Deep-Sea Research II 58 (2011) 899–908

Contents lists available at ScienceDirect

Deep-Sea Research II

journal homepage: www.elsevier.com/locate/dsr2

The geomorphology of the flanks of the Lord Howe Island volcano, Tasman Sea, Australia

D.M. Kennedy [1,*], B.P. Brooke [2], C.D. Woodroffe [3], B.G. Jones [3], C. Waikari [4], S. Nichol [2]

[1] Department of Resource Management & Geography, The University of Melbourne, Parkville Vic 3010, Australia
[2] Marine and Coastal Environment Group, Geosciences Australia, GPO Box 378, Canberra ACT 2601, Australia
[3] School of Earth and Environmental Sciences, University of Wollongong, Wollongong NSW 2522, Australia
[4] School of Geography, Environment & Earth Sciences, Victoria University of Wellington, PO Box 600 Wellington, New Zealand

ARTICLE INFO

Article history:
Received 27 October 2010
Accepted 27 October 2010
Available online 18 November 2010

Keywords:
High Island
Volcano
Tasman Sea
Lord Howe Rise
Slump
Isotope

ABSTRACT

The flanks of mid-ocean volcanoes are inherently unstable features especially in the constructional phase of development when the volcano is active. Lateral and vertical stresses are placed on the volcanic edifice as it builds, with the flanks continuing to be unstable up to at least 1 Ma after volcanism has ceased. The flanks of the Lord Howe Island volcano record this period of greatest instability and a subsequent period in which marine and subaerial erosion have dominated its geomorphic evolution. Lord Howe Island lies in the Tasman Sea of the Southern Pacific region and is the subaerial remnant of a Miocene mid-ocean volcano. The island has only recently entered reef building seas and therefore has been subject to marine erosive processes over the past 5–6 Ma. The island is unique as it sits on the stable drowned continental crust of the Lord Howe Rise rather than oceanic crust like many other mid-plate basaltic islands. Multibeam sonar bathymetry data were collected to a depth of 3500 m where the island flanks grade into the surrounding planar sea floor. Several slump features are evident, the largest being over 130 km^2 in area. These features are inferred to be old (late Tertiary) based on an extensive cover of marine sediment as indicated by low multibeam backscatter intensity and subdued topography. Most likely the slumps formed during the immediate post-eruptive stage of volcano evolution, before the bulk of the subaerial portion of the volcano was removed by marine erosion. Flank processes are now dominated by the deposition of carbonate sediment composed of mollusc and foraminiferal remains. Based on radiocarbon and stable isotope analyses of a sediment core (760 m depth) collected on a trough in the centre of the volcanic edifice, Quaternary sediment was likely deposited predominantly during glacial periods. The erosional morphology, sediment cover and tectonic stability of the region suggest that the flanks of the volcano are at present relatively stable.

1. Introduction

Most mid-ocean volcanic islands and seamounts are the result of hotspot eruptive processes and their irregular shape is often an outcome of subsequent marine erosion and deposition. Volcanic islands and seamounts are inherently unstable due to the formation of steep slopes as a result of the successive build up of erupted material. This is compounded by eruptions often repeatedly occurring along defined volcanic vents which place lateral stress on the volcanic edifice through the formation of dyke complexes (Carracedo, 1994). Instability and erosion are also induced by eruptions occurring along radiating and flank rift zones, caldera collapse, slope oversteepening induced by volcanic intrusions, the presence of low-strength layers (e.g. hyaloclastite deposits), coalescence of volcanoes and seismic movement (Mitchell, 1998; Bulmer and Wilson, 1999; Chaytor et al., 2007).

The net effect of these stresses on the volcanic edifice is that lateral slides are common, with mass wasting processes being an integral element of island evolution (Carracedo, 1994). Such mass wasting events can be catastrophic. The classic example is the collapse of the southern flank of Kilauea, Hawaii, where a slump of approximately 5200 km^2 has been mapped up to 50–200 km off the present coastline (Moore et al., 1989; 1994). Large landslides have also been reported from the Canary Islands (Masson, 1996; Carracedo et al., 1999, Ablay and Hürlimann, 2000; Hürlimann et al., 2004; Coppo et al., 2009; Llanes et al., 2009) and Cape Verde Islands (Day et al., 1999), with such events appearing to be common in the post-eruptive phase of oceanic basaltic volcanoes (Siebert, 1984). The hazards posed by such events are significant, especially when considering any resultant tsunami that a displacement may cause (e.g., Waythomas et al., 2009). Coral deposits on Molokai and Lanai Islands, Hawaii, have been related to the collapse of the south flank of Kilauea, an event known as the Lanai Tsunami (c. 220 ka),

* Corresponding author.
E-mail address: davidmk@unimelb.edu.au (D.M. Kennedy).

0967-0645/$ - see front matter © 2010 Elsevier Ltd. All rights reserved.
doi:10.1016/j.dsr2.2010.10.046

although some controversy exists around the tsunamigenic origin of these sediments (Felton, 2002; Crook and Felton, 2006). Smaller slides can occur on centennial scales which also cause large wave events. For example an earthquake in 1975 in Hawaii, resulted in the seaward translocation of an area of seafloor 70 km long and 30 km wide by up to 8 m horizontally and 3.5 m vertically (Lipman et al., 1985; Morgan et al., 2000). This slide resulted in a tsunami estimated to be 14.6 m high on the Halape–Apua Point coast of Hawaii (Goff et al., 2007).

Landslides on volcano flanks appear to be very common during the constructional phases of island development. Little, however, is known about how these systems evolve after volcanism has ceased and where post-emplacement processes such as marine and subaerial erosion, sediment deposition and tectonic movement dominate landscape evolution (Mitchell et al., 2002). Marine and subaerial processes abrade the shoreline and dissect the high parts of the island respectively, termed skeletonisation by Cotton (1969); however, antecedent factors such as the age of the crust on which the island sits may also affect geomorphic evolution. For example, eruptions through older seafloor may produce volcanoes that are less stable than those formed on younger plates because lava is emplaced on marine sediment that overlies the plate, which can induce flank instability as the island grows (Bulmer and Wilson, 1999). In addition, the mass of the island and crustal buoyancy will influence the rate of island subsidence and therefore its exposure to erosive processes at and above sea level (Menard, 1983; Nakada, 1986).

Lord Howe Island is a mid-plate basaltic island that lies at the transition between temperate and tropical marine carbonate sedimentation. Volcanism that formed the island ceased around 6 Ma (McDougall et al., 1981) and since this time marine erosion has formed a shelf around the island remnant approximately 20 km wide (Kennedy et al., 2002). The island also sits on top of a section of submerged continental crust called the Lord Howe Rise (Fig. 1). The shelf is presently the locus of carbonate sediment accumulation, coralline algal sediments dominating the modern shelf, with reef formation occurring in shallow (<50 m) waters. An incomplete history of carbonate accumulation over the past 300 ka is preserved in calcarenite on the island (Brooke et al., 2003a, b), which indicates an increasing contribution from tropical carbonates (Kennedy and Woodroffe, 2000; Woodroffe et al., 2005, 2006). This study examines the submarine flanks of the island to better understand slope evolution after volcanism has ceased. Current flank stability is particularly important to assess because flank slumps may pose a tsunami hazard to the coasts of eastern Australia and western New Zealand.

2. Regional setting

Rifting of the Tasman Sea began during the Late Cretaceous (c. 85 Ma), representing the final stage of the breakup of Gondwana, with most active seafloor spreading occurring between 60–80 Ma (Willcox et al., 1980). As New Zealand separated from Australia a fragment of continental crust also detached and subsequently drowned to form the present day Lord Howe Rise (Stagg et al., 2002; Exon et al., 2007). Lord Howe Island (31° 33′S, 159° 04′E), located approximately 500 km east of the Australian coast, sits on the margin of this foundered continental crust. The island and Balls Pyramid, located 20 km to the south, form the subaerial remnants of the Miocene Lord Howe volcano. These islands are the only subaerial exposures in the Tasman Sea of a north–south trending line of seamounts related to Tertiary hot-spot volcanism (Quilty, 1993) (Fig. 1). The subaerial geology of Lord Howe Island is dominated by alkaline basalt, hawaiites, and tuffs (Standard, 1963; McDougall et al., 1981), while at lower (<100 m) elevations

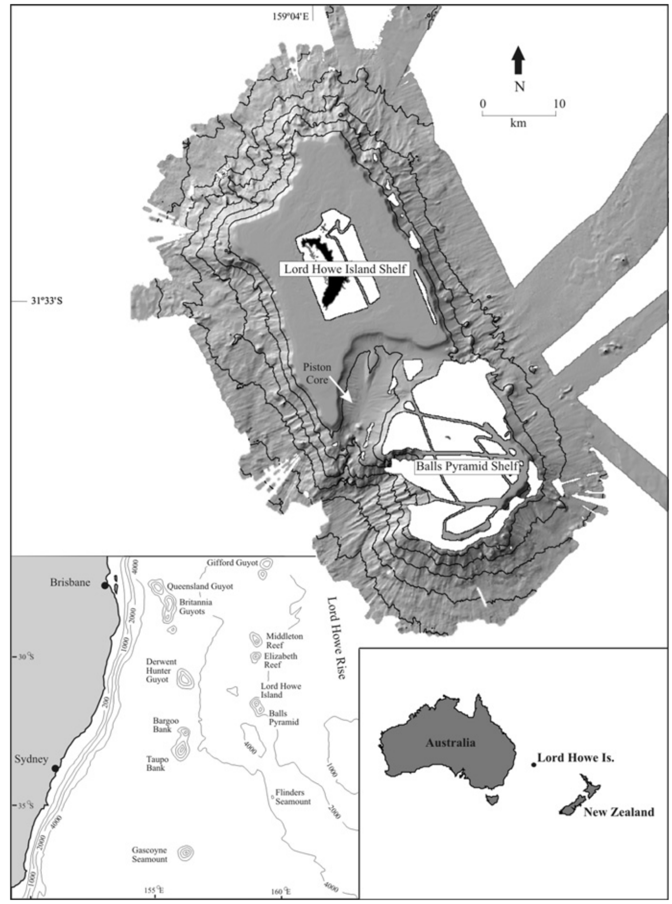

Fig. 1. Location of Lord Howe Island and Balls Pyramid on the western edge of the Lord Howe Rise within the Tasman Sea. The island shelves and flanks are presented as a hillshade view of the dataset in 100×100 m cells. Bathymetry is at 500 m intervals.

a veneer of Quaternary eolianite has been deposited during sea-level highstands, dating back to at least oxygen isotope stage 7 (Brooke et al., 2003a; b). The oldest lithologies on the island are the Roach Island Tuff, which is overlain by a series of tholeiitic lava flows of the North Ridge basalt, the latter dating around 6.6-7.2 Ma (McDougall et al., 1981). Caldera collapse is inferred to form the next youngest unit, the Boat Harbour Breccia, with the final stage of volcanism being represented by the Mount Lidgbird basalt (c. 6.4 Ma) a caldera infill composed of horizontally-bedded alkaline olivine basalts and hawaiite (McDougall et al., 1981). Balls Pyramid is also composed of a similar mid-ocean basaltic sequence and although it has not been mapped in detail it is younger than Lord Howe Island.

Marine erosion of these basalts has dominated island evolution since the cessation of volcanism (McDougall et al., 1981), with Balls Pyramid representing the penultimate stage of marine planation (Woodroffe et al., 2006). This has led to the development of a rhomboidal shaped shelf around Lord Howe Island, 24 km wide and 36 km from north to south, and an oval shelf around Balls Pyramid, 15 km wide, and 22 km from north to south. Both of these shelves have an average water depth of 50 m across their surface and are relatively flat ($<1°$) (Kennedy et al., 2002; Fig. 1). The Australian Plate on which the islands sit is moving north at 5-6 cm/yr (Quilty, 1993), implying that only since the Middle Pleistocene have they moved into an oceanographic environment which supports sub-tropical and tropical carbonate sedimentation (Woodroffe et al., 2005). At present, subtropical coralline algae dominate the shelf sediments, with coral reefs restricted to the shallow margin of Lord