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Evolution of reefs and islands, northern Great Barrier Reef: synthesis and interpretation

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This paper brings together the evidence from shallow coring, surface geomorphology, lithology of exposed rocks, superficial sediment accumulations, vegetation patterns, and the historical record derived from radiometric dating to suggest a sequence of reef and island development on the northern Great Barrier Reef in Holocene time. Reefs initially grew vertically as the sea rose rapidly from glacial low levels. This continued until vertical growth was limited by the air/sea interface as the rate of sea level rise slowed. Vertical growth was then replaced by reef flat formation at low intertidal levels, and by the lateral extension of reefs, especially to leeward. Superficial sediment accumulations on the reef flat define a series of changing habitats for further organic growth, and also record the sequence of Holocene events. Controls of the transition from vertical to horizontal reef growth will be discussed and some comments offered on latitudinal variation in reef form along the Great Barrier Reef.

INTRODUCTION

The work of the 1973 Royal Society and Universities of Queensland Expedition has thrown new light on the history and morphological development of reefs of the northern Great Barrier system in response to sea level change since the last glacial low stand of the sea. In this final discussion we wish to review some of the main conclusions from these new data, in the context of our present knowledge of reef development elsewhere in the world.

Before this work was undertaken, we inferred that as the continental ice sheets melted between 14000 and 7000 a B.P., the level of the world ocean rose over several thousand years at a mean rate of 1 m/100 a and also that sea level continued to rise after the greater part of the ice had disappeared, partly because of 'isostatic decantation' of water from areas rebounding after ice load had been removed and partly because of differential adjustments of the crust to the new water load (Bloom 1971; Walcott 1972). The complex interplay of these factors led to continuing change of sea level after 6000 a B.P., and probably also to differences in the record of sea level at different localities, both on a continental and a local scale, thus casting doubt on the heuristic utility of the concept of a general eustatic sea level curve. Much of the controversy in the coral seas over the Holocene record of sea level change has concerned the interpretation of events during these last 6000 years, which is also the period during which the surface morphologies of existing reefs have developed.

Sea level and reef growth

We also suspected that the growth potential of reefs was such that they could not keep pace with the transgressing sea during the major part of its last rise. Chave, Smith & Roy (1972), in a first approximation based on calcification rates of common organisms and their relative abundance on reefs, suggested a gross potential equivalent to a vertical accretion rate of 0.7 m/100 a, but a net rate of only 0.1 m/100 a. Subsequent estimates of carbonate production, based on analyses of sea water alkalinity, yield mean figures equivalent to accretion on reef flats of 0.3–0.5 m/100 a and in lagoons of 0.06–0.1 m/100 a (Smith & Kinsey 1976). Although some rates of reef growth derived from the recent geological record exceed these figures (see for example, Chappell & Polach 1976), the disparity is not great, and the data are consistent with the inference by Mesolella, Sealy & Matthews (1970) from the raised reefs of Barbados that during times of rapid sea level rise reefs are drowned, and that only during stable or slowly rising sea levels do reefs have the capacity to grow vertically from suitable substrates to the surface and then to expand laterally.

We knew further that, once emersed by a negative movement of sea level, reefs are relatively persistent structures, subject to superficial karstic modification and to diagenetic fabric changes but retaining their gross forms. Estimates based on the solubility of carbonate minerals suggested vertical erosion rates of 0.5–1.0 mm/100 a, but these have been revised upwards by Trudgill's direct measurements on Aldabra Atoll to 26 mm/100 a for subaerial and 50–400 mm/100 a for marine erosion (Trudgill 1976*a*, *b*). These rates are nevertheless extremely slow by comparison with the magnitudes of reef structures and the periods available for karst erosion during low glacial sea levels.

Hence it is not surprising that many workers have concluded that modern reefs thinly veneer older karst-eroded reef structures; indeed, in many parts of the world eroded remnants of pre-Recent reefs protrude through more recent accretionary sequences (Stoddart 1969, 1973). Subsurface exploration in the Marshall Islands first revealed evidence of stratigraphic discontinuities between pre-Recent and Recent limestones, marked by diagenetic changes and marked radiometric age disparities between adjacent units; and the Eniwetok bores showed that such horizons occurred several times in the reef column, corresponding to successive periods of emersion (Schlanger 1963). Similar records have been found by drilling in the Tuamotus, the Leeward Hawaiian Islands, New Caledonia and British Honduras, and demonstrate that many reefs have formed by incremental accretion of thin limestone sequences on periodically submerged but intermittently emerged and karst-eroded older reefs, a model first fully envisaged by Tayama (1952, p. 170) and recently beautifully demonstrated at Aldabra Atoll by Braithwaite, Taylor & Kennedy (1973). Purdy (1974*a*, *b*) went further by using shallow cores to interpret seismic profiles, and was able to show that phases of reef accretion are preferentially located on topographic highs, and hence that reef topography mimics that of underlying karst features.

We have similar evidence for the northern Great Barrier Reef. On the shelf itself the seismic record shows an extensive subsurface discontinuity at depths of 40–100 m, while the bore on Bewick Island passed from Holocene accretionary sediments into altered older limestones at only 5.6 m below datum (the comparable depth at Stapleton was 14.6 m); the material above this discontinuity is less than 6000–7000 years old, and that below is Pleistocene. These figures compare with minimum depths to this 'Thurber Discontinuity' in other reef areas of 13 m at

Eniwetok, 31 m at Funafuti, 37 m at Midway, 10 m in New Caledonia, 6–11 m at Mururoa in the Tuamotus, and 9–26 m in British Honduras, though at many reef localities the discontinuity rises above present sea level to outcrop as pinnacles of *feo*, *ironshore*, *makatea* or *champignon*. This broad comparability is perhaps surprising in view of the probable diversity of erosional environment in different parts of the tropics during the last low stand of the sea. Not enough is yet known of Pleistocene climates in the northern Great Barrier Reef area to make any precise statements about the character of karst erosion. There is some agreement that full glacial periods were arid, with rainfalls cut by one-half compared to the present, and that there was a moister period about 8000 a B.P., with rainfall increased by a similar amount (Webster & Streten 1972; Bowler *et al.* 1976).

Bewick is one of the most 'mature' of the low wooded island reefs, in the sense of having the greatest proportion of its reef top covered with Holocene supratidal sediments and mangroves. This advanced stage may result from the existence of an unusually high reef residual on which the Recent reefs have grown, whereas other reefs with deeper foundations and perhaps greater area, requiring greater volumes of carbonates to bring them to present sea level, have lagged in stage of development. The Bewick and Stapleton records may also be compared with that at Heron Island in the extreme south of the Barrier Reef, where, as interpreted in terms of topography, stratigraphy and seismic structure, the pre-Recent unconformity lies at a depth of 15 m (Davies 1974; Davies, Radke & Robison 1976; Flood 1976). Unfortunately no uppermost Pleistocene and Holocene records are available from any of the other Barrier Reef bores. There is a further point of difference in that the shelf surrounding Bewick is up to 30 m deep, and that in the Capricorn Group is at 35–40 m. Subtracting the known Holocene reef increment, we infer that up to 26 m in the north and 20–25 m in the south represents the thickness of pre-Wisconsin reef growth. Little is known about this sequence, but the record of accretion and erosion that it represents is undoubtedly complex.

Nowhere on the Great Barrier Reef is Pleistocene reef framework exposed above present sea level. The 'raised reef' described by Jukes (1847, vol. I, pp. 340–342) near Raine Island is a deposit of storm boulders, and there is no evidence at Raine Island itself, in spite of frequent references to the contrary, of any emergence. This absence of Pleistocene reefs is in marked contrast to the situation in Western Australia, much of the Indian Ocean, mobile Indonesia, parts of Polynesia and the Caribbean. The generally ubiquitous last interglacial reef limestones, aged 90000–130000 years, are not found, although individual corals of this age (112000–143000 a) have been reported from the mainland coast of eastern Australia at 28° and 34 °S, beyond the present limits of the Great Barrier Reef (Marshall & Thom 1976). The absence of outcropping Pleistocene reefs with the Barrier Reef province itself leads to the inference that subsidence rather than stability has dominated reef development in that area.

Sea level and reef-top morphology

Microatolls and the cessation of vertical growth

The sea reached its present level on the northern Great Barrier Reef by about 6000 a B.P., as shown by dated microatolls (Fisher 6310 ± 90; Low Wooded 6080 ± 90; Houghton 5850 ± 170; Leggatt 5800 ± 130). On several reefs extensive fields of microatolls grew on reef tops, many within the vertical range of modern corals, between 6300 and 4800 a B.P. This confirms the conclusion previously reached from other evidence that in Australia, sea level had reached its present level substantially earlier than appears to have been the case in Micronesia and Northern Hemisphere areas, where the record appears lagged by up to 2500 years (Thom & Chappell 1975).

The highest of these fossil microatolls reaches 0.7 m above the present maximum height at which modern moated microatolls form (1.55 m, or approximately h.w.n.), and is dated at 3700 ± 90 a B.P. Some at least of the fossil microatolls may not have been moated, but may have lived on open reef tops (the amount of sediment accumulated on the flats cannot have been great in the initial stages of flat formation), and we must infer either a marginally higher sea level or a slightly greater tidal range than at present at about 3500 a B.P. Hopley (1975) has described similar fossil microatolls on a fringing reef at Middle Island, near Bowen, at about 1.6 m (approximately m.s.l.), with ages of 5210–5290 a B.P.

The microatoll is a diagnostic feature of a critical transition in the mode of reef growth, from unconstrained vertical upgrowth when the reef top is still below sea level, to inhibited vertical and extended lateral growth when the top reaches sea level (or more precisely the level of low water neaps). We have no information from any Queensland reef of the internal facies distributions resulting from this transition, but the age scatter of microatolls from 6000 a B.P. to the present reveals that the transition is diachronous, as a result of differences in basement depth and topography, reef size, and reef environment and ecology. Many extensive surface reefs of apparently simple structure probably conceal complex growth histories. Marshall & Orr (1931, pp. 117–123) showed at Low Isles how much of the reef volume is sedimentary infill rather than growth framework, and there are several cases in the Northern Province of irregular reefs in the process of apparent coalescence to form larger and simpler reefs (Turtle II and East Hope are examples). Conversely it is possible, at least in the Princess Charlotte Bay area, that some large platform reefs are thin veneers on flat-lying Mesozoic sedimentary rocks, and have simpler histories.

It is instructive to compare the Barrier Reef record of reef growth with other areas. A fringing reef in Panama shows a period of rapid vertical growth beginning about 7000 a B.P. at -15 m, with reef flat formation about 2500 a B.P.; similarly a reef on Oahu started growing about 7000 a B.P. at -10 to -15 m, and its reef flat formed about 2000–3000 a B.P. (Macintyre & Glynn 1976; Easton & Olson 1976). In both cases the approach of the reef top to the sea surface led to the replacement of dominantly vertical growth by lateral growth, with consequent ecological and facies changes. During the last 7000 years, in both localities, the sea itself rose by 10–15 m. Thus reef growth began when the sea flooded the -15 m level, and was presumably successful because the rate of sea level rise had by then decelerated sufficiently for incipient reefs not to be drowned. At Eniwetok in the Marshall Islands, however, the formation of a reef flat with microatolls occurred between 4980 and 1885 a B.P., when active coral growth was replaced by rubble accumulation (Tracey & Ladd 1974; Buddemeier, Smith & Kinzie 1975) the Eniwetok record is closer in time to that of the Great Barrier Reef than that in Panama and in Oahu.

Rampart and platform formation

Shingle ramparts on the windward sides of inner reefs of the northern Great Barrier are storm-deposited materials of varying ages; records in this century demonstrate their variability in form and location. Platforms are their lithified equivalents and occupy comparable positions on the reefs. Steers (1937) showed the widespread distribution of 'upper' and 'lower' platforms, though the former are more restricted in extent than the latter; in accepting his distinction, we do not necessarily wish to imply in all cases either accordance, contemporaneity, or even constant relation between the two features so named. Steers believed that both platforms had similar origins as cemented ramparts, but that the upper owed its greater elevation to its formation during a former higher sea level in the Holocene.

The upper platform stands at 2.6–3.8 m (datum m.l.w.s.) and is often being dissected by erosion; radiometric ages on constituents range from 3050 ± 70 to 4420 ± 90 a B.P. and cluster in the interval 3300–3600 a B.P. Two dates on cement suggest that lithification took place about 1000 years after the formation of the clasts. The lower platform is usually separated from the upper, where both occur together, by an interval of 1–1.5 m. Its height range is 1.6–2.4 m and ages range from 380 ± 80 to 1460 ± 70 a B.P. More recent rampart deposits are unlithified, and probably overlap with younger lower platform deposits in age. The modern ramparts, in process of formation, and the lower platform are clearly related to present sea level: we could, therefore, infer, as did Steers, that the upper platform derived from ramparts built during a sea level up to 1 m higher than present, thus reinforcing the inference already made from the existence of microatolls of similar age (*ca.* 3500 a B.P.) up to 0.7 m above their modern counterparts. Two points of caution must be noted: (*a*) there is uncertainty about the level to which platform cementation can take place with respect to sea level, and (*b*) it is curious that if sea level did stand higher, no elevated reef framework, as distinct from moated microatolls, has been found on reef tops.

Neither ramparts nor platforms are unique to the northern Barrier Reef; indeed there is some danger that the use of a distinctive terminology implies an unwarranted degree of singularity. Steers (1940) drew attention to loosely comparable 'promenades' on the Pedro Cays, Jamaica, and comparison can also be made between the Barrier Reef platforms and the widespread rock ledges of windward coasts in the Carolines, Marshalls, central Polynesia, and the Tuamotus. These are, however, always single features, and being in microtidal situations they are less spectacular than the Queensland examples; their upper surfaces are at or slightly above the level of h.w.s. At Mururoa, Tuamotus, such a ledge rises to + 3 m and is dated at 3610 a B.P.; dates generally range from 1270 to 4350 a B.P., and most cluster between 2000 and 3000 a B.P. (as at Ebon, Jaluit, Ailinglapalap, Aitutaki and Mopelia: Curray, Shepard & Veeh 1970; Guilcher *et al.* 1969; Stoddart 1975). The age range taken as a whole is greater than that for the Queensland examples, nor do most of the ledges necessarily suggest formation during any sea level higher than the present (Curray *et al.* 1970; Newell & Bloom 1970). They are best interpreted as lithified storm-built ridges, which in the absence of any marked storm periodicity or sea level change need not show any marked clustering in heights or ages.

Sand cays

Sand cays are one of the most studied features of the coral reefs, but very little is known of their stratigraphy or history, and the few radiometric dates hitherto available have been either ambiguous or uninformative. We have shown that sand cays on the northern Great Barrier consist of an extensive high core surrounded in some cases by a discontinuous lower terrace. The high core effectively delineates the present gross topography of the cay. Its elevation varies from 5 to 7.3 m, and ages of constituent sediments range from 3020 ± 70 to 4380 ± 80 a B.P., averaging about 3500 a B.P. Hence the cays formed after 6000 a B.P., when the reef flats first appeared at the surface, but were essentially complete in shape and size by 3000 a B.P. The lower terrace, where present, stands at 3.5–4.5 m; constituent sediment ages range from 21 90 ± 7 to 3280 ± 80 a B.P., with an average of approximately 2700 a B.P. However, the cartographic evidence of Steers's surveys and our own shows that on at least two islands where the sediment ages are greater than 2000 a B.P., the lower terrace as a topographic feature has formed in large part since 1936.

Nevertheless the clustering of dates for the two levels, supported by sedimentological, pedological and vegetational differences, and the remarkably constant height interval between them, requires explanation. The lack of cay sediments dating as younger than 2000 a B.P., if not caused by sampling bias, is notable: even the sediments of an unvegetated and ephemeral cay (Pickersgill) are 2330 ± 70 a old. The only younger sediments are gravels from unconsolidated storm-deposited banded shingle ridges, which on a number of islands show a scatter of ages from 510 ± 50 to 1550 ± 70 a; comparably scattered dates have been published by Hopley for islands near Townsville and Bowen (Hopley 1971, 1975).

Beach-rock

Beach-rock is extensive on sand cay shores; it has an aragonite cement and forms intertidally. On islands with both upper and lower terraces there are also two distinct generations of beach-rock: an older beach-rock on the shores of the higher part of the cay, and a younger, less consolidated rock on the shores of the lower terrace. On some islands only higher beach-rock is found, generally forming narrow horizontal shelves; on some smaller islands only the lower beach-rock occurs, forming, as on lower terrace beaches, characteristic inclined ledges. The higher rocks, which are well consolidated, have height ranges up to 2.5–3.0 m (m.h.w.s. 2.3 m), and their ages range from 2030 ± 70 to 2670 ± 70 a B.P. The lower rocks, between 1 and 2.3 m, are modern. The existence of these two types of beach-rock is thus consistent with the evidence of two aggradation levels on the cays and of two periods of rampart formation in the rampart rocks.

Rocks interpreted as old beach-rocks have been found on islands further south near the mainland coast between Bowen and Cairns, and especially in the Palm Islands. These have maximum elevations of 4.1–7.8 m, and ages from 3240 to 5250 a B.P. On at least two of these islands there is evidence of two levels of beach-rock, with a height difference of about 3.5 m (Hopley 1971). In the Bowen area three levels of beach-rock *sensu lato* have been reported in the height range 4.8–6.4 m, with ages in the range 3350–6020 a B.P. (Hopley 1975). The sea level record in these islands, however, appears to be different from that of Barrier Reef islands to both north and south, possibly for tectonic reasons (Hopley 1974, 1975).

Summary

Any interpretation of the development of reefs and islands on the northern Great Barrier Reef and any reconstruction of sea level history must therefore take account of four main groups of facts.

(1) Reef flats formed, as shown by the presence of microatolls, close to present sea level, on several reefs between 6000 and 5000 a B.P., between 12° 14′ S and 15° 06′ S. Not all reefs formed horizontal surfaces at sea level simultaneously, and some are still doing so, but this period represents the earliest during which substrates were available for reef islands to form on. Hopley (1977) has subsequently obtained similar dates for reef flat formation on outer ribbon reefs.

(2) Three groups of features formed in the time interval centred on the period 4000–3000 a B.P. Storm ramparts (now represented by upper platform rocks), often overlying fossil

microatolls, formed over the range 4420–2050 a B.P. and cluster during 3600–3300 a B.P. The core areas or upper terraces of sand cays formed by normal wave refraction rather than by storm activity were built from sediments dated 4380–3020 years, and clustered round 3500 a B.P. High beach-rock formed round the shores of these high-standing cays during the period 4380–2030 a B.P. Each of these features stands at higher elevations than comparable features associated with present sea level: in broad terms this height anomaly reaches 1 m for the ramparts represented by the upper platform, 1.5 m for the high cays, and 0.7 m for the high beach-rock; the corresponding anomaly for the highest fossil microatolls (dated 3700 a B.P.) is 0.7 m.

(3) Three groups of features, which appear to be related to present sea level, have substantially younger dates. Storm ramparts represented by the lower platform, with elevations within present tidal range, date between 1460 and 380 a B.P. The lower terrace of sand cays, with heights of 3.4–4.5 m, is formed of sediments dated between 3280 and 2190 a B.P. The storm-deposited banded shingle ridges formed between 1550 and 510 a B.P.

(4) Contemporary changes shown by cartographic evidence over the last 45 years include minor adjustments of cay shores, both erosional and aggradational; the often substantial relocation of rampart tongues and ridges during storms; and the episodic extension of mangroves on reef tops.

Interpretation

Similar clusterings of dates for reef features have been found in other parts of the world, notably in the Marshall Islands (Buddemeier *et al.* 1975) and in Micronesia (Curray *et al.* 1970). The factor most frequently invoked to explain such data is Holocene sea level change: in the reef seas this was emphasized by Daly (1920) and powerfully reinforced by the influential work of Fairbridge (1961). It is, however, only one of several factors which need to be considered. These may be broadly grouped as *external* and *internal* controls of changing reef morphology.

External controls

Of these controls, sea level is the most obvious, but in a macrotidal situation such as that of north Queensland it is often difficult to relate particular reef features to specific tidal levels. Thus we have shown that corals can grow in moated situations to levels more than 1 m above those of open-water reef flats. Further ambiguities arise in considering the relations between lithified features such as platforms, the sedimentary accumulations from which they are formed, and cementation processes in relation to tidal levels. Thus the surfaces of upper platforms stand up to 1.5 m above still-water h.w.s. tide levels (2.3 m), but the highest astronomical tides reach 2.9 m, storms may carry local levels considerably above this, salt spray during high-tide rough weather may wet deposits higher still, and where interstices of deposits are mud-packed, capillarity may carry internal water levels in ramparts (and hence potential cementation levels) above those of external tides. We are, therefore, hesitant in too readily adopting a higher sea level explanation for the cluster of features dated 4000–3000 a B.P., for at least four further reasons:

(*a*) we have nowhere found constructional reef framework of this age exposed on the northern Great Barrier Reef;

(*b*) as Hopley (1974, 1975) has indicated, it is possible that tectonic movement associated with

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structural lineations on the mainland could have differentially affected reef levels along the Queensland shelf;

(*c*) on theoretical grounds Chappell (1974) has proposed that reef levels across the shelf may also have been affected by transverse deformation;

(*d*) any interpretation is at present based on assumed constancy of tidal range. However, if tidal range had contracted in the Holocene this would explain both the existence of high-standing fossil microatolls and the absence of high-standing reef framework, without recourse to changes in mean sea level.

Clustering of ages of sediments could also be explained by absolute variations in storminess in Holocene times, affecting the rates of mechanical destruction of corals and the process of lodgement of debris on the reef flat. We know that over the period 1909–69 decennial frequencies of cyclones in the area north of Cooktown varied by a factor of two. Further, in the same period, the area south of Cooktown had nearly 45% more cyclones than that to the north (Coleman 1971), so that latitudinal shifts over time as well as changes in absolute frequency could lead to variations in sediment supply. The only unambiguous evidence for cyclone frequencies, however, would come from the ages of large storm-deposited reef blocks. We have dates on two such blocks on the reef flat at Ingram (4310 ± 100 a B.P. for a block 130 m from the reef edge, 640 ± 70 a B.P. for a block 80 m from the edge), and also for a block (2420 ± 70 a B.P.) in upper platform deposits at Howick, but these are quite inadequate for any conclusions to be drawn. Peaks of storminess could explain periods of rampart building and hence clusters of high corals in rampart-bounded moats; but they are less likely to explain periodicities in sand cay formation, since these result from normal swash and wind processes rather than storms.

There is no evidence in the reefs studied of secular changes in wind direction over the last 6000 years, of the kind identified in this century in Indonesia by Verstappen (1954).

Internal controls

The processes of reef and island formation involve not only responses to external controlling factors but also complex interrelations between processes, some of which are self-damping and others self-reinforcing. The existence of these negative and positive feedback linkages has rarely been explicitly considered in analyses of reef geomorphology, but they add a further complication to the assessment of reef responses to presumed changes in external factors. Three such linkages may be considered.

First, as Neumann (1972) suggested, since reef growth lags behind a rising sea level, there will be a period immediately after sea level has stabilized but before reef crests reach the surface when wave activity will not be damped by the coral baffle or excluded by coral barriers. This 'Holocene energy window' will be closed by the growth of the reefs themselves, leading to an apparent decrease in energy both across reef tops and on mainland coasts protected by reefs. This is a concept more relevant to reef-bordered continental coasts than to the tops of small patch reefs, and on the northern Barrier where reefs have reached the surface at different times over the last 6000 years it does not necessarily imply a finite time interval.

Second, we might infer that as greater proportions of a reef top are covered with sedimentary accumulations such as ramparts, cays and mangroves, the area available for active carbonate production is reduced, and that ultimately a steady-state situation is reached where a given amount of sediment formed during a high-production period is resorted and relocated on the reef and where any additional sediment supply is balanced by sediment export. This could account for the absence of young sediments (other than recent storm shingle ridges) and the clustering of most sediments in the period 4000–3000 a B.P. when the reef flats were still open and productive. Not only would sediment supply be affected by such a mechanism, but, by an extension of Neumann's principle, wave energy on reef tops would be diminished by the spread of islands and mangroves. Hence this might account, for example, not only for clustering of sediment ages but also for the formation of discrete upper and lower levels on cays. That this cannot be a complete explanation is shown by the fact that upper and lower terraces with high and low beach-rock are found on reefs which are almost completely open (such as Ingram and East Hope), and the lower terrace may be absent on reef flats almost completely covered by land (such as Bewick). Further, in assigning a dominant rôle in sediment production to corals on reef tops, we may well underestimate the contributions of other calcifying organisms, notably Foraminifera and algae, in pools and grass beds. If, however, sea level had fallen marginally below its present level since 3000 a B.P., the reef tops would have been emergent and calcification completely inhibited.

Both the above are examples of negative feedback linkages. The most obvious positive feedback process is the way that once sedimentation is initiated, particularly by storm deposition of rubble and shingle on windward reef edges, these deposits both form the nucleus of further deposition and also reduce energy conditions on the reef top so that mangroves can colonize. Reef-top levels are thus rapidly raised from the level of open-water coral growth (low water neaps, 1.2 m) to the maximum of mangrove accumulation just above high water springs (2.5 m). The mangroves themselves serve as an energy baffle leading to further accumulation of shingle ridges on their windward sides. These incremental growth processes, once initiated, act rapidly, and lead to the replacement of bare reef flats by the complex array of high depositional features associated with the low wooded islands. The crucial initiating step is the formation of storm ramparts, which, *contra* Spender (1930), is less a function of reef height (which is an effect rather than a cause of the changes) than of varying energy conditions across the shelf: on outer reefs coarse sediments are spread across the flats because wave energy is high, whereas on inner, more protected reefs the coarse material is simply lodged on the windward edge, where it triggers the sequence of developmental changes (Stoddart 1965). This initiating trigger may be considered a random event in that it is contingent both on existing reef morphology and on individual cyclone behaviour, and this, together with the rapidity of subsequent morphological changes, may explain some of the spectacular differences between the surface features of closely adjacent reefs. Since sand cays are built by normal rather than by storm wave processes, through refraction at the leeward ends of reefs, their formation is independent of the sequence of events leading to the development of low wooded islands, and they have indeed a much wider distribution.

Conclusion

These investigations by the Royal Society and Universities of Queensland Expedition to the northern Great Barrier Reef have revealed both the complexity of recent reef history and also the ambiguity of the many kinds of evidence available for reconstructing Holocene history. Indeed we believe that no useful purpose would be served by attempting to fit our data to pre-existing sea level curves, such as that most recently revised by Fairbridge (1976). Rather we would like to suggest three areas of further enquiry which might help to resolve some of the difficulties which we have found.

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First, we require a detailed knowledge of the internal structure of a coral reef of the Great Barrier system, to determine its response to changing sea level. This knowledge can only be achieved by closely spaced drilling and radiometric dating; and ideally it is needed for outer ribbon reefs, islandless platform reefs and fringing reefs, as well as low wooded island reefs.

Secondly, we need to test the hypothesis of fluctuating storm frequency during the last 6000 years by radiometric dating of sequences of storm-deposited reef blocks, both on individual reefs and on different sectors of the Barrier.

Thirdly, we need to compare the data from the reefs with the record from the mainland coast of Queensland. Hopley (1970) began this work on aggradational sequences with his study of the Burdekin delta to the south, and it now needs to be extended to the beach ridge sequences of the lower Normanby round Princess Charlotte Bay and to the Pleistocene coastal dunes north of Cooktown. Steers (1929, 1937) also noted the possible correspondence between features on the reefs and wave-cut platforms on the mainland coast and high islands, and these require investigation as integral parts of the problem of the Great Barrier Reef.

Finally, it is salutary to recall that all the events which we now so painstakingly attempt to reconstruct were witnessed by man himself. People have lived in Queensland for at least 18000 years, and undoubtedly settled on the coastal shelf, possibly in rock shelters at the foot of limestone hills which now form the reefs, before it was flooded by a rapidly transgressing sea about 8000 years ago. There is much to be learned about the features of the shelf itself to supplement the ideas put forward by Maxwell (1968), and it is possible that the archaeological record may supplement the geological and geomorphological here as well as on the islands.

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