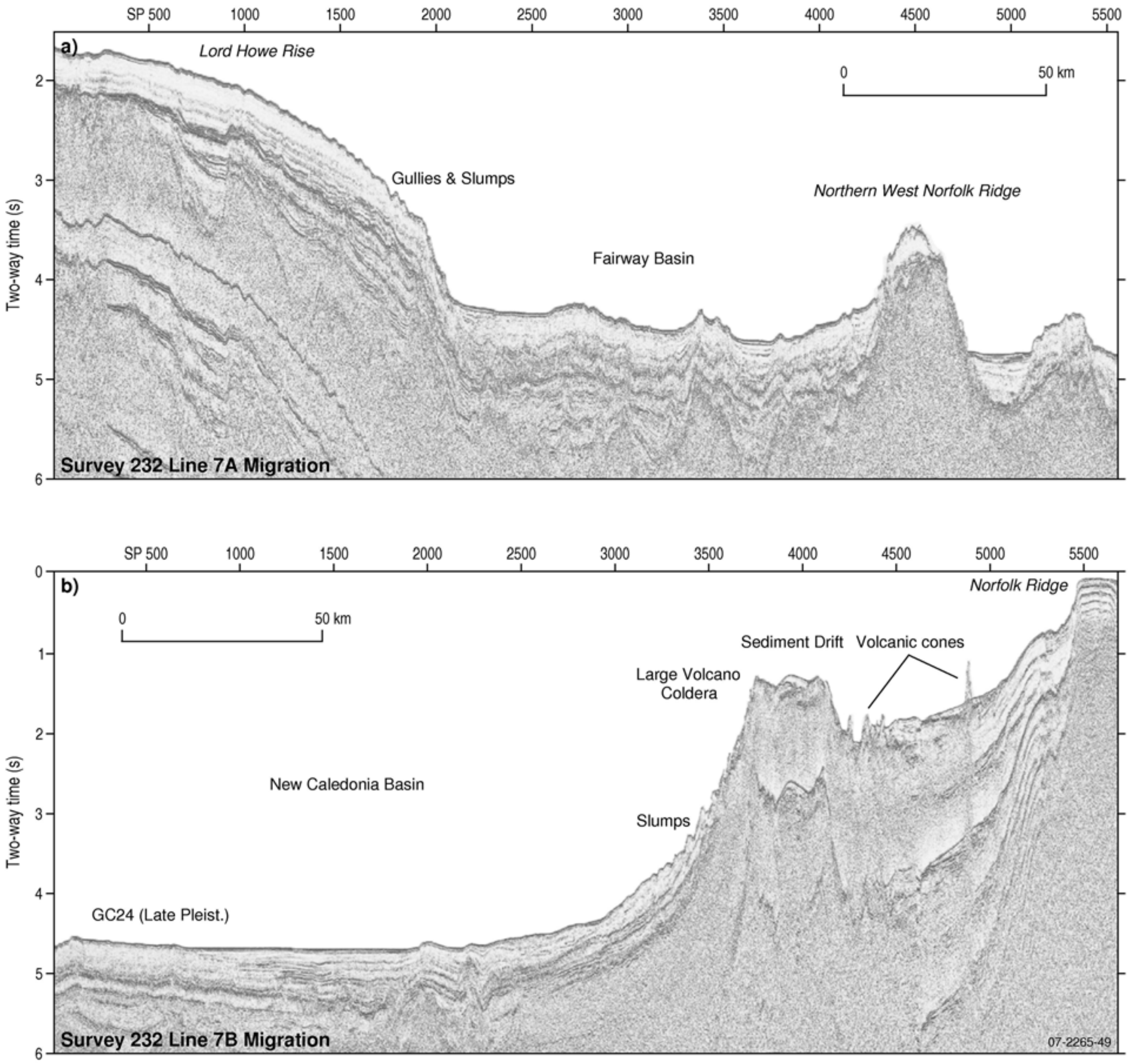


Figure 3.56.a)Seismic profile from Lord Howe Rise across the South Fairway Basin to the Northern West Norfolk Ridge showing hills in the basin and steep ridges and b) Seismic profile from the New Caledonia Basin to Norfolk Ridge showing volcanic cones, sediment drifts and slumping. Lines 232-7A and 7B, Exon et al., (2004b).

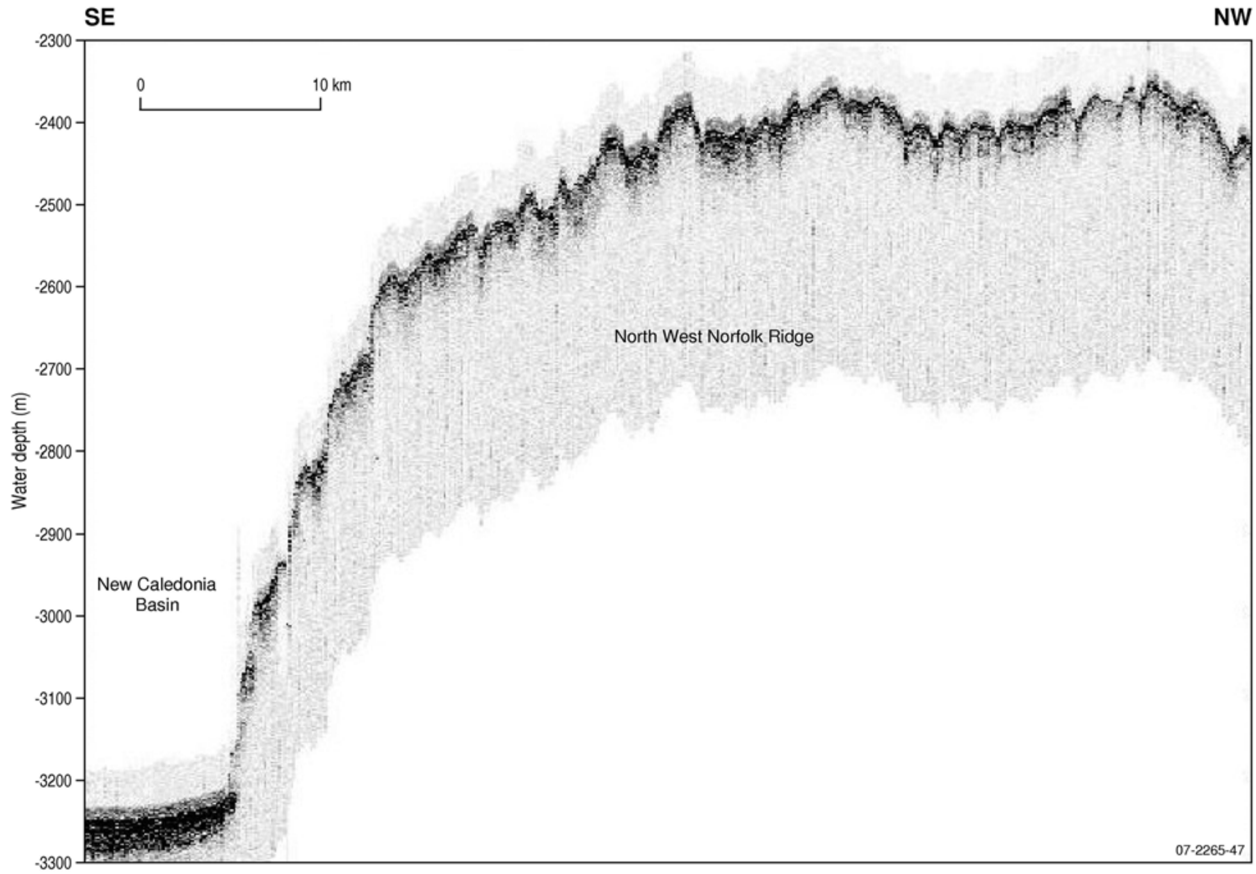


Figure 3.57. Profile of North West Norfolk Ridge showing rough topography on ridge, steep lower slope and flat floor of the Fairway Basin. Location in Figure 3.54. Colwell, (2006).

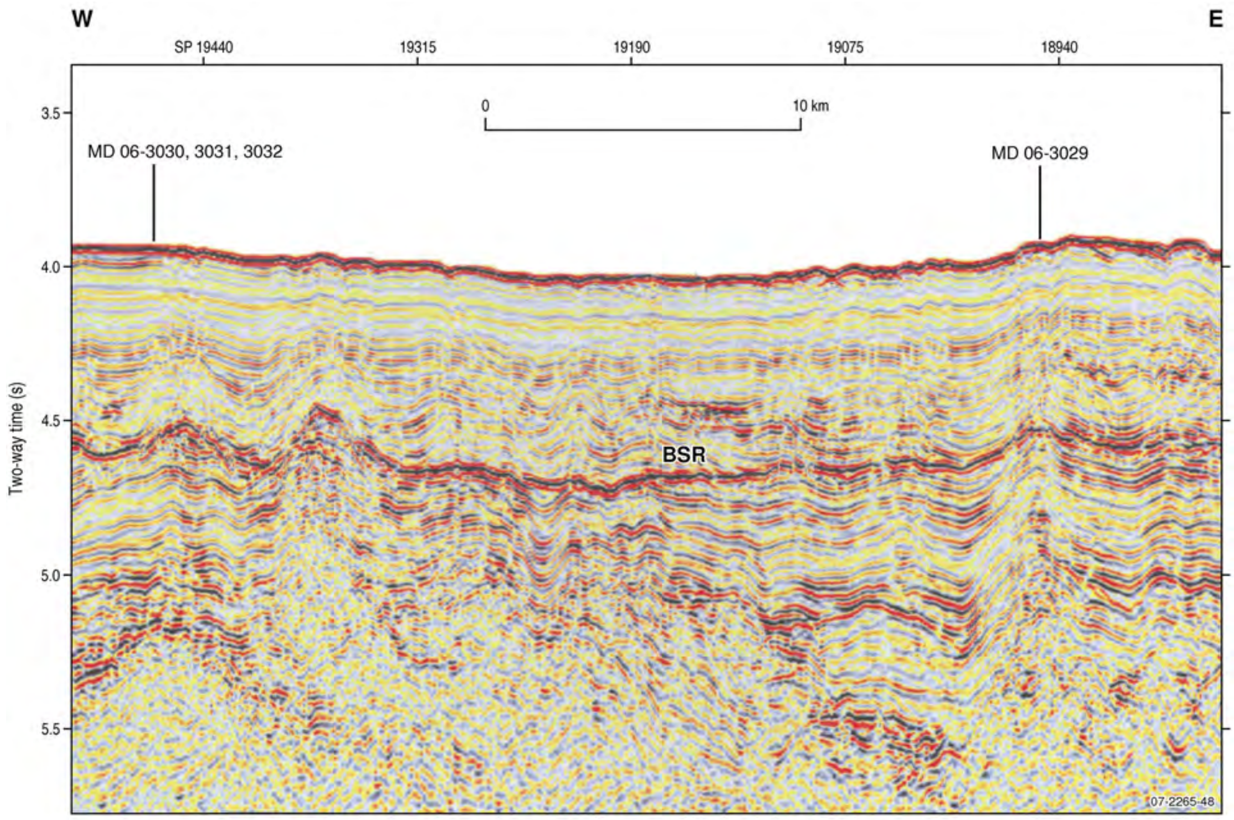


Figure 3.58. Seismic profile at eastern flank of central Fairway Basin, showing the distinctive bottom simulating reflector (BSR) and location of cores. GA line 177/LHRNR-BA, Location in Figure 3.54. Colwell et al., (2006)

4. Quantitative Description of the EMR

4.1 GEOMORPHOLOGY

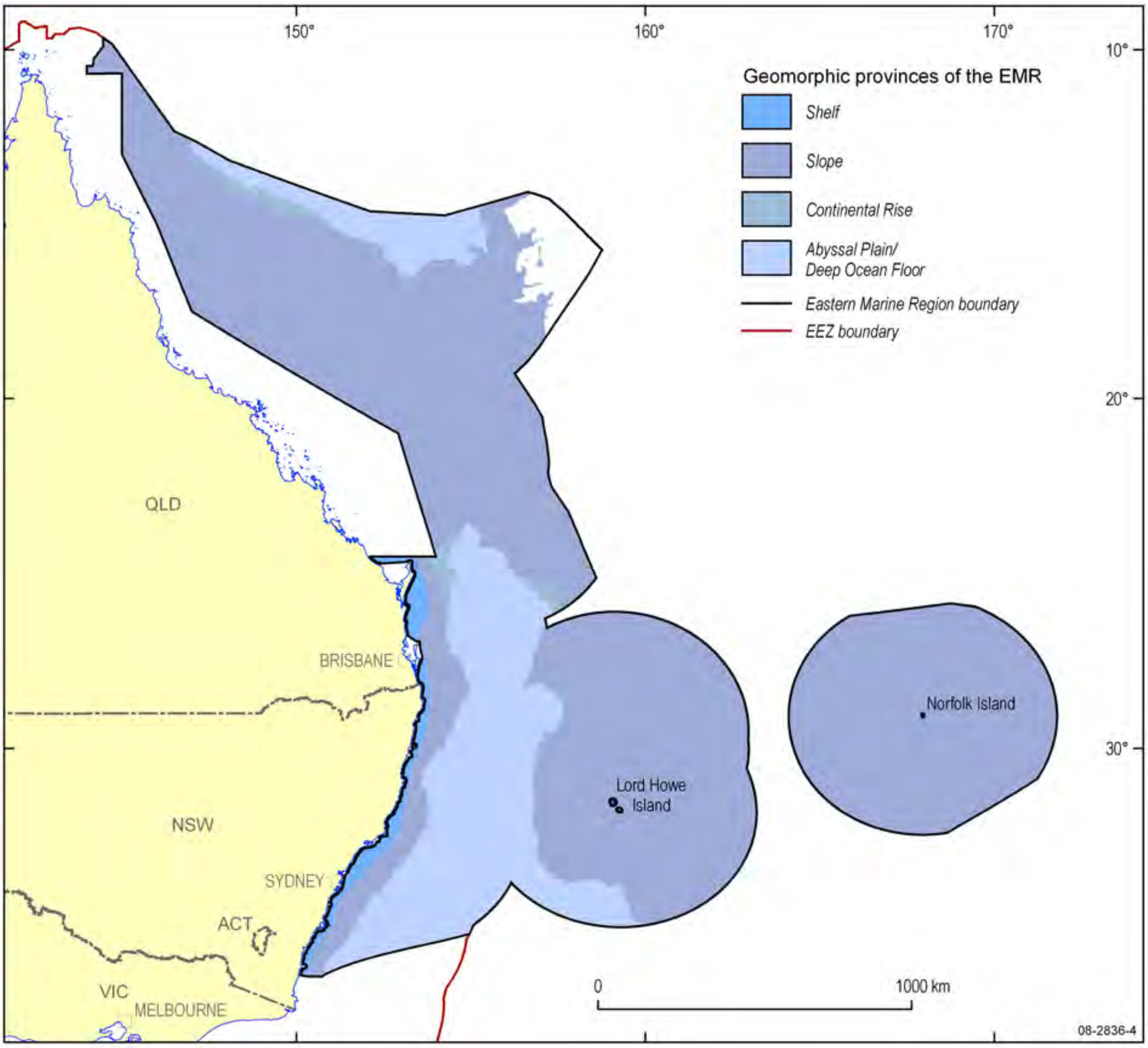
Four geomorphic provinces occur in the EMR (Fig. 4.1; Table 4.1). Slope makes up the largest area (77%, 1,837,670 km^2), followed by abyssal plain/deep ocean floor (20%, 485,660 km^2), shelf (2%, 38,080 km^2), and rise (1%, 30,430 km^2) (Table 4.1). Relative to the rest of Australia's EEZ the EMR has a significantly larger percentage of slope, and far lower percentage of shelf. The EMR contains approximately 31% of area of slope in the entire AEEZ (Fig. 4.1; Table 4.1).

Of the 21 geomorphic features defined on the Australian margin, 18 are represented in the EMR. Tidal sand wave/sand banks and escarpments are not represented (Fig 4.2; Table 4.1).

Large areas of the shelf, slope, rise and abyssal plain/deep ocean floor in the EMR have no geomorphic features identified within them. These areas comprise 26% of the total EMR area (shelf = 1%, slope = 9%, continental rise = 1% and abyssal plain/deep ocean floor = 15%). Geomorphic features covering significant areas of these provinces with quantitative data include basins, deepwater trenches/troughs, shallow and deep water terraces and plateaus, which combined comprise 1,534,590 km^2 or 62% of the area of these provinces in the EMR. Geomorphic features with no quantitative data available include apron/fans, deep/hole/valleys, canyons, knoll/abyssal hills/mountains/peaks, saddles, pinnacles, reefs and seamounts/guyots, which together comprise 200,250 km^2 or 8% of the area of these provinces in the EMR.

The EMR contains a large proportion of the total area of several geomorphic features over the EEZ. Relative to the entire EEZ, the EMR contains relatively large areas of plateaus, saddles, basins and trench/troughs. Plateaus in the EMR cover 1,027910 km^2 or 69% of the total area of plateaus in the EEZ, followed by saddles (94,610 km^2; 65%); basins (366,190 km^2; 51%); and trench/troughs (82,160 km^2; 47%) (Fig 4.2; Table 4.1).

a)



150°
160°
170°
10°
20°
30°
Geomorphic provinces of the EMR
Shelf
Slope
Continental Rise
Abyssal Plain/
Deep Ocean Floor
Eastern Marine Region boundary
EEZ boundary
QLD
BRISBANE
NSW
SYDNEY
ACT
VIC
MELBOURNE
Lord Howe
Island
Norfolk Island
0
1000 km
08-2836-4

b)

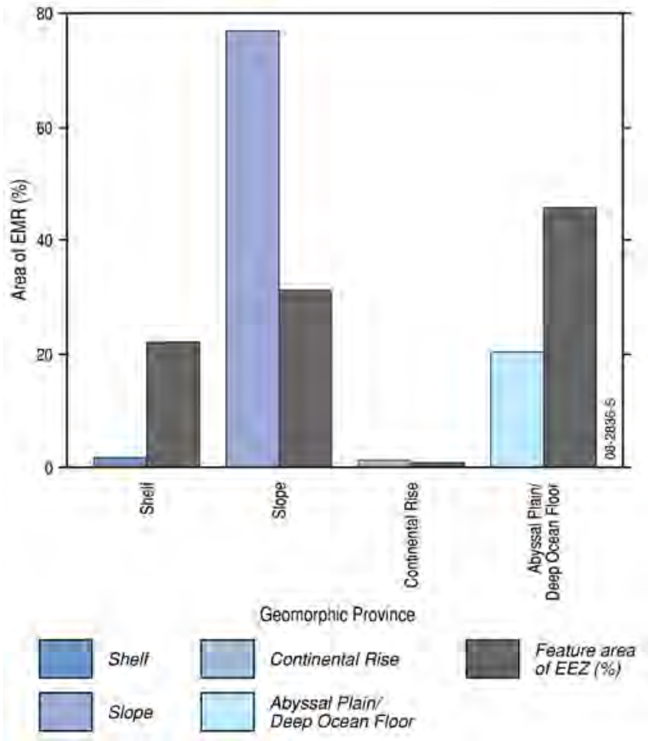
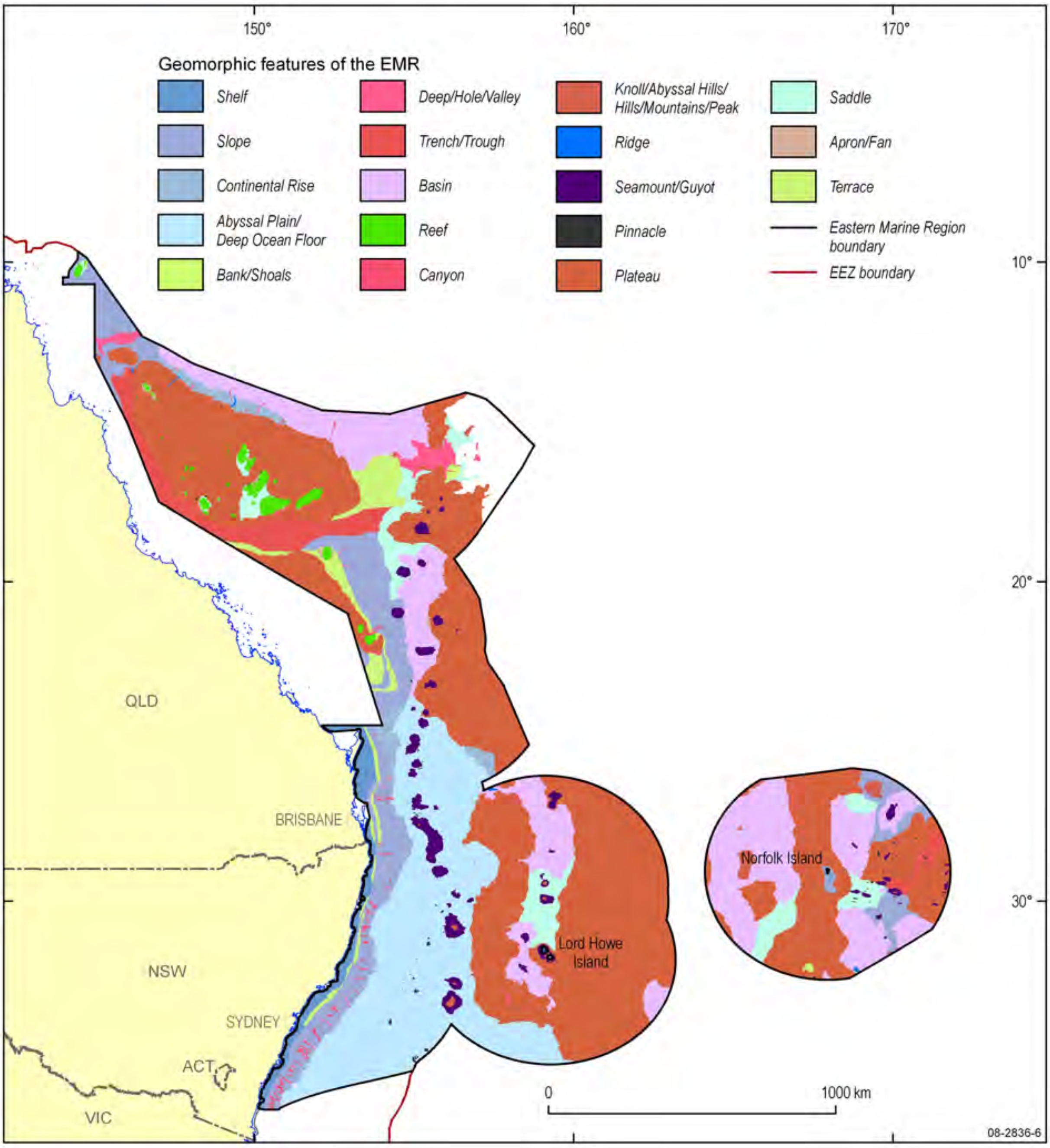


Figure 4.1. a) Geomorphic Provinces of the East Marine Region (EMR); and b) Percentage area of each geomorphic province within the EMR and EEZ.

a)



150°
160°
170°
10°
20°
30°
Geomorphic features of the EMR
Shelf
Slope
Continental Rise
Abyssal Plain/
Deep Ocean Floor
Bank/Shoals
Deep/Hole/Valley
Trench/Trough
Basin
Reef
Canyon
Knoll/Abyssal Hills/
Hills/Mountains/Peak
Ridge
Seamount/Guyot
Pinnacle
Plateau
Saddle
Apron/Fan
Terrace
Eastern Marine Region
boundary
EEZ boundary
QLD
BRISBANE
NSW
SYDNEY
ACT
VIC
Lord Howe
Island
Norfolk Island
0
1000 km
08-2836-6

b)

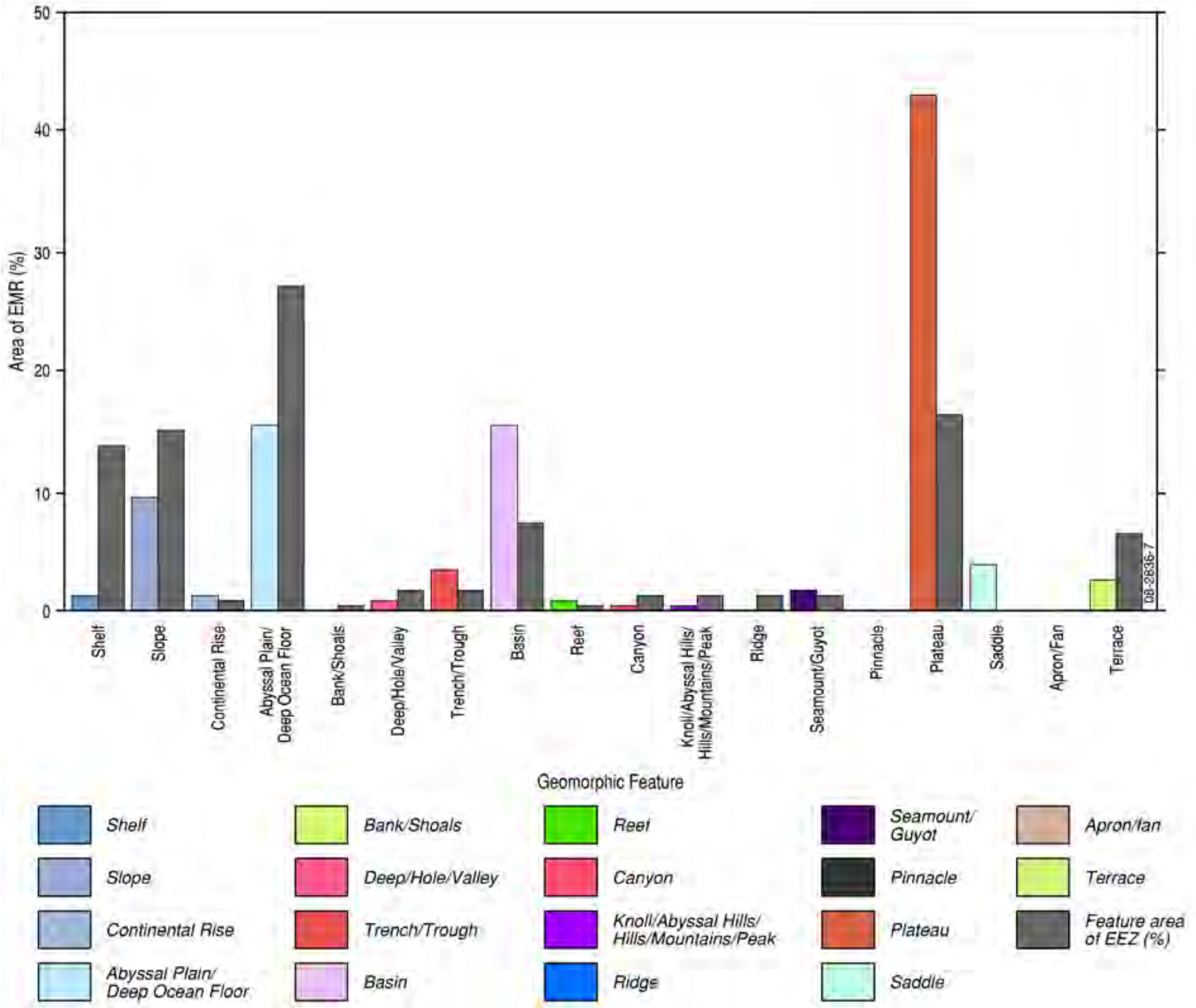


Figure 4.2. a) Geomorphic Features of the East Marine Region (EMR); and b) Percentage area of each geomorphic feature within the EMR and EEZ.

Table 4.1. Statistics of geomorphic provinces and features of the EMR.

Feature	Area in EMR	% total* EMR Area	% EEZ Area	Proportion of the total EEZ area of this geomorphic feature that is located in EMR	Water Depth Range** in EMR (m)
Geomorphic Provinces					
Shelf	38,080	1.59	21.91	1.39	2 - 250
Slope	1,837,670	76.83	31.31	48.40	230 – 5,270
Rise	30,430	1.27	1.08	30.16	1,925 – 4,665
AP/DOF*	485,660	20.30	45.71	20.15	125 – 5,270
Geomorphic Features					
Shelf (unassigned)	34,110	1.39	13.69	2.75	2 - 255
Slope (unassigned)	221,760	9.03	15.17	16.15	25 – 5,270
Rise (unassigned)	29,180	1.19	1.11	28.92	2,480 – 4,655
AP/DOF* (unassigned)	370,130	15.07	26.03	15.70	1,585 – 5,270
Apron/fan	2,650	0.11	0.07	40.10	15 – 3,195
Bank/shoal	710	0.03	0.56	1.39	205 – 1,420
Basin	366,190	14.91	7.89	51.26	260 – 4,955
Deep/hole/valley	18,870	0.77	1.88	11.07	85 – 4,445
Canyon	9,820	0.40	1.78	9.22	85 – 4,885
Knoll/abyssal hills/hills/ peak	9,690	0.40	1.24	8.63	4 – 4,480
Saddle	94,610	3.85	1.61	64.62	55 – 3,875
Pinnacle	2,470	0.10	0.06	44.73	75 – 5,055
Plateau	1,027,910	41.84	16.45	69.04	90 – 5,140
Reef	20,010	0.81	0.54	41.12	230 – 2,290
Ridge	1,110	0.05	1.30	0.94	960 – 4,070
Seamount/guyot	42,130	1.72	1.11	41.78	25 – 5,150
Terrace	58,330	2.37	6.37	10.12	50 – 4,540
Trench/trough	82,160	3.34	1.94	46.78	480 – 3,900
TOTAL	*4,783,680*				

* AP/DOF = Abyssal plain/deep ocean floor.
** Does not include areas designated as land and water shallower than 10 m totalling 7,300 km².

4.2. BATHYMETRY

Water depths in the assessed area of the EMR range from 230 to 5,269 m (Fig. 4.3). The EMR is relatively deep with >80% of the total area in water depths between 1,000 m and 5,000 m, although water depths >5,000 m comprise <1% of the EMR (Fig. 4.3). Areas with shallow water depths (<500 m) cover less than 6% of the EMR and comprise 3% of the total EEZ area for these water depths. This depth distribution reflects the major slope and abyssal plain/deep ocean floor provinces that occur along much of the eastern Australian margin.

Some features in the EMR occur in water depths at which they are not commonly found elsewhere in the EEZ. Compared to occurrences elsewhere on the Australian margin, reefs in the EMR occur in greater water depths (230- 2,290 m) (Table 4.1). Across the entire EEZ, reefs occur mainly in water depths <500 m, and all reefs in water depths >500 m are located in the EMR (Fig. 4.4). Across the entire EEZ, deep/holes/valleys, basins, and pinnacles occur at a range of water depths. However, in the EMR these occur mainly at water depths >3,000 m and occasionally include areas with depths > 5,000 m. (Fig. 4.4; Table 4.1). Approximately 65% of the area of deep/hole/valleys in the EEZ in water depths between 2,000 and 4,000 m occurs in the EMR. Approximately 25% of the area of basins in the EEZ in water depths between 3,000 and 5,000 m occurs in the EMR. More than 35% of the area of pinnacles in the EEZ in water depths between 2,000 and 5,000 m occurs in the EMR.

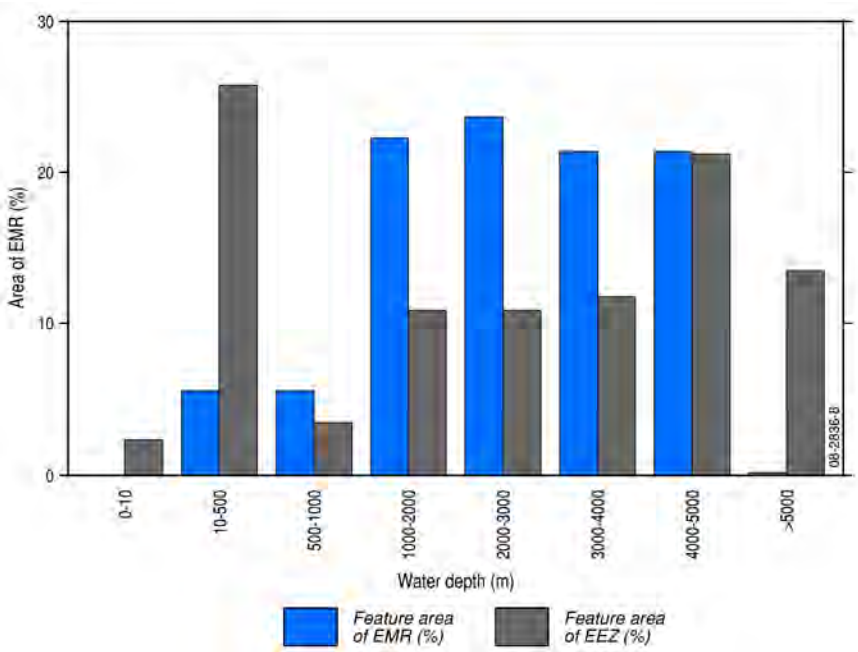
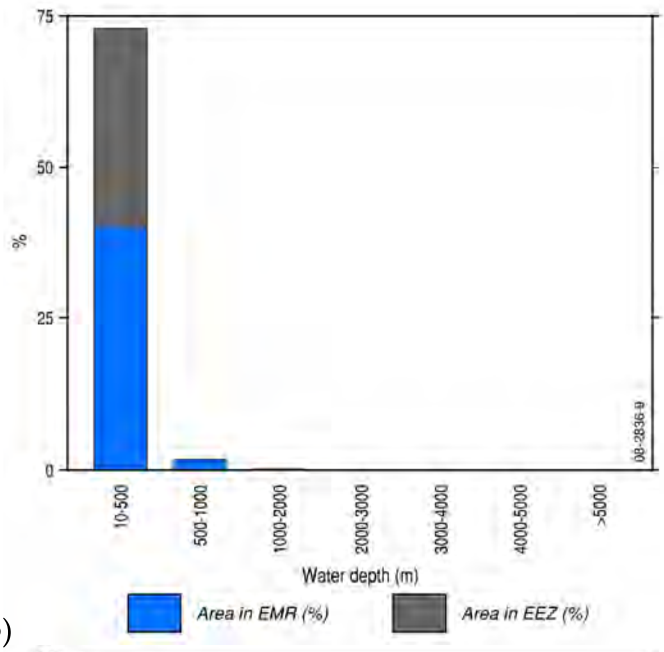
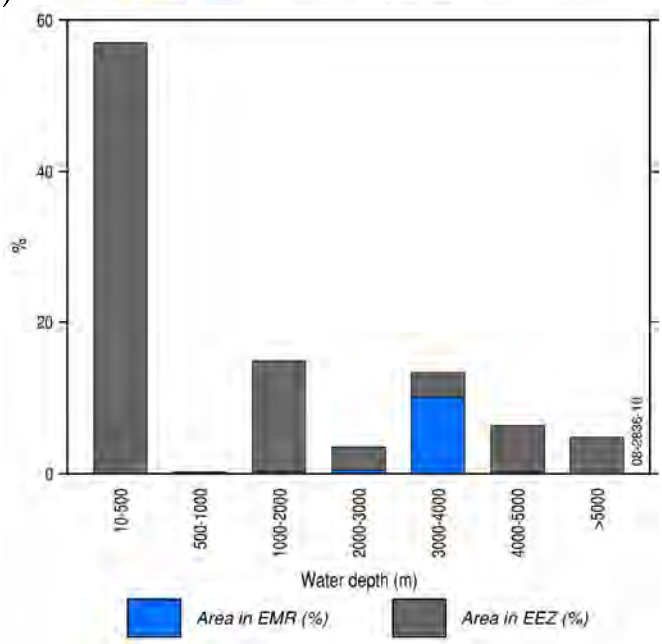


Figure 4.3. Distribution of water depth classes by percentage area within the East Marine Region in comparison to water depths of the whole EEZ.

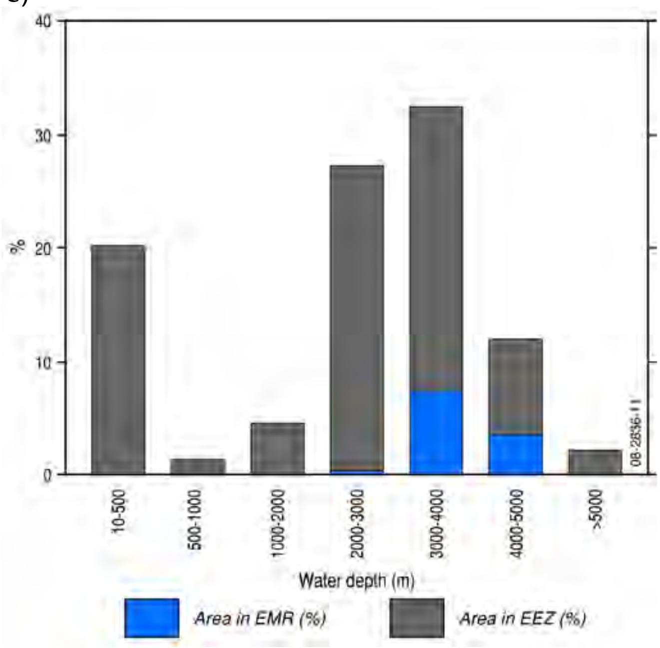
a)


75
50
25
0
%
08-2836-9
10-500
500-1000
1000-2000
2000-3000
3000-4000
4000-5000
>5000
Water depth (m)
Area in EMR (%)
Area in EEZ (%)

b)


60
40
20
0
%
08-2836-10
10-500
500-1000
1000-2000
2000-3000
3000-4000
4000-5000
>5000
Water depth (m)
Area in EMR (%)
Area in EEZ (%)

c)



d)

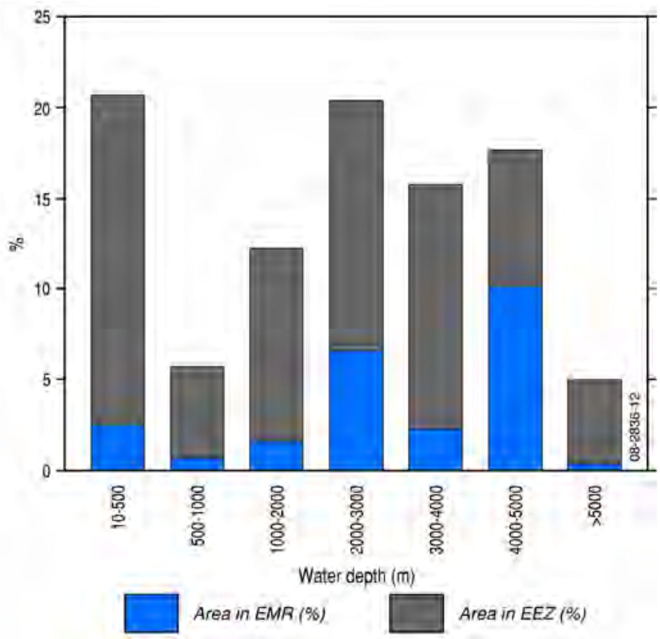


Figure 4.4. Distribution of water depths for a) reefs, b) deep/hole/valleys, c) basins and d) pinnacles in the EMR and the contribution of these to the total area of each feature in the EEZ.

4.3. QUANTITATIVE DESCRIPTION OF SEDIMENT DATA COVERAGE IN THE EMR

4.3.1. Quantitative Textural and Compositional Data

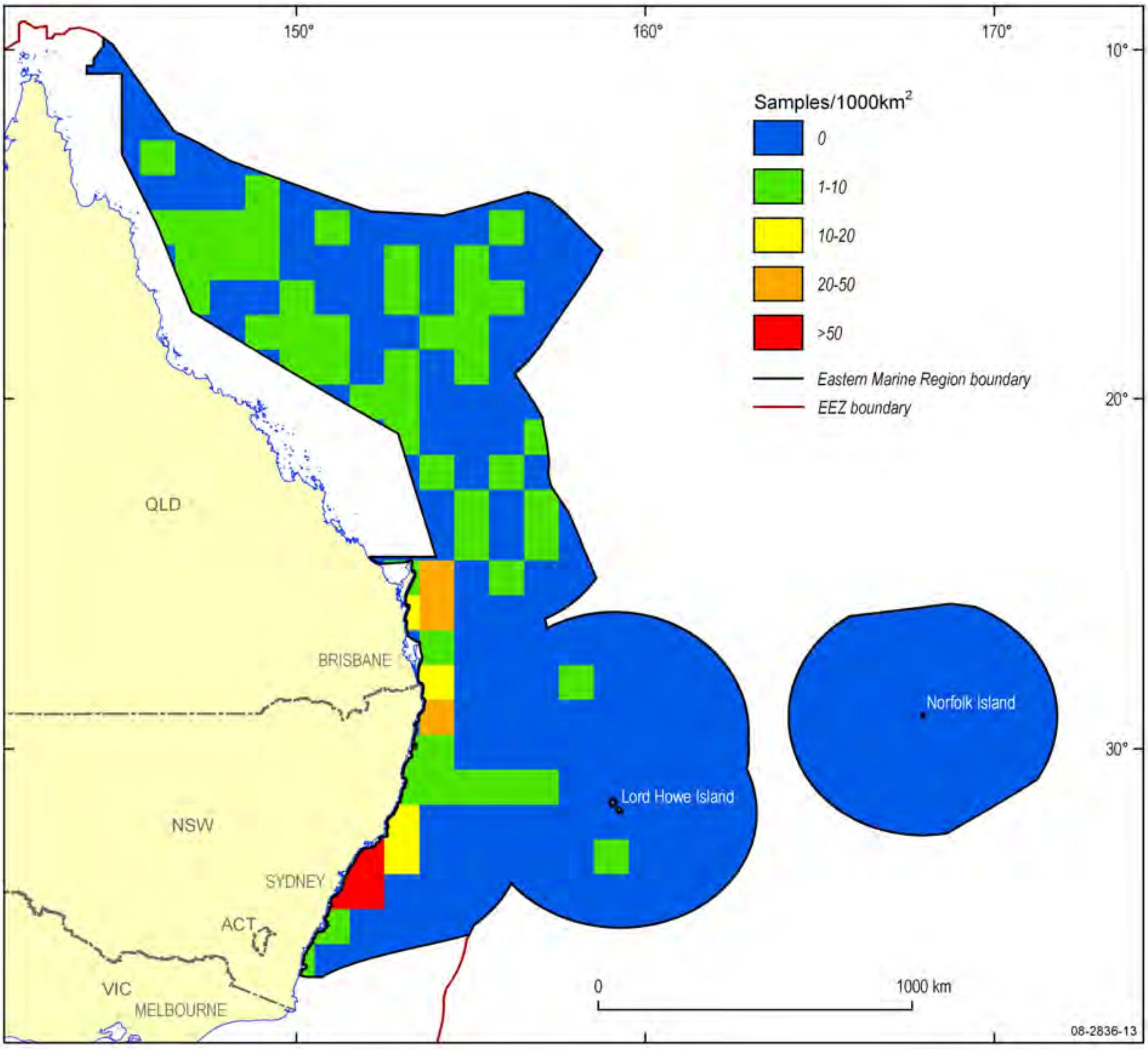
Sample density varies significantly across the EMR (Fig. 4.5). Sample density exceeds 10 samples per 1,000 km² for approximately 4% of the total area of the EMR. Sample density does not attain 1 sample per 1,000 km² for approximately 73% of the EMR (Fig. 4.5). Samples are clustered as a result of collection on surveys in shallow water areas (shelf and upper slope) or targeting significant topographic features offshore. In general, sample coverage is most dense in the southeast of the EMR where the boundaries include a narrow area of shelf (Fig. 4.6). Samples are very sparse in deepwater areas which cover the majority of the EMR, particularly in the southeast (Fig. 4.7).

A total of 539 samples (62% of samples) occur clustered on the relatively small area of shelf (38,080 km², <2% of the EMR area) that occurs in the southwest along the EMR boundary. This results in a relatively high average density of approximately 15 samples per 1,000 km² for the shelf (Fig. 4.6; Table 4.2). A total of 140 (16%) samples occur on the slope resulting in an average density of approximately 0.2 samples per 1,000 km². A total of 14 (<2%) samples occur on the rise and abyssal plain/deep ocean floor. These provinces form approximately 515,000 km² (22%) of the EMR area and have an average sample density of <0.04 samples per 1,000 km² (Fig. 4.6). Samples achieve coverage considered sufficient to assess the sedimentology in 8 of the 18 geomorphic features present in the EMR. No samples were collected from bank/shoals, reefs, knoll/abyssal hill/hill/mountain/peaks, ridges, pinnacles and apron/fans. Together, these features cover approximately 36,640 km² (<2%). Less than three samples were collected from each of rise, deep/hole/valley, canyon and seamount/guyot features. Together, these features cover approximately 100,010 km² (4%) of the EMR (Table 4.1).

Average sample density exceeds 1 sample per 1,000 km² in only shelf (unassigned), and terrace features. These cover approximately 4% of the EMR. For other features containing adequate samples for analysis, highest sample densities were achieved for slope (unassigned), trench/troughs and canyons (densities >0.2 samples per 1,000 km²) (Table 4.2). Low numbers of samples and/or clustering of samples on some features mean that assays may not be representative of seabed properties for the entire feature across the EMR. Low numbers of samples may significantly affect results for rise, Deep/hole/valley, Canyon and Seamount/guyot features. Clustering may significantly affect results for canyons, ridges, and terraces. Sample coverage at depths >4,000 m was achieved for slope, basin, plateau and abyssal plain/deep ocean floor features.

Despite targeted addition of data points, coverage remains poor for some areas of the abyssal plain/deep ocean floor (1:22,080 km²) and rise (1:30,430 km²), and (<0.1 sample per 1,000 km) for many features, particularly those located in deep water. Additional data improved coverage of plateaus and trench/troughs occurring in up to 4,280 m water depth. However, data does not achieve coverage of apron/fans, knoll/abyssal hills/hills/peaks, pinnacles, ridges, canyons and saddles occurring in deepwater areas of the slope, rise and abyssal plain/deep ocean floor in the EMR. It is important to note that average densities and areas given for these will vary depending on the scale (Marine region/province/feature) at which density is being assessed).

a)



b)

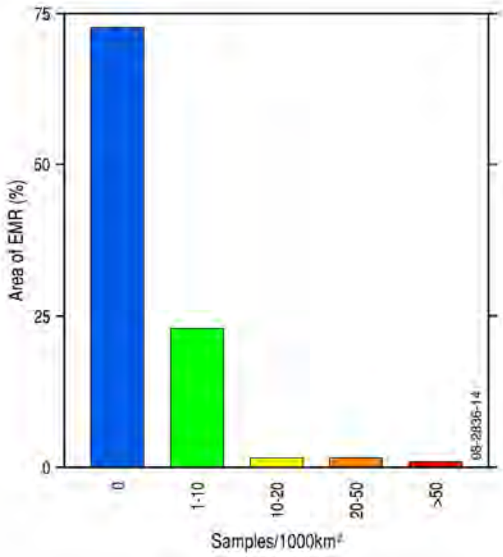


Figure 4.5. a) Sample density distribution across the EMR, and b) Frequency distribution of sample density.

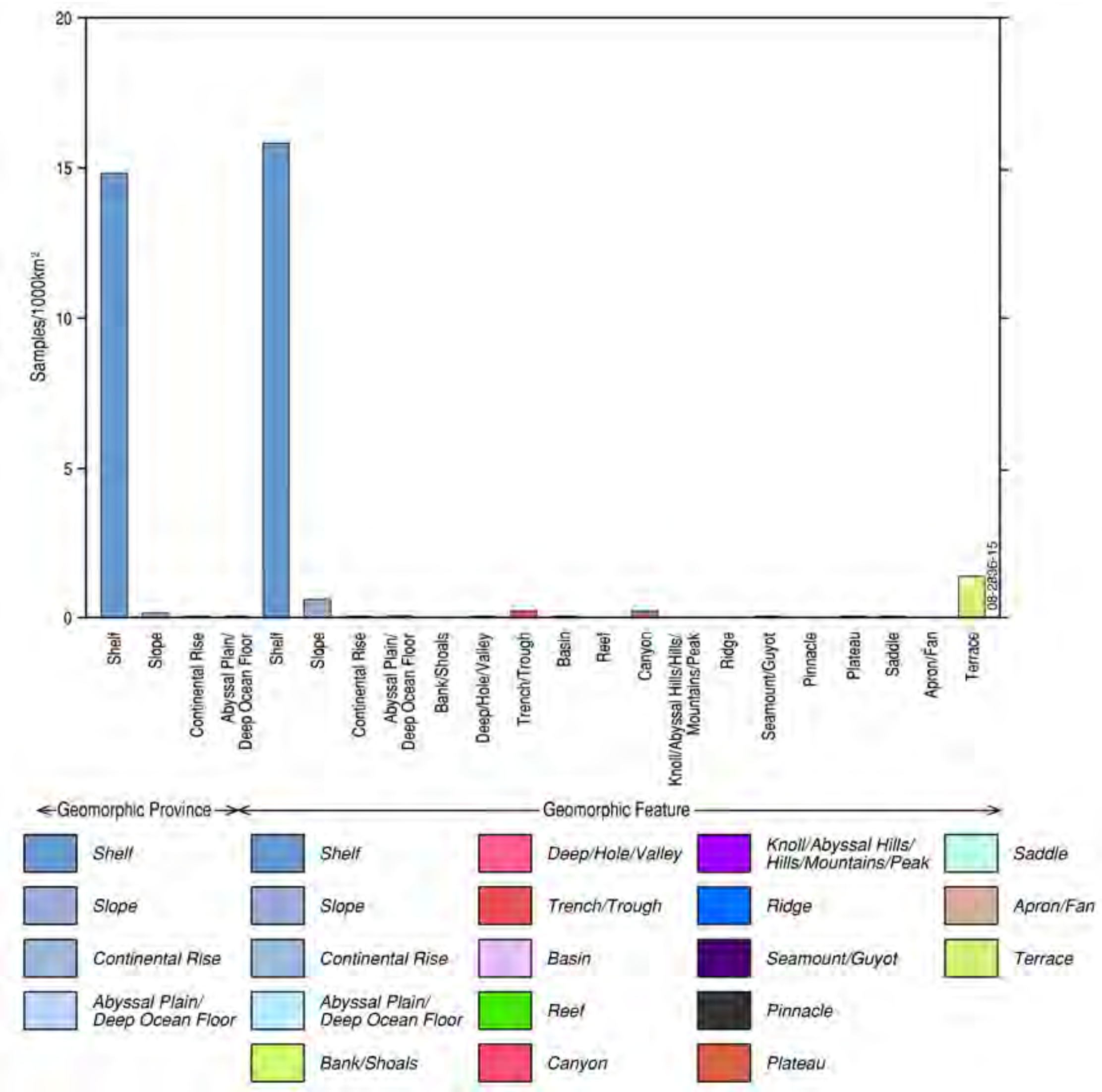


Figure 4.6. Sample density in each geomorphic province and feature of the EMR (y axis shows average density measured as samples per 1,000 km²).

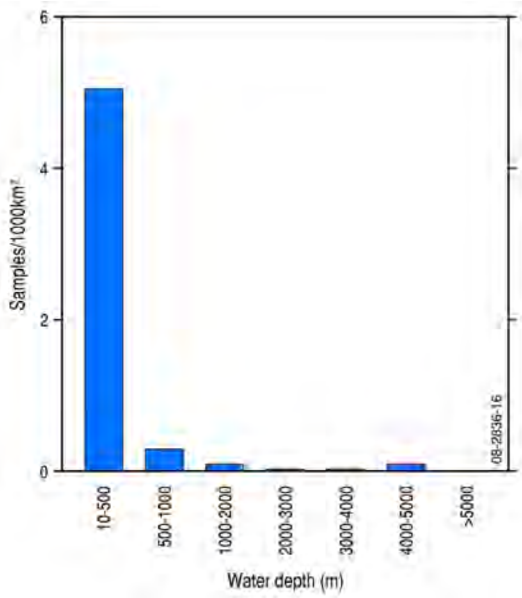


Figure 4.7. Sample density in each water depth class for the entire EEZ (sample density measured as samples per 1,000 km^2).

Table 4.2. Description of average density per geomorphic provinces and features containing samples.

PROVINCE/ # Feature	No. sample points	% EMR Area	Average sample density (samples per 1,000 km²)
Geomorphic Province			
Shelf	565	1.55	14.84
Slope	283	74.80	0.15
Rise	1	1.24	0.03
AP/DOF*	22 + 14 in deepwater outside EEZ	19.77	0.05
Geomorphic Features			
1 Shelf (unassigned)	539	1.43	15.80
2 Slope (unassigned)	140	9.27	0.63
3 Continental Rise (unassigned)	1	1.22	0.03
4 AP/DOF (unassigned)	13	15.47	0.04
5 Bank/Shoals	0	0.03	0.00
6 Deep/Hole/Valley	1	0.79	0.05
7 Trench/Trough	19	3.44	0.23
8 Basin	8	15.31	0.02
9 Reef	0	0.84	0.00
10 Canyon	2	0.41	0.20
11 Knoll/Ah/M/P	0	0.41	0.00
12 Ridge	0	0.05	0.00
13 Seamount/guyot	2	1.76	0.05
14 Pinnacle	0	0.10	0.00
15 Plateau	62	42.98	0.06
16 Saddle	4	3.96	0.04
17 Apron/Fan	0	0.11	0.00
20 Terrace	80	2.44	1.37

4.4 QUANTITATIVE REGIONAL SEDIMENT DISTRIBUTION IN THE EMR

4.4.1. Overview of Distribution and Properties

Sample assays indicate that the seabed in the EMR is characterised by a range of sediment types. The majority of samples are located on the shelf where sand is the most dominant size fraction. A total of 626 samples (83%) contain >50% sand, and 515 (70%) contain >80% sand (Fig 4.8). Only 19 samples contained <10% sand. Sand is most dominant on the shelf and upper slope.

A total of 83 samples (11%) contained >50% mud and 299 samples (40%) contained <10% mud. Mud is absent from 207 (28%) samples. Mud is the dominant size fraction on the slope and abyssal plain/deep ocean floor. Samples containing <10% mud occur less frequently on the shelf and upper slope, particularly offshore of Port Douglas and Bowen.

Gravel is detected in 406 (54%) samples, but is the dominant size fraction in only 19 (3%) samples. Gravel forms a minor component (<10%) in 316 samples (42%) and is absent in 345 samples. Gravel occurs most frequently on the slope and is generally absent in deep water areas. The abundance and distribution of sediment containing gravel is likely to be understated in the data due to sparse sample coverage of areas on the slope and rise.

Carbonate is the dominant constituent of sediments in the EMR with 452 (66%) of samples containing >50% carbonate, and 219 (32%) containing >80% carbonate (Fig 4.9). Carbonate forms <10% of sediment in 24 samples (4%) and is absent from three samples. Carbonate content generally decreases with increasing water depth and increasing distance from the coast. The highest bulk carbonate contents occur in sediment located within the shallow reaches of shelf and upper slope (Fig. 4.14). Bulk carbonate contents are consistently lower on the lower slope, rise and abyssal plain/deep ocean floor with around 73% of samples from these areas containing <50% carbonate. An exception to this is in the Coral Sea Basin, where carbonate content ranges from 16 to 86%.

All size fractions are dominated by carbonate grains (Fig 4.9). Carbonate content of mud was analysed for 87 samples and attains 50% in 78 (92%) of these. Carbonate content of mud generally shows variation over large distances with assays of similar concentrations observed to be clustered even at a regional scale. The majority of the samples (64) with mud containing >50% carbonate are located in plateau, slope and trench/trough features.

Carbonate is the dominant constituent of the sand sized fraction. Sand carbonate content was generated for 172 samples and attains >50% in 128 (74%) of these. The carbonate content of sand varies most in areas in close proximity to the coast, where carbonate content of sand is <25% at 24 sites. On the shelf and upper slope offshore Hervey Bay, Maroochydore and Sydney, highly variable carbonate content of sand is well documented by data.

Carbonate content of the gravel size fraction was assessed for 65 sites and attains 50% in 56 (86%) of these. Carbonate content of gravels <20% occur at 4 sites. These all occur on the Kenn Plateau within the Kenn Transition and Kenn Province. High carbonate contents of gravels (78 and 100%) generally occur on the shelf and upper slope. Gravel carbonate content decreases with increasing water depth and distance from the coast, with contents generally not exceeding 20% on the lower slope, rise and AP/DOF.

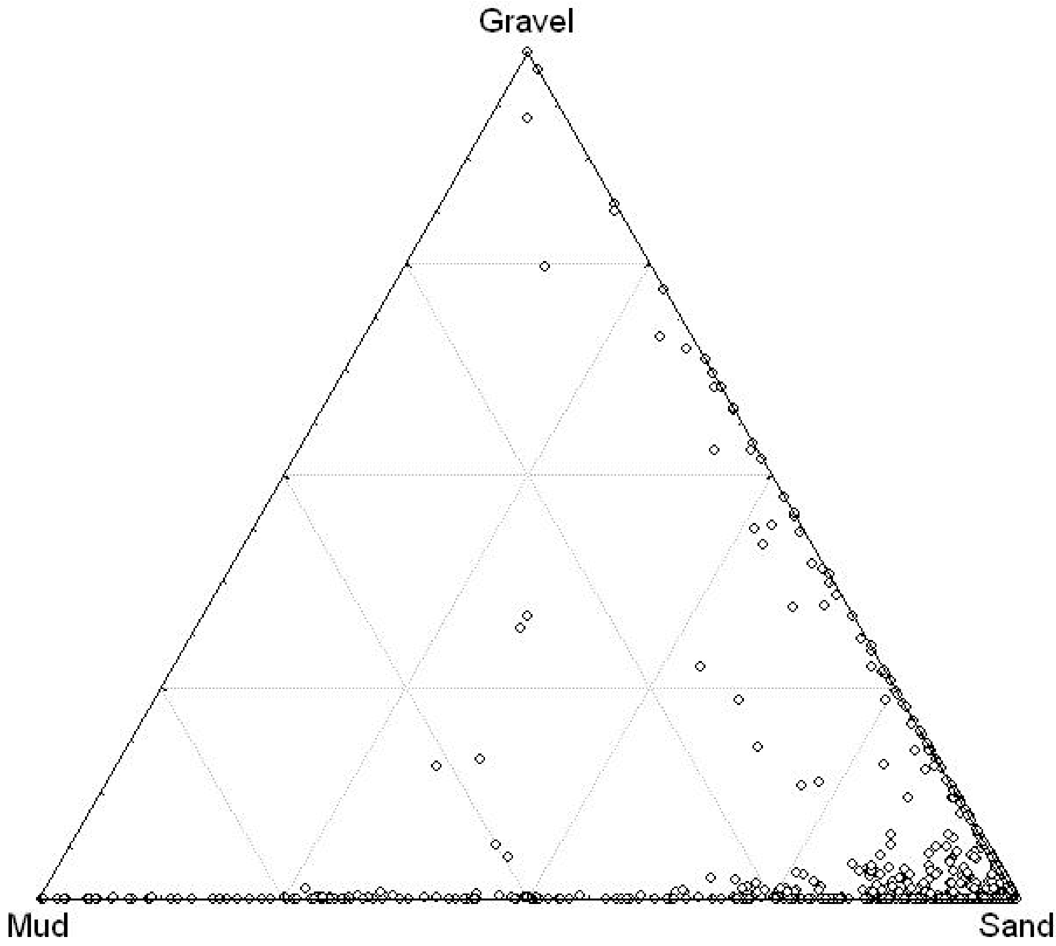
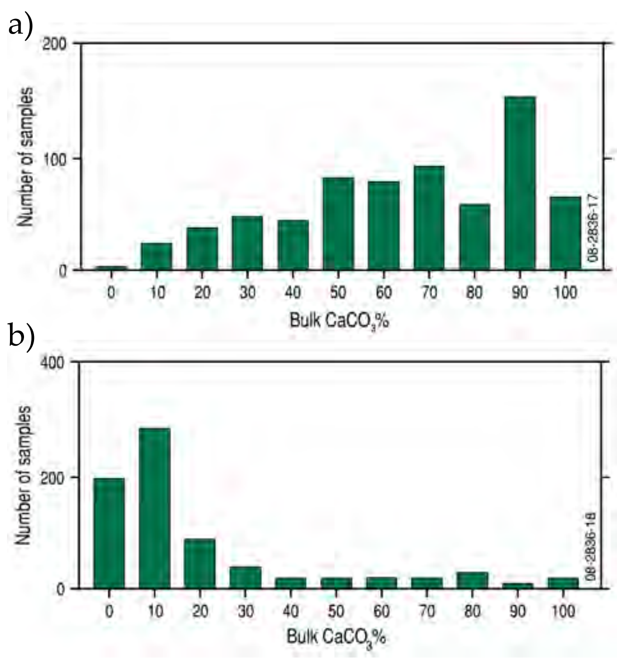
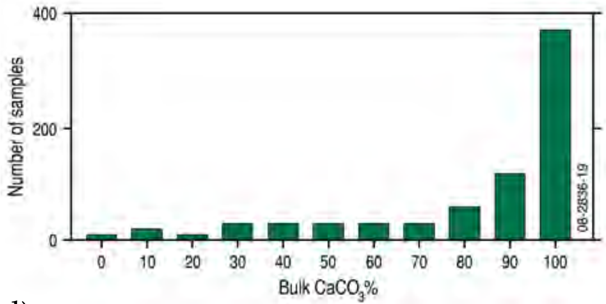


Figure 4.8. Textural composition (mud:sand:gravel ratio) of sediments within the EMR.



c)



d)

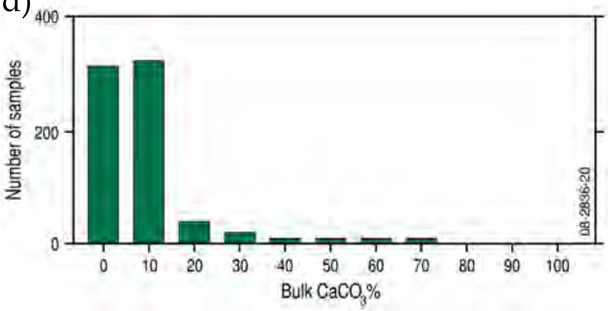


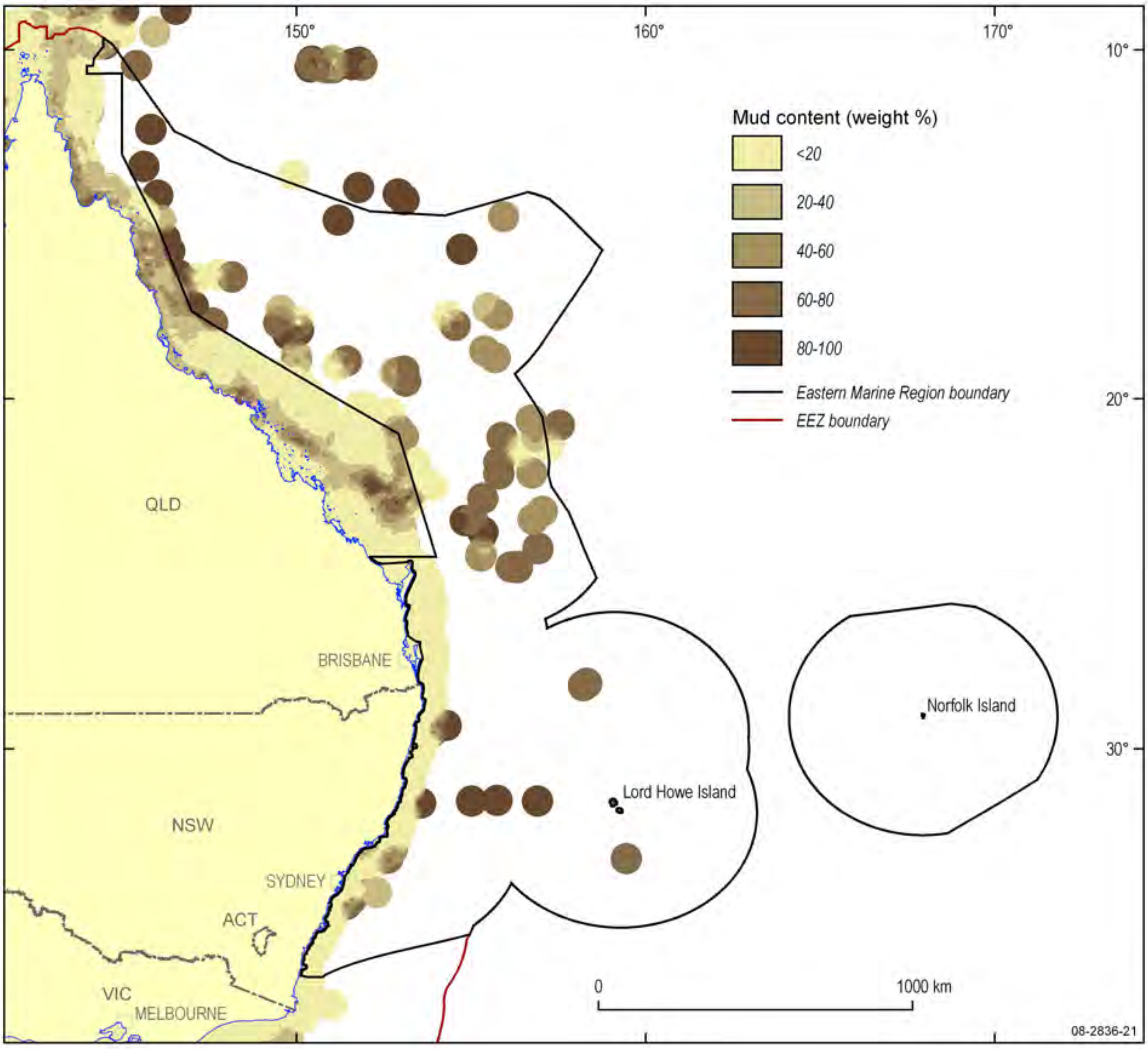
Figure 4.9. Carbonate content of the a) bulk fraction; b) mud fraction; and c) sand fraction; and d) gravel fraction of sediments in the EMR.

Sediment assays were interpolated using the methods described in Chapter 2 to give an estimate of regional distribution of sediment properties in the EMR. Interpolated grainsize data achieves coverage of around 454,800 km² (20%) of the total EMR (Figs. 4.10 - 4.12). Uneven distribution of data points in the region means that interpolated sediment data covers 37,400 km² (98%) of the shelf; 362,700 km² (20%) of the slope and 52,800 km² (10%) of the rise/ abyssal plain/deep ocean floor. Interpolated bulk carbonate data and folk classified data cover similar areas of each geomorphic province (Figs. 4.13 & 4.14).

The interpolated sediment maps give an interpretation of possible regional distribution of sediment properties. Areas with the highest sand (50-100% sand) and lowest mud (<50% mud) content are predicted to occur on the inner- to mid-shelf (Figs 4.10 & 4.11) except offshore of Hervey Bay where large variations of gravel (10-90%) and sand (0-100%) content occur. Mud contents increase significantly with water depth, with the highest mud contents occurring on the lower slope, rise, and abyssal plain/deep ocean floor. High gravel contents occur on the slope and locally in the north of the Tasman Basin (Fig. 4.12).

From the Folk Classification (Fig 4.14), gravelly sand (gS) and sand (S) with smaller quantities of slightly gravelly sand ((g)S) were most common on the shelf and upper slope. The lower slope is dominated by muddy sand (mS) and gravelly muddy sand ((g)mS). Muddy sand (mS), mud (M), and sandy mud (sM) become more common as water depth increases, occurring most frequently on the abyssal plain/deep ocean floor. An exception to this occurs offshore between Lorna Doone and Bambaroo, with highly variable sediment texture and relatively high (~70%) carbonate content. This change is best observed in the Folk Classification (Fig. 4.14) where an increase in gravelly muddy sand (gmS) is observed.

a)



b)

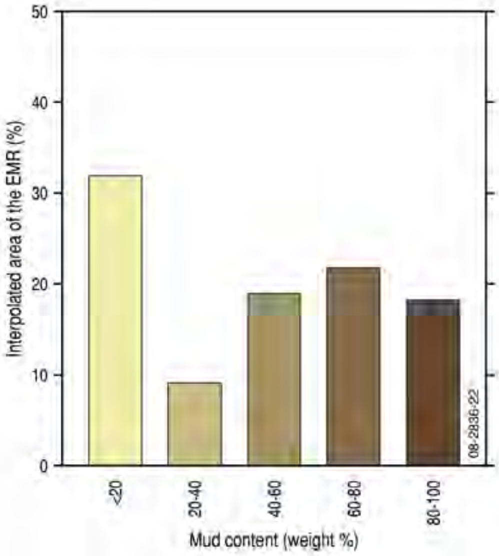
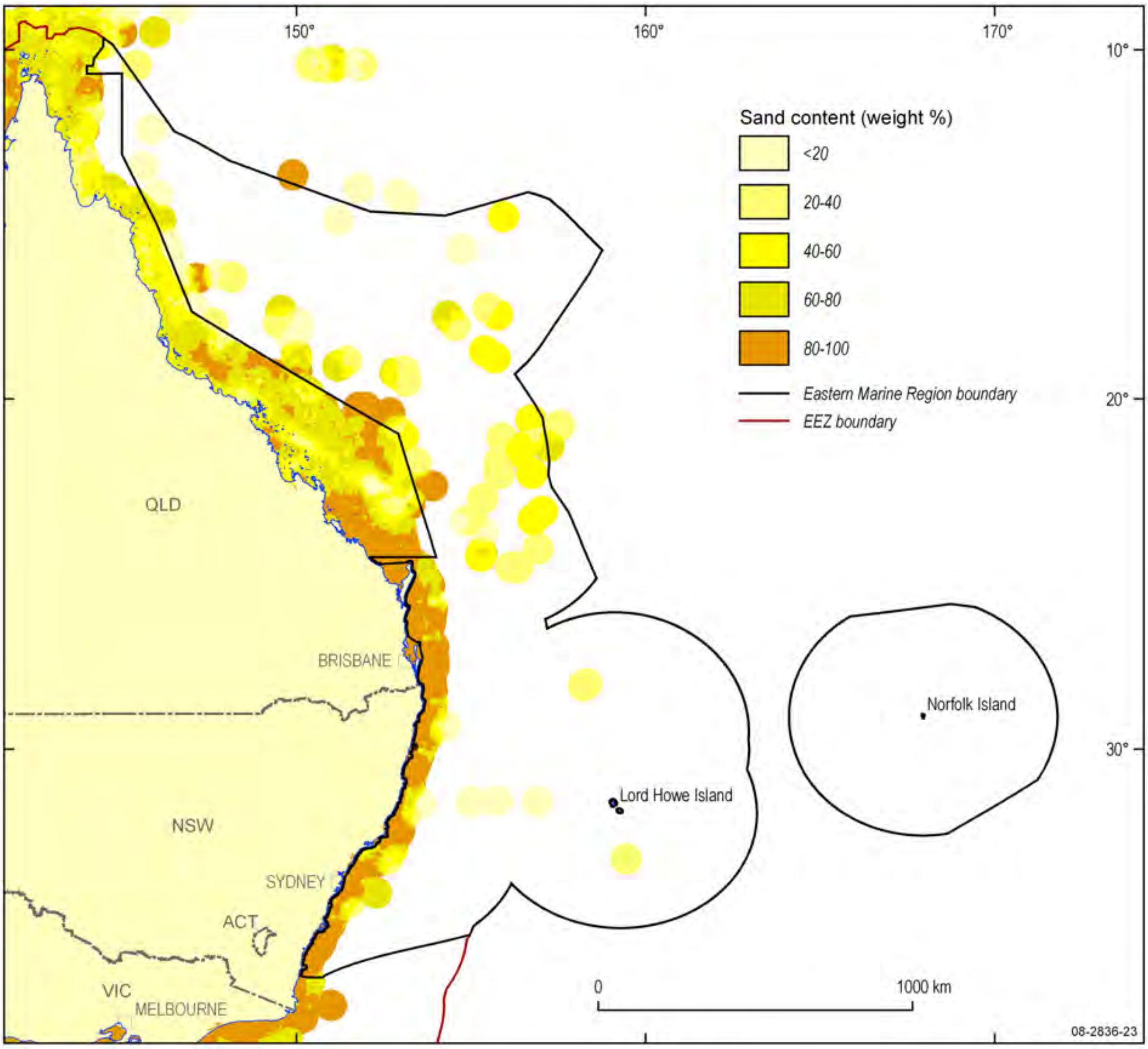


Figure 4.10. a) Distribution of mud in the EMR; b) % area of mud classes within the EMR derived from interpolated data.

a)



b)

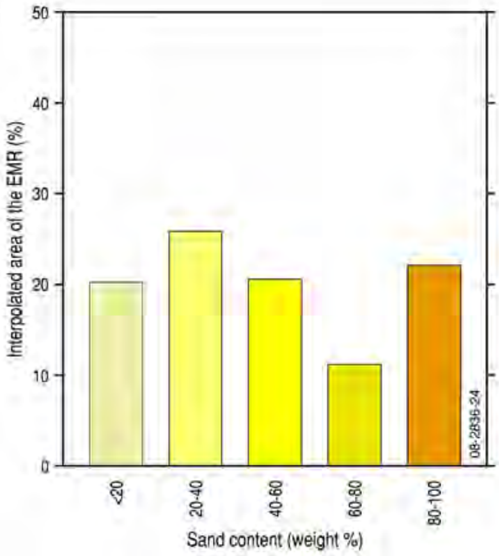
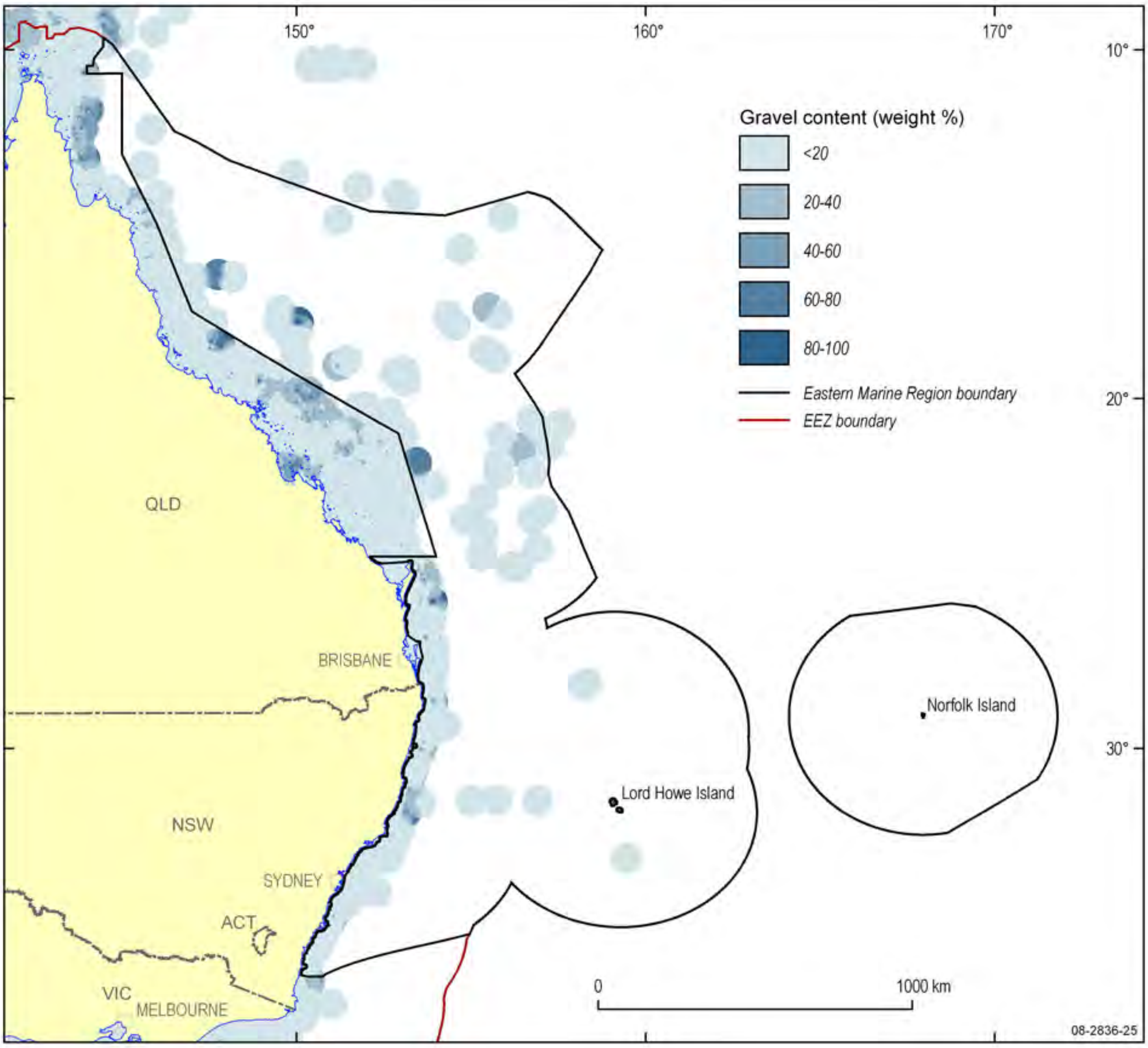


Figure 4.11. a) Distribution of sand in the EMR; b) % area of sand classes within the EMR derived from interpolated data.

a)



b)

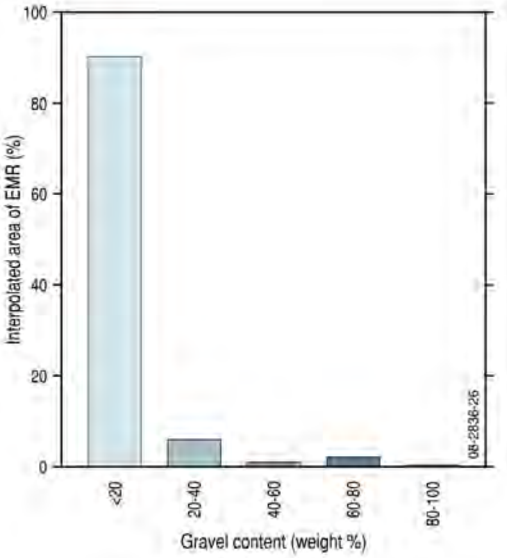
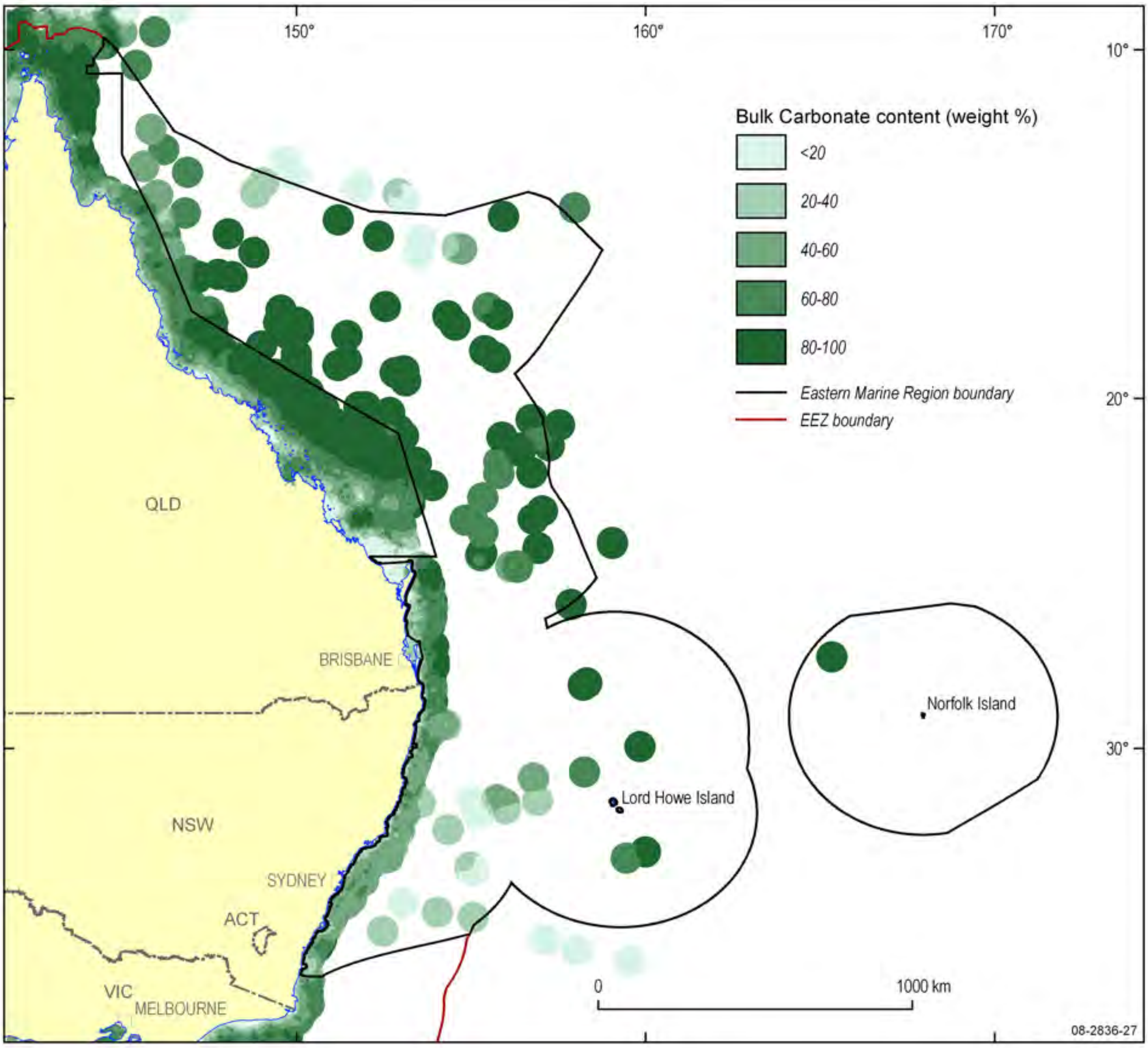


Figure 4.12. a) Distribution of gravel in the EMR; b) % area of gravel classes within the EMR derived from interpolated data.

a)



b)

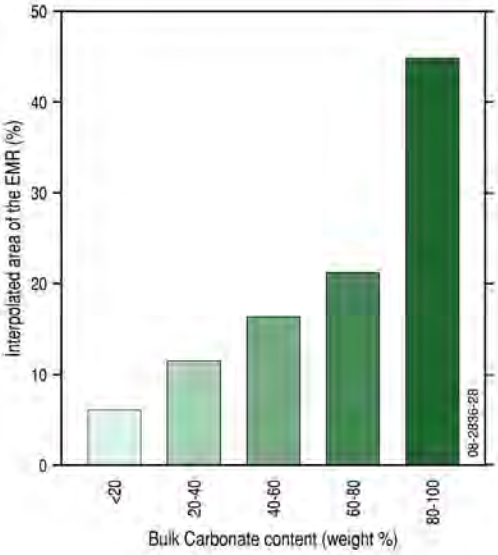
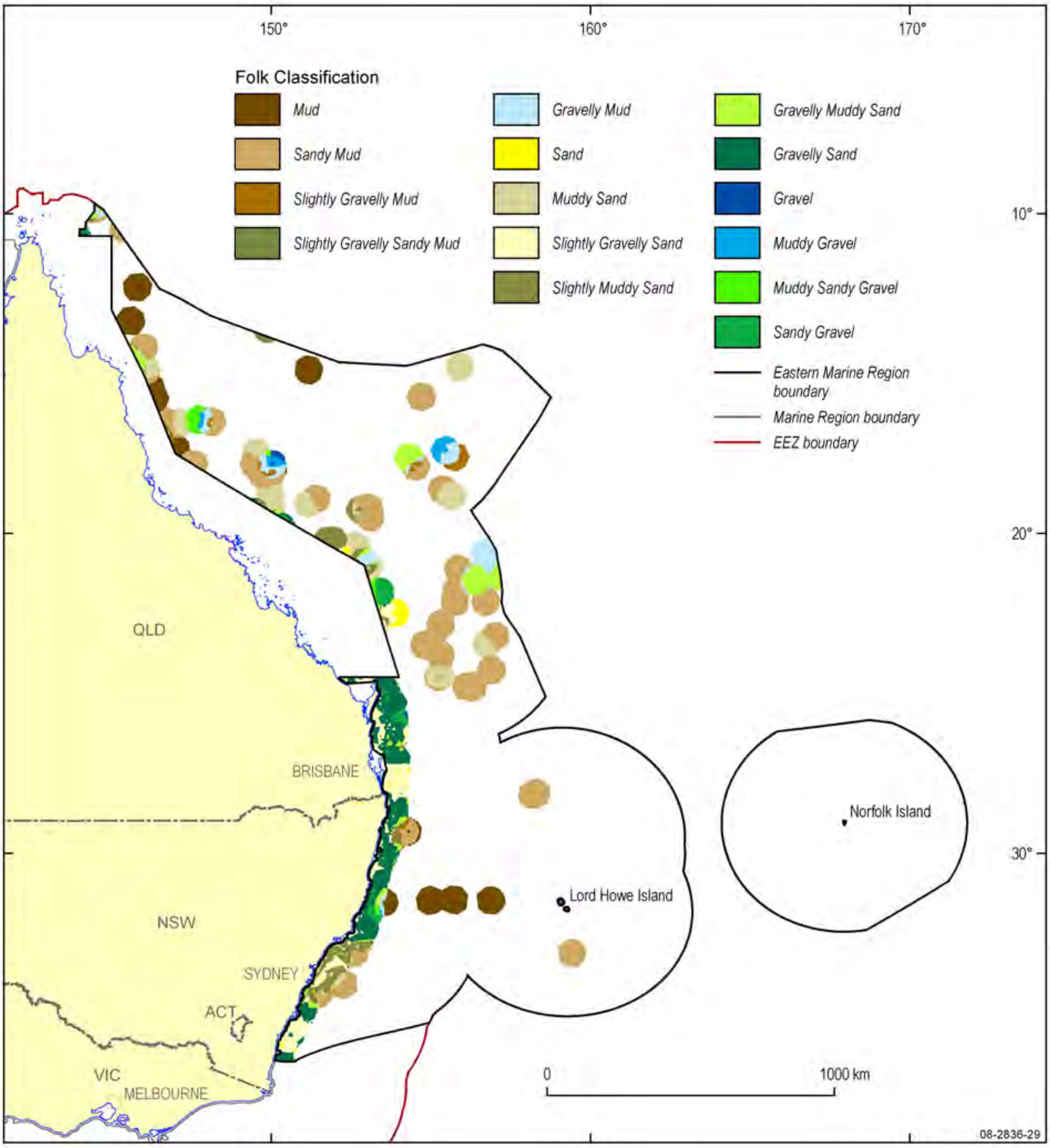


Figure 4.13. a) Distribution of bulk carbonate in the EMR; b) % area of carbonate content classes within the EMR derived from interpolated data.

a)



150°
160°
170°
10°
20°
30°
Folk Classification
Mud
Sandy Mud
Slightly Gravelly Mud
Slightly Gravelly Sandy Mud
Gravelly Mud
Sand
Muddy Sand
Slightly Gravelly Sand
Slightly Muddy Sand
Gravelly Muddy Sand
Gravelly Sand
Gravel
Muddy Gravel
Muddy Sandy Gravel
Sandy Gravel
Eastern Marine Region boundary
Marine Region boundary
EEZ boundary
QLD
BRISBANE
NSW
SYDNEY
ACT
VIC
MELBOURNE
Lord Howe Island
Norfolk Island
0
1000 km
08-2836-29

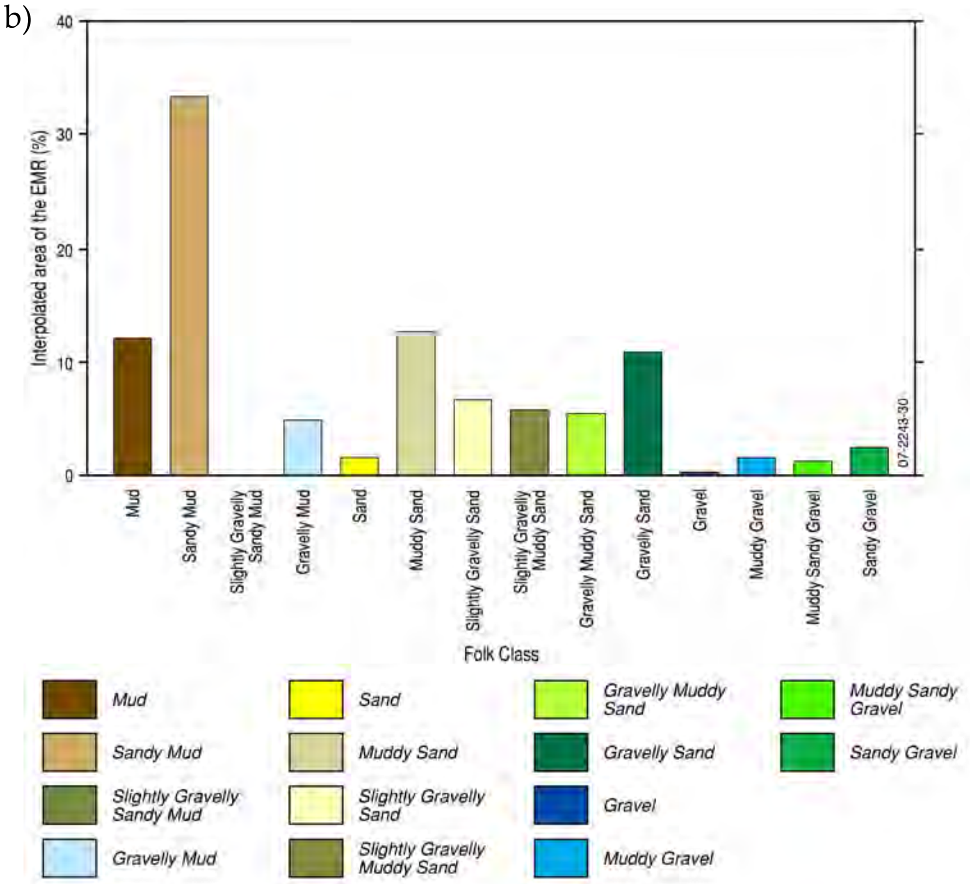


Figure 4.14. a) Distribution of folk classified sediments in the EMR; b) % area of folk classes within the EMR derived from interpolated data.

4.4.2. Sedimentology of Geomorphic Provinces and Significant Features of the EMR

Quantitative sedimentology is reported for features judged significant at a planning region scale that attain adequate sample coverage. Where coverage is only of local occurrence for a feature, sedimentology is assessed at a bioregion scale. Where occurrences of a feature form distinct groups based on morphology or water depth, each group has been described separately. Where a feature is judged as significant, but does not attain adequate data coverage, features are noted as significant at a planning region or bioregion scale. Properties and distribution of sediment within these features is, where possible, assessed from previous literature and summarised in Chapter 6. Significant features with no sedimentological assays include; rise (unassigned), apron/fans, deep/hole valleys, canyons, knoll/abyssal hill/hills/peaks, saddles, pinnacles, reefs and seamounts/guyots, but will be summarised in Chapter 6.

4.4.2.1. Shelf Province

The shelf in the EMR contains 485 grainsize and 375 carbonate assays. Over most of the area of the shelf, seabed sediment is characterised by sand (>60), with lesser gravel and/or mud (Fig 4.15a & Fig. 4.16a). Sand forms >60% of sediment at 459 (94%) sites sampled, and >90% at 322 (66%) sites. A total of 41 (8%) samples contain >25% mud, and 19 (4%) contain >25% gravel. Sediment containing significant proportions of mud (>25%) is generally restricted to the inner- to mid- shelf, however samples containing up to 75% mud occur locally on the outer shelf. Samples containing >25% gravel occur most frequently on the outer shelf, however samples containing up to 80% gravel occur locally on the mid-shelf off the east coast of Fraser Island.

Bulk carbonate content of sediment exceeds 75% in 99 (26%) samples from the shelf, and exceeds 90% in 26 (7%) samples (Fig. 4.16). Sediments containing significant proportions (>35%) of carbonate occur most frequently on the middle and outer shelf. Bulk carbonate contents are generally lower (<50%) in areas of the EMR closest to the shore.

Where the carbonate content of the gravel size fraction was measured, carbonate content exceeds 40% at all except two sites within Hervey Bay and Coffs Harbour. Where carbonate content of mud was measured, carbonate content does not exceed 37% except two sites located in Port Macquarie and offshore between Sydney and Gosford. Where carbonate content of sand was measured, carbonate exceeds 75% in 82 (22%) samples and exceeds 90% in 16 (4%) samples. Carbonate content of sand frequently exceeds 50% in three areas: 1) The mid- to outer shelf offshore of Fraser Island. Samples containing <30% carbonate sand are also common in this area. 2) Samples on the mid- to outer shelf offshore of Caloundra. Samples containing <20% carbonate sand also occur in this area. 3) Samples on the mid- to outer shelf offshore between Sydney and Gosford. While carbonate sand dominates locally in this area, sand carbonate content of sand is more commonly <40%.

4.4.2.2. Slope Province

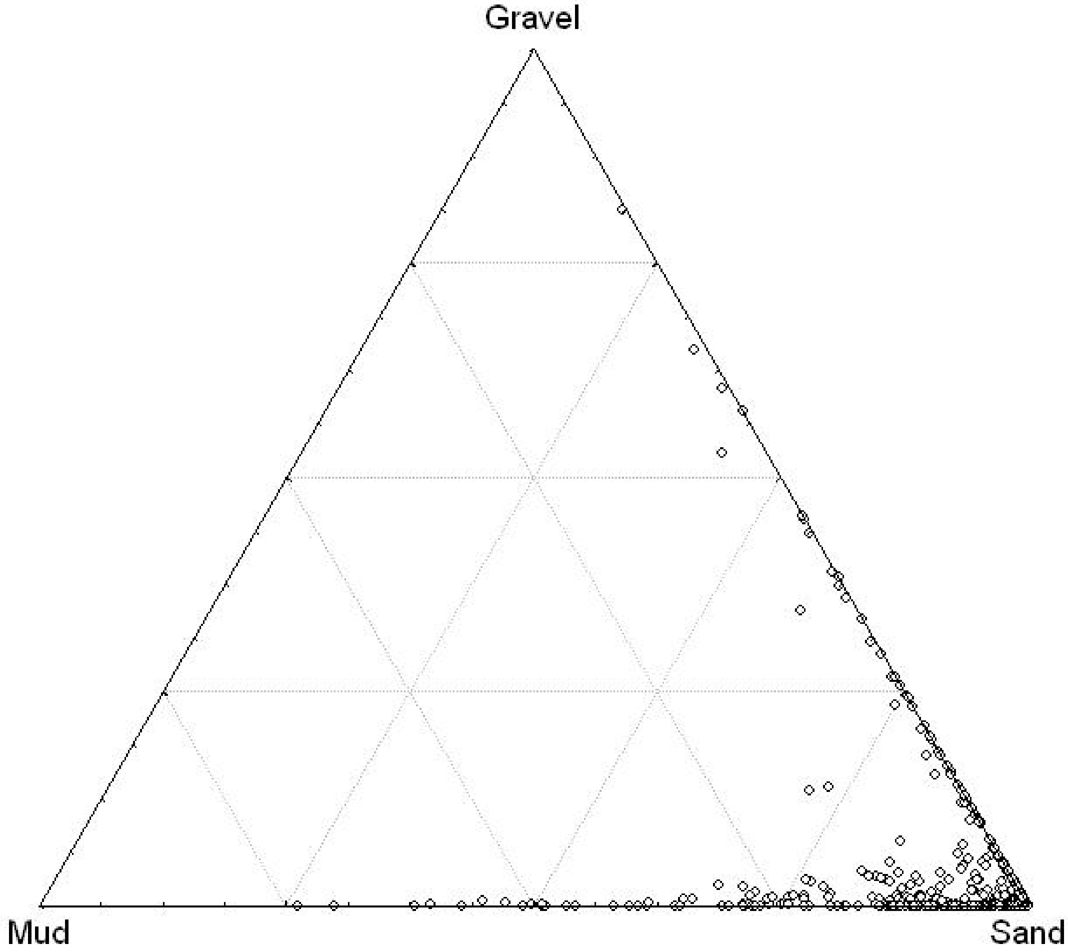
The slope in the EMR contains 113 grainsize and 118 carbonate assays; these are located mainly near the shelf break. Mud is interpreted to be the dominant size fraction across most of the area of the slope, though spatial clustering of assays means that overall statistics for the province do not reflect this (Fig. 4.15b). Sediment texture is zoned with water depth, with gravel and sand contents decreasing and mud content increasing with increasing water depth. Mud content of samples on the slope varies from 1 to >90%, however sediment composition in this area is highly variable with around 26% of samples containing <1% mud (reflecting sedimentology on the adjacent shelf), and around 19% of samples containing >70% mud. Gravel forms <1% of sediment at 63 (56%) sites sampled. All samples containing gravel >1% were collected within 100 km of the shelf break, gravel content at these sites ranges from 1 to 63%, with 102 (90%) samples containing <20% gravel. Sand content of slope sediment ranges from 3 to 100%, but rarely exceeds 30% in areas more than 100 km from the shelf break. Sand forms between 1 and 30% of sediment at >90% of sites sampled on the mid- and lower slope.

Carbonate content on the slope varies from 21-96%, with sediment on the upper slope generally containing >70% carbonate and mid to lower slope sediment showing highly variable bulk carbonate content with no apparent zoning with water depth (Fig. 4.16b). Carbonate mud content on the slope shows zoning with distance from the coastline of the Australian continent. On the upper slope it generally exceeds 60%, and attains >80% carbonate at nearly 50% of sites. Carbonate gravel content exceeds 80% at 80% of sites. Carbonate sand content varies between 42 and 93%, and exceeds 80% at 69% of sites sampled. Sand containing >80% carbonate was distributed across all areas of the slope.

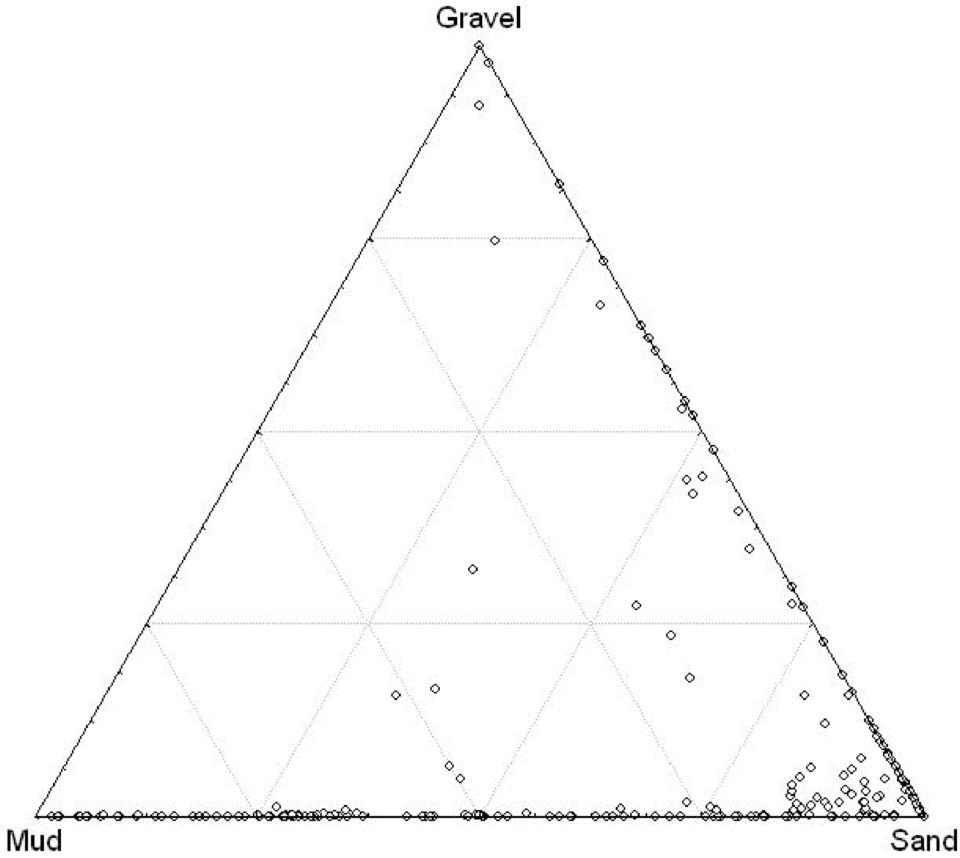
4.4.2.3. Abyssal plain/deep ocean floor province

The abyssal plain/deep ocean floor in the EMR contains two grainsize and 13 carbonate assays. Mud dominates sediment on the abyssal plain/deep ocean floor with both samples containing >95% mud, with the residual sediment made up of sand (Fig. 4.15c). Gravel was not detected on the abyssal plain/deep ocean floor. Bulk carbonate varies from 5 to 55% with the highest carbonate contents (45 to 55%) in two samples from the Tasman Basin Province (Fig. 4.16c). Carbonate mud content was available for two samples and ranged from 22 to 52%. Carbonate content of sand was not analysed.

a)



a)



b)

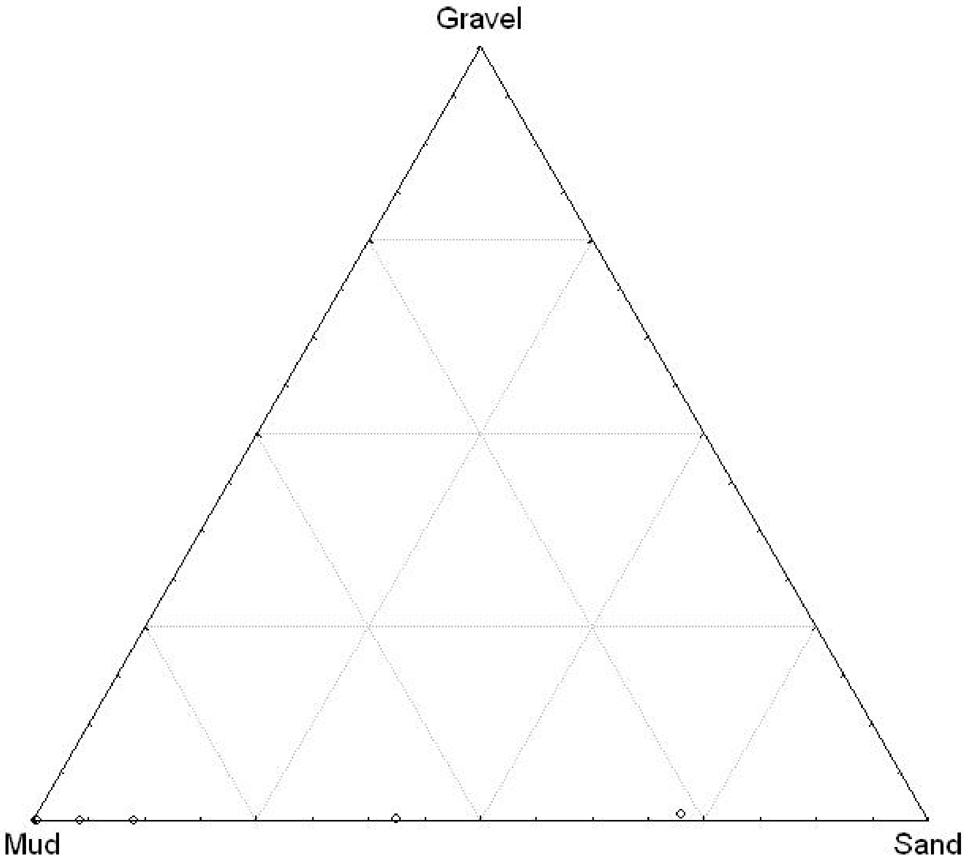


Figure 4.15. Textural composition (mud:sand:gravel ratio) for geomorphic provinces in the EMR, a) shelf; b) slope; and c) abyssal plain/deep ocean floor.

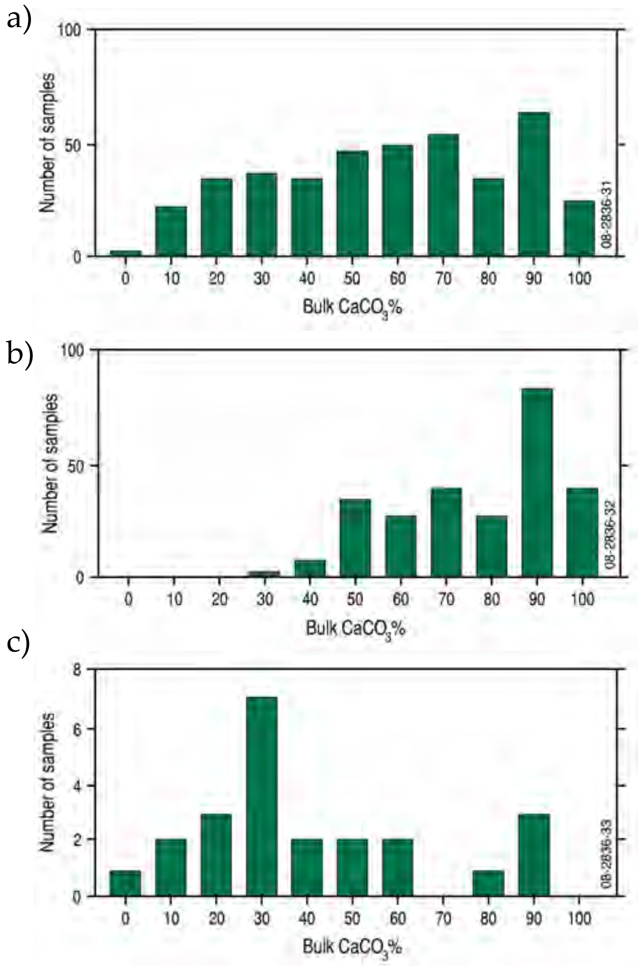


Figure 4.16. Carbonate content for geomorphic provinces in the EMR, a) shelf; b) slope; and c) abyssal plain/deep ocean floor.

4.4.2.4. Basins

A total of eight samples were obtained from basins. Basins contain seven grainsize and eight carbonate assays, all samples were located within the basin and coincident to the east and north boundary of NET and NEP, respectively. Five of the seven grainsize samples have only mean grainsize data. Mud is the dominant fraction with contents ranging from between 88 and 99%. Sand is the next most abundant sediment fraction with contents ranging between 1 and 11%. No gravel was detected. Bulk carbonate content ranges from 16 to 86% with four out of eight samples attaining 75%. Carbonate mud content ranges from 58 to 89% in two out of eight samples. Carbonate gravel and sand content were not analysed.

4.4.2.5. Deep water trenches/troughs

Deep water trenches/troughs contain 19 grainsize and carbonate assays. Mud is the dominant fraction with contents ranging between 21 and 97% (Fig. 4.17a), with six (32%) samples containing >75% mud. Sand is the next most abundant fraction, with contents ranging from 2 to 79%. Gravel content is <2% in 11 (58%) samples. Bulk carbonate content ranges from 37 to 94% with nine (47%) samples attaining 75% (Fig. 4.18f). The gravel fraction of sediment is entirely composed of carbonate in seven samples. Carbonate content of sand was analysed for six samples and ranges from 93 to 97%. Carbonate content of mud was analysed for 13 samples and ranges from 51 to 95%.

4.4.2.6. Shallow water terraces

Shallow water terraces in the EMR contain 22 grainsize and 29 carbonate assays. Sand is the dominant fraction with contents ranging from 31 to 100%, and exceeding 75% in 19 (86%) samples (Fig. 4.17b). The remainder of sediment is composed of gravel generally ranging from 0 to 53%, and smaller amounts of mud, ranging between 1 and 4%. Bulk carbonate content varies between 42 and 90% (Fig. 4.18). Carbonate sand ranges between 52 and 86% in six (27%) samples. No carbonate mud or gravel contents were measured.

4.4.2.7. Deep water terraces

Deep water terraces in the EMR contain 38 grainsize and 49 carbonate assays. Sand is the dominant fraction with contents ranging from 4 to 100%, and exceeding 75% in 31 (82%) samples (Fig. 4.17c). Gravel is the next most abundant fraction with contents attaining 40% in 31 (82%) of samples, and attaining 60% in three (8%) samples. Mud content generally ranges from 1 to 35%, but attains 60% in one sample. Bulk carbonate content of sediment ranges from 39 to 94%, and exceeds 75% in 22 (45%) samples (Fig. 4.18). Carbonate content of size fractions is available for 22 samples. Carbonate content of sand ranges from 43 to 94%, and exceeds 75% in five (23%) samples. Carbonate content of mud exceeds 40% in five (23%) samples and exceeds 75% in two (9%) samples. Carbonate content of gravel exceeds 30% in six (27%) samples and exceeds 75% in five (23%) samples.

4.4.2.8. Plateaus

A total of 62 samples were obtained from plateaus. Plateaus contain 55 grainsize and 62 carbonate assays which are located mainly in areas of these features that occur on the slope. Mud is the dominant size fraction with contents ranging from 6 to 98% and exceeding 50% in 29 (53%) of samples (Fig. 4.17d). Sand is the next most abundant fraction with contents ranging from 2 to 92%, and exceeding 50% in 14 (25%) samples. Gravel content is generally <30%, although three samples contain gravel ranging from 72 to 82%. Bulk carbonate content of sediment ranges from 32 to 82% and exceeds 75% in 49 (89%) samples (Fig. 4.18d). Carbonate content of sand ranges from 80 to 98% and exceeds 90% in 33 (55%) samples. Carbonate content of mud ranges from 34 to 96% and exceeds 90% in 11 (20%) samples. Carbonate content of gravel ranges from 5 to 100%. Gravel is composed entirely of carbonate clasts in 23 (42%) samples.

4.4.2.9. Pinnacles

A total of three samples were obtained from pinnacles. At a planning region scale, pinnacles show a common sedimentology that distinguishes them from other geomorphic features. Gravel is the dominant size fraction with contents generally ranging from 56 to 100%. Sand is the next most abundant fraction ranging from 14 to 40%. Mud content is <10% for all samples. Bulk carbonate content consistently exceeds 75%. The small number of assays for this feature means that results may not represent all sediments present in pinnacles in the EMR.

4.4.2.10. Plateaus on the Shelf or near the Shelf Break

A total of 48 samples were obtained from plateaus located on the shelf or near the shelf break. Sand is the dominant fraction with contents generally ranging from 25 to 100% (Fig. 4.17e) and exceeding 50% in 38 (79%) of samples. An exception to this is a single sample collected adjacent to a pinnacle that contains <5% sand. Mud is the next most abundant fraction with contents generally ranging from

3 to 49%. A total of 19 (40%) samples contained <20% mud and 12 (25%) contained no mud. Gravel content ranges from 1 to 100%, with 31 (65%) samples containing <20% gravel and 6 (13%) containing no gravel. Bulk carbonate content generally ranges from 17 to 99% and exceeds 50% in 40 (83%) samples (Fig. 4.18).

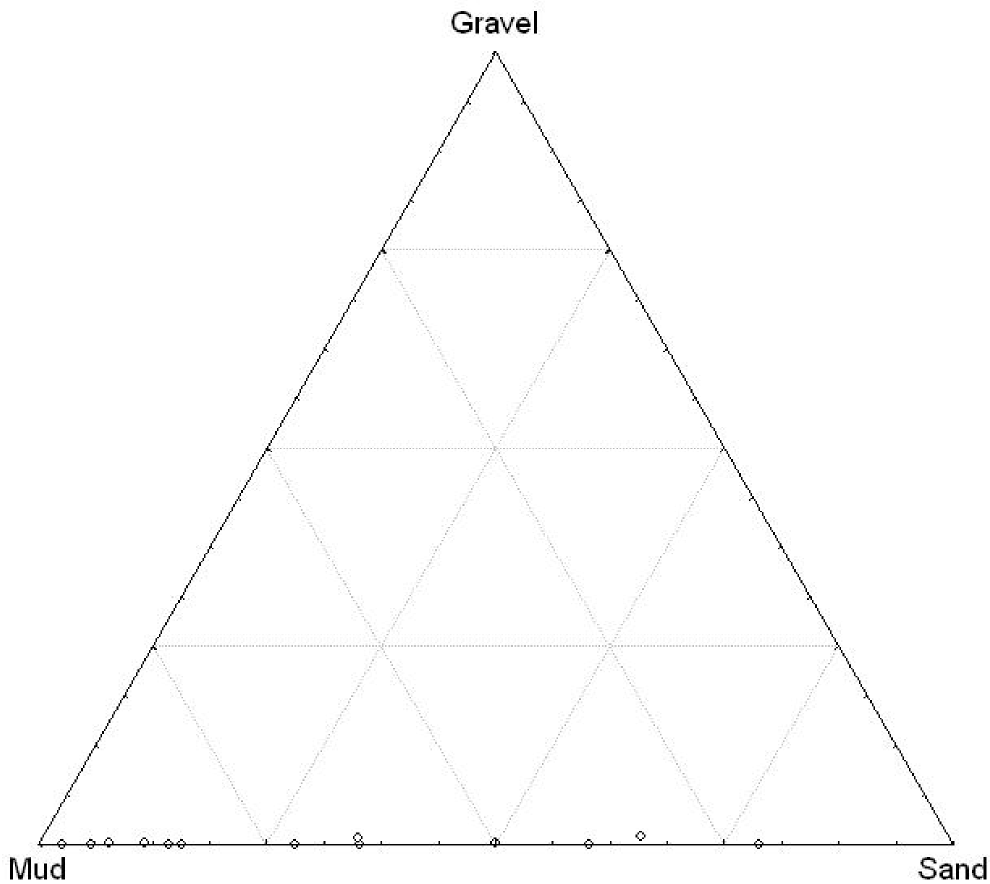
4.4.2.11. Offshore Plateaus and Terraces

A total of 78 samples were collected from offshore plateaus and terraces. Sand is the dominant size fraction in sediments with contents generally ranging from 2 to 100% (Fig. 4.17f), and exceeding 50% in 38 (49%) samples. Mud is the next most abundant fraction with contents ranging from 3 to 97%, with 10 (13%) of samples containing no mud. Gravel content is <25% in 32 (41%) samples, and <1% in 48 (62%) samples. Three samples (4%) contain between 27 and 45% gravel. Where data are available, bulk carbonate content generally ranges from 50 to 99% (Fig. 4.18).

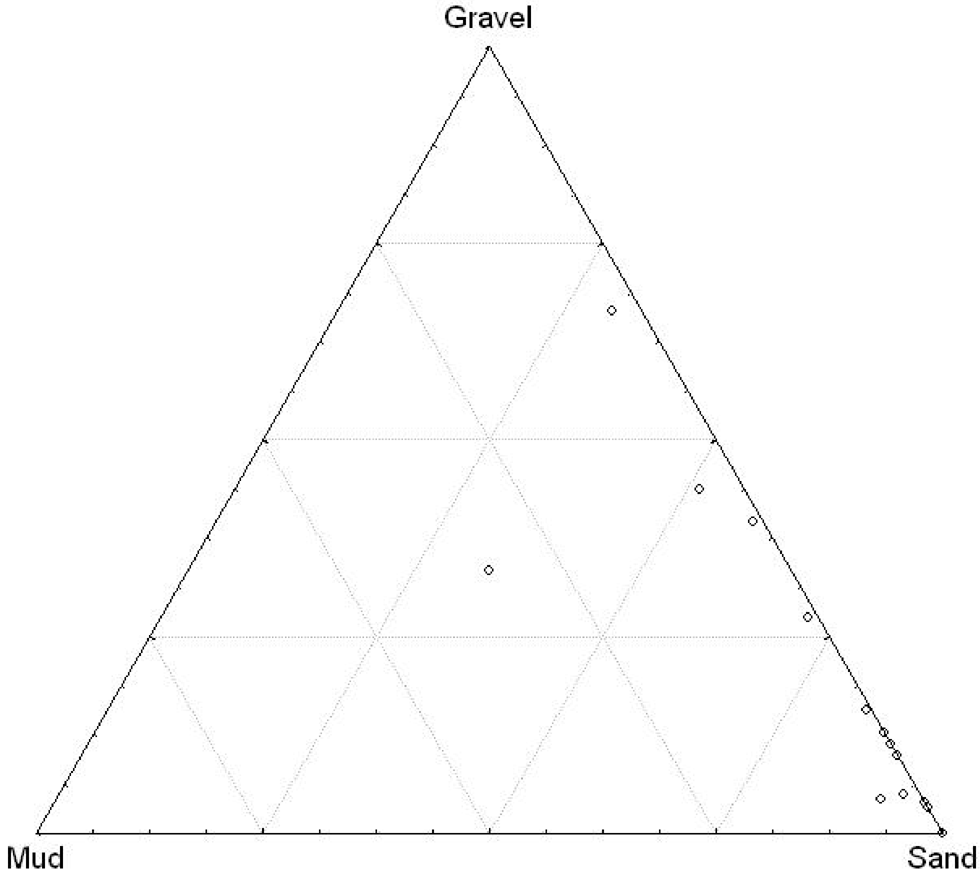
4.4.2.12. Terraces Located on the Shelf or Near the Shelf Break

A total of 210 samples were collected from terraces located on the shelf or near the shelf break. Sand is the dominant fraction in sediment with contents generally ranging from 8 to 100% (Fig. 4.17g) and exceeding 50% in 159 (76%) samples. Mud is the next most abundant fraction although the content is highly variable, ranging from 1 to 91%. Mud content exceeds 50% in 27 (13%) samples, while 37 (18%) contain no mud. Gravel content ranges from 1 to 88% in 75% of samples, and attains 100% in one sample. Where analysed, bulk carbonate content of sediment generally ranges from 42 to 99% (Fig. 4.18).

a)



b)



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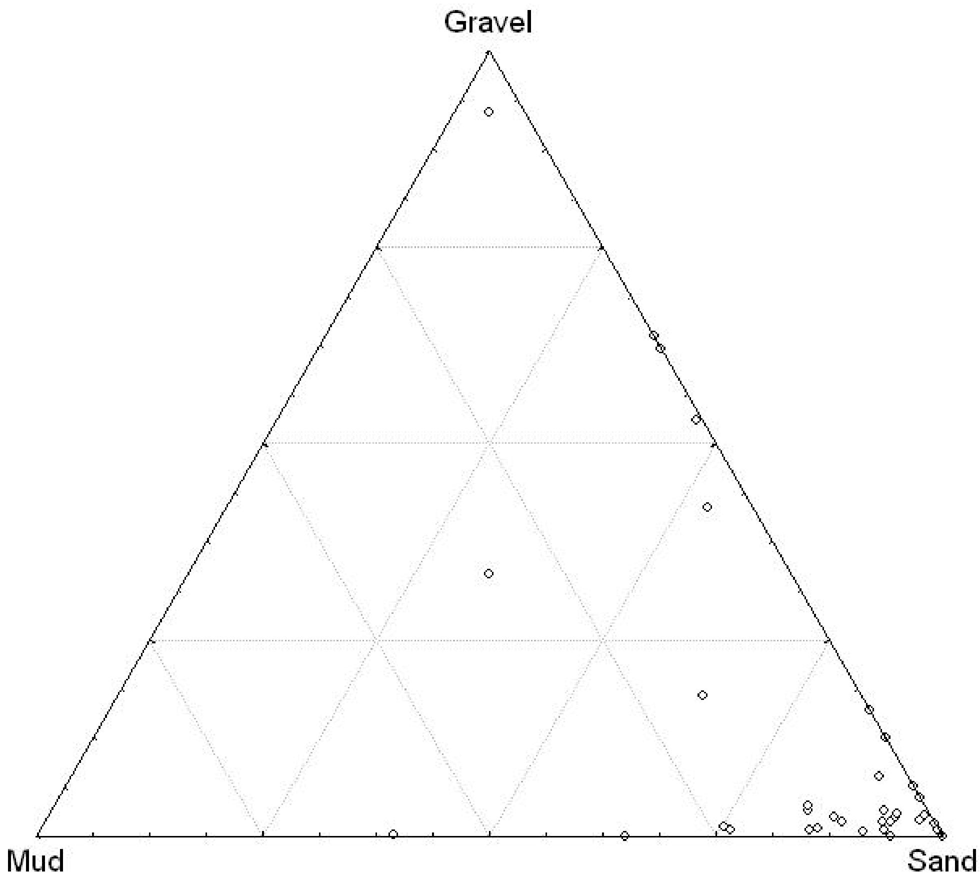
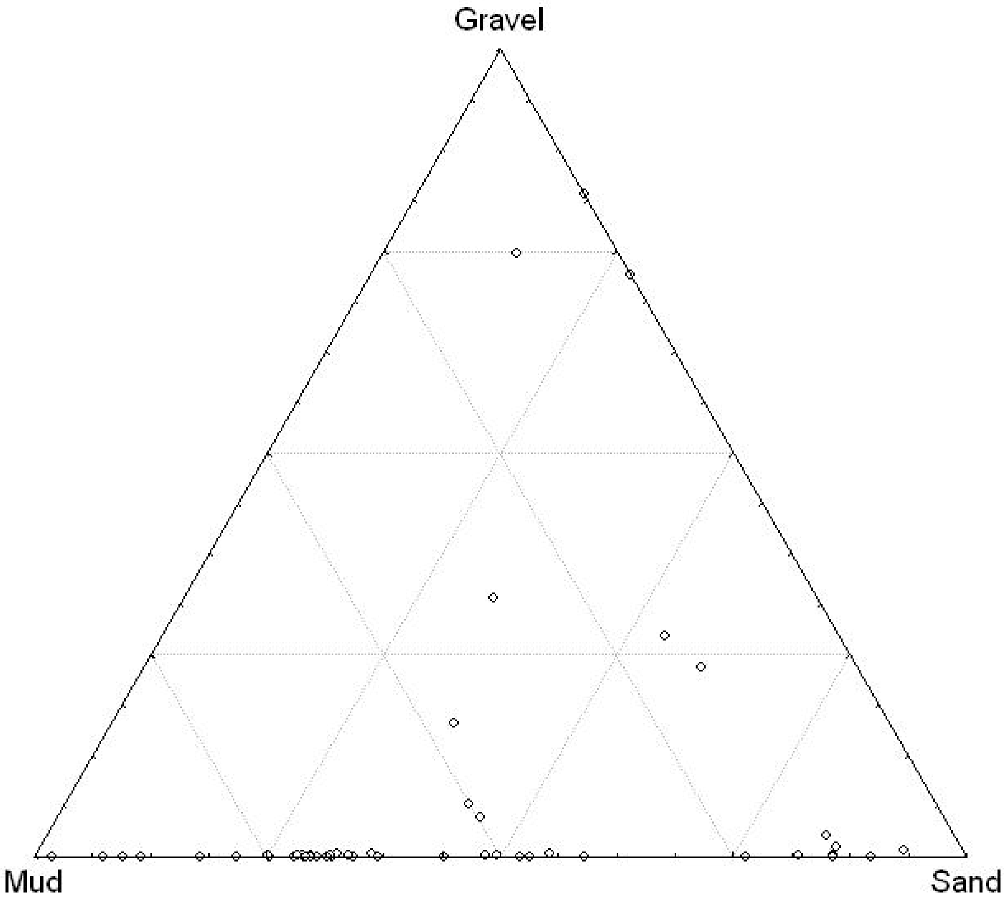
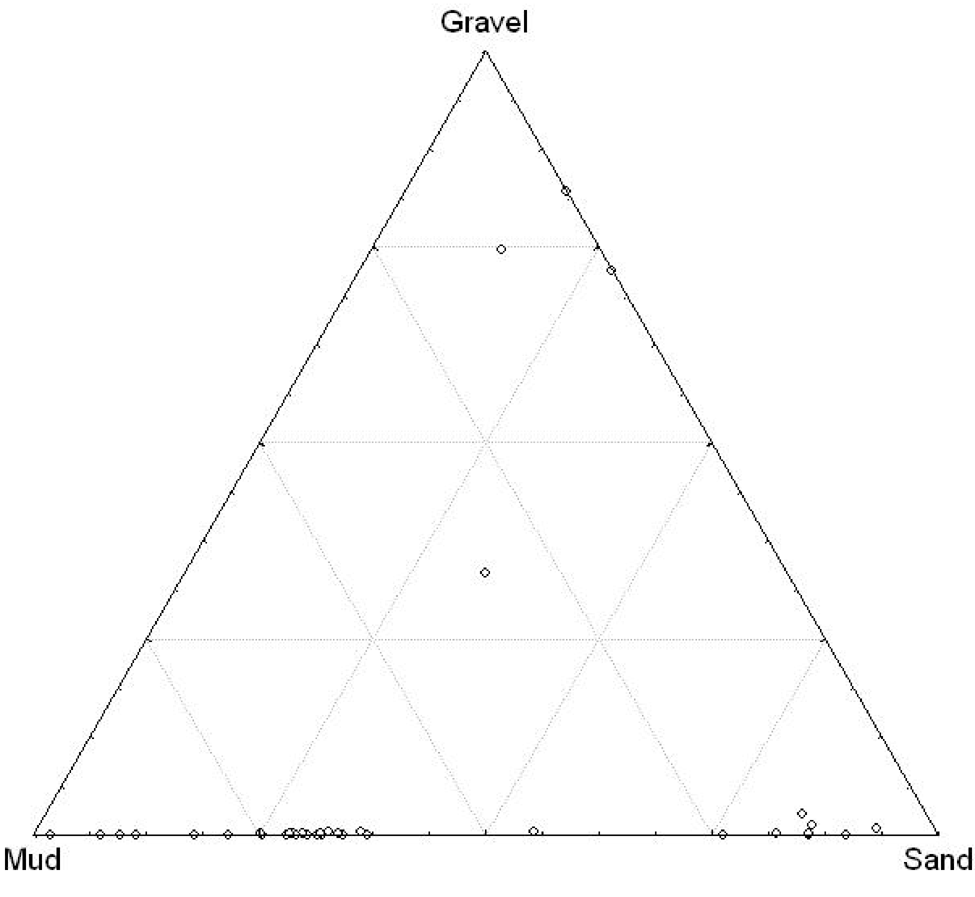


d)



e)



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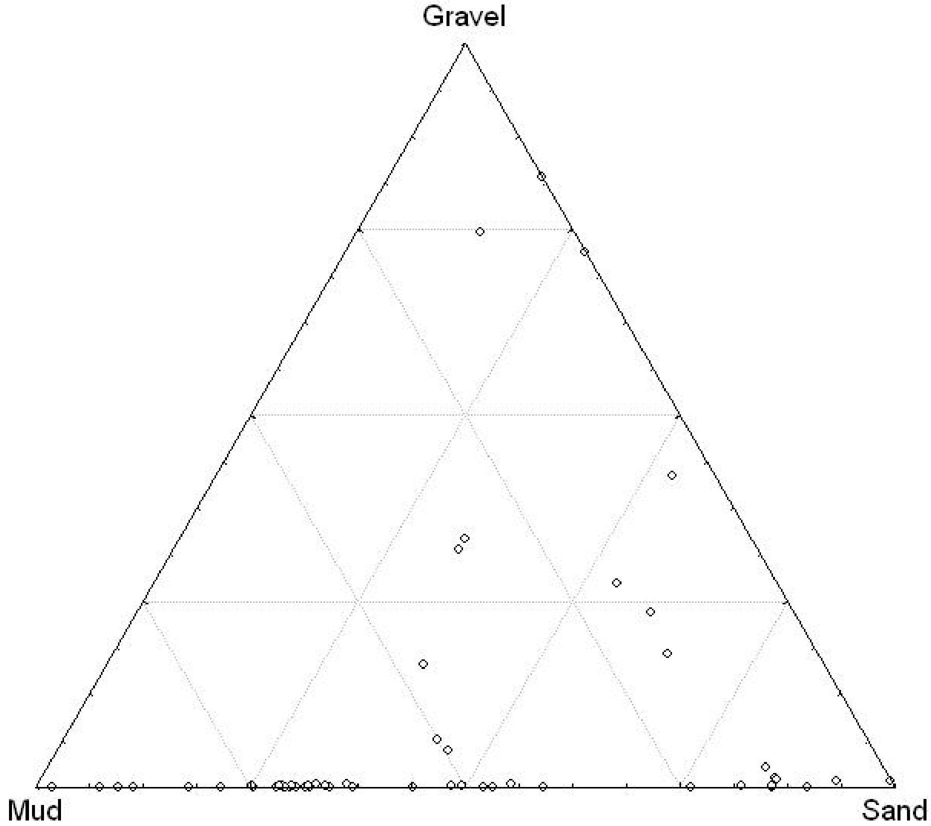


g)

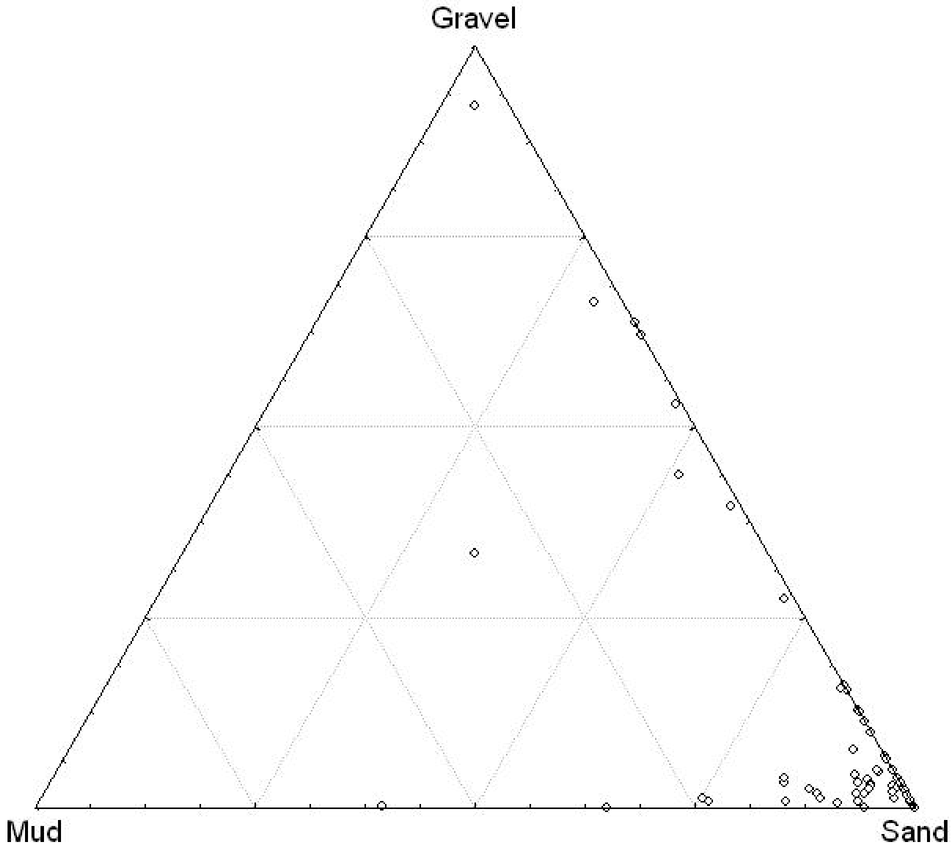
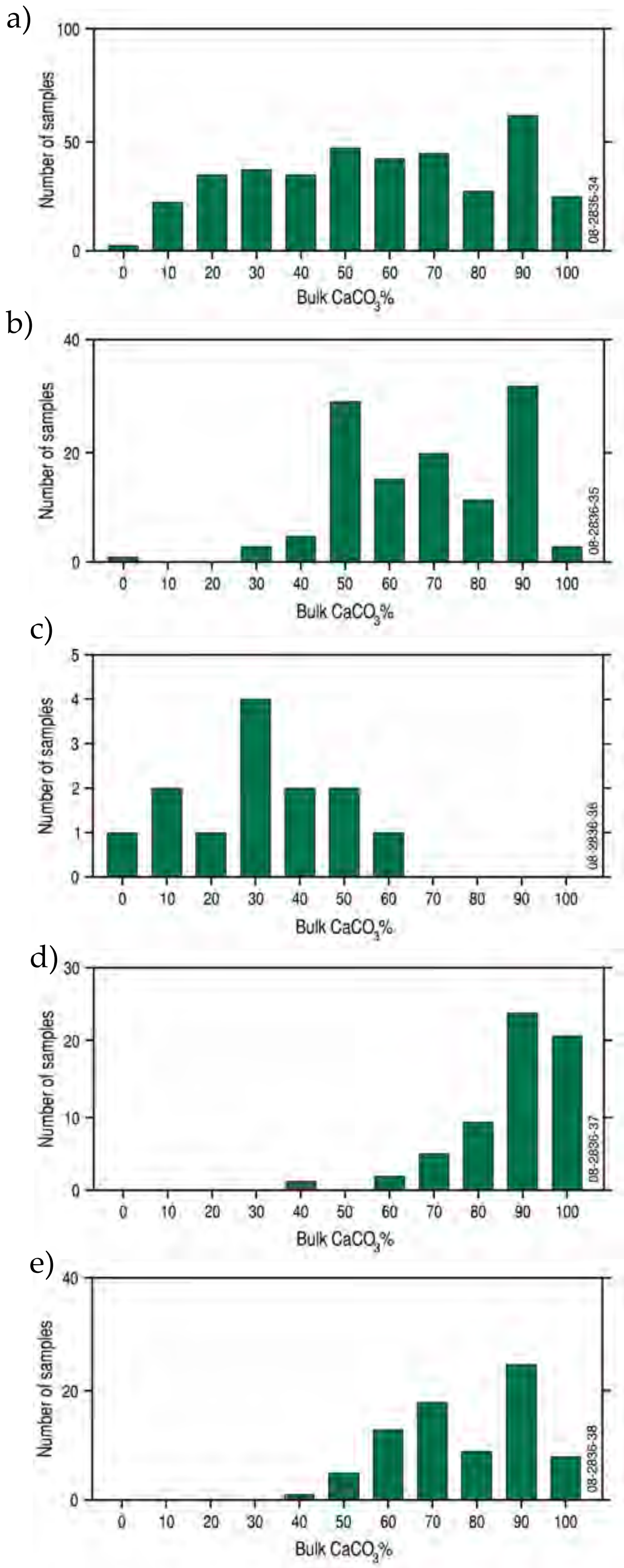


Figure 4.17. Textural composition (mud:sand:gravel ratio) of geomorphic features in the EMR, a) deepwater trench/trough; b) shallow water terrace; c) deepwater terrace; d) plateau; e) plateaus on the shelf or near the shelf break; f) offshore plateaus and terraces; and g) terraces located on or near the shelf break.


a)
b)
c)
d)
e)
Number of samples
Bulk $CaCO_3$%
0
10
20
30
40
50
60
70
80
90
100
08-2836-34
08-2836-35
08-2836-36
08-2836-37
08-2836-38

f)

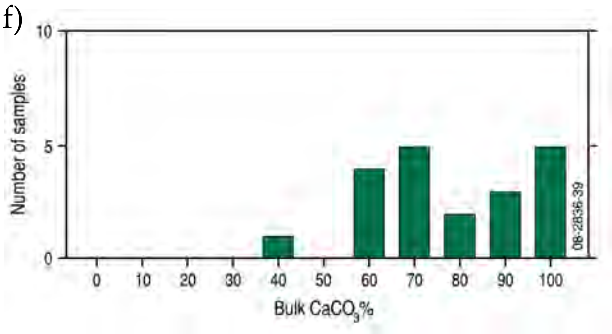


Figure 4.18. Carbonate content of geomorphic features in the EMR, a) shelf (unassigned); b) slope (unassigned); c) abyssal plain/deep ocean floor; d) plateau; e) terrace; and f) trench/trough.