The Development of Talus Slopes around Lord Howe Island and Implications for the History of Island Planation

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ABSTRACT  Lord Howe Island is a small eroded remnant of a Late Miocene shield volcano. A fringing coral reef dissipates wave energy along a portion of the shoreline, but the remainder of the coast is rugged with spectacular high basaltic sea cliffs. This paper investigates the evolution of talus slopes that occur beneath the loftiest cliffs, and places this analysis within the context of a longer history of island planation that has resulted in a wide truncated shelf around the island. During the Last Glacial, when the sea level was lower than at present, talus slopes accumulated around the extent of the island’s cliffed coast because material eroded from cliffs by subaerial processes could not be removed by marine processes. The survival of these slopes during the Holocene has depended on a balance achieved between rates of subaerial and marine erosion. This balance is fundamentally influenced by cliff height, as cliffs higher than 200 m are plunging or veneered by talus slopes, whereas lower cliffs have erosional shore platforms. On comparison with published erosion rates from inland basalt scarps it appears that marine processes may account for over 90 per cent of the total cliff retreat that has occurred at Lord Howe Island, yet contemporary coastal morphology attests to the significance of subaerial processes in recent times. It is likely that marine cliffing was very rapid soon after volcanism ceased, but rates of erosion decreased through time as wave energy became increasingly attenuated across a widening planation surface, and as increasing cliff heights yielded greater quantities of talus that provided protection from rapid marine erosion.

KEY WORDS  Talus slope; oceanic island; sea cliff; plunging cliff; shore platform; rock coast geomorphology

Introduction

Lord Howe Island (31° 33’ S, 159° 05’ E) is a small basaltic oceanic island in the northern Tasman Sea (see Figure 1). The island is crescent-shaped, approximately 11 km long and 0.5–2.5 km wide. Basalt lava flows, separated by surface breccias, form the prominent hills of Mount Gower (875 m) and Mount Lidgbird (777 m) at the southern end of the island, as well as Transit Hill (121 m) and the North Ridge (100–200 m) at the northern end. The Admiralty Islands, Malabar Point and Stevens Point are composed of Roach Island Tuff, whereas much of Intermediate Hill (250 m) consists of Boat Harbour Breccia. The low-lying middle portion of the island is veneered by Late Pleistocene eolianite that consists of calcarenite with isolated beach facies (Brooke et al. 2003). The island is surrounded, in water that is generally less than 50 m deep,
by a near-horizontal shelf that measures c.28 km NW–SE by 23 km NE–SW; the width of the shelf indicates that considerable truncation of the island has occurred since it was formed during the Late Miocene (McDougall et al. 1981). Midway across the shelf, in
water about 30 m deep, there is an extensive fossil coral reef that nearly encircles the island, but the chronology of this feature is not known (Kennedy et al. 2002).

The island has a temperate to subtropical maritime climate with mean summer temperatures between 26°C and 20°C and mean winter temperatures between 20°C and 13°C. Average annual rainfall is 1676 mm with a maximum of 2870 mm and a minimum of 1000 mm (Bureau of Meteorology 1997; Pickard 1983). The island has essentially unlimited fetch and experiences strong winds throughout the year; accordingly, the wave climate is very powerful. On the basis of linear wave theory, 28 years of daily wave-observation data (compiled by the Australian Bureau of Meteorology) were analysed with reference to nearshore bathymetry to establish the occurrence and breaking characteristics of different wave heights around the island (Dickson 2002; Dickson in review). The analysis showed that east, northeast and southeast wind and swell waves occur most frequently, but the largest waves come from the southwest as well as the east.

The coastal geomorphology of Lord Howe Island is highly diverse owing to lithological variability as well as contrasts in wave energy associated with a fringing coral reef that occurs along a portion of the western coastline (Kennedy & Woodroffe 2000). The southern coastline consists of spectacular, near-vertical cliffs up to 800 m high that either plunge into the sea or are mantled by talus slopes. Much of the remainder of the coastline consists of cliffs less than 200 m high that have basal shore platforms, but the reef-protected coastline consists largely of depositional sandy beaches. Six talus slopes occur around Lord Howe Island (see Plate 1). Four of the talus slopes occur at the foot of cliffs surrounding Mount Gower: The Big Slope and a small pocket of talus immediately south of The Big Slope occur on the eastern face, The Little Slope occurs on the western face, and a smaller talus slope occurs on the western side of King Point. One talus slope occurs on the western face of Mount Lidgbird and is partially protected from marine erosion by the fringing coral reef. The northernmost talus slope occurs between Malabar Hill and Kims Lookout at the base of the North Ridge cliffs.

This paper examines the evolution of talus slopes around Lord Howe Island. Such an analysis is important as it provides an insight into the long history of processes that have resulted in the current form of the island’s shoreline. This insight emerges through observations of the current gross morphology of talus slopes and the cliffs from which they are sourced, the morphogenetic contrast provided by adjacent landforms, and the morphology of the truncated shelf that sets the context of a long history of island planation.

**Methods**

Talus slopes and gross cliff morphology were mapped using vertical colour aerial photographs with details of particular stretches of coast clarified using aerial video reconnaissance, or visited in the field by small boat with ground survey possible when sea conditions were relatively calm. The study focused on broad-scale aspects of talus evolution, but occasional field observation revealed that talus slopes are well vegetated and consist largely of boulders and cobbles with relatively little sand and gravel. In contrast to adjacent resistant basalt cliffs, the talus slopes may be viewed as accumulations of boulders with relatively little strength provided by supporting fabric. Digital terrain analysis was undertaken using a Digital Terrain Model in ArcView GIS. A Triangulated Irregular Network was generated from digital contours, and maps of slope angles were produced and overlain on this to generate three-dimensional images (see
PLATE 1. Beneath the spectacular high cliffs of Mount Gower are (a) The Big Slope, (b) a talus slope near King Point, and (c) The Little Slope. (d) A talus slope on the flanks of Mount Lidgbird is partially protected from marine erosion by the fringing reef, in contrast to a small talus slope beneath the North Ridge cliffs (e).

Figure 2). Talus source areas were defined as areas of cliff face where slopes exceed 50°, whereas talus slopes were noted as coinciding with slope angles between 25° and 40°. Figure 2 illustrates that talus slopes generally occur beneath sub-vertical cliffs at Lord Howe Island, such that a demarcation between talus and cliff face (plunging cliffs) could be made readily on the basis of slope angles. Digital interpolation, visualisation and area and volume calculations utilised ArcView extensions Spatial Analyst, 3D Analyst, and X-tools. Nearshore sea-bed morphology was determined from bathymetric data derived from hydrographic survey as well as CEEDUCER echo-sounder traverses that were undertaken for this project. The morphology of the planated shelf was derived from various data sources including Admiralty surveys, RV Franklin soundings, and Laser Airborne Depth Sounder data.

Processes of talus accumulation

At least two major mechanisms are apparent through which talus slopes might form at the base of sea cliffs: weathering processes, which result in gradual talus accumulation through small rock falls, and rapid talus accumulation that results from cliff collapse, which is dependent on processes such as marine undercutting through the development of extensive notches (Sunamura 1992) or the coalescence of sea caves (Cotton 1967).
The Development of Talus Slopes

FIGURE 2. Three-dimensional visualisation of Lord Howe Island from the east with slope angles draped over a Digital Terrain Model. A clear demarcation is apparent between talus slopes and sub-vertical cliff faces.

In the case of oceanic islands, it is possible also that mega-landslides could leave scarps mantled by talus material (Holcomb & Searle 1991), but these landslides are often triggered by earthquakes or volcanic activity during the growth of volcanic islands or soon after (Masson et al. 2002), whereas Lord Howe Island has been volcanically inactive for the last 6 million years (McDougall et al. 1981) and the island sits in the middle of a broad shelf well away from the shelf-edge where mega-landslide activity would probably have been concentrated (see Figure 1). It is more likely that talus slopes at Lord Howe Island have accumulated gradually through continued subaerial erosion of cliffs, or through marine undercutting.

The relative contribution of marine and subaerial erosion processes to gradual rock coast evolution has proved to be a difficult research problem (Trenhaile 1987). In a generalised classification Emery and Kuhn (1982) described cliffed profiles as active, inactive or former, and related these to the relative efficacy of marine and subaerial erosion. The classification depicts cliff profiles as very steep where marine processes dominate, whereas the top-of-cliff profiles are rounded where subaerial processes dominate. Emery and Kuhn (1982) note that lithology exerts an important constraint on the classification, and this is plainly apparent at Lord Howe Island where hard basalt sea cliffs are sub-vertical in profile regardless of whether they have basal talus slopes, and are therefore currently eroded solely by subaerial processes, or shore platforms that are generally linked to marine erosion. Hence, on the basis of cliff morphology it is difficult to differentiate the relative contributions of marine and subaerial erosion at Lord Howe Island. An attempt to address the problem is provided in this paper through specific examination of whether gradual subaerial erosion processes alone could have yielded sufficient eroded material to have formed the talus slopes that occur around the
island today. Implicitly, this approach also provides for consideration of the relative contributions of marine and subaerial erosion in gradual, hard-rock cliff retreat.

Douglas et al. (1991, 1994) investigated the specific mechanism by which subaerial processes erode precipitous basaltic sea cliffs along the coastline of Northern Ireland. Unlike weathering processes on gentle slopes that result in soil profiles, erosion on steep basalt freefaces was observed to occur through a continuous evolution of rock properties. First, microfractures develop and are eroded by processes such as swelling, oxidation of iron, hydration, and freeze-thaw. Gradual enlargement of the microfractures then leads to instability, and rock falls occur through gravity. It was further noted that small-scale changes induce instability in larger-scale discontinuities or local overhangs that periodically favour large rock falls, such that the small-scale changes must be regarded as the fundamental cause of all instability in the basalt freefaces.

The probable rate of gradual cliff retreat through subaerial erosion can be inferred from published erosion rates of basalt scarps that are subject only to subaerial erosion processes. In southeast Australia, scarps have generally retreated at rates of 12–28 m/million years (Young 1983; Young & McDougall 1985), although Nott et al. (1996) noted considerably slower erosion rates at the Shoalhaven Gorge (0.3 m/million years). Basalt scarps in central Queensland, Australia, have retreated at rates of up to 130 m/million years (Young & Wray 2000), and rates of 50–95 m/million years have been recorded at Drakensberg, South Africa (Fleming et al. 1999). Very rapid rates of 320 m/million years have been reported from basalt scarps at the Isle of Skye, Scotland, but the data set incorporates erosion following deglaciation 17 500 years BP, and therefore probably includes the effect of frost shattering under periglacial conditions (Hinchliffe & Ballantyne 1999).

Like many sea cliffs, coastal cliffs at Lord Howe Island are likely to be particularly prone to weathering of discontinuities by salt crystallisation and expansion (e.g. Johannessen et al. 1982), whereas freeze-thaw and other periglacial processes do not occur on subtropical Lord Howe Island, and they were probably relatively ineffectual even during the Last Glacial. Hence, while the free availability of sea salts may have engendered more rapid rates of subaerial cliff erosion at Lord Howe Island than that reported from inland sites in Australia and South Africa, the lack of periglacial processes has probably resulted in long-term erosion rates being somewhat slower than those reported for the Isle of Skye. Present erosion rates at Skye are thought to be about 100 m/million years (Hinchliffe & Ballantyne 1999), and it is unlikely that erosion rates at Lord Howe Island would be much faster.

It is quite possible that large landslides/rock falls may have periodically removed sizeable sections of cliff face from around Lord Howe Island, and in this context it is worth noting that the eastern face of Mount Gower above The Big Slope is somewhat embayed, as is the cliff face above the North Ridge talus slope and the talus slope near Little Island. Perhaps these talus slopes have been supplied with material from large-scale mass movements as opposed to talus slopes such as The Little Slope which occur beneath cliff faces that are more or less convex. Regardless of the size of erosion events, material eroded from the face of cliffs falls to the cliff foot where it may accumulate into talus slopes. Accumulation of talus is dependent on several factors, but the most important of these relates to the level of the sea and the relative balance between wave-attrition of talus material and the efficacy of the subaerial processes that supply talus. This paper assumes that talus production occurs reasonably constantly through time, but repeated rises and falls of sea level have periodically altered conditions of talus accumulation. In this context, it is notable that for most (c.85 per cent) of the last 1
million years, sea level has been 25 m or more below its present position (see Figure 6) thereby providing favourable conditions for talus accumulation.

**Cliff height and talus source area**

The size of talus slopes that accumulated during periods of lower sea level depends on factors such as the intensity of subaerial erosion, the nature of the talus and bedrock eroded, and the height of the cliff from which the talus was sourced. Other factors being equal, the highest cliffs would be associated with the largest talus slopes, as high cliffs have a larger source area of cliff face that is exposed to subaerial erosion.

Talus slopes at the base of cliffs have the effect of protecting the cliff toe from wave erosion. Consequently, one might expect to find relationships between coastal morphology and the height of adjacent cliffs, because erosion of higher cliffs yields greater quantities of protective talus. In a number of instances the width of shore platforms has been found to be negatively correlated with cliff height (e.g. Edwards 1941, 1958; Trenhaile 1999), but some sites show considerable ambiguity (e.g. McKenna 1990) perhaps because of landform inheritance. Lord Howe Island is surrounded by cliffs that vary enormously in scale and this provides a clearer illustration of the important role that cliff height can exert on coastal morphology. For instance, in Figure 3, where the average width of shore platforms has been plotted against cliff height, it is apparent that platform width is inversely correlated with cliff height, implying that erosion of the cliff toe is impeded in rough proportion to the amount of talus that falls from adjacent cliffs. However, of more significance is the fact that shore platforms do not occur when cliffs exceed 200 m in height; such cliffs are either plunging or they have cliff-foot talus. It is probable that cliffs greater than 200 m high erode so slowly by marine processes that a platform has not had time to form in the last 6000 years. Furthermore, the absence of erosion over this period would appear to be a reflection of the quantity of protective talus that results from erosion of very high cliffs, because the absence of platforms is not overtly related to variability in rock resistance or the assailing force of waves (Dickson *et al.* 2004).

**Evolution of talus slopes at Lord Howe Island**

It is possible that the talus slopes around Lord Howe Island accumulated wholly within the Last Glacial period when lower sea levels prevented the removal of talus material by marine processes. The rates of gradual subaerial erosion necessary to amass such talus slopes may be estimated, because erosion rate over time is a function of present talus volumes (m$^3$) as well as respective talus source areas (m$^2$). The estimate assumes that the Last Interglacial highstand removed any pre-existing talus, that talus was not removed by marine erosion during interstadial highstands, and that marine erosion during the Holocene removed no more talus than was contributed by subaerial erosion over the same period. It is also apparent that accumulation of talus must have gradually reduced the source areas exposed to subaerial erosion, but minimum and maximum erosion rates can be provided with the accompanying assumptions that only the source area presently above the talus slope was subject to erosion, or that the entire cliff face was subject to erosion throughout the period in question; the true erosion rate must be between these extremes.

Table 1 summarises estimates of the gradual subaerial erosion rates that would have been necessary to amass the talus slopes that occur around Lord Howe Island today.
The size of the talus slope near Little Island results, at least in part, from the partial protection from marine erosion that it has received by the fringing coral reef during the Holocene. As such, it is excluded from the discussion that follows. The remaining estimates imply that all but one of the exposed talus slopes could have accumulated entirely within the Last Glacial period, because erosion rates of 18–59 m/million years are similar to those reported from inland basalt scarps in Australia and South Africa. By contrast, the rapid rates of erosion required to have formed The Big Slope indicate that it probably did not accumulate wholly within the Last Glacial period. It may be that The Big Slope is the only talus slope that pre-dates the maximum of the Last Interglacial. In support of this interpretation is the fact that the only deposits of eolianite preserved on talus slopes around Lord Howe Island occur on the northern end
TABLE 1. Estimates of the subaerial erosion rates necessary for talus slopes to have accumulated between the Last Interglacial and 6000 years BP when the sea attained its current level

<table>
<thead>
<tr>
<th>Location</th>
<th>Present talus source area (m²)</th>
<th>Source area in the absence of talus (m²)</th>
<th>Approximate subaerial volume of present talus slope (m³)</th>
<th>Erosion rate necessary to deposit entire talus slope between 125 ka and 6 ka (m/million years)</th>
</tr>
</thead>
<tbody>
<tr>
<td>The Big Slope</td>
<td>319 660</td>
<td>596 620</td>
<td>76 995 000</td>
<td>129–241</td>
</tr>
<tr>
<td>Talus between The Big Slope</td>
<td>64 051</td>
<td>95 100</td>
<td>3 800 400</td>
<td>40–59</td>
</tr>
<tr>
<td>and King Point</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Talus slope on the western face of King Point</td>
<td>17 230</td>
<td>25 900</td>
<td>567 600</td>
<td>22–33</td>
</tr>
<tr>
<td>The Little Slope</td>
<td>364 680</td>
<td>451 570</td>
<td>7 966 800</td>
<td>18–22</td>
</tr>
<tr>
<td>Little Island talus slope</td>
<td>202 350</td>
<td>428 820</td>
<td>49 530 800</td>
<td>116–245</td>
</tr>
<tr>
<td>North Ridge talus slope</td>
<td>14 400</td>
<td>24 900</td>
<td>699 300</td>
<td>28–49</td>
</tr>
</tbody>
</table>

of The Big Slope in the form of an outcrop 125 m in length that was probably deposited during oxygen isotope stage 5 (Brooke 1999).

It seems unlikely that the eolianite on The Big Slope was emplaced prior to formation of the talus slope, as eolianite would currently occur in greater volumes along the face of the slope and it would be overlain by large quantities of basaltic talus. Equally, if the dunes were deposited on an existing talus slope, the slope could not have been larger than it is now because the eolianite currently occurs at its base, but it is unlikely that it would have been smaller as greater expanses of eolianite would be apparent today. The most likely evolution of The Big Slope is summarised in Figure 4. The conceptual model suggests that during the Penultimate Glacial, and perhaps during preceding glacial periods, talus accumulation increased the size of The Big Slope beyond its present dimensions, whereas in the maximum of the Last Interglacial (oxygen isotope stage 5e) marine erosion trimmed the talus back to near its present size. In subsequent interstadials (substages 5c and/or 5a) sea level was lowered by an amount that allowed large volumes of eolianite to be deposited on the low-lying middle portion of the island (Brooke et al. 2003). It is likely that climbing dunes coated the base of The Big Slope around this time, and over the following 85 000–105 000 years marine processes had little effect on the talus slope as sea levels were considerably lower than at present. Subaerial erosion of cliffs continued, however, depositing a veneer of basalt material on the eolianite dunes. In the current sea-level highstand, talus has been trimmed by marine erosion in a similar manner to that which occurred during substage 5e, and most of the eolianite deposits have been removed.

During the Last Glacial period of lower sea level, talus slopes must have accumulated around the extent of the island’s cliffed shoreline, but only pockets of that talus have persisted to the present day. In order to account for this differential preservation there is a need to consider relationships between the efficacy of marine processes that reduce talus slopes, and subaerial processes that replenish them. This relationship was investigated by contrasting the surface areas of talus slopes and the source areas that occur
above the slopes (see Figure 5). It is notable that the only talus slope in the north of the island (the North Ridge talus) occurs adjacent to its highest peaks. Talus probably accumulated along the length of the North Ridge cliffs during the Last Glacial, but the lower cliffs west of Malabar Hill resulted in smaller accumulations. Hence, during the current highstand the talus slope beneath Malabar Hill has been fed with sufficient
The Development of Talus Slopes

FIGURE 5. Surface area of talus slopes and talus source areas.

material to maintain the slope to the present day (although it may be considerably smaller than it was), whereas other sectors of the North Ridge cliffs have source areas that were unable to sufficiently offset marine attrition of talus during the postglacial transgression and stillstand. The relationship between source area ($14\ 400\ m^2$) and talus slope size ($10\ 000\ m^2$) at the North Ridge may be close to a threshold that demarcates the survival or removal of talus slopes that accumulated during the Last Glacial. It is notable in this respect that on the western side of King Point in the south of the island, a talus slope that has a surface area of $9500\ m^2$ and a source area of about $17\ 200\ m^2$ has been cliffed by marine erosion processes; it may be that this source area/slope size relationship is also close to a threshold that accounts for preservation of talus to the present day.

In summary, in interglacial periods when the sea is at the base of cliffs, there is a critical relationship between the amount of talus that is removed from the cliff toe by marine processes and the amount of material that is supplied by subaerial erosion. At Lord Howe Island, even beneath the loftiest cliffs that yield the largest quantities of talus, marine depletion of talus during sea-level highstands appears to be faster than the supply of material. Indeed, of the talus slopes currently exposed to open-ocean waves, The Big Slope appears to be the only one to have survived the Last Interglacial highstand, while the others accumulated entirely through subaerial erosion during the Last Glacial. During the most recent postglacial marine transgression and highstand, marine erosion has entirely removed talus slopes that accumulated in areas where cliffs are less than 200 m high and where cliff source areas are less than about $15\ 000\ m^2$. 
Island planation and relative contributions by marine and subaerial erosion

Since the Last Interglacial, and perhaps well before that, marine processes have contributed little to overall cliff retreat along stretches of shoreline at Lord Howe Island where cliffs are higher than 200 m. However, the island is now only about 3 per cent of its original area, and current process–form interactions are not necessarily representative of the much longer planation history.

The extensive truncated shelf that surrounds Lord Howe Island provides an indication of the extent of planation that has occurred. There is a distinct break at the edge of the shelf in approximately 60 m water depth that separates the gentle slopes of the shelf and the steeper slopes of the original volcano. The post-eruptive form of the island was presumably a smooth, gently sloping shield volcano. Assuming that the 60 m isobath approximates the position of the original shoreline, and that the volcano had subaerial flank slopes of about 5°, the original height of the volcano may be estimated to have been in the order of 970–1600 m. The highest point on Lord Howe Island today is 875 m, and approximately 6 million years have elapsed since volcanism ceased. This implies that vertical denudation processes have lowered the island at a rate of 15 to 120 m/million years, which is broadly similar to denudation rates reported from basaltic oceanic islands in the Hawaiian group (e.g. 40–190 m/million years at Hawaii, 61–105 m/million years at Lanai, 130 m/million years at Kaneohe Bay; see Li 1988; Wentworth 1927; Moberly 1963). By contrast, most of the coastline has retreated about 10 km, equating to long-term average erosion rates of 1700 m/million years. This implies that vertical denudation has proceeded 14–113 times slower than coastal retreat. Moreover, if one accepts that subaerial erosion of basalt cliffs at Lord Howe Island has probably proceeded at a long-term rate of no more than 100 m/million years, subaerial cliff erosion may account for only 5–10 per cent of total sea cliff retreat, with 90 per cent or more of total retreat attributable largely to marine erosion.

It is likely that wave erosion and clifffing of the newly formed volcano of Lord Howe Island was very rapid initially (cf. Bird & Rosengren 1984). This would have been facilitated by several factors including the depth of water close to the shoreline and the fact that gently sloping flanks erode into low cliffs that yield smaller amounts of protective talus. The first major restriction on rapid clifffing would probably have been the development of wide shore platforms in front of the retreating cliffs. Up until the mid-twentieth century, researchers generally believed that waves could abrade the seabed to great depths (c.180 m) thereby allowing for constant high wave energy and interminable cliff erosion. This view ran contrary to Darwin’s (1844, 1846) observations of Saint Helena some 100 years previously in which he argued that marine processes could not have eroded large volumes of rock beneath sea level while simultaneously depositing a smooth seabed of fine sediment. Dietz and Menard (1951) eventually dispelled notions of wave abradion at great depths and confirmed Darwin’s suggestion that the limit of vigorous wave abradion on the seabed is about 10 m. It is now clear that rather than the continued subsidence of landmasses that Darwin supposed, eustatic sea-level fluctuations provide worldwide conditions suitable for ongoing marine erosion of cliffs (Cotton 1969). Without these sea-level movements, continuous vigorous clifffing is prevented by the development of wide shore platforms; Bradley (1958) and Flemming (1965) agreed that the maximum extent of cliff retreat at the current stillstand is about 500 m.

Figure 6a shows a generalised sea-level curve for the last 1.2 million years together with approximations of the amount of time that the sea has been at different levels. In
The Development of Talus Slopes

235

FIGURE 6. (a) Sea-level curve and oxygen isotope stages for the last 1.2 million years (after Chappell & Shackleton 1986; Shackleton 1987; Chappell et al. 1996; Pillans et al. 1998); (b) cross-section showing shelf break, shelf, fossil reef, and coastline of Lord Howe Island. The grey hatched lines show the percentage of time during the past 1.2 million years that sea level has been at or above present, up to 25 m below present, 25–50 m below present, 50–75 m below present, 75–100 m below present, and greater than 100 m below present.

Figure 6b, the percentage of time that the sea has been at different levels has been transposed onto two cross-sections of the planated shelf around Lord Howe Island. There have been many sizeable sea-level fluctuations over the last 1 million years, but for about 50 per cent of this period, erosion of the island’s shoreline has been restricted as the sea has been below the level of the planated shelf. Assuming that rates of island planation were greatest soon after volcanism ceased, but decreased through time owing to increasing cliff heights and the development of a widening shelf over which wave energy was increasingly attenuated (Menard 1983, 1986), the possibility arises that planation of the shelf may have occurred largely in Tertiary times when sea levels were mainly above present or less than about 75 m below present (Vail & Hardenbol 1979; Haq et al. 1987; Pillans et al. 1998). Moreover, the current average level of the shelf (c.30–50 m) would appear to be a manifestation of the integrated level of the sea during the period in which the shelf was cut.

Further clues regarding the history of planation are provided by an extensive fossil reef that almost encircles the island in 30 m water depth. It is possible that the reef formed after the shelf was planated, as wave energy on the landward side of the reef would appear to be incapable of eroding bedrock to the depths observed between the fossil reef and the modern coastline. Likewise, the high cliffs that surround Mount Gower seem likely to have eroded prior to establishment of the fossil reef as many of the cliffs currently plunge into deep water and are resistant to wave erosion, while during a lower sea level in which the cliffs might be subject to erosive breaking waves, much of the wave energy arriving at the coastline would have been attenuated across the shallow fossil reef. The origin and age of the fossil reef are problematic as efforts to date material obtained from it were unsuccessful, but it has been suggested that it probably formed at a time in which much more prolific coral growth was possible at this
latitude than during the Holocene or the late Quaternary (Kennedy et al. 2002). At a minimum the fossil reef appears to be older than oxygen isotope stage 5e, but it may be considerably older.

**Summary**

Marine processes may account for over 90 per cent of the total cliff retreat that has occurred at Lord Howe Island, yet contemporary coastal morphology attests to the significance of subaerial processes in recent times. Marine cliffing was probably particularly rapid soon after volcanism ceased as there was deep water close to the shoreline, low cliffs, and repeated sea-level movements. However, erosion rates decreased as the coastline receded towards the summit of the volcano with higher cliffs yielding greater quantities of protective talus, and wave energy becoming increasingly attenuated across a widening planation surface. At some point an extensive coral reef formed around the island and provided considerable protection from marine erosion. Today, in the penultimate stages of island planation, marine erosion occurs only in those areas where cliffs are less than about 200 m high. Some cliffs higher than 200 m plunge into deep water such that erosion by breaking or broken waves is very limited, whereas protective talus slopes occur in front of high cliffs where talus source areas are in excess of 15 000 m². While the rate of marine cliff erosion may have decreased through time, the rate of subaerial cliff erosion has probably been more consistent such that subaerial processes have assumed greater relative importance in recent times. During the Last Glacial, when the level of the sea was much lower than at present, subaerial cliff erosion resulted in the accumulation of talus material around the extent of the island’s coast. However, in the postglacial transgression and current highstand, wave erosion removed talus from all sectors of coast where source areas were not sufficiently large to offset marine attrition of talus. On comparison with published rates of retreat of inland scarps it appears that most of the talus slopes around Lord Howe Island could have accumulated wholly within the Last Glacial. By contrast, the gross morphology of The Big Slope and the presence of preserved eolianites on it imply that this talus slope pre-dates the Last Interglacial and it may be considerably older.

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