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Quaternary

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Abstract

Despite their limited economic utility, Queensland's Quaternary deposits have attracted considerable attention. This is because studies of these deposits could shed light on processes and mechanisms of past periods of climate change, and therefore are particularly pertinent in contemporary debates on human impacts, future climate change and environmental fluctuation.

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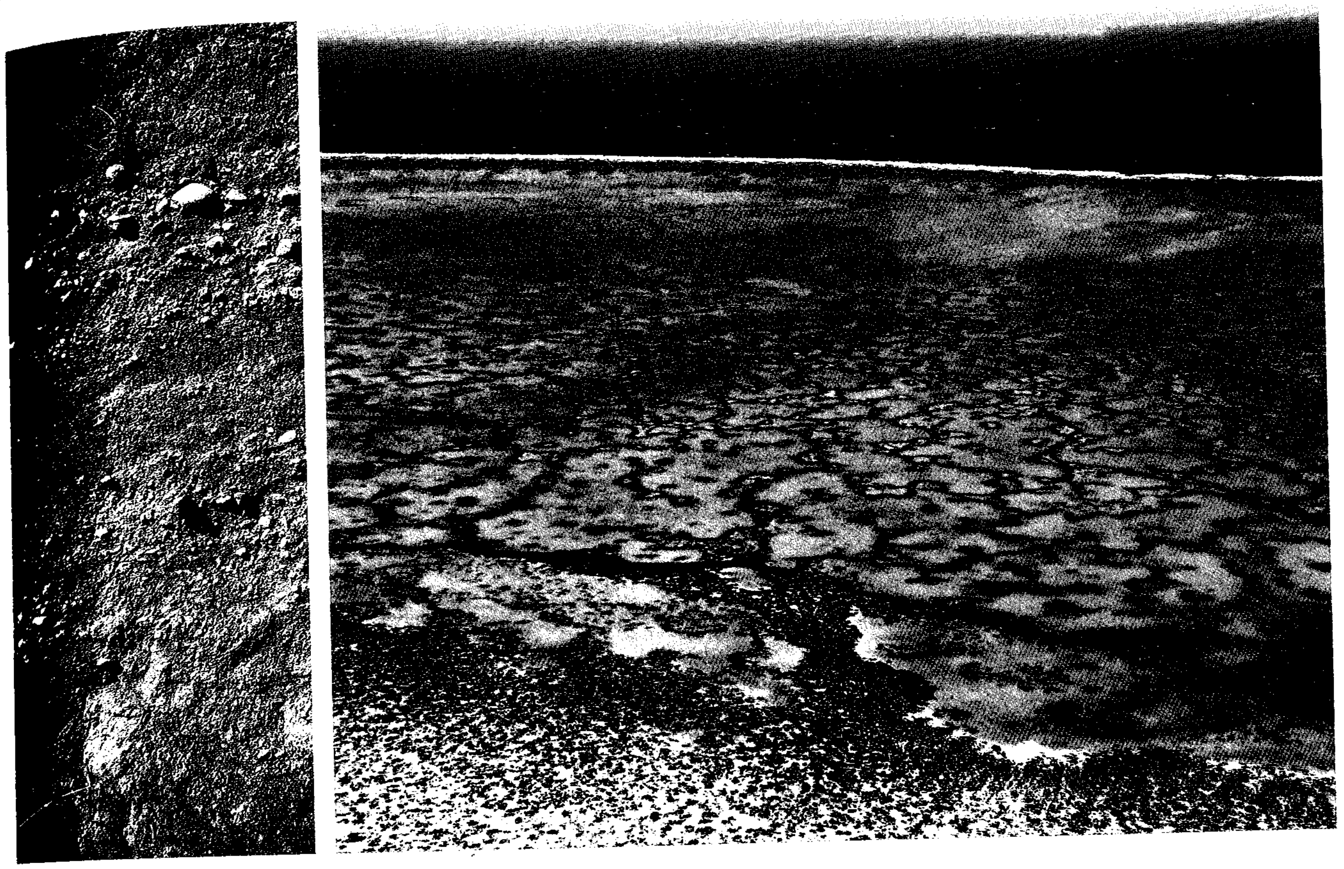
In New Guinea, following the late middle – early late Miocene unconformity, rise in sea level and continued uplift of the northern highlands saw clastic sediments fill the foredeep then spread over the platform, with late Miocene – early Pliocene carbonates restricted to the southern margin of the Papuan Basin and tropical coral reefs re-established on topographic highs. The Torres Strait – Maer lava-field volcanism began in the late Pliocene (~3 Ma) and extended into the Pleistocene.

Along the continental shelf, the late middle Miocene lowering of sea level led to a further sequence of a low sea-level progradational unit (probably late Miocene) followed by a high sea-level onlapping marine unit (Pliocene) outbuilding the shelf. This was followed in the late Pliocene by another low sea-level prograding unit. Subsidence continued to outstrip sedimentation in the Capricorn Basin, with late Miocene and Pliocene deepwater (>500 m) limy mudstone and marl deposited. The marginal troughs and oceanic basins received conformable successions of deepwater oozes, turbidites and slump deposits. In the late Miocene the Queensland Platform underwent increased subsidence, with the carbonate build-ups continuing on the elevated parts but backstepping and becoming restricted. Similarly, on the Marion Platform, subsidence and higher sea levels saw the platform flooded and carbonate sedimentation resume on the southern build-up but not on the northern. The southern one was drowned by the end of the Miocene or early Pliocene due to a major rise in sea level (>150 m). Hemipelagic drift deposits are found across the platform away from the build-ups and continued to the Pleistocene, but were subjected to strong currents probably associated with the Eastern Australian Current.

Australia came into being in the Mesozoic, with separation from Antarctica and New Zealand through opening of the Southern Ocean and the Tasman Sea, respectively. The Cenozoic framework of offshore Queensland provides evidence of the breakup of Australia's northeastern continental margin and its stratigraphy affords sea level and palaeotemperature data. Queensland's Cenozoic volcanic history allows its path and rate of northerly migration to be determined. Its onshore stratigraphy provides a record of cycles of geological activity alternating with periods of stability and deep weathering. Dated weathering profiles and weathered volcanic rocks suggest that Queensland changed from a green land, containing abundant vegetation and a rich fauna despite its high latitude, to a redder, drier environment to which animals and plants had to adapt. Watertables withdrew, rivers dried, and salt-lake deposits developed. All Cenozoic ore and industrial material deposits in Queensland are related to weathering and erosion, or concentration by fluvial and aeolian processes.



Chapter 9

Quaternary

by GJ Price

with contributions from
DI Cendón, AR Chivas,
A García and JS Jell

9.1 Introduction

Despite their limited economic utility, Queensland's Quaternary deposits have attracted considerable attention. This is because studies of these deposits could shed light on processes and mechanisms of past periods of climatic change, and therefore are particularly pertinent in contemporary debates on human impacts, future climate change and environmental fluctuation. We are in an interglacial period and face similar climatic cycling regimes to those patterns of climate change evident in the Quaternary geological record. Our ability to reliably predict the potential effects of hypothesised future climate and environmental changes is greatly enhanced by developing a thorough understanding of Quaternary geological and climatic processes. The vast majority of studies on Queensland's Quaternary history have therefore been directed towards sedimentary assemblages with a focus on palaeoclimatic and palaeontological reconstruction.

A wide range of previously used stratigraphic terms for Queensland's terrestrial Quaternary deposits were mentioned by Woods (1960), but many are considered invalid and few are in modern usage (Molnar & Kurz 1997). By necessity, the following discussion focuses on the spatial and temporal occurrence of broad geological features, rather than specific formations and basins. This chapter will cover geological, palaeoclimatic and palaeontological aspects of:

- cave deposits
- fluvial systems
- lacustrine settings
- aeolian features
- weathering and soil formation
- volcanism
- offshore deposits.

For the 1960 review of the geology of Queensland (Hill & Denmead 1960), few numerical dating methods existed that were applicable to Quaternary deposits. Allen et al. (1960) recognised the difficulties of developing reliable chronologies for Queensland's upper Cenozoic, noting that it was rarely possible to distinguish between Pleistocene and Holocene deposits. However, dating methods now available, including short- and long-lived radioisotopes (e.g. radiocarbon, Ar–Ar, K–Ar, U-series, cosmogenic isotopes), radiation exposure dating (e.g. TL, OSL, electron spin resonance and fission-track dating), relative dating (e.g. amino acid racemisation) and annually banded records (e.g. dendrochronology), as well as a greater understanding of biostratigraphic ranges of key taxa (e.g. of the extinct wombat-like marsupial *Diprotodon*), allow for ready age differentiation between such epochs. Despite lacking high geochronological resolution as in other continents, it is still possible to develop a reasonably reliable geochronological framework for Queensland's terrestrial Quaternary deposits. Dating of offshore events and sequences have also improved greatly with the Ocean Drilling Program including sensitive biostratigraphic, palaeomagnetic and geochemical chronological scales as well as correlation with established time curves for sea level and various geochemical proxies for oceanic environmental parameters.

9.2 Quaternary: definition and framework

The Quaternary encompasses the youngest of three Cenozoic periods (Figure 9.1). Revision of the International Geologic Time Scale has seen the base of the Quaternary extended from 1.8 Ma to 2.58 Ma to encompass the Gelasian Stage, formerly attributed to the Pliocene (Gibbard et al. 2010). The Quaternary comprises the Pleistocene (2.58 Ma–11.7 ka) and Holocene (11.7 ka–present) epochs (Figure 9.1). For the purposes of this review, the Pleistocene is informally divided into the early Pleistocene (2.58 Ma–780 ka), middle Pleistocene (780–126 ka) and late Pleistocene (126–11.7 ka) as proposed by Gibbard et al. (2010).

Relative to the past 65 million years of geological history, the speed and amplitude of global temperature oscillations during the Quaternary were unprecedented (Zachos et al. 2001). The climate of the Quaternary was dominated by repetitive glacial cycles (ice ages) and interglacial cycles. Between 2.7 Ma and 800 ka such cycles were dominated by 41 thousand-year oscillations, and since 800 ka, by 100 thousand-year oscillations (Ashkenazy & Tziperman 2004). Such transitions from glacial to interglacial episodes were driven by Milankovitch orbital forcing caused by variations in Earth's eccentricity, axial tilt and precession (Paillard 2001). In the oceans, sea levels fluctuated markedly, periodically encroaching and retreating across coastal plains. Exposed land bridges commonly connected mainland Australia to both Papua New Guinea and Tasmania during low sea-level phases associated with glacials (Lambeck & Chappell 2001). Continental glaciers and ice sheets developed in numerous regions of the Northern Hemisphere and contracted or expanded in response to glacial–interglacial cyclicity (Hubberten et al. 2004). In regions of the Southern Hemisphere where glaciers were not as prevalent, such as Australia, extensive arid zones developed in response to cool, dry and windy conditions (Bowler 1986; Hesse, Magee & van der Kaars 2004). It was during the Quaternary, under a backdrop of intense and varied climatic change, that the modern biota, ecosystems and landscapes emerged.

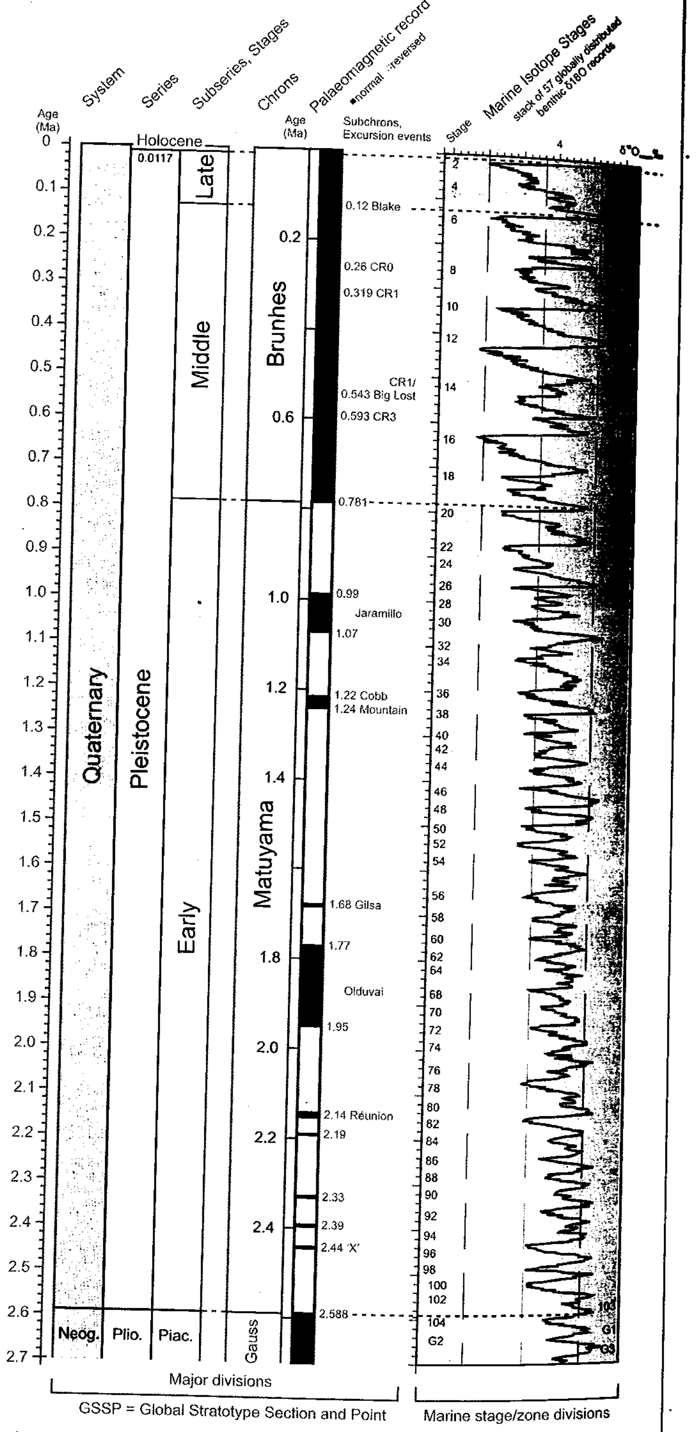


Figure 9.1 Major divisions of the Quaternary and Marine Isotope Stages (modified from Cohen & Gibbard 2010).

9.3 Caves

By far the best known and studied Quaternary caves of Queensland are those formed within limestone host rock. This is because:

- Cave-hosting limestones are common, especially along the length of the Tasmanides in the east of the state.
- Limestone caves and associated cave decorations (i.e. speleothem) are attractive geological features and have been generally well documented, especially by regional caving clubs (Sprent 1970; Chillagoe Caving Club 1988).
- Limestone caves are commonly valuable sources of palaeoclimatic and palaeontological information.

9.3.1 Mount Etna region

In the Mount Etna area north of Rockhampton (Figure 9.2), numerous caves containing fossil deposits formed in limestone of the Early Devonian Mount Alma Formation (Barker et al. 1997a; Murray et al. 2012). Associated cave deposits are mostly brecciated and clay-rich, with minor gravels and cobbles, commonly with interlayered flowstones. The deposits are dominated by extant and extinct vertebrate remains, including Pleistocene megafauna (large-bodied terrestrial mammals, birds, lizards and turtles). Over 20 fossil-yielding cave deposits have been identified at Mount Etna (Figure 9.3), as well as the adjacent Limestone Ridge (Hocknull 2005). An extensive U–Th dating study has been instrumental in placing the deposits into a reliable geochronological framework (Hocknull et al. 2007; Price et al. 2009a; Price & Hocknull 2011). Significantly, the results demonstrate that the region's 500–280 ka deposits contain extremely diverse faunas, including forms typically associated with modern rainforests of northeastern Queensland and New Guinea (e.g. tree kangaroos, striped possums, cuscuses and giant white-tailed rats). However, between 280 ka and 205 ka, a major faunal turnover event occurred that reflected the loss of 80% of taxa, including the extinction of numerous rainforest-adapted vertebrates. Deposits dating from 205 to 170 ka are typically dominated by more open-woodland-adapted forms and include species such as bilbies and pig-footed bandicoots, taxa that have been recorded in the historic period only from central Australian arid and semi-arid environments. Another major faunal turnover event is evident between 170 ka and the late Pleistocene, where there appears to be a shift from arid-adapted forms to more mesic-adapted taxa reminiscent of living faunas.

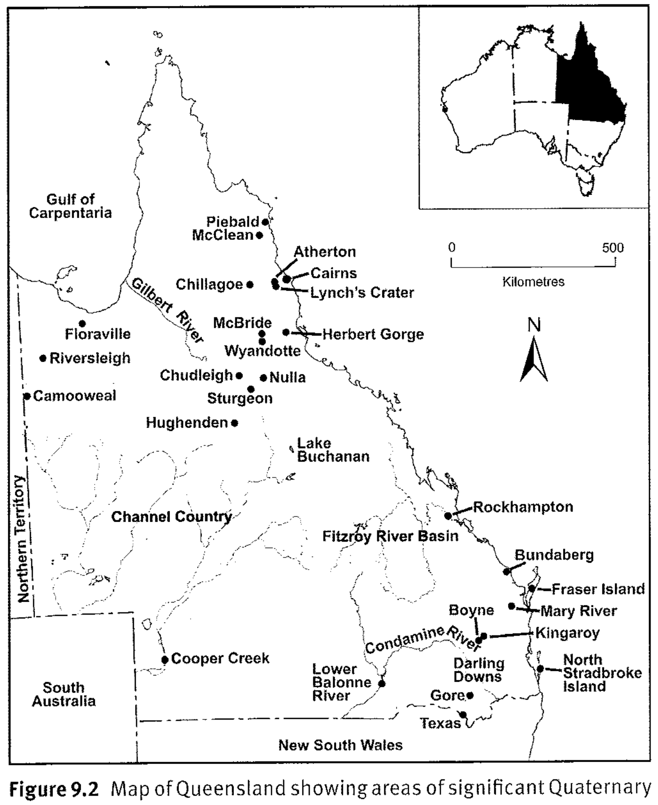


Figure 9.2 Map of Queensland showing areas of significant Quaternary interest.

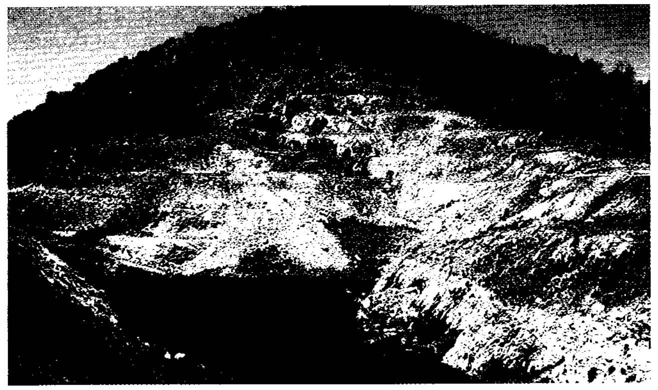


Figure 9.3 Mount Etna, central eastern Queensland, has produced several middle Pleistocene vertebrate fossil deposits.

Hocknull et al. (2007) also dated numerous speleothem formations from the region using U–Th methods. The results demonstrated that 70% of specimens formed prior to 300 ka, implying significantly wetter conditions before that time. Collectively, the results suggest that the Quaternary ecosystems of the region have been dramatically shaped by significant middle–late Pleistocene climate changes. A major decrease in humidity and precipitation during 280–205 ka led to the extinction of numerous rainforest vertebrates and development of more open habitats. A later shift towards subtropical conditions developed a more mesic local climate similar to that of today.

The middle Pleistocene faunal turnover of Mount Etna included the last dated occurrences of numerous species including extinct megafauna (e.g. the marsupial lion *Thylacoleo hilli*). A lack of younger fossil records of such taxa, both on local and continental scales, suggests that they may have become extinct long before the hypothesised late Pleistocene megafaunal extinction event at ~50–40 ka (Roberts et al. 2001). In contrast, Prideaux et al. (2007) found no evidence for a middle Pleistocene megafaunal turnover or extinction event in the Naracoorte Caves region of southeastern South Australia, suggesting that most extinctions were clustered in the late Pleistocene.

9.3.2 Texas Caves region

Numerous caves in the Texas region of southeastern Queensland (Figure 9.2) occur within the Texas beds, in limestone lenses up to 400 m wide and 1000 m long (Grimes 1978). Additional karst forms identified in the region include dolines (sinkholes) and small-scale solution pipes. Most of the region's caves were flooded following the construction of the Glenlyon Dam in the mid-1970s.

Formation of the caves and their exposure to the surface occurred progressively as the adjacent Pike Creek slowly incised through the limestone. Grimes (1978) suggested that the majority of the caves were carved from the host rock under low-energy phreatic conditions, as evidenced by the irregular forms of the passages and chambers and occurrence of several solution hollows, pockets and blades. Following formation of the caves, deposition and erosion of sediments and speleothem genesis were the dominant geological processes. The Quaternary deposits within the caves are similar to those from the Mount Etna region in that they are dominated by brecciated clays. In contrast, however, speleothems are much less common than at Mount Etna.

Fossil vertebrate faunas from the Texas Caves (Archer 1978; van Dyck 1982) are significantly different to those from Mount Etna in that they lack faunal elements characteristic of rainforest environments. However, they are diverse, with more than 40 species of mammals mostly typical of open-woodland habitats.

Archer (1978) described two major fossil assemblages from The Joint and Russenden (rear) Cave on Viator Hill. The Joint contains a diverse assemblage of extinct Pleistocene megafauna including the land crocodile (*Quinkana*), marsupial 'rhinoceros' (*Zygomaturus trilobus*), short-faced kangaroos (*Sthenurus*, *Simosthenurus* and *Procoptodon*), and long-faced forest wallabies (*Protemnodon* and *Macropus* sp. cf. *M. agilis siva*) and kangaroos (*Macropus* sp. cf. *M. thor* and *Macropus* sp. cf. *M. giganteus titan*). In contrast, Russenden (rear) Cave contains only *Protemnodon*, *Macropus* sp. cf. *g. titan*, *Macropus* sp. cf. *M. a. siva* and one species of *Sthenurus*. The remainder of the Russenden (rear) Cave assemblage is composed of extant and recently extinct taxa (e.g. thylacine).

A combination of geomorphological and palaeontological inference and numerical dating has been used to establish the ages of the local fossil deposits. Grimes (1978) argued that as the ancestral Pike Creek cut down through the limestone, the topographically highest caves would have been conducive to allowing faunal input long before caves positioned lower on Viator Hill. On that basis, Grimes (1978) hypothesised that The Joint (>410 m above mean sea level) fossil deposit is older than that of Russenden (rear) Cave (~402 m above mean sea level). Archer (1978) noted that more now-extinct megafauna species are found in The Joint assemblage (11 species) than in Russenden (rear) Cave (5 species). Because of this, and the higher proportion of extant forms in Russenden (rear) Cave, Archer (1978) argued that The Joint was the older of the two assemblages. Both assemblages contain taxa considered to be exclusive to Pleistocene deposits (Archer 1978). Although the original radiocarbon dating was unsuccessful (Archer 1978), Price et al. (2009b) applied U–Th dating of associated speleothems and teeth directly to determine the age of the deposits. The U–Th dating supported both Grimes (1978) and Archer (1978) and demonstrated that The Joint is >292 ka (but certainly middle Pleistocene), whereas Russenden (rear) Cave is ~55 ka (Price et al. 2009b). As for Mount Etna, the results broadly imply a long-term decline in biological diversity through the middle–late Pleistocene in southeastern Queensland.

9.3.3 Northeastern Queensland

Numerous caves are known from the Chillagoe, Mitchell-Palmer and Broken River areas of northeastern Queensland, but comparatively little work has been directed towards understanding their palaeontological and palaeoclimatic records.

At Chillagoe, numerous fossil faunas have been recovered from cave and fissure deposits within Devonian limestone (Figure 9.2). The majority of deposits are dominated by clays with minor gravels. A dated chronology for the majority of local deposits has not been established, although the faunas are characteristic of Quaternary species (Molnar 1977, 1981; Muirhead & Godthelp 1995; Cramb & Hocknull 2010) and include extinct land crocodiles (*Quinkana fortirostrum*) and carnivorous kangaroos (*Propleopus chillagoensis*) as well as extant forms.

At Tea Tree Cave, Chillagoe, Price et al. (unpublished data) produced 28 U–Th dates for a fossil maxilla of the marsupial 'tapir' (*Palorchestes azael*). The purpose of the study was to determine the age of the specimen and to demonstrate the utility of the direct fossil dating approach of museum specimens using U-series methods. The specimen was collected from the cave in 1977 and curated into the fossil collections of the Queensland Museum before being sequestered for dating. The results demonstrate that the age of the specimen is between ~137 and 199 ka, predating the hypothesised time of final megafaunal extinctions. The result is significant because it is the most northerly mainland dated record for any of the extinct Australian megafauna and represents one of the youngest reliably dated records for the species. The stratigraphic relationship of the dated specimen to other fossils from the cave is unclear.

Speleothem-based palaeoclimate reconstructions have also been established using samples collected from Chillagoe (Figure 9.4). Xia, Zhao and Collerson (1998) examined a stalagmite that formed ~20–7.8 ka (chronology established using U–Th methods). The stalagmite grew during the Last Glacial Maximum, through the period of late Pleistocene global warming and into the early portion of the current interglacial. Several systematic variations can be seen in the stalagmite, providing evidence of rapid climatic oscillations associated with the deglaciation (Xia et al. 1998). An increase in growth rate and decrease in $\delta^{18}O$ at around 10.2 ka is likely due to enhanced activity of the northern monsoon (Xia, Zhao & Collerson 1998).

Nott (2007) and Nott et al. (2007) explored the prehistory of northeastern Australian tropical cyclones based on an annually banded stalagmite also collected from Chillagoe. The specimen, dated at 1228–2003 AD, was sampled at high resolution to obtain a $\delta^{18}O$ record. Low $\delta^{18}O$ values were interpreted to indicate that cyclones had passed over the region (Nott 2007). The analyses demonstrated that over the past ~800 years, not only were there significant variations in the frequency of tropical cyclones; there were also marked centennial-scale regimes. For example, 1600–1800 AD was a period of intense tropical cyclone activity. Conversely, the record shows that tropical cyclone activity in northeastern Queensland has been in a period of quiescence since the 1800s (Nott 2009). Understanding the history and frequency of past tropical cyclones is critical for predicting their behaviour in the advent of global warming and enriched atmospheric CO_2 (Nott 2011).

9.3.4 Other cave and fissure deposits

Many other Queensland caves contain fossils of extant vertebrates, suggesting that the associated deposits are Quaternary in age, although only a few have actually been extensively dated using radiometric methods. For example, the Camooweal Caves, Carrington's Cave and Message Stick Cave (near Riversleigh) in northwestern Queensland (Figure 9.2) contain the remains of extant taxa, but such records have not been documented stratigraphically, nor have they been isotopically dated (Tyler 1990; Archer et al. 2006; Cramb & Hocknull 2010). Sedimentary characteristics of such deposits are also unknown. At Marmor in central eastern Queensland, several fossil teeth have been directly dated using U-series methods and demonstrate that the assemblage is middle Pleistocene (Price et al. 2009a).

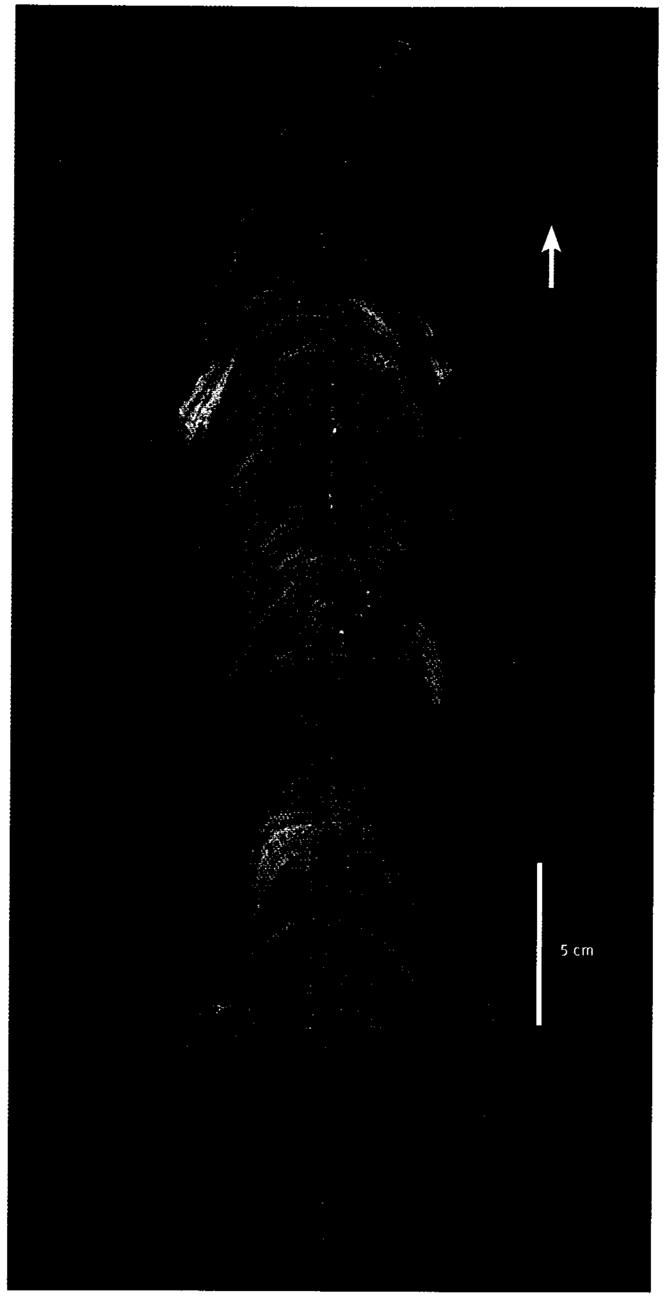


Figure 9.4 Late Quaternary stalagmite that formed ~20–7.8 ka. Arrow indicates growth direction.

Cave and fissure deposits from Cement Mills, Gore, southeastern Queensland, have been the focus of palaeontological investigations (Figure 9.2). The deposits occur in restricted limestones of the Paleozoic New England Orogen, thought to be part of the Texas beds (Siemon 1973). Over 30 taxa have been identified from the cave earths and include extant lizards and marsupials as well as extinct Pleistocene megafauna (Bartholomai 1977). Direct U–Th dating of megafaunal fossils yielded a minimum age of 53 ka (Price et al. 2009a). A maximum age for the faunas is yet to be established. The deposits are the type locality for the extinct giant koala, *Phascolarctos stirtoni* (Bartholomai 1968), which lived alongside the smaller extant koala (*P. cinereus*) at least until the late Pleistocene, when it became extinct (Price 2008a). The youngest appearance age for the giant koala on the entire continent is that for the Cement Mills specimen. Significantly, this young record predates the hypothesised late Pleistocene megafauna's extinction 'window' (40–50 ka). Price (2012a) noted that late Pleistocene pre-human climate-driven habitat changes were likely responsible for a massive retraction in the geographic range of the modern koala; it is possible that the giant form became extinct during coeval environmental disturbances.

9.4 Fluvial systems

Although Queensland has numerous active rivers, creeks and streams, particularly in the higher rainfall zones of the northern and eastern margin, the Quaternary history of fluvial systems in semi-arid and arid environments has attracted the greatest attention. Although most studies of Queensland's Quaternary fluvial systems have largely had a palaeoclimatic focus, they have also been important for developing a greater understanding of sedimentological processes of inland rivers.

9.4.1 Channel Country

The Channel Country of southwestern Queensland (Figure 9.2) encompasses the catchments of the Georgina River, Diamantina River and Cooper Creek (Figure 9.5), which drain into the Lake Eyre Basin of the central arid zone (Figure 9.6). Although such catchments may lie dormant for several years at a time, several high-energy fluvial episodes have been identified during the Pleistocene. These fluvial episodes are among the most extensively studied for the Quaternary in Queensland (Rust & Nanson 1986; Nanson, Price & Short 1992; Gibling, Nanson & Maroulis 1998; Maroulis et al. 2007; Nanson et al. 2008).

OSL and TL dating (both targeting sediments) have been the key to establishing the chronology of late Quaternary fluvial deposition. For example, Rust and Nanson (1986), Nanson, Chen and Price (1992) and Nanson et al. (2008) dated numerous palaeochannel and overbank sequences in the Channel Country.

Figure 9.5 The Cooper Creek flood plain, channels and adjacent waterholes.

With an enlarged dataset of dated sequences from other sites within the Lake Eyre Basin (>150 OSL and TL dates), Nanson et al. (1992b, 2008) demonstrated that high-energy fluvial conditions dominated Marine Isotope Stage (MIS) 7 (250–230 ka). Mean bank-full discharges on Cooper Creek are estimated to be up to seven times larger than those of today (Nanson et al. 2008). Numerous other central rivers then became laterally restricted between MIS 7 and MIS 5 (130–74 ka). In the late Pleistocene, peak fluvial activity was greatest at around 110 ka (MIS 5d), rather than at the height of the last interglacial maximum of MIS 5e (132–122 ka), occurring at a time when sea levels and temperatures were lower than today. Nanson et al. (2008) noted a return to fluvial conditions in MIS 4 and 3 (~65–50 ka), although not to the degree seen in MIS 5. Dating of stranded beach ridges from Lake Eyre can also be used as a proxy to examine episodes of flow from the Channel Country. For example, Magee et al. (2004) showed that lake-full conditions were progressively less common through the late Pleistocene, with each subsequent filling event failing to reach the same level as previous lake-full episodes (Magee et al. 2004). Cohen et al. (2011, 2012) also demonstrated that the last time Lake Eyre was connected to Lake Mega-Frome in the south (filled partly from floodwaters from the Channel Country via overflow from Lakes Eyre, Gregory, Blanche and Callabonna) was between 47 and 50 ka. After that time, fluvial activity in the Channel Country was greatly reduced in the lead-up to the Last Glacial Maximum (Nanson et al. 2008).

Significantly, the results suggest that although broad Milankovitch climate-driven shifts have oscillated between wet and dry conditions, there has been a progressive decline in precipitation since MIS 7. Moreover, the chronology of late Quaternary fluvial activity closely matches that of southeastern Australia, rather than northern Australia (Nanson, Price & Short 1992). Combined with the observation that the MIS 5d peak in fluvial activity occurred at a time when temperatures were lower than today, the results suggest that the northern monsoon was not a major source of moisture in the region. Nanson et al. (2008) proposed an alternative model where a western Pacific warm pool was trapped adjacent to eastern Australia, establishing a semipermanent La Niña, with westerly trade winds irrigating the Channel Country and the Lake Eyre Basin.

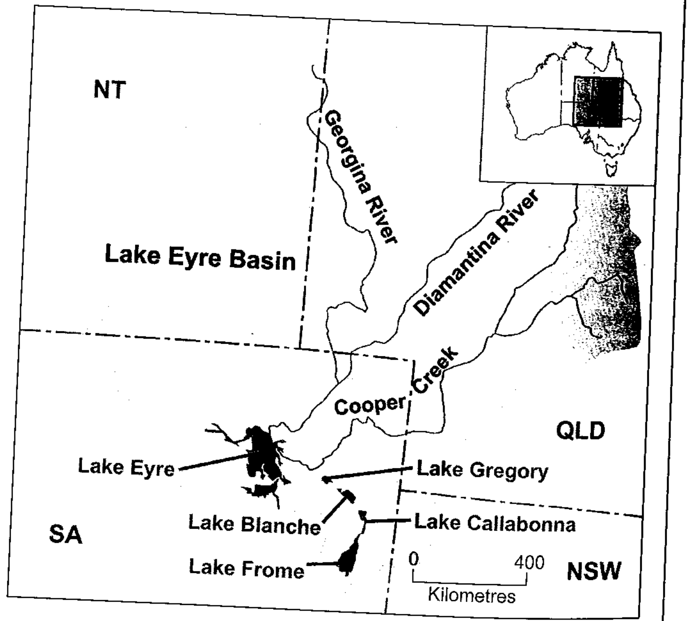


Figure 9.6 The Lake Eyre Basin showing the Channel Country catchment of southwestern Queensland.

Studies of flow dynamics in the region's modern channels have also been important for understanding fluvial processes in arid environments (Rust 1981; Knighton & Nanson 1994; Knighton & Nanson 2002). Gibling, Nanson and Maroulis (1998) examined sedimentary processes of the anastomosing rivers of the Channel Country (Figure 9.7). Many sedimentological features of the modern channel were recognised from older local deposits, indicating that anastomosing rivers have been gently aggrading in the region for the past 100 000 years.

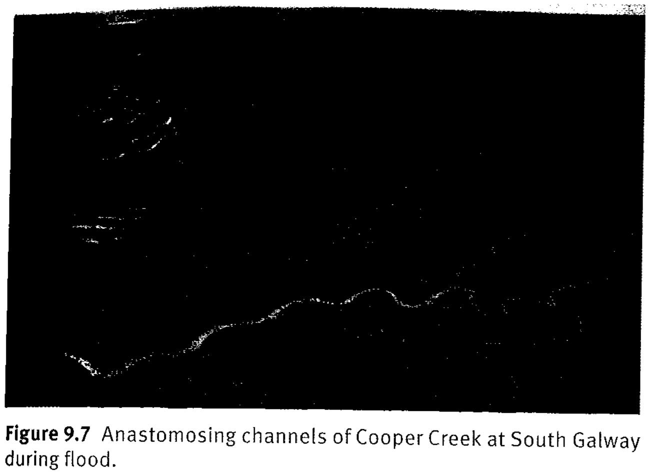


Figure 9.7 Anastomosing channels of Cooper Creek at South Galway during flood.

9.4.2 Fitzroy River Basin

The Fitzroy River Basin of central eastern Queensland comprises the six major modern subcatchments of the Isaacs, Comet, Nogoa, McKenzie, Dawson and Fitzroy rivers (Figure 9.2). Modern river patterns include anabranching, single-channel meandering and confined meandering systems (Amos et al. 2008). Although the Quaternary fluvial history of the region is poorly known relative to that of the Channel Country, the region is critical for understanding long-term changes in a transitional zone that is influenced by both the northern monsoon and southern trade winds. Moreover, investigations from the region offer an alternative proxy of late Quaternary climate changes that can be tested directly against the palaeontological record from the nearby Mount Etna region.

In one of the few studies in the region, Croke et al. (2011) produced 41 preliminary OSL ages for fluvial bedload deposits within the broader Fitzroy River Basin. Fluvial activity is broadly coincident with that observed for the Channel Country, with peak activity occurring during well-defined and discrete phases at 110–90 ka (MIS 5), 50–40 ka (MIS 3) and 30–10 ka (MIS 2/3). Few deposits postdating 14 ka have been recorded in the basin, suggesting a reduction in stream power during the Holocene (Croke et al. 2011). This is also associated with a change from meandering to anabranching in the Nogoa subcatchment. Overall, fluvial activity does not appear to have been synchronous between the Fitzroy River Basin's subcatchments (Croke et al. 2011). Today, the subcatchments of the basin exhibit highly variable mean annual discharges reflecting, in part, highly variable spatial patterns in local precipitation (Amos et al. 2008). This may also extend back through the late Pleistocene. Overall, there appears to be a marked decline in system energy over the past 100 ka, similar to the observation for the Channel Country.

9.4.3 Northern Queensland

In comparison to rivers from arid and temperate zones, there has been little focus on the Quaternary history of rivers in the tropics (Tooth & Nanson 1995). This is partly due to difficulties in obtaining easily datable materials from deposits that occur within humid and extremely seasonal climates. The problem is compounded by the fact that the tropics are home to highly oxidising environments, making datable samples difficult to source (Nanson et al. 1991).

Despite the obvious challenges in working in the tropics, Nanson et al. (1991) developed a TL and U–Th based geochronological history for a large fan delta of the lower Gilbert River, northwestern Queensland (Figure 9.2). They identified an extensive sand body dated at 85 ka, representing a peak period in fluvial activity that has not been seen since. Although a maximum age for the deposit is unclear, Nanson et al. (1991) suggested that the sand body could be as old as 120 ka; if that interpretation is correct, the peak fluvial activity slightly predates a similar peak identified for the Lake Eyre Basin (Nanson, Price & Short 1992; Nanson et al. 2008). Muds and fine sands are superimposed on the larger basal sand body, intersected by a thin medium sand unit dated at 40–50 ka. A series of early Holocene sandy distributaries overlie the fan surface, suggesting a higher energy fluvial phase than at present (Nanson et al. 1991). An early Holocene increase in the intensity of the northern monsoon is also interpreted on the basis of a geochronological and isotopic study of a stalagmite from the nearby Chillagoe Caves (Xia, Zhao & Collerson 1998).

Using TL and accelerator mass spectrometry (AMS) ^{14}C dating, Nott, Thomas and Price (2001) identified extensive late Pleistocene alluvial fan and debris flow deposits from near Cairns in the wet tropics of northeastern Queensland (Figure 9.2). Deposition occurred during 27–14 ka, coincident with the driest phase of the last glacial cycle. A reduction in stream activity during that period led to the development of the fan (Thomas, Nott & Price 2001). Slope instability was likely a result of climatic desiccation and reduced vegetative cover (Nott, Thomas & Price 2001). Although alluvial fan development and landslides have also occurred throughout the Holocene, a return to greater precipitation and concomitant re-establishment of local forests has meant that such depositional processes have been less frequent than during the late Pleistocene.

koala (*P. cinereus*) at least until the late Pleistocene, when it became extinct (Price 2008a). The youngest appearance age for the giant koala on the entire continent is that for the Cement Mills specimen. Significantly, this young record predates the hypothesised late Pleistocene megafauna's extinction 'window' (40–50 ka). Price (2012a) noted that late Pleistocene pre-human climate-driven habitat changes were likely responsible for a massive retraction in the geographic range of the modern koala; it is possible that the giant form became extinct during coeval environmental disturbances.

9.4 Fluvial systems

Although Queensland has numerous active rivers, creeks and streams, particularly in the higher rainfall zones of the northern and eastern margin, the Quaternary history of fluvial systems in semi-arid and arid environments has attracted the greatest attention. Although most studies of Queensland's Quaternary fluvial systems have largely had a palaeoclimatic focus, they have also been important for developing a greater understanding of sedimentological processes of inland rivers.

9.4.1 Channel Country

The Channel Country of southwestern Queensland (Figure 9.2) encompasses the catchments of the Georgina River, Diamantina River and Cooper Creek (Figure 9.5), which drain into the Lake Eyre Basin of the central arid zone (Figure 9.6). Although such catchments may lie dormant for several years at a time, several high-energy fluvial episodes have been identified during the Pleistocene. These fluvial episodes are among the most extensively studied for the Quaternary in Queensland (Rust & Nanson 1986; Nanson, Price & Short 1992; Gibling, Nanson & Maroulis 1998; Maroulis et al. 2007; Nanson et al. 2008).

OSL and TL dating (both targeting sediments) have been the key to establishing the chronology of late Quaternary fluvial deposition. For example, Rust and Nanson (1986), Nanson, Chen and Price (1992) and Nanson et al. (2008) dated numerous palaeochannel and overbank sequences in the Channel Country.

Figure 9.5 The Cooper Creek flood plain, channels and adjacent waterholes.

With an enlarged dataset of dated sequences from other sites within the Lake Eyre Basin (>150 OSL and TL dates), Nanson et al. (1992b, 2008) demonstrated that high-energy fluvial conditions dominated Marine Isotope Stage (MIS) 7 (250–230 ka). Mean bank-full discharges on Cooper Creek are estimated to be up to seven times larger than those of today (Nanson et al. 2008). Numerous other central rivers then became laterally restricted between MIS 7 and MIS 5 (130–74 ka). In the late Pleistocene, peak fluvial activity was greatest at around 110 ka (MIS 5d), rather than at the height of the last interglacial maximum of MIS 5e (132–122 ka), occurring at a time when sea levels and temperatures were lower than today. Nanson et al. (2008) noted a return to fluvial conditions in MIS 4 and 3 (~65–50 ka), although not to the degree seen in MIS 5. Dating of stranded beach ridges from Lake Eyre can also be used as a proxy to examine episodes of flow from the Channel Country. For example, Magee et al. (2004) showed that lake-full conditions were progressively less common through the late Pleistocene, with each subsequent filling event failing to reach the same level as previous lake-full episodes (Magee et al. 2004). Cohen et al. (2011, 2012) also demonstrated that the last time Lake Eyre was connected to Lake Mega-Frome in the south (filled partly from floodwaters from the Channel Country via overflow from Lakes Eyre, Gregory, Blanche and Callabonna) was between 47 and 50 ka. After that time, fluvial activity in the Channel Country was greatly reduced in the lead-up to the Last Glacial Maximum (Nanson et al. 2008).

Significantly, the results suggest that although broad Milankovitch climate-driven shifts have oscillated between wet and dry conditions, there has been a progressive decline in precipitation since MIS 7. Moreover, the chronology of late Quaternary fluvial activity closely matches that of southeastern Australia, rather than northern Australia (Nanson, Price & Short 1992). Combined with the observation that the MIS 5d peak in fluvial activity occurred at a time when temperatures were lower than today, the results suggest that the northern monsoon was not a major source of moisture in the region. Nanson et al. (2008) proposed an alternative model where a western Pacific warm pool was trapped adjacent to eastern Australia, establishing a semipermanent La Niña, with westerly trade winds irrigating the Channel Country and the Lake Eyre Basin.

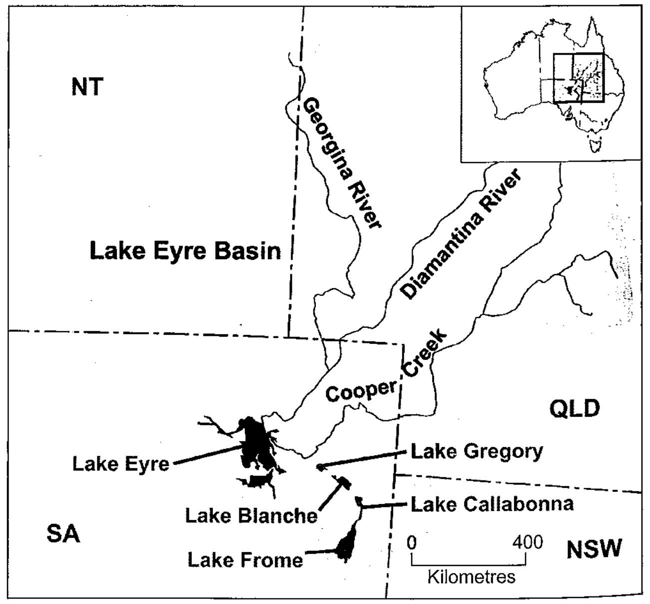


Figure 9.6 The Lake Eyre Basin showing the Channel Country catchment of southwestern Queensland.

Studies of flow dynamics in the region's modern channels have also been important for understanding fluvial processes in arid environments (Rust 1981; Knighton & Nanson 1994; Knighton & Nanson 2002). Gibling, Nanson and Maroulis (1998) examined sedimentary processes of the anastomosing rivers of the Channel Country (Figure 9.7). Many sedimentological features of the modern channel were recognised from older local deposits, indicating that anastomosing rivers have been gently aggrading in the region for the past 100 000 years.

Figure 9.7 Anastomosing channels of Cooper Creek at South Galway during flood.

9.4.2 Fitzroy River Basin

The Fitzroy River Basin of central eastern Queensland comprises the six major modern subcatchments of the Isaacs, Comet, Nogoa, McKenzie, Dawson and Fitzroy rivers (Figure 9.2). Modern river patterns include anabranching, single-channel meandering and confined meandering systems (Amos et al. 2008). Although the Quaternary fluvial history of the region is poorly known relative to that of the Channel Country, the region is critical for understanding long-term changes in a transitional zone that is influenced by both the northern monsoon and southern trade winds. Moreover, investigations from the region offer an alternative proxy of late Quaternary climate changes that can be tested directly against the palaeontological record from the nearby Mount Etna region.

In one of the few studies in the region, Croke et al. (2011) produced 41 preliminary OSL ages for fluvial bedload deposits within the broader Fitzroy River Basin. Fluvial activity is broadly coincident with that observed for the Channel Country, with peak activity occurring during well-defined and discrete phases at 110–90 ka (MIS 5), 50–40 ka (MIS 3) and 30–10 ka (MIS 2/3). Few deposits postdating 14 ka have been recorded in the basin, suggesting a reduction in stream power during the Holocene (Croke et al. 2011). This is also associated with a change from meandering to anabranching in the Nogoa subcatchment. Overall, fluvial activity does not appear to have been synchronous between the Fitzroy River Basin's subcatchments (Croke et al. 2011). Today, the subcatchments of the basin exhibit highly variable mean annual discharges reflecting, in part, highly variable spatial patterns in local precipitation (Amos et al. 2008). This may also extend back through the late Pleistocene. Overall, there appears to be a marked decline in system energy over the past 100 ka, similar to the observation for the Channel Country.

9.4.3 Northern Queensland

In comparison to rivers from arid and temperate zones, there has been little focus on the Quaternary history of rivers in the tropics (Tooth & Nanson 1995). This is partly due to difficulties in obtaining easily datable materials from deposits that occur within humid and extremely seasonal climates. The problem is compounded by the fact that the tropics are home to highly oxidising environments, making datable samples difficult to source (Nanson et al. 1991).

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Wohl (1992) examined the occurrence and frequency of slackwater deposits to reconstruct a late Holocene flood history for the Herbert Gorge, south of Cairns (Figure 9.2). Step-backwater modelling of water-surface profiles indicates that at least six flows, estimated between 11 000 and 17 000 m^3/s, occurred in the gorge over the past 900 years. Recognition of the processes and occurrence of such high-energy fluvial phases is instrumental in understanding the geomorphological factors that have shaped the modern Herbert Gorge.

Numerous other Quaternary fluvial deposits have been identified in northeastern Queensland, but have largely been the focus of palaeontological investigation. For example, the fluvial deposit at Terrace Site, Riversleigh, in northwestern Queensland (Figure 9.2), contains the remains of extinct Pleistocene megafauna. Although conventional ^{14}C dating of charcoal associated with the fossils suggests deposition around 24 ka (Davis & Archer 1997), the results are considered unreliable (Gillespie, Brook & Baynes 2006).

Extensive Pliocene–Pleistocene river terrace deposits occur in the Floraville region just south of the Gulf of Carpentaria (Figure 9.2). The discrete deposits are largely dominated by sands and gravels adjacent to the Leichhardt River (Archer 1982, 1984). The deposits are stratigraphically complex and it is difficult to trace and correlate individual deposits over wider areas. Many of the local fossil assemblages contain abundant remains of the megafaunal marsupial *Diprotodon* (Figure 9.8), a taxon commonly considered to be a biochronological marker of Pleistocene deposits (Mackness & Godthelp 2001). However, some of the local *Diprotodon*-yielding deposits also contain the remains of the Pliocene *Euryzgoma*, the hypothetical direct ancestor of *Diprotodon* (Archer 1977; Price & Piper 2009). Stratigraphic and taphonomic aspects of such deposits are poorly known, and so it is unclear whether the associated assemblages are temporally mixed or both taxa were truly coeval when alive. Resolving such a dichotomy will be useful for future biochronological studies. In a preliminary dating study, Price (unpublished data) produced direct U–Th ages of ~100 ka for a fossil maxilla of the giant extinct goanna, *Megalania* (*Varanus priscus*). Although the fossil faunal assemblages appear to be diverse and abundant, they have not been studied extensively. Archer (1982) described the small fossil marsupial *Sminthopsis floravillensis* from the region, while Tyler, Archer and Godthelp (1994) identified three species of frogs on the basis of fossils. One of the frog species, the spotted marsh frog (*Limnodynastes tasmaniensis*), is today extinct in the region.

A diverse Pleistocene fossil fauna has also been identified from the Wyandotte Formation, southwest of Cairns (Figure 9.2). Eleven discrete fossil units have been identified in the formation outcropping along the banks of Wyandotte Creek. The deposits are typically indurated, being dominated by gravels, sands and clays. McNamara (1990) suggested that on geomorphological grounds, the fossil-bearing deposits must be younger than nearby basalt from the McBride Province (Section 9.8.1) dated at 410 ka (Griffin & McDougall 1975) and may be younger than 200 ka. Two major fossil horizons have been identified: the basal Unit A and overlying Unit B (McNamara 1990). The base of Unit A is older than 45 ka, while the basal portion of Unit B was dated to 30 ka (conventional ^{14}C dating). However, Gillespie, Brook and Baynes (2006) suggested that such dates are unreliable. Both extant and extinct Pleistocene megafaunal taxa have been identified from the deposits (Gaffney & McNamara 1990; McNamara 1990). Interestingly, the stratigraphically older Unit A is noticeably more abundant in megafaunal taxa than Unit B (six v. two taxa, respectively), mirroring similar long-term declines in Quaternary biological diversity as seen in the Mount Etna, Texas Caves and Darling Downs regions.

9.4.4 Darling Downs

The Darling Downs, situated just west of the Great Dividing Range in southeastern Queensland (Figure 9.2), is bounded to the north by the Bunya Mountains and to the south by the Herries Ranges; it slopes gently westwards into the plains of central Australia. The Great Dividing Range forms a significant watershed in eastern Australia and separates several significant drainage basins. The modern catchments of the Darling Downs are dominated by perennial meandering creeks and channels that drain into the ephemeral Condamine River. The Condamine River forms the headwaters of the Darling River Catchment, which eventually connect to the Murray River to form the Murray–Darling Basin, draining into the southern ocean in southeastern South Australia.

During the late Paleogene – early Neogene, a sequence of volcanic eruptions around the Toowoomba region produced a series of predominantly basaltic flows overlying Mesozoic sandstones (Thompson & Beckmann 1959). Subsequent dissection and erosion of the sandstones and basalts developed several landforms in the region including the Toowoomba Plateau, basaltic uplands, steep eastern slopes of the Great Dividing Range, and alluvial flood plains (Thompson & Beckmann 1959). Deep lateritic soils developed from erosion of basalts on the eastern edge of the range. Additional erosion of underlying sandstones and basalts led to the development of dark clay, sand and silt with abundant calcrete nodules, and is characteristic of the extensive alluvial plains in the region (Woods 1960; Gill 1978; Price & Sobbe 2005; Price & Webb 2006). The sediments are alkaline, as evidenced by the high amounts of preserved calcareous materials such as fossil molluscs and calcretes (Webb et al. 2007).

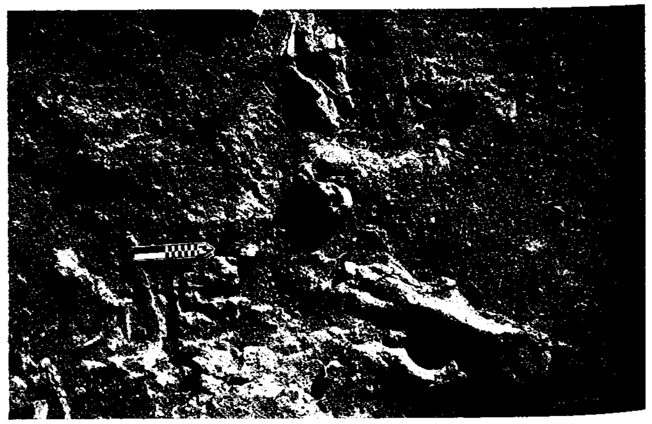


Figure 9.8 *Diprotodon* skeleton eroding from fluvial sediments from Floraville, northern Queensland. The black and white photo scale is 250 mm long.

The Darling Downs is remarkable for the diversity and abundance of Quaternary fluvially derived fossil deposits. Molnar and Kurz (1997) identified over 50 specific localities where fossils have been recorded from the relatively small region (~200 × 80 km), although the true number of fossil sites would likely exceed several hundred as every local catchment produces discrete fossiliferous units in great abundance (Price 2012b). The fossil-bearing portion of the Darling Downs is divided into two sections (Molnar & Kurz 1997). The eastern Darling Downs is underlain predominately by Pleistocene deposits and is situated to the east of the Condamine River. The older Pliocene Chinchilla Local Fauna is located in the western Darling Downs (Bartholomai & Woods 1976). It is possible that Pleistocene deposits also occur in the Chinchilla region.

Macintosh (1967), Gill (1978) and Sobbe (1990) provided limited stratigraphies for sections along the creeks of the eastern Darling Downs, introducing the units termed 'Tollburra silt', 'Talgai Pedoderm' and 'Ellinthorpe Clay'. However, those names have not been subsequently used (Price & Sobbe 2005; Price, Tyler & Cooke 2005; Price & Webb 2006) and are not considered to be valid stratigraphic units (Molnar & Kurz 1997).

The deposits have been a traditional collecting area for Pleistocene megafauna with numerous specimens sent to England in the 1800s for study by the renowned anatomist Richard Owen. The deposits include the type localities for many now extinct megafaunal taxa (e.g. giant horned land turtle *Ninjemys oweni* and giant short-faced kangaroo *Simosthenurus pales*) and smaller bodied species (e.g. Sobbe's long-nosed bandicoot, *Perameles sobbei;* de Vis 1895; Gaffney 1992; Price 2002; Prideaux 2004).

By far the greatest focus has been on fossil deposits of the Kings Creek Catchment, southern Darling Downs (Roberts et al. 2001; Price 2002, 2005, 2012b; Price & Hocknull 2005; Price & Sobbe 2005, 2011; Price, Tyler & Cooke 2005; Price et al. 2011; Price & Webb 2006; Webb et al. 2007). Several assemblages have been located in the catchment and include the geologically youngest record of articulated megafaunal skeletons on the continent (Neds Gully). The fluvial sediments are typical of most deposits of the Darling Downs (Figure 9.9), composed of fine clays (particularly floodplain deposits), grading towards sands and gravels (higher energy proximal overbank and channel deposits) and cobbles (mostly confined to channel deposits; Price & Sobbe 2005; Price & Webb 2006). Fossils are most common within the higher energy laterally accreted deposits, rather than low-energy vertically accreted sediments.

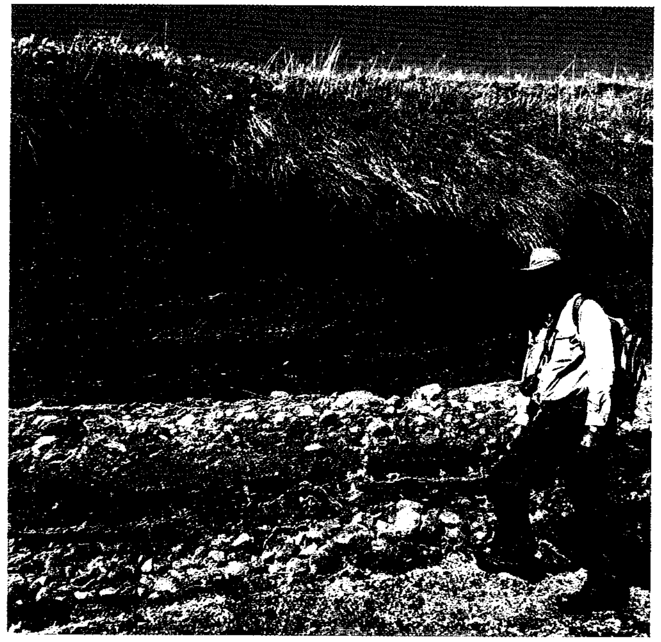


Figure 9.9 Late Quaternary palaeochannel showing alternating high- and low-energy deposition exposed in cross-section at Neds Gully, Darling Downs, southeastern Queensland.

Major fossiliferous units identified in Kings Creek Catchment (Price & Sobbe 2005; Price & Webb 2006) preserve evidence of a long-term local decline in faunal diversity (including both megafauna and smaller taxa). OSL and U–Th dating demonstrates that the oldest and most diverse fossiliferous horizon dates to around 120 ka. Local megafaunal diversity declined progressively from 15 species at 120 ka, to just 4 taxa by 80 ka (Price, Tylor & Cooke 2011). The established chronology of fossil deposits is coincident with warm, wet peaks of MIS 5, consistent with deposition during more humid conditions at those times.

The local decline in taxonomic diversity is not related to body size, but rather appears to have been habitat specific (Price, Tyler & Cooke 2005; Price 2005). That is, taxa that had habitat preferences of scrubby vine thickets and closed wooded vegetation were more likely to have declined in abundance, and so the youngest deposits are dominated by forms adapted to open habitats. Taphonomic analyses provide evidence for drought-like mortality profiles of the megafaunal taxa. Similarly, abundant calcrete formation, flashy discharge features, scour surfaces and abundant abrupt lithofacies shifts suggest highly seasonal changes in local climates over the period of deposition (Price & Webb 2006). Significantly, the youngest OSL-dated deposit (46 ± 6 ka; 1σ error; Roberts et al. 2001) in the region (Neds Gully in the upper Kings Creek Catchment) contains the remains of only two articulated species of now extinct Pleistocene megafauna (Figure 9.10). The long-term decline in diversity appears to be part of a longer term (pre-human) trend towards reduced precipitation, increased aridity, retraction of woodlands and opening of local habitats (Price 2012b). Kernich et al. (2009) noted a similar long-term trend towards increasing aridity through time in the Lower Balonne River, part of the broader Condamine River Catchment, just west of the Darling Downs.

While there is evidence that late Pleistocene climate changes had detrimental impacts on local Darling Downs megafaunal populations, human beings (arriving on the continent 45–50 ka) cannot yet be excluded as potential contributors to the extinctions, at least for late surviving forms. Also, it is important

Figure 9.10 Late Pleistocene *Diprotodon* skull (crushed dorsally) and lower jaws from Neds Gully, Darling Downs, southeastern Queensland.

to note that the results from the Darling Downs reflect only local conditions and cannot be used, at least in isolation, to explain megafaunal extinctions in other parts of the continent as traditionally has been the case (Miller et al. 1999; Roberts et al. 2001).

9.5 Lacustrine deposits

With typically long and commonly continuous sedimentological records, lakes have proven to be among the most valuable sources of high-precision palaeoclimatic information for the Quaternary of Queensland (Petherick, Moss & McGowan 2011). By far the most intensively studied Quaternary lakes are those from northeastern Queensland.

9.5.1 Lynch's Crater

Lynch's Crater, northeastern Queensland, has been pivotal in developing a thorough understanding of Quaternary changes in precipitation, temperature and vegetation in Australia's tropics (Figure 9.2). The crater itself is relatively small (500 m diameter) and formed during the middle Pleistocene following an explosive eruption that created an 80 m deep maar. Since that time, the crater has been accumulating lake and swamp sediments, although the surface has been lowered in the historic period due to local anthropogenic agricultural activity and peat mining.

The Quaternary palaeoclimatic record from Lynch's Crater is unmatched for any other terrestrial archive in Australia. Since the initial site description (Kershaw 1974), the lake has been the focus of numerous studies that have investigated its sedimentology, stratigraphy, geochemistry, palynology, taphonomy, chronology and palaeoclimatic signatures (Kershaw 1976, 1983; Bohte & Kershaw 1999; Turney et al. 2001, 2004; Kershaw, Bretherton & van der Kaars 2007; Muller et al. 2008; Coulter et al. 2009; Rieser & Wüst 2010). In the most recent extensive review, Kershaw, Bretherton and van der Kaars (2007) extended the earlier drilling of the lake to develop a long, mostly continuous record dating to ~230 ka (Figure 9.11). This record is significant in that it spans the three most recent glacials and interglacials, through the period of human colonisation, megafaunal extinction and the Last Glacial Maximum. Glacials and interglacials are clearly demarcated on the basis of marked changes in local vegetation, with the warm and wet interglacials dominated by rainforest angiosperms, whereas the cool, dry glacials are characterised by araucarian forest and sclerophyllous vegetation (e.g. *Eucalyptus*). It is clear that orbital forcing on the vegetation has been a controlling factor (Kershaw, Bretherton & van der Kaars 2007).

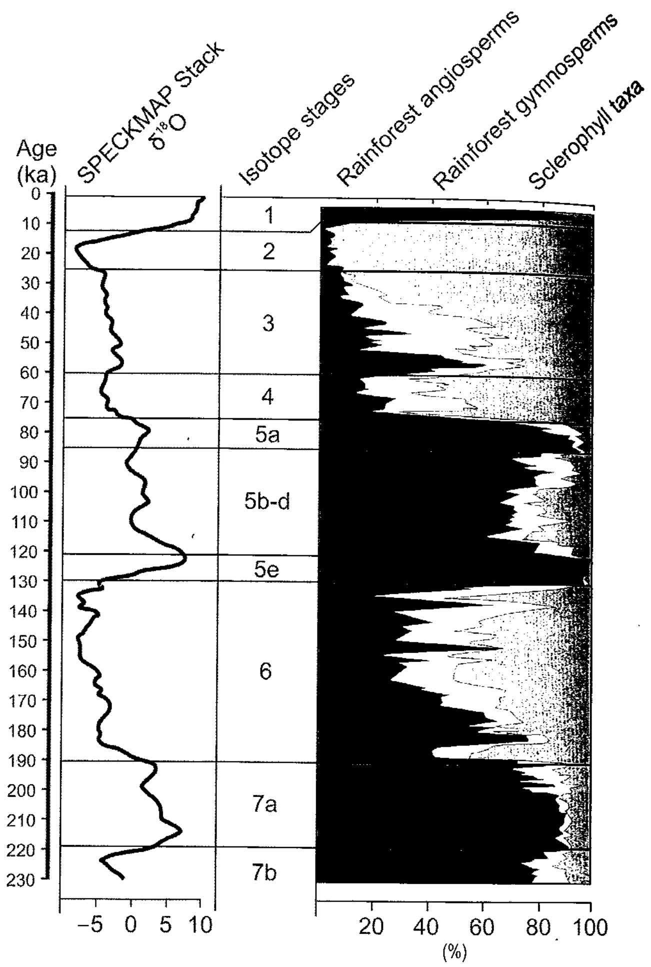


Figure 9.11 Summary diagram showing relative proportions of rainforest and sclerophyll taxa, relative to oxygen isotope curve and isotope stages since 230 ka, at Lynch's Crater, northeastern Queensland (modified from Kershaw, Bretherton & van der Kaars 2007).

An apparent increase in charcoal after 45 ka has been interpreted to be directly related to anthropogenic burning of the landscape (Kershaw et al. 2002). This marked incidence of charcoal may serve as a proxy for indicating earliest human colonisation in the area (Kershaw 1976). This hypothesis has not seen universal support (White & O'Connell 1982), partly because the onset of significant burning predates the earliest archaeological records for the region by >5000 years (David et al. 2007). In any case, an increase in charcoal after 45 ka is associated with a substantial reduction in rainforest cover, major expansion in sclerophyllous vegetation and generally more open habitats. Kershaw (1976, 1983), Kershaw et al. (2002) and Kershaw, Bretherton and van der Kaars (2007) suggested that the vegetative characteristics of the local habitat was then maintained through anthropogenic burning, at least between the late Pleistocene and early Holocene. However, as an independent test, Stevenson and Hope (2005) examined a 130 000-year pollen core from New Caledonia and noted that a similar shift after 45 ka towards an expansion in sclerophyllous vegetation and reduction in rainforest was not associated with an increase in the incidence of fire. Significantly, New Caledonia was not populated by human beings until 3 ka, so the results suggest that late Pleistocene vegetation changes must have been driven solely by climatic processes (Stevenson & Hope 2005). Mooney et al. (2011) examined 233 independent sedimentary charcoal records across Australasia to determine temporal and spatial patterns of burning for the past 70 000 years. The combined record shows that the incidence of fire is closely associated with climatic shifts and that there was no distinct change in fire regimes at the hypothesised time of human arrival. There is no compelling evidence to suggest that human beings exerted any marked control on the landscape through burning, at least until the latest Holocene (Mooney et al. 2011).

The view from the more complete pollen record at Lynch's Crater demonstrates that a longer term decline of fire-sensitive vegetation may have been initiated around 130 ka (Kershaw, Bretherton & van der Kaars 2007). *Eucalyptus* became more dominant from around the beginning of MIS 4 (~70 ka), predating the onset of burning (Kershaw, Bretherton & van der Kaars 2007). These observations suggest that climate change (including insolation forcing and the intensification of El Niño southern oscillation) largely controlled the late Pleistocene vegetation.

9.5.2 Lake Carpentaria (AR Chivas, A García and DI Cendón)

The modern drainage catchment of the Gulf of Carpentaria (Figure 9.2) comprises more than 25 monsoon-fed rivers that collectively provide 25% of the total fluvial run-off of the Australian continent (Figure 9.12). The Gulf is bounded to the east by Torres Strait (maximum water depth 12 m) and by the Arafura Sill (water depth 53 m) to the west. The deepest, eastern central portion of the Gulf today is 71 m (Figure 9.13). Accordingly, during episodes of low Quaternary sea levels, the Gulf was separated from the open waters of the Indian and Pacific oceans and formed the large freshwater-brackish Lake Carpentaria (Torgersen et al. 1983, 1985, 1988; Jones & Torgersen 1988). When most recently filled, the lake had a maximum depth of 15–18 m and lateral dimensions of about 600 × 300 km, with overflow westwards into the Arafura Sea (Chivas et al. 2001). Among modern lakes, only the Caspian Sea has a larger surface area. Throughout the past glacial cycle of ~125 000 years, there was at least one land bridge between Australia and New Guinea for more than 90% of this time. The seismic profile across the Gulf (Figure 9.14) shows multiple unconformities that represent lake deposits, cut by outlet channels and formed at times of low sea level transgressed by successive marine inundations. The depocentre has migrated progressively eastward. It is possible that many of the past low sea-level lacustrine episodes of the Quaternary are preserved in superposition in this slowly subsiding continental margin (Edgar et al. 2003; Sandiford 2007; Sandiford et al. 2009) as a marine/lacustrine equivalent of the Chinese terrestrial loess sequences. However, only the uppermost cycle of the past 130 000 years, represented by 15 m of sedimentation, has been cored (Chivas et al. 2001; Reeves et al. 2008). The Arafura Sill is composed of sediment rather than bedrock and has been cut and infilled to various depths in the past (seismic sections in Edgar et al. 2003). Accordingly, the water depth and lateral extent of Lake Carpentaria will have varied during past lowstand episodes of lake filling. Indeed, each lacustrine episode need not have filled its available basin volume, depending upon the hydrological balance, as is evident by the palaeosols at lake margins during the most recent cycle.

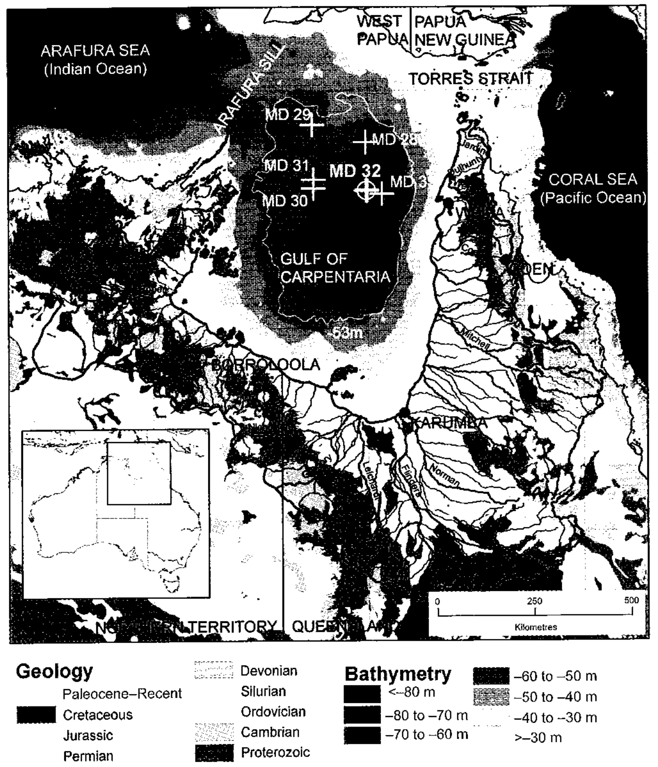


Figure 9.12 Drainage catchment of the modern Gulf of Carpentaria (from Playa et al. 2007) showing the potential maximum extent of palaeolake Carpentaria (white bathymetric contour at −53 m).

Numerous short sedimentary cores (35 cores, commonly to 2 m) were collected by Torgersen in 1982, and longer cores (4–15 m) were collected in 1997 (Chivas et al. 2001). The latter have been extensively studied and dated by AMS ^{14}C, OSL, TL and amino acid racemisation techniques (Chivas et al. 2001; Reeves et al. 2008). The sedimentary record of the past 130 000 years includes cycles of subaerial exposure, marine transgression, regression, and freshwater- and saline-lake phases. This diverse sedimentary record has been used as a testbed for new analytical techniques to deduce palaeosalinity, including by trace-element (de Dekker et al. 1988) and strontium-isotope (McCulloch, de Dekker & Chivas 1989) geochemistry of ostracods, and boron isotopes (Vengosh et al. 1991). Detailed studies of ostracods (Reeves et al. 2007), coccoliths (Couapel et al. 2007), palynomorphs, foraminifers, charophytes and molluscs have also been instructive. Of note is the discovery of the new mollusc *Lentidium origolacus* (Hallan & Willan 2010), which was the most abundant mollusc throughout the 130 000-year history of Lake Carpentaria, and which has since been found extant in the slightly saline refugia of several estuaries in the southeast of the Gulf of Carpentaria.

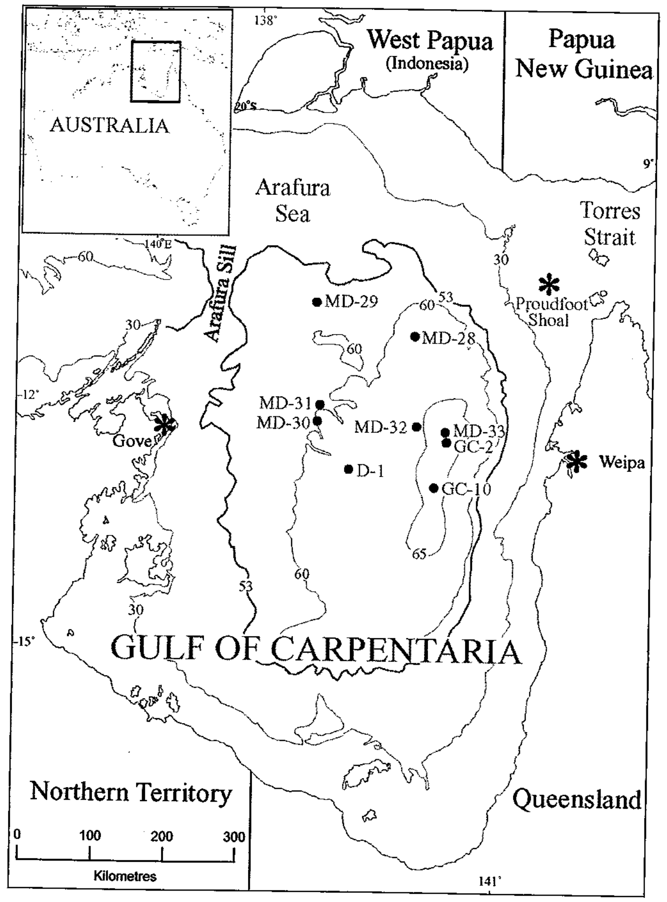


Figure 9.13 Bathymetric map (from Chivas et al. 2001) of the Gulf of Carpentaria (bathymetry from Grim and Edgar 1998) showing the position of the Arafura Sill and core locations. MD indicates long cores collected by the French research vessel *Marion Dufresne*. D-1 refers to oil exploration well Duyfken 1, which penetrated 210 m of Pliocene–Pleistocene strata and terminated in Proterozoic rhyolite at 1113 m.

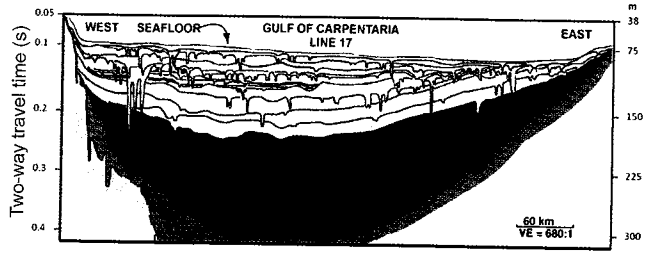


Figure 9.14 Interpretative line drawing of a seismic section (line 17 of US Geological Survey transects of 1993–94) from Gove to Weipa (modified from Chivas et al. 2001; Edgar et al. 2003). The west–east distance is ~600 km. Also shown are basin-wide reflections and incised channels (width not to scale) at exposure surfaces. The right-hand scale shows the approximate depth in metres, based on the velocity of sound in water.

During a late phase of MIS 6, the lake was subaerially exposed with extensive rivers meandering across the Carpentaria Basin. A significant marine transgression occurred during MIS 5 (130–84 ka) followed by a marine/estuarine phase up until 70 ka (Reeves et al. 2008). During 70–40 ka, a protracted lacustrine episode meant that the lake varied in depth and salinity, and at times was almost completely desiccated (Reeves et al. 2008). During that episode (MIS 4), extensive evaporites formed, taking on varve-like morphology (Playà et al. 2007). The thickness of the calcite–gypsum couplets closely resemble those from modern evaporitic environments, and along with evidence for numerous alternations between dry and wet environments, the results suggest that the region experienced reduced monsoon rainfall patterns during those evaporitic phases (Playà et al. 2007).

Freshwater conditions were largely maintained during 40–12.2 ka (Reeves et al. 2007, 2008). Palynological analyses of the sedimentary cores suggest that savanna grasslands formed a dominant habitat around the lake over the late Pleistocene – early Holocene transition (Torgersen et al. 1988), with no evidence of lowland rainforest as in southern regions of modern New Guinea (Chivas et al. 2001). Peak freshwater conditions occurred ~14 ka (Reeves et al. 2008) at a time when the lake is estimated to have been near full, reaching 600 × 300 km and a depth of 15 m (Chivas et al. 2001). The lake-full episode is likely to have been caused by the intensification/restoration of the Australian monsoon, but was short-lived (Reeves et al. 2008). A subsequent marine transgression (above the –53 m bathymetric contour, relative to present-day sea level) inundated the lake, with full marine conditions being restored by 10.5 ka and breaching of Torres Strait at ~8 ka (Reeves et al. 2008). Coral reefs were developed in southern areas of the Gulf of Carpentaria after 10.5 ka (Harris et al. 2004a, 2008), but were seemingly drowned by 7 ka as the sea level continued to rise.

9.5.3 Other northern lakes (GJ Price)

Numerous other lakes have contributed significantly to our understanding of late Quaternary climatic and vegetation changes in northeastern Queensland (Figure 9.15). Although records from lakes such as Bromfield Swamp (Kershaw 1975), Quincan Crater (Kershaw 1971), Lake Euramoo (Kershaw 1970; Haberle 2005; Tibby & Haberle 2007), Lake Eacham (Constable & McElhinny 1985), Lake Barrine (Chen 1988; Dimitriadis & Cranston 2001; Walker 2007, 2011), Strenekoff's Crater (Kershaw et al. 1991), Three-Quarter Mile Lake (Luly, Grindrod & Penny 2006) and Whitehaven Swamp (Genever, Grindrod & Barker 2003) typically span only parts of the latest Pleistocene through to the Holocene, they have been instrumental in supporting and testing hypotheses established for the more intensively studied record from Lynch's Crater. Such records confirm that glacial–interglacial cyclity has been paramount in controlling local vegetation, with anthropogenic activities potentially imprinting on the late Quaternary record of change. Interestingly, unlike at Lynch's Crater, rainforest taxa are commonly absent from many of the other northeastern lake records at the Last Glacial Maximum (Kershaw 1994; Haberle 2005; Moss & Kershaw 2007). Although it is widely accepted that rainforests declined markedly in the lead-up to the Last Glacial Maximum, the paucity of rainforest pollen may also partly be the result of a taphonomic bias where low pollen producing rainforest angiosperms are not sampled as frequently as high pollen producing woodland species. Thus, a paucity of rainforest pollen may not necessarily reflect the complete decimation of such habitat at the Last Glacial Maximum (Pickett et al. 2004; Petherick, Moss & McGowan 2011).

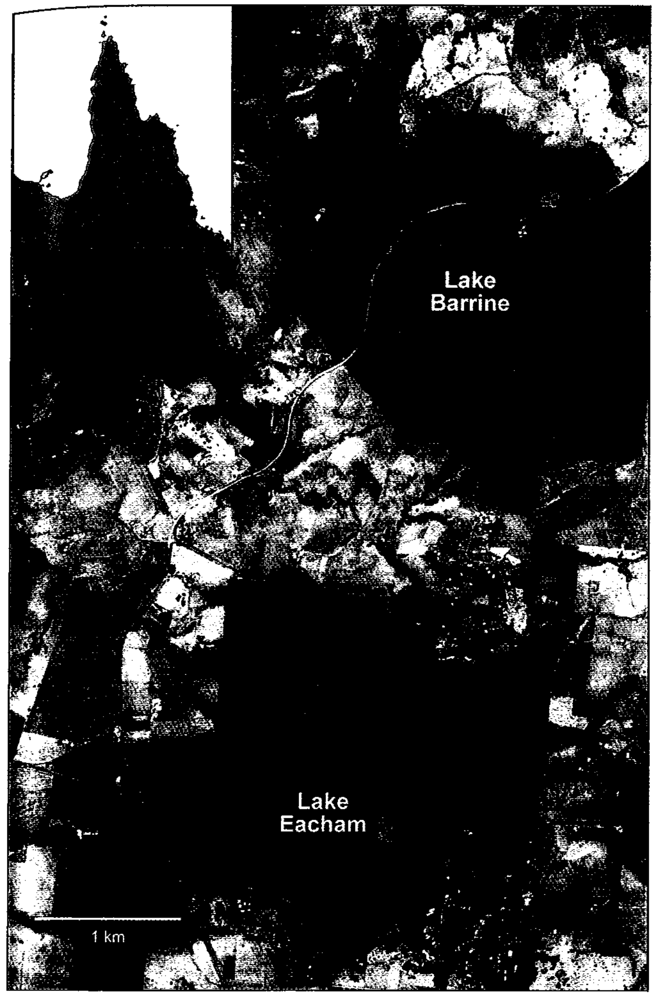


Figure 9.15 Aerial photograph of Lake Barrine and Lake Eacham (both crater lakes) in northeastern Queensland.

9.5.4 Lake Buchanan

Although Lake Buchanan has not been as extensively studied as those lakes and swamps of northeastern Australia, it is the oldest dated Quaternary lake in Queensland. Situated adjacent to the Great Dividing Range in central eastern Queensland (Figure 9.2), Lake Buchanan is an intermontane playa of the semi-arid zone. Although the modern lake is only 23 × 7 km, stranded beach ridges and cliffed lacustrine sediments suggest past episodes when the lake was substantially larger (Chivas et al. 1986). A 15 m core was extracted from the modern basin and dated using palaeomagnetic methods by Chivas et al. (1986). The analyses identified the Brunhes–Matuyama boundary (780 ka; Pillans 2003) at 5 m, while the top of the Jaramillo subchron (920 ka) was identified at 6.3 m. Thus, sedimentation of only ~7 m was calculated for the last million years (Chivas et al. 1986). Chivas et al. (1986) estimated that the base of the core dates to 1.5–2 Ma. Fossil assemblages are dominated by ostracods and charophytes, with gastropods, foraminifers and vertebrates also present. For the early history, conditions varied between a playa and shallow lake. Abundant palaeosols in the cores imply long and frequent episodes of desiccation, with four major wet phases since 780 ka (Chivas et al. 1986). Overall, the Lake Buchanan record points to an increasingly arid climate through the Quaternary.

9.5.5 Lakes on southeastern sand islands

Freshwater lakes on the sand islands off southeastern Queensland have provided key palaeoenvironmental and palaeoclimatic information over the late Quaternary (Figure 9.2). Numerous lakes have been documented; these include Lake Allom and Old Lake Coomboo Depression on Fraser Island (Longmore 1997; Longmore & Heijnis 1999; Donders, Wagner & Visscher 2006) and Tortoise Lagoon, Native Companion Lagoon, Blue Lake and Myora Springs on North Stradbroke Island (McGowan, Petherick & Kamber 2008; Petherick, McGowan & Moss 2008, 2009, 2010; Petherick et al. 2008; Moss et al. 2009). Each record spans at least the late Pleistocene (generally <40 ka) and/or Holocene.

The perched Old Lake Coomboo Depression is the oldest lake in the region. A combination of ^{14}C and U–Th dating, combined with sedimentation rate modelling, suggests that a basal age for the extracted core is around 600 ka (Longmore & Heijnis 1999). A pollen and sedimentary analysis demonstrates a shift from an *Araucaria*-dominated rainforest and deepwater lake at 600 ka to a drier *Podocarpus*-dominated rainforest and intermediate lake levels by 350 ka (Longmore 1997).

Records from other lakes in the region show major habitat changes in the lead-up to the Last Glacial Maximum, principally a shift from rainforest to open woodland vegetation. For example, the Native Companion Lagoon pollen record demonstrates that prior to 31.4 ka, a mixed habitat dominated by *Eucalyptus* and some rainforest taxa (e.g. Araucariaceae, Arecaceae, Sapotaceae and *Syzygium*) occurred proximal to the lake. On the basis of sediment moisture content, charcoal and pollen analyses, two significant cold events were identified, at ~30.8 ka and 21.7 ka (Petherick, McGowan & Moss 2008). During the Last Glacial Maximum, rainforest taxa declined significantly while a Casuarinaceae-dominated woodland saw marked expansion presumably reflecting drier local conditions (Petherick, McGowan & Moss 2008). More humid conditions returned during the Holocene and reflected the expansion of *Eucalyptus* woodlands and some rainforest taxa (Petherick, McGowan & Moss 2008). It is thought that intensification in El Niño activity during the Holocene resulted in greater climatic variability in the region leading to an increase in heterogeneous habitats (Donders, Wagner & Visscher 2006).

9.6 Aeolian features

Although deserts today occupy approximately 40% of the Australian continent, they showed marked development and expansion during the Quaternary, especially during glacial phases. Despite receiving far less attention than fluvial and lacustrine systems, Queensland's Quaternary aeolian deposits have been the source of important palaeoclimatic information.

9.6.1 Inland deserts

At 130 000 km^2, the Simpson Desert of central Australia is the largest desert on the continent and partly encompasses the Channel Country of southwestern Queensland (Figure 9.2). Today it is dominated by broad-crested linear and typically well vegetated dunes (Brookfield 1970) that are commonly associated with sand provided from interior river systems (Maroulis et al. 2007). Although dune-building episodes have been tentatively identified as far back as ~1 Ma (Rhodes et al. 2005), due to typically high instances of reworking and obliteration, the best studied dunes have been those of the late Quaternary. For example, although numerous arid phases (commonly associated with dune building) have been suggested for the late Quaternary in the western Simpson Desert, the majority of dunes are Holocene in age and possibly reworked from older underlying late Pleistocene dune systems (Nanson, Chen & Price 1992; Nanson, Price & Short 1992; Twidale et al. 2001).

In an integrated OSL dating study of both fluvial and aeolian deposits in the Cooper Creek area of southwestern Queensland (Figure 9.2), Maroulis et al. (2007) identified significant building periods of source-bordering dunes in late MIS 5 (85–80 ka) and MIS 3 (50–30 ka). Both episodes are associated with a significant decline in fluvial activity, so much so that after 40 ka sand channel activity ceased, with the fluvial channels isolating the dunes as emergent features on the landscape (Figure 9.16). Such dunes have undergone little migration since that time, demonstrating remarkable stability over the past 40 000 years. Collectively, the data demonstrate a progressive trend towards increasingly arid climates since 100 ka (Maroulis et al. 2007).

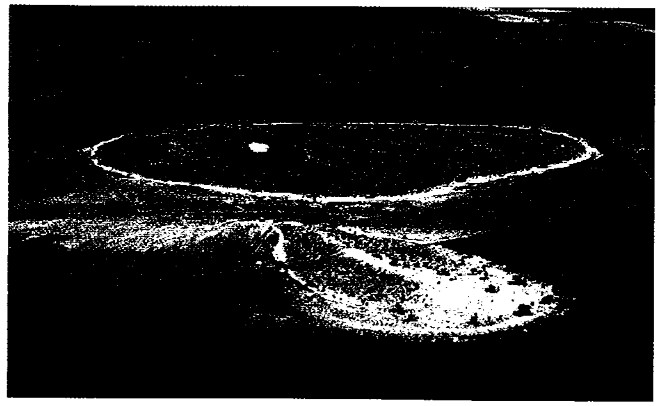


Figure 9.16 Isolated dunes on the Cooper Creek flood plain.

9.6.2 Southeastern sand islands

In addition to providing information about palaeohabitat changes, innovative geochemical 'fingerprinting' techniques of aeolian sediments extracted from cores of freshwater lakes of southeastern Queensland (Figure 9.2) have allowed reconstruction of dust transport pathways for the late Quaternary. The premise behind the approach is to develop a dataset of trace elements of surface sediments from potential source areas with a view to comparing that dataset to the trace-element signatures of aeolian dusts extracted from the cores (Marx, Kamber & McGowan 2005).

McGowan, Petherick and Kamber (2008), Petherick, McGowan and Moss (2008) and Petherick et al. (2008) applied the technique to aeolian sediments deposited in Native Companion Lagoon on North Stradbroke Island. Combined with a robust AMS ^{14}C-dated chronology, they were able to identify four significant aeolian dust transport pathways for the late Pleistocene. Collectively, peak dust input into the lake occurred during 33–18 ka (Figure 9.17). An increase in dust transport at that time implies reduced precipitation and vegetation cover, and an increase in surface sedimentation susceptibility to erosion in the source areas (McTainsh & Strong 2007). During 33–32.1 ka, the dominant dust supply area was the Channel Country, with additional input from the Paroo River of southwestern Queensland. During 31.4–29.8 ka and 22.2–21.7 ka, the source of the aeolian sediments was the Channel Country along with the Murray–Darling Basin and southeastern South Australia. Between 29.7 ka and 23.2 ka, the dominant dust source areas were the Darling River and western New South Wales, with minor input from southwestern Queensland, the Channel Country and Lake Eyre. Towards the peak of the Last Glacial Maximum (20.7–18 ka), dust was sourced predominantly from the Murray–Darling Basin with minor input from Lake Frome, central South Australia, and southwestern Queensland. Overall, the datasets point to distinct variations in atmospheric circulation patterns during the period of deposition at Native Companion Lagoon (Petherick et al. 2008). In combination with independent evidence from the associated pollen record, the data suggest that the two major cold events identified at Native Companion Lagoon were characterised by meridional dry southwesterly winds (Petherick, McGowan & Moss 2008).

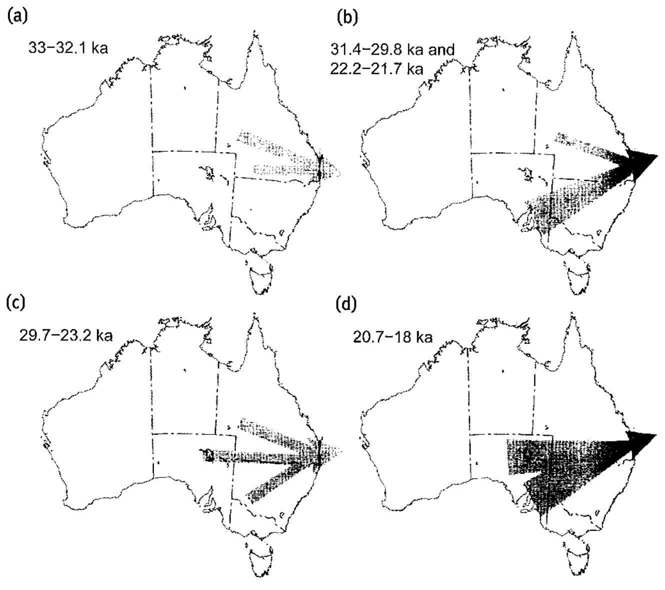


Figure 9.17 Hypothesised aeolian dust source areas and transport pathways for the late Pleistocene at North Stradbroke Island, southeastern Queensland (modified from Petherick, McGowan & Moss 2008).

9.7 Weathering and soil formation

Weathering plays a major role in the geomorphic evolution of Earth's surface, generates unconsolidated material for erosion and thus provides a major contribution to sedimentary basins. Despite this, few studies have specifically investigated the Quaternary history of weathering in Queensland. This is partly due to the greater emphasis on the nature of post-weathering palaeoclimatic events (e.g. subsequent input into fluvial systems), as well as the difficulties associated with establishing the chronology of weathering episodes.

Significant advances have been made, however, due to the emergence and development of new dating techniques (Vasconcelos et al. 1994). For example, Feng and Vasconcelos (2001) used laser heating ^{40}Ar/^{39}Ar analyses of Mn oxides to determine the timing of weathering episodes for the Mary River area of southeastern Queensland (Figure 9.2). Well-defined plateau ages for the Mn oxides were identified and ranged from 346 ± 15 to 291 ± 14 ka (2σ error). Humid conditions during that period are implied on the basis of coeval precipitation of supergene cryptomelane (Feng & Vasconcelos 2001). This palaeoclimatic interpretation is supported by comparison to other independent middle Pleistocene (400–275 ka) palynological and palaeontological records for eastern Queensland that also suggest relatively humid conditions at that time (Longmore 1997; Hocknull et al. 2007; Moss & Kershaw 2007). The weathering episode corresponds with the MIS 9 interglacial (Imbrie et al. 1984).

Feng and Vasconcelos (2007) applied ^{40}Ar/^{39}Ar dating to the Upper Kandanga Mn deposit (Mary River area) and extended the chronology of weathering in the region to almost 1 Ma, with significant plateau ages centred on 1000–800 ka, 630–510 ka, and 313 ± 4 ka and 213 ± 7 ka. Again, the peaks are associated with warm wet periods. Feng and Vasconcelos (2007) interpreted the results to suggest that the rates of weathering vary through time in the region, being partly controlled by climate, with warm and humid conditions typically resulting in enhanced weathering.

The relationship between climate and weathering was also the focus of an earlier study by (Pillans 1997), who investigated the rate of soil development from late Cenozoic basalt in a semi-arid region of northern central Queensland. Flow ages, using K–Ar dating (Coventry, Stephenson & Webb 1985) for basalts from the Nulla and Sturgeon provinces (Figure 9.2), were 10 ka, 890 ka, 2.64 Ma, 3.4 Ma and 5.59 Ma. The sequences show a progressive trend in soil development of only 0.3 m/1 000 000 years, 10–1000 times slower than soil formation on glacial, dune and alluvial deposits (Pillans 1997). Palaeomagnetic dating methods show that weathering rates were greater before 780 ka due to local higher summer rainfall and humidity. Higher rates of weathering of ~3.8 m/1 000 000 years were estimated for an older sequence (dominantly Miocene–Pliocene) in the Dugald River further to the west, during a comparatively warmer and more humid period (Vasconcelos & Conroy 2003). Since 780 ka, the soils of central Queensland have shown incredible stability and are virtually unchanged from chemical weathering or physical disturbance (Pillans 1997). Coventry, Stephenson and Webb (1985) also noted that planation surfaces of the nearby Hughenden Plain (Figure 9.2) have barely been modified by erosional or depositional processes since its formation ~1 Ma (K–Ar date of associated basalts).

9.8 Volcanism

Despite Queensland having no modern active volcanoes, earlier periods of the Quaternary saw multiple episodes of igneous activity. The 35-million-year southward migration of the locus of volcanic magmatism in eastern Australia resulted from northward migration of the continent over a deep mantle plume hot spot (Duncan & McDougall 1989). However, the model cannot account for Quaternary volcanism in Queensland. Tectonically, Australia is a relatively fast moving continent (Debayle, Kennett & Priestley 2005), so eastern Australia exhibits a high degree of asthenospheric shear. This, and heterogeneous viscosity, a near solidus asthenosphere and shear flow—causing melting within the asthenosphere—most likely explains the intraplate Quaternary volcanic magmatism of Queensland (Conrad et al. 2011).

Quaternary basalt flows, although laterally restricted, occurred in distinct provinces in northern and southeastern Queensland. K–Ar dating has shown that magmatism has occurred in eastern Queensland through much of the Cenozoic (Vasconcelos et al. 2008). Although the volcanic assemblages mentioned here may date from the Paleocene to the Holocene (Chapter 8), the following discussion concentrates only on their Quaternary history.

9.8.1 Northern Queensland

Numerous basaltic provinces have been identified in northern Queensland, including Piebald, McLean, Atherton, McBride, Chudleigh, Sturgeon and Nulla (Figure 9.2). Collectively, such areas constitute the greatest activity of Quaternary volcanism in Queensland (Figure 9.18). Three main types of volcanic provinces are typical for eastern Australia—mafic strongly undersaturated in high potassium, central volcano and lava field (Wellman & McDougall 1974). However, only the latter two have been identified in northern Queensland. In most cases, the rocks are hawaiite, with some transitional to alkali olivine basalt and several unsaturated varieties including basanite and nepheline hawaiite (Griffin & McDougall 1975; Stephenson & Griffin 1976b). Tholeiites have not been recorded. Most flows are chemically similar, differing only in alkalis (especially potash). Some remarkably long flows extend up to 160 km. The large volume of lava, combined with channelling into narrow dry watercourses, probably contributed to the high rates of effusion (Stephenson & Griffin 1976b). The lavas range from glassy to coarse and well crystallised (Stephenson, Griffin & Sutherland 1980).

Fourteen vents have been identified at Mount Piebald, many preserved as pyroclastic cones. K–Ar dating at Mount Piebald and the adjacent Bald Hills have ages of 1.2 Ma and 1.6 Ma respectively (Morgan 1968), confirming an early Pleistocene age for their development. Undated cones and lavas from the larger McLean Province are significantly younger than dated Pliocene flows, possibly <1 Ma (Stephenson 1989).

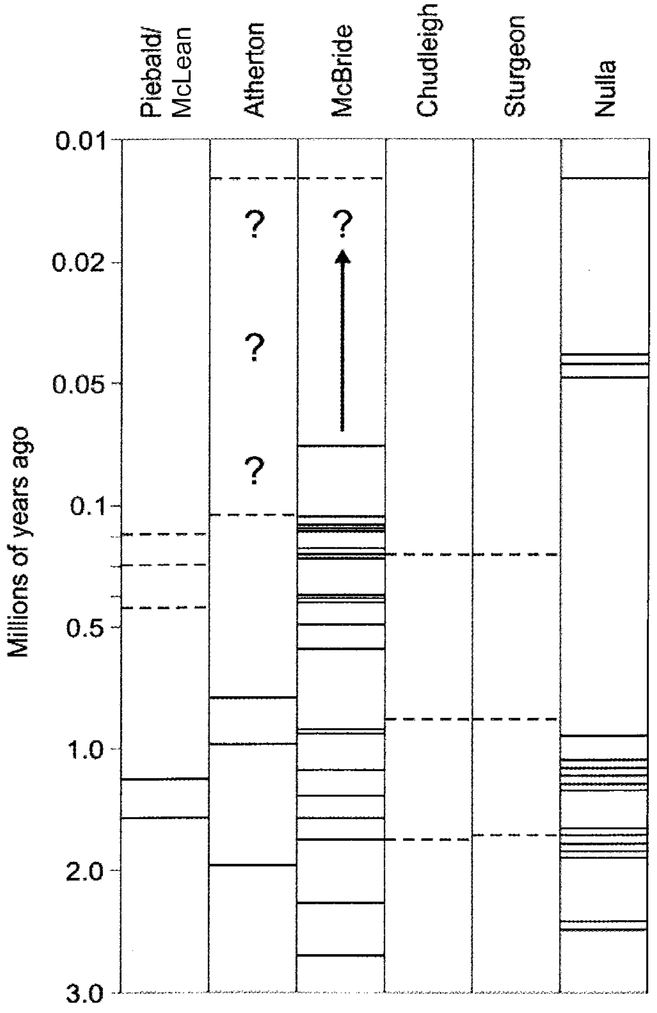


Figure 9.18 Episodes of Quaternary volcanism in northern Queensland provinces (temporal series not to scale). Broken lines indicate estimated ages based on comparative erosion (modified from Stephenson, Griffin & Sutherland 1980).

Numerous maars in the Atherton area, such as Lakes Eacham and Barrine (Figure 9.15), have yielded important palaeoclimatic information from their lake sediments (Constable & McElhinny 1985; Walker 2007). The rims are composed of a pyroclastic mixture of basalts and metamorphic rock. The age of the Atherton maars is unclear, but most likely <1 Ma (Head, Taylor & Walker 1994). Dating of the associated lake sediments provide at least minimum ages for their formation and suggest that volcanic activity may have occurred as recently as the Holocene (Whitehead et al. 2007). The Atherton Province also contains broad lava plains on the tableland, with associated cones dated at 1.7–1.8 Ma. Other younger volcanoes have been identified in the Atherton Province in the Gordonvale and Innisfail districts (990–640 ka). Although volcanism commenced in the region at ~7.1 Ma and continued into the early Holocene, Quaternary activity was greatest during 2–1 Ma (Whitehead et al. 2007).

Volcanic activity occurred sporadically in the McBride Province over much of the Quaternary. Vents in the McBride Province range from large shield volcanoes to smaller pyroclastic and scoria cones. A maximum K–Ar age of 50–70 ka for the Kinrara volcano (Griffin & McDougall 1975) makes it the youngest volcano in the region. The preserved cone is ~50 m high and the crater is up to 300 m in diameter. The volcano and associated flows contain an abundance of ropy lava surface features (Stephenson, Griffin & Sutherland 1980). The McBride Province also includes the famous Undara Lava Tubes. The Undara Crater is a steep-sided depression 340 m across and 20 m higher than the surrounding lava field (Stephenson & Griffin 1976a). Numerous extensive lava flows have been recorded as long as 90–160 km and dated to 190 ka (Stevens & Atkinson 1976; Webb, Joyce & Stevens 1993; Atkinson & Atkinson 1995). The associated lava tubes are up to 100 km long. Several of the basaltic rocks adjacent to and lining the lava tubes have unusual primary oxidised features, as well as a distinctive mineralogy comprising oxide-pigmented forsterite phenocrysts, primary hematite and yellow clinopyroxene (Atkinson, Griffin & Stephenson 1976; Griffin 1977).

The Chudleigh Province adjacent to the Great Dividing Range is characterised by lava plains that have been dissected by various features including cones (pyroclastic and composite) as well as lava shields (Stephenson 1989). Several of the younger volcanoes in the region have been dated at 250–700 ka (Stephenson & Griffin 1976a, 1976b). A flow from one of the volcanoes, Barkers Crater, can be traced down the Einasleigh River for at least 160 km and has been dated at 265 ka (Withnall & Grimes 1995).

The Sturgeon is the most southerly of the volcanic provinces in northern Queensland. Volcanic activity was greatest during the Pliocene, with the youngest flows 920 ka (Stephenson & Griffin 1976b; Coventry, Stephenson & Webb 1985). Such young flows have been used to constrain the age of development of the Hughenden Plain (Coventry, Stephenson & Webb 1985). Erosion data demonstrate that the plain has lowered a maximum of only 20 m since its development (Coventry, Stephenson & Webb 1985).

The Toomba volcano of the Nulla Province is thought to be one of the youngest volcanoes in northern Queensland, dating to around 13 ka (Stephenson, Polach & Wyatt 1978). The Toomba volcano has a vent complex of fissures and pyroclastic cones that trend northeasterly (Stephenson 1989), with the associated flow being unique in that it is the only one in the region where the original pahoehoe surface is preserved (Stephenson & Griffin 1976a). It is comprised of numerous segregations including pipes, sheets, veins and vesicles (Stephenson, Zhang & Spry 2000). The rocks at the Toomba cone are porphyritic in plagioclase, olivine and sector-augite, most likely erupted after the Toomba flow (Stephenson & Griffin 1976b). Several of the older basalt flows from the Nulla province have greatly influenced the development of the region's rivers. For example, Nulla's Kanerong, Birdbush and Hann Creek porphyritic basalt flows extended into the course of the Burdekin River, damming and diverting it on four separate occasions during the Pleistocene (Stephenson, Polach & Wyatt 1978).

9.8.2 Southeastern Queensland

Quaternary volcanism in southeastern Queensland was much less extensive than that in the north. Only the Bundaberg and adjacent Boyne provinces represent volcanic activity during the Quaternary (Figure 9.2). The rocks are mafic and typically include vesicular olivine basalt flows (Robertson, 1993a). Wellman (1978) produced a series of K–Ar ages spanning 1.1–0.9 Ma for the Hummock Basalt near Bundaberg. The basalt, up to 55 m thick and outcropping over 215 km², was extruded from a vent at Sloping Hummock, part of the Capricorn Basin. Other flows from near Bundaberg, such as the Gin Gin Basalt, are also Pleistocene (Robertson 1985).

South of Bundaberg, the Barambah Basalt near Coalstoun Lakes was dated to 600 ka (Wellman 1978; Figure 9.19). The lava flowed from three vents and spread out over a valley 3–5 km wide and almost 100 km long (Robertson, Sutherland & Hollis 1989). Various features including lava tunnels and crater lakes are associated with the Barambah Basalt (Figure 9.20). Although the Barambah Basalt represents the youngest episode of basaltic volcanic activity in southeastern Queensland, slightly younger (380–450 ka) explosive outbursts from near Boyne produced brecciated deposits (Robertson, Sutherland & Hollis 1989).

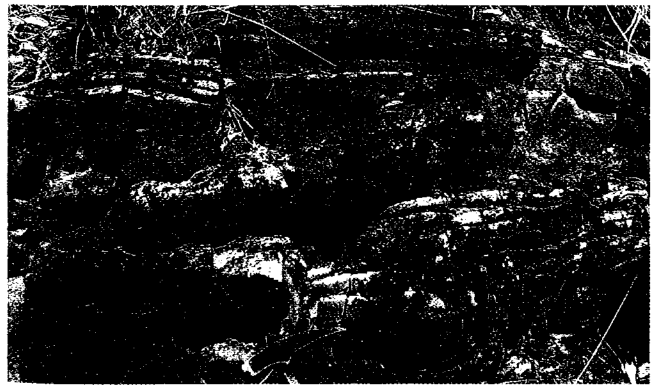


Figure 9.19 Pahoehoe lava at Barambah Creek Bridge downstream from Mount Le Brun, southeastern Queensland.

Figure 9.20 Mount Le Brun Crater with northern Coalstoun Lake (dry), southeastern Queensland.

9.9 Offshore eastern Queensland (JS Jell)

The continental margin off the east coast of Queensland hosts the largest extant tropical mixed carbonate–siliciclastic depositional system in the world (Maxwell & Swinchat 1970; Davies & Peerdemann 1998; Dunbar, Dickens & Carter 2000; Mathews, Heap & Woods 2007). Hopley, Harvey and Pye (cited in Coventry et al. 1980) comprehensively reviewed the Quaternary of offshore northern Queensland and Davies (2011) outlined recent studies of the Great Barrier Reef and adjacent seas. Most knowledge of offshore areas comes from geophysical investigations tied to a very limited number of drill holes. There is a great need for more offshore drill holes to detail the geological history.

Rifting and sea-floor spreading during the Cretaceous–Neogene produced the basal Pleistocene structure of the continental margin (Chapter 8). Orogenesis continued in the north with uplift and erosion of the New Guinea Highlands supplying huge quantities of sediment into the Gulf of Papua. Isern, McKenzie and Feary (1996) showed that early Pleistocene surface sea temperatures on the Queensland Plateau, which may be extrapolated to the central and southern Great Barrier Reef, were 20–24 °C and only suitable for restricted coral growth. By the middle Pleistocene (600–700 ka), surface sea temperatures had risen above 24 °C due to the establishment of a western Coral Sea warm water pool. This rise induced increased coral growth rates, enabling reef growth to match or exceed relative sea-level rises (Isern, McKenzie & Feary 1996). The area continued broadly subsiding with only minor epeirogenic activity in the form of reactivation along some faults. Fluctuating sea level related to repetitive glacial–interglacial cycles was the major controlling factor on the marine margin and it also influenced climate on land.

9.9.1 Gulf of Papua

By late Pliocene, the proximal foredeep of the New Guinea Orogen had been filled and terrigenous deposition (Era beds) moved south during the late Pliocene–Holocene to bury the carbonate platform in the Gulf of Papua (Pigram et al. 1989, 1990). Throughout the Quaternary, the Fly River has delivered terrigenous sediment onto the Papuan Shelf (Walsh et al. 2004) and prograding east into the deeper eastern Gulf of Papua, where a suite of volcanics and volcaniclastics (200 ka – 1.6 Ma) is interbedded (Jablonski, Pono & Larsen 2006). The last glacial sea-level fall exposed much of the Fly Platform, over which lowstand deltas prograded to the shelf break. Currently well-developed, tidal dominated deltas have produced large clinoforms into the deeper waters to the east (Harris et al. 2004b). In TAI Anchor Cay 1, on the northernmost coral reef of the Great Barrier Reef, the upper 300 m including the Holocene–Pleistocene and Pleistocene–Pliocene boundaries were not cored. The western shelf of the Gulf of Papua is devoid of extant coral reefs (Whitehouse 1973).

9.9.2 Great Barrier Reef Province

The World Heritage listed Great Barrier Reef Province consists of >3500 reefs extending 2300 km from Bramble and Anchor cays in the Gulf of Papua (9° 15' S) to the channel between Lady Elliot Island and Breaksea Spit (24° 10' S). The province covers >250 000 km², 10% of which is occupied by reefs (Figure 9.21). It extends east to the edge of the continental shelf, which is furthest east in the Swains complex (152° 50' E), 290 km from the mainland. The province narrows to a minimum of 21 km off Cape Melville north of Cooktown. More than 86% of the reefs occur in the 25–65 km wide Reef Zone (Maxwell 1969) along the margin of the shelf. The Nearshore Zone is separated from the Reef Zone by the Shelf Lagoon. Hopley, Smithers and Parnell (2007, ch. 5) provided detailed spatial data of the reefs for the Great Barrier Reef Marine Park portion of the province based on available bathymetric surveys and satellite imagery.

Pleistocene antecedent foundation: At the earliest Pleistocene (2.6 Ma), the continental shelf was far enough north and in appropriate neritic water depth for reef development, but reefs did not form until the middle Pleistocene (500–700 ka). In their model of the Cenozoic evolution of the central Great Barrier Reef, Symonds, Davies and Parisi (1983) considered the pre-reef Pleistocene as the lower part of their seismic depositional Sequence A, which has a relatively uniform thickness of ~180 m but thins to <90 m beneath the inner shelf, where it onlaps basement and thickens to 600 m beneath the upper slope. Beneath the outer shelf and slope, the upper part of the underlying Sequence B may be early Pleistocene.

Initiation of the Great Barrier Reef: Speculated age of the Great Barrier Reef varies greatly, even to the Miocene. Marshall (1983) and Symonds, Davies and Parisi (1983) suggested that the reef may be no older than Pleistocene and seismic data and sediment cores from the continental slope (ODP Leg 133) suggested initiation ~500 ka (Davies 1991; Davies & McKenzie 1993; Davies & Peerdeman 1998). In 1995, deep cores recovered from Ribbon Reef 5 and Boulder Reef drill holes constrained initiation of the reef to ~452–365 ka (International Consortium 2001; Alexander, Wilson & Cooper 2001; Webster & Davies 2003; Braithwaite et al. 2004; Braithwaite & Montaggioni 2009). Those dates were questioned by Dubois et al. (2008), who suggested initiation for the southern central Great Barrier Reef at 560–670 ka on the basis of sediment cores from the Marion Plateau obtained on ODP Leg 194. Debris in the Ribbon Reef core and in ODP Leg 133 holes suggests carbonate deposition on the Great Barrier Reef Province ~770 ka and reef building starting ~400–500 ka, which coincides with the middle Pleistocene Climate Transition (Hays, Imbrie & Shackleton 1976; Clark et al. 2006) or MIS 11 (Webster & Davies 2003). MIS 11 was one of the warmest Quaternary interglacials when 100 000-year climate cyclicity (i.e. glacial to interglacial) was fully established along with higher sea-level amplitudes.

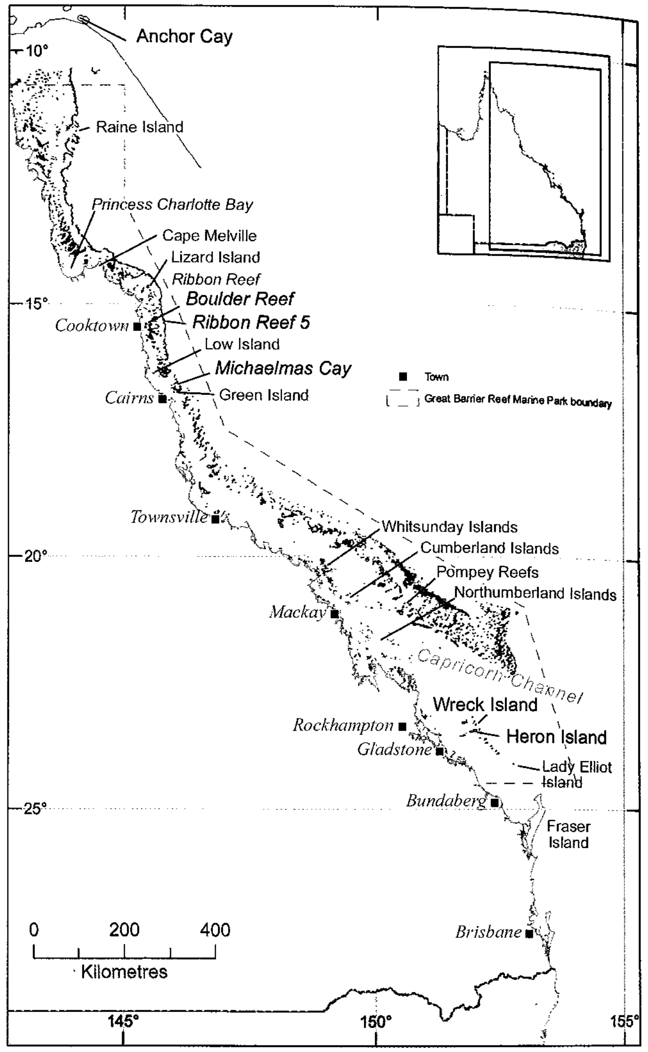


Figure 9.21 The Great Barrier Reef Province showing major locations and the boundary of the Great Barrier Reef Marine Park.

Reefs were initiated on siliciclastic fluviatile and deltaic low sea-level foundations, but whether they occurred preferentially on raised geomorphic features is uncertain (Symonds, Davies & Parisi 1983). Nowhere have they grown from karst antecedent surfaces. In the six reefs that have been drilled to their foundations (Table 9.1), the reef development is relatively thin at <200 m. Apart from an Eocene build-up in the Grafton Passage (Symonds, Davies & Parisi 1983), buried reefs are rare in inter-reefal areas of the Great Barrier Reef Province, indicating that reefs have grown on the same sites since their initiation.

Table 9.1 Deep boreholes through reefs of the Great Barrier Reef Province

Borehole or reef	Latitude (S)	Longitude (E)	Total depth (m)	Thickness (m) of Quaternary reefal sections	Reference
Anchor Cay	9° 22'	144° 07'	3623	Nil	Oppel (1970)
Ribbon 5	15° 22'	145° 47'	210	0–96, 117–130, 130–138	Webster and Davies (2003)
Boulder	15° 24'	145° 26'	86	0–34,coral material 47–56, 60–64	Webster and Davies (2003)
Michaelmas Cay	16° 35'	146° 02'	183	115	Richards and Hill (1942)
Wreck Island	23° 20'	151° 58'	575	121	Derrington (1960)
Heron Island	23° 27'	151° 55'	223	154	Richards and Hill (1942)

Ribbon Reef 5 provides the most information (Figure 9.22) on reef development (International Consortium 2001; Webster & Davies 2003; Braithwaite et al. 2004; Braga & Aguirre 2004). The oldest unit (210–158 m) is non-reefal grainstones and packstones interpreted as debris flows and/or turbidites. The overlying unit (158–96 m) is rhodolith-dominated with two thin coral framework reefal intervals. The upper 96 m is almost entirely reefal. Of six units in this upper interval, the upper four are separated by disconformities associated with glacial lowstands (Webster & Davies 2003). Dissolution surfaces and palaeosols were recognised. Four reef units in the upper 34 m and two coral rubble horizons at 55 m and 63 m were recognised in the Boulder Reef core (Webster & Davies 2003). Flood (1993) recognised six reefal units separated by solution unconformities in the Heron Reef research core. Pleistocene development of the Great Barrier Reef Province reefs was aggradational, consisting of stacked highstand reef packages separated by unconformities associated with intervening lowstands when the reefs may have been emergent. Shallow seismic refraction (Harvey 1977a, 1977b; Davies et al. 1981; Harvey, Davies & Marshall 1979; Searle & Harvey 1982) identified the Pleistocene/Holocene boundary in a number of reefs but lower reflectors are difficult to correlate.

With initiaton of reef growth, sedimentation changed from dominantly terrigenous to mixed siliciclastic/carbonate. Seismic profiles and shallow coring indicate continued shelf sedimentation under high and low sea-level conditions (Searle 1983). The inner shelf was characterised by prodeltaic sedimentation associated with coastal wave-dominated deltas. Carbonate, shed from the reefs during highstand growth and lowstand exposure, was incorporated into outer shelf sediments. Quaternary successions have extensive and repeated channelling, multigeneration infill and levee-bank deposits indicative of fluvial systems. The development of reefs on the edge of the shelf led to channelling of major sediment loads off the shelf through inter-reef passes and changed the style of sedimentation on the slope and in adjacent basins.

Sediments of the continental shelf: Holocene inter-reefal sea-floor sediments (Maxwell & Maiklem 1964; Maxwell 1968, 1973b; Maxwell & Swinchatt 1970; Belperio 1983; Marshall 1977; Flood, Orme & Scoffin 1978; Orme & Flood 1980; Flood & Orme 1988; Mathews, Heap & Woods 2007) are composed of terrigenous material with fluvial association, aeolian inputs from onshore dunes and remobilisation of Pleistocene sediments, particularly ancient dune systems and carbonate grains produced *in situ* or derived from adjacent intertidal, subtidal and reefal deposits. The sediments form longitudinal zones related to shelf depth: an inner shelf terrigenous zone to 22 m, a sediment-starved middle shelf from 22 to 40 m and a carbonate outer shelf below 40 m (Belperio 1983). In the southern Great Barrier Reef Province, the 300 m deep Capricorn Channel that deepens southwards between the Swains Reefs and the Capricorn–Bunker groups received a considerable thickness of hemipelagic sediment mixed with fine clays delivered from the inner shelf by large floods. Ooids in the succession (Marshall 1977) were used by Yokoyama et al. (2006) to establish a palaeoshoreline 100 m lower than today at 16 ka, immediately prior to a period of accelerated melting of the global ice sheet.

Sedimentological studies of the Great Barrier Reef Province and adjacent basins have shown that the classic reciprocal model for passive margin siliciclastic/carbonate shelf sedimentation is not appropriate (Dunbar, Dickens & Carter 2000; Page, Dickens & Dunbar 2003; Francis et al. 2007; Bostock et al. 2009). The trangressive shedding model (Page, Dickens & Dunbar 2003) proposes that terrigenous sediment supplied in the last glacial lowstand (31–14.7 ka) is sequestered on the middle shelf landward of the main reef tract, possibly as a result of reduced fluvial discharge and stream power, and little was transported to the reef slope (Wolfe et al. 1998). During the postglacial transgression (14.7–6.5 ka), when the shelf flooded, the siliciclastic sediment stored on the mid-shelf was transported by waves, tidal currents or cyclones off the shelf to the continental slope (Dunbar, Dickens & Carter 2000). In the late stages of the transgression, intense reworking of the stored sediment may have combined with a higher riverine input, resulting in a large off-shelf siliciclastic flux. During the highstand (6.5–0 ka), siliciclastic sediment was confined to the inner shelf by the southeast winds and northward longshore currents, and with renewed biogenetic carbonate production mainly on the outer shelf, carbonate sedimentation prevails in this zone (Figure 9.23).

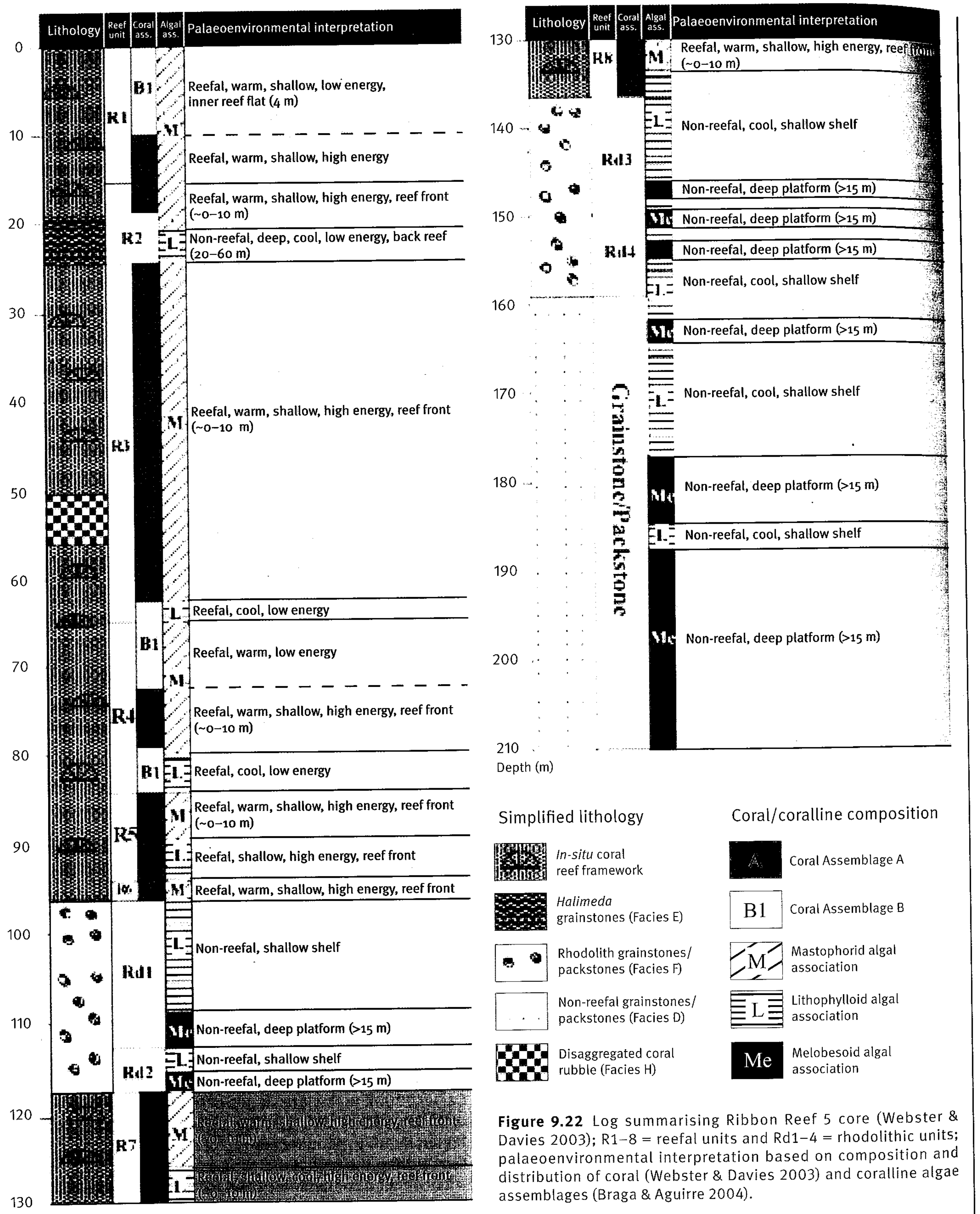


Figure 9.22 Log summarising Ribbon Reef 5 core (Webster & Davies 2003); R1–8 = reefal units and Rd1–4 = rhodolithic units; palaeoenvironmental interpretation based on composition and distribution of coral (Webster & Davies 2003) and coralline algae assemblages (Braga & Aguirre 2004).

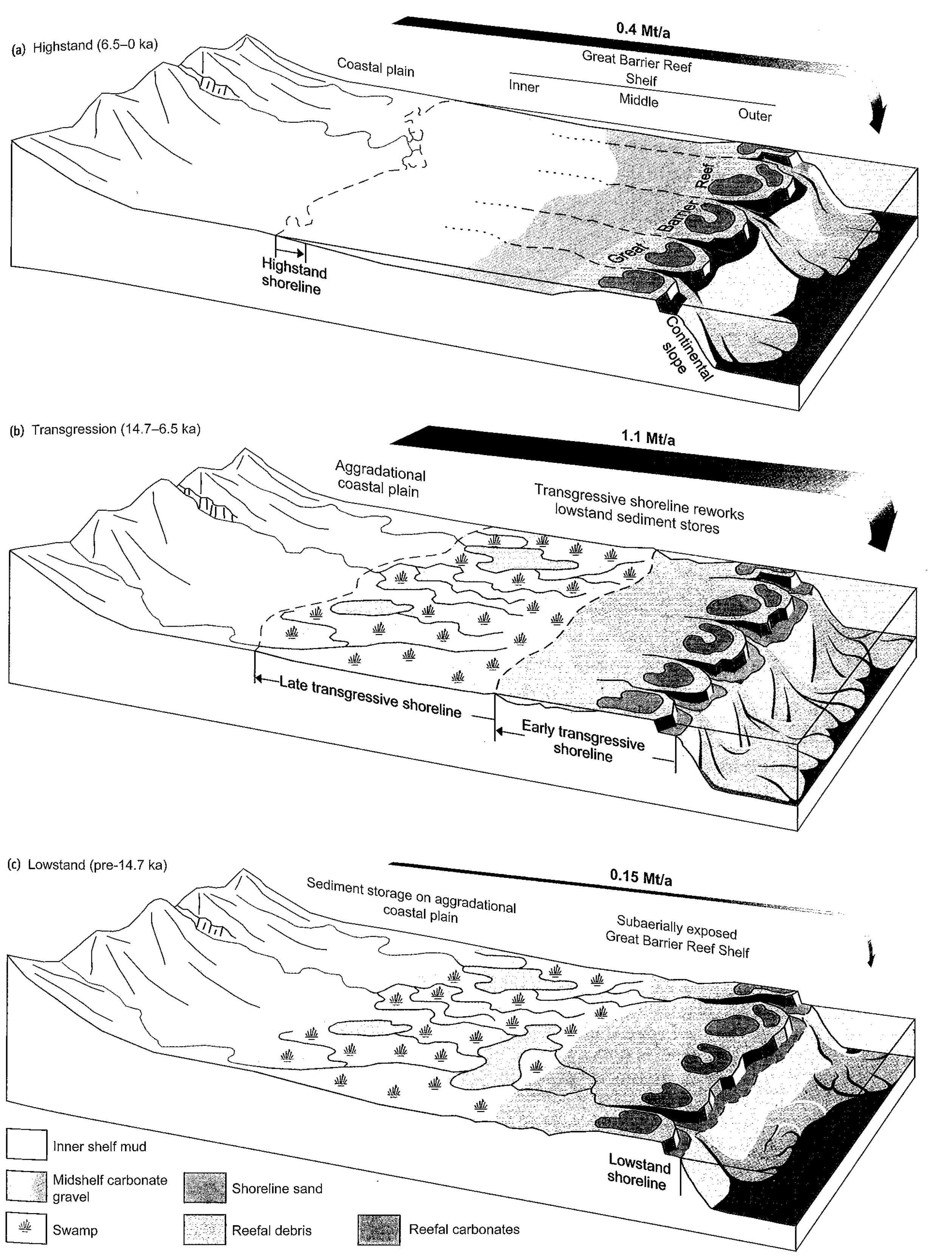


Figure 9.23 Model of shelf depositional environments and morphology during the past 15 ka (after Mathews, Heap & Woods 2007). Diagram shows the changing shelf environments and movement/behaviour of siliciclastic sediments during **(a)** highstands, **(b)** transgression and **(c)** lowstands, and the corresponding amount of siliciclastic sediments transported off the shelf during each stage.

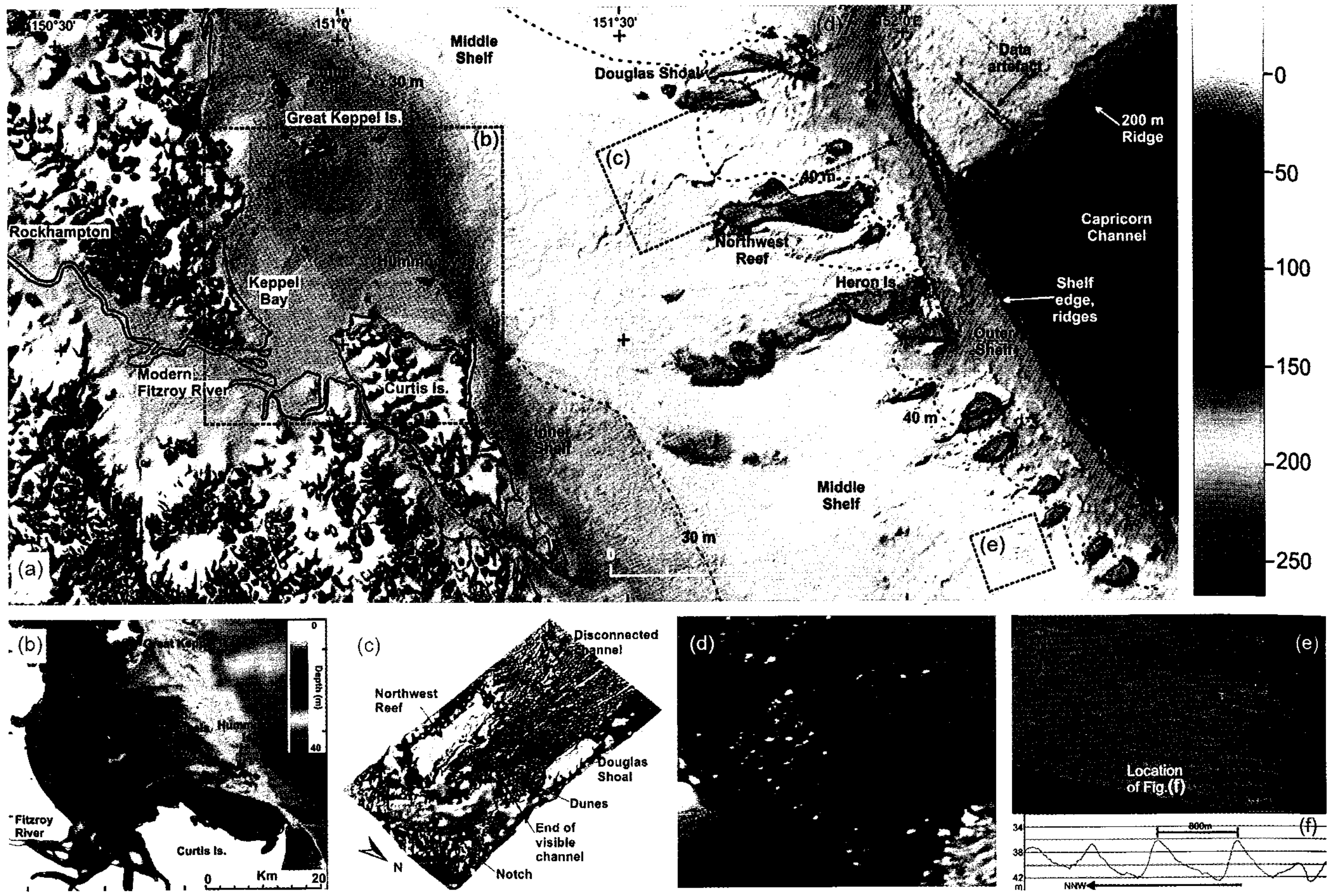


Figure 9.24 Bathymetry of the Capricorn Shelf showing trace of the palaeo–Fitzroy River crossing the continental shelf and entering the palaeo-ocean between Northwest Reef and the Douglas Shoals (after Ryan et al. 2007).

Holocene reef development: During the last glacial, the continental shelf was a coastal plain crossed by river systems with their incised channels recognisable using bathymetric and geophysical data (Figure 9.24) and with the Pleistocene reefs exposed as islands undergoing karstification. Drowned reefs are known at depths of 100 m, 90 m, 60–50 m and 35–40 m (Veeh & Veevers 1970; Marshall 1977; Carter & Johnson 1986; Harris & Davies 1989; Hopley 2006; Hopley, Smithers & Parnell 2007; Beaman, Webster & Wust 2007). Integrated ODP Expedition 325 drilled a number of these structures, showing they are postglacial reef developments (Abbey, Webster & Beaman 2011; Yokoyama et al. 2011; Webster et al. 2011). These submerged reefs occur on terraces cut into the shelf edge at depths of 40–130 m. They occur as terrace systems of pinnacle and barrier reefs and palaeolagoons (Abbey, Webster & Beaman 2011) and are dominantly composed of coragal-microbialite and coralgal boundstones up to 30 m thick with pockets of *Halimeda*, large benthic foraminifers, molluscs, red algae, bryozoans and echinoderms, similar to the shelf reefs today. Preliminary dating suggests that in most of the deeper holes, reef material provides dates consistent with the Last Glacial Maximum and ranges 25–20 ka but in shallower holes this is 13–14 ka, showing the reef development tracked sea-level rise (Yokoyama et al. 2011). Some older ages suggest that reefs predating the Last Glacial Maximum may be available for interpreting Pleistocene development.

By ~12 ka, the shelf began to flood, with Pleistocene reefs becoming limestone islands. The continental shelf, which is a shallow (0 –90 m, average 40 m), partially rimmed, high-energy platform, sloping gently (<1–4 m/km) away from the coast and deepening southward, would have been completely flooded by 7 ka (Hopley, Smithers & Parnell 2007). Start-up growth began on eroded Pleistocene reefal limestone foundations during 8.6–6.6 ka on the outer shelf, 9.1–6.7 ka on the middle shelf and 7.8–6.3 ka on the inner shelf (Davies 2011).

Once established, reefs reacted to water depth, sea-level rise and rate of reef growth in four patterns of growth—start-up, catch-up, keep-up and give-up (Davies, Marshall & Hopley 1985; Davies & Montaggioni 1985; Neumann & Macintyre 1985). More than one growth pattern can occur on the same reef as local environmental factors affect the growth response. Hopley, Smithers and Parnall (2007) analysed available growth rate data for the Great Barrier Reef Province and showed framework accretion rates up to 16 m/1000 years with a modal value at 7–8 m/1000 years and detrital accretion rates up to 10 m/1000 years. Growth rate is depth-dependent and thus annual growth is generally slow (3–4 mm) to start, becomes more rapid (>8 mm) as the reef trails sea level and slows in high-energy shallow water. Sea level first reached its present level 6.5 ka (possibly a little higher ~5 ka) and a number of reefs reached that level at that time, whereas many are still growing to reach it. The growth rate v. water depth curve varies between inner, middle and outer shelf reefs, with the greatest variation in the upper 6 m—the outer declining to 7 m/1000 years and inner shelf 4–5 m/1000 years. Growth rates do not vary with latitude.

Reef types and distribution: There is great variation in most parameters applying to reefs of the Great Barrier Reef Province (Fairbridge 1950; Maxwell 1968; Hopley 1982; Hopley, Smithers & Parnell 2007) but a classification into four main types is possible (Figure 9.25):

- Submerged reefs formed on the continental slope during periods of lower sea level but were unable to keep up with the rapidly rising sea level and drowned. They occur along the length of the Great Barrier Reef from east of Cooktown (15° 22') to the Pompey Complex (21°; Hopley, Smithers & Parnell 2007). Terraces may be palaeofringing reefs, pinnacles may be round or oval patches of reef growth or palaeoplatform reefs and reefs may be palaeoribbon reefs (Abbey, Webster & Beaman 2011). Based on their height and complexity, under-developed juvenile reefs and well-developed mature reefs are recognised and the distribution of these suggests that terraces were not ideal environments for extensive vertical growth (Abbey, Webster & Beaman 2011).
- Ribbon or linear reefs grow along the shelf edge and grow landward if water depth increases too rapidly on the seaward side. They occur over 700 km between 11° S and 17° S as narrow discrete reefs (10–28 km long) separated by channels (1 km wide). Their zonation tends to be parallel to the reef front and typically with no repetition of zones on the lee margin. With the great tidal flows through the channels, the reef crest commonly curves around the end of the reef and in some cases behind the reef so defining the wall, cuspate, open-ring and closed-ring reefs of Maxwell (1968).
- Platform reefs rise up from the shelf and may potentially grow in all lateral directions. The influence of size and shape of the antecedent topography is well illustrated by the boomerang-shaped reefs off Princess Charlotte Bay based on fossil sand dunes. Many of the reefs of the Torres Strait are elongated in the direction of the strong tidal flows through the straits (Woodroffe et al. 2000). Hopley (1982) classified platform reefs as:
  - juvenile—mainly start-up reefs with enhancement of the antecedent platform
  - mature—reefs with strong growth on the windward side and horizontal extension to leeward around the perimeter of the foundation to eventually enclose a lagoon
  - senile—planar reefs with lagoons filled in, produced by growth and sediment infilling the lagoons and masking the original relief.
- Fringing reefs grow seaward from land or continental islands with poorly developed zones parallel to the shoreline (Fairbridge 1950; Kennedy & Woodroffe 2002; Hopley 1982). They are most common between 20° S and 22° S and may be classified as headland, bayhead, narrow beach-based and nearshore shoals (Hopley, Smithers & Parnell 2007). They occur in turbid water and live corals are restricted to a narrow band near the seaward edge. Typically, the area behind the living coral zone is higher as many of the reefs formed when sea level was ~1.5 m higher than today.

Other reef types are various modifications of the above basic types:

- Low wooded island reefs, such as Low Isles, occur close to shore on the narrow northern section and are characterised by shingle ramparts, mangrove swamps and sand cays.

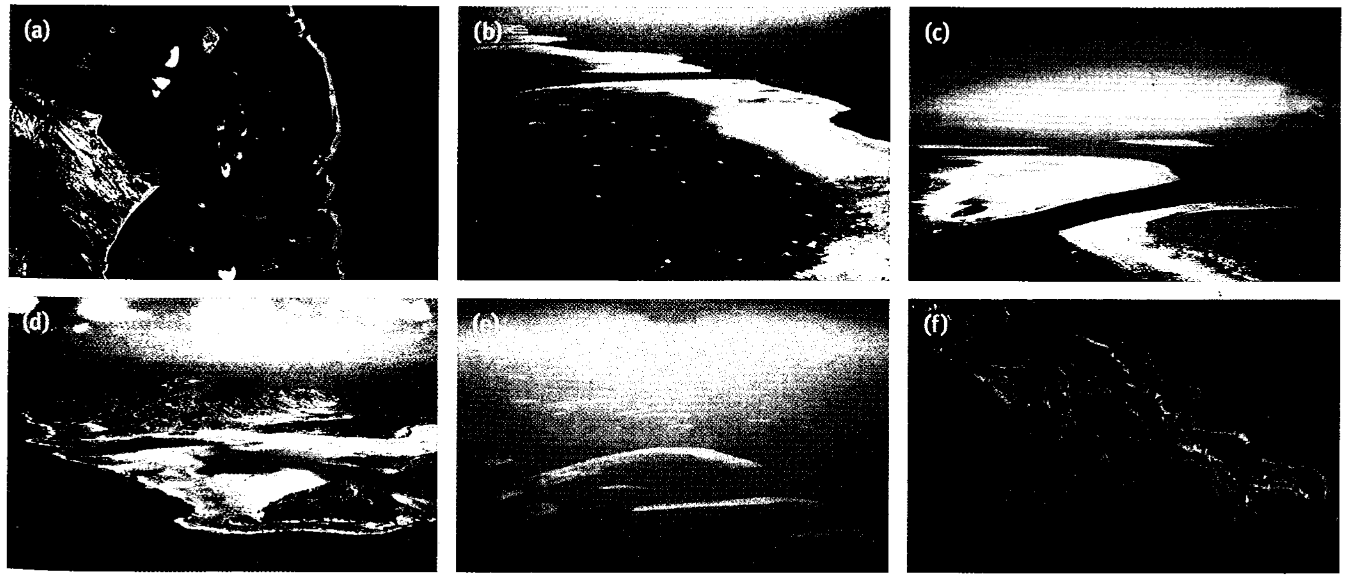


Figure 9.25 Reef types within the Great Barrier Reef Province. **(a)** Northern Great Barrier Reef showing the Ribbon Reefs, and platform and fringing reefs. **(b)** The Ribbon Reefs paralleling the reef edge off Cooktown. **(c)** Platform reefs of the Capricorn Group: Wistari Reef in the foreground, Heron with Sykes and One Tree reefs in the background. **(d)** Fringing reefs around Lizard Island. **(e)** Reefs of Princess Charlotte Bay. **(f)** Deltaic reefs of the Swains Reefs.

- Detached reefs, such as Raine Island Reef, rise steeply from deep water east of the shelf edge off Cape York. They appear to have grown on small continental blocks separated from the main shelf, possibly due to faulting.
- Deltaic reefs of the Pompey Complex parallel the shelf edge in the central region and are set back ~20 km from it (Figure 9.25(f)). They experience the greatest tidal currents of the province and become large complex platform reefs (giganto reefs of Davies 2011), with reefal deltas formed about the bifurcating channels between reefs. Commonly a triangular plug reef, with its apex pointing seaward, grows in the bifucation of these channels (Maxwell 1970). Deltaic reefs also occur in Torres Strait.

Reef morphology: Reef development is controlled by interplay of:

- construction of a skeletonised framework
- destruction of the framework to produce sediment
- transportation of the sediment
- deposition of sediment, including settling, baffling, trapping and protecting
- unification of the framework and sediment, including binding, encrusting and cementing.

The relative importance of these processes, and the organisms involved in each, vary with the energy of the local environment and water depth. The predictable distribution of these factors produces a reef morphology recognisable in most reefs of the Great Barrier Reef Province.

Maxwell (1968) and Hopley (1982) described reef morphologies of the province with the attendant ecological and sedimentological zones that parallel the line of intersection of the reef and the water surface at low tide, differentiating the reef into the reef slope and reef top. Five main zones may be recognised (Figure 9.26):

- **reef rim**—narrow, exposed to the waves and swells, forms the highest part of the reef top and gently slopes seaward, made mainly of coralline algae encrusted corals and strewn with coral rubble
- **reef flat**—shallow, horizontal, at low water level (at the ponded low water level on enclosed reefs), consists of outer coral and inner sandy subzones
- **lagoon or back reef**—calm, up to 10 m deep, has some patch reefs
- **reef front**—seaward of the surf zone, upper part of the reef slope bears good coral cover, commonly featuring spur and groove structure and terraces
- **lower reef slope**—less steeply inclined, has less coral cover and less coral and other detritus.

Maiklem (1968) and Jell and Flood (1978) described the zonation of reefs of the Capricorn Group, and Jell and Webb (2012) further subdivided some of the zones (Figure 9.26).

Holocene reef sediments: Reef sediments include those of the reef slope (about which very little is known) and the reef top (Flood & Scoffin 1978; Tudhope & Scoffin 1984; Scoffin & Tudhope 1985; Maxwell, Day & Fleming 1961; Maxwell, Jell & McKellar 1964; Orme, Flood & Ewart 1974; Flood 1974, 1977; Flood & Jell 1977; Flood & Orme 1977; Jell & Flood 1978; Davies & Kinsey 1977). Orme and Flood (1980) reviewed sedimentation in the Great Barrier Reef Province.

Corals, coralline algae, foraminifers, molluscs and *Halimeda* generally account for ~90% of reef top sediment. Distribution of skeletal detritus is controlled by decreasing wave action across the reef top producing a size gradient from gravel and coarse sand to very fine sand, from the windward reef rim to the lagoon or back reef. This may be modified by tidal currents when the reef top waters are isolated from the open ocean. The hydraulic properties of the grains vary according to the structure of the organisms from which they are derived and consequently the compositional patterns reflect the textural patterns. Corals (aragonitic) are rapidly broken down into distinct size modes (shingle sticks, very coarse sand, fine sand to very fine sand) with coral dominating the gravel at the rim and the fine sands of the calm waters to leeward. Coralline algae (calcitic) are relatively resistant to mechanical breakdown and contribute to the very coarse sands and gravels on the reef rim. Sediment thickness varies from commonly <10 cm on the reef flat to 4 m below the shallow lagoon and >5 m in the deep lagoon (Smith et al. 1998).

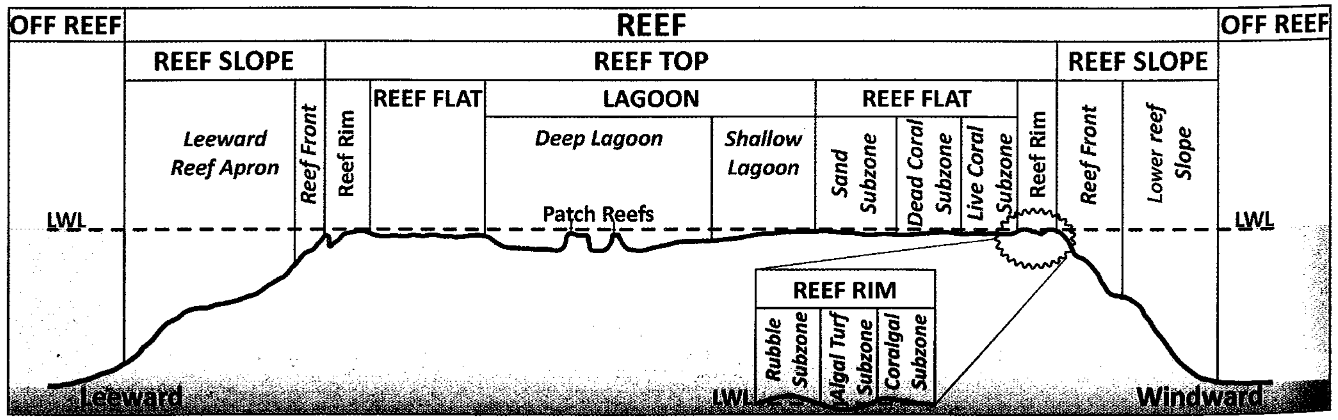


Figure 9.26 Generalised cross-section of a lagoonal platform reef such as Heron Reef from windward to leeward showing the recognisable zones and subzones.

Holocene reef structure: Maxwell (1968) interpreted the internal reef structure by extending the surface sedimentary facies into the subsurface. Shallow seismic refraction, shallow coring and extensive dating of cores have allowed recognition of facies and structure of reefs, and the development of growth models (Hopley, Smithers & Parnell 2007; Montaggioni & Braithwaite 2009).

Seven broad facies in Holocene reefs of the Great Barrier Reef Province are:

- coralline-algal facies characterised by boundstone of laminated crusts on coral and rubble with substantial amounts of microbialite; best developed on the windward margin but encrust most dead or dying corals across the reef
- branching coral facies of branching *Acropora*, *Pocillipora* and *Porites*, generally heavily encrusted by coralline algae with little-bound coarse debris but containing open cavities between horizontal ledges of wedged debris and/or encrusting corals and packed with microbialite
- platy coral facies of platy coral framestone with *Acropora hyacinthus* Group and encrusting coralline algae and foraminifers, coarse debris and microbialite within a well-connected cavity system
- head coral facies of massive coralline framestone/rudstone with massive corals in growth position or overturned encrusted with coarse debris and an open cavity system
- rubble facies, typically rudstone, of broken massive, platy and branching coral, in some cases thinly encrusted, heavily bioeroded, in some cases cemented, and with an extensive but irregular cavity system containing microbialite; commonly on the inner reef margin or reef slope
- sand facies of medium to coarse grainstone–packstone but with some floatstone, variously sorted, commonly heavily bioturbated; extensive sheets across the reef flat and infilling the lagoon
- muddy sand facies of fine sand to silt packstone–wackestone heavily bioturbated and variously interrupted laterally by framestones (patch reefs); not common, restricted to closed lagoon.

Less common facies including noncarbonate ones occur in fringing and inner shelf reefs.

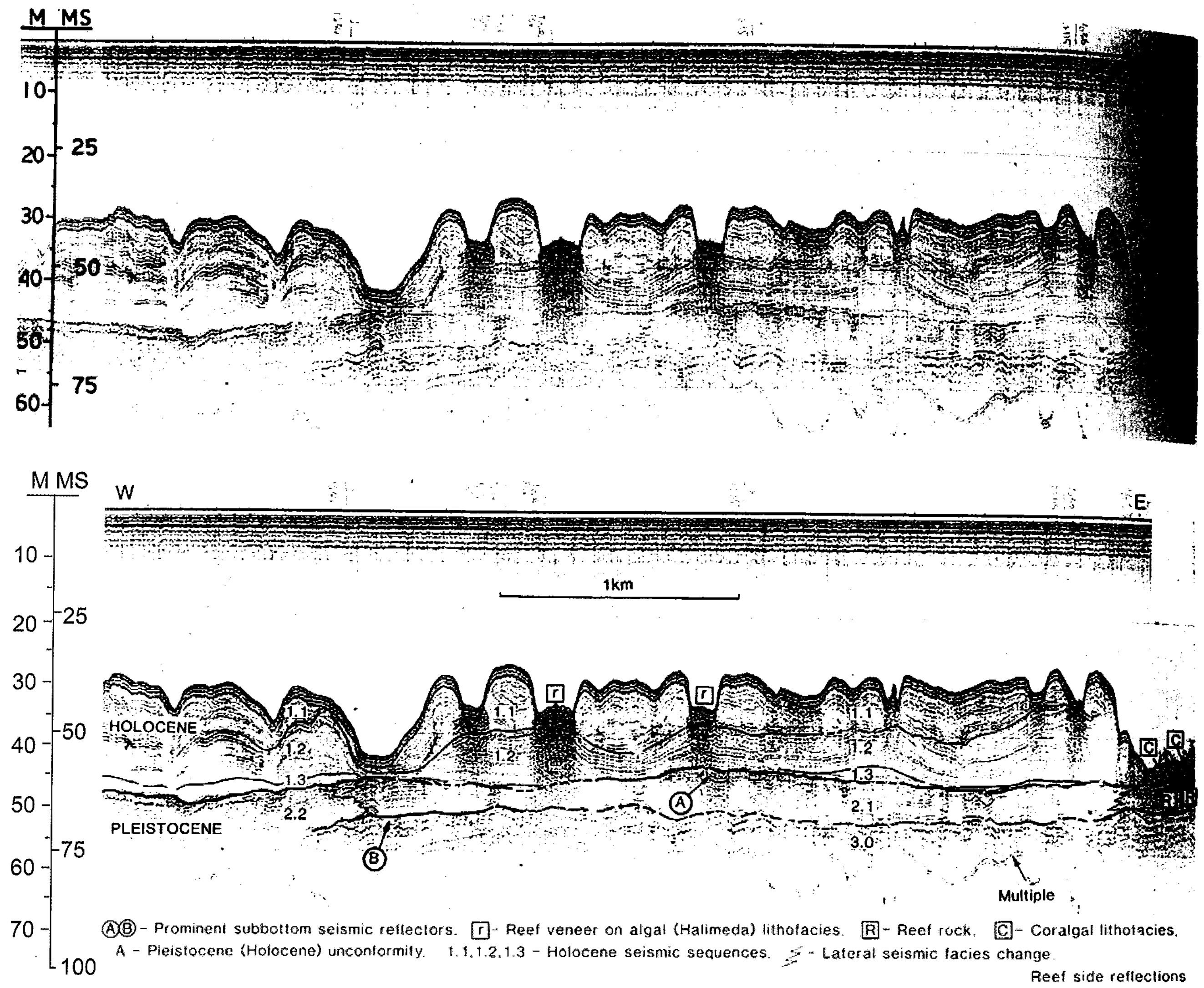
The generalised reef growth model of Marshall and Davies (1982) suggests that following Pleistocene substrate inundation ~8 ka, high-energy coral head facies and lower energy branching coral facies aggraded on the windward and leeward sides of the reef, respectively, with both initially lagging behind sea-level rise and the windward facies attaining modern sea level sooner than the leeward after sea level began to stabilise. Once sea level was attained on the windward margin, progradation was directed primarily towards the back reef. Davies (2011) provided growth models for three reef types, but others will be necessary for different settings (tectonics, tidal range, wave energy, storm activity, etc.) in different regions (Jell & Webb 2012).

***Halimeda* banks (bioherms):** Along the length of the Great Barrier Reef Province, a narrow strip on the outer shelf in waters 50–100 m deep hosts large meadows of the green calcareous alga *Halimeda* (Figure 9.27). The resulting sediment is rich in *Halimeda* grains (Orme 1985; Drew & Abel 1985; Davies & Marshall 1985; Scoffin & Tudhope 1985; Searle & Flood 1988; Hopley, Smithers & Parnell 2007). In places, particularly in the lee of northern ribbon reefs, banks almost exclusively of *Halimeda* debris occur in waters 20–50 m deep and are up to 16 m thick, 100 m wide and 150 m long (Orme, Flood & Sargent 1978). The banks appear to lie directly on the Pleistocene that in one core was overlain by a thin peat layer, which is in turn overlain by carbonate sands and *Halimeda* gravel. The peat has an age of 10 070 ± 180 years (Orme 1985), suggesting that the *Halimeda* banks began forming before the ribbon reefs (Marshall & Davies 1988). Orme and Salama (1988) suggested that late Pleistocene succession consisted of cemented *Halimeda* and the meadows spread over the transgressed Pleistocene surface, developing into banks where vertical growth was enhanced by nutrient-rich waters spilled onto the shelf.

Holocene deepwater algal build-ups: Davies et al. (2004) recognised deepwater algal build-ups east of the Capricorn–Bunker groups (23° 15' S–23° 45' S) in 80–120 m of water. They occur in a 4 km wide strip of broken ground characterised by terraces, valleys, scarps and mounds. The mounds are up to 4 m high and occur on top of a prominent sub-seabed reflector over a distance of 40 km. Dredge samples showed that the mounds consist of thick coralline algal boundstones composed of thin encrusting coralline algae growing one over another, which then divide, creating a leafy open structure. Their tops are covered by live forms, indicating that the mounds are still growing. Radiocarbon dates are 9060 years for a coral from the base and 6980 years for the boundstone, indicating that they started growing in water at least 30 m shallower than at present.

9.9.3 Southern Queensland continental margin

The southeastern Queensland coast consists of a series of long sandy zetaform embayments separated by headlands and is characterised by large Quaternary sand islands. The Quaternary coastal deposits include sandy beaches, beach ridges, foredunes and transgressive dunes seaward of estuaries and bays with tidal flats, deltas and prodelta fans (Figure 9.28). Coral reefs grew in the past but today only a veneer of living coral lives on exposed basement rocks in a few places. On the narrowed continental shelf off southeastern Queensland (south of Breaksea Spit at 24° S), marine sedimentation continued alternating with periods of erosion, non-deposition or fluviatile sedimentation at times of lowered sea level. Like the Great Barrier Reef Province, this part of the shelf is a mixed carbonate–siliciclastic depositional system, but the carbonate is from subtropical red algae and restricted coral faunas (Marshall et al. 1998).



ⒶⒷ - Prominent subbottom seismic reflectors. [r] - Reef veneer on algal (Halimeda) lithofacies. [R] - Reef rock. [C] - Coralgal lithofacies. A - Pleistocene (Holocene) unconformity. 1.1,1.2,1.3 - Holocene seismic sequences. - Lateral seismic facies change.

Figure 9.27 Interpretation of *Halimeda* banks morphology just east of Lizard Island based on high-resolution seismic profiles and bathymetry (redrawn from Orme & Salama 1988).

Stephens (1982) reviewed coastal geological studies, reinterpreted all depositional units and their ages and identified major factors influencing coastal sedimentation. Seismic surveys were carried out by Searle (1982) and Nunn (1982).

Coastal zone: The coastal zone is a wave-dominated coastline in a microtidal to mesotidal system (Hayes 1975). The most prominent features are the sweeping strand plains and the barrier sand islands (Fraser, Moreton, Bribie, North Stradbroke and South Stradbroke). They are the product of one of the most extensive sediment transport systems in the world, transporting sand 1500 km from the Sydney Basin north to these sand islands and along the way collecting large volumes of the weathering products from the New England Batholith granitoids brought to the coast by the northern rivers of New South Wales. These islands protect Moreton Bay (Figure 9.29) and Hervey Bay from the strong southeast swell. The Gold Coast, south of the Stradbroke islands, and the Sunshine Coast, between Moreton and Fraser islands, consist of long embayments infilled with unconsolidated Quaternary sands between basement headlands.

Strand plain deposits: Beaches and spits with associated barrier ridges, foredunes and back-barrier swamps occur along the sweeping coast from Point Danger to Cape Moreton and from Bribie Island to Breaksea Spit, north of Fraser Island. The beaches face the open ocean at the top of a nearshore zone, characterised by high-energy wave activity and continuous sediment transport bringing instability to the beaches. Barrier ridges and foredune development varies greatly, being widest at the southern ends of zetaform embayments (Stephens 1982). South Stradbroke Island consists solely of beach ridges and foredunes behind the sweeping sandy beach. Bribie Island likewise is low lying, consisting of beach ridges, minor dunes and tidal delta deposits. The best developed back-barrier swamp is Eighteen Mile Swamp on North Stradbroke Island (Figure 9.30), which is 600 years old (Flood, Grant & Frankel 1986). Back-barrier swamps and Pleistocene beach ridges are known from the adjacent coastal plain, such as at Coolangatta, Palm Beach and Labrador (Stephens 1982).

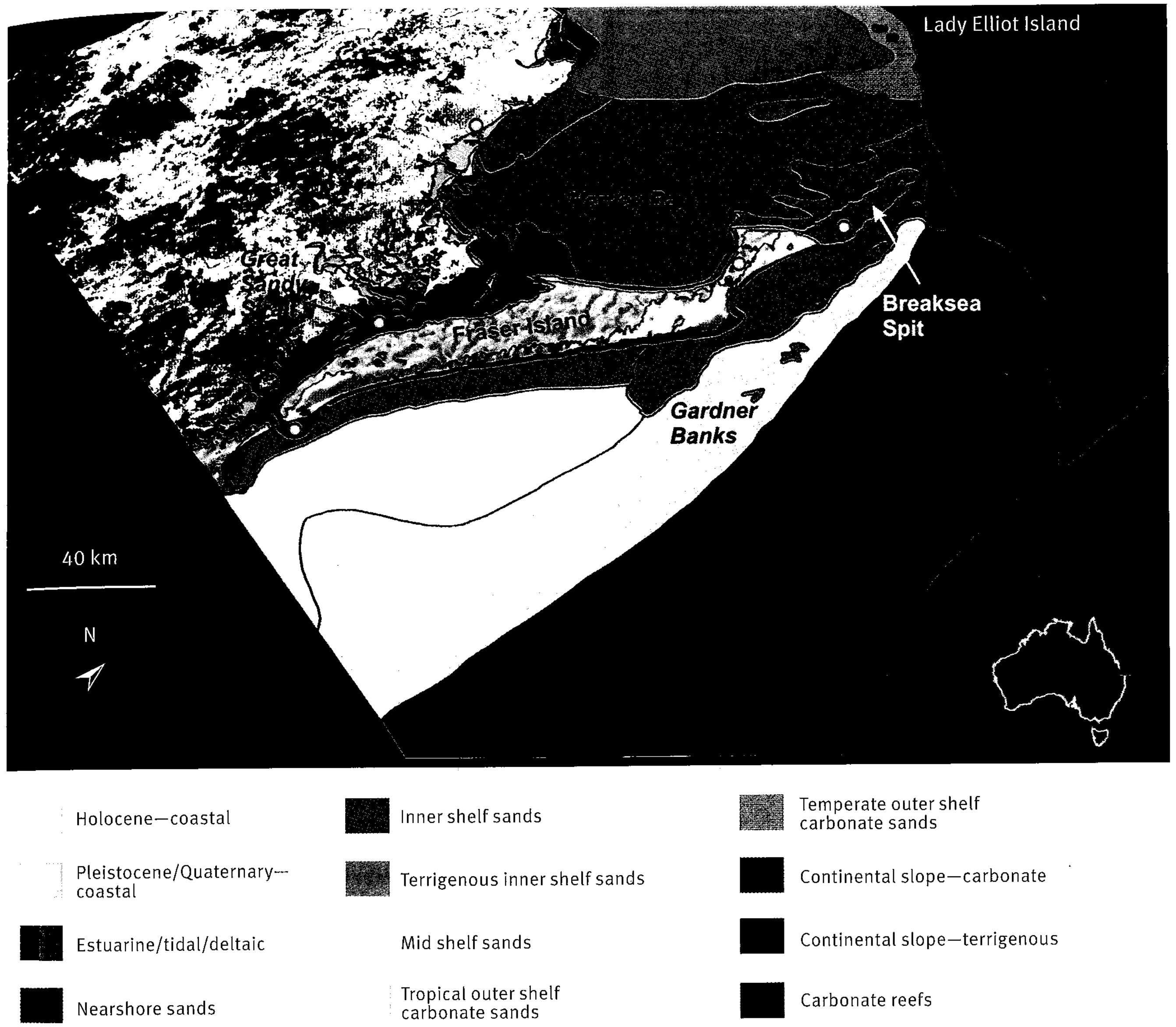


Figure 9.28 Map of the major Quaternary environments in the Hervey Bay – Fraser Island area (from Boyd et al. 2004).

Barrier islands: South Stradbroke, North Stradbroke, Moreton, Bribie and Fraser islands are the biggest sand islands in the world. The largest is Fraser Island, which has accumulated ~200 km^3 of mainly quartz sand on an island 124 km long, up to 16 km wide and with dunes up to 244 m high (Boyd et al. 2008). The islands developed in the same positions as they exist today, anchored by headlands and underlying topography of basement rocks, and sculptured predominantly by wave and wind action during several sea-level oscillations. In Sandy Cape 1–3R drill hole, 103 m of loose well-sorted medium quartz sand similar to the overlying dune sands developed on a Neogene – early Pleistocene succession (Grimes 1982). On North Stradbroke Island, sand extends 30–50 m below sea level, overlying ridges of Paleozoic and Mesozoic basement (Laycock 1975a, 1975b; Grimes 1982), although it is not certainly identified as aeolian. Nine episodes of parabolic dune formation with intervening episodes of strandplain development have been mapped (Ward & Grimes 1987; Ward 2006). Strandplains of beaches, beach ridges, foredunes and back-barrier swamps formed during transgressions by wave action on the coast, resulting in erosion of previous dunes and onshore transport of large quantities of reworked sand with some landward wind transport. During stillstands, wave action continued the onshore transport of sand, declining as the offshore sediment was depleted. Dunes have been variously reported as forming during lowstands (Ward 1978, 2006; Roy & Thom 1981; Thom, Hesp & Bryant 1994), transgressions (Stephens 1982; Pye & Bowman 1984; Jones 1992) and highstands (Kelly & Baker

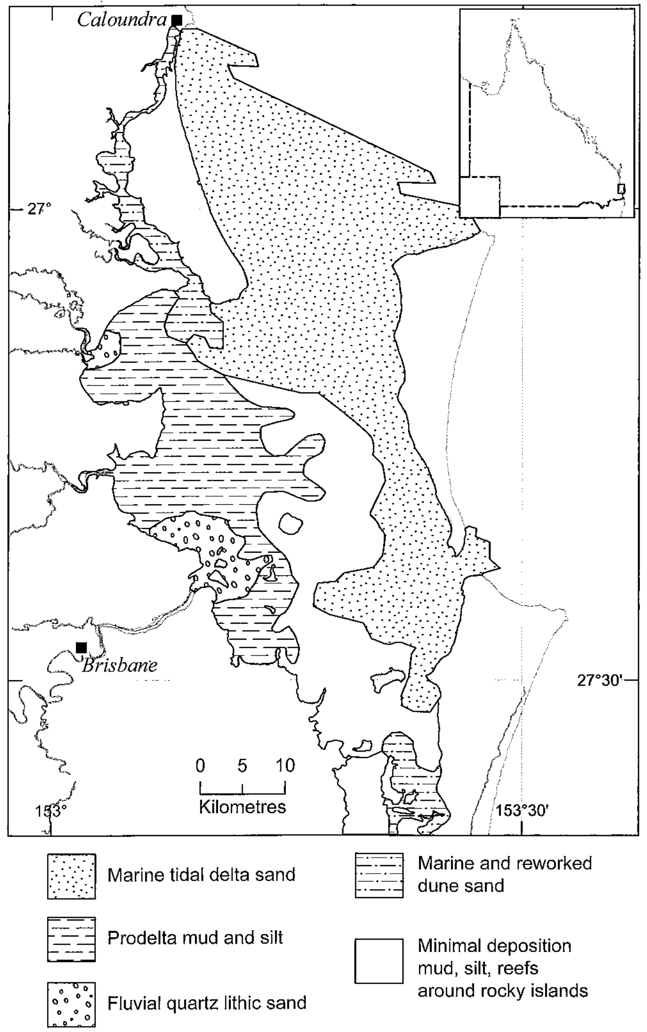


Figure 9.29 Sedimentary environments of Moreton Bay.

1984, Pickett et al. 1984, 1985, 1989). Ward (2006) proposed that dune formation started with more energetic winds in arid glacial periods marking a correlation between the sand-blowing episodes and increased dust content in the Antarctic ice cores, and persisted to interglacial times.

High dune fields have also been blown up between Noosa, Double Island Point and Rainbow Beach south of Fraser Island, forming the Cooloola sandmass. Its history is similar to that of the sand islands.

The sand islands are Pleistocene–Holocene but their initiation has not been dated. The Pliocene–Pleistocene boundary has not been recognised, but in Sandy Cape 1–3R bore a calcreted sandstone interval 127.3–109 m beneath the quartz sand contains an early Pleistocene foraminiferal assemblage (Grimes, 1982; Palmieri 1984). On North Stradbroke Island, U-series dates for three coral species recovered by dredging during sandmining from beneath the large degraded parabolic Yankee Jack dune (Ward 2006) yielded $^{230}Th/^{234}U$ isotope ages of 101–108 ka (Pickett et al. 1985), since revised to 119–132 ka (Pickett et al. 1989). Stratigraphic relationships imply that the coral was buried by a prograding shore as the dune advanced into the ancestral Moreton Bay, and that the dune sands and corals were contemporaneous. Ward (1985) questioned the dates and interpretation, as he recognised three episodes of sand blowing and three further high shores before the Holocene. Along the coast he recognised four interglacial shoreline deposits, correlating them with MIS 5, MIS 7, MIS 9 and MIS 11, allowing estimation of the ages of the dunes and beach deposits by cross-correlation with the dust content in the ice cores from Antarctica. His oldest age on North Stradbroke Island was 360 ka for the Kankee Jack dune overlying the dated corals. TL dating of the quartz sand suggests that in places like Cooloola, the dunes may be as old as 750 ka (Tejan-Kella et al. 1990). The Pleistocene–Holocene boundary has not been distinguished in the Sandy Cape 1 3R bore.

Figure 9.30 Oblique aerial view of the east coast of North Stradbroke Island with the northern end of 18 Mile Swamp in view.

Large sheltered embayments: Moreton and Hervey bays, in the lee of the barrier islands, are significantly populated regions and have been well studied. Neil (1998) reviewed the geomorphological history of Moreton Bay, and Lang et al. (1998) and Lockhart, Lang and Allen (1998) summarised sedimentological and geophysical studies.

Wedge-shaped Moreton Bay occupies ~1500 km², has maximum dimensions of ~80 × 35 km, opens to the ocean in the north, has maximum water depths of 40 m in the north to 6 m in the south and has an average tidal range of ~2 m. Four tidal entrances (at Southport, Jumpinpin, South Passage and North Passage) contain extensive tidal deltas. The Pleistocene Moreton and Stradbroke islands protected the northern bay from vigorous wave attack through interglacial times. The southern sector developed as a back-barrier lagoon behind Holocene South Stradbroke Island.

Four Pleistocene pre-glacial successions in the northern bay represent the last three Pleistocene high sea levels. They suggest >340 ka for the early sequence, 340–318 ka and 240–170 ka for the following periods of high sea level and 140–120 ka for the last interglacial sequence (Evans, Stephens & Shorten 1992). From 119–28 ka, sea level fell gradually with minor fluctuations allowing rivers to produce incised valleys up to 46 m deep and 3 km wide. Most of the streams, including the Logan, Tingalpa, Pine and Caboolture rivers, joined the Brisbane River and discharged ~20 km north of North Passage. The Pimpama and Coomera rivers incised >60 m and flowed southeasterly beneath present-day South Stradbroke Island to the shelf edge. During the glacial lowstand at 28–18 ka, the bay was completely dry.

Hervey Bay occupies 4000 km² in the lee of Fraser Island, is funnel-shaped to the north, has maximum dimensions of 120 × 70 km, opens to the ocean in the north and has a maximum water depth of 50 m. Limited muddy sediment enters from the Mary, Elliot and Burnett rivers, and quartzose sands enter from the Great Sandy Strait in the south.

The Quaternary geological history of Hervey Bay is similar to that of Moreton Bay and an ancient incised subaerial drainage system extends from the bay to the continental shelf (Payenberg et al. 2006). The incised valley, up to 50 m deep and 600 m wide, formed during the Last Glacial Maximum. In high sea-level periods, it transported nearshore quartz sands to the Capricorn Channel and beyond to the Tasman Basin (Boyd et al. 2008), explaining why large quartz sand masses are not known north of the Breaksea Spit.

Continental shelf: The continental shelf in southeastern Queensland has a steep continental slope and is ~80 km wide, narrowing off the tips of the sand islands to be 16 km off Breaksea Spit. The shelf is starved of sediment owing to the limited terrigenous input and the combined effects of the northerly longshore drift promoting loss via submarine canyons and, on the outer shelf, the powerful Eastern Australian Current, which has been measured at >100 cm/s on the shelf edge. Four sedimentary zones are:

- the **nearshore slope** with a thick quartz sand wedge, the profile of equilibrium of the surf zone, from the beach to 20–30 m at a gradient of 0.5–1°
- the **inner shelf**, marked by a well-defined change in slope to a relatively flat plain to 40 m, with or without a thin covering of fine quartz sand
- the **middle shelf**, a subhorizontal plain at 40–100 m with numerous banks (Gardiner Banks off Fraser Island, Barwon Banks off Noosa) and carbonate hardgrounds
- the **outer shelf** with calcareous sands, a gently sloping (1–3°) plain to the shelf break at up to 200 m (Figure 9.31).

East of Fraser Island a nearshore sediment wedge of quartz sands grades into a thin sand layer beneath the inner shelf and under the middle shelf, and a thicker massive poorly reflecting body (Figure 9.32) interpreted as a carbonate build-up grades eastwards into at least three downlapping packages prograding to the outer shelf (Nunn 1982; Marshall et al. 1998). The carbonate build-ups are beneath and seaward of Gardiner Banks on the older Fraser Platform and form the scarp at the middle–outer shelf boundary at 105 m. The build-ups are at least 65 m thick and Gardiner Banks rise to within 25 m of sea level and are 20 m above the surrounding sea floor. Dredge samples from the 105 m cliff and the top of Gardiner Banks include coral-algal boundstones, rhodolith floatstones to

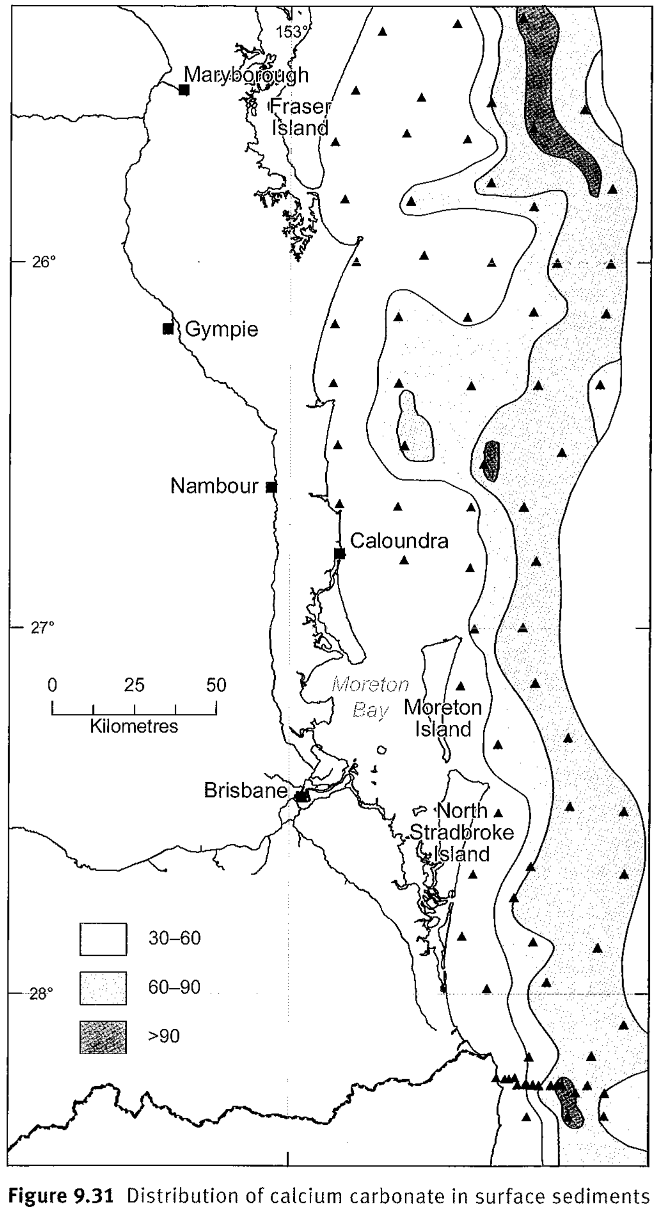


Figure 9.31 Distribution of calcium carbonate in surface sediments on the southern Queensland continental shelf (wt%).

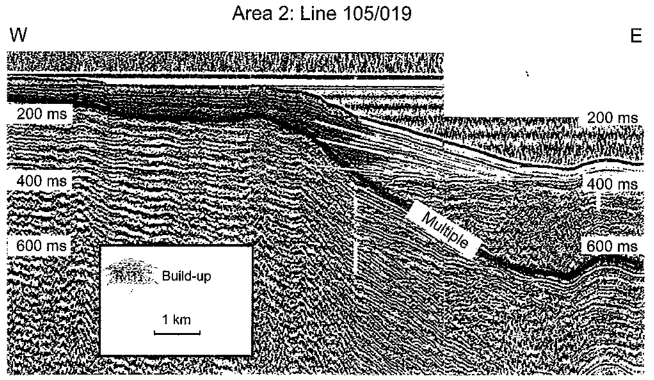


Figure 9.32 Part of a seismic profile showing the three downlapping sequences in front of an upper carbonate build-up platform east of Fraser Island (from Marshall et al. 1998).

rudstone and bioclastic grainstones–packstones. Coralline algae and hermatypic corals dominate (Marshall et al. 1998). The build-ups are Quaternary with the oldest nannofossils indicating 390 ka and appear to have been backstepping from the 105 m cliff to their present site (Stephens 1982; Harris et al. 1996). Although their present bathymetry is between supposed highstand and lowstand levels, the material so far retrieved shows no evidence of meteoric diagenesis. It is suggested that the limestones have been lithified since the last postglacial transgression.

A sequence stratigraphic model for the shelf east of South Stradbroke Island (Lockhart 2001) includes five sequences. The lower three represent early lowstand fluvial deposits filling topographic irregularities and trangressive middle Pleistocene deposits. The basal reflector has been correlated with reflector S_1 of Davies (1975), which he equated to a middle–early Pliocene unconformity. Mounds on the lower boundaries of these sequences represent relic barriers formed during transgressions identified by Searle (1982). Overlying these subhorizontal sequences is a set of three progradational subsequences that become progressively older with their tops benched seawards. They are interpreted as regressive deposits perched on the shelf during the three interstadials (100 ka, 80 ka, 60 ka) of the last glacial sea-level fall. The erosional surface of the Last Glacial Maximum is identified and the overlying sequence is very thin or absent. It is thickest seaward of the tidal deltas where a wedge-shaped sequence downlaps onto the erosional surface such as off Jumpinpin.

9.9.4 Marginal platform and basins

The various deep ocean drilling programs and Geoscience Australia's surveys of the continental margin have added greatly to our knowledge of offshore platforms and deep basins, but our understanding of the Quaternary sequences is still very sketchy. The geological framework of the region east of the continental shelf was already set by the Quaternary. The Queensland Platform was separated from the Great Barrier Reef Shelf to the west by the Queensland Basin and from the Marion Platform to the south by the Townsville Basin. The Marion Platform is a deeper water extension of the Great Barrier Reef Shelf to the west. The platforms and basins are bordered to the north, east and south by oceanic basins. Tectonically, the region was subsiding throughout the Cenozoic but experienced a significant pulse between 2 Ma and 3 Ma. The early Pleistocene was also a period of relatively high global sea levels. By the beginning of the Quaternary, the region was almost in its present position latitudinally, experiencing subtropical to tropical conditions. Palaeoceanographic changes and variations in relative sea level were the main controlling factors on carbonate production and ultimately on sedimentation across the region.

The **Queensland Platform** is an isolated, partially drowned carbonate platform, essentially isolated from terrigenous influx throughout the Quaternary. Its Quaternary sequence consists of biogenic carbonate ooze, reefal detritus and coral reefs. Prior to 2.6 Ma, global sea level was high and the platform was in shallow tropical waters with flourishing coral reefs well established as large pinnacles (Flinders, Holmes, Bougainville and Osprey reefs) on fault blocks along the western edge of the plateau and reef complexes developed across the higher parts of the platform (Lihou, Tregrosse, Coringa, Willis and Diana reefs). Numerous drowned reefs have been reported (Davies et al. 1989). The accelerated subsidence across the Pliocene–Pleistocene boundary increased relative water depth markedly, leading to contraction of reef growth (Katz & Miller 1993). The present reef complexes are surrounded by prominent terraces at 40–50 m, outlining the Neogene reef developments on which they sit. The pinnacle reefs continued growth but became smaller in area. The reefs have not been drilled, so there are no data on their Quaternary development, but Davies et al. (1989, fig. 9c) provided a schematic cross-section of the plateau from seismic profiles, showing their relationship to the underlying Cenozoic stratigraphy.

Descriptions of the extant reefs are provided by Orme (1977) and of the coral faunas by Done (1982). These reefs are typical high-energy reefs similar to those of the Great Barrier Reef and show similar geomorphic characteristics (Jell & Neil 1998). Reef-derived material is found in front of many reefs. Away from the reefs, hemipelagic oozes predominate.

The **Marion Platform** forms a deeper water extension of the continental shelf and is not as isolated from terrigenous input as the Queensland Platform. Reef growth is restricted to Marion Reef on the northeastern corner and Saumarez Reef at the southeastern end of the plateau. Davies et al. (1989, figs 9c, 9e) show the relation of these extant reefs to the older carbonate build-ups of the platform. The geological development of these reefs has not been investigated. The rest of the plateau is covered by a thin layer of skeletal grainstone composed of planktonic foraminifera that reflects modern current-swept conditions of the sea floor. It overlies a relatively thin (<50 m) unconsolidated Pleistocene–Miocene hemipelagic foraminiferal packstone heavily bioturbated of mostly very fine sand-sized planktonic foraminifers in a clay-rich carbonate matrix. The lower boundary of this unit is at 194 m.

The **Queensland Basin** is a long, narrow, deep depression east of the siliciclastic/carbonate Great Barrier Reef Platform and west of the carbonate-dominated Queensland Platform. It is the repository for biopelagic sediment as well as terrigenous and carbonate sediments from the adjoining platforms. The Quaternary comprises the upper part of Sequence A (late Pliocene – present) of Symonds, Davies and Parisi (1983), which can be mapped across the basin. ODP Leg 133 drilled four sites (Sites 819–822) on the slope from the Great Barrier Reef Platform bottoming in the early Pleistocene. Site 823 in the axis of the basin and Sites 811/825 and 824 on the western slopes of the Queensland Platform intersect what appears to be a continuous Quaternary sequence. Site 823, the base of the Pleistocene (2.58 Ma) has been identified by Watts, Varga and Feary (1993) at 242–243 m below the sea floor on the occurrence of *Discoaster tamalis* in the lower part of their Unit II. The Quaternary succession is composed of hemipelagic muds, which are compositionally variable—admixtures of terrigenous mud, pelagic carbonate, aragonitic periplatform ooze and turbidites, debris flows and slump deposits from the slopes bringing intermittent influxes of coarser siliciclastic and bioclastic material downslope from the adjoining platforms. These submarine gravity mass flow deposits are controlled predominantly by sea-level oscillations and provide insights into the history of the adjoining platforms. Davies and McKenzie (1993), comparing results from the western slope holes with the basinal sequence, recognised up to 18 events (many recognisable at all sites). In general, turbidites characterise both high and low sea levels and debris flows characterise the lowstands, whereas slump deposits define periods of major slope instability during lowstands and occasionally during transgressions. Turbidites and other gravity deposits (such as those at Site 823) do not necessarily represent submarine fan deposits, particularly if they are composed of hemipelagic sediments reworked from drowned platforms and slopes.

The early Pleistocene reflects rapid sedimentation and progradation on the Great Barrier Reef Shelf, whereas from 750 ka onwards aggradation became dominant (Feary et al. 1993a). An increase in carbonate content at about 500 ka in Site 822 at the base of the slope might signal the initiation of the Great Barrier Reef. Once the reefs were established, the shelf became rimmed and its supply of sediment to its slope and adjoining basin changed. A high-resolution stratigraphic record for the last 330 ka, developed from the ODP cores and 35 shallow piston cores across the basin, allowed mass accumulation rates for siliciclastic and carbonate sediment to be calculated for the basin over the last 31 ka (Dunbar, Dickens & Carter 2000). The results demonstrated that the maximum and minimum carbonate accumulations occur during highstand and lowstand respectively, and maximum and minimum siliciclastic accumulations occur during transgression and lowstand respectively. Page, Dickens and Dunbar (2003) analysed six cores along the western slope of the basin and showed that the maximum siliciclastic flux to the basin since 25 ka was 11–7 ka, corresponding to the later stage of transgression. Sidescan sonar of the Queensland Basin showed four sediment gravity flows originating at the basin margin and bending northwards into the trough axis, and no evidence of submarine fans (Dunbar, Dickens & Carter 2000). Sediment deposits in this setting reflect tectonic and eustatic events that caused slope instabilities, rather than migration of different submarine fan facies.

The **Townsville Basin** is in water depths of 200–2000 m between the Marion and Queensland platforms and abutts the Great Barrier Reef Platform in the west. In addition to the pelagic sediment supply, it received bioclastic carbonate material from all three platforms and some siliciclastic sediment from the continent that would have bypassed the reefs of the latter platform. Only seismic profiles of the basin are available and these show a relatively uniform sequence (200–700 m thick) of Pliocene to present pelagic oozes, terrigenous and calcareous turbidites and slump deposits throughout the basin (Struckmeyer & Symonds 1997). ODP drill sites on the edges of the adjacent platforms show that most of the sediment is of oozes and gravity flows (Struckmeyer & Symonds 1997).

The Quaternary of the **Coral Sea Basin** is a thin layer (100–140 m) widely distributed across the basin. In DSDP 210 it consists of 112 m of graded cycles (typically 11 cm thick) of silt and clay with interbeds of nannofossil ooze (Burns & Andrews 1973). These cycles are interpreted as turbidites and their mineralogy, as well as the basin morphology, indicates they are sourced from New Guinea. The biogenic beds may also be turbidites and are thought to be sourced from the west or southwest, including the Great Barrier Reef region. The **Cato Basin** and northern **Tasman Basin** have not been drilled.

9.9.5 Economic significance

The Great Barrier Reef is one of Australia's major tourism attractions and is well managed by the Great Barrier Reef Marine Park Authority. The dune fields contain large freshwater reserves and are important water resources for a number of coastal towns and communities (Chapter 13). Rutile, zircon, monazite and ilmenite have been mined from the sand deposits of the southern coast (Brown 1971; Colwell 1982) and silica sand is mined on North Stradbroke Island for glassmaking. Unconsolidated sands from the bays are used for construction and special projects such as the Brisbane airport development, expansion of the Port of Brisbane at the mouth of the Brisbane River and beach replenishment programs (Jones & Holmes 1993).

9.10 Synthesis (GJ Price)

Development of numerous Quaternary dating methods has been paramount in the construction of a reliable geochronological framework for Queensland's geological, climatic and biological histories. Results of several studies not only highlight orbital forcing as a major controller of Quaternary climates, but also suggest that there has been a shift towards progressively drier conditions since 2.6 Ma. The climatic history since 350 ka is by far the best known period (Figure 9.33), including all or parts of the chronological application 'windows' of numerous dating methods such as Ar–Ar, K–Ar, U–Th, ^{14}C, OSL, TL, electron spin resonance and amino acid racemisation.

Prior to 240 ka, much of Queensland experienced a warm, humid climate as evidenced by enhanced fluvial activity in the Channel Country and high rates of weathering and extensive rainforest cover on the eastern margins. Diverse vertebrate faunas also existed in central and southeastern Queensland at that time. Low rates of weathering and soil development demonstrate more arid climates in some parts of central Queensland.

The Penultimate Glacial Cycle ~190–126 ka saw a significant shift towards drier conditions. Weathering rates were greatly reduced in southeastern and central Queensland. A major shift towards reduction in rainforest cover and expansion in woodlands is evident from northern pollen records. Similarly, while arid and open-habitat adapted vertebrate faunas became dominant in central eastern Queensland, numerous rainforest adapted forms suffered either extinction or significant geographic range

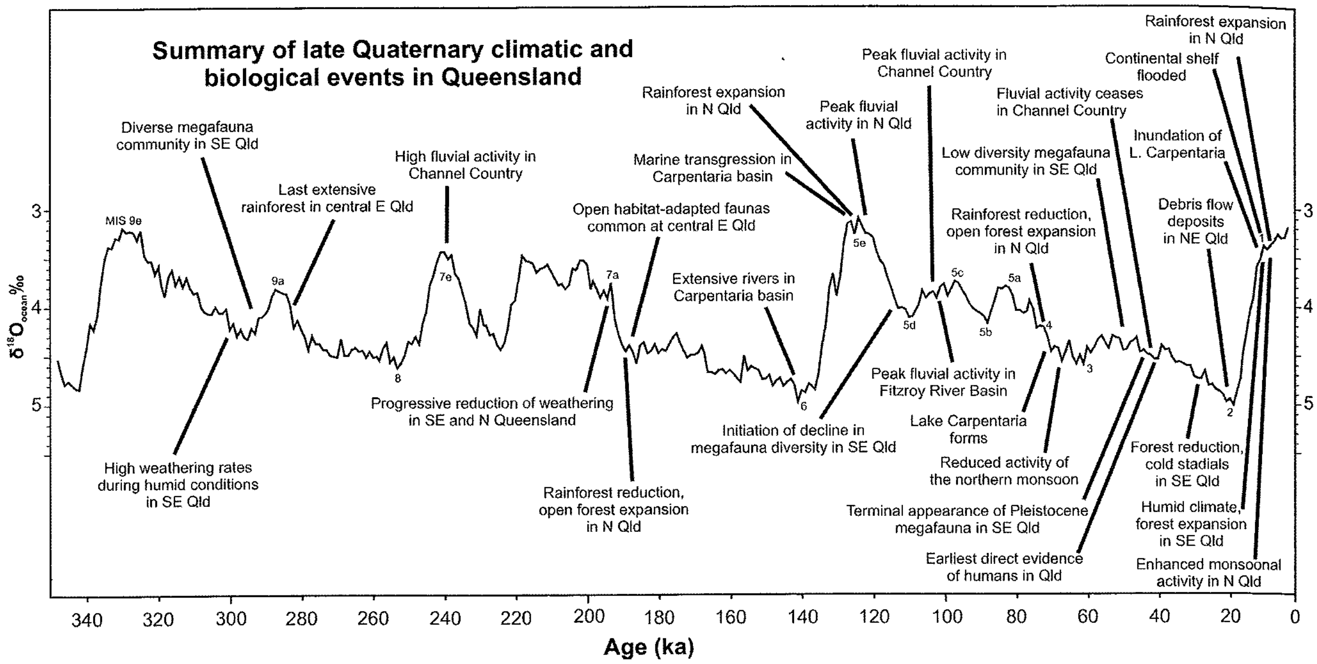


Figure 9.33 Summary of major climatic and biological events for the last 350 ka in Queensland relative to the oxygen isotope curve.

shifts. At the peak of the Penultimate Glacial Maximum around 140 ka, sea levels were so low that numerous rivers extended across the Carpentaria basin and the eastern continental shelf.

The late Pleistocene (126–11.7 ka) was an episode of intense climatic change, culminating in the most arid phase of the Quaternary. The start of the last glacial cycle saw a return to widespread warm and humid conditions, as evidenced by the re-expansion of northern rainforests, a peak in fluvial activity in the north and in the Channel Country, and inundation of the Carpentaria Basin and the Great Barrier Reef Shelf as a result of a marine transgression. The interglacial/interstadial of MIS 5 also saw a progressive decline in biological diversity of southeastern Queensland terrestrial faunas, evidenced by the late MIS 5a communities (~80 ka) being markedly less diverse than the older 5d–e assemblages (120–130 ka). Around 70 ka, a major reduction in the northern monsoon occurred as sea levels fell and Lake Carpentaria re-emerged. The change in precipitation was also a catalyst for expansion of sclerophyllous vegetation and reduction of rainforests in northern Queensland. Late Pleistocene faunal communities experienced all-time lows in diversity after that time, leading to the terminal appearances of some species of now-extinct megafauna by 40 ka. A major reduction in precipitation occurred just prior to the final megafaunal appearance, as indicated by the cessation of fluvial activity and emergence of dunes in the Channel Country. An intensification of El Niño Southern Oscillation and reduction in the northern monsoon is also evident at that time. Such a shift occurred immediately prior to the oldest direct evidence of humans in Queensland. After 45 ka, rainforest again retracted, while open sclerophyll woodland expanded in the north. A similar vegetation shift is also apparent from deposits of that age in southeastern Queensland. Widespread cool, dry and arid conditions define the Last Glacial Maximum (~30–18 ka), as evidenced by extensive aeolian activity and reduced fluvial episodes in the south, and extensive debris flow deposition and alluvial fan building in the north.

After the peak of the Last Glacial Maximum, a significant warming episode leading into the Holocene is associated with a return to higher precipitation and, especially, the re-intensification of the northern monsoon. A marine incursion occurred in the Carpentaria Basin, while rainforests expanded in the northeast. Similarly, an increasingly humid climate and expansion of forest is evident in the southeast. The early Holocene also marked the last episode of volcanic activity in Queensland. By 12 ka the sea began to flood the continental shelf with the older Pleistocene reefs forming limestone islands, which by 9 ka began to be submerged again, allowing initiation of Holocene growth of the Great Barrier Reef.

Queensland's geological record has been critical for determining the nature of past climatic change and associated landscape and ecological evolution as well as sea-level and oceanographic changes. Importantly, it provides a baseline understanding of climate variability; this is crucial for distinguishing anthropogenic environmental impacts from natural inherent climate change. However, despite significant advances in our understanding of Queensland's Quaternary history, data remain patchy and several questions remain unresolved. For example:

- Few early Pleistocene archives exist, so it is not possible to examine geological, climatic and biological evolution during the establishment of the glacial–interglacial cycles, nor the shift between 41 and 100 thousand-year cyclicity. This is compounded by the limited number of Quaternary geochronological methods that allow dating of deposits older than 500 ka.

- Although the role of orbital forcing and its influence on ecosystem and landscape evolution is relatively well established for the late Quaternary, we still have only a limited understanding of the history of the development and nature of a variety of other significant climatic processes (such as El Niño Southern Oscillation) and formation of related climate-controlling phenomena (such as the Western Pacific Warm Pool). This is due mostly to a rarity of high-resolution palaeoclimate archives.
- While Queensland's Quaternary geological and palaeontological record demonstrates a major trend towards increased aridification and generally drier climates since 350 ka, it is becoming increasingly evident that such events were not synchronous across Australia. For example, changes in precipitation and humidity led to the retraction of rainforests, expansion of open sclerophyllous habitat and extinction of numerous vertebrate taxa in central eastern Queensland during the middle Pleistocene. In contrast, similar climate-driven vegetation changes and faunal extinctions have not been identified from temporally coeval deposits in southeastern Australia. Although those observations suggest that distinct climatic processes were operating in both regions during the middle Pleistocene, the nature of such climatic controls remains unclear. Better spatial control and correlation of climatic data across the continent are required.
- An understanding of how Quaternary biota (such as megafauna) responded to various episodes of climate change and environmental perturbations is fundamental. Yet only a few reliably dated fossil records exist, and these are mostly restricted to central eastern and southeastern Queensland. Although it is evident that both megafauna and smaller-sized species were detrimentally affected by late Quaternary climate changes, the biological and ecological nature of their response to such episodes remains unclear. A lack of extensive spatially and temporally constrained records for most Pleistocene megafaunal species makes it difficult, if not impossible, to comprehensively test leading extinction hypotheses. Similarly, a lack of data limits our understanding of the timing of the emergence of the modern biota in the late Quaternary.
- In many cases where well-dated faunal records do exist, they are rarely associated with independent local palaeoclimate archives (such as pollen-bearing sediment cores and speleothems), so it is not possible to independently establish the nature of local climatic conditions. Diverse, well-dated palaeoenvironmental archives are the key to understanding climatic influences on Quaternary ecosystems and biota.
- Although knowledge of Quaternary history of offshore eastern Queensland has greatly increased in the last 50 years, there have been few attempts to correlate changes on land to those seen in the marine record. The latter record, especially our understanding of the Pleistocene structure and development of the Great Barrier Reef, is also still incomplete. This will only be remedied by systematic drilling through the total reefal succession.

Finally, the ability to integrate independent palaeoclimate records from both the terrestrial and marine (e.g. reef) realms in relation to tectonic activity would significantly enhance understanding of the development of Quaternary climate systems. This would be fundamental for examination of the array of climatic processes that have operated at varying regional, continental, and global scales through the Quaternary. Also, it would no doubt lead to a better understanding of the specific climate mechanisms that have influenced Quaternary ecosystems in Queensland. Ideally, such knowledge gaps can be filled through the development of higher resolution palaeoclimate and palaeontological archives, especially on wider spatial and temporal scales. However, this will require further advancements in dating methods, development of new geochronological techniques and an increase in the scope and focus of Quaternary studies in general, especially with integration of expertise and datasets across traditional scientific boundaries.