ELSEVIER

Quaternary Science Reviews 23 (2004) 107–135

QSR

Cyclone pumping, sediment partitioning and the development of the Great Barrier Reef shelf system: a review

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Received 20 December 2002; accepted 10 October 2003

Abstract

The modern Great Barrier Reef (GBR) is part of the world's largest and best known mixed terrigenous-carbonate continental margin. The GBR shelf contains three shore-parallel sedimentary belts: an inner shelf zone of terrigenous sedimentation at depths of 0–22 m; a middle shelf zone of sediment starvation at depths of 22–40 m; and an outer shelf reef tract with its inner edge at ca. 40 m depth. These zones are controlled by the dynamics of northward, fair-weather, along-shelf drift, driven by southeasterly trade winds, and by the regular passage of tropical cyclones. Cyclones cause wind-driven north-directed middle shelf flows in excess of 130 cm/s, which erode the seabed, concentrate the sparse mobile sediment into sand ribbons, and advect suspended load onto the outer part of the nearshore terrigenous sediment prism and into inter-reef depocentres within the outer shelf reef tract. Cyclones largely control the input of new sediment into the GBR system, via river flooding, seabed erosion or reef breakage. They also help to control the partitioning and dispersion of the three main shore-parallel belts of sediment, and hence stratigraphic accumulation. Acting as a sediment pump, especially during interglacial highstands, cyclones have exerted great control on the development of the modern GBR province and its sediments by maintaining a broad shelf-parallel zone of episodically mobilised sediment and scoured seabed upon which coral reefs have been unable to form. Cyclones may also have partly controlled the timing of initiation of the first GBR at ~0.6 mybp. Contrary to current models, GBR storm beds are most likely to be preserved intact close to the shoreline, and become coarser-grained away from the shoreline. For the central GBR, "highstand shedding" only applies to carbonate sediment at the scale of local reefs; system-wide, oceanographic controls cause high rates of carbonate sedimentation on the slope during both sea-level rise and highstand; concomitantly, terrigenous sediment accumulates fastest on the slope during sea-level rise, and slowest during sea-level lowstand and highstand.

1. Introduction

Understanding the relative impact on shelf sediments of daily low-energy versus episodic high-energy phenomena (e.g. cyclones, tsunami) is crucial to our understanding of how shallow water sedimentary systems function (e.g. Siringan and Anderson, 1994). Around 30–40% of today's continental margins lie in the tropics and sub-tropics, where cyclones are major controls of sediment supply to the shelf, and sediment transport upon it. The geological development of carbonate and mixed carbonate shelves is especially important for the petroleum industry, and has been summarised by a unique sequence stratigraphic model characterised by highstand shedding (cf. Vail et al., 1991 with James and Kendall, 1992; Schlager et al., 1994). Understanding the sedimentary dynamics of tropical shelves at various stages of sea level is therefore a fundamentally important issue. However, current sedimentation models for tropical shelves are strongly influenced by studies of ocean plateaux such as the Bahamas (see, for example, references in Isern and Anselmetti, 2001), and often do not fit well with the characteristics displayed by the geologically common mixed terrigenous-carbonate systems. Furthermore, and despite papers which describe the influence exerted by cyclones on reefs and reef lagoons (e.g. Stoddart, 1973, 1974; Done, 1992; Scoffin, 1993; Beanish and Jones, 2002), established models of tropical sedimentation take little account of the effects of cyclones on sedimentation.

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0277-3791/$ - see front matter
doi:10.1016/j.quascirev.2003.10.003

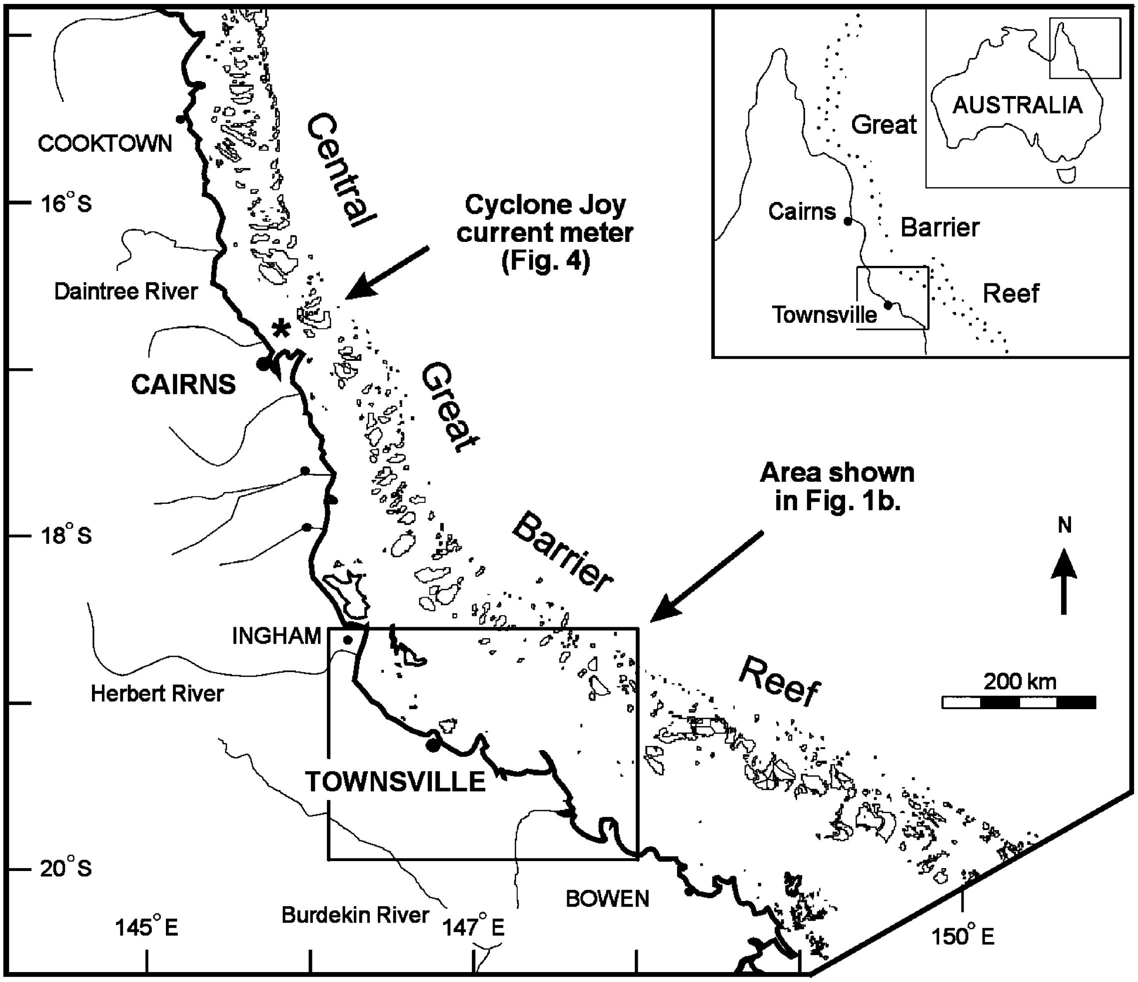
We aim to show in this paper that cyclones are a critical control on sediment partitioning and reef development on the tropical Great Barrier Reef (GBR) continental margin. Given the low seafloor-gradients of <1:1000, which characterise the GBR shelf platform, the effects of cyclones are at least as important as those of slope (i.e. gravity) in controlling shelf development and sediment dispersion. At times of sea-level highstand, cyclones exercise a powerful shelf-parallel sediment pumping effect, and at lowstand—when the shelf is exposed and the shoreline lies on the upper continental slope—their power is concentrated along a narrow coastal zone which contains unconsolidated sediment prone to direct downslope remobilisation to the deep sea. Furthermore, because the highstand GBR shelf is a rimmed shelf, it is only at sea-level lowstands that the shoreline system is exposed to the unrestricted impacts of ocean swell and wave-induced offshelf downslope transport. Puzzlingly, however, lowstand is also a time of low rates of terrigenous sediment input onto the GBR slope (Dunbar et al., 2000).

Our model of GBR sediment transport and deposition emphasises the importance of tropical storms (cyclones, typhoons, hurricanes) as sedimentary agents, which—in association with repeated cycles of sea-level change—control the character of the GBR and probably other reef-rimmed carbonate shelves. We stress the importance of *along-shelf transport* for the distribution of sediment on the middle shelf (cf. Swift et al., 1986), in contrast with the focus on processes of *cross-shelf transport* that characterises many earlier studies of shelf storm-layer transport (e.g. many citations in Myrow and Southard, 1996). We report new data on cyclone-associated flows and on the nature and distribution of sedimentary bedforms on the central GBR shelf, and infer the nature of both highstand and lowstand shelf-wide sedimentary processes. The initiation of the GBR at ca. 600–400 kyr BP, i.e. about the time of the 'mid-Pleistocene transition' (MPT) between 41 and 100 kyr-long climatic cycles (Berger and Jansen, 1994; Berger et al., 1994), may relate in part to the development then, during the highstand of marine isotope stage 11 (MIS 11), of a relatively wide and deeper water shelf, which was able to sustain both a cyclone corridor and a reef tract.

2. The GBR shelf

The GBR shelf (Fig. 1a) is the largest modern tropical mixed carbonate-siliciclastic shelf system on Earth, with a well-understood tectonic setting (e.g. Symonds et al., 1983). The GBR has had its older stratigraphy delineated by scientific drilling, and therefore also comprises our best known example of the geological development of a tropical passive continental margin (Davies et al., 1989; McKenzie et al., 1993; International Consortium, 2001; Webster and Davies, 2003). Much of the earlier work on the GBR was concentrated on the geomorphology of the reefs themselves, with Fairbridge (1950) concluding that understanding "… the Australian shelf reefs require(s) the utilisation, at least of some parts, of (a) the subsidence theory promulgated by Darwin, Dana, Davis, (b) the antecedent platform theory as set forth by Wharton, Agassiz, Andrews, Vaughan, Hoffmeister and Ladd, and (c) the glacial control theory of Penck, Daly and others". In a comprehensive summary of this and more modern research, Hopley (1982) concluded that "… diversity itself is a basic characteristic of the GBR", and stressed the "…strong control of present morphology by antecedent-platform relief" together with "… the nature of the Holocene transgression, exposure conditions, incidence of tropical cyclones, protection by distance from the mainland from freshwater flushing, and at least a mesotidal range …". This diversity precludes any one, simple factor from explaining fully the "origin" of the GBR. Nonetheless, we believe that some earlier studies have focussed too narrowly on the reef tract itself, to the exclusion of the wider environment within which the reefs have developed. In this paper, therefore, we consider the development of the central GBR within a wider natural system which includes the shelf and shoreline regions contiguous with the reefs themselves, and which we term the GBRscape. In other words, we take as our context the changing physical, geological and biological processes which have shaped the entire modern GBR shelf, not just the reef tract, with a special emphasis in this paper on the role played by cyclones.

The central region of the modern GBR shelf comprises three distinct shelf-parallel sedimentary zones (Maxwell, 1968; Belperio, 1988) (Figs. 1b and 2). The *inner shelf* (0–22 m depth) comprises a terrigenous inner shelf prism (ISP) of mixed sand and mud, which is most commonly shoreface-attached. The prism is commonly 5–10 m thick near sediment point sources and thins along-shelf and seawards (Johnson and Searle, 1984; Carter et al., 1993). Commonly, the tapering seaward edge of the ISP is located 15–20 km offshore, in water depths of 20–22 m. Pervasive bioturbation results in rapid homogenisation of storm beds and hence poor preservation of bedding within the ISP (cf. Gagan et al., 1988). The *middle shelf* (22–40 m depth) is starved of terrigenous sediment and is generally devoid also of coral reefs, except for local fringing reefs which occur around high islands which rise from the middle shelf, such as the Whitsundays. A thin veneer, usually <1 m thick, of poorly sorted, shelly, muddy sand and shell hash overlies weathered Pleistocene clay, which locally outcrops at the seabed (Harris et al., 1990; Ohlenbusch, 1991; Carter et al., 1993). The *outer shelf* (40–80 m depth) is also starved of terrigenous sediment, but encompasses scattered accumulations of coralgal reef



(a)

Fig. 1. (a) Locality map for the central Great Barrier Reef shelf, showing the location of the Cyclone Joy current meter. (b) Bathymetry and the terminology of the inner, middle and outer shelf in the central GBR region, and location of LADS image of Fig. 6.

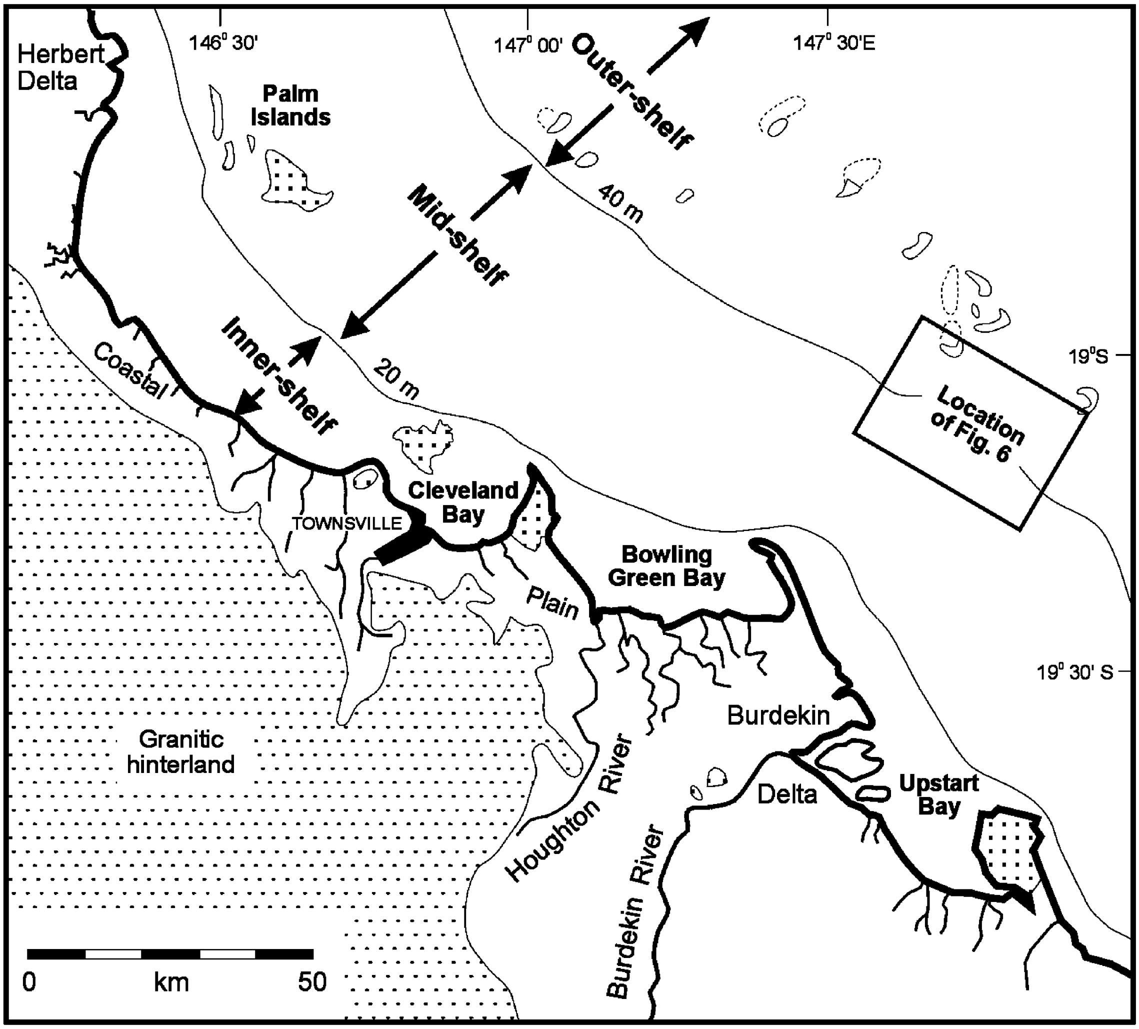
framework and associated detrital carbonate sediments where modern reefs rise to the surface (e.g. Maxwell, 1968). Locally, carbonate mud or *Halimeda* banks up to several metres thick accumulate between reefs (Orme, 1985; Orme and Salama, 1988; Dye, 2001), whilst elsewhere the flat outer shelf plain carries a thin condensed cover of shelly calcsand (Scoffin and Tudhope, 1985).

3. Modern sedimentary processes and products

The three sedimentary zones described above are the end product of sedimentary processes and reef growth which have operated throughout the transgressive (18–5.5 kyr BP) and highstand (5.5–0 kyr BP) phases of the postglacial sea-level cycle, as superimposed upon inherited shelf physiography. Distinct sedimentary processes now dominate in each shelf-parallel zone, as described below, although the boundaries of these zones have of course evolved through space and time. (All radiocarbon dates quoted in this paper are given in conventional radiocarbon years, corrected where appropriate for the marine reservoir effect, as in Larcombe et al., 1995a. Calibrations to calendar years have not been performed on the data).

3.1. GBR inner shelf

During fair-weather conditions, shelf sediment transport results primarily from wave-induced resuspension, combined with trade-wind-driven currents and coastal longshore drift (Larcombe et al., 1995b; Orpin et al., 1999). Suspended sediment concentrations of 10–100 mg/l, caused by resuspension of seabed mud, occur in a 2–10 km wide coastal belt, which drifts north under the influence of the wind-driven northward along-shelf current. Bedload transport of sand is mostly restricted to the net northward-moving beach-shoreface system at depths less than ~5 m (e.g. Beach Protection Authority



(b)

Fig. 1 (*continued*).

of Queensland, 1984; Jones, 1985). Cyclones are a common seasonal feature of the GBR shelf, and typically occur two to three times per summer at latitude 20°S (Fig. 3). Cyclones have well-recognised coastal (Hopley, 1974, 1982, 1984; Chappell and Grindrod, 1983; Chappell et al., 1983; Nott and Hayne, 2001) and reef (Massell and Done, 1993) impacts, and can break and remove all branching corals down to depths of 12 m or more (references cited in Scoffin, 1993). Most importantly, however, it is generally only during the passage of episodic cyclones that sandy and gravelly sediments are able to be mobilised on the shelf (Chappell et al., 1983; Chappell and Grindrod, 1983; Gagan et al., 1988).

In December 1990, Cyclone Joy (category 3; cf. Fig. 3) passed across the GBR shelf, incidentally crossing a seabed current-meter moored in 12 m of water off Cairns (Fig. 1a). The cyclone was accompanied by a 9-day period of along-shelf wind-driven currents, which flowed to the NW at sustained speeds of 60 cm/s and instantaneous speeds up to 140 cm/s *near the bed* (residual of 130 cm/s) (Fig. 4). These flows are fast enough to form longitudinal bedforms on either sand-gravel or mud substrates (e.g. Belderson et al., 1982; Flood, 1983).

In Halifax Bay, north of Townsville, in water depths of 8–12 m, occurs an 8 km-long field of ~13 transverse, very large subaqueous dunes (*sensu* Ashley, 1990) of gravelly shelly sand, as mapped by 3.5 kHz profiling, and by grab and vibrocore sampling (Fig. 5). The individual seabed dunes rest on a substrate of transgressive mangrove mud (9160–7310 ybp), are up to 2 m high with spacings of ~100–300 m, and are today moribund during fair-weather conditions. The dune sediments comprise crudely interlayered beds of moderately well sorted sandy shell gravel, and more poorly sorted, muddy shell gravel (Kirsch, 1999). We infer that the moderately sorted gravelly sands correspond to periods of dune mobility, and that the muddier gravel interbeds correspond to the later downward bioturbation of a

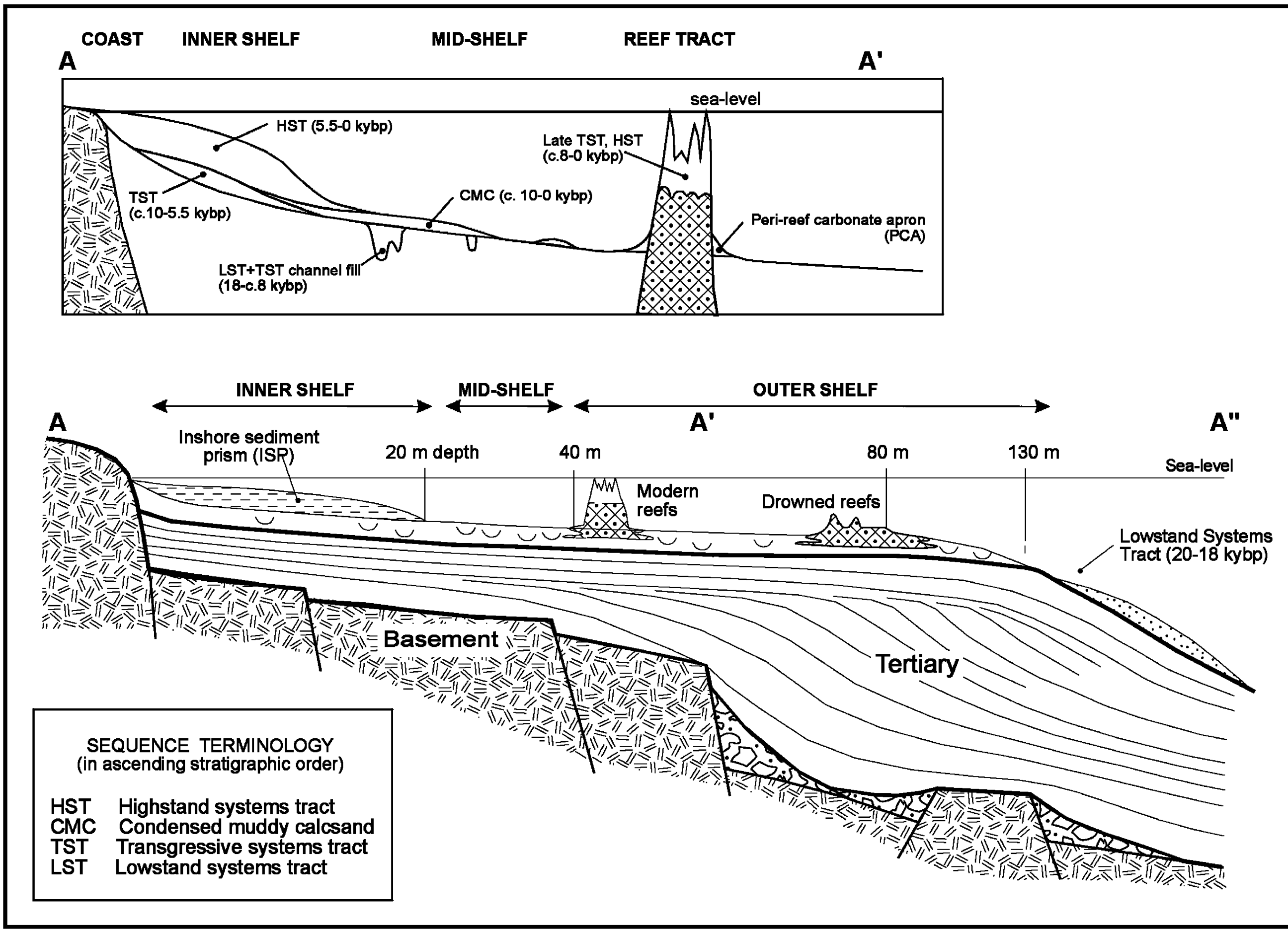


Fig. 2. Reconstructed cross-section through the postglacial sediment facies of the central Great Barrier Reef shelf (modified after Larcombe and Carter, 1998). From landward to seaward, the three shore-parallel sediment provinces are the terrigenous inshore sediment prism (ISP), the mixed middle shelf condensed muddy calcsand (CMC), and the peri-reef carbonate apron (PCA) which occur within the reef tract. Other carbonate facies (not shown) which occur on the outer shelf include inter-reef micrite pools (Dye, 2001), and *Halimeda* calcsand bodies (Orme, 1985).

post-transport mud drape (cf. Gagan et al., 1988, 1990). Muddy sands, similar to those which form the bulk of the inner shelf deposits regionally, occupy the interdune areas. The gravelly sand present in the dunes requires bed shear stresses of $>0.6\,N/m^2$ for mobilisation, and therefore the dunes could only have been activated by cyclonic waves and currents (cf. Orpin et al., 1999). Significantly, the dune crestlines are almost shore-normal, indicating that bedform generation and movement resulted from broadly shore-parallel flow, consistent with the Cyclone Joy flow data presented above.

3.2. GBR middle shelf

Maximum surface tidal current speeds on the middle shelf are generally less than 30 cm/s (Church et al., 1985; Wolanksi and Pickard, 1985). However, when augmented by the regional wind-driven northerly flow, the combined open-shelf surface water speed along the middle shelf commonly reaches 50 cm/s or more (e.g. Kalangi et al., 2000). These fair-weather surface speeds do not cause appreciable sediment transport at the seabed (Orpin et al., 1999). The short fetches available inboard of the main reef tract mean that waves generated locally by small cyclones are limited in size, to around <7 s period and <5 m significant wave height (Hardy et al., 2000, see also http://tsunami.jcu.edu.au/atlas/wave_atlas.shtml), and therefore unable by themselves to cause appreciable bedload transport of gravel and coarse sand. However, cyclones also drive along-shelf currents (Gagan et al., 1988; Wolanski, 1994). During Cyclone Winifred, these currents reached 60 cm/s at the surface at a distance of ~ 80 km from the cyclone centre, i.e. up to twice the speed of typical fair-weather shelf currents (Wolanski and Ridd, 1990). Nearer a cyclone centre, or during more intense cyclones, current enhancement will be much stronger. At such times, sediment unmixing from the seabed results in mud being resuspended and transported northward along-shelf, with the subsequent deposition of a post-storm suspension mud drape (<1 cm thick) on top of the sandy storm bed (Gagan et al., 1988). Within

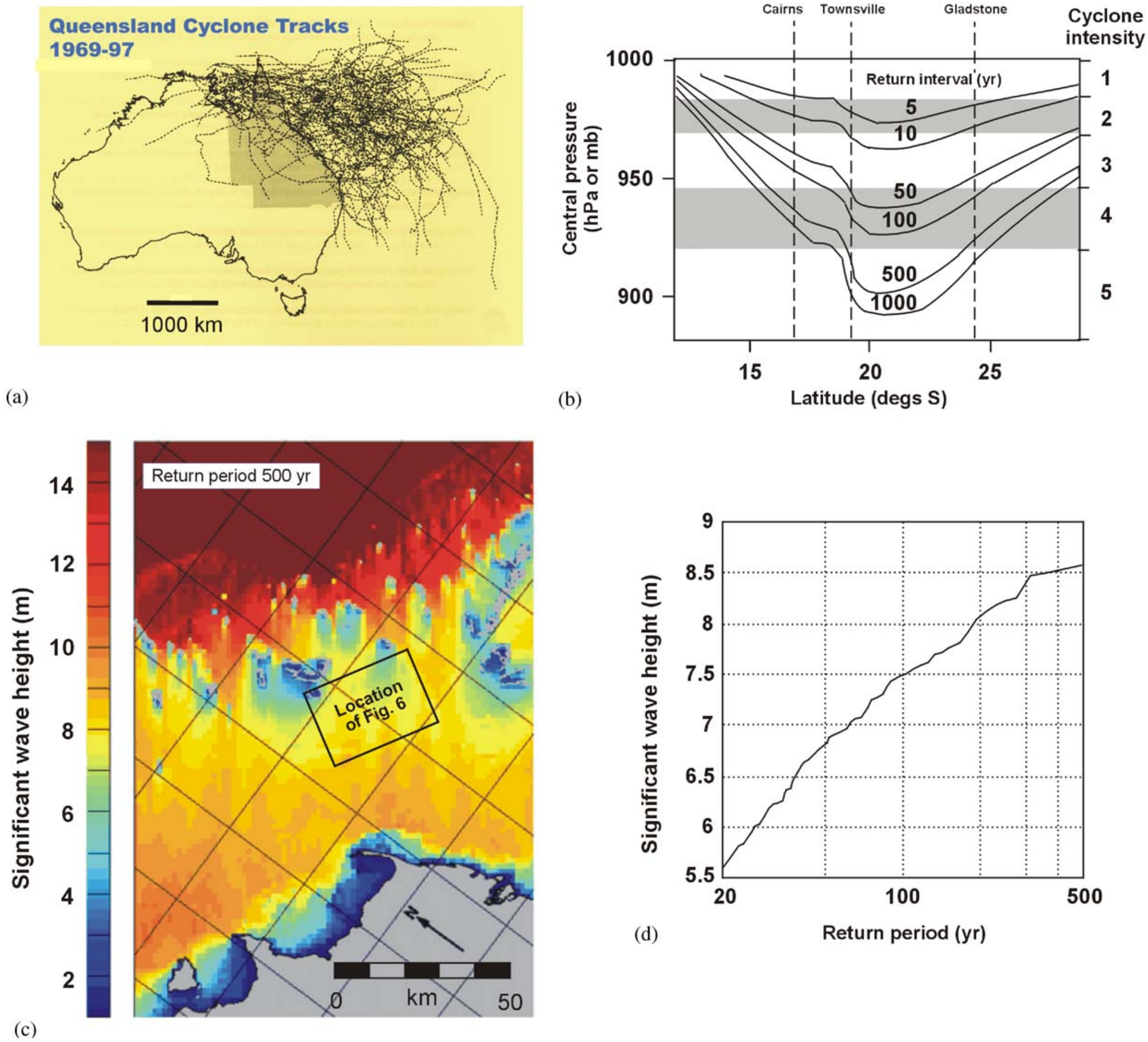


Fig. 3. (a) Map of cyclone tracks for the Great Barrier Reef region for the period 1969–97 (after Puotinen et al., 1997). (b) The expected return period for cyclones of indicated intensities across tropical latitudes on the east coast of Australia (after Massell and Done, 1993). Note that the shortest return intervals of intense cyclones (category 4 and 5) lie between about latitudes 16°S and 24°S, i.e. coincident with the location of the central Great Barrier Reef shelf, between Cairns and Gladstone. (c) Significant wave height as predicted by the deterministic model of Hardy et al. (2000) generated by an intensity 5 cyclone (return period 500 yr) on the central GBR shelf off Cape Bowling Green; area of LADS image (Fig. 6) indicated. Note the wave heights of 12–14 m are characteristic of the open ocean outside the reef tract, whereas heights within the lagoon increase landward from 7–9 m, but decrease sharply across the inner shelf. Location of LADS image of Fig. 6 indicated. (d) Return intervals of significant wave height (also from Hardy et al., 2000) from the wave field illustrated in (c) for a mid-shelf location indicated on (c).

3 months of the passage of Cyclone Winifred bioturbation was well advanced, and after 12 months the mud drape deposited after the cyclone had been fully integrated with the sediments beneath by pervasive bioturbation.

Where sufficient unconsolidated sand–gravel sized sediment exists, along-shelf cyclone-driven flow produces shelf-parallel bed-sediment transport as large dunes, sediment ribbons or small dune fields on all parts of the inner and middle GBR shelf. On the middle to outer shelf between Townsville and Cairns, sand-ribbons, crag-and-tail bedforms and dunes observed by side-scan sonar and Laser Airborne Depth Sounder (LADS) are all consistent with northward movement of bedload sediment (Fig. 6; see also Carter et al., in prep. A). Over time, these strong near-bed cyclone-driven flows of 130–200 + cm/s have resulted in the erosion of the Pleistocene substrate and its exposure at the seabed beneath a widely distributed but patchy middle shelf veneer of muddy calcsand and shell hash. This veneer represents the integrated record of ~10,000 yr of cyclone-influenced sedimentation (Carter et al., in prep. B). Erosion of the top few centimetres of the muddy calcsand veneer during the passage of a cyclone results in unmixing, with mud in suspension being transported along-shelf and advected shorewards (Gagan et al., 1988, 1990), leaving bedload fields of bioclastic sand-ribbons up to about 15 cm thick on the middle shelf (cf.

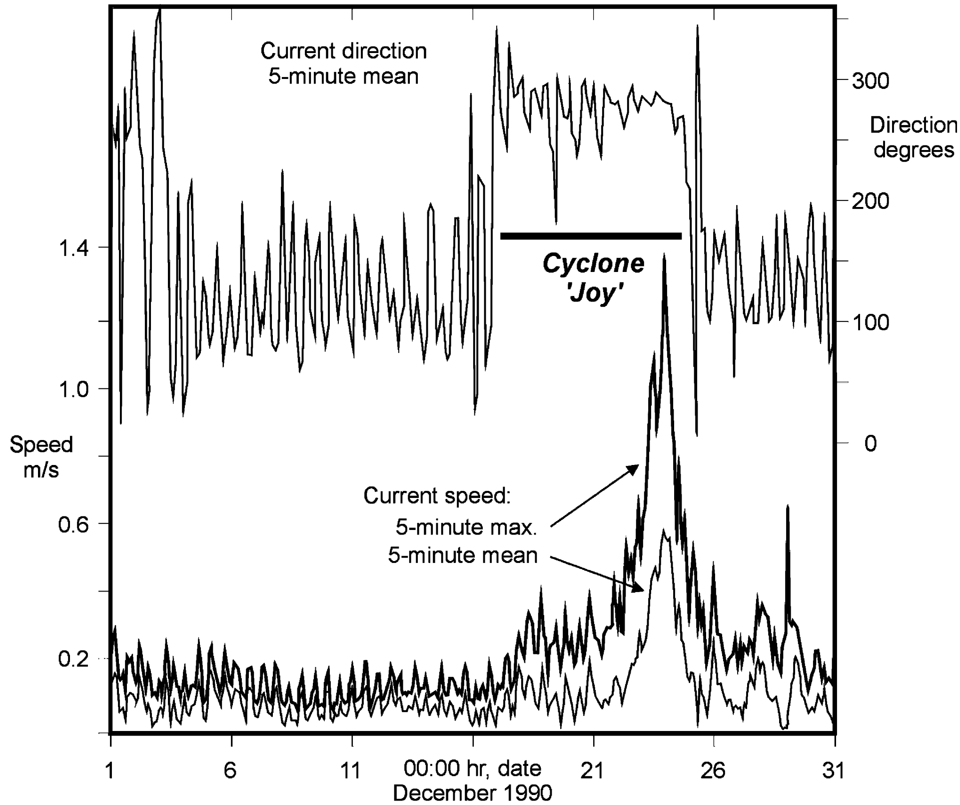


Fig. 4. At 1 month record of current strength and direction measured 1 m above the seabed in 12 m water depth, Trinity Bay, Cairns (data courtesy Cairns Port Authority). Note that the approach and passage of Cyclone Joy (category 3) was associated with a 10-day period of along-shelf currents which peaked at 140 cm/s.

Kenyon, 1970), and—preferentially located near reefs—large and very large dunes (using the notation of Ashley, 1990).

3.3. Cross-shelf transport of terrigenous mud

Though some seaward-spreading surface flood plumes do reach from the mainland out to the reef tract (Devlin et al., 2001), they are extremely dilute (generally <5 mg/l of suspended sediment; cf. Taylor, 1996) and contribute only tiny amounts of suspended sediment to the middle or outer shelf (Fig. 7). Cyclone-induced flood plumes, therefore, do not cause significant transport of terrigenous sediment from the inner to the middle and outer shelf. But what about near-bed transport? The presence of a GBR shelf-wide bottom nepheloid layer (BNL) (e.g. Sahl et al., 1987, for the eastern Texas shelf) has not been observed after cyclones (cf. Carter et al., in prep. A), nor, for that matter, in fair-weather. Instead, carbon isotope and other data for cyclone Winifred (Fig. 8) show that resuspended middle shelf material was moved *shoreward* and *along-shelf*, and that 10–30% of the inner-shelf storm layer was composed of mud derived from the middle shelf (Gagan et al., 1990). Seaward-directed turbid underflows, which might be caused by coastal setup within the broad raised lens of shelf water beneath a cyclone, would rapidly become incorporated into the northward-flowing mass of water along the middle shelf, and therefore probably never penetrate even to the reef tract, let alone beyond that to the shelf edge itself. In keeping with this, sediment samples from the reef tract rarely contain more than a few percent of terrigenous mud (e.g. Orme and Flood, 1978), even in the northern reef sector where the reefs are relatively nearshore (Davies and Hughes, 1983), and acid insoluble sediment residues from the outer reef actually contain more arc-derived volcanic material than they do terrigenous material of Australian continental origin (Okubo and Woolfe, 1995).

These facts notwithstanding, in places the middle shelf veneer of shelly calcsand may contain up to 30% mud, of which 30–70% is terrigenous (Gagan et al., 1988). However, the maximum veneer thickness of $\sim$2 m, much of which comprises bioclastic material, implies a net mud accumulation rate at least an order of magnitude less than that of coastal depocentres. For instance, a 2 m core containing say 60 cm of mud dispersed through 6000 yr of calcsand accumulation would imply a total mud sedimentation rate of 10 cm/kyr, compared with rates up to 100 cm/kyr or more near the coast. However, most of the 18–42 cm (30–70%) non-carbonate portion of this mud is probably derived

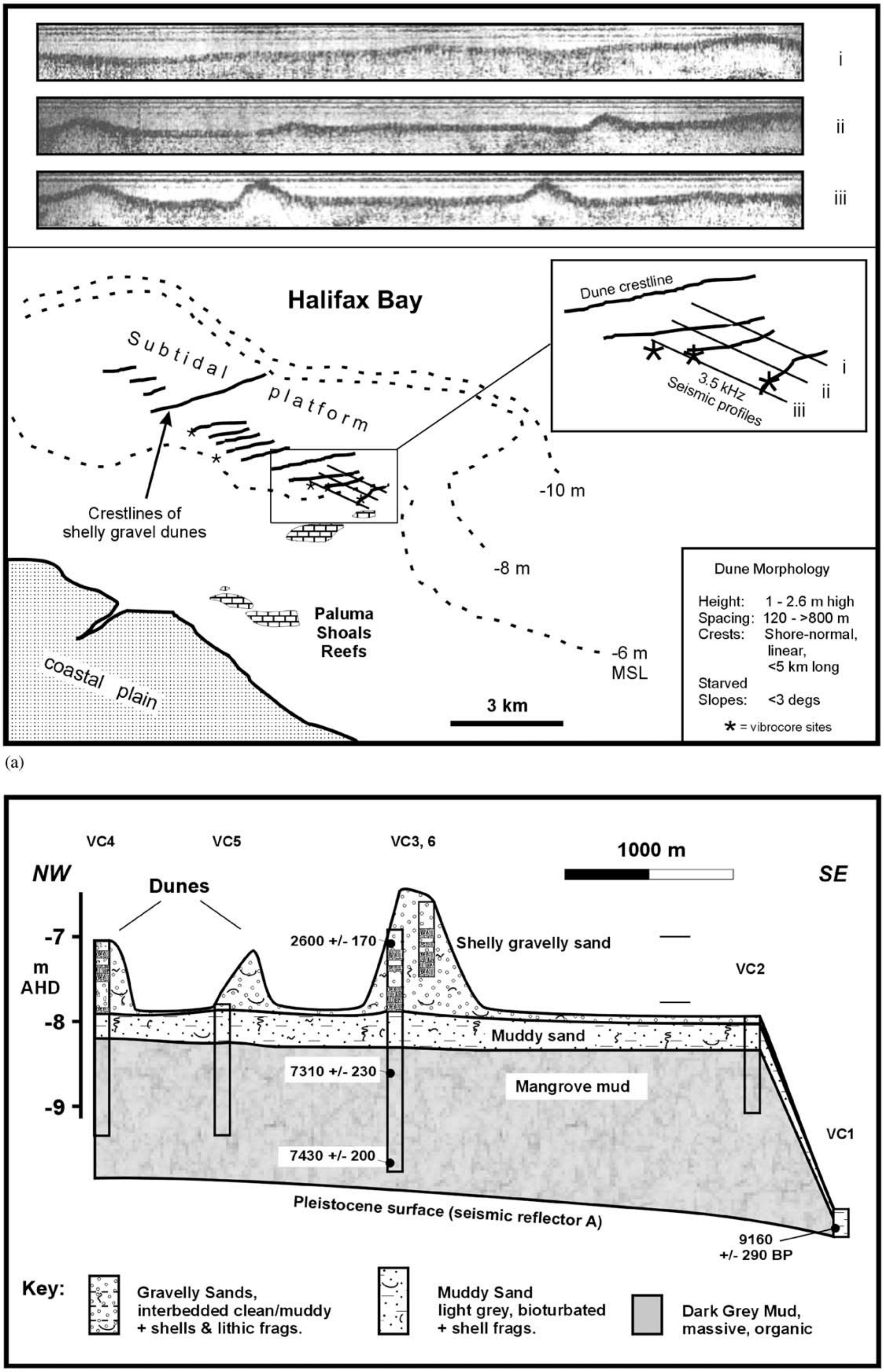


Fig. 5. (a) Map showing the location and orientation of a field of shelly gravel dunes up to 2.5 m high on the inner shelf, Halifax Bay, central GBR shelf; (b) shore-parallel cross-section through the dunes and their underlying palaeo-mangrove substrate, with radiocarbon dates indicated (in ^{14}C years and reservoir-corrected as appropriate and following Larcombe et al., 1995a). Cores have been corrected for estimated compaction.

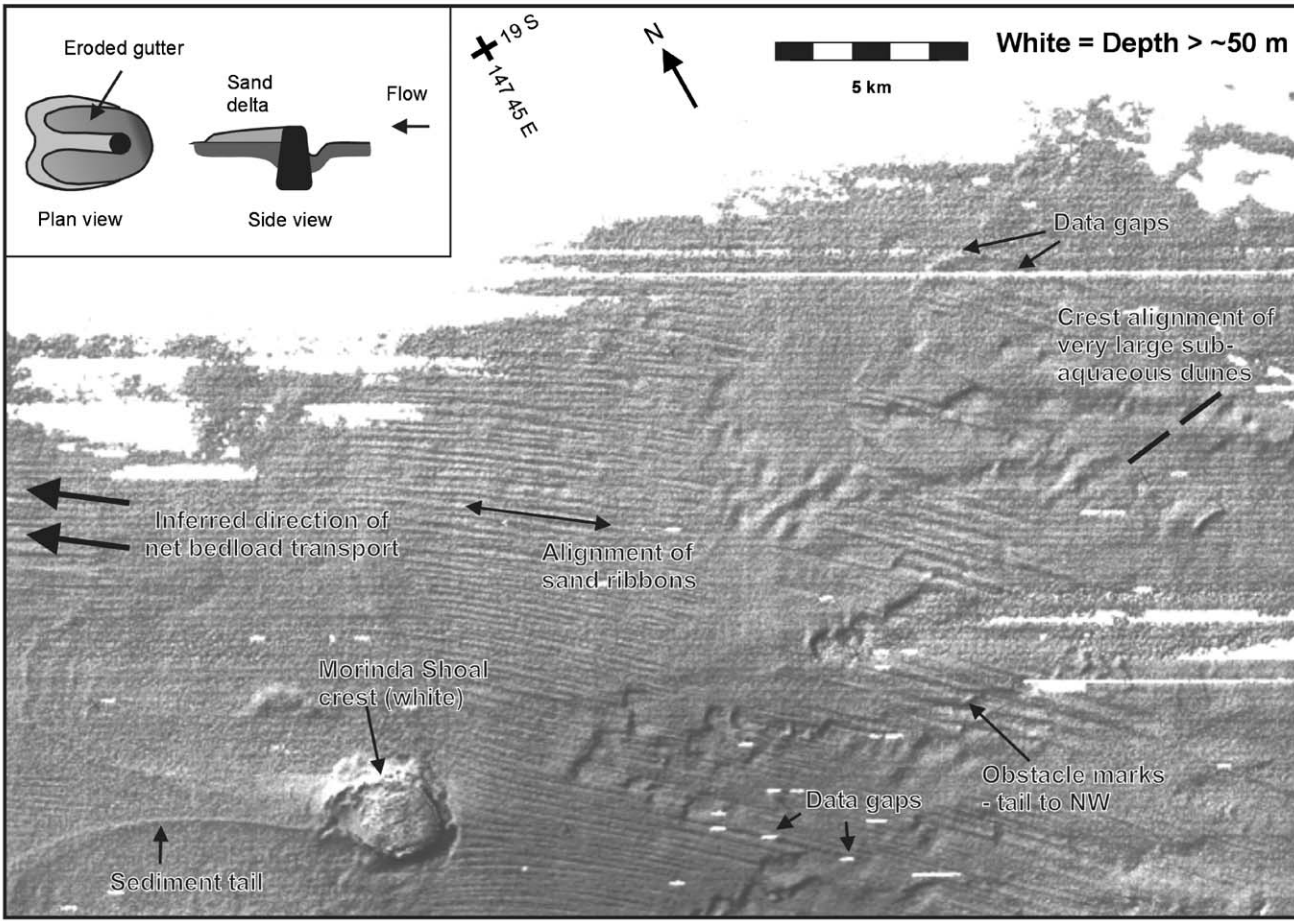


Fig. 6. Seabed morphology as mapped by LADS measurements, in the vicinity of Morinda Shoal, off Cape Bowling Green, central GBR (data courtesy Royal Australian Navy). Note the widespread development of an erosive seabed, with obstacle marks and sediment ribbons. Note also the upcurrent moat and downcurrent sediment delta ("crag and tail" effect) associated with Morinda Shoal. Inset: depiction of crag and tail effect on an erosive seabed (after Allen, 1984), which compares well with Morinda Shoal and indicates that long-term bedload transport is to the NW.

by erosion of the Pleistocene clay substrate within the cyclone corridor, with only a minor contribution from coastal river plumes (Gagan et al., 1987, 1988, see also Fig. 8); accordingly, the land-sourced mud contribution to the middle shelf is probably only about 1 cm/kyr. We conclude that the overwhelming majority of land-sourced terrigenous mud which has been contributed to the GBR shelf during the Holocene is sequestered within the ISP. One unsurprising consequence of this is that turbid-zone fringing reefs developed at the coast or on the inner shelf (Woolfe and Larcombe, 1998; Larcombe and Woolfe, 1999; Larcombe et al., 2001) usually exhibit a reef matrix that contains >50% terrigenous mud (Orr and Moorhouse, 1933; Johnson and Carter, 1987; Johnson and Risk, 1987; Smithers and Larcombe, in press).

3.4. GBR outer shelf reef complex

A 1-month record of waves and currents at 35 m depth recorded at the outer edge of the northern GBR reef tract near the Pandora wreck-site indicates that fair-weather sediment transport here is controlled by tidal and other unidirectional currents, which at up to 100 cm/s are strong enough to form small dunes in biogenic sand (Ward et al., 1999). The south-flowing East Australian Current, which attains surface velocities between 30 and 100 cm/s (Church, 1987), also commonly impinges on the GBR outer shelf (e.g. Burrage et al., 1996). Although the characteristics of the shelf outside of the central GBR reef tract remain extremely poorly known, regional high-resolution seismic data (Dye, 2001) show that the outer shelf has an almost flat surface, punctuated by occasional drowned reef pedestals of unknown age (Harris and Davies, 1989), and largely devoid of significant thicknesses of postglacial sediment (cf. Scoffin and Tudhope, 1985; Harris et al., 1990). It is also noteworthy that sediments with up to 20% of relict grains are common in inter-reef passages of the central GBR outer shelf. This high proportion of relict material was interpreted by Scoffin and Tudhope (1985) as resulting from a combination of biological processes and shrimp trawling in areas close to the occurrence of past river channels. Our data now indicate

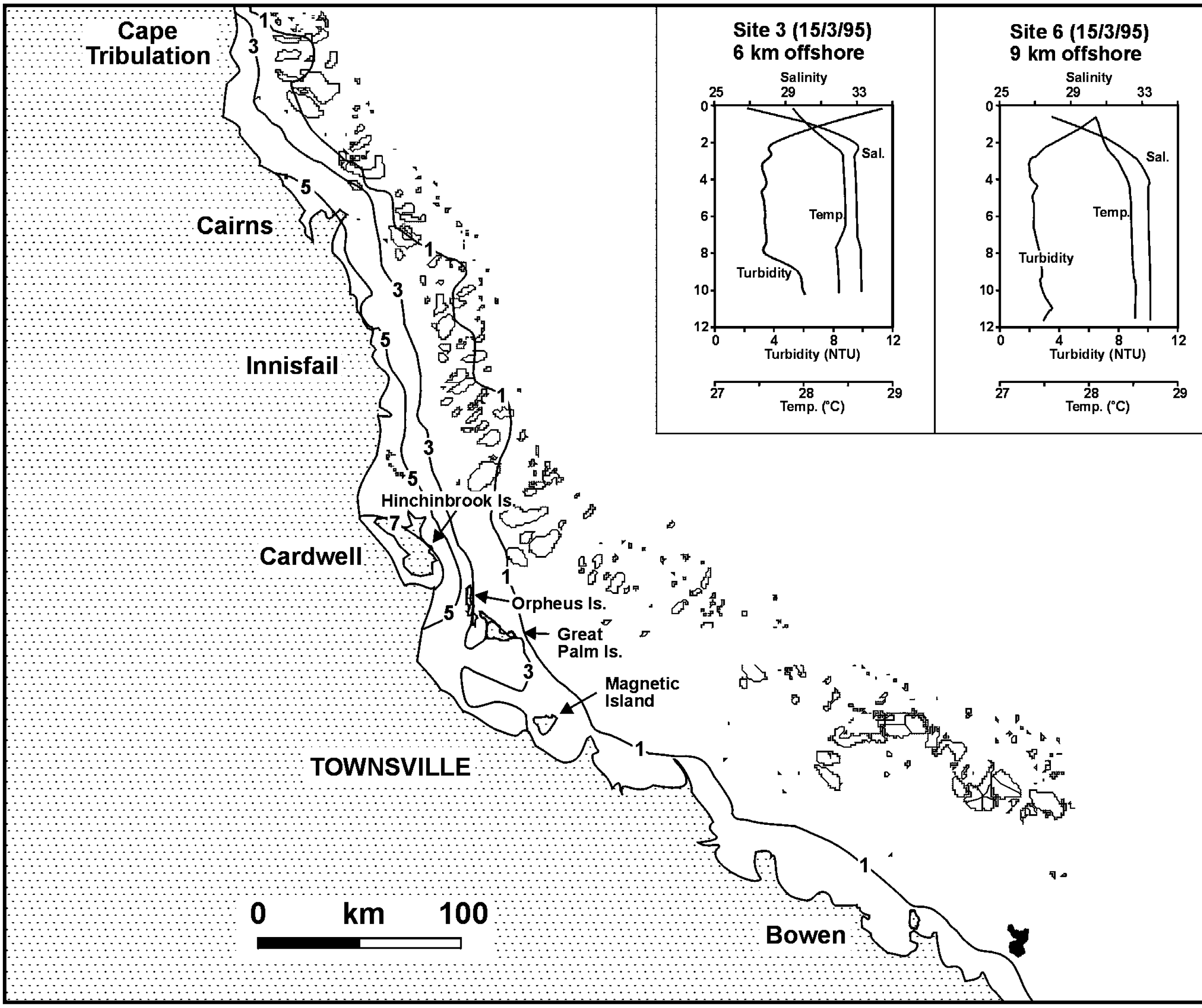


Fig. 7. Simplified contours of location of surface river flood plumes generated on the central GBR shelf by post-cyclone river flooding. Contours are the number of plumes during the period 1991–2000 (redrawn after Devlin et al., 2001). Note that plumes do not reach the reef tract at all over most of the central region, and that even in the Cairns region, where the lagoon narrows, only the dilute seaward edge of the largest plumes reach the reef tract. Inset: water turbidity (NTU) and salinity content of Barron River flood plume, 6 km off the river mouth (after Taylor, 1996). Note the maximum turbidities of 10 NTU (~10 mg/l).

that storm-associated erosion of the seabed in inter-reef passages is probably a major source for the reworked sedimentary grains.

3.5. Storm-driven currents on other shelves

High wind-driven current velocities in an along-shelf direction have been recorded during storms on shelves worldwide. For instance, during the passage of category 3 hurricane Delia (1968) in the Gulf of Mexico, wind-driven along-shelf currents at depths of 18 m peaked at 160 cm/s, and are estimated to have exceeded 200 cm/s for more intense storms such as Carla and Camille (Camille, 1969, category 5; Forristall et al., 1977). In the North Sea, Gienapp (1973) analysed a 1965 winter storm that affected the German Bight. A combination of tidal and storm-forcing resulted in current velocities 2 m above the seabed of up to 150 cm/s, and in some areas an alongshore single-layer current greater than 100 cm/s lasted for more than 6 h. Of course, at the seabed such unidirectional, wind-forced storm flows act in concert with the oscillatory motion driven by storm waves (Grant and Masden, 1979; Cacchione and Drake, 1990). This can cause a shear stress in the current direction that is three to 10 times greater than that exercised by the current alone (Silvester, 1974), which in turn allows the suspension and transport of very coarse grained bedload particles. In the Gulf of Mexico, hurricane Diana (1984, category 3) generated waves 6 m in height and 10 s in period, implying orbital seafloor velocities of 125 cm/s in water depths of 25 m (Mearns et al., 1988). Morton (1988) estimated that waves caused by hurricane Camille

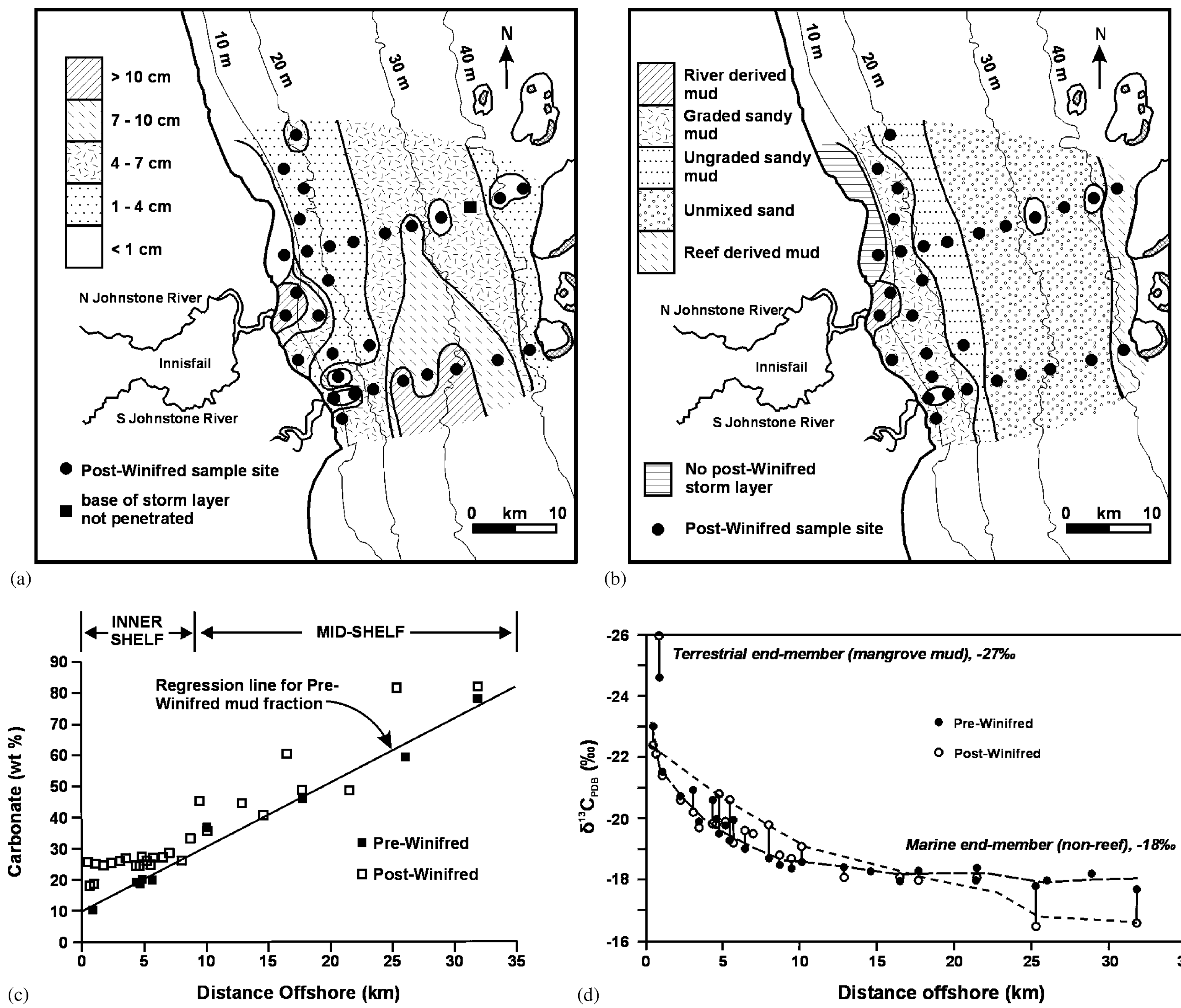


Fig. 8. Sedimentary effects of Cyclone Winifred (after Gagan et al., 1987, 1990) (a) thickness and distribution of storm bed; (b) surficial sedimentary facies of storm bed; (c) shore-normal transect, comparing pre-and post-cyclone carbonate percentage of mud fraction; note the enhanced carbonate concentration in post-cyclone samples, consistent with onshore transport of mud from the mid-shelf; (d) shore-normal transect, comparing pre- and post-cyclone carbon isotope composition of the organic carbon of the <2 mm sediment fraction; note the lighter post-cyclone values within 10 km of the coast, little change across the middle shelf, and heavier post-cyclone values near the reef tract, consistent with limited terrestrial input to the inner shelf, and off-reef transport near the reef tract.

would have created orbital velocities up to 500 cm/s in 20 m of water and 300 cm/s in 45 m of water, and Hubbard (1992) reported sand-erosive seabed oscillatory flows up to 400 cm/s during the passage of hurricane Hugo (1989), St. Croix. Theoretical modelling is consistent with these observations and calculations. Keen and Slingerland (1993) used a numerical model to hindcast four historical hurricanes in the Gulf of Mexico and concluded that a common response to storm passage was "along-shelf flow (which) transports finer sediment in deep water and coarser sediment in shallow water". Further, in the geostrophic model they used, the depth (thickness) of the friction-dominated, wind-mixed layer exceeded the water depth everywhere over the shelf, causing the upper and lower boundary layers to overlap and thereby forcing vertically uniform along-shelf flow.

We conclude that vertically uniform, unidirectional, along-shelf, wind-driven currents commonly attain sustained velocities greater than 100 cm/s during severe storms, and intermittently burst as fast as 300 cm/s or more. These velocities are adequate to entrain and transport grains up to granular and pebbly sand in size, and to drive the development of characteristic bedforms such as sand ribbons (cf. Belderson et al., 1982), and small ("megaripples") and large sand dunes (Ashley, 1990). Some of these bedforms are generally most conspicuous in sediment-starved locations. For example,

middle shelf seafloor with strong along-shelf fabric (sand ribbons, dunes, etc.), located outside and sub-parallel to a terrigenous inshore sediment prism, occurs along the Agulhas shelf, South Africa (Flemming, 1978, 1980; Ramsay, 1994), the northeastern Brazilian shelf (Vianna et al., 1991; Testa and Bosence, 1999), and the Otago (Carter et al., 1985) and Coromandel shelves (Bradshaw et al., 1994) in eastern New Zealand, as well as on the GBR shelf as described here. In all these cases, the major bedforms and sediments are developed in an environment which is relatively starved of terrigenous sediment (and is therefore carbonate-enriched in relative terms), and are a response to oceanic or wind-driven along-shelf currents. Generalisations which suggest that "the midshelf region is recognized worldwide as the location where modern sediment accumulates, usually muddy" (Nittrouer and Wright, 1994), are only true for shelves which are characterised by high terrigenous input, and on a worldwide scale such shelves are probably in the minority.

3.6. Preservation of storm-induced bedforms

Before-and-after sidescan sonar surveys of hurricane Diana on the north Carolina shelf (Mearns et al., 1988) and Cyclone Winifred on the GBR shelf (Gagan et al., 1988; Carter et al., in prep. A) revealed, perhaps surprisingly in view of the category 3 intensity of the storms, little discernible change in the respective major patterns of seabed facies distribution. Nevertheless, at a smaller scale, the immediate post-Winifred survey delineated shelf-parallel sand ribbons and small (1.6 m wavelength) dunes which were sharply imaged compared with their much more subdued and "fuzzier" pre-Winifred counterparts. The surveys also revealed that degradation by bioturbation of the Winifred bedforms themselves commenced on the middle shelf within 3 months of the storm, and was advanced at 12 months post-storm, by which time the shallower shelf seabed will also have been modified by the development of small-scale bedforms of fair-weather wave, wind-driven and tidal current origin. Similarly, rapid seabed modification within 3 months post-storm occurred for the hurricane Alicia storm-bed in the Gulf of Mexico (Morton, 1988).

Clearly, which bedforms we observe and describe from the seabed on storm-influenced shelves depends to a significant degree on the time which has elapsed since the occurrence of the last major storm. With the passage of a storm, increasing then decreasing seabed shear stresses will produce a time-varying set of bedforms, as recently captured by Hume et al. (2000) in 1997 for tropical Cyclone Gavin in northern New Zealand waters. Gavin was accompanied by winds up to 220 km/h and waves up to 5.5 m high; sustained wave heights exceeded 3 m height and 11 s period, implying seafloor velocities which peaked at over 100 cm/s in water depths of 25 m. During the passage of Gavin, an instrumented tripod (ALICE-IMAGINEX) recorded changing bedforms on a sandy seabed at 25 m depth. Small wave-orbital ripples (5 cm high, 40 cm wavelength, 70 cm crest length) generated early in the storm were replaced by steeper-sided small dunes (15 cm high, 80–100 cm wavelength, 15 m crest length) at the height of the storm. The degree to which such bedforms survive in the months and years after the storm is a function of the waning processes of the storm, the subsequent rate of bioturbation, and the degree to which the seabed is later remobilised by fair-weather processes, or buried by younger sediment.

3.7. Summary

Away from the strong tidal flows which occur locally nearshore or near reef passages, coarse sand and gravel on all parts of the inner and middle GBR shelf are emplaced and moved mainly by north-directed along-shelf flows created during the passage of cyclones. These conclusions are based on real time observations of the GBR shelf, on before-and-after studies of particular storms, and on reported observations from shelves worldwide which are characterised by a limited sediment supply. Care must be exercised in inferring storm transport motions using small-scale bedforms preserved at the modern seabed, because such bedforms often represent waning-flow or fair-weather, rather than peak storm, conditions.

4. The highstand cyclone-pump

The modern GBR middle shelf, located between depths of 20–40 m, is a low-gradient erosional surface cut onto Pleistocene clay, with a local bed armour of muddy calcsand or shell hash. The middle shelf comprises a corridor through which fast along-shelf currents move during and after the passage of a cyclone. Direct measurements during Cyclone Joy at the inner edge of this corridor show near-bed currents of up to 140 cm/s at 12 m depth (cf. Fig. 4), and sand-ribbons, crag-and-tail bedforms and dunes are observed by side-scan sonar and LADS at depths up to 25–50 m across the corridor (cf. Fig. 6). Further, a clear divergence in sand ribbon crests occurs near the southeastern margin of the reef-capped Morinda Shoal, a scoured moat is located on its inferred up-current side, and a cuspate shadow sand-delta occurs to leeward (cf. Allen, 1984, Vol. 2, Figs. 5–10). Together, these facts are consistent with seabed erosion and with a regional and long-term net northward movement of bedload sediment, driven by storm events. Henceforward, we use the term 'cyclone corridor' to refer to the mainly non-depositional, current-swept area of seabed in modern depths of

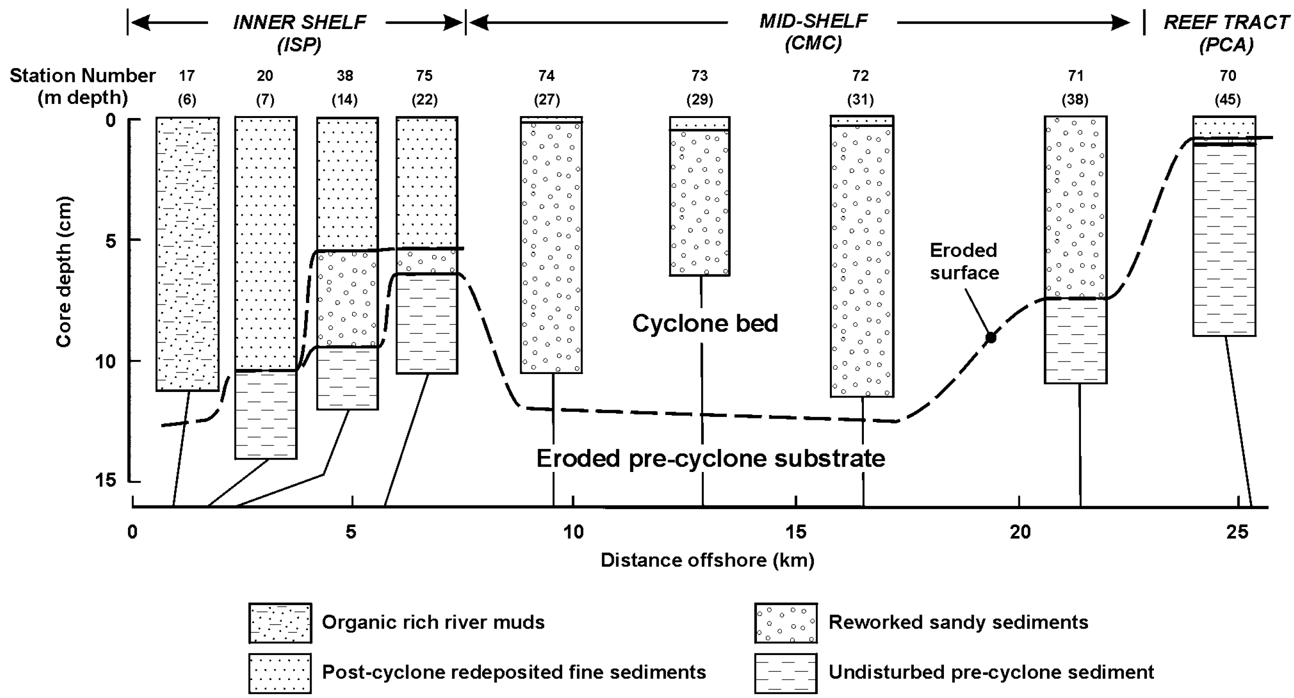


Fig. 9. Cross-shelf transect of short cores collected after Cyclone Winifred, showing the sedimentary composition of the storm bed. Note the seaward thinning trend of the inner-shelf storm bed which, like the ISP itself, is restricted to within about 8 km of the coast, and the thin to absent post-cyclone mud drape over the middle shelf mixed calcsand bed.

20–50 m, and which is located between the ISP and the outer-shelf reef carbonate complex. The corridor therefore corresponds to the outer part of the GBR lagoon (e.g. Maxwell, 1968; Hopley, 1984), and closely approximates to the middle shelf as defined on sedimentary grounds by Belperio (1978, 1988).

The main cyclone corridor is delimited seawards by the inner edge of the reef tract, and landwards by the seaward slope of the ISP. Each of these boundaries represents an important energy fence, across which fine-grained sediment (especially) is advected, to settle out later. Thus, significant deposits of micrite occur in sheltered locales between outer-shelf reefs off Innisfail (J. Dye, pers. comm.), and *Halimeda* banks in a similar location contain up to 25% terrigenous mud (Davies and Marshall, 1985), much or all of which may derive from storm erosion of the middle shelf seabed rather than from shoreline erosion or coastal flood plumes. The terrigenous ISP typically contains 10–20% carbonate, in the form of shells and micrite (Aliano, 1978; Gagan et al., 1990), rising to 25% or more at the seabed immediately after passage of a cyclone. We infer that most of the fine-grained carbonate has been advected landwards onto the ISP from the middle shelf, after cyclones had first (i) dislodged sediment from the reef tract onto the middle shelf, and/or (ii) eroded carbonate nodules and the shell veneer along the Pleistocene middle shelf substrate (cf. Fig. 8). When a tropical cyclone makes landfall, it typically degenerates into a rain depression which causes strong river flooding. Such events are responsible for most of the ~4–5 Mt of terrigenous sediment which is estimated to have entered GBR coastal waters annually prior to European settlement (Moss et al., 1993; Neil and Yu, 1996; Neil et al., 2002). Thereafter, during fair-weather, bedload sand is reworked north along the beaches and shoreface by coastal drift, whilst mud plumes from the river mouths are generally spread northwards, and advected shorewards, under the influence of SE trade winds (Wolanski and van Senden, 1983; King et al., 2001).

Cyclones therefore act to contribute sediment to the shelf system in three discrete ways:

- by direct breakage of reef material during cyclone passage across the outer shelf reef tract, by which means carbonate gravel and sand is added to the peri-reef sediment aprons; such damage is generally patchy rather than regionally devastating (e.g. Hubbard, 1992; Done, 1992);
- by erosion of Pleistocene clay, and breakage of shells and other biogenic material, at the seabed within the middle shelf cyclone corridor; and
- by rainfall-induced input of terrigenous detritus at river-mouth point-sources.

In effect, cyclonic activity on tropical, reefed shelves acts as a *sediment pump*. Cyclones control the production of all three major types of sediment detritus, which

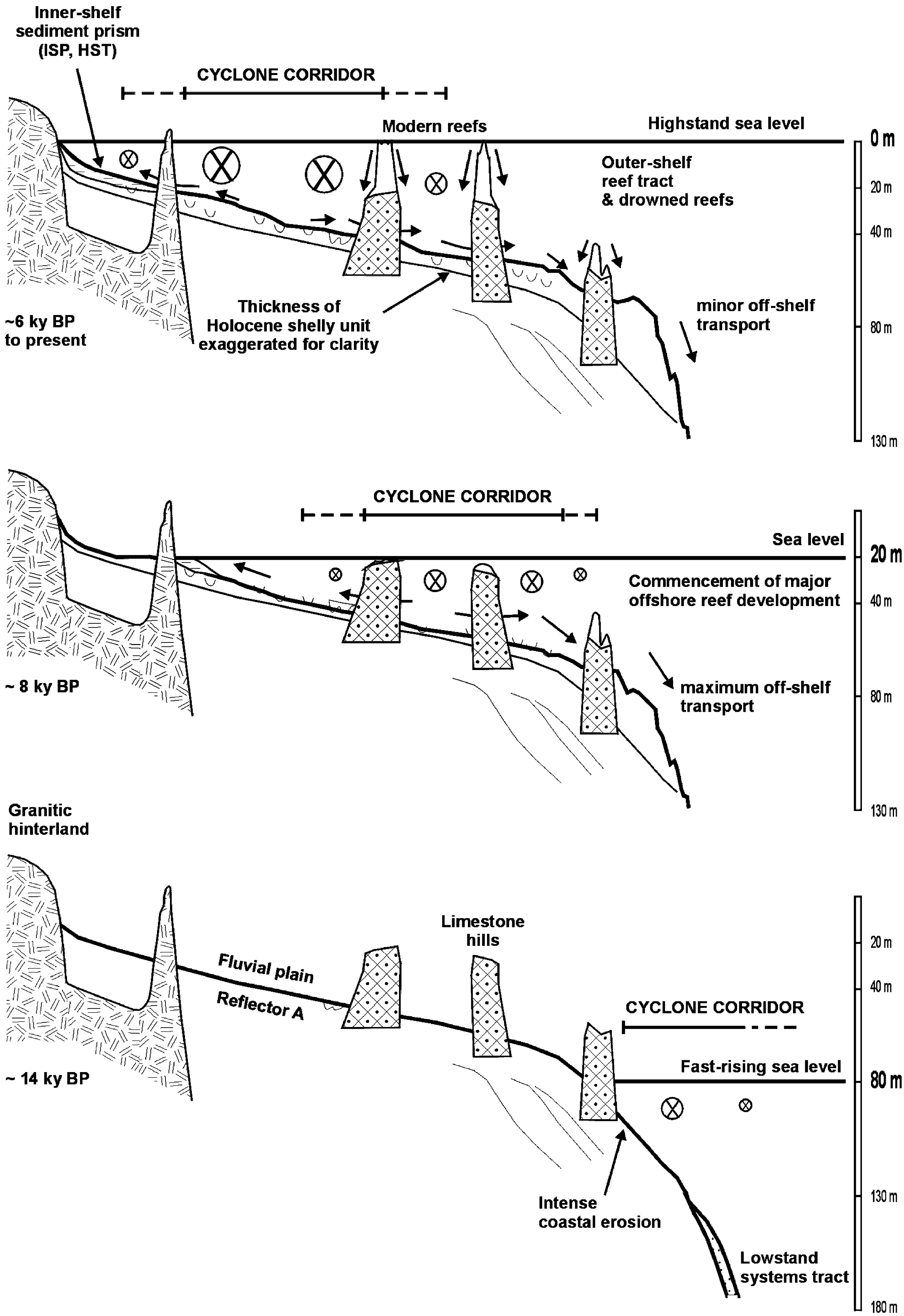


Fig. 10. The cyclone corridor at various stages of the postglacial sea-level rise, based on a scaled cross-section of the shelf off Halifax Bay, Townsville region. The lower, middle and upper parts illustrate different stages of the postglacial transgression and the shifting location of the 'cyclone corridor'. See text for full explanation. Crossed circles denote along-shelf currents flowing into the page and larger circles indicate faster currents.

is then redistributed by gravity (peri-reef apron), flood plumes (river jetting), and strong along-shelf flow (wind forcing) within the cyclone corridor. Constrictions, such as that which occurs between the projecting Cape Grafton and the reef tract off Cairns (cf. Fig. 1a), cause acceleration of the flow and probably also act as valves, through which bedload sediment is passed northward, and displaced laterally, but back through which little sediment will return (cf. Flemming, 1981, on the Agulhas current-swept east African shelf). Available current meter data (Wolanski and Ridd, 1990, this study, Ridd and Larcombe unpublished data; see also the aerial study of Devlin et al., 2001 and the modelling of King et al., 2001) show no evidence of southerly directed cyclonic flow, either forced directly by wind or occurring as a relaxation flow following landfall of the cyclone, and all seabed sedimentary data indicate only bedload transport to the northwest. A cyclone-pump sedimentation model is consistent with:

- cyclones being by far the highest energy input at the bed of the middle shelf, and to the shelf generally;
- the eroded, flat, middle shelf seabed of low gradient (1:1000), which incidentally then permits only weak offshore sediment flux;
- strong sediment partitioning, whereby most terrigenous sediment is sequestered into the ISP, in a totally distinct belt to the high-carbonate sediments of the offshore reef tract; which explains
- the absence of terrigenous highstand deposits in the middle shelf cyclone corridor and on the outer shelf.

4.1. Summary

Cyclone pumping probably exerts a strong controlling effect on sediment facies distribution on many tropical 'mixed' carbonate-terrigenous systems, of which the GBR shelf forms but one example. The nature and disposition of similar facies belts on other tropical shelves will vary from place to place according to the hemisphere, the geographic orientation of the particular shelf and coastline, and the nature and history of local sediment sources.

5. The stratigraphic record of GBR storm deposition

Near to their source, episodic high-energy sedimentary agents such as storms and turbidity currents are characteristically erosive in nature. At one extreme, therefore, the sedimentary record of a high-energy event may comprise only a bedding-plane diastem or an erosive disconformity. More usefully, and especially where significant sediment transport has occurred away from the source, a more or less complete stratigraphic record of the event may be preserved. Thus, a major continental slope slump is represented at distant locations by a characteristic turbidite (e.g. Elmore et al., 1979), or an episodic storm by a graded storm bed (e.g. Gagan et al., 1988). In these and most similar circumstances, the event bed records the decreasing sediment transport power which accompanied the waning of the energy associated with the actual event. A fining-upward, graded sedimentary unit is therefore the almost universal characteristic of event beds which have been water-lain.

Considered as a sedimentary agent on the central GBR shelf, cyclones conform to these generalisations. Cyclone-induced seabed erosion is undoubtedly a dominant process along the 20–30 km-wide, middle shelf cyclone corridor. At the same time, many cyclones which pass across the GBR shelf do leave a sedimentary record of their passage, in one of the following three main ways.

5.1. Inner-shelf prism

In cores, the sediments which make up the terrigenous prism on the GBR inner shelf comprise strongly bioturbated, poorly sorted, muddy sand and sandy mud with occasional molluscan shells (Carter et al., 1993; Woolfe et al., 2000). Sedimentary bedding is rarely preserved, and grain-size segregation is represented mainly by burrow-fills and clots of moderately sorted fine to medium grained sand, or by lenses of shell fragments. The sediment is conspicuously polymodal, with sand, silt and very fine-grained silt modes. In contrast to this, Gagan et al. (1988) described an immediate post-Winifred storm bed from the inner shelf which comprised a moderately well sorted, graded bed of terrigenous sand-mud, locally with a basal shell lag, and 5–20 cm in thickness (Figs. 8a and b). Twelve months after the cyclone, this storm bed had been homogenized within the substrate by bioturbation. Thus, we infer that ancient storm beds within the ISP are represented mainly by the polymodal nature of its muddy-sandy sediment (cf. Orpin and Woolfe, 1999), and by scattered lenses of shells and sand. Graded sand-mud beds, such as the cyclone Winifred storm bed, will be preserved only if they are either thick enough on formation, or are buried fast enough beneath later sediment, to preclude their bioturbation. Graded storm beds are therefore only likely to be preserved where rapid post-storm burial occurs, perhaps at locations immediately adjacent to a terrigenous sediment source such as a river-mouth flood jet.

Isotope measurements on organic carbon from pre- and post-Winifred seabed samples in the vicinity of the Johnstone River (Gagan et al., 1987) indicate that the isotopically light terrigenous sediment borne in the river mouth jet was deposited over a seaward-broadening fan-shaped drape beneath the plume, and within 12 km of

the shoreline, i.e. entirely on the ISP (Fig. 8d). At the same time, isotopically heavier carbon derived from the middle shelf was moved shorewards to be deposited widely across the ISP in locations north and south of the river mouth flood jet. The movement of muddy sediment from offshore onto the terrigenous prism was also indicated independently by an estimated systematic 3–15% increase in the proportion of carbonate mud in the mud drape on the ISP, deduced from seabed samples collected before and after the passage of the cyclone (Gagan et al., 1990) (Fig. 8c).

We conclude that much of the sediment within the Holocene ISP has been *emplaced* onto the GBR inner shelf from both sides, during storm (and flood) events; most of the sand and mud is derived from river flooding, but up to 30% of the total mud present is emplaced by advection from offshore, after seabed erosion within the middle shelf cyclone corridor during each storm. Between storms, bioturbation and fair-weather physical processes act to destroy a storm bed by homogenising it within the substrate, thus the ISP generally contains no easily interpretable record of the passage of individual storms.

The Winifred storm bed averaged 10 cm thick over the nearshore part of the inner shelf, and the ISP averages ~4 m in thickness near to riverine sources. Thus, the prism is the stratigraphic equivalent of about 40 category 3 storm events. Over the ~5500 yr which have elapsed since the mid-Holocene sea-level high, and assuming that every storm left an equal record, this would imply a recurrence interval for category 3 storms of 140 yr. This estimate is similar to the 177–280 and 120–150 yr estimated between major cyclones by Nott and Hayne (2001) and Carter et al. (2002), based respectively on measurements of successive storm-emplaced chenier ridges and foreshore shellbeds. In contrast, a 50 yr recurrence interval is indicated for category 3 cyclones by meteorological hindcasts (Walker and Reardon, 1986). The meteorological and geological figures can, however, be brought into agreement by assuming that category 3 and greater cyclones exercise both erosion and deposition, which in combination have eroded and advected an average of 1.2 m of inner shelf seabed sediment every 1000 yr. Though these estimates are inevitably generalized, they do give an internally consistent indication of the general way in which the mid-Holocene and younger ISP has developed.

5.2. Middle shelf condensed shell hash

The middle shelf, between 20 and 40 m water depth, broadly corresponds to the modern location of the main cyclone corridor, along which strong seabed-erosive flow and sediment transport occurs during the passage of each storm. After Cyclone Winifred, a 10–15 cm thick, graded sand-mud storm bed was deposited widely over the middle shelf off Innisfail (Gagan et al., 1988) (Figs. 8a, b and 9). The storm bed sand is well sorted, has a mixed terrigenous-bioclastic composition, and is inferred to have been derived from erosive unmixing from the seabed. As for its terrigenous counterpart on the inner shelf, the upper (muddier) part of the middle shelf Winifred storm bed was already well bioturbated within 12 months of the occurrence of the cyclone.

Away from the few infilled palaeo-channels which traverse the GBR shelf, the veneer of mixed terrigenous-carbonate sediment within the middle shelf corridor rarely exceeds 1–2 m in thickness, and is often represented by only a shell-rich armour. The veneer comprises extremely poorly sorted, structureless, muddy, sandy shell gravel, which rests on weathered Pleistocene clay (Carter et al., in prep. B). However, and even when it is bioturbated throughout, radiocarbon dates show that the condensed shell hash facies retains a coherent internal stratigraphy (Larcombe and Carter, 1998). In keeping with this, some rare cores have survived pervasive bioturbation and contain up to seven or more 15–20 cm thick, graded storm bed cycles (Gore, 2001). Each such cycle is similar to the storm bed deposited during and after Cyclone Winifred, but is distinguished by being thicker overall and by having a bioturbated muddy sand (instead of an intact mud drape) as its upper part. We infer that these beds have been preserved from complete bioturbation or erosion by their thickness and particular seabed location. For instance, a series of five radiocarbon dates from core V9 near Cairns, located in 40 m of water and just landwards of the Green Island cay (Carter et al., in prep. A), show (i) a gap of ~7.7 kyr between the marine flooding of R^A (Reflector A of Johnson and Searle, 1984) and deposition of the oldest (deepest) storm bed at ~1800 cal yr BP (during which time, was exposed, or intermittently covered and then re-exposed, at the seafloor); and (ii) deposition of several succeeding storm beds at an average interval of 360 yr. This recurrence interval is the same as that predicted for category 4 or greater cyclones at Cairns from meteorological records (Massell and Done, 1993; see Fig. 3b).

5.2.1. Storm bed models

Gagan et al. (1988, 1990) showed that the Cyclone Winifred storm bed illustrates the inverse of the commonly accepted storm sedimentation model, which predicts sediment fining and thinning away from the shoreline (e.g. Wright and Walker, 1981; Allen, 1984; Seilacher and Aigner, 1991). Considering the inner and middle shelf storm beds as a single entity, the Winifred bed becomes markedly *coarser*-grained, and *thickens* where it passes from the inner to the middle shelf, i.e. away from the shoreline (cf. Figs. 8 and 9). Post-cyclone observations also show that the storm bed was only

preserved close inshore, where immediate post-storm sedimentation buried it below the depth of bioturbation.

5.3. Island-perimeter or mainland chenier ridges

The third type of sedimentary deposit produced by cyclones is a supratidal storm ridge, as a sand- or shell-based chenier. The elevation of the sea surface above that produced by tides alone is a combination of the barometric effect, wind-set up and wave set-up (Pond and Pickard, 1983; Allen, 1997), and the resultant "storm surge" varies from up to 4 m for a category 3 cyclone to as much as 7 m for an extreme category 5 cyclone (Table 1). In combination with a high tide, storm surges can cause widespread flooding of the coastal plain, during which the associated coastal waves may erode shoreface sediment and emplace it in a new chenier ridge located landward of the beach. Historic observations exist of cheniers forming during individual storms (e.g. Maragos et al., 1973; Scoffin, 1993), but local factors—such as a coincidence between cyclone landfall and low tide, or a low availability of coarse sediment—often preclude chenier formation during the passage of particular, even large, cyclones (e.g. Woodroffe and Grime, 1999). Much of our knowledge regarding storm cheniers is therefore inferential, and drawn from geomorphic or geologic study.

It is unusual for chenier ridges to exhibit strong vertical stratigraphic superposition. Rather, they occur as a series of low, laterally adjacent, seaward-prograding and younging, shore-parallel ridges along a coastal plain or around an island or cay perimeter. Studies of chenier on the GBR coastline (Bird, 1972; Chappell et al., 1983; Jones, 1985; Chivas et al., 1986) demonstrate that they have been emplaced over the last few thousand years, with an average return interval which varies across sites between about 150 and 300 yr. In a recent and comprehensive study, Nott and Hayne (2001) showed that 22 chenier on Curacao Island in the central GBR encompass the entire period since the mid-Holocene sea-level high at ~5.5 ka, with an average return interval of 280 yr—for a similar series of cheniers on the mainland coastal plain at Princess Charlotte Bay, the return interval was 177 yr. Using a modelling approach, Nott and Hayne (2001) concluded from these data that each chenier was emplaced by a Category 5 "supercyclone", of central pressure <920 hPa. Meteorological hindcasting suggests a broad range of return intervals for category 5 cyclones in the central GBR of 200–1000 yr (Walker and Reardon, 1986; Massell and Done, 1993). An alternative possibility, therefore, is that the Curacao Island cheniers were emplaced by cyclones of varying size (including some less intense storms with a shorter recurrence interval), which shared as their common characteristic not intensity 5 status but rather a landfall which coincided with high tide. It should also be noted that Nott and Hayne's (2001) calculations of cyclone conditions for the mid-Holocene cheniers apparently did not take into account the higher sea level (up to 1–1.5 m) at the time of their formation (Chappell et al., 1983; Beaman et al., 1994; Larcombe et al., 1995a), which would have led to an overestimate in calculated cyclone intensity. Given the range of chenier periodicity which has been observed in the GBR province, and allowing for the fact that the annual number of historic cyclones is a function of latitude (peaking at 20–22°S, Fig. 3), it is thus probable that some GBR cheniers represent the passage of cyclones with an intensity as low as 3 on the Saffir-Simpson scale.

5.3.1. Summary

A varied stratigraphic record exists of the passage of cyclones across the GBR inner and middle shelf, but the spatial and temporal coverage provided is patchy and spasmodic, and mostly only available for the period since ~5.5 kyr BP. Nonetheless, stratigraphic information indicates an average return interval for major cyclones (intensity 3 or higher) of between 150 and 300 yr from shelf core and chenier data. The generally erosive nature of the middle shelf cyclone corridor is

Table 1
Characteristics of typical north Queensland tropical cyclones, according to the Saffir–Simpson scale (after Walker and Reardon, 1986 and other sources)

Saffir–Simpson Scale	Saffir–Simpson Magnitude	Return interval (N. Queensland) Years	Pressure (hPa)	Max. wind gust Knots	Max. wind gust (m/s)	Wind-driven current (cm/s)	Surge (m)	Wave, H_s (m)	Wave, T_s (s)	Wave, L (m)	U_{max} (cm/s) (deep)	U_{max} (cm/s) (shallow)
1	Mild	5	>990	40–60	20–30	40–60	0.0–1.0	8.3	11.2	193	204	75
2	Moderate	10	970–985	70–90	35–45	70–90	1.5–2.5	8.5	11.3	197	212	77
3	Severe	50	950–965	100–120	50–60	100–110	3.0–4.0	9.0	11.6	207	234	81
3	Severe	100	945–950	120–130	60–65	120–130	4.0–4.5	9.2	11.8	214	245	83
4	Very severe	500	930–945	130–150	65–75	130–150	4.5–5.5	9.7	12.1	225	269	88
5	Catastrophic	1000	<925	160–180	80–90	160–180	6.0–7.0	10.0	12.2	229	280	90

3—Althea, Winifred; 4—Tracy; 5 = Camille.
Listings indicate the numerical magnitudes of some named cyclones.

well indicated by the longer average interval of ~600 yr, which occurs between cyclone beds emplaced there. Somewhat earlier cyclonic activity (~9–5.5 ka) may be represented by the stratigraphic gap which occurs in many cores between the time of flooding of the top-Pleistocene unconformity and the oldest overlying Holocene sediment. This includes the basal part of cores through the reef tract, many of which comprise fragmented coral detritus rather than reef framework material (e.g. Davies and Hopley, 1983; Kleypas and Hopley, 1992).

6. Implications of the cyclone-pump model

6.1. Shaping the postglacial GBR shelf

Modern coralgal reefs occur largely between 32°S and 34°N, leading to a conventional presumption that their distribution is controlled by latitude and warm currents (e.g. Ziegler et al., 1984; Kiessling, 2001). Alternatively or additionally, the location of reefs may be controlled by mechanisms related to latitude. Cyclones, for example, by providing episodic nutrient replenishment, comprise a mechanism which could play a partly controlling role in coralgal reef development.

The inner and outer edges of the GBR cyclone corridor comprise energy fences which have exerted great influence on the demarcation of the present shelf physiography and its associated sedimentary deposits (GBRscape). The modern boundaries between sedimentary zones have only been in their present position for the last few thousand years. Prior to then, at 18 kyr BP, the shore-parallel cyclone corridor was located on the upper continental slope, just seaward of the lowstand shoreline and above deep (and colder) water (Fig. 10, lower). At the Last Glacial Maximum (LGM), West Pacific sea-surface temperatures were 3–4°C colder than today (Patrick and Thunnel, 1997). Therefore, for the LGM GBR, cyclones may have been were probably less severe and less common than now—although perhaps not necessarily so, given the complexities of the upper atmosphere and ENSO-type variations in the regional generation of Pacific cyclones (Henderson-Sellers et al, 1998)—and their associated along-slope flow may have had a weaker impact on the (deep) offshore seabed and a greater one on the LGM shoreline itself. During the postglacial transgression, the cyclone corridor must have moved progressively across the lip of the shelf, past the dead limestone pedestals of the modern reef (now surmounted by parts of the Holocene reef tract) (Fig. 10, middle), to arrive finally at its modern position (Fig. 10, upper), evolving in nature as it did so because of the changing nature of the underlying seabed physiography and water mass.

In addition to a range of atmospheric factors, a significant mass (at least on the order of 50 m depth) of warm water >26.5°C is required as the energy source to sustain a cyclone on the shelf itself (Gray, 1968, 1979; Ryan et al., 1992; see also Holland, 1997). Therefore, during postglacial transgressions, slow-moving cyclones would generally have weakened as they crossed the shelf break into shallow shelf waters, until the shoreline had advanced far enough westwards for the shelf-break to have been inundated to a reasonable depth, such that warm shelf waters could sustain the cyclone in its passage along or across the shelf. Such movements of the cyclone corridor, and the increasing magnitude of its effect, must have strongly influenced the development of the physiography, biology and sedimentology of the whole GBRscape; for instance, recent modelling (Sugi et al., 2002) indicates that even modest future global warming may result in *fewer* cyclones in the SW Pacific, but with little change in maximum intensity. The power of the cyclone corridor to shape the shelf system is indicated by the almost complete absence of modern coral reefs located in the modern GBR lagoon throughout the 2500 km length of the reef province, and by the strong partitioning which is observed in the distribution of Holocene sediment facies on either side of the corridor.

Models can be envisaged whereby the frequency of cyclones through the Holocene transgression and highstand increased sharply (Fig. 11a) or gradually (Fig. 11b) around 10 kyr BP, or instead increased slowly

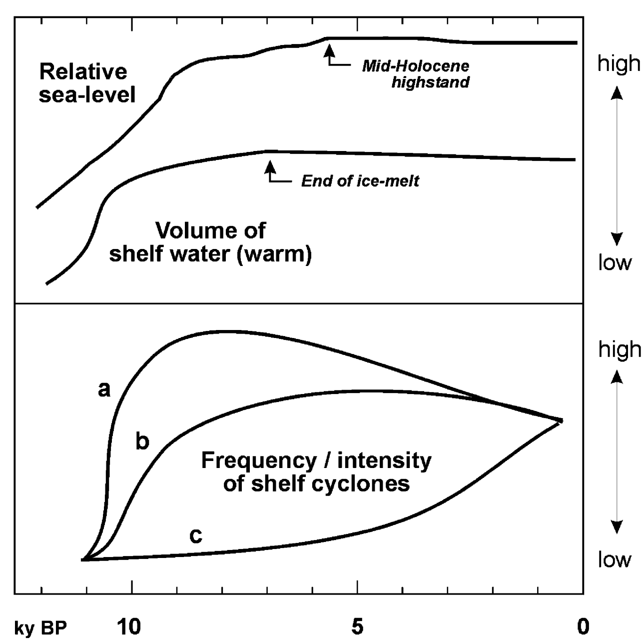


Fig. 11. Conceptual model of the possible effects of changing relative sea level on the frequency and intensity of cyclones. Above, relative sea-level curve for the last 12 ka, and projected volume of shelf water. Below, three alternative paths of development of frequency and intensity of shelf cyclones, which depends, in part, upon the volume of shelf water available. See text for discussion.

up to modern times (Fig. 11c). High-resolution palaeoclimatic studies, such as those of Kershaw (1986) and Gagan et al. (1994, 1998), may eventually allow discrimination to be made between these and other possible models. It is likely that a line such as (c) can be discounted because data indicates that SSTs in the GBR lagoon at the mid-Holocene highstand were about 1° warmer than present (Gagan et al., 1998). The significance of this issue is that the frequency, intensity and favoured paths of cyclones have undoubtedly influenced the development of the GBR shelf physiography as it occurs today. Although some investigators have commented on the frequency and intensity of GBR cyclones during the Holocene highstand (Chappell and Grindrod, 1983; Nott and Hayne, 2001; Hayne and Chappell, 2001), no link at all has yet been established between changing shelf conditions and cyclone activity during the development of the last postglacial transgression.

6.2. *Development of the Holocene reef tract*

One major unexplained finding from Holocene reef drilling on the GBR is a ubiquitous 1–2 kyr gap which occurs between the marine inundation of reef foundations, which most commonly comprise Pleistocene limestone at levels 25–30 m below modern sea level, and subsequent coral growth (e.g. Davies and Hopley, 1983; Johnson et al., 1984) (Fig. 12a). The first transgressive (early or mid-Holocene) sediment atop reef pedestals often comprises carbonate gravel and rubble, and coralgal framework facies are restricted to the upper (youngest) parts of many cored sequences. After inundation of the Pleistocene pedestals, it is probable that initial coral growth would have been affected by vigorous reworking in the cyclone corridor (cf. Massell and Done, 1993), which between 11 and 8 kyr BP would have been slowly traversing shorewards across the location of the modern reef tract. In keeping with this effect at the reef, cores from the middle shelf also exhibit a 2–3 kyr diastem between the time of marine flooding and their basal ages (Fig. 11b) (Carter et al., in prep. B). Finally, reworking and erosion within a slowly shoreward-moving palaeo-cyclone corridor may also explain why the evidence on the GBR shelf for drowned postglacial shorelines is so patchy (cf. Carter and Johnson, 1986) as is the presence of lowstand fluvial deposits (Fielding et al, 2003).

6.3. *What switched on the GBR?*

The separation of the Australian plate from Antarctica began in the late Jurassic, accelerated markedly in the mid-Cenozoic, and continues today at a rate of 6–7 cm/yr (Veevers, 2000, p. 65). Concomitantly, the north Australian climate has progressively warmed, and the central Queensland continental shelf has been bathed by tropical waters suitable for coralgal reef development since about 15 My BP (Davies et al., 1989). Yet, surprisingly, ODP leg 133 and later reef drilling indicates that reef development on the GBR shelf did not commence until about 0.6 My BP (International Consortium, 2001). Therefore (and as for the development of the Holocene reef (see discussion above)) a mysterious, and in this case many million year long, gap exists between the establishment of apparently suitable climatic conditions and reef growth. Why is this? Peerdeman et al. (1993) and Davies and Peerdeman (1998) suggest that the initiation of major GBR coral growth coincided with a 3–5° increase in water temperature indicated by an oxygen isotope shift at 75 mbsf (MIS 9; 600–400 ka) at ODP site 820A. We suggest an additional factor for consideration.

The lip of the GBR shelf occurs at the relatively shallow depth of ~80 m, and the shelf plain landward of the shelf-break is extremely flat, with a seaward gradient of less than 1:1000 (Maxwell, 1968; Hopley, 1982). Accordingly, conditions on the shelf at different times in the recent past were critically controlled by the magnitude of particular Quaternary glacio-eustatic cycles. During the late Pliocene and early Pleistocene, these cycles were of magnitude 60–100 m at a frequency of 40 kyr, and interglacial highstands seldom lasted for more than 5 kyr (e.g. Shackleton et al., 1990) (Fig. 13). By the end of the MPT at about 0.6 My BP, the climatic cycles had lengthened to 100 kyr, their magnitude had increased to 120–130 m, and their shape had become saw-toothed in reflection of slow, episodic sea-level fall into the glaciations and more rapid sea-level rise as the climate cycle warmed into the 5–10 kyr long interglaciations (e.g. Broecker and van Donk, 1970). Oceanic oxygen isotope data and depositional evidence (Rohling et al., 1998; Hearty et al., 1999, and references therein) indicate that MIS 11 and 9 were "super-interglacials", marked by the highest negative anomalies in the Pleistocene, and therefore coincident with particular climatic warmth and high sea-level (see also Hall et al., 2001). Further, MIS 11 was of long total duration, perhaps as long as 37 kyr (Karner et al, 1999; Poli et al., 2000). Because the available GBR deep drillhole data are extremely limited, we do not rule out the possibility that earlier reef growth occurred during the MIS 15, 33 or 37, which were also relatively warm interglacial periods, but the available evidence indicates that nonetheless the main development of the GBR commenced in MIS 11 and 9 (cf. Howard, 1997). Then, for the first time, the higher (and prolonged) sea-level highstand would have allowed a wide continental shelf to develop, with the significant feature that there became room for the establishment of a reef tract seawards of the contemporary cyclone corridor.

The nature of the Holocene reef tract shows that there are two main prerequisites for an offshore reef belt to

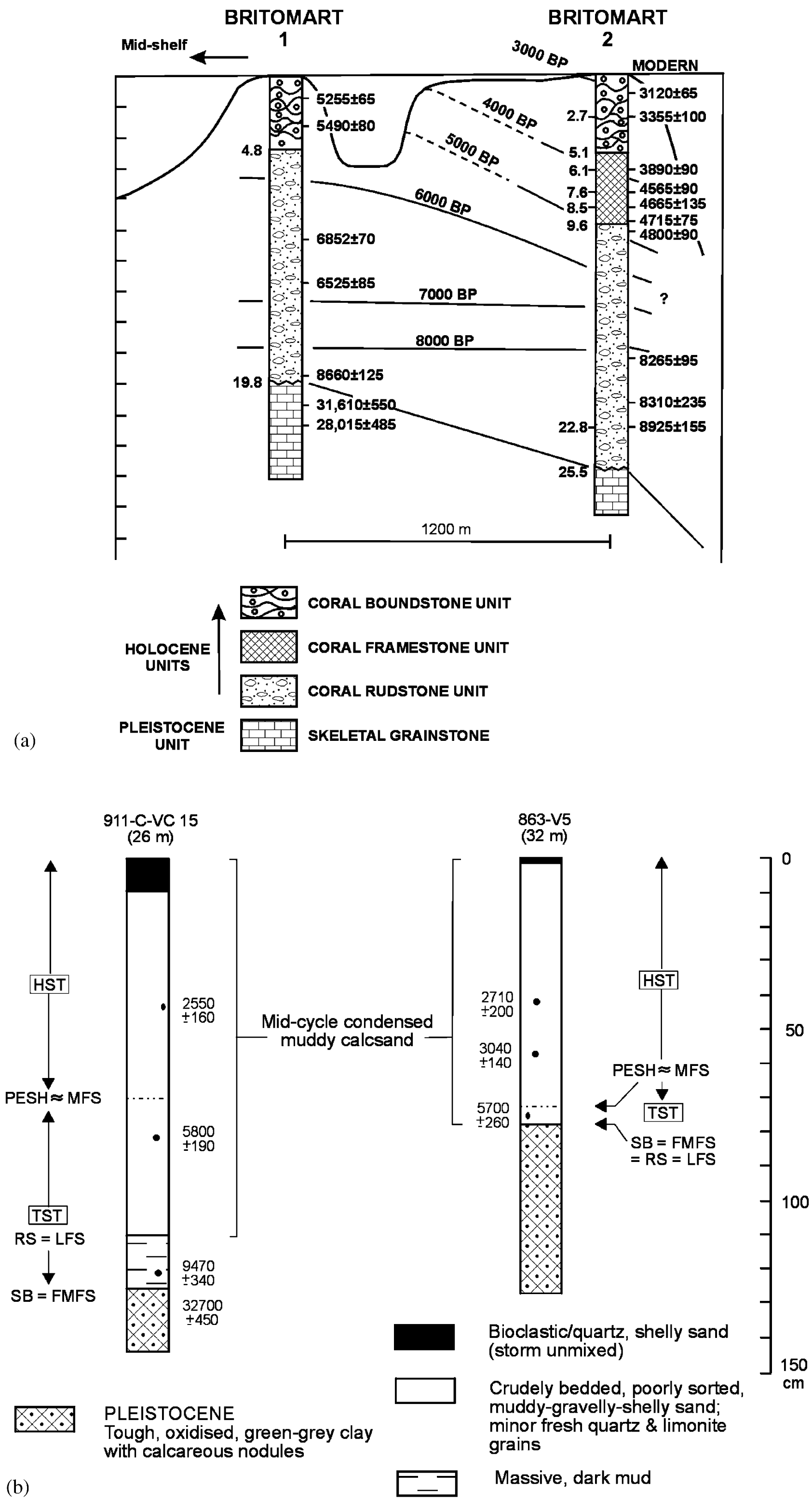


Fig. 12. Stratigraphy of: (a) a typical middle shelf reef (Britomart, after Johnson et al., 1984) and (b) a typical middle shelf core through the condensed muddy calcsand veneer (after Larcombe and Carter, 1998), showing in each case the significant time gap that occurred between marine flooding and the commencement of sediment accumulation.

become established, namely (i) the presence of (older limestone) pedestals in an offshore location to serve as a substrate, and (ii) a highstand period at least several thousand years long. Obviously, the first precursor reefs of the GBR tract could not meet the first condition, and probably instead became established on isolated

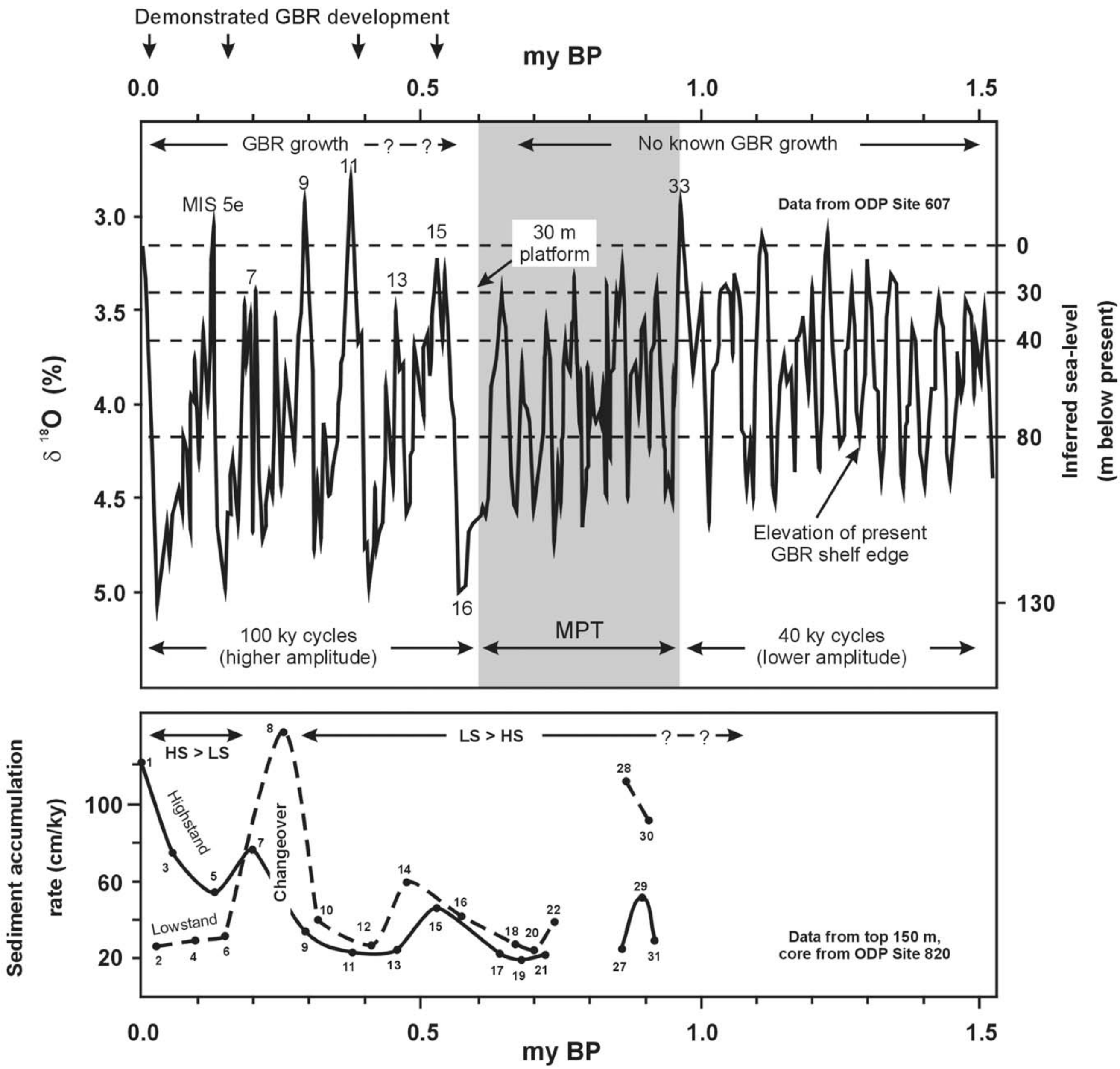


Fig. 13. Above, proxy sea-level curve for the last 1 million years (oxygen isotope data from ODP Site 607; after Raymo, 1992). Note the change in frequency and amplitude of the climatic, and by inference sea level, fluctuations after the MPT at 0.9–0.6 My BP, and that the GBR shelf edge corresponds to the −80 m elevation; other elevations of −40 and −30 m are also plotted. Below, Pleistocene–Holocene sediment accumulation rates from ODP Site 820, east of Cairns (after Davies and Peerdeman, 1998). Note between MIS 9 and 6 the extended duration of sea level between elevations of 30 and 80 m below present, during which period sediment transport processes operating in the cyclone corridor would have enhanced erosion of surficial bed sediments. Further, note that following MIS 9, continental slope sedimentation became dominated by highstand deposits. See text for full explanation.

drowned bedrock pinnacles on the former coastal plain, or on patches of relatively stable gravel sediment, as seen for modern nearshore turbid-zone reefs (Larcombe and Woolfe, 1999; Larcombe et al., 2001). Alternatively, the antecedent reefs to the Holocene outer-shelf reef tract could have first become established as fringing shoreline reefs (cf. Maxwell, 1968) during a particularly favourable interglacial or interstadial sea-level of intermediate elevation, thus providing reef pedestals for an offshore reef tract at the next succeeding full interglacial highstand. For example, the widespread GBR -30 m platform from which so many Holocene reefs arise (Maxwell, 1968; Carter and Johnson, 1986) may represent the eroded top of the former MIS 7 reef tract (cf. Fig. 13).

Irrespective of the nature of the first precursor reefs, which inevitably remains speculative, it is likely that the GBR became established during or shortly after the

MPT (International Consortium, 2001). During the shorter 40 kyr cycles which precede the MPT, the cyclone-corridor will have repeatedly swept back and forth across a relatively narrow, shallow shelf, and thereby acted as a major inhibition to the establishment of an extensive reef tract. In contrast, during the longer and asymmetric 100 kyr cycles which followed, and as instanced by the Holocene example, reef initiation and development would have been encouraged by the higher (and longer) highstands and concomitantly wider and deeper shelf. Shoreward advection from the cyclone corridor on the middle shelf will have enhanced the tendency for terrigenous sediment to remain confined near the shoreline, and thereby enhanced the environment for reef growth in clearer waters and on relatively coarse and stable substrates further offshore.

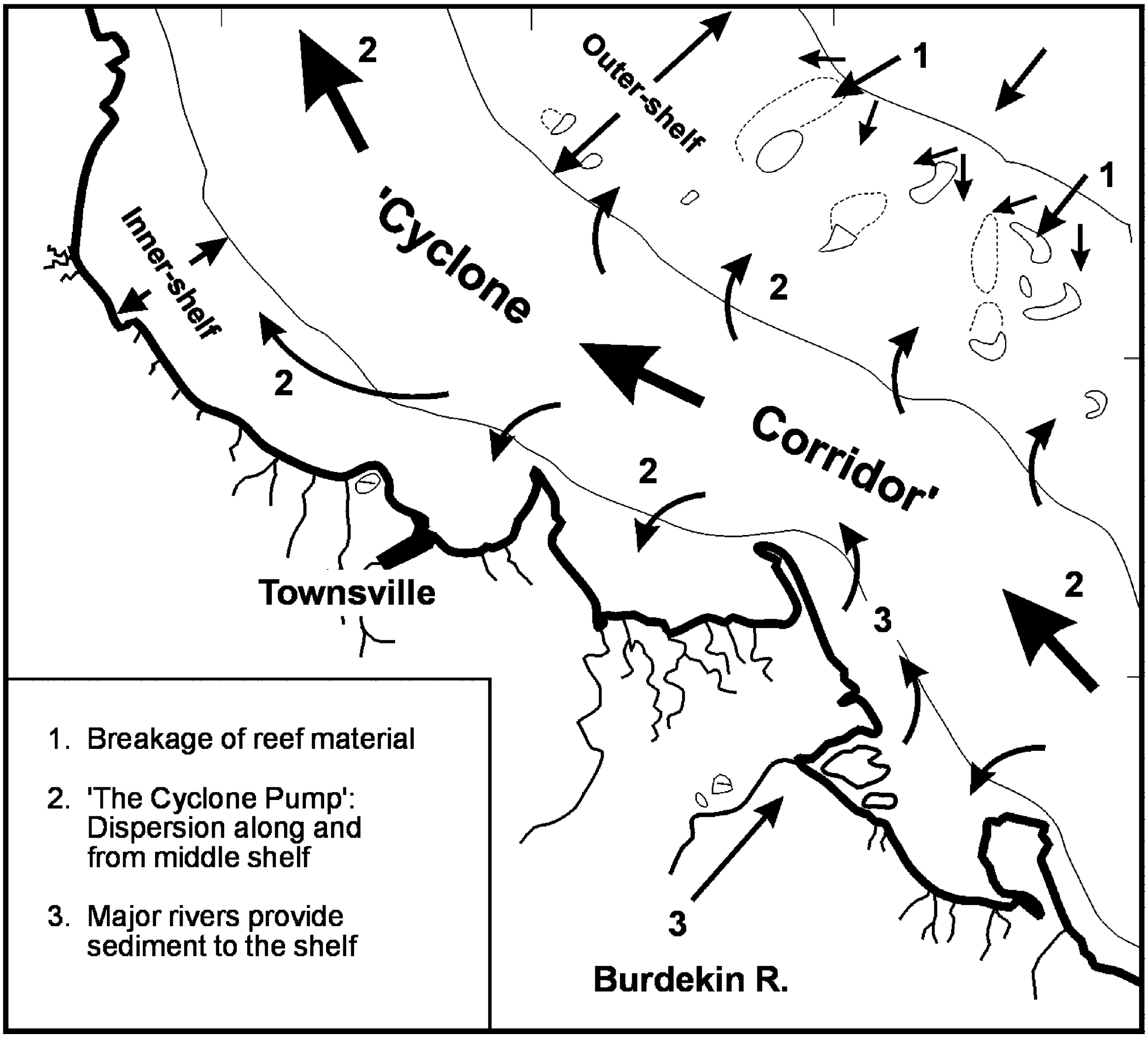
6.4. Sequence stratigraphic implications

The application of sequence stratigraphic models (SSM) to ocean platform carbonate sediments has generated considerable discussion and controversy (for instance, Schlager et al., 1994). However, it has become generally accepted that different models apply to terrigenous- and carbonate-dominated basins. The carbonate model encompasses the concept of "highstand shedding", whereby a greater volume of sediment is input to the deep sea during sea-level highstands (when the reef "carbonate factory" is maximised), than during lowstands (when the reef and shelf are subaerially exposed, and provide only their eroded products to the surrounding fluvial plain). In contrast, the conventional SSM for terrigenous margins predicts greater sediment input to the deep sea during lowstands, when rivers deliver their sediment load directly at or near the shelf-break (e.g. Vail et al., 1991).

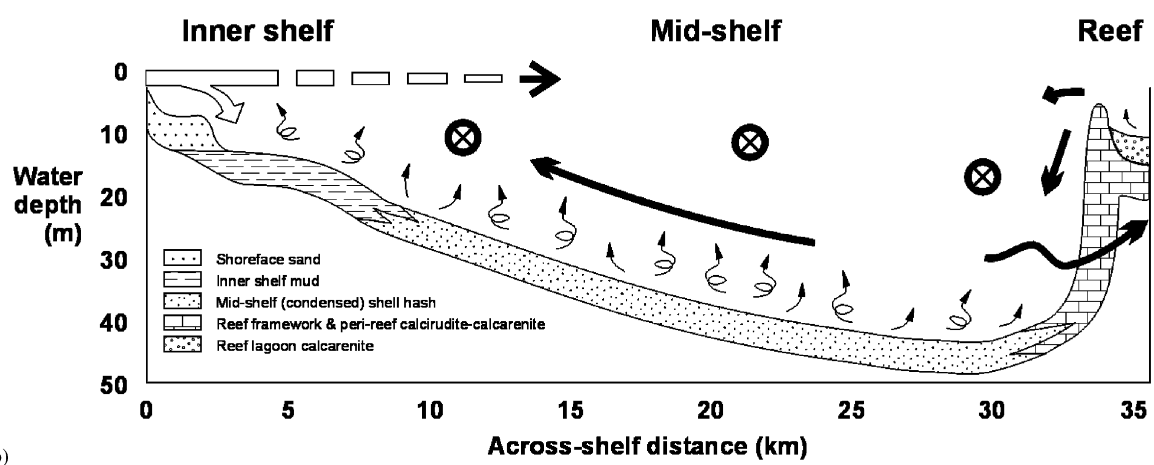
Both of these SSMs have become overgeneralised. First, the carbonate SSM model is strongly based upon examples such as the oceanic Bahamas plateau, where (unusually, for the more general geologic case) no significant terrigenous source exists, and where the carbonate source is located in close proximity to steep ocean-basin slopes. Second, the standard terrigenous SSM does not take account enough of the fact that rates of terrigenous sediment delivery to the deep sea are as much controlled by the local width of the shelf and by accidents of geography, tectonics, climate and oceanography as they are by the highstand or lowstand position of the shoreline. In reality, individual continental margin (or "sedimentary basin") systems most often require their own SSM variant which is specifically tailored to the existing local controls. In such a context, "highstand shedding" may occur within terrigeous systems, especially those characterised by a narrow or bypassed shelf (e.g. parts of the eastern North Island, New Zealand, shelf; Pantin, 1966; Lewis, 1973), and "lowstand shedding" within carbonate systems (e.g. the outer shelf Flower Garden Banks, Gulf of Mexico; Bright, 1977; Rezak, 1977), depending upon the particular local circumstances which prevail.

A sequence stratigraphic model for the central GBR shelf must recognize the mixed nature of the sediment sources (i.e. both terrigenous and carbonate), and the intensely dynamic nature of such tropical sedimentary systems. Three major dynamic controls exist. First, in fair-weather, alongshore and along-shelf currents driven by southeasterly trade winds cause shoreward advection and strong northward movement of terrigenous sediment in littoral and shallow shelf environments. The production of new sediment is mostly confined to in situ calcification along the reef tract (corals, coralline algae), or in areas between reefs (molluscs, foraminifera, *Halimeda*). Second, during the passage of cyclones, major amounts of new terrigenous sediment are introduced by river flooding, and more is derived from erosion of the middle shelf seafloor; concomitantly, new carbonate sediment is produced by storm-induced reef breakage. Cyclone pumping (Fig. 14) then causes much of this sediment to be moved northwards, and part of the finer fraction to be advected towards the coast where it is incorporated into the ISP. Third, the 70–120 km width and the extremely low seawards slope of the central GBR shelf platform together preclude the development of cross-shelf turbid underflows which might otherwise have delivered terrigenous sediment directly to the slope.

The inescapable result of these dynamic processes is: (i) strong sediment partitioning, whereby modern terrigenous sediment is confined almost exclusively within the ISP, and reef carbonate sediment within the reef tract, situated respectively inboard and outboard of the cyclone corridor; (ii) development of the reef tract at the outer edge of the cyclone corridor, in waters largely free from coastal turbidity; and (iii) the delivery to the upper continental slope of mainly carbonate sediment during the later stages of transgression and at highstand (Dunbar et al., 2000), accompanied by small amounts of terrigenous material which most probably represents reworked transgressive or lowstand sediment stored in outer shelf locations. ODP Leg 133 drilling results (Davies and Peerdeman, 1998; Site 820, just east of the GBR off Cairns) demonstrate that during MIS 31 through 9, i.e. prior to establishment of the GBR, the lowstand sediment accumulation rates exceeded those during highstands (Fig. 12, inset). Conversely, for MIS 6 through 1, highstand sediment accumulation rates greatly exceeded those during lowstands. Against this background pattern, MIS 8–7 comprise the last cycle to exhibit higher lowstand rates, with MIS 8 itself exhibiting the highest rate of sedimentation of any of the lowstands cored. This anomaly, and reversal in sediment accumulation rates, corresponds with the



(a)



(b)

Fig. 14. (a) Simple conceptual model of the operation of the cyclone pump on the central GBR shelf, summarising the three main phases of introduction of 'new' sediment to the shelf, and typical sediment transport paths driven by cyclones. (b) Synoptic model of water motions and sediment transport paths induced within the cyclone corridor during the passage of a major cyclone (after Gagan et al., 1989). Crossed circles denote along-shelf currents flowing into the page.

estimated time of origin of the GBR, indicating perhaps that MIS 8 at Site 820 contains the record of the "cleaning off" of the shelf after the first development of the reef tract in MIS 11 and 9.

In summary, then, the Leg 133 data show that strong "highstand" shedding characterised the Cairns region of the GBR outer shelf only during MIS 5, 3 and 1. Later results indicate that the sediment packages intepreted as highstand by Leg 133 scientists in fact encompass both transgressive and highstand materials. Using piston-core data and tight stratigraphic control, Dunbar et al. (2000) have shown that highstand shedding into deep water (as opposed to into peri-reef sediment aprons) is not volumetrically significant on the GBR shelf. Regionally, terrigenous sediments on the north Queensland continental slope accumulated fastest (1.0 Mt/yr) during the postglacial transgression (14.7–6.5 kyr BP) and slowest (<0.1 Mt/y) at lowstands; and accumulation of carbonates was high during both transgression *and* highstand (1.1–1.4 Mt/yr). The low rates of sediment input during lowstand reflect that at that time (i) reef development was restricted to a narrow coastal fringe on the upper slope; and (ii) the low gradient rivers which traversed an arid coastal plain of expanded width deposited most of their reduced sediment load before they reached the lowstand coast (cf. Woolfe et al., 1998). Similar results have been reported by Troedson and Davies (2001) for subtropical upper slope cores from off Noosa, a little south of the southern end of the GBR at latitude 26.5° and in 840–1020 m water depth. These cores show that high rates of carbonate deposition occurred during the middle of the postglacial transgression (11–7 ka), peaking between 9.5 and 7.5 ka; terrigenous deposition was highest between 13–10 ka, and was then sustained at variable but generally high levels through to the present.

The action of energetic marine processes on a newly flooded shallow shelf, and especially the effects of cyclones are probably the critical factors which cause sediment input to the continental slope of tropical and subtropical Queensland to peak during postglacial transgressions. These processes will have caused vigorous erosion and reworking of the outer shelf Pleistocene substrate, accompanied by the direct introduction of new terrigenous detritus in sediment-laden river mouth jets. We present in Figs. 10 and 14 a model for the dynamic, mixed-source GBR shelf system that we have described. With appropriate local modifications, this model may apply also to other modern and ancient mixed terrigenous-carbonate tropical systems.

7. Conclusions

1. The GBR shelf is a dynamic sedimentary system in which fair-weather sediment transport is dominated by nearshore mud resuspension and northward littoral sand drift. The passage of intermittent cyclones creates northward along-shelf currents of 100–300 cm/s, which cause erosion of the middle shelf seabed, transport of mobile bedload, and mainly shoreward advection of suspended sediment.
2. Three major shore-parallel belts of clastic sediment eventuate. At depths of 0 to $\sim$22 m, an inner shelf, terrigenous, shore-connected, sediment prism; at depths of $\sim$22 to $\sim$40 m, a middle shelf zone of sediment starvation, erosion and northward transport, marked by seabed erosion and a thin veneer of mixed terrigenous-carbonate sand ribbons and sand dunes; and at depths of $\sim$40 to 80 m, an outer shelf zone of reef-perimeter (and sometimes inter-reef) carbonate sediment.
3. During highstands such as the Holocene, cyclones act as the main mechanism which builds and maintains the three shore-parallel sedimentary belts. Cyclones dislodge carbonate detritus from the reef-tract by direct breakage, and erode terrigenous clay and sand from the middle shelf Pleistocene substrate. After a cyclone makes landfall, strong rainfall and river flooding contributes new terrigenous sediment to the nearshore shelf, and more mud is contributed to the inshore sediment prism (ISP) by the shoreward advection of material resuspended from the middle shelf seabed.
4. Contrary to current models, (i) GBR storm beds become coarser grained, and less likely to be preserved entire in the sedimentary record, further away from the shoreline; and (ii) on the central GBR, high rates of carbonate sedimentation occur on the slope during both sea-level rise and highstand; concomitantly, terrigenous sediment accumulates fastest on the slope during sea-level rise, and slowest during sea-level low, i.e. despite the carbonate-rich nature of the outer shelf margin, a simple model of "highstand shedding" does not apply.
5. Over longer periods of time, movements of the cyclone corridor across the shelf and upper slope have played a guiding role in development of the physiography of the north Queensland shelf, the 'GBRscape'. The longer and more extreme climatic cycles which have occurred after the MPT may also have helped guide the establishment of the GBR at sometime after ca. 0.6 My BP. Individually and in combination, the higher sea-levels, wider shelf, more numerous stadials, and longer interglacials have all acted to favour sediment partitioning and reef development during recent interglacial periods.

Acknowledgements

The Cyclone Joy current meter data in Fig. 4 and the LADS data in Fig. 6 are reproduced with the respective

permission of the Cairns Port Authority and the Royal Australian Navy, for which we are grateful. We thank Jane Dye, Mike Gagan, Alex Kirsch and Anne Gore for help with field and laboratory studies, and Peter Ridd, Doug Haywick and Russell Jaycock for their ready advice. Scott Smithers, Colin Jago and Ed Purdy gave welcome constructive criticism of an early version of the manuscript, and Cathy McKeagney was superlative in finishing off the figures. Financial assistance for this research was provided by the Australian Research Council (including an Australian Research Fellowship to PL) and by James Cook University. PL thanks the Department of Geography, University of Durham for hosting him as a Visiting Overseas Fellow in 2002 during which time the manuscript was largely completed. We thank Chris Perry and an anonymous journal reviewer for their positive comments.

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