Temporal and spatial complexity in post-glacial sedimentation on the tectonically active, Poverty Bay continental margin of New Zealand

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Abstract

On the eastern Raukumara Ranges of the New Zealand East Coast, active tectonics, vigorous weather systems, and human colonisation have combined to cause widespread erosion of the mudstone- and sandstone-dominated hinterland. The Waipaoa River sedimentary dispersal system is an example that has responded to environmental change, and is now New Zealand’s second largest river in terms of suspended sediment discharge. This paper presents new sediment accumulation rates for the continental shelf and slope that span century to post-glacial time scales. These data are derived from radiochemical tracer, palynological, tephrostratigraphic, and seismic methods. We hypothesise on the temporal and spatial complexity of post-glacial sedimentation across the margin and identify the broad extent of sediment dispersal from the Waipaoa system. The \(15 \text{ km}^3\) Poverty Bay mid-shelf basin lies adjacent to the mouth of the Waipaoa River, reaching a maximum thickness of \(45 \text{ m}\). A post-glacial mud lobe of an additional \(3 \text{ km}^3\) extends through the Poverty Gap and out onto the uppermost slope, attaining \(40 \text{ m}\) thickness in a structurally controlled sub-basin. Here, an offset in the last-glacial erosion surface indicates that deposition was sympathetic with fault activity and the creation of accommodation space, implying that sedimentation was not supply limited. Contrary to classical shelf sedimentation models, the highest modern accumulation rate of \(1 \text{ cm y}^{-1}\) occurs on the outer-shelf sediment lobe, approximately \(2\) times the rate recorded at the mid-shelf basin depocentre, and \(10\) times faster than the excess \(^{210}\text{Pb}\) rates estimated from the slope. Pollen records from slope cores fingerprint Polynesian then European settlement, and broaden the spatial extent of post-settlement sedimentation initially documented from the Poverty Bay mid-shelf. Changes in sub-millennial sedimentation infer a \(2\)–\(3\)-times increase in post-settlement accumulation on the shelf but a smaller \(1\)–\(2\)-times increase on the slope. Over longer time scales, seismic evidence infers slower but steady sedimentation since the last transgression, and that significant cross-shelf sediment pathways pre-date the increase in sedimentation resulting from colonisation and deforestation. From a summation of coastal bedload, shelf and slope sediment mass accumulation, the total sediment budget for the Holocene is \(1\) Mt y\(^{-1}\). Under modern conditions a larger proportion of the Waipaoa sediment dispersal system likely extends onto the slope and beyond.

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1. Introduction

Anthropogenic and natural forcings in catchment watersheds affect sedimentation on continental shelves, particularly those associated with small mountainous rivers (e.g., Milliman and Syvitski, 1992; Syvitski et al., 2005). On the Raukumara Ranges of the New Zealand East Coast, colonisation brought with it widespread destruction of native forests for pasture which, when combined with high rainfall events, caused accelerated erosion of the mudstone- and sandstone-dominated hinterland (e.g., Gomez et al., 1999; Marutani et al., 1999; Page and Trustrum, 1997; Hicks et al., 2004). By the early 20th century, nearly all the forest cover had been removed, landslides proliferated in steep headwaters, and rivers were aggrading rapidly (e.g., Gage and Black, 1979). Simultaneously, these surficial processes occur against a background of rapid tectonic uplift and margin instability driven by oblique subduction. Coastal sedimentary dispersal systems respond to such changes. Today, approximately 70 Mty⁻¹ of suspended sediment is carried from the Raukumara Ranges to the sea, primarily by three rivers, the Waipaoa, Uawa/Hikwai, and Waiapu (Hicks and Shankar, 2003; Hicks et al., 2004). Despite a small catchment of 2205 km², the Waipaoa River delivers 15 Mty⁻¹ of suspended sediment (1% bedload) to coastal Poverty Bay (Hicks et al., 2000), with sediment yields in its upper catchment among the highest recorded on Earth (Walling and Webb, 1996). These factors contributed to the Waipaoa River and the Poverty Bay margin being selected as part of a larger international initiative to examine the terrestrial and marine sedimentary response to natural and human impacts on dispersal systems at mud-dominated coasts, called MARGINS Source-to-Sink.

High-yield rivers on active margins contribute the majority of terrestrial material flux to the world oceans, yet the sediment dispersal patterns of such rivers are not well understood, and appear to contrast markedly with those of the world’s relatively well-studied large rivers such as the Amazon (Nittrouer et al., 1995, 1996) or the Yangtze (Chen et al., 2000; Saito et al., 2001), which typically drain passive margins. A mechanistic understanding of the transport and fate of river sediment on continental margins is needed in order to model and predict a range of challenges that are of scientific and societal relevance, such as: the influence of climate and landscape changes on stratigraphic evolution of the margin; cycling and burial of biogeochemically active constituents such as carbon, nitrogen and iron; and sedimentation effects on benthic and pelagic communities. Whereas many of the world’s large river systems produce prominent shoreline-attached clinoform mud deposits which may be greatly attenuated in the alongshore direction of dominant sediment transport (e.g., Amazon, Yangtze), smaller rivers less commonly produce clinoforms, and appear in many cases to have a more direct connection with the deep sea, such as the Sepik River (Kuehl et al., 2004). These smaller rivers typically show event-dominated sediment transport. For example, at the muddy Eel River margin off California, flood discharge promotes the formation of off-shore directed, high-turbidity bottom flows, lasting perhaps a day or less (e.g., Ogston et al., 2000; Puig et al., 2003). Estimates of the sediment mass accumulation through time are required if the impact of landscape change, and the delivery of terrestrial material to the oceans are to be assessed quantitatively. Notwithstanding the scale of post-colonisation landscape changes that have occurred, Holocene and post-colonisation sediment mass accumulation rates have not been determined previously on a margin-wide scale adjacent to a major New Zealand river, encompassing both the shelf and slope depositional systems. Moreover, changes in post-colonisation sedimentation patterns and off-shelf sediment accumulation have been described in few localities globally, yet such deposits are critical to the wider understanding sediment dispersal (e.g., Nittrouer, 1999; Walsh and Nittrouer, 2003; Syvitski et al., 2005). Short-term (century scale or less) sedimentation and anthropogenic impacts are particularly important at tectonically active margins adjacent to muddy rivers (e.g. Sternberg, 1986).

This paper presents new, semi-quantitative information on sediment accumulation rates for the Poverty Bay continental shelf and slope that span century to post-glacial time scales. The results are derived from radiochemical tracer, palynologic, tephrostratigraphic, and seismic methods. We hypothesise on the temporal and spatial complexity of post-glacial sedimentation across the margin and identify the broad extent of sediment dispersal for the Waipaoa sedimentary system. Furthermore, this study extends the seismic characterisation of the outer shelf in particular and presents a revised Holocene sediment budget for the Poverty Bay margin. The insights derived from our observations of the Waipaoa sedimentary system are likely characteristic of small mountainous rivers in general.
1.1. Regional setting

1.1.1. Geology, morphology, and post-glacial seismic architecture

The Poverty Bay continental shelf and slope are located on the tectonically active northern Hikurangi margin of New Zealand, where oceanic crust of the Pacific Plate is being subducted obliquely beneath the Raukumara Peninsula and its eastern margin. (Fig. 1). Inboard (west) of the subduction margin lies the rhyolitic Taupo Volcanic Zone which is a prolific source of geochemically distinct tephras that punctuate the terrestrial and offshore stratigraphic record throughout the Quaternary (e.g., Froggatt and Lowe, 1990; Carter et al., 1995), and are used in this paper as a geochronological tool. The Hikurangi margin lies within a zone of active deformation that straddles the top of...
the Neogene imbricated margin and forearc basin (e.g., Lewis, 1980; Lewis and Pettinga, 1993); faults are ubiquitous on the margin. Subduction-related underplating beneath the Raukumara Peninsula is uplifting the axial ranges at an estimated maximum rate of 3 mm y\(^{-1}\), actively uplifting an allochthonous Palaeogene slab and overlying Neogene cover sequence (e.g., Reyners and McGinty, 1999).

The region encompasses the 2205 km\(^2\) Waipaoa River basin, which drains the eastern flanks of the axial Raukumara Range. The regional geology is structurally and stratigraphically complex. Rocks and sediments range in age from Cretaceous to Recent and the dominant lithologies are sandstone, argillite, and mudstones (Mazengarb and Speden, 2000). The strongly jointed and clay-rich lithology results in highly unstable landforms, manifest as slumps, landslides, and extensive gully erosion (e.g., Berryman et al., 2000; DeRose et al., 1998), and lead to very high sediment yields for the Waipaoa River catchment. The alluvium base along the Waipaoa River has an exponential form to within ~25 km of the coast, whereupon regional neotectonism causes a transition from uplift to subsidence (Brown, 1995; Berryman et al., 2000).

Accretionary tectonics and plate convergence have produced imbricate-thrust faults and folded Neogene slope sediments as a deforming backstop, with only a narrow accretionary prism locally forming in places at the toe of the slope (Lewis and Pettinga, 1993; Collot et al., 1996). The 1500 km\(^2\) Poverty indentation is a major continental margin depression extending from a re-entrant in the deformation front at the Hikurangi Trough to the continental shelf (Collot et al., 1996). The bathymetry of the Poverty indentation is complex and comprises six basic morphologic components, in order of increasing water depth (Fig. 2): (i) a heavily gullied upper slope; (ii) a gently sloping mid-slope trough (Paritu Trough); (iii) the beheaded Poverty Canyon system; (iv) margin-parallel lower slope ridges (North and South Paritu Ridges); (v) a V-shaped structural re-entrant in the deformation front at the canyon mouth; and (vi) the largely flat expanse of the Hikurangi Trough, seamounts, and Hikurangi Channel seaward of the deformation front (Fig. 1).

Mud is accumulating on the Poverty shelf in a broad depositional basin centred at mid-shelf water depths of 30–70 m, with a long axis that is parallel with the NE–SW orientation of the coastline and shelf break. The basin is subsiding at 2–4 mm y\(^{-1}\) (Foster and Carter, 1997). Regionally, the post-glacial shelf architecture can be broadly characterised by two major seismic reflectors, namely the uppermost erosional unconformity regarded as the last-glacial transgressive surface (termed “W1” after Lewis, 1973) (Fig. 3) and a conformable strong seismic reflector in the top 15 m of the deposit (“H”) (Fig. 3). The W1 last-glacial erosion surface crops out on both the landward and seaward flanks of the emergent Lachlan and Ariel ridge anticlines, where it defines the unconformity between unconsolidated post-glacial mud and the underlying deformed and indurated Neogene strata (cf. Lewis, 1973; Foster and Carter, 1997; Barnes et al., 2002). This surface has not been dated directly at Poverty Bay, but elsewhere along the East Coast it is taken to be contemporaneous with the last glacial maximum (ca. 18 ka BP) and/or the early stages of marine transgression approximately 18–14 ka BP (cf. Lewis, 1973; Barnes et al., 2002). A younger age for the erosion surface is unlikely on the Poverty outer shelf because, according to Barnes et al. (2002), water depths greater than around 30 m will limit the abrasion potential of waves in the area. Like the W1 reflector, the “H” reflector can be traced regionally (Lewis, 1973; Foster and Carter, 1997) and is probably early Holocene in age (ca. 8000 y BP) after carbon dating by Pantin (1966) from just below a presumed exposure of the reflector in Hawke’s Bay. Additionally, the occurrence of Tuhua Tehpra (6970 y BP—note all ages herein are in calendar years) at 14.13 m depth in the 16-m long Calypso core MD97-2122 (Gomez et al., 2004) from the Poverty Bay mid-shelf basin, confirms that the immediately underlying “H” reflector ~1–2 m below the core base is early Holocene in age. A pre-Holocene age for the H reflector is unlikely given that there is no evidence for a hiatus in post-glacial sedimentation on the Poverty Bay shelf or the East Coast region (cf. Lewis, 1973; Foster and Carter, 1997).

Further offshore, the Poverty slope has received significant amounts of terrigenous sediment, accumulating predominantly in the Paritu Trough and a smaller lower-slope basin (Orpin, 2004). On gently sloping terrain on the upper- and mid-slope sediment cores suggest that hemipelagic muds are the dominant lithology, inter-layered with discrete sandy volcanic tephra of mid- and late-Holocene age. Hummocky and irregular seafloor bathymetry is associated with avalanche deposits (cf. Lewis.
et al., 1998; Orpin, 2004), as observed elsewhere on the Hikurangi margin (Collot et al., 2001) and globally (e.g., Masson et al., 2002; Normark et al., 2004).

1.1.2. Climate, hydraulic regime, and floods

In the Gisborne region, there is a predominance of orographically influenced north and northwest winds, which occur for 80 and 110 days per year,
respectively (Table 1, NIWA electronic climate archive 1989–2002). Mean wind speeds are >10 km h\(^{-1}\) for all months (Table 1) and the mean annual wind speed is 12 km h\(^{-1}\). In contrast, at East Cape 120 km to the north, winds are considerably stronger with mean wind speeds around 24 km h\(^{-1}\).
occurring throughout the year (electronic climate archive 1961–1985) and the predominant wind direction is variable from the westerly quarter with a strong southerly influence.

Climatically, the East Coast is typified by a temperate but vigorous maritime climate, punctuated by slow-moving deep depressions of subtropical origin that generate easterly winds, intense rain, and floods. The averaged annual discharge data summarised in Foster and Carter (1997) indicate that higher flows are typically in late summer (March), with a secondary high in early spring (September). In general, March–May is the wettest period of the year (Page et al., 2001) and floods are frequent but annual stage height data from 1932 to 2002 at Kanakanaia Bridge on the lower reaches of the Waipaoa River show that large (>1900 m$^3$s$^{-1}$) floods occur throughout the year (data modified after Reid, 1999). A comparison of wind direction and Waipaoa flood heights indicates that the timing of major flood events >2000 m$^3$s$^{-1}$ is synchronous with easterly winds and the passage of the centre of meteorological cyclonic depressions responsible for the rainfall.

The regional oceanography plays a major role in the seaward dispersal of sediment from the Waipaoa River. Along the eastern North Island, seaward of the northward flowing Wairarapa Coastal Counter-current (WCC) is the southward flowing East Cape Current (ECC), forming part of the Subtropical Inflow to New Zealand (Heath, 1985; Chiswell and Roemmich, 1998; Chiswell and Booth, 1999).

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<td>Percent annual occurrences of monthly wind direction, mean wind speed,</td>
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<th>East Cape 1961–1985 Jan</th>
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Data were extracted from the National Institute of Water and Atmospheric Research (NIWA) electronic climate archive.
With a variable volume transport of $1-2 \times 10^7 \text{ m}^3 \text{ s}^{-1}$, the ECC is nominally positioned seaward of the 1000 m isobath, although the geotrophic cross-section of Stanton et al. (1997) and current meter data of Chiswell and Roemmich (1998) from south of East Cape indicate the ECC extends inshore almost to the shelf break and to at least 1300 m depth. In this case, the current’s passage across Poverty Bay margin may be modified by the topography of the indentation, for example, causing possible intensification between Ariel Bank and Tuaheni Ridge. The ECC will transport suspended sediment south. A third component of the coastal circulation is large eddies formed by complex interactions between the southward-flowing ECC, the Wairarapa Eddy, and the East Cape Eddy (Chiswell and Roemmich, 1998; Roemmich and Sutton, 1998; Chiswell, 2000). New analysis by Chiswell (2005) reveals a dynamic situation along northeastern New Zealand, with eddies moving southwest along the outer margin at a rate of 2-3 per year, stalling or merging with older features, and at times these eddies may be responsible for moderate currents at $>2000$ m depth (P. Sutton, NIWA unpublished data).

1.1.3. Anthropogenic impact on landscape erosion

Using fires, Polynesian settlers began to clear the Raukumara Ranges of thick temperate rain-forest 500-700 y BP (e.g., McGlone et al., 1994; McGlone and Wilmshurst, 1999), and forest clearing accelerated significantly with European colonisation in the mid-18th century (Pullar, 1962). By 1880, most of the hinterland had been cleared, and by 1920 all but a few percent of the land had been converted to pasture. Following deforestation an intense phase of landscape erosion was initiated in the upper reaches of the Waipaoa catchment around the beginning of the 20th century (Allsop, 1973). Since then the river has aggraded rapidly in response to increased sediment yield (Gage and Black, 1979; Gomez et al., 1999). Detailed data from the mid-shelf core MD97-2122 suggest that European deforestation was the major factor leading to an estimated four-fold increase in vertical sediment accumulation (Gomez et al., 2001, 2004). In this core record, the tephra chronology indicates that during the mid-Holocene, between the Whakatane (5580 y BP), Waimihia (3472 y BP), Taupo (1718 y BP), and Kaharoa (665 y BP) eruptions, vertical sedimentation rates are consistently around $0.15 \text{ cm y}^{-1}$ (Gomez et al., 2004), or approximately $0.1 \text{ g cm}^{-2} \text{ y}^{-1}$.

Palynology has been used in the East Coast marine record as an indicator of Holocene landscape and climate change (e.g., Wilmshurst et al., 1999; Carter et al., 2002). In general, the record from the mid-Holocene until around 650 y BP shows that the region was dominated by mature podocarp-beech forest with Dacrydium cupressium (Rimu) and Prumnopitys taxiflora (Matia) as the main species (e.g., Wilmshurst et al., 1999). Lowland alluvial swamp forests of Dacrycarpus dacrydioides (Kahikatea) were also abundant in the Gisborne area but were not deforested until the arrival of European settlers early in the 19th century (Wilmshurst et al., 1999). A dramatic decrease in podocarp and beech pollen, and punctuated, sharp increases in Pteridium esculentum (bracken) spores have been considered indicators of volcanic damage (burning, basal surge, wind-fall, or smothering by ash) and land clearance by Polynesian settlers (e.g., Wilmshurst et al., 1999; Carter et al., 2002; M. Elliot, NIWA, unpublished data), the latter of which is broadly coincident with the Kaharoa Tephra (665 y BP) in the offshore Poverty Bay record. More recently, Pinus radiata (Monterey Pine) was introduced in the early to mid-1800s upon European settlement but increased significantly upon the initiation of exotic forest plantations in the 1930s (Guthrie-Smith, 1969).

2. Methods and data

To evaluate relative mass accumulation rates over the last century, sediment samples were analysed for naturally occurring $^{210}\text{Pb}$ and bomb Cs$^{137}$, taken from NIWA cores U2304, U2305, U2306, W697, W698, and W699. Samples were obtained from 1 cm ($\sim 70 \text{ cm}^3$, 100 g) core slices of wet mud extruded from a core barrel, with a maximum core length of $\sim 60$ cm. Sediments were weighed, oven-dried ($60^\circ \text{C}$), then ground to a powder, sealed in 30 ml polypropylene jars and left to equilibrate. Total $^{210}\text{Pb}$ activity in slope sediment samples was directly determined by measuring the 46.5 KeV gamma peak following the techniques described in Alexander et al. (1993). $^{137}\text{Cs}$ activities were determined by measurement of its 661.6 KeV gamma peak (Kuehl et al., 1986). $^{137}\text{Cs}$ (half-life 30 y) is an impulse tracer (produced from atmospheric nuclear tests), which was first introduced in significant amounts in about 1954 and reached a peak in 1963. Shelf sediment samples were analysed for $^{210}\text{Pb}$ through measurement of its granddaughter, $^{210}\text{Po}$, assuming secular
equilibrium. Dried sediment samples were spiked with a known quantity of $^{209}$Po and successively leached in concentrated HNO$_3$ and 6N HCl. Polonium isotopes were plated onto a silver metal disc suspended in a weak HCl leachate, following the basic procedure of Flynn (1968). The $^{209}$Po and $^{210}$Po isotopes were then measured by their characteristic alpha decay using standard silicon surface barrier detectors. Supported $^{210}$Pb activities were estimated by averaging values deep in sediment cores that had reached low and uniform levels. Accumulation rates were estimated from the linear regression of the log of excess $^{210}$Pb activity below the mixed surface layer (core top of relatively homogeneous distribution of excess $^{210}$Pb resulting from rapid physical and biological mixing).

A total of 20 samples were taken for palynological analysis from cores W697, W698, and W699 (the same cores also used for radiochemical analysis) to assess the relative mass accumulation rates on the slope over sub-millennial time scales. Selected 1 cm ($\sim$70 cm$^3$, $\sim$100 g) wet mud slices were prepared and data reported in accordance with standard laboratory procedure (refer to Mildenhall, 2003). Preparation and analyses were undertaken by D.C. Mildenhall and E.M. Crouch, GNS Science, Lower Hutt (pers. comm., 2003). A total of 300 pollen grains were counted in each sample, irrespective of the appearance of each grain, except for obviously extinct taxa. Palynological age control was established on the basis of the following criteria, essentially founded on Polynesian and European land disturbance indicators (cf. McGlone et al., 1994; Wilmshurst, 1997; McGlone and Wilmshurst, 1999; Wilmshurst et al., 1999): (1) a decline in lowland tall forest taxa, an increase in charcoal (from fires), and an increase in _Pteridium_ (bracken) spores and other scrub taxa such as _Aristotelia_, _Coriaria_, _Coprosma_, _Leptospermum_ and grass pollen occurred after the deposition of the Kaharoa Tephra (665 y BP), and after permanent Polynesian settlement and land clearance in the source area of the palynomorphs; (2) spores of hornworts and liverworts (Anthocerotales) signify an increase in erosion since these taxa live on damp exposed soil; (3) swamp clearance by the 1920s is indicated by a decrease in _D. dacyrioides_ (Kahikatea) and taxa like _Cyperaceae_ and _Restionaceae_ (Pullar, 1962); (4) _P. radiata_ plantation forests proliferating in the early 1930s; and (5) grass pollen increased as a result of land clearance and farming the 1960s, in addition to an increase in exotic pollen from _Cupressus_, _Plantago_, _Rumex_, _Salix_, _Taraxacum_, and _Trifolium_.

The 3.5 kHz seismic-reflection data were obtained using a hull-mounted ORE 140 profiler that incorporates a 16-element transducer array. Navigation was provided by GPS and Transit Satellite data. Seismic data were collected on NIWA research cruises 2011 and 3021 (summarised in Foster and Carter, 1997), 3044, TAN0005 (Carter, 2000), and TAN0106 (Lewis, 2001).

The sediment mass accumulation rate over long-time scales was calculated from the total post-glacial sediment volume for the shelf deposit and outer shelf lobe, using 3.5 kHz seismic profiles from Foster and Carter (1997), supplemented with shelf profiles collected on cruises 3044, TAN0005, and TAN0106. To estimate the mass of solid sediment contained within the post-glacial sediment deposit, the volume of water was calculated from the exponential decrease in porosity with depth, using the relationship after Athy (1930) modified by Slater and Christie (1980). Porosity variation to 50 m depth for Poverty Bay was modelled analytically, guided by down-core physical properties measured in mud-dominated sequences from northeastern New Zealand, namely cores ODP 181–1124 (Carter et al., 1999), MD97-2121 (Carter et al., 2002), and shorter piston cores described in Barnes et al. (1991). The model assumes an initial porosity at the surface of approximately 70% based on published near-surface measurements from the region (e.g., Barnes et al., 1991). GIS interpolation between seismic lines was used to determine the volume of the basin, along with the basin area and mean depth. Porosity at the mean depth is taken as representative of the integrated porosity. Therefore, the dry weight equivalent of sediment can be calculated by multiplying the estimated total solid sediment volume by the grain density ($\sim$2.65 g cm$^{-3}$).

3. Results

3.1. Seismic stratigraphy defines the shelf post-glacial sediment deposit

The landward flank of the basin is defined by the actively rising coast (Ota et al., 1987). At the seaward flank the basin is bounded by the emergent Lachlan and Ariel anticlinal ridges, where synclinal muds lie unconformably on the uppermost erosion surface, regarded as the last-glacial/early
transgression erosion surface (W1) (e.g., profiles A and C, Fig. 3; cf. Foster and Carter, 1997). The emergent anticlinal shoals show truncated, steeply dipping tertiary strata of mudstone (e.g., profiles E and G, Fig. 3; cf. Barnes et al., 2002). Units thin seaward towards the Lachlan and Ariel ridge anticlines as a result of tectonic growth of the ridge and progressive rotation of the limbs, contemporaneous with sedimentation and subsidence of the shelf basin, and consistent with that recorded to the south in Hawke’s Bay (Barnes et al., 2002). On the landward side of Lachlan anticline, conformable landward dipping reflectors occur below W1, representing previous lowstand events (profile C, Fig. 3, cf. Lewis, 1973). At the feather edge of Lachlan anticline, the geometry of these older mud deposits is not clear on 3.5 kHz records. However, confirmation of the relative position of the W1 reflector can be established from tracing the reflector in profiles that cross the basin axes. The maximum measured thickness of post-glacial sediment on the Poverty Bay shelf is around 35 m and a maximum projected thickness of ~45 m is attained in the basin depocentre, at approximately 50 m water depth (Fig. 4; cf. Foster and Carter, 1997). In the central portion of the Poverty mid-shelf basin, at sub-seafloor depths of 8–15 m, internal reflectors are commonly masked by gas (profiles A, B and C, Fig. 3).

Seaward of the Poverty Gap, the post-glacial mud sequence is not broken by emergent anticlinal ridges and extends onto the outermost shelf, forming a post-glacial mud lobe of approximately 140 km² and ~3 km² that separates the two shoals (profiles B, D–G, Fig. 3). The isopach surface shows that sediment thickness generally increases seaward of the Poverty Gap (Fig. 4) and extends onto the uppermost slope to ~170 m water depth. A sub-basin occurs in the portion of the lobe south of an upper-slope gully head (profile D, Fig. 3), within which a maximum measured sediment thickness of 40 m is attained above the W1 reflector. A relative offset in the H and W1 reflectors indicates that the northern extent of this outer-shelf sub-basin is bounded by a fault (profile F, Fig. 3), which has created additional accommodation space concomitant with sedimentation. Elsewhere, sediment lobe thickness is predominantly <15 m tapering to the northeast (Fig. 4), and eroded Neogene strata crop out on the sea floor at the northeastern-most edge (profile G, Fig. 3).

3.2. Century-scale accumulation rates from geochemical tracer chronology

Sedimentation rates determined from excess $^{210}$Pb profiles indicate that vertical accumulation rates increase towards the shelf break, reaching a maximum of 0.93 cm y$^{-1}$ at the outer shelf (Fig. 5). No net accumulation is apparent over the last century at the inner-middle shelf. A rate of 0.42 cm y$^{-1}$ was determined at the mid-shelf depositional basin, which is consistent with a rate of 0.55 cm y$^{-1}$ using excess $^{210}$Pb from core MD97-2122 (Gomez et al., 2001, 2004). Sedimentation rates are an order of magnitude slower on the slope, ranging from 0.13 cm y$^{-1}$ on the upper Paritu Trough (W697) to 0.09 cm y$^{-1}$ on the lower Paritu Trough (W699) and flanks (W698), giving an average mass accumulation rate for the slope of approximately 0.06 g cm$^{-2}$ y$^{-1}$.

An X-radiograph (Fig. 5) of the inner mid-shelf core U2304 shows intercalated sand laminae, an absence of bioturbation, and no measurable accumulation rate. These characteristics are indicative of an energetic setting, with wave-induced sediment resuspension and redistribution. In the mid- and outer shelf cores, U2305 and U2306, respectively, no physical structures are evident, making it difficult to determine whether the homogenous structure is massively uniform because of steady deposition or bioturbation, but small burrows suggest the latter. Hence, in the absence of a second tracer, e.g., $^{137}$Cs for verification, the $^{210}$Pb rates are maximum rates because of the possibility of diffusive mixing (i.e., bioturbation).

X-radiographs (Fig. 5) from slope cores W698 and W699 exhibit sedimentary structures consistent with significant bioturbation: W698 shows evidence of large burrowers while W699 contains small burrows. Hence, $^{210}$Pb rates are again maximum estimates of the true rates. $^{137}$Cs activities in the same cores are just above, at or below detection limits, and $^{137}$Cs does not appear to penetrate as deeply as $^{210}$Pb, suggesting that the rates are not dominantly due to mixing, but these results are inconclusive.

3.3. Palynological record of colonisation and slope sedimentation

Spores and pollen type varied in preservation characteristics, and as a result estimates of abundance from pollen counts (excluding pollen from herbaceous taxa) are semi-quantitative. The samples
Fig. 4. Isopach map (in metres) of the last-glacial/transgressive erosion surface (W1) on the Poverty Bay continental shelf. Interpreted contours are based on 3.5 kHz profiles from NIWA cruise 3021 (modified after Foster and Carter, 1997), and new seismic data collected from NIWA cruises TAN0005 (Carter, 2000) and TAN0106 (Lewis, 2001). Sediment thicknesses have been inferred where seismic subbottom penetration is limited by gas. A fault appears to offset the W1 reflector in the southern portion of the outer shelf sediment lobe, forming a small sub-basin. The orientation of key seismic profiles (A–G) shown in Fig. 3 are indicated.
contain abundant charcoal fragments (from fires) and palynomorphs including: spores, pollen, dinoflagellates, fungal spores and hyphae, and other algal cysts. In addition, amorphous and cellular plant material was present, including cuticles, wood, fibres, etc. Angiosperm pollen was conspicuous by its scarcity, and fern spores and conifer pollen by their over-representation. The uppermost portions of all three multicores (Fig. 6) show a clear palynological fingerprint of post-European
deforestation of native podocarp forests, land clearance for pasture, and the establishment of exotic timber forests.

In general, the uppermost few centimetres of the cores contain high concentrations of Pinus (pine), bracken, grasses, some Cupressus (macrocarpa), evidence of soil erosion (hornworts and liverworts), and no Kahikatea pollen. Land disturbance indicators are minimised near the base of the cores with little or no indication of Polynesian settlement. All samples from core W697 show evidence of accelerated erosion and fires indicating human occupation and deforestation (Fig. 6). Early-European influences first appear at 21–22 cm depth. In addition, Salix (willow) pollen is common in the top few centimetres of W697, and at 11–12 cm depth typical disturbance indicators occur but still include Kahikatea pollen, suggesting that significant stands existed in lowland alluvial swamps. The deepest sample from 52 cm contains a reduced concentration of Pteridium (bracken) spores.

European disturbance indicators first appear in core W698 at about 14 cm depth (Fig. 6). The middle section of the core generally has a pollen assemblage consistent with pre-European, but post-Polynesian settlement. However, the absence of scrub taxa such as Coriaria spp. (Tutu), the scarcity of Pteridium spores, and the relative lack of bryophytes from 41 to 42 and 55 to 56 cm depth, respectively, is more consistent with an age of about 665 years or younger, when forests were only minimally affected by Polynesian burning and erosion had not exposed bare tracts of land. Apparent sedimentation rates fluctuate throughout the sequence. In addition dinoflagellates are far more numerous relative to spores and pollen, and
their poor preservation is probably reflected in the higher proportion of thick walled *Cyathea* spores relative to the much thinner walled *Pteridium* spores. The frequency and diversity of recycled material in this sequence is similar to that of multicore W698. *Coriaria*, numerous lycopod and bryophyte spores (evidence of erosion and open muddy surfaces), and abundant *Pteridium* spores occur at 14 cm, providing the first evidence of substantial changes in the lowland podocarp forests by Polynesian burning and erosion. However, evidence of fires occurs throughout the sequence. The taxa at the base of the cores W698 and W699 (51–54 cm depth) are generally representative of an undisturbed and stable landscape prior to the arrival of Polynesian settlers, containing an assemblage of tall podocarp forests (including Kahikatea), and an absence of scrub taxa such as *Coriaria*, grasses, bracken, and soil erosion products.

4. **Discussion**

4.1. **Changes in sub-millennial-scale sedimentation across the Poverty Bay margin**

An excess $^{210}$Pb vertical accumulation rate of $\sim 0.9 \text{ cm yr}^{-1}$ (equivalent to a mass accumulation rate of $\sim 0.6 \text{ g cm}^{-2} \text{ yr}^{-1}$) on the outer shelf lobe, seaward of the Poverty Gap, is approximately 2–3 times greater than the estimated post-glacial and Holocene sediment mass accumulation on the mid-shelf from teprostratigraphy (Gomez et al., 2004) and seismic thickness. An explanation for this dispersal pattern is supported by the regional oceanography. A 45 day-long current meter record from the middle shelf off Poverty Bay has a northeastward residual current of $3.4 \text{ cm s}^{-1}$ for the upper ocean. This direction is in accord with the WCC. However, at the seabed, the residual current is southeastward at $5.9 \text{ cm s}^{-1}$, according to Stephens et al. (2001). Thus, there was a persistent off-shelf flow which presumably would facilitate dispersal of near-bed suspended load towards the Poverty Gap. Furthermore, during floods associated with the eastward passage of low-pressure systems across northern New Zealand, winds rapidly revert via a clock-wise swing to prevailing northwest offshore winds, which result in the offshore migration of surface waters, enhanced by surficial stratification from freshwater inputs (Stephens et al., 2001).

Farther offshore across the Paritu Trough (Fig. 2) a similar pattern of elevated post-colonisation sedimentation is evident, but with excess $^{210}$Pb rates 3–10 times less than those recorded from the shelf. A smaller 1–2 times increase in post-settlement accumulation is recorded on the slope when compared to teprostratigraphy (cf. Orpin, 2004). These spatial and temporal differences in accumulation rate can be readily portrayed as a bar chart spanning the different time scales and margin settings (Fig. 7). Collectively, these observations...
emphasise a consistent pattern of broad, margin-
wide mud dispersal.

Sub-millennial-scale slope sedimentation rates and off-shelf sediment dispersal can also be determined from the pollen record. Drainage and clearance of swamps for pasture had finished by 1920 (Pullar, 1962), which suggests that the pollen assemblage observed in slope core W697 at 11–12 cm depth is representative of the early 1900’s. Using this age, the accumulation rate for the last 100 years is approximately 0.1 cm y\(^{-1}\), which is compatible with the excess \(^{210}\)Pb rate (Fig. 7). Farther offshore in the Paritu Trough, the high pollen and spore concentrations in core W699 relative to the lycopodium standard suggest that here too sedimentation rates increase up core. The base of cores W698 and W699 have a pollen assemblage consistent with pre- to early-Polynesian settlement (Fig. 6). Significant European disturbance indicators are limited to the top 10 cm of the core, inferring that ~40 cm of sediment represents the time interval from approximately 1300–1900 AD. Hence, the post-Polynesian settlement accumulation rate is ~0.06 cm y\(^{-1}\), which is compatible with early- to mid-Holocene rates determined from tephra chronology (cf. Orpin, 2004), as summarised in Fig. 7. Confirmation of the post-Polynesian mass accumulation rate on the upper slope is shown in core W703, where the occurrence of Kaharoa Tephra (~665 y BP) at 30 cm depth yields an estimated sedimentation rate of 0.05 cm y\(^{-1}\) to the core top (Orpin, 2004). This analysis of the pollen record suggests that the sediment accumulation rate on the mid-slope increased by approximately two times upon European settlement and deforestation of the Waipaoa River catchment, and thus broadens the spatial extent of post-settlement sedimentation, documented previously from pollen records from the Poverty mid-shelf only (Wilmshurst et al., 1999; Gomez et al., 2004; Mike Elliot, NIWA, unpublished data).

Tephrostratigraphic indicators of sedimentation over millennial scales across the margin show that sediment delivery and mass accumulation rates were steady during mid-Holocene, both for the Poverty mid-shelf basin (Gomez et al., 2004) and in the Paritu Trough (Orpin, 2004), as summarised in Fig. 7.

4.2. Tectonic control of Holocene mud deposition on the outer shelf

Concave-up cross-shelf profiles visible in the southern portion of the shelf (profile C, Fig. 3; and Foster and Carter, 1997) suggest that the mid-shelf basin has maintained accommodation space concomitant with ongoing sediment supply. The cross-shelf gradient through the Poverty Gap exhibits a flatter profile (profile B, Fig. 3), suggesting that net sedimentation is higher in the central area of the shelf. This pattern of accumulation is consistent with the depocentre of the post-glacial sediment basin (Fig. 4). However, in general, the post-settlement accumulation rates on the mid-shelf are about 2–3 times higher than those determined from the stratigraphic thicknesses. The increase might be up to five times higher on the outer shelf where vertical \(^{210}\)Pb accumulation rates are up to ~1 cm y\(^{-1}\) (Fig. 7). Evidence of significant sedimentation seaward of the mid-shelf is indicated by the thickened outer shelf sediment lobe (profiles B, D–G, Fig. 3), particularly in its southern portion. Here, an offset in the H and W1 reflectors suggest that Holocene (and perhaps transgressive) sedimentation occurred sympathetically with fault activity and the creation of accommodation space, implying that sedimentation was not supply limited, and reinforces the important role of tectonics in shelf sediment distribution. We also infer that significant cross-shelf sediment pathways pre-date the increase in sedimentation resulting from colonisation and deforestation. Note, however, that the outermost shelf could conceivably receive sediment from other sources to the north or south. This part of the shelf will be exposed to changes in coastal circulation driven by dynamic and complex interactions of the main coastal flows (e.g., WCC and ECC) with the regional bathymetry and coastal promontories, Mahia Peninsula in particular.

Tectonic influence on post-glacial sedimentation on the Eel margin shows several similarities with that of the Waipaoa dispersal system. Like the Poverty shelf, recent accumulation of dispersed fine-grained Eel River sediment on the relatively smooth sea floor has been focussed on the mid- to outer shelf (e.g., Sommerfield and Nittrouer, 1999). Lowstand wave planation has truncated structures (anticlines) which have subsequently deformed syndepositionally upon transgressive and highstand deposition. Recent sediment accumulation, measured from excess \(^{210}\)Pb analysis, generally occurs over areas of thick transgressive deposits on the Eel shelf (Sommerfield and Nittrouer, 1999) and slope (Alexander and Simoneau, 1999). However, detailed seismic analysis by Spinelli and Field (2003) suggests that short-term sediment accumulation is not strongly controlled by
the underlying structure and spatially varying accommodation space that results from long-term neotectonism. In contrast, $^{210}$Pb data on the Poverty shelf shows that century-scale sediment accumulation is broadly consistent with areas of the shelf that have undergone long-term syndepositional tectonic subsidence, namely the mid-shelf basin and the outer shelf sediment lobe. This could imply that subsidence rates have kept pace with sediment supply on the Poverty shelf, at least for the Holocene, creating loci for sedimentation that are coincident with underlying structure and accommodation. Currently, this interpretation is limited by the poor spatial coverage of $^{210}$Pb data (particularly along shelf) and the lack of age control on deposits immediately overlying the glacial/early-transgressive erosion surface.

### 4.3. A revised post-glacial sediment budget for the Poverty continental margin

Coastal accretion rate rates from Smith (1988) suggest that sandy beaches along the Poverty Bay coast have grown a total of $\sim 55,000 \text{m}^3 \text{y}^{-1}$, a component of $47,000 \text{m}^3 \text{y}^{-1}$ from the modern Waipaoa River and $8000 \text{m}^3 \text{y}^{-1}$ sourced from surrounding sea cliffs and the Turanganui River (Fig. 2). Using a porosity of 49% (cf. Pryor, 1973) yields a dry bulk density of $\sim 1.4 \text{g cm}^{-3}$, and an accretion rate of $\sim 0.08 \text{Mt y}^{-1}$, which accounts for approximately half of the bedload (1% of suspended load—Hicks and Shankar, 2003) from the modern Waipaoa River. A sand component evident in grain size analyses described in Foster and Carter (1997) suggests that a small amount of bed load escapes the beaches and is transported offshore to at least the inner mid-shelf, but in general bedload can be considered volumetrically insignificant in this mud-dominated dispersal system.

For the Poverty Bay shelf and outer-shelf lobe the volume of the sediment deposit above the last-glacial erosion surface (identified as W1 from the modern Waipaoa River and 8000 m$^3$ y$^{-1}$ sourced from surrounding sea cliffs and the Turanganui River (Fig. 2). Using a porosity of 49% (cf. Pryor, 1973) yields a dry bulk density of $\sim 1.4 \text{g cm}^{-3}$, and an accretion rate of $\sim 0.08 \text{Mt y}^{-1}$, which accounts for approximately half of the bedload (1% of suspended load—Hicks and Shankar, 2003) from the modern Waipaoa River. A sand component evident in grain size analyses described in Foster and Carter (1997) suggests that a small amount of bed load escapes the beaches and is transported offshore to at least the inner mid-shelf, but in general bedload can be considered volumetrically insignificant in this mud-dominated dispersal system.

For the Poverty Bay shelf and outer-shelf lobe the volume of the sediment deposit above the last-glacial erosion surface (identified as W1 from 3.5 kHz seismic sections, Fig. 3) is $\sim 18 \text{km}^3$, with a surface area of 980 km$^2$, and a total solid sediment mass of 15,300 Mt. This approximates to a sediment mass accumulation over the last 18,000 y of $\sim 0.9 \text{Mt y}^{-1}$ (of dry sediment). This equates to an average vertical accumulation rate at the mid-shelf depocentre of $\sim 0.2 \text{cm y}^{-1}$ (Fig. 7) and an average mass accumulation of $\sim 0.2 \text{g cm}^{-2} \text{y}^{-1}$. Given the lack of age control on the basal section of post-glacial fill, a younger transgressive age of 14,000 y increases the accumulation rate by 20% to $\sim 1.1 \text{Mt y}^{-1}$. Combined with an uncertainty of $\pm 10 \text{m}$ ($\sim 20\%$) in the deep regions of the isopach surface from gas-masking, the error in this mass accumulation estimate is large, perhaps $\pm 40\%$.

Adopting a similar approach to calculate the Holocene mass accumulation, the volume of the sediment deposit above the early-Holocene (8000 y BP) seismic reflector H (Fig. 3) to the modern sea floor is $7 \text{km}^3$ with a mass of 5600 Mt (using an average dry bulk density of 0.8 g cm$^{-3}$). This equates to an early-Holocene sediment mass accumulation of $\sim 0.7 \text{Mt y}^{-1}$. This estimate is better constrained than the post-glacial volume due to improved seismic imaging and accuracy of the isopach surface ($\pm 2 \text{m}$) and age control ($\pm 1000 \text{y}$ or 10%) of the early-Holocene reflector, resulting in an error in the Holocene mass accumulation rate of $\pm 20\%$.

From tephrostratigraphy Orpin (2004) estimated that the average sediment mass accumulation rate on the Paritu Trough is $\sim 0.05 \text{g cm}^{-2} \text{y}^{-1}$ (0.04–0.07 cm y$^{-1}$), and by applying this rate over a 450 km$^2$ area of the slope, identified as being zones of hemipelagic sedimentation from echo character and acoustic imagery, speculated that the total mass accumulation to the slope is around 0.2 Mt y$^{-1}$. This estimate is likely to be conservative if a wider area of the slope flanks also receives any hemipelagic sediment. Equally, some fraction of the hemipelagic drape could conceivably be sourced from rivers further north on the Raukumara Peninsula (e.g., Uawa/Hikuwai and Waiaupu Rivers), and transported south by the ECC.

From a summation of coastal bedload (0.08 Mt y$^{-1}$), shelf (0.7 Mt y$^{-1}$) and slope (0.2 Mt y$^{-1}$) sediment mass accumulations, the total sediment mass accumulation for the Holocene is approximately 1 Mt y$^{-1}$. This mass accumulation estimate is an order of magnitude smaller than the current Waipaoa sediment supply of 15 Mt y$^{-1}$ (cf. Hicks et al., 2000), and approximately half the post-glacial basin accumulation rate first estimated by Foster and Carter (1997), due largely to the application of a dry versus a wet bulk density in this study. The wider implication of this analysis is that riverine sediment supply today far exceeds that suggested from millennial-scale margin sedimentation. Given the extent of post-settlement deforestation and erosion this trend would appear broadly consistent. Following a similar line of reasoning for century-scale sedimentation, even applying the highest $^{210}$Pb sediment accumulation rate of 0.6 g cm$^{-2} \text{y}^{-1}$ recorded in this study (core U2306, Fig. 5) across
the whole shelf (980 km$^2$) gives an estimated maximum mass accumulation rate of $\sim 6$ Mt y$^{-1}$, accounting for less than half of the modern sediment supply. This approach suggests that, in contrast the Holocene, under modern conditions a larger proportion of the Waipaoa sediment dispersal system extends onto the slope and beyond.

Consistent with millennial-scale sedimentation rates determined from tephrostratigraphy, post-glacial accumulation rates from basin volume estimates, and measured from maximum Holocene seismic thickness (Fig. 4) infer that sedimentation on the shelf, and by implication shelf sediment retention, was for the most part steady since the start of the last glacial transgression (Fig. 7). Here on the Poverty shelf, high fluvial input to a relatively narrow shelf would provide a consistent supply of mud to the mid-shelf, in concert with steady subsidence of the Poverty shelf basin, growing anticlinal ridges, and beheading of escape canyons. Other regional records from northeastern New Zealand indicate that this pattern of sedimentation may be peculiar to Poverty Bay. Evidence from core MD97-2121 at 2314 m water depth off southern Hawke’s Bay, imply an increase in shelf retention of terrigenous sediment since the mid-Holocene (Carter et al., 2002). On the wide Hawke’s Bay shelf, sediment was likely reworked and lost during the early stages of the transgression, but by the early Holocene an outer shelf barrier in the form of a proto-Lachlan Ridge favoured shelf accumulation. Again, this contrast reinforces the significance of tectonic-sediment interactions along this portion of the northeastern New Zealand margin (e.g., Lewis et al., 2004), and the localised effects of syndepositional subsidence adjacent to large muddy rivers observed elsewhere (e.g., Burger et al., 2001).

5. Conclusions

A mud lobe occurs on the outermost shelf, extending through the Poverty Gap seaward of the mid-shelf, and forms a post-glacial deposit up to 40 m thick in a structurally controlled sub-basin. This sediment lobe occurs immediately upslope of a canyon head and several upper-slope gullies. A relative offset in reflectors suggests that sedimentation occurred sympathetically with tectonic activity and the creation of accommodation space, implying that sedimentation was not supply limited.

Across the margin, the highest accumulation rate of 0.9 cm y$^{-1}$ measured in this study occurs on the outer shelf sediment lobe, approximately an order of magnitude higher than accumulation rates estimated from the slope.

The pollen record from the Paritu Trough fingerprints Polynesian, followed by European settlement, and subsequent land-use intensification and deforestation in the 20th century. These results broaden the spatial extent of post-settlement sedimentation initially documented from the Poverty mid-shelf.

Changes in sub-millennial sedimentation infer there is a 2–3 times increase in post-settlement accumulation on the shelf but a smaller 1–2 times increase on the slope, reinforcing the broad extent of Waipaoa sediment dispersal across the Poverty Bay margin. However, over longer time scales, seismic evidence of syndepositional subsidence infers that significant cross-shelf sediment pathways pre-date the increase in sedimentation resulting from colonisation and deforestation.

From a summation of coastal bedload, shelf and slope sediment mass accumulation for the Poverty Bay margin, the total sediment mass accumulation for the Holocene is approximately 1 Mt y$^{-1}$.

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