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Published in:
Australian Journal of Earth Sciences

Document Version:
Peer reviewed version

Queen's University Belfast - Research Portal:
Link to publication record in Queen's University Belfast Research Portal

Sedimentology, structure and age estimate of five continental slope submarine landslides, eastern Australia

S. Clarke[a], T. Hubble[a], J. Webster[a], D. Airey[b], E. De Carli[a], C. Ferraz[c], P. Reimer[d], R. Boyd[e,f], J. Keene[a], and Shipboard Party SS12/2008

[a] Geocoastal Research Group, School of Geosciences, University of Sydney, Sydney, NSW 2006, Australia

[b] School of Civil Engineering, University of Sydney, Sydney, NSW 2006, Australia.

[c] Divisão de Geologia e Georecursos Marinhos, Instituto Português do Mar e da Atmosfera (IPMA), Rua C do Aeroporto 1749-077, Lisboa, Portugal.

[d] Centre for Climate, the Environment & Chronology ([14]CHRONO), Queen's University, Belfast BT9 6AX, UK

[e] University of Newcastle, Callaghan, NSW 2308, Australia.

[f] ConocoPhillips, Houston, TX 77252-2197, United States.

Title of Team: Shipboard Party SS12/2008

samantha.clarke@sydney.edu.au

T +61 2 9114 1425

M +61 (0)410 106 196

Abstract

Sedimentological and Accelerator Mass Spectrometry (AMS) ^{14}C data provide estimates of the structure and age of five submarine landslides (~0.4–3 km^3) present on eastern Australia's continental slope between Noosa Heads and Yamba. Dating of the post-slide conformably deposited sediment indicates sediment accumulation rates between 0.017 mka^{-1} and 0.2 mka^{-1}, which is consistent with previous estimates reported for this area. Boundary surfaces were identified in five continental slope cores at depths of 0.8 to 2.2 m below the present-day seafloor. Boundary surfaces present as a sharp colour-change across the surface, discernable but small increases in sediment stiffness; a slight increase in sediment bulk density of 0.1 gcm^{-3}, and distinct gaps in AMS ^{14}C ages of at least 25 ka. Boundary surfaces are interpreted to represent a slide plane detachment surface but are not necessarily the only ones or even the major ones. Sub-bottom profiler records indicate that: 1) the youngest identifiable sediment reflectors upslope from three submarine landslides terminate on and are truncated by slide rupture surfaces; 2) there is no obvious evidence for a post-slide sediment layer draped over, or burying, slide ruptures or exposed slide detachment surfaces; and 3) the boundary surfaces identified within the cores are unlikely to be near-surface slide surfaces within an overall larger en masse dislocation. These findings suggest that these submarine landslides are

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geologically recent (<25 ka), and that the boundary surfaces are either: a) an erosional features that developed after the landslide, in which case the boundary surface age provides a minimum age for the landslide; or b) detachment surfaces from which slabs of near-surface sediment were removed during landsliding, in which case the age of the sediment above the boundary surface indicates the approximate age of landsliding. While an earthquake triggering mechanism is favoured for the initiation of submarine landslides on the eastern Australian margin, further evidence is required to confirm this interpretation.

Keywords: mass-failure, multibeam, seafloor geomorphology, continental margin, southeast Australia, continental slope, passive margin, sedimentation rates, submarine landslide

Introduction

Submarine landslides are common features along continental slopes and oceanic islands (Hampton, Locat & Lee, 1996; Lee, 2009). Their volumes range over 5 orders of magnitude from small shallow slides of less than 0.1 km^3 to more than 2400 km^3, such as the Storegga slide off the Norwegian coast (Haflidason et al., 2004; Masson, Harbitz, Wynn, Pedersen & Lovholt, 2006). The larger slides are generally thought to be capable of generating damaging or catastrophic tsunami (Harbitz et al., 2013) and the suggested landslide triggers include earthquake loading (Fine, Rabinovich, Bornhold, Thomson & Kulikov 2005), pore pressure effects (Locat & Lee, 2002; Masson et al., 2006), gas generation (Maslin, Mikkelsen, Vilela, & Haq, 1998; Sultan, Cochonat, Foucher, Mienert & Sejrup, 2004), storm waves (Prior & Coleman, 1984), and rapid sedimentation (Masson et al., 2006). While a number of slides have been identified and examined in detail, for example, the Storegga Slides in Norway (Haflidason et al., 2004), the Brunei Slide in Borneo (Gee, Uy, Warren, Morley & Lambiase, 2007), the Goleta Slide in California (Greene et al., 2006), slides in Angola (Gee et al., 2006), the Gulf of Mexico (Silva, Baxter, LaRosa & Bryant, 2004), the Hawaiian Islands (McMurtry et al., 2004), Canary Islands (Masson et al., 2006), and slides along the Hikurangi Margin in the Southwest Pacific Region (Lamarche, Joanne & Collot, 2008), the physical processes that generate and facilitate submarine landslides are not well-constrained or understood (Bardet, Synolakis, Davies, Imamura & Okal, 2003; Locat & Lee, 2002; Mosher et al., 2010; Urlaub, Talling, & Masson, 2013). One of the principal reasons for the lack of definitive explanations for this phenomenon is the relative dearth of data on the physical and mechanical properties of the sediments, particularly sediments representative of the failure surface, as few examples of sediments of this type have been recovered (Urlaub et al., 2013).

Jenkins and Keene (1992) first identified submarine landslides on the eastern Australian continental margin, south of Sydney, in GLORIA swath maps produced in the late 1980's. Subsequent mapping of this margin with the next generation of high resolution, multibeam echo-sounding equipment between 2006 and 2013, has provided sufficient information and morphological data to establish that there are a surprisingly large number of submarine landslides evident on the continental slope (Boyd et al, 2010; Glenn et al, 2008; Hubble, Airey, Sealey, De Carli & Clarke, 2013) given this margin's well-recognised low sedimentation rates (Boyd, Ruming & Roberts, 2004) and passive margin tectonic setting. Instead of the relative quiescence and stability that might be expected, recent studies have demonstrated that submarine sliding should be considered to be a common and ongoing characteristic of the continental slope offshore eastern Australia (EA) south of Fraser Island (Boyd et al., 2010; Clarke et al., 2012; Hubble et al., 2012).

This paper presents a study of upper continental slope sediments collected from the EACM and is part of a larger body of work intended to determine the frequency and consequences of submarine

landsliding in this part of the margin (c.f. Boyd et al., 2010; Clarke, 2014; Clarke et al., 2012; Hubble et al., 2012, 2016). This paper aims to establish the geological and sedimentological characteristics of the materials in which geologically recent (<25 ka) submarine landsliding has occurred (Boyd et al., 2010; Clarke et al., 2012) and examines material sampled from ten gravity cores from the upper slope between 2–5 m long that were collected onboard *RV Southern Surveyor* in November 2008 (*SS2008-V12*; Boyd et al., 2010). Five cores that contain boundary surfaces are identified by a sharp, colour-change boundary, small increases in sediment stiffness, slight increases in sediment density, and distinct gaps in AMS ^{14}C age of at least 25 ka. These boundary surfaces are interpreted to represent detachment surfaces or slide-plane surfaces. We interpret sediment below the boundary features to have been previously buried to depths between 50–220 m prior to detachment of the overlying material with the sediment above the boundary surface represents recent sediment drape. We report sedimentological data and AMS ^{14}C isotopic dates for the cored sediments and then interpret these data in the context of the morphology of the landslides, as evident in the multibeam bathymetry and constrained by sub-bottom profiler transects.

Study area

The study area is located along the EACM (Figures 1, 2), offshore from Noosa Heads in southern Queensland and Yamba in northern New South Wales). It is situated ~ 30 to 70 km seaward of the present coastline in water depths of 150–4500 m (Figure 2).

Five slides have been identified in the study area: 1) Bribie Bowl Slide (centre of mass water depth 600 m); 2) Coolangatta-2 Slide (centre of mass water depth 900 m); 3) Coolangatta-1 Slide (centre of mass water depth 600 m); 4) Cudgen Slide (centre of mass water depth 600 m); and 5) Byron Slide (centre of mass water depth 800 m) (see Figures 2–4; Table 1). These features are representative of slope failures that occurred in the two dominant slope morphologies present in the study area (see Figure 3) identified by Boyd et al. (2010), Clarke et al. (2012), and Clarke (2014): a) the relatively steep (3–7°) and canyon-incised slope (Bribie Bowl Slide and Bryon Slide); and b) the relatively gentle slope (1–3°) of the Nerang Plateau (Coolangatta-1, -2 slides and Cudgen Slide). Ten gravity cores from the upper continental slope (<1200 m) were examined in this work (Figure 2), with at least one gravity core recovered from each of the five submarine landslide scars investigated.

Of the ten cores examined, seven (GC2, GC3, GC5, GC6, GC8, GC11, and GC12) were collected from five submarine landslide scars (see Figures 2, 3; Table 1) and three (GC1, GC4, and GC7) were recovered from adjacent sites in slopes that do not present obvious slide features or morphologies.

Geologic setting and margin structure

The EACM represents a distinct oceanographic, climatic and geological province (Boyd et al., 2004). It extends 1500 km from the Great Barrier Reef to Bass Strait and forms the western continental physiographic boundary of the Tasman Sea. The margin is relatively narrow and typically presents a steep, thinly sedimented upper and middle slope with continental bedrock exposed on its lower slope (Figure 5) (Boyd et al., 2010; Keene, Baker, Tran & Potter, 2008; Marshall, 1978, 1979). The present day morphology of the slope is dominated by erosion and mass wasting (Boyd et al., 2010; Clarke et al., 2012; Glenn et al., 2008; Hubble et al., 2012).

The continental shelf is 20 km wide offshore from Byron Bay in the south of the study area and gradually increases to 70 km offshore from Noosa Heads in southern Queensland in the north. The shelf break is located an average distance of 50 km from the shore at depths between 100 and 150 m. A bedrock high may be present beneath the shelf break but, more commonly, the bedrock basement surface is located beneath the outer shelf and upper slope presents as an inclined,

irregular surface that drops away to the east and joins the oceanic basement reflector at the base of the continental rise (Figure 5).

The slope of the EACM is relatively steep at the shelf break compared with typical (<5°) Atlantic-style passive margins (McAdoo, Pratson & Orange, 2000) and presents average slopes of 5–10° in water depths ranging from 150 m at the shelf break to 4500 m at the abyssal plain (Boyd et al., 2010). The relatively steep inclination of the slope is ascribed to the combination of asymmetric passive-margin rifting and low rates of sedimentation (Boyd et al., 2004). The original continental rift that preceded sea-floor spreading in the Tasman Sea Basin between 74 and 52 Ma before present (Gaina, Muller, Brown & Ishihara, 2003) is thought to have breached on its western flank (Falvey, 1974; Lister, Etheridge & Symonds, 1986; Shaw, 1978) resulting in a particularly steep continental slope. The lack of sediment on the lower slope is suggested to be a consequence of the absence of major river systems delivering material to the slope and submarine erosion during the Neogene (Hubble et al., 2012; Keene et al, 2008). In any case, the slope is generally regarded to be sediment deficient, especially when compared with many other passive margins of similar age around the world (Boyd et al., 2004, 2010).

The upper continental slope sediments are mostly deposited as a thin layer of material above continental basement rocks (see Figure 5). These deposits are generally less than one-kilometre thick and commonly less than 500 m thick (Conolly, 1969; Keene et al., 2008; Ringis, 1972), which is thin in comparison with other passive margin deposits such as those of the North Atlantic (Heezen, 1974) whose sediments are commonly an order of magnitude thicker. Previous studies are few, but they consistently indicate that these upper slope sediments are mixed siliciclastic-carbonate muds (Glenn et al., 2008; Hubble & Jenkins, 1984a, b; Hubble et al., 2012; Troedsen & Davies, 2001).

Only one published study on four cores of continental slope sediments from the Eastern Seaboard of Australia (depth range 350–2500 m) is currently available (Troedsen & Davies, 2001), and only 40 cores, all obtained by gravity corer and <5 m in length, have been described for the ~1500 km length of continental slope (Glenn et al., 2008; Hubble & Jenkins, 1984a, b; Troedsen, 1998; Troedsen & Davies, 2001). None are reported to have penetrated slide surfaces, although the cores collected by Glenn et al. (2008) were taken as part of a submarine landslide study. Sedimentation rates range from 0.01–0.02 m/ka (long-term estimates based on current margin sediment thickness and the margin's age) to 0.02–0.24 m/ka (directly determined rates from the limited core samples; Troedson & Davies, 2001). These rates are based on limited data and further investigation is needed to produce a more robust pattern of sedimentation rate variation for the margin.

Data and methods

Three types of data were interpreted for this study: 1) high-resolution multibeam bathymetry data, sub-bottom profiling data and archived seismic reflection data; 2) sediment recovered from the margin; and 3) radiocarbon dating.

Bathymetry

Approximately 13 000 km^2 of bathymetric data were acquired using a 30-kHz Kongsberg EM300 multibeam echosounder (Boyd et al., 2010). The multibeam data were processed to produce a 50 m gridded digital elevation model (DEM) covering the region investigated (Figure 2). Using *Fledermaus V7.3.3b* software (http://www.qps.nl/), the DEM was used to examine the continental slope in the study area and the five individual slide sites from where the 12 gravity cores were taken.

Landslide thickness (*t*), length (*L*), width (*W*), and water depth at landslide centre of mass (h_0) were determined for each of the selected features. Landslide thickness is the maximum thickness within

the landslide scar assuming the surface is continuous without the apparent landslide feature (McAdoo et al., 2000). Landslide length is the distance from landslide head to landslide toe. Landslide width is the average of a set of width measurements determined at 500 m intervals perpendicular to the landslide's downslope axis. Water depth is determined at the landslide's centre of mass. These parameters define the size of each submarine landslide and amount of sediment available during failure. They also relate to the size of tsunami that could be generated.

Submarine landslide morphology is described and classified following the definitions and nomenclature in McAdoo et al. (2000) and summarised in Figure 6.

Sub-bottom profiles

Sub-bottom profile data were acquired in water depths less than ~1500 m using a *Topas PS18* parametric sub-bottom profiler (15–21 kHz). The raw *Topas* data were processed and converted into SEG-Y format and displayed using *SeiSee V2.16.1* software (http://www.dmng.ru/seisview/). Two-way travel time was converted to water depth and sediment thickness using a constant sound velocity of 1500 ms^{-1}.

Sub-bottom profile penetration depended on the weather conditions and on the response of the seabed. Performance was best between 100 to 500 m water depth, but adequate performance was recorded in water depths up to 1500 m. Seismic penetration ranged from ~40 m and up to 100 m commonly in shallower water (<1200 m) and, in deeper water (1200–2000 m), when the sea state and swell direction were favourable.

Core collection and sediment properties

Coring was carried out using a gravity corer with a 1000 kg head-weight and 5 m long steal barrel containing plastic liners. Core sampling locations were chosen after inspection of the multibeam bathymetric data and sub-bottom profiles in order to: 1) identify submarine landslide scars (Boyd et al., 2010), and 2) examine the stratigraphy of the slide scars and their adjacent slopes. Despite limited seismic penetration (usually ≤50 m), various slide characteristics, morpho-structural features, layer geometry, and the physical removal of slabs of layered sediment can be inferred from the sub-bottom profiles. Figure 7a shows the location of the sub-bottom profiles.

Sediment cores were cut into one-metre lengths, split longitudinally, photographed, logged, and sub-sampled at regular intervals between 10–40 cm spacing, depending on test type. Additional samples were taken on either side of lithological boundaries. Two cores were left unsplit for geotechnical testing.

Grain size and mean grain-size distribution were determined using a laser particle sizer (Malvern Mastersizer 2000) using standard operating procedures (Malvern Instruments, 1999), and statistically analysed using GRADISTAT 8.0 (Blott & Pye, 2001). Distributions for each sample represent the average of the three sample runs. Grain-size subsamples were taken at ~ 30–40 cm intervals.

Carbonate and organic carbon contents were determined using the sequential loss on ignition (LOI) method, which gives results to within a maximum error of 2% (Heiri, Lotter & Lemcke, 2001). Average carbonate and organic carbon contents were determined for each 1 m core section, based on three equi-spaced samples taken per metre. High-resolution changes were determined for three cores (GC8, GC11, GC12), where ten even-spaced samples were taken from each 1 m core section (c.f. Clarke, 2014).

Dry and bulk density, unit weight, void ratio, and water content were determined by testing sediment samples of a known volume using classical, soil mechanics testing methods and techniques

(c.f. ASTM methods; Head, 1982) that are fully described in Clarke (2014). Three samples were taken from each 1 m core section. For simplicity, only unit weight values are shown on included core logs (see Figures 8, 9), however these values directly correlate to the dry and bulk density, void ratio, and water content (c.f. Head, 1982).

^{14}C radiocarbon dating

Radiocarbon dates were determined by accelerator mass spectrometry (AMS) on 49 samples taken from eight gravity cores (GC1, GC2, GC5, GC6, GC7, GC8, GC11 and GC12; see Table 2). Dates were obtained from both bulk sediment samples and planktonic foraminifera assemblages extracted from the hemipelagic sediments. Twenty five bulk and 24 foraminifera sub-sampled radiocarbon ages were determined. Samples were run at three laboratories: 1) the CHRONO Centre, Queen's University, Belfast UK (Lab ID: UBA); 2) Australian Nuclear Science and Technology Organisation (ANSTO), Lucas Heights, Australia (Lab ID: OZP); and 3) Radiocarbon Laboratory, University of Waikato, New Zealand (Lab ID: Wk).

Conventional ^{14}C yrs BP were converted into calibrated calendar ages following Stuiver and Reimer (1993). Median calibrated ages (BP) were calculated with CALIB V6.1.1 (Stuiver, Reimer & Reimer, 2005) using marine calibration curve Marine09.14c data set (Reimer et al., 2009) with a reservoir correction (ΔR) value of 11 ± 13 yr (the average for eastern Australia; see http://calib.qub.ac.uk/marine/; Ulm, Petchey & Ross, 2009) and reported here with 2σ errors. Some dates could not be calibrated as they fell outside the calibration range (0–50 ka). Results are shown in Table 2.

Ages from GC1 and GC2 were specifically taken to determine a more robust sedimentation rate for the continental slope in this region, while ages from GC8, GC7, GC11, and GC12 were taken primarily to investigate boundary surfaces; however all ages are useful in constraining sedimentation rates.

Planktonic foraminifera assemblage samples

Extraction of suitable planktonic foraminifera was conducted by disaggregating each core slice sample of ~1–4 cm of sediment in water and washing over a 63 µm wet sieve. The coarser fraction was dried at 40ºC, weighed and used for foraminiferal analysis. A minimum of ~6 mg of well-preserved planktonic foraminifera shells were identified and hand-picked under a binocular microscope and included in each sample (e.g. Murray, 1991). Pristine foraminifera shells that showed minimal or no indication of erosion or abrasion through transport were selected from the coarse fraction bulk sediment samples, thereby minimising disparity between death of skeletal organism and time of deposition (Woodroffe, Samosorn, Hua & Hart, 2007), and foraminifera containing secondary cements or chamber infilling were rejected. For the identification of foraminifera we followed the methods of Loeblich and Tappan (1988) and Jones (1994). The remains of planktonic foraminifera from the mixed surface layer of the water column were favoured for dating due to uncertainties in the deep-water reservoir correction associated with dating benthics.

A mixed assemblage of planktonic foraminifer species (maximum 2 to 3 species; polyspecific sample) were picked for each sample, consisting primarily of *Globigerinoides ruber and Pulleniatina obliquiloculata*. Where sufficient material was available, single specie (monospecific) samples were used, reducing the possibility of different species giving different ages. Species selection was consistent with foraminifera used in previous sedimentological studies from the east Australian margin (Troedson & Davies, 2001, samples located ~ 100 km north of the study area; and Glenn et al., 2008, samples located ~ 700 km south of the study area). Planktonic foraminifera species *Neogloboquadrina, Pulleniatina, Globigerinoides rubber, Globigerinoides* spp. and *Globoquadrina* spp. were used for radiocarbon dating.

Results

Bathymetry and sub-bottom Profiles –- relationship of core locations to slide morphology

Four cores are located within, or adjacent to, slide features from the steep canyon dissected slopes (3–7°) of this segment of the margin (Bribie Bowl Slide, north – 2 cores within the slide, 1 core adjacent to the slide; Byron Slide, south – 1 core). Eight cores are located within, or adjacent to, slide features on the southern end of the gently sloping, dissected Nerang Plateau (1–3°) (Coolangatta-2 Slide – 1 core; Coolangatta-1 Slide – 1 core; and Cudgen Slide – 4 cores within the slide, 2 cores adjacent to the slide) (see Figures 2, 3; Table 1). Multibeam bathymetry indicates that the slides sampled are dominantly translational except for the Cudgen Slide, which presents sub-bottom profile features more typical of rotational slides (i.e. the area of disturbed compressed sediment evident in Figure 10b). It is also apparent that the Cudgen Slide presents morphologic features indicating that near-surface slabs have detached (see Figures 10, 11 and description below). The Bribie Bowl and Byron slides within the canyon regions are significantly thicker (>100 m) in comparison to those developed on the adjacent Nerang plateau (~20–50 m thick).

The translational slides are retrogressive features with headwall scarps exposing layers that are "unsupported" (i.e. truncated at the headscarp). In map view the crown/headwall scarps present a distinctive semicircular shape (see Figure 2) characteristic of circular failures (c.f. Varnes, 1978). Detached slide slabs are slope-parallel planar blocks. Sub-bottom profiles indicate that the surficial seismic reflectors, upslope of the slide, are truncated by the slide rupture surfaces (see Figures 7, 10, 11). Undeformed, parallel-bedded sediment reflectors are observed upslope of the headscarp, while contorted/distorted sediment reflectors are observed downslope (see Figure 11).

Sub-bottom profiles show no clear evidence (such as continuous overlaying reflectors) of a post-slide sediment drape burying the slide ruptures or exposed slide detachment surfaces; i.e. any post-slide sediment layer is too thin to detect in the sub-bottom profile records. This indicates that these translational slides are either a) relatively young, b) the upper slope has experienced a continual erosive removal of material since the occurrence of the slide, maintaining exposure of the detachment surface, and/or c) the sedimentation rate is too slow to deposit detectable post-slide sediments.

Bribie Bowl Slide (Cores GC1, GC2, GC3)

GC1 is a reference core, taken adjacent to the Bribie Bowl Slide, whereas GC2 and GC3 are located within the slide scar (Figure 2a). The Bribie Bowl Slide is a translational slab failure located within the northern steeper canyon region (3–7° slope) and presents a thicker slide (>100 m) in comparison to those developed on the adjacent Nerang plateau. The Bribie Bowl Slide has an average slope of ~12° along the majority of the seafloor surface, increasing to 33° in the crestal amphitheatre region at the head of the slide scar.

Approximately 8 km north of the Bribie Bowl Slide, sub-bottom profile line L58a (Figure 7b) shows the subsurface sediment layering geometry of the upper slope between 300 and 1100 m. The line crosses a 100 m high escarpment, which is interpreted to be the crown headscarp of a slide, at ~800 m water depth. This escarpment is visible in the bathymetry (see Figure 4a) and extends into the headscarp of the Bribie Bowl Slide. Two prominent reflectors (a, b), recognised in the upper slope of the profile, one at 15 m and the other at 30 m below surface slope (320–400 m), onlap a third reflector (c) at the western end of the line. The a, b, and c (?) reflectors terminate on the western side of the slide scarp. A very similar geometric package of reflectors is present in the near-surface, contorted sediment layers located immediately downslope (to the east) of the slide scarp. Line L58a provides context for the GC1 core and demonstrates the relative geological youth of the prominent

scarp, which is suggested by the feature's bathymetric expression, inter-level character and overlapping outline, but the lack of downslope line length reduces its utility in confirming the slide as a translational slab slide type.

Coolangatta-2 (Core GC9) and Coolangatta-1 (Core GC8) slides

Cores GC9 and GC8, located in the slide scar of the Coolangatta-2 and Coolangatta-1 slides, respectively, on the Nerang Plateau (Figure 2b). The Coolangatta-2 and Coolangatta-1 Slides are relatively shallow translation slab slides. The removed material is relatively thin (<50 m) and is representative of the numerous upper slope failures that are present ubiquitously on this very gently dipping (2–3°) plateau. The seafloor within the scars is hummocky, gently concave downwards, with average slopes of ~ 3.5° within the failure plane, increasing to 7.5° at the head scarp. The ubiquitous presence of submarine landsliding on the Nerang Plateau suggests a continuous shedding of material off the plateau slope, with each successive failure retrogressively retreating towards the shelf break.

Sub-bottom profile GC8-SN is orientated parallel to the bathymetric contours across the Coolagatta-1 Slide and through the location of gravity core GC8 (Figure 7c). The unfailed area to the south of the slide is located in water ~1100 m deep and is characterised by gently undulating, well-bedded material, whereas the sediments present on the north side of the scarp are irregularly distorted sediments with some obvious layering on the slide margin. The youngest identifiable seismic reflectors upslope of these slides terminate on the scarp. The seafloor within the slump area is at 1150 m water depth, supporting the interpretation that a slab of material ~ 50 m thick has been removed. The within-slide surface topography is irregular and contains numerous point diffraction sources upslope (to the south) of the southern headwall. Distorted layering is also evident in the subsurface material. Core GC8 penetrated 2.52 m and, at this location, the strong continuous sub-bottom reflector was about 5 m below the surface (see Figure 7c, insert).

Cudgen Slide (Cores GC5, GC6, GC7, GC11)

Six cores are located within, or adjacent to, the Cudgen Slide at the southern end of the gently dipping (2–3°) Nerang Plateau (GC4, GC5, GC6, GC7, GC10, GC11; Figure 2c). Two of the cores are reference cores, that is, located outside of the boundaries of the inferred Cudgen Slide (GC4, GC7), and the remaining four were retrieved within the boundaries of the Cudgen Slide (Table 1).

Sub-bottom profiles of the Cudgen Slide (Figures 10, 11) indicate it is a large (50 km^2), deep rotational slide from which near-surface slabs have been detached by translational sliding. The current slide depression shown in the bathymetric data is relatively thin (<50 m; see Figures 3c, 4c) and is representative of the numerous upper slope failures that have occurred on this very gently dipping plateau. The seafloor is hummocky within the failure region, with a gently concave downwards geometry. It has average slopes of ~ 3.5° within the failure plane, and up to 7.5° at the head scarp.

Four sub-bottom transects (three west–east transects and one north–south traverse tie) are available over the Cudgen Slide (Figures 10, 11). The three downslope transects all image the landslide scarp in detail and the two longer lines (transects L79a-WE and GC6-WE) display very irregular topography, dominated by diffractions, and highly disturbed material at the downslope terminations (Figure 11), consistent with the hummocky terrain at the equivalent position in the bathymetric image. The apparent continuation of the upslope, above-scarp sediment layering in the near-surface below-scarp landslide mass is consistent with and corroborates the interpretation of the Cudgen Slide's bathymetric expression as a rotational failure.

West–east line L79a-WE down the Cudgen Slide shows a transition from well-bedded soft sediments above the main slide, to irregular distorted sediments with little obvious layering, below a drop-off

(especially > 5 m penetration), with the sea floor dropping from ~875 to 950 m water depth (Figure 10a). A 0.5 km long slab of material has been removed from the upper slope just above the slide's headscarp (~840 m water depth). This slab would have been ~20 m thick and there is no evidence for its incorporation as blocks deposited on the top of the rotated slump mass. This slab has apparently detached from its basal surface and either travelled beyond the extent of the both the sub-bottom profile and bathymetric image, or completely disaggregated during failure. Core GC11 is taken from the within-slide distorted sediments and has penetrated a distinct but non-continuous sub-bottom reflector surface at ~2 m below the surface (see Figure 10a, insert).

Similarly, profile GC6-WE shows well-bedded soft sediments upslope of the slide's headscarp and irregular distorted sediments with little obvious layering downslope of the scarp (especially > 5 m penetration), with the sea floor dropping from ~850 to 950 m water depth (Figure 10b). Three slabs of material have been removed from the upper slope above the main slide scarp, with headscarps at ~690, 770 and 810 m water depth (*1, 2, 3* on Figure 10b). The slabs are ~3–5, 5, 10 m thick, and ~700, 400, 450 m long, respectively but are not present as blocks on the slide surface in the lower hummocky terrain-slumped material, although they may be incorporated in this material. GC6 is taken from the contorted layers located within the main body of the slide where there a strong non-continuous, sub-bottom reflector surface present about 3 m below the seafloor surface but the core probably does not sample this surface as it is only 1.91 m long (Figure 10b, insert).

Profile GC4-and-5-WE (Figure 11) also presents well-bedded soft sediments above the slide's headscarp, but does not show disturbed material below as this line does not traverse the hummocky terrain evident in the bathymetry (see Figure 11a). As with the other two downslope transects, a slab of material appears to be missing from the upper slope above the slide (headscarp at ~670 m water depth). This slab is ~10 m thick, ~900 m long, and is not obviously present in the downslope slumped material. Core GC5 was taken from the within slide sediment and penetrated a distinct but non-continuous sub-bottom reflector surface at ~1.2 m below the surface, which correlates directly with the depth of the identified boundary surface at 1.16 m depth within the core (see Figure 9). GC5 terminates directly above a semi-continuous sub-bottom reflector surface at ~3 m below the surface (Figure 11a, insert).

South–north tie line GC6-and-7-SN similarly shows a transition from well-bedded soft sediments in the sidewalls of the slide, to irregular distorted sediments with little obvious layering below a drop-off (especially > 5 m penetration), with the slumped sea floor surface dropping from ~900 to 1000 m water depth (see Figure 11b). Cores GC6 and GC11 were taken from the distorted sediment in the slide scar and penetrated a distinct but non-continuous sub-bottom reflector surfaces at ~1 m and 2 m below the surface, respectively, correlating with the depth of the boundary surfaces at 1.83 m (GC6) and 2.11 m (GC11) for each core (see Figure 9). Both cores terminate directly above a strong continuous, but distorted, sub-bottom reflector surface at ~3–4 m below the surface (Figure 10b, insert). Core GC7 is taken from parallel-bedded sediment adjacent to the slide, and penetrates at least three distinct continuous sub-bottom reflector surfaces at ~1.5, 2, and 4 m below the surface, respectively (Figure 11b, insert).

Byron Slide (Core GC12)

One southern canyon core was collected from within the Byron Slide (core GC12; Figure 2d). There are no sub-bottom profiles available for the Byron Slide but this feature's bathymetric expression indicates it is a translational slab failure on the upper slope. It is located within the steeper canyon region (3–7°) and presents as a significantly thicker slide (>200 m) in comparison with those developed on the adjacent Nerang Plateau (~20–50 m thick). The seafloor within the Byron Slide has an average slope of ~ 12°, increasing to 33° in the crestal amphitheatre region at the head of the slide

scar. The Byron Slide displays both shallow slab slide failures on the upper slope that have failure planes parallel to the adjacent apparently unfailed slope segments, and deeper slump scars that seem to suggest rotational failure towards the bottom of the lower slope.

Core descriptions and sedimentology (physical properties)

A particular focus of the sedimentology is the investigation of changes across visually identified transition surfaces (hereafter called boundary surfaces) suspected to represent slide planes, which separate looser material from stiffer more compacted material.

Lithological units

Of the two main lithological units that can be observed in the cores: Unit 1 is defined as the upper unit, which comprises homogenous, bioturbated, hemipelagic, clay-bearing sandy silt to silty sand or clay–sand-bearing silts. The sediment is dominantly biogenic, sandy carbonate mud, with some terrigenous silt and clay. The sand fraction is typically composed of bioclastic materials, mostly foraminifera. The colour of the sediment generally grades through olive grey at the top to dark olive grey at the base of the unit. The upper 5–20 mm is a thin oxidised yellowish-brown layer in all cores. Burrows, laminae and mottling are variably present. Unit 1 thickness ranges from ~0.25 to 4.4 m in the collected cores and is interpreted to be a sediment drape above the slide plane. Unit 2 is the lower unit present beneath a distinct break and is a homogenous, bioclastic, hemipelagic, clay-bearing sandy silt to silty sand, which appears in most cases (GC6, GC7, GC8, GC11, GC12) to be noticeably firmer/stiffer in comparison to the overlying Unit 1. Unit 2 generally has a uniform texture, with some faint laminations apparent and is composed of foraminifera, shell fragments and other carbonate detritus. The colour of the material grades through light grey at the top of the boundary into a dark grey right at the base with faint to moderate mottling and rare bioturbation also present. Unit 2 thickness ranges from ~0.25 m to 1.6 m in the collected cores.

Cores GC1, GC2, GC3, GC9 and GC10 only penetrate Unit 1, while cores GC5, GC6, GC7, GC8, GC11, and GC12 penetrate both Units 1 and 2 (Figure 8, 9, 12). The difference between these two units is physical rather than compositional in that the lower Unit 2 is measurably stiffer and usually denser than the associated upper unit. Cores that sample both units tend to be shorter than those only penetrating Unit 1 most likely due to the stiffer Unit 2 sediments, with the exception of GC7.

Cores collected from the northern region of the study area (GC1, GC2, GC3; hereafter referred to collection as northern region sediments) have a thicker Unit 1 than cores collected from the southern region of the study area (GC6, GC7, GC8, GC11, GC12; hereafter referred to collection as southern region sediments).

Carbonate content (TIC) and Total Organic Carbon (TOC)

TIC remains fairly constant throughout both regions at around 17–22 wt% (average 20 wt%) and TOC content ranges between 4–12 wt% in all cores (average 7 wt%) (c.f. Clarke, 2014). These values of TOC are within a range to be considered high enough to have some impact on the strength of the sediment, with Bishop (2010) showing that TOC values as low as 3 wt% can affect mechanical strength.

Grain size

Sediments comprise mixtures of calcareous and terrigenous sand (15–40 vol%), silt (50–65 vol%) and clay (10–20 vol%). While relative proportions of clay, silt, and sand vary between each core, the general uniformity of the sediments is evident across all cores, with the majority classified as either clay/sand-bearing silts, clay-bearing sandy silts, or clay-bearing silty sands (Figure 13).

Northern region sediments are characterised by a lower sand content (<30 vol%) compared with southern region sediments (up to 60 vol%); clay content remains secondary in all samples (<25 vol%), with silt the dominant grain size. Sediments from above and below boundary surfaces generally display no significant change in grain-size characteristics.

Core sediments can be grouped into three main types:

   i.   Clay/sand-bearing silt – northern cores
   ii.   Clay-bearing sandy silt – southern cores
   iii.   Clay-bearing silty sand – southern cores

Density change

Small but distinct changes in some sediment physical properties (bulk density, water content, unit weight) were recorded above and below individual boundary surfaces. If no boundary surface were detected in the core and a uniform sequence of hemipelagic layers was present, a gradual increase in bulk density and unit weight, and decrease in water content, is seen down core (see Figures 8, 9). The variation and apparent disconnection of these properties above and below the boundary is interpreted as related to different burial depths, with sediments below the boundary presenting densities consistent with compaction due to burial at 5–10 m greater than their present depth below the sea floor (see Clarke, 2014).

The increase in density and stiffness of the sediment of the lower unit (Unit 2) when compared with the upper unit (Unit 1) was measured in four cores (GC7, GC8, GC11, GC12), with unit weight values increasing between 0.05–0.14 gcm^{-3} across the boundary (see unit weight in Figures 8 and 9 core logs). There was insufficient sediment below the boundary surface to test this phenomenon in GC6, although its close proximity to GC11 makes it reasonable to infer that the trend would follow. In contrast, GC5 presents a slight decrease in unit weight directly below the boundary but the sediment density then increases down the core (unit weight decreases from 1.62 gcm^{-3} to 1.55 gcm^{-3} across the boundary and returns to 1.65 gcm^{-3} 30 cm further down core).

Colour changes between Unit 1 and Unit 2 (that is, across the boundary surfaces) are inferred to be a consequence of decreasing water content that corresponds directly to increasing unit weight (c.f. Clarke, 2014). Decreases in water content across the boundary surfaces ultimately reflect differences in the burial depth of the sediment above (<1 m burial) and below (~5–10 m burial) the boundary features (c.f. Clarke, 2014).

Radiocarbon ages – dating the boundary surface and determining sedimentation rates

Age discontinuities – dating the boundary surface

Multiple age discontinuities are present across identified boundary surfaces in five different cores (GC8, GC7, GC5, GC11, GC12) taken from four separate slides (Coolangatta-1, "GC7 Slide", Cudgen, and Byron). Ages taken from either side of boundary surfaces indicate a significant time gap in deposition, suggesting distinct units. Sediment sampled from the basal layer of the overlying Unit 1 sediment returned ages between 23.7 to 12.3 ka, while sediment sampled from the underlying Unit 2 directly below the boundary surface dates at > 50 to 45.6.1 ka (see Table 2; Figures 8, 9, 12). The Unit 2 dates are at the limit of the ^{14}C technique and could be radiocarbon dead. We also suspect that bioturbation has introduced a small amount of modern sediment into the older sediment below the boundary, which could potentially result in a younger age being determined.

Sedimentation rate

[14]C ages were used to establish sedimentation rates for the northern NSW/southern QLD continental slope using material from six gravity cores: five cores from within submarine landslides (GC2, GC7, GC8, GC11 and GC12) and one adjacent reference core (GC1) (Figure 14; see Figure 2 for core locations).

Assuming continuous and constant sedimentation, the upper slope offshore Bribie Island within the northern canyon region (southern Qld) was determined to have a sedimentation rate between 0.21 mka^{-1} (GC1; 4 samples) and 0.24 mka^{-1} (GC2; 5 samples) (Figure 14). The upper slope offshore Coolangatta/Cudgen on the plateau region (northern NSW) returned sedimentation rates of 0.04 mka^{-1} (GC8; 6 samples), 0.02 mka^{-1} (GC7; 2 samples), and 0.12 mka^{-1} (GC11; 5 samples). The upper slope offshore Byron within the southern canyon region (northern NSW) returned a sedimentation rate of 0.06 mka^{-1} (GC12; 4 samples). These rates are applicable for the last ca 20 ka (ages used to determine the sedimentation rate are all <23 ka and all located above the boundary features).

Comparison of bulk sample dates with picked foraminifera dates

Age correspondence between bulk sediment and foram picked samples was investigated using five sample pairs (see Table 2) to test the difference/error between the two different sampling techniques. Whether bulk ages are acceptable depends on the depositional processes of the sediment and the type of material tested (Harney, Grossman, Richmond & Fletcher III, 2000). If sediment re-working and/or turbidite deposition were a characteristic of the site, then errors associated with bulk samples would be much greater than those reported here. There is no obvious physical re-working of the sediment evident. Neither is graded bedding, characteristic of turbidite deposition, evident in any of the cores from this region, apart from GC7, which does contain a reworked unit (Unit 2, 28–48 cm). There is reasonable correspondence between the ages of foram-picked samples and bulk samples for three out of the five sample pairs tested (GC8, GC11, GC12), with ages differences ranging between 2.18 to <1 ka for those three pairs. One sample pair (GC8) showed moderate correspondence, with an age difference of 3.2 ka. The fifth sample pair (GC7) showed little to no correspondence with a recorded age difference of ca 8.4 ka. This sample pair was taken from GC7 Unit 2 (41 cm down core) within a reworked unit and is therefore expected to have a greater error range when using a bulk sample. We conclude that the bulk sample ages can be considered acceptable and representative of the age of the sediments, whereas samples are not taken from within reworked units.

Discussion

Sedimentation style and rates

There is no major differences in the texture of the deposited material, either down individual cores or between the 10 different sites located in this area. Grain size, carbonate and organic carbon content present no significant change between units across the boundary features. While displaying an obvious change in colour, stiffness, density and water content between the upper and lower units, grain-size analysis reveals that the sediment composition remains generally uniform with all materials in the cores being poorly sorted, medium-to-coarse sandy silt, with relatively low clay and relatively high silt and sand percentages. TIC and TOC content remains fairly uniform, with carbonate content varying between 17% and 22% by weight and organic carbon content less than 10% by weight. These trends in the physical and mechanical properties are consistent across all 10 split gravity cores.

Dating of the conformably deposited post-slide sediment indicates sediment accumulation rates between 0.017 mka^{-1} and 0.2 mka^{-1}, which is consistent with previous estimates reported for this area. Given the approximate age of the eastern Australian margin since the initial phase of rapid rifting, subsidence associated with the opening of the Tasman Sea ended (>50 Ma), and sediment thickness of between 500 to 1000 m (Boyd et al., 2010), the average sedimentation rate for the margin is between about 0.01 to 0.02 m/ka. This low rate has been ascribed to a combination of factors such as low mainland sediment flux, limited accommodation space, and reworking by the strong, southward-moving sediment transport by the East Australian Current (Keene et al., 2008). Since rifting ceased ca 52 Ma, the southeast section of the east Australian margin has experienced minimal subsidence (Boyd et al., 2004), which is expressed by the convergence and apparent combination (or "pinching-out") of post-rift seismic reflectors on the shelf and uppermost slope (Keene et al., 2008; Marshall, 1979; Figure 5) indicating that the present-day inner-shelf had been maintained at an elevation near sea level (within 200 m) during the Cenozoic (excluding the Quaternary).

Troedson and Davies (2001) directly determined sedimentation rates for late Quaternary to Recent upper slope sediment deposits offshore Noosa Heads (southern Queensland) and Sydney (central NSW) from three gravity core samples (<4.5 m). Offshore from Noosa Heads, a sedimentation rate of 0.08 mka^{-1} was determined for the last glacial lowstand (ca 21–19 ka), while the last post-glacial transgression produced more rapid sedimentation rates of 0.15–0.24 mka^{-1}. Offshore Sydney, mean sedimentation rates have been reported to be between 0.02 and 0.05 mka^{-1} over the last 71 ka, with mean combined glacial/interstadial rates higher than Holocene rates by a factor of 1.36 (Troedson & Davies, 2001). Sedimentation rates around 0.02–0.11 mka^{-1} have been calculated using radiocarbon ages from gravity cores described by Glenn et al. (2008) from the central NSW continental slope. It is of interest to note that the directly determined sediment accumulation rates here are generally an order of magnitude higher than the average whole-of-Cenozoic rate for the eastern Australian margin discussed above, but are quite low in comparison with reported rates worldwide (Urlaub et al., 2013); e.g. 0.2 mka^{-1} for the Atlantic margins of North and South America, 0.17–36.0 mka^{-1} for Europe and 0.12–0.25 mka^{-1} for Western Africa.

The late Pleistocene and Holocene sedimentation rates reported in this work are comparable with previous directly determined sedimentation rates reported for this margin (c.f. Glenn et al., 2008; Troedson & Davies, 2001) and are of the same order of magnitude. We found that offshore central and northern NSW have slightly lower sedimentation rates (~0.02–0.12 mka^{-1}), with sediments offshore southern Queensland being deposited at a slightly higher rate again (~0.15–0.24 mka^{-1}). In comparison with approximate long-term determination of the rate of sediment accumulation on the upper slope of 0.01–0.02 m/ka, the directly determined rates for the last 50 ka from southern Queensland (~0.15–0.24 m/ka) are an order of magnitude higher.

As the rate of sedimentation is expected to have been higher in the geologically recent past (i.e. Holocene and latest Pleistocene) due to lower sea levels (~ –120 m at last glacial lowstand; Yokoyama, Lambeck, De Deckker, Johnston & Fifield, 2000), the consistent rates of sedimentation over the last 25 ka suggest that either these are in some way anomalously high, or that removal of sediment accumulated on the slope is a normal process. If the rates were representative or similar to the long-term Neogene average surface accumulation then sediment removal has been a consistently occurring event since the formation of the margin at 60 Ma. Given that the sediment wedge deposit is generally less than 500 m thick (Boyd et al., 2004, 2010) and there is abundant evidence for submarine landsliding (Boyd et al., 2010; Clarke et al., 2012; Hubble et al., 2012; Hubble et al. 2016; and this work) then it follows that the main mechanism for sediment removal on this section of the margin is the eastern Australian current and mass wasting. Assuming constant

sedimentation rates of between 0.02–0.20 mka^{-1} on the margin since its formation, a sediment deposit between ~1–3 km is "missing" from the margin, suggesting that much of the sediment has moved off the shelf and slope and down to the abyssal plain. If these rates were indicative of average conditions during the Pleistocene, a full glacial cycle (ca 100 ka) would be represented by somewhere between 8 to 24 m of sediment (~20 ms ± 10 ms of sub-bottom profiler record) (e.g. Figure 11a).

Significance of boundary surfaces

One slide-adjacent core, and four cores from within the Coolangatta-1, Cudgen, and Byron slide scars in the south of the study area all contain a boundary surface, located at depths of between 0.8 to 2.2 m below the present-day seafloor. All have a sharp, colour-change boundary, a small increase in sediment stiffness, and a slight increase in sediment bulk density of 0.1 gcm^{-3}. Distinct gaps in ^{14}C age of at least 20 ka are recorded across the boundary surface, in most cases. In all five cases these boundaries are paraconformities or obvious disconformity surfaces representing a hiatus. Coupled with the small but distinct down-core increases in sediment density, these observations when taken together suggest that the surfaces probably represent mass removal of material.

The sub-bottom profiles of the Coolangatta-1 (Figure 7c) and the Cudgen slides (Figures 10, 11), demonstrate that the boundary surfaces sampled in the cores are unlikely to be the detachment surface along which the slide blocks moved. Nevertheless, near-seafloor surface sediment was removed from above these boundary surfaces. It is inferred that this removal took place as either: a) displacement of small, thin slabs; b) liquefaction and downslope flow; or c) scour erosion (Özener, Özaydın & Berilgen, 2009; Steedman & Sharp, 2001). The boundary surfaces therefore do not enable a direct estimation of the time of slide occurrence. They do, however, enable dating of a relatively recent disturbance event, and the termination of the youngest reflectors in the sub-bottom profiles across the Cudgen Slide scar suggest that these slope failure events are geologically recent enough to be the event that removed significant amounts of material downslope. We suggest that the boundary surfaces can be used to constrain the timing of the overall, larger-scale sliding events with some confidence and that the boundary surfaces themselves probably date either the large-scale disturbance associated with slide motion or a penecontemporaneous post-slide erosion event.

If a disturbance/hiatus relationship or a post-disturbance erosion event were accepted, it can be inferred that the consistent dates for the boundary surfaces at difference sites (ca 12 ka, GC7; ca 15 ka, GC12; and ca 23–20 ka, GC5, GC8, GC11) suggest a regional erosional or disturbance event. The truncation of the material shown in sub-bottom profiles shows that the slides are young (Figures 7, 10, 11). Slide headscarps are shown to rupture the relatively modern bedding indicating that the maximum age the thinner (20 m) translational slides can be is ca 100 ka (this is based on the calculated recent surface sedimentation rates, that a deposit ~20 m thick has a maximum age of ca 100 ka).

The surfaces identified in the Coolangatta-1 and Cudgen within-slide cores are either: a) erosional features that developed after the occurrence of the landslide, in which case the hiatus surface age provides a minimum age for landslide occurrence or b) detachment surfaces from which slabs of near-surface sediment were removed during landsliding in which case the post-hiatus sediment dates indicates approximately when landsliding occurred. In either case, it is reasonable to suggest that these two spatially adjacent slides occurred penecontemporaneously ca 20 ka years ago.

A number of thin planar translational slab slide blocks have been removed from the seafloor upslope of the major slide scarps of the Cudgen Slide (Figures 10, 11). These translational slides are significant because they indicate that subsequent to the large rotational Cudgen Slide failure, there has been a continued retrogressive dislocation and removal of young sediment slabs, which were present prior

to the formation of the Cudgen rotational slide surface. The translational slides are retrogressive features with headwall scarps comprising "unsupported" layers (i.e. truncated at the headscarp) and could possibly slide in the future with further retrogressive failure.

Possible triggers

A variety of causes for the initiation of submarine landslides have been suggested (e.g. Locat & Lee, 2002; Masson et al., 2006). These include: earthquakes, storm-wave loading, erosion and, in particular, slope over-steepening, rapid sedimentation leading to under-consolidation, the presence of weak layers, gas dissociation, sediment creep, sea-level change, glaciation/isostatic uplift, volcanic activity, and diaper intrusion. It is widely accepted that a combination of these factors is required to initiate a landslide, especially where these occur on low inclination slopes slopes. There are data indicating that several large landslides have coincided with earthquakes, (e.g. Bardet et al., 2003; Masson et al., 2006; Synolakis et al., 2002; Tappin, Watts, McMurtry, Lafoy & Matsumoto, 2001), and the role of weak layers, oriented parallel to the sedimentary bedding, has long been used to explain the scale of some large slides, with the importance of weak layers in controlling sliding at all scales, having been suggested by several authors (see Masson et al., 2006).

The absence of weak clay layers in any cores, despite the widespread occurrence of submarine landslides, leads us to suspect that weak clay layers are not a major causal component in the case of the eastern Australian margin slides. Observations suggest earthquakes or erosive over-steepening of slopes (cf. Fletcher, 2015, Hubble et al, 2016) as more likely triggering mechanisms, for reasons explained below. The sediments of this area are ubiquitously uniform without discrete interbeded clay, and the similarity of hemipelagic muds from a relatively large geographic area has been noted (40 cores – Glenn et al., 2008; Hubble & Jenkins, 1984a, b; Troedsen, 1998; Troedsen & Davies, 2001).

The wide occurrence of upper slope slides across the east Australian margin (c.f. Clarke, 2014) indicates that submarine sliding should be considered to be a common characteristic of this passive continental margin. This indicates that one or more of the potential triggering mechanisms listed above can operate in passive margin settings to destabilise the slope. The processes leading to a failure are suspected to be most likely include: 1) reduction of the shear strength of the upper-slope sediments to very low values, possibly induced by creep or a build-up of pore-pressure; 2) long-term modification of the slope-geometry, i.e., sedimentation on the head of the slope and/or erosion of the toe of the slope; and/or 3) seismic events large enough to trigger sediment liquefaction or a sudden increase of pore-fluid pressure.

The major contenders – earthquake induced failure, over-steepening and loss of sediment strength – for triggering submarine landslides can be investigated with respects to the east Australian margin as a set of the following three questions:

1. Is submarine landsliding along the east Australian margin related to sea-level change?

The effects of climate change over hundreds of years, particularly with regards to sea-level fluctuations, have previously been identified as a possible long-term triggering mechanisms for submarine landslides and slumping, contributing to slope instability (Masson et al., 2006; Sultan et al., 2004). With this in mind, it is important to determine when major sea-level fluctuations have occurred in the past within the study area and whether the timing of these events correspond to the dates of the sliding events.

Lee (2009) suggested that landslides were more frequent during and just after the last glacial maximum (LGM ca 21–19 ka) than they are today. One of the reasons for this is that glacial maxima

coincide with periods of low relative sea level. The relative sea-level curve for Australia for the last 0.5 Ma (Waelbroeck et al., 2002; see Figure 15b for last 150 ka) indicates that sea levels have been over 100 m below present on several occasions, and at times are associated with glacial maxima. Lowered sea level can increase the likelihood of sliding because it results in the shoreline migrating closer to the shelf edge, leading to increased erosion and higher rates of sedimentation offshore directly onto the slope. Lower water pressures (and possibly, changes of temperature) could lead to release of gas from gas hydrates increasing pore pressures and reducing strength (although there is little evidence of gas hydrates in this section of the east Australian margin) and related hydrostatic unloading of the crust can increase seismic activity (Clark, 2010; King, Stein & Lin, 1994). Additionally, increased groundwater flows from underlying rocks could contribute to reduced strength. It has also been suggested (Hubble et al., 2012, 2016) that changes to ocean currents associated with glaciations can contribute to toe erosion and slope over-steepening.

[14]C ages determined for sediment sampled directly above the boundary surfaces do not appear to correlate sliding with any one particular sea-level event (Figure 15b). The oldest above-boundary surface age (ca 20 ka) correlates to around the time of the most recent lowstand; the next above-boundary surface age (ca 15 ka) correlates to the beginning of the rising limb of sea-level rise; and the third above-boundary surface age (ca 12 ka) corresponds to mid-way up the rising limb of sea-level rise, moving towards more stable sea levels (Figure 15). These data are not conclusive, especially given the limit in both the datable age of the sediment (50 ka) and maximum depth for the sediment cores (5 m), but are not consistent with lowered sea level as a primary factor inducing sliding. This is consistent with Urlaub et al.'s (2013) recent conclusion that sea-level change does not appear to correlate with submarine landslide occurrence and that sea-level changes are a contributing factor only if sediments cannot dissipate pore pressure. However, the age database for submarine landslides is still extremely limited.

2. Do gas hydrates play a role in triggering submarine landsliding along the east Australian margin?

Gases formed by the dissociation of methane hydrate can dissipate though porous permeable sediments that drain relatively freely (such as granular silts). Gas hydrates are generally located much deeper beneath the seafloor (i.e. >500 m), and while gas build-up is recognised as a contributing causal factor for some submarine landslides (e.g. Storegga Slide or Ruatoria; Collot, Lewis, Lamarche & Lallemand, 2001) they generally relate to deep, very large, slope failures (Collot et al., 2001). This process is not considered to be likely in the case of the eastern Australian upper slope due to the more shallow occurrence of the slides, the low clay content (<20%), relatively loose packing (low density), and high porosity of the materials found by this study, along with the relatively thin sediment wedge (<0.5 km thick) present on the eastern Australian continental margin.

Our bathymetric data show no obvious evidence of gas hydrate dissociation ("pockmarks") for this section of the continental slope. There are circular depressions, referred to as pockmarks, reported for the continental margin offshore Newcastle (~ 400 km south of the study area), which are believed to be associated with gas leakage from the underlying Permian coal measures (Glenn et al., 2008); however, similar features are not observed in the study area.

Biogenic (shallow) gas formation could impact the stability of the slopes, causing near-surface, superficial failure of the slope (Canals et al., 2004). Biogenic gas is usually identifiable in high-resolution seismic profiles by "bright" spots (strong reflections); however, to date, no such evidence has been observed.

3. Do the grain-size distributions of the sediments from the east Australian margin make the slopes prone to earthquake-induced strength loss, triggering submarine landsliding?

Earthquake loading (also known as cyclic loading or seismic loading) is the most commonly indicated submarine mass failure trigger (Masson et al., 2006). It is a short-term trigger, with slope failure occurring during the earthquake or shortly afterwards (e.g. Aitape submarine landslide; Tappin, Watts, & Grilli, 2008). During an earthquake, the ground accelerations may be of a sufficient magnitude to overtop static gravity forces (Bardet et al., 2003). Mechanically, there are three aspects to this phenomenon: the sediments exposed to seismic loading (where there can be an application of acceleration in both the horizontal and vertical directions) generally experience a decrease in stiffness, decrease in shear strength, and increase in pore pressure (Coleman & Prior 1988; Sultan et al., 2004). If the sediment were loosely-packed, compaction takes place at a rate controlled by permeability of the material. Seismic loading usually takes place over a time period that is too short to allow pore water drainage and the consequent pore-pressure increase produces an equivalent decrease of the effective confining stress. In some cases, this process can reduce the vertical effective stress to zero with a corresponding loss shear strength, which may produce a complete loss of shear strength in a non-cohesive granular sediment (Sultan et al., 2004). All sediment types (not over-consolidated or dense) encounter on continental slopes are expected to experience pore pressure build up due to cyclic loading. Silty sand and, to a lesser extent, sand can be very sensitive to pore pressure build up and can loose strength and collapse due to this process (Mitchel & Soga, 2005), which can lead to large ground deformation and possibly liquefaction (Coleman & Prior, 1988).

The non-cohesive, granular sediments recovered from the eastern Australian margin (i.e. silty sand/sandy silt) described in this study present grain-size distributions similar to those materials demonstrated to be prone to earthquake-induced strength loss and liquefaction when subjected to the prolonged ground-shake that occurs in larger earthquakes (i.e., ~Mw >7.0) (Özener et al., 2009; Steedman & Sharp, 2001; Wiemer et al., 2012). This, combined with the ubiquity of submarine landslides on the east Australian margin and a suspected increased incidence of east coast earthquakes during the Pliocene and Pleistocene due to the ongoing collision of the NW Australia with SE Asia (Hillis, Sandiford, Reynolds & Quigley, 2008; Müller, Dyksterhuis & Rey, 2012) as suggested by Hubble et al. (2012), favours an earthquake-triggering mechanism for the initiation of submarine landslides on this passive continental margin. The recurrence interval for large eastern Australian surface-breaking earthquakes (i.e. those that rupture on an individual fault source) is not well-constrained and estimates range over three orders of magnitude from one moment magnitude (Mw) >7.0 event every ten to twenty thousand years to one every million years (i.e. a range in recurrence interval of 10 ka to 10^3 ka) (see Burbidge, 2012; Clark, 2010; Clark, McPherson & Van Dissen, 2012; Leonard, 2008).

Conclusions

Specific conclusions are as follows:

1. The sediments collected in this study are typical of the materials found elsewhere on southeastern Australia's upper continental slope. Calcareous hemipelagic muds, composed of mixtures of calcareous and terrigenous clay (10–20 vol%), silt (50–65 vol%) and sand (15–40 vol%) and generally uniform in appearance, have been sampled in cores taken from within, or adjacent to five submarine landslides on eastern Australia's continental slope in 450 to 1150 m of water depth.

2. Dating of conformably deposited material indicates a range of sediment accumulation rates between 0.017 mka^{-1} and 0.200 mka^{-1}, which are consistent with previous estimates reported for this area.

3. Sub-bottom profiler records of transects through three within-slide core sites and their nearby landslide scarps (Coolangatta-1 and Cudgen slides) indicate that the youngest identifiable seismic reflectors located upslope of these slides terminate on and are truncated by slide rupture surfaces and that the studied slides are geologically recent, therefore < 25 ka. There is no obvious evidence in the sub-bottom profiles for a post-slide sediment layer draped over or otherwise burying slide ruptures or exposed slide detachment surfaces.

4. Boundary surfaces were sampled in one slide-adjacent core, and four within-landslide cores at depths of 0.8 to 2.2 m below the present-day seafloor. These boundary surfaces are identified by a sharp, colour-change boundary; discernable but small increases in sediment stiffness; and slight increases in sediment bulk density of 0.1 gcm^{-3}. Distinct gaps in AMS ^{14}C age of at least 25 ka are recorded across the boundary surfaces.

5. The boundary surfaces are either: a) erosional features that developed after the removal of a large landslide mass, in which case the hiatal surface age provides a minimum age for landslide occurrence; or b) detachment surfaces from which slabs of near-surface sediment were removed during landsliding, in which case the post-hiatus sediment dates indicate approximately when landsliding occurred. In either of these scenarios, it is reasonable to suggest that these two spatially adjacent slides occurred penecontemporaneously at ca 20 ka.

6. While we currently favour an earthquake triggering mechanism for the initiation of submarine landslides on the eastern Australian margin this causal mechanism cannot be conclusively demonstrated to be the only trigger. Further investigation of these features may well show that erosive scour and over-steepening is sufficient to induce slope failure.

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Acknowledgments

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indicated with a dashed red line. The boundary feature is indicated with a dashed black line. Yellow boxes show close up location of sediment core photos (Figure 12). Legend is shown in Figure 9.

Figure 9 Core logs of gravity cores (GC) analysed in this study collected from the plateau region of the study area: (a) GC5, (b) GC6, (c) GC7, (d) GC8, and (e) GC11. Core logs show radiocarbon ages, core photos, and physical properties with depth. Physical properties include mean grain size (μm) and unit weight (kNm^{-3}). The projected unit weight trend assuming no boundary feature is indicated with a dashed red line. The boundary feature is indicated with a dashed black line. Yellow boxes show close up location of sediment core photos (Figure 12).

Figure 10 Panel's a and b show uninterpreted (left) and interpreted (right) sub-bottom profile lines. (a) Sub-bottom profiling line L79a-WE downslope across the Cudgen Slide, through GC11 (southern plateau). The location of GC11 is indicated in red. Missing sediment slab location is shown in blue (see text for details). Line locations are shown on Figure 7a. Insert shows distorted sediment layers within the slide scar. (b) Sub-bottom profiling line GC6-WE from west to east down the Cudgen Slide, through GC6 (southern plateau). The location of GC6 is indicated in red. Locations of three missing sediment slabs are shown in blue (see text for details). Line locations are indicated in Figure 7a. Insert shows undisturbed layers above to the west of the slide scarp and distorted layers below.

Figure 11 Panel's a and b show uninterpreted (left) and interpreted (right) sub-bottom profile lines. (a) Sub-bottom profiling lines GC4-and-5-WE down the Cudgen Slide, through core GC5 (southern plateau). The location of GC5 is indicated in red. Missing sediment slab location is shown in blue (see text for details). Line locations are indicated in Figure 7a. Insert shows sub-parallel sediment layers within the slide scar. (b) Sub-bottom profiling line GC6-and-7-SN tie line across the Cudgen Slide, through cores GC6, GC11, and the adjacent core GC7 (southern plateau). Gravity core locations are indicated in red. Line locations are shown on Figure 7a. Note the dark near surface reflector of modern sediments above the failure surface.

Figure 12 Close up images and interpretation of the boundary features in each of the six gravity cores (GC) analysed in this study: (a) GC5, (b) GC6, (c) GC7-i, (d) GC7-ii, (e) GC8, (f) GC11, and (g) GC12. See Figures 1, 2 for core locations and Figures 8, 9 for location of the photo down each core. Panel's a–g show uninterpreted (left) and interpreted (right) core sections. The interpreted boundary feature is indicated with a dashed black line. Radiocarbon ages for each core are shown in yellow (ka = thousand years before present, RCD = radiocarbon dead).

Figure 13 Grain-size analysis results for 142 sediment samples from 12 cores: (a) sand:silt:clay ratio diagram; and (b) grain-size distribution plot (log scale). Coloured markers (a) or lines (b) represent "typical" responses for each main sediment type: Type 1: clay–sand bearing silts (green); Type 2: clay bearing sandy silts (blue and orange); and Type 3: clay bearing silty sands (red).

Figure 14 Sedimentation rates determined from calibrated ^{14}C ages (see Table 2) within 6 gravity cores (GC1, GC2, GC7, GC8, GC11, GC12). Sedimentation rate values range between 0.21–0.24 mka^{-1} for the northern cores (GC1, GC2), and 0.02–0.12 mka^{-1} for the southern cores (GC7, GC8, GC11, GC12). These rates apply over the last ca 20 ka (ages used to determine the sedimentation rate are all <23 ka and all located above the boundary features – see text for details).

Figure 15 (a) ^{14}C age frequency distribution plot: a total of 49 samples are plotted (31 samples above the boundary surface, 18 samples below) into 10-year bins. Samples located above the boundary surfaces cluster in <30 ka age ranges, while samples taken below the boundary surfaces fall only in the 50–41 ka age range. (b) Eustatic sea-level record from Waelbroeck et al. (2002) (using foraminifera δ^{18}O records) over the last 150 ka. Stars highlight the location of

approximate ^{14}C age groupings from the gravity cores samples: 12 ka (blue), 15 ka (orange), 20 ka (yellow), and 50 ka (green). See Table 2 for the full list of ^{14}C ages.

Table 1 Summary of locations, depth, total recovery length, stratigraphy and target feature of the sediment cores retrieved from the study area.

Table 2 Radiocarbon ages of 49 core samples. Ages were determined by accelerator mass spectrometry (AMS) and were obtained from both bulk sediment samples and planktonic foraminifera assemblages extracted from the hemipelagic sediments. Conventional ^{14}C yrs BP were converted into calibrated calendar ages following Stuiver and Reimer (1993). Median calibrated ages (BP) were calculated with CALIB V6.1.1 (Stuiver et al., 2005) using marine calibration curve Marine 09.14c data set (Reimer et al., 2009) with a reservoir correction (ΔR) value of 11 ± 13 yr and reported here with 2σ errors (see text for details).