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POSTGLACIAL (AFTER 18 KA) DEEP-SEA SEDIMENTATION ALONG THE HIKURANGI SUBDUCTION MARGIN (NEW ZEALAND): CHARACTERISATION, TIMING AND ORIGIN OF TURBIDITES

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ABSTRACT

Recent sedimentation along the Hikurangi subduction margin off northeastern New Zealand is investigated using a series of piston cores collected between 2003 and 2008. The active Hikurangi Margin lies along the Pacific-Australia subduction plate boundary and contains a diverse range of geomorphologic settings. Slope basin stratigraphy is thick and complex, resulting from sustained high rates of sedimentation from adjacent muddy rivers throughout the Quaternary. Turbidites deposited since c. 18 ka in the Poverty, Ruatoria and Matakaoa re-entrants are central to this study in that they provide a detailed record of the past climatic conditions and tectonic activity. Here, alternating hemipelagite, turbidite, debrite and tephra layers reflect distinctive depositional modes of marine sedimentation, turbidity current, debris flow and volcanic eruption, respectively. Turbidites dominate the record, ranging in lithofacies from muddy to sandy turbidites, and include some basal-reverse graded turbidites inferred to be derived from hyperpycnal flows. Stacked turbidites are common and indicate multiple gravity-flows over short time periods. The chronology of turbidites is determined by collating an extremely dense set of radiocarbon ages and dated tephra, which facilitate sedimentation rate calculation and identification of the origin of turbidites. Sedimentation rates range from 285 cm/ka during late glacial time (18.5-17 ka) to 15 to 109 cm/ka during postglacial time (17-0 ka). Turbidite deposition is controlled by: (1) the emplacement of slope avalanches reorganizing sediment pathways; (2) the postglacial marine transgression leading to a five-fold reduction in sediment supply to the slope due to disconnection of river mouths from the shelf edge, and (3) the Holocene/Pleistocene boundary climate warming resulting in a drastic decrease in the average turbidite grain-size. Flood-induced turbidites are scarce: nine hyperpycnites are recognized since 18 ka and the youngest is correlated to the largest ENSO-related storm event recorded onland (Lake Tutira). Other turbidites contain a benthic foraminiferal assemblage which is strictly reworked from the upper slope and which relate to large earthquakes over the last c. 7 ka. They yield a shorter return time (270-430 years) than the published coastal records for large earthquakes (c.670 years), but the offshore record is likely to be more complete. The deep-sea sedimentation along the New Zealand active margin illustrates the complex interaction of tectonic and climate in turbidite generation. Climate warming and glacio-eustatic fluctuations are well recorded at a millennial timescale (18 ka), while tectonic deformation and earthquakes appear predominant in fostering turbidite production at a centennial timescale (270-430 years).

Keywords: hyperpycnite; earthquake; debris avalanche; marine transgression; triggering mechanism; sediment cores.

1. INTRODUCTION

Gravity-driven flows are ubiquitous and fundamental process that control sediment dispersal where steep bathymetric gradients, enhanced tectonic activity and voluminous terrigenous sediment supply prevail such as at active margins. They range from submarine avalanches, cohesive debris or grain flows, liquefied and fluidized flows and turbidity currents (Stow and Mayall, 2000; Stow et al., 1996). Such processes can generate complex sets of sedimentary structures from a variety of triggering mechanisms and scales including giant avalanches consisting of >100 km^3 of lithified sediment (e.g. Collot et al., 2001; Canals et al., 2004), thick successions of density-variable turbidites (Bouma, 1962; Stow and Shanmugam, 1980; Lowe, 1982), to centimetre-thick hyperpycnites that can be linked to individual flood events (Mulder et al., 2003). As such, gravity flow deposits contain invaluable information about past stratigraphic, climatic and tectonic history (Adams, 1990; Goldfinger et al., 2003; St Onge et al., 2004; Blumberg et al., 2008; Noda et al., 2008; Nakajima et al., 2009). However, due to the geomorphologic complexity of active margins lateral correlation of events is often problematic, both in terms of dealing with the spatial variability of gravity events and recognising synchronous “event assemblages”.

The active Hikurangi convergent margin, New Zealand is an excellent locality for the study of gravity-driven events because of the diversity of geomorphological settings, the intense tectonic activity (e.g., Lewis and Pettinga, 1993; Collot et al., 1996) and the high rates of sedimentation that produce an expanded stratigraphic record at an exceptional resolution. As the Hikurangi Margin lies along the Pacific-Australia subduction plate boundary, it is subjected to intense seismic activity. Here, a well documented upper-plate earthquake record exists for magnitude $M_w < 7.8$ (Reyners, 1998; Webb and Anderson, 1998) but only a poorly documented record of inferred plate interface ruptures capable of generating ~M_w 8.8 earthquakes (Reyners, 1998; Reyners and McGinty, 1999; Wallace et al., 2009; Cochran et al., 2006). At the northern extent of margin, intense mass wasting and margin-collapses activity is manifested as large morphological re-entrants in the continental slope (Collot et al., 2001; Lamarche et al., 2008a; Pedley et al., 2010). Due to the vigorous maritime climate, floods are a common feature of northeastern New Zealand (Hicks et al, 2004). Some prehistoric catastrophic floods have been inferred from river flood-plains and continental shelf sediments (Brown, 1995; Gomez et al., 2007; Brackley et al., 2010) which might be capable of rapidly transporting sediment directly from the coast to slope basins via hyperpycnal flows. The occurrence of numerous tephra originating from the Central Volcanic Zone (Fig. 1) provide excellent chronological control in the offshore stratigraphic record (e.g. Carter et al., 2002). The northern Hikurangi Margin was intensely studied over the last 20 years, and contributed to a robust understanding of long- and short-time scale tectonic deformation (Collot et al., 1996; Reyners, 1998; Reyners and McGinty, 1999), sedimentary processes and stratigraphy (Foster and Carter, 1997; Joanne et al., 2010; Orpin, 2004, Gomez et al., 2007; Paquet et al., 2009; Kniskern et al., 2010) and Holocene sediment budgets (Orpin et al., 2006; Alexander et al., 2010; Gerber et al., 2010; Paquet et al., 2011). But the thick and complex suite of Quaternary turbidites that infill the slope basins remain largely understudied and their event stratigraphy underutilised.

In this paper, we use a series of sediment cores collected in the Poverty, Ruatoria and Matakaoa re-entrants along the northern Hikurangi Margin to identify and characterise a complete and comprehensive series of turbidite events. We generate a chronology of catastrophic sedimentation over the last 20,000 years for the northeastern Hikurangi Margin, and detailed characterisation of turbidites is used to compare and contrast depositional patterns. The excellent chronological control afforded by tephra and radiocarbon dating allows us to develop a methodology for investigating turbidite origin, and determine the relative contribution of trigger and controlling mechanisms. The balance of these processes are likely to be applicable to active margins globally. The study suggests that large earthquakes, catastrophic floods and volcanic eruptions are the principal triggering mechanisms of turbidites in the deep water sedimentary systems, and that over the past 20 ka, turbidite systems activity was primarily controlled by glacio-eustatic fluctuations and basin morphology.

2. Geological and Sedimentological Settings

2.1. Geomorphology

The Hikurangi Margin marks the region where the oceanic crust of the Pacific Plate is being subducted obliquely beneath the Raukumara Peninsula (Fig. 1). The zone of active deformation covers from east to west, the Hikurangi Trough, the continental slope and shelf and the east coast of the North Island of New Zealand (Lewis, 1980; Lewis and Pettinga, 1993; Collot et al., 1996). Subduction-related underplating beneath the Raukumara Peninsula is actively uplifting the axial ranges at an estimated maximum rate of 3 mm/a (e.g. Reyners and McGinty, 1999). A narrow accretionary prism forms locally at the toe of the slope. To the west lies the rhyolitic Central Volcanic Zone which is a prolific source of geochemically-distinct tephra that punctuate the terrestrial and offshore stratigraphic record throughout the Quaternary (Lowe et al., 2008).

The northern Hikurangi Margin includes a flat, 20-30 km-wide continental shelf, a steep sediment-starved slope, and a 3500 m-deep subduction trough (Fig. 1). Tectonic erosion has produced three large slope avalanches: the 30-50 km-wide, Poverty re-entrant (Pedley et al., 2010); the 30-40 km Ruatoria re-entrant (Collot et al., 2001); and landward of the trench wall and immediately north of the Raukumara Peninsula, the Matakaoa passive margin contains the 50 km-wide Matakaoa re-entrant (Lamarche et al., 2008a). Elsewhere smaller debris slides, slumps and head-wall scarps are abundant, indicating ongoing slope instability (e.g. Lewis et al., 1998). The current study focuses on sediment cores within the Poverty, Ruatoria, and Matakoa re-entrants. The 1500 km^2 Poverty re-entrant is a major continental margin depression resulting from successive margin collapses since 1,500±500 ka (Pedley et al., 2010). The bathymetry of the Poverty re-entrant is complex and comprises several basic morphologic components (Orpin, 2004) including: a heavily gullied upper slope; the beheaded Poverty Canyon System; the gently sloping mid-slope Paritu Trough; margin-parallel North and South Paritu Ridges that are cross-cut by a small canyon feeding into the Lower Paritu Basin (Fig. 2). The Paritu Trough is filled with the Poverty Debris Avalanche (PDA), which is blanketed by sediments. Although the PDA is undated, the surface of the avalanche is rough and hummocky suggesting a recent event. The 3300 km^2 Ruatoria re-entrant formed following a giant debris avalanche 170±40 ka ago (Collot et al., 2001). The re-entrant consists of a gullied upper slope, a vast highly chaotic debris avalanche composed of individual blocks of several cubic kilometres in size, and the subduction trough (Fig. 3). The 1000 km^2 Matakaoa re-entrant resulted from multiple mass transports events, which occurred between 1,300 and 35 ka ago (Carter, 2001; Lamarche et al., 2008a; Joanne et al., 2010) (Fig. 4). The eastern half of the re-entrant is infilled by the Matakaoa Turbidite System (MTS), which developed subsequently to the Matakaoa Debris Avalanche, 600±150 ka ago (Joanne et al., 2010). The MTS is a classical channelized turbidite system with a canyon incising into the shelf break, a well-developed channel/levee turbidite plain and a fan growing in the Raukumara Plain.

In the Hikurangi Trough, the 2000 km-long Hikurangi Channel drains large turbidity currents parallel to the North Island East Coast (Lewis et al., 1994; 1998; Lewis and Pantin, 2002) (Fig. 1). At the latitude of the Poverty re-entrant, the channel is redirected sharply eastward (Fig. 1). There, well developed overbank sediment waves grew over the last 2 Ma, due to the combined effect of centrifugal and southern hemisphere Coriolis force. Sediment waves in the channel axis are comprised of stacked coarse turbidites overlain by a hemipelagic drape, suggesting limited activity during interglacial periods with episodic flows contained into the channel.

2.2. Sedimentology

Up to a kilometre of Quaternary sediment fill accumulates in ponded basins along the northern Hikurangi Margin continental shelf (Lewis et al., 2004) and slope (Orpin, 2004; Orpin et al., 2006; Paquet et al., 2009) as well as in the Hikurangi Trough (Lewis and Pettinga, 1993) and Raukumara basin (Kohn and Glasby, 1978) (Fig. 1). The mass accumulation rate along the margin is generally high over the last 1 My (4 Mt/a in Hawkes Bay), with millennial variations over glacio-eustatic cycles (Carter and Manighetti, 2006; Paquet et al., 2009). Over the last 30 ka, the highest rates were recorded during last-glacial lowstand through to the early highstand stage (30-7 ka). The Holocene highstand period (7-0 ka) shows a declining flux to the lower continental slope as more sediment is retained in subsiding shelf basins (Gerber et al., 2010) and baffled in intra-slope basins bounded by imbricate thrust ridges (Lewis et al., 1998; Paquet et al., 2011), where the hemipelagic flux is around 60 cm/ka since the mid–late Holocene (Orpin, 2004).

Driven by the vigorous maritime climate across the Raukumara Ranges, the present day sediment flux delivered to the adjacent shelf and slope basins is 70 Mt/a. Forest clearing by early Polynesian settlers 500-700 y BP and then by European colonisation in the mid-eighteenth century resulted in present day river sediment fluxes an order of magnitude greater than pre-human colonization (e.g. McGlone et al., 1994; McGlone and Wilmshurst, 1999). Paquet et al. (2009) estimated an increase of 110-250% in Hawkes Bay whereas Kettner et al. (2007) calculated a rise of 660% for the Waipaoa river alone (Fig. 1).

Regional oceanography plays a major role in the offshore dispersal of sediments from Raukumara rivers (Fig. 1). On the continental shelf, swell waves, wind direction, the northward-flowing Wairarapa Coastal Current (WCC) and large ephemerical gyres affect current direction (Foster and Carter, 1997; Chiswell, 2000). Beyond the shelf break, the southward-flowing East Cape Current (ECC) is the dominant current affecting the region during the Holocene (Stanton, 1998; Stanton et al., 1997; Carter et al., 2002). During the Last Glacial Maximum (LGM), the ECC strength decreased while the proto-WCC, flowing northward near the shelf break, increased (Carter and Manighetti, 2006). Deep circulation in the Hikurangi Trough is influenced by the Southwest Pacific Deep Western Boundary Current (DWBC). The main flow of the DWBC is confined by the northeast scarp of the Hikurangi Plateau, but a shallower westward-flowing branch reaches the Hikurangi Trough at Poverty Bay where it deviates northward and joins with the main DWBC over the Kermadec trench (McCave and Carter, 1997).

The New Zealand terrestrial and marine climate record over the past 30 ka shows three climatic intervals: (1) the Last Glacial Cold Period between 28 and 18 ka, which includes the Last Glacial Maximum at 21±3 ka (Mix et al., 2001; Barrows et al., 2002); (2) the Last Glacial Interglacial Transition extending from 18 to 11.6 ka, including the Late Glacial Climate Reversal (13.5-11.6 ka), which extends from the early Antarctic Cold Reversal to the end of the Younger Dryas; and, (3) the Holocene Interglacial stage from 11.6 ka to present (Alloway et al., 2007). Glaciers during the Last Glacial Cold Period did not reach the Raukumara Ranges (McArthur and Shepherd, 1990; Pillans et al., 1993; Brook and Brock, 2005). Palynological studies demonstrate a strong climatic impact on East Coast vegetation, with grass and shrub dominating during cold and dry conditions at the Last Glacial Cold Period and large stands of podocarp and hardwood forest prevailing during warm and moist conditions of the Holocene (McGlone, 2001; Okuda et al., 2002; Mildenhall and Orpin, 2010). The protection provided by vegetative cover is an important control on erosion rates in the region (Page et al., 2004; Litchfield and Berryman, 2005).

3. DATA AND METHODS

3.1. Collection of sediment cores

Sixteen sedimentary cores are used in the current study, collected in water depths ranging from 650 to 3520 m below sea level (mbsl; Table 1, Fig. 1). Four of these are giant piston cores collected from the Poverty and Ruatoria re-entrants during the MD152 MATACORE voyage of *R.V. Marion-Dufresne* (Proust et al., 2006). Twelve short piston cores were acquired in the Ruatoria and Matakaoa re-entrants onboard *R.V. Tangaroa* research voyages TAN0314 (Carter et al., 2003) and TAN0810 (Lamarche et al., 2008b).

High-resolution 3.5 kHz seismic reflection data and multibeam bathymetry were systematically acquired prior to coring in order to ascertain the suitability of the sampling sites, providing subsurface stratigraphic information up to 20 m below the seafloor with a vertical resolution of <1 m. The bathymetry is compiled from data acquired during the Geodynz survey using the 12 kHz EM12 echo-sounder of R/V L'Atalante (Collot et al., 1996) and a large number of surveys using the 30 kHz Kongsberg EM300 echo-sounder of R/V Tangaroa with an optimal accuracy of ~0.2 % of the water depth. The margin morphology is provided by Digital Terrain Models (DTM) generated from the multibeam bathymetry database maintained at NIWA (CANZ, 2008).

Sedimentary cores targeted recent gravity sedimentary activity, including intra-slope basins fed by turbidite flows, aprons of avalanche debris and intra-canyon levees. In the Poverty re-entrant, two giant piston cores were collected in the Paritu Trough (MD06-3003) and the Lower Paritu Basin (MD06-3002, Fig. 2). In the Ruatoria re-entrant, sediment cores were collected on the gullied upper slope (Tan0810-1, -2, -3, -5), on the Ruatoria Debris Avalanche (MD06-3009) and in the Hikurangi Trough (MD06-3008, Tan0810-6) (Fig. 3). In the Matakaoa re-entrant, short cores were collected along the Matakaoa Turbidite System (MTS), on the canyon floor (Tan0314-86), in the channel/levee complex (Tan0810-9 and 12 in channel; Tan0810-10, -11 and -13 in levees) and in the deep-sea fan (Tan0314-8) (Fig. 4).

3.2. Sedimentological analyses

Detailed logs were generated for all cores and analyses were undertaken to further characterise the turbidites. Geotek Multi-Sensor Track (MST) analyses were run at University of Otago (New Zealand) to provide continuous gamma density, magnetic susceptibility and P-wave velocity measurements as well as high definition photos of split cores. These measurements were complemented by X-Ray radiographs of split cores, performed using a Varian PaxScan 4030E veterinary digital imaging system from NIWA, to characterise the internal structure of sediments. We performed grain-size analyzes of selected samples using a Beckman-Coulter LS 13 320 Lasersizer (size range of 0.38-2000µm). Physical properties complement the visual descriptions of turbidite events and are critical to refining the location of their boundaries.

Compositional analysis of the silty-clay fraction was undertaken to characterise the transition between turbidite tails and hemipelagite sediments. The coarse and dense silt fraction was extracted by decantation and analysed with a stereomicroscope to provide a semi-quantitative estimate of the main component. The composition of the sand fraction (>53µm) of turbidites was determined following the same semi-quantitative approach on wet sieved 2 cm-thick samples taken at the base of selected turbidites. Benthic foraminifers were then extracted from the medium sand fraction (125-

500µm), to determine the source of the sediments deduced from the distribution of modern benthic foraminifers in New Zealand (Hayward et al., 2010; Camp, 2009).

3.3. Age Dating

Timing and age downcore are provided using tephrochronology and ^{14}C radiochronology. All cores are densely dated with one age every 0.5 to 1.4 m of core. Tephra were systematically sampled and characterised by glass chemistry, mineralogy and stratigraphic position and identified by comparing with the data bank of well-established terrestrial occurrences (Shane, 2000). Tephra ages follow the convention proposed by Lowe et al. (2008). In the channel-levee complex of the MTS, tephra were sampled in three out of the five cores (Tan0810-9, 10 and 12). In the two neighbouring cores (Tan0810-11, 13), identified tephra were correlated using geophysical data and stratigraphic position.

Radiocarbon dating was performed on handpicked mixed planktonic foraminifers at the Rafter Radiocarbon Laboratory, GNS Science. The 0.7-1.0 cm-thick samples were collected in hemipelagite layers, 0.7-1.0 cm below gravity-flow deposits to prevent any contamination and mixing by bioturbation. AMS ^{14}C were calibrated to calendar years by using the MARINE09.14 calibration curve (Reimer et al., 2009) in CALIB Rev 6.0 program (Stuiver and Reimer, 1993), applying an average regional reservoir age of 395±57 years calculated from published East Cape reservoir age (Higham and Hogg, 1995; Kalish, 1993; Calib database at http://calib.qub.ac.uk/marine/). A reservoir age of 800±110 years has been applied for the time of the Waiohau tephra deposition (13,635 cal. yr BP; Table 3), as defined by Sikes et al. (2000) and Carter et al. (2008). The ^{14}C radiochronology calibration is adequate for most of the samples from the Marion Dufresne cores as shown by the good correlation with tephrochronology. Two samples on MD06-3002 located less than 10 cm above the Waiohau Tephra suggest a stratigraphic reversal. Reservoir age modification during that period may explain the inconsistency between AMS ^{14}C and tephra ages. For this study, we prefer the tephra age and discarded the two AMS ^{14}C samples.

3.4. Sedimentation Rates

Based on lithofacies identification, we distinguish uncorrected and corrected sedimentation rates. Uncorrected sedimentation rates includes the total sediment thickness from all lithofacies, whereas corrected sedimentation rate includes only the hemipelagite. Corrected sedimentation rate is calculated by subtracting the thickness of the turbidites and tephra layers from the total sediment thickness, and assumes limited erosion at the base of the turbidite layers. Corrected rate is used here to estimate the age of turbidites. Hemipelagites represent a continuous and steady mode of deposition whereas gravity-driven depositional events are emplaced instantaneously. The Terrigenous Accumulation Rate (TAR) is the difference between uncorrected and corrected sedimentation rates, representing the cumulated thickness of gravity-driven deposits (mostly turbidites) through time. Because of the high density of dated samples in each core, deformation in the piston core does not significantly influence our results and interpretations. Deformation is localised and easily identified in the age model by a change in the slope of the age curve.

4. RESULTS

We define four end-members facies: tephra, debrites, hemipelagites and turbidites. In this section, we describe these facies and provide a detailed description of the turbidites in terms of their composition, foraminiferal content and facies. We subsequently provide age models and sedimentation rates for the Poverty, Ruatoria and Matakaoa re-entrants.

4.1. End-members facies

4.1.1. *Tephra*

All cores contained several tephra composed of 1 to 2 cm-thick (rarely up to 6 cm-thick), pinkish, normally graded silts, capped by a clay-rich bioturbated horizon (Fig. 5). They are exclusively composed of volcaniclastic debris (glass shards and pumiceous lapilli mostly) and identified by their typical colour and high values of magnetic susceptibility (>40 SI). In places, tephra are thoroughly reworked by intense bioturbation, which suggests that the original depositional layer was less than 1 cm-thick, preventing asphyxia of the benthic fauna (Hess and Kuhnt, 1996). In this case, corrected sedimentation rates are calculated using a 1 cm thickness for highly bioturbated tephra. In this study, we assume that all tephra originate from ash-fall coincident with volcanic eruptions (Wiesner et al., 1995; Carter et al., 1995).

In places, tephra are made up of a cm-thick normally-graded lapilli layer, which differs from other tephra by their coarser grain size. These layers are composed of >90% of volcaniclastic grains of monomagmatic origin i.e. coming from the same volcanic eruption, and correspond to primary monomagmatic turbidites as defined by Schneider et al. (2001). Hence, primary monomagmatic turbidites can be treated like airfall tephra as they emplace directly after the eruption, and are therefore datable.

4.1.2. *Debrites*

This end-member facies consist of < 35 cm thick chaotic intervals of dark olive-grey silty-clay with sand, granules, pebbles and occasionally deformed stratified lithoclasts (Fig. 5). The sand to granule size material shows weak reverse grading. Debrites are composed of quartz, volcaniclastic clasts, bivalve and gastropod shells and 2-3 cm-large clasts of poorly laminated silty clays and laminated fine- to medium-clayey silts. This facies is rare, representing only five events in two cores (Tan0810-5 and MD06-3003). The chaotic facies, the absence of well-defined basal erosion and the matrix supported texture, suggest a mass transport deposit from a debris flow (Mulder and Alexander, 2001).

4.1.3. *Hemipelagite*

Hemipelagites consist of heavily bioturbated light olive-grey silty-clay. The silt fraction typically show more than 50% of volcaniclastic grains, mainly pumiceous lapilli and less than 20% of quartz grains (Figs. 5 and 6A). Foraminiferal content shows low and stable values like in turbidite tails and thus cannot be used to distinguish facies. Hemipelagites usually have the finest grain size (<10 µm). This facies is interpreted as the result of deposition by pelagic rain in stable, deep offshore environments. It represents 20% to up to 90% of sediment volume in cores.

4.1.4. *Turbidites*

Together with hemipelagites, turbidites dominate the sedimentary record. There are a maximum of 101, 89 and 20 single turbidites per core in the Poverty, Ruatoria and Matakaoa re-entrants, respectively. Turbidites are recognized by their coarser grain size and a typical fining upward trend (Fig. 5). Thickness ranges from 1–75 cm. Turbidites are usually interbedded with hemipelagites. The basal boundaries are easily identified from a change to coarser-grain size, darker color and increase in density, magnetic susceptibility and P-wave velocity. The top boundary is progressive with bioturbated contact from the turbidte tail, grading into the hemipelagite background. Compositional analysis shows a doubling in quartz grain concentrations in turbidite tails (> 50%) and slightly higher values of rock fragments and micas than in hemipelagite (Fig. 6A). All turbidites in this study are interpreted as deposited by low to medium density turbidity current as defined by Stow and Shanmugam (1980) and Bouma (1962).

Several turbidites can be stacked in sequences over up to 75 cm thick bounded by hemipelagites. These are termed herein “stacked turbidites”, as opposed to “isolated turbidites”, which consist of a single gravity-flow deposit under- and overlain by hemipelagite (Fig. 5). The small thickness of individual turbidites, the lack of thick coarse grain basal unit (<20cm) and the thick uppermost silty-clay unit in stacked turbidites suggests very low erosion at the base of individual gravity events. Hence we infer that the lack of intervening hemipelagite in stacked turbidites is due to non-deposition rather than erosion. These conditions suggest only a short duration of time between successive turbidites.

4.2. Turbidite composition and facies

4.2.1. *Sand Composition*

The turbidite sand fraction is predominantly composed of volcaniclastic grains and angular to rounded light mineral grains of quartz with rare feldspar (Fig. 6B). Volcaniclastic grains include angular and massive type glass shards with rare inner bubbles, and coarse and rounded pumiceous lapilli. Bubble-wall type glass shards are rare. All volcanic glass is fresh implying rapid emplacement after volcanic eruptions and reduced storage time onland or on the shelf. All cores contained a small amount of rock fragments. Other detritic grains include wood fragments, micas and heavy minerals such as pyroxene and hornblende. This class shows generally low values, but high concentrations occur in core MD06-3008 in the Hikurangi Trough where some mica-rich turbidites have been described. Rare bioclastic grains include well-preserved benthic and planktic foraminifers and shell fragments.

The turbidite composition varies between cores in the Ruatoria re-entrant, whereas it is homogeneous in Poverty re-entrant (Fig. 6B). Poverty re-entrant turbidites have higher concentrations of quartz grains compared with the Ruatoria re-entrant turbidites, probably due to the proximity of the coastal rivers (70 km and 100 km, resp.). In terms of water depth, the deepest turbidites in Poverty and Ruatoria re-entrants (Lower Paritu Basin and Hikurangi Trough, respectively) show higher concentrations of light minerals and foraminifers and reduced amount of volcaniclastic grains. In the Matakaoa re-entrant, composition is dominated by volcaniclastic grains with only minor concentrations of light minerals.

4.2.2. *Foraminiferal assemblages*

We identified 28 benthic foraminifera species, of which *Uvigerina peregrina, Bulimina marginata f. aculeata, Evolvocassidulina orientalis, Notorotalia depressa, Bolinita quadrilatera, Globobulimina pacifica* and *Quinqueloculina auberina* largely dominate. These species are indicative of a variety of environments from the inner shelf to the abyssal plain. However, most of them are characteristic of environments seaward of the shelf break (>150m).

We defined four benthic foraminiferal associations from their living water depth (Fig. 7A). The associations are basin dependant. Association 1 (0–200 m) includes shelf species and indicates remobilization of shelf sediments, such as might be expected by storm waves or hyperpycnal flows. Association 1 is only present in the Poverty re-entrant and on the Ruatoria Debris Avalanche. Association 2 (0–600 m) includes species from the shelf and the upper slope, and is characteristic of turbidites from the Matakaoa Turbidite System. Association 3 (0-1200 m) includes species with a depth range shallower than the base of the upper slope and is observed in small quantities in all basins. Association 4 (200–5000 m) has deep water species only and is characteristic of turbidites from Poverty and Ruatoria re-entrants.

The proportion of planktic foraminifers shows a constant increase with depth, and therefore distance from shore, ranging from 32% at c. 1100 mbsl in the Matakaoa re-entrant to 87% at c. 3500 mbsl in the Hikurangi Trough (Fig. 7B).

4.2.3. *Turbidites facies*

Five turbidite facies were determined based on grain size, internal structures, sand composition and foraminiferal assemblage, namely muddy turbidites (T I); silt laminae turbidites (T II); silty turbidites (T III); sandy turbidites (T IV); and reverse-graded basal turbidites (T V) (Fig. 5, Fig. 6C). These are summarised below.

Muddy turbidites (T I) are characterised by dark olive-grey silty-clays, which differ from the hemipelagic background by being more darker and coarser grained (10 to 22 µm) (Fig. 5). Muddy turbidites are 1–40 cm-thick, fining upward sequences with a sharp basal contact and gradational upper boundary, in places overprinted by bioturbation. The occurrence of occasional wavy basal contacts suggest some basal erosion. Typically, muddy turbidites are composed of poorly-laminated silty-clays with occasional basal silt laminae (<1 cm-thick), overlain by massive silty-clay. The composition of sand grains shows a predominance of light minerals (83%), negligible volcaniclastic grains (3%) and a relatively high percentage of foraminifers (10%). The foraminiferal content is predominantly planktic species (85%), with benthic species only occurring as Association 4 (Fig. 6C). Muddy turbidites are interpreted as the upper subdivisions Td, Te of medium density turbidites (Bouma, 1962) or T4 to T8 subdivisions of low density turbidites (Stow and Shanmugam, 1980) deposited by very low density turbidity currents.

Silt laminae turbidites (T II) consist of irregular, <40 cm-thick sequences of interbedded, thinning and fining upward clay and silt laminae (Fig. 5). Silt laminae are usually <1 cm-thick and stacked in sets of 2 to 10. Weak cross-stratification in the basal silt laminae merge up section to planar lamination, followed by poorly laminated, then homogeneous silty-clay. Sand grains are predominantly light minerals (59%) and volcaniclastic grains (30%). Foraminiferal assemblages show a majority of planktic species (80%) and a benthic assemblage dominated by Association 4 (81%) with rare species from Associations 2 (11%) and 3 (8%) (Fig. 6C). Silt laminae turbidites are interpreted as fine-grained turbidites deposited by a low density turbidity current (T2 to T8 subdivisions of Stow and Shanmugam (1980)). They differ from stacked turbidites by their specific grain-size showing a single sequence (Fig. 5).

Silty turbidites (T III) are composed of 0.5–55 cm-thick, fining upward clayey silt sequences (Fig. 5). The basal contact is usually sharp, with little evidence of erosion. A complete graded sequence shows from base to top: (1) a massive coarse clayey silt base; (2) laminated coarse clayey silt; (3) cross-stratified coarse clayey silt; (4) laminated fine to medium clayey silt; and in places, (5) a laminated silt and clay and (6) poorly laminated to homogeneous silty-clay top. Sand grains within silty turbidites are composed of volcaniclastic grains and light minerals (33% each) and a relatively high proportion of foraminifers (9%). In core MD06-3008, some mica-rich silty turbidites (>90% of mica) artificially increase the average value of the class “other detritic grains” (Fig. 6C). Silty turbidites are also characterised by 69% of planktic foraminifers and benthic assemblage is composed of Association 4 (73%) with a minor component from other associations (from 6 to 12% each). These clayey silt sequences are interpreted as the upper subdivisions Tc to Te of medium density turbidites (Bouma, 1962) deposited by low density turbidity currents.

Sandy turbidites (T IV) are characterised by a clean sand at the base, fining upward to clayey and silty sand sequences, typically <75 cm-thick (Fig. 5). A complete graded sequence shows from base to top: (1) massive coarse to fine-grained sand base; (2) laminated fine sand; (3) cross stratified very fine sand; (4) laminated clayey silt; and, (5) poorly laminated to homogeneous silty-clay. In places, the laminated clayey silt interval (i.e. 4) is graded and thick, and shows cross stratification passing upward to horizontal lamination and silt laminae. The basal contact is usually erosive with some evidence of scour. The composition of sand grains show high proportions of light minerals (57%) and volcaniclastic grains (29%), similar to silt laminae turbidites. However, sandy turbidites contain more rock fragments (6%) and other detritic grains (17%) than the other facies. Sandy turbidites have the lowest planktic foraminifers content (49%). Benthic foraminiferal assemblages show a high concentration of Association 4 (51%) with 19% of Associations 2 and 3 and 11% of Association 1 (Fig.

6C). These sequences are interpreted as the Ta to Te subdivisions of medium density turbidites (Bouma, 1962).

Basal reverse-graded turbidites (T V) range from 3 to 45 cm in thickness and are characterised by a reverse graded silty-sand basal unit overlain by a fining upward silty to sandy unit (Fig. 5). The basal contacts are usually sharp. Sharp to irregular contacts are observed between the reverse and normally grading units. The normally-graded upper unit consists of T II, T III or T IV. The T V turbidites are interpreted as deposits from a waxing then waning flow (Kneller, 1995). Approximately half of the T V turbidites display a basal unit composed of light coloured clayey silts with horizontal laminations, abundant large foraminifers (over 20% of the sand fraction) and sparse plant debris. These particular turbidites are labelled T Va, and show a basal texture finer than that of the hemipelagite background (eg. 920-990 cm in MD06-3009, Fig. 5). The boundary between the reverse and the normal graded sections in T Va show a sand grain composition and planktic foraminiferal concentration similar to silt laminae turbidites (TII). The benthic foraminiferal assemblage has high concentrations of Association 2 (49%) and a relatively low concentration of Association 4 (31%) compared to other turbidites (Fig. 6C). We interpret T Va as hyperpycnites as described by Mulder et al. (2003).

4.3. Age controls and sedimentation rates

Ages recovered from core material are compiled in Table 2 and 3 and summarised in Fig. 8. In the Poverty re-entrant core MD06-3003 offers a continuous chronology from c.1 ka to c.16.5 ka, but in MD06-3002 a truncated range from c.6 to c.17 ka was recovered. In the Ruatoria re-entrant MD06-3009 presents the longest record from c.1 ka to c.18 ka, whereas ages in core MD06-3008 range from c. 0.5 ka to c.16.5 ka. All short cores in the Ruatoria and Matakaoa re-entrants exhibited dates spanning a shorter period from c.8 ka to present day except for core Tan0314-8, which shows a truncated range from c.5 to c.17 ka.

Uncorrected sedimentation rates since 17 ka are highly variable on the Hikurangi Margin (Fig. 8A). They range from 15 cm/ka in the MTS deep-sea fan to 109 cm/ka in the Hikurangi Trough. These rates contrast with the 285cm/ka calculated before 17 ka on the Ruatoria Debris Avalanche. The hemipelagite corrected sedimentation rate throughout the Holocene (11.7 ka to present) has a considerably tighter range of 34 to 38 cm/ka along the margin and is variable during the Late Pleistocene (17-11.7 ka) ranging from 8 cm/ka in the Poverty re-entrant to 21 cm/ka in the Ruatoria re-entrant.

We re-calibrated the ages from *Marion Dufresne* core MD97-2121 in southern Hawkes Bay (Carter et al., 2008), following our methodology to yield a revised sedimentation rate (Fig. 8B) of ~37 cm/ka, constant for the last 40 ka. Since, MD97-2121 reportedly only contains hemipelagite sediments (Carter et al., 2008), this value corresponds to the corrected sedimentation rate in that location. This rate is similar to the corrected Holocene sedimentation rates that we calculated in this study suggesting that turbidite deposition did not significantly affect the background sedimentation record. For the Late Pleistocene, the rate of 37 cm/ka contrasts with the observed fluctuations in Poverty and Ruatoria re-entrants (8-21 cm/ka), suggesting either differential erosion by successive gravity flows or a localised decrease of hemipelagite sediment fluxes. Considering the coarser grain size of turbidites during that time, the basal erosion hypothesis is preferred here.

4.4. Main patterns of turbidite activity in the different re-entrants

4.4.1. *Poverty re-entrant*

The recent (<17ka) turbidite activity of the Poverty re-entrant is sampled in the two Paritu mid-slope basins. Upslope over the last c. 17 ka, the Paritu Trough records variable Terrigenous Accumulation Rate (TAR), turbidite frequencies and grain size with less than 20-30% of stacked turbidites. A change in recorded sedimentation appears at c. 12 ka with high TAR (95 cm/ka), high turbidite frequency (6.2 turb/ka; turb for turbidites), and a majority of coarse grained turbidites from 17 to 12 ka, and moderate TAR (49cm/ka), moderate turbidite frequency (3.7 turb/ka) and a majority of fine grained turbidites from 12 ka to present day (Figs. 9 and 10, Table 4).

Downslope, during the truncated period sampled c. 17 – 6 ka, the Lower Paritu Basin shows a high and constant TAR (69 cm/ka) with 30% of stacked turbidites. As in the Paritu Trough, a change at c. 12 ka is discernible with moderate turbidite frequency (4.4 turb/ka) and coarse grained turbidites from 17 to 12 ka as opposed to high turbidite frequency (6.7 turb/ka) and fine grained turbidites from c. 12 to 6 ka (Figs. 9 and 10, Table 4).

These results shows two distinct periods of sedimentation in the Poverty re-entrant (Table 4): (1) the 17-12 ka period characterised by coarse grained, silty to sandy turbidites and (2) the 12-0 ka period characterised by fine grained, silt laminae to silty turbidites (Figs. 9 and 10). This change in turbidite grain size is punctuated by a sharp spatial migration of high turbidite frequencies and TAR from the Paritu Trough during the 17-12 ka period to the Lower Paritu Basin through the 12-0 ka one.

4.4.2. *Ruatoria re-entrant Turbidite Sedimentation*

The three areas sampled in the Ruatoria re-entrant show distinct sedimentation patterns, with turbidites fining and increasing in frequency basinward.

(1) The gullied upper slope shows a stable depositional pattern from 7.9 ka to the present day. However, turbidite sedimentation varies laterally depending on geomorphology. The sampled channelized area comprises from north to south (Fig. 3): a northern channel (Tan0810-5) characterised by a stack of debrites and turbidites with the youngest dated at c.0.8 ka, a levee (Tan0810-2) with low TAR (28 cm/ka) and turbidite frequency (2.1 turb/ka), variable turbidite grain size and up to 40% of stacked turbidites, and a southern channel (Tan0810-3) characterised by extremely low TAR (3-18 cm/ka) and turbidite frequency (0.4cm/ka), and fine grained turbidites. These results suggest that turbidite activity is concentrated in the northern channel. In contrast, the isolated plateau northern of the channelized area (Tan0810-1) records no turbidites except one, the Taupo primary monomagmatic turbidite.

(2) Small troughs on the Ruatoria Debris Avalanche (MD06-3009) record strong variations in turbidite sedimentation over the last c.18.5 ka with a sharp boundary at c. 17 ka (Figs. 11 and 12, Table 5). The period 18.5-17 ka is characterised by extremely high TAR (261 cm/ka) and turbidite frequency (9.1 turb/ka), and coarse grained turbidites while the period 17-0 ka shows moderate TAR, low turbidite frequencies and fine grained turbidites. A minor change is supposed at c. 7 ka with slightly lower TAR (56 cm/ka) and turbidite frequency (1.5 turb/ka) from 17 to 7 ka than during the 7-0 period (62cm/ka and 1.9 turb/ka).

(3) The Hikurangi Trough shows temporal variations in turbidite sedimentation over the last c. 17 ka as well as spatial fluctuations depending on geomorphology. At the mouth of the Ruatoria channel (MD06-3008, Fig. 3), a boundary at c. 7 ka separates the 17-7 ka period characterised by high TAR (91 cm/ka) and turbidite frequency (5 turb/ka), variable turbidite grain size with 30% of stacked turbidites, and the 7-0 ka period characterised by moderate TAR (56 cm/ka) and turbidite frequency (2.4 turb/ka), fine grained turbidites with only 5% of stacked turbidites. During the first 17-7 ka period, minor changes in turbidite grain size are recorded at c. 12 ka with coarse grained turbidites from 17 to 12 ka and fine grained turbidites occurring from 12 to 7 ka. In isolated areas not fed by large channels (Tan0810-6), turbidite sedimentation during the period 7-0 ka is different, characterised by low TAR (28 cm/ka), low turbidite frequency (1.4 turb/ka), coarse grained turbidites and 50% of stacked turbidites.

Consequently, the turbidite sequence in the re-entrant can be divided into three periods (Table 5): (1) 18.5-17 ka characterised by high TAR and turbidite frequencies, a majority of stacked turbidites and coarse grained, silty to sandy turbidites; (2) 17-7 ka characterised by a decrease in the overall TAR and turbidite frequencies, fewer stacked turbidites and a progressive fining upcore; and, (3) 7-0 ka characterised by a moderate TAR and low turbidite frequencies, generally high proportions of stacked turbidites except in the Hikurangi Trough and a dominance of fine grained turbidites.

4.4.3. *Matakaoa Turbidite System Sedimentation*

Geographically, we differentiate three regions, from the shelf edge to the deep basin (Figs. 5 and 13, Table 6):

(1) The canyon floor (Tan0314-86) which contains a stack of turbidites. No ages are available to date these deposits, but the absence of hemipelagite drape at the top of the core argues for very recent deposits and a high frequency of flows.

(2) The channel-levee complex which contains c. 90% of hemipelagites from ~5 ka to Present both on levees (cores Tan0810-10, -11 and -13) and into the channel (cores Tan0810-9 and -12). This period shows extremely low TAR (8 cm/ka in the channel and 3 cm/ka on the levees), low turbidite frequency (up to 1.8 turb/ka on levees) and relatively rare stacked turbidites (<25%). Turbidite grain size ranges from sandy to muddy turbidites.

(3) The deep-sea fan (Tan0314-8) which records turbidites in a truncated period starting before 17 ka to 5 ka. A sharp boundary at c. 17 ka is identified. The period pre-17 ka contains only a stack of fine grained turbidites. The lack of datable material below 16,677 cal. yr BP (Fig. 13, Table 3), prevents an estimate of sedimentation rate and turbidite frequencies. The period 17-5 ka is characterised by extremely low TAR (7 cm/ka) and turbidite frequency (0.9 turb/ka), with rare stacked turbidites (10%). Turbidite grain size and occurrence show a period from 12 to 9.5 ka free of turbidites, that separates silty turbidites from 12 to 17 ka from muddy turbidites from 9.5 to 5 ka. A thin hemipelagite drape is recorded younger than 5 ka, suggesting that no turbidite deposition occurred since that age.

These results show three depositional periods in the MTS (Fig. 13, Table 6). (1) The first period, older than 17 ka, is characterised by a deep-sea fan with a stack of turbidites implying a constant growth with continuous activity of the MTS. (2) The period 17-5 ka is characterised by a decrease in turbidite frequency and TAR in the fan, and a progressive fining-up texture. (3) The period 5-0 ka is characterised by hemipelagite sedimentation in the turbidite plain and deep-sea fan, low TAR concentrated in the channel and a moderate turbidite frequency, with thin and fine turbidite layers infilling the head of the canyon.

5. DISCUSSION

The detailed characterisation of the turbidites in this study enables a wider re-examination of the parameters that control turbidite deposition at a c. 18 ka timescale along northeastern New Zealand including: (1) changes in slope morphology, (2) glacio-eustatic sea-level variations and (3) changes in sediment supply. Triggering mechanisms generating turbidites at a centennial timescale are then examined with a particular discussion on the relative contribution of climate (floods) and tectonic (earthquakes).

5.1. Control parameters on turbidite sedimentation

5.1.1. *Local changes in slope morphology*

Large continental slope failures, such as the one that formed the Matakaoa re-entrant, are known to reorganize down-slope sediment pathways controlling basin sediment supply (e.g. Joanne et al., 2010). In the Poverty re-entrant, the successive margin collapses have created two distinct sedimentary systems : the Poverty Canyon System and the Paritu mid-slope basins (Orpin, 2004; Pedley et al., 2010). In the Paritu Trough, the Poverty Debris Avalanche (PDA) has also most likely impacted the sedimentary dynamics of the mid-slope basins. Our data provide evidence of reorganisation of the sedimentary system over the last 17 ka and incidentally allow a first age estimation of the PDA.

Since the emplacement of the PDA in the Paritu Trough, gravity flows descending from the upper-slope gullies are pseudo-channelized along the Paritu Channel, constrained by the relief of the PDA, and flow downslope to the Lower Paritu Basin (Fig. 2). Core data confirm this dynamic with overbank deposits in the Paritu Trough and fan lobes deposits in the Lower Paritu Basin (Table 4). This channelized activity has been constant for the last c.12 ka. Prior to that time, sediment was mainly captured in the Paritu Trough, as demonstrated by the higher TAR and higher frequency of coarse-grained turbidites in the Paritu Trough, compared with that of the Lower Paritu basin (Table 4). This change in the sedimentary routing at c.12 ka is not recorded in the nearby Ruatoria re-entrant, suggesting that it is not controlled by global climatic drivers. We interpret this change to a local reorganization associated with the emplacement of the PDA. Such event would have also eroded the widespread Waiohau tephra (13,635 cal. yr BP) in the Paritu Trough. Based on these observations, we estimate the age of the PDA at 12-13.5 ka. Similar basin evolutions have been interpreted in Miocene trench-slope basins along the Hikurangi Margin (Bailleul et al., 2007) with : (1) the pre-PDA period corresponding to a low-gradient submarine ramp system, when the Paritu Trough is the main collecting basin, (2) the emplacement of the PDA to a large submarine slide, and (3) the post-PDA period to a fine-grained sand-rich submarine fans system, when the two basins are connected.

5.1.2. *Influence of glacio-eustatic sea-level variations*

During the late lowstand – early transgressive period (LLET; 18.5-17 ka), the coastline along the northern Hikurangi margin was c.120 m below present day sea level (Gibb, 1986; Pillans et al., 1998)(Fig.1). The very high sedimentation rates recorded in MD06-3009 (261 cm/ka and 9.1 turb/ka being minimum values since the core is located on a perched basin; Table 5) and the high proportion of rock fragments contained in turbidites are consistent with strong connectivity between rivers and continental slope (Fig. 6D). Although no sedimentation rates could be derived for the Matakaoa re-entrant (Fig. 13), the thick sequence of stacked turbidites indicates sustained high sediment supply from the nearby rivers. Here, the paleo-Waiapu River was very likely connecting to the Matakaoa Turbidite System (MTS) through the Matakaoa Canyon (Joanne et al., 2010) (Figs. 1 and 4) and to the Ruatoria re-entrant through the dense network of upper slope gullies (Fig. 3). Similarly, in the Poverty re-entrant, the Waipaoa River likely directly fed the Poverty Canyon System bypassing the Paritu Trough and Lower Paritu Basin (Lewis et al., 1998; Orpin, 2004).

During the marine transgression (17-7 ka), the coupling between East Coast rivers and slope basins was progressively cut-off. Terrigenous sedimentation rates and turbidite frequencies decreased noticeably in the MTS (7 cm/ka and 0.9 turb/ka) and in the Ruatoria re-entrant (56 cm/ka and 1.5 turb/ka). The proportion of rock fragments also drastically decrease (Fig. 6D). This net decrease in TAR to the slope basins is consistent with the large volume of postglacial sediment trapped on the shelf (Lewis et al., 2004; Orpin 2004, Paquet et al. 2009, Gerber et al, 2010). Further seaward in the Hikurangi Trough, the high level of activity in the channel (MD06-3008) compared to the low activity in the channel-levees overbank (MD06-3009) suggests that the size of gravity flows during the transgression was smaller than during the LLET period and contained within the channel.

During the highstand (7-0 ka), Paquet et al. (2009), Gerber et al. (2010) and Wolinsky et al. (2010) showed that riverine sediments were stored in fluvial valleys onland and on the inner shelf. The constant and uniform rates of accumulation of terrigenous material (TAR) in the Poverty and Ruatoria re-entrant from the time of the Whakatane tephra (cal. 5530 BP), corroborate this observation. But a TAR value of 55 cm/ka in slope basins and in the Hikurangi Trough also suggests that riverine dispersal extends beyond the shelf basin to deliver sediments to the upper slope. This is in slight contrast with the near-full shelf capture scenario proposed by Gerber et al. (2010) for the

Poverty shelf, but consistent with observations in the Ruatoria outer shelf. There, Addington et al. (2007) and Kniskern et al. (2010) showed that part of Holocene river sediments by-pass locally the shelf and reach the upper slope. This is confirmed in this study by sedimentological and morphological evidence showing differential upper slope channelized activity: the active northern channel (Tan0810-5 and -2) connects upslope to incised V-shape gullies where sediment bypassing is observed while the inactive southern channel (Tan0810-3) connects upslope to smooth gullies where shelf depocenters trap shelf sediments. In the MTS, the thin layer of hemipelagite draping the deep-sea fan indicates that no turbidites reach the fan. In the channel-levee complex, low TAR (3 to 8 cm/ka) and low turbidite frequency (1.8 turb/ka), together with the dominance of hemipelagites, indicate that the MTS is mainly inactive.

From these observations we propose that during the LLET period, rivers were closely-coupled to submarine canyons, supplying point source sedimentary systems such as the MTS and the Poverty Canyon System. At this time, linear source systems, such as the Poverty mid-slope basins, may record low activity as they were nourished via advective dispersal and longshore transport. The Ruatoria re-entrant represents a composite system because of its high activity and linear source morphology associated to a short-lived connection with the Waiapu River. As sea level rises, the coupling between riverine supply and submarine canyons wanes with a widening distance between river mouths and canyon heads and the sequestration of the sediment load on the shelf. Point source systems activity sharply decreases while linear source activity increases as rivers sediments are stored on the shelf edge and not flushed out to deep sea through the canyons. During highstand conditions, point source systems are mainly inactive because of the lack of direct sediment input: the head of the Poverty Canyon is partly buried by Late Holocene shelf sediments (Walsh et al., 2007); the Matakaoa Canyon is filled by thin turbidites which episodically reach the channel-levee complex. Linear source systems record stable activity because of the large width of their source area, active slumping at the shelf edge (Lewis et al., 2004), sediment by-passing the shelf (Addington et al., 2007; Alexander et al., 2010; Kniskern et al., 2010) and the presence of numerous gullies, preventing thick sediment wedge accumulation.

5.1.3. *Impact of changes in sediment supply to the slope*

Coarse sandy and silty turbidites deposited older than c. 12 ka contrast with the concentration of fine silty to muddy turbidites during 12-0 ka, with no variations in the TAR (Figs. 10, 12 and 13). The change in turbidite facies at c.12 ka, coeval of the Pleistocene/Holocene climatic boundary, is very likely a consequence of the combined effect of abrupt climatic control on sediment supply and glacio-eustatic fluctuations.

Deposition of coarse turbidites in deep basins is consistent with cold and dry climatic conditions favouring high erosion onland (McGlone, 2001; Okuda et al., 2002; Mildenhall and Orpin, 2010). The subsequent coarse eroded material is transported by Raukumara rivers to a narrow, 5-15 km wide, continental shelf with relatively low trapping efficiency. On the contrary to this, fine turbidites are associated with warm and moist conditions of the Holocene and a wide (20-30 km wide) continental shelf with high trapping efficiency (Paquet et al., 2009; Gerber et al., 2010). Despite the large sediment load provides by river incision (Litchfield and Berryman, 2005), the coarse material is trapped on the wide highstand continental shelf and cannot reach shelf edge and upper slope to generate coarse turbidites.

This change in turbidite facies is well represented and abrupt in deep sea cores MD06-3002, -3003 and -3008 (Figs. 10 and 12). For core MD06-3009 collected on the Ruatoria channel levee 250 m above the main sediment pathway, the fining in turbidite texture occurs at c.17 ka. Since channel levees construction is primarily controlled by flow volumes, their drastic decrease accompanying the disconnection between the Waiapu River mouth and the upper slope at the initiation of the marine transgression (c.17ka) have reduced the overbank deposition efficiency. As a consequence, the climatic impact in that core is less discernible and glacio-eustatic fluctuations may have primarily control the source and delivery of sediments.

5.2. Turbidite generation and triggering mechanisms

Known triggering mechanisms of turbidites generally involve large earthquakes (Goldfinger et al., 2003; St Onge et al., 2004; Blumberg et al., 2008; Noda et al., 2008; Beck, 2009), tsunamis (Shanmugam, 2006), storm waves (Mulder et al., 2001; Puig et al., 2004), volcanism (Schneider et al., 2001) and catastrophic floods (Mulder et al., 2003; St Onge et al., 2004; Beck, 2009). Other possible mechanisms such as sediment overloading or gas hydrate destabilization are assumed to be indirect effects of regional changes like increase of sediment delivery or glacio-eustatic sea-level variations. Because large earthquakes, tsunamis and storm waves are all able to originate turbidity currents on the upper slope, the distinction by using their subsequent deposits is extremely hazardous. However, large storm and tsunami waves affect the seafloor up to a maximum depth of 80-120 mbsl (Mulder et al., 2001; Puig et al., 2004; Shanmugam, 2006). Since the present-day and highstand shelf edge is lying between 150 and 200 mbsl, these processes are unlikely to generate turbidites over the last 7 ka at the difference of earthquakes. Only six large volcanic eruptions are directly associated with primary monomagmatic turbidites in the last c. 18 ka (Figs. 9, 11 and 13), demonstrating that volcanism is a minor process for turbidite generation and will not be discussed fully in the following. Finally, catastrophic floods generating particular turbidites easily distinguishable from others are discussed below.

5.2.1. *Catastrophic floods*

Flood-induced turbidites, recorded as hyperpycnites (Mulder et al., 2003), have been recognized in deep sea basins as far as 700 km away of a river mouth when there is close spatial coupling to deeply incised canyon heads (Nakajima, 2006; Nakajima et al., 2009). In contrast, on the Hikurangi Margin, rivers were connected to canyon heads only during the last sea-level lowstand and disconnected during the present day highstand. Over the last 150 years of catchment deforestation on the Raukumara Peninsula, river discharges attain the threshold for hyperpycnal flows around once a year for the Waiapu River and every 40 years for the Waipaoa River (Hicks et al., 2004). The 1-in-100 year flood caused by Cyclone Bola in 1988 led to thick deposits on the Poverty shelf speculated to have been the result of hyperpyncal flows (Foster and Cater, 1997; Brackley et al., 2010). The 7,200 year sedimentary record of Lake Tutira (Orpin et al., 2010; Page et al., 2010) suggested seven catastrophic storms greater in magnitude than Cyclone Bola (Sinclair, 1993), and twenty-five large storm periods (40-400 year-long) similar to Cyclone Bola representing an average frequency of one storm period every 290 years. Such periods generate intense erosion of the catchment and increase sediment delivery to the shelf. Consequently, there is compelling circumstantial evidence that several hyperpycnal flows could have occurred in the Waipaoa and Waiapu Rivers since 18 ka, even under forest cover.

We have identified a total of nine hyperpycnites since the LGM along the northern Hikurangi Margin. They are dated at 2,930±190, 7,657±137, 11,544±184, 12,863±288, 14,011±347 and 15,681±624 cal. yr BP, in Poverty re-entrant and at 9,266±170, 9,594±218, and 10,882±255 cal. yr BP in Ruatoria re-entrant. None have been identified in the Matakaoa Turbidite System. They represent only ~3% of the total turbidites identified in Poverty and 4% in Ruatoria. Most of them are recognized during the marine transgression (17-7 ka). Another flood event is inferred to occur just after the Taupo eruption when we observe a primary monomagmatic turbidite, locally overlying the Taupo tephra and containing macroscopic wood fragments. This turbidite is interpreted as related to a catastrophic flood washing over the thick volcanic drape deposited all over the North Island.

Only one hyperpycnite is recorded during the present day highstand. Compared to onland climate proxies, this 2,930±190 cal. yr BP hyperpycnite (MD06-3003, Fig. 9) is contemporaneous of a thick heavy rainfall-related bed recorded in Lake Tutira at c. 2,950 cal. yr BP. This bed is the thickest one since 4 ka and the second thickest since the origin of the lake at 7.2 ka, and is part of a suite of seven thick storm beds interpreted as the result of catastrophic rainfall events over 500 mm/day (Orpin et al., 2010). These seven events exceeded in magnitude the 1988 Cyclone Bola (300mm/day; Sinclair, 1993). The deep-sea record of only one of these seven events shows that there is no systematic relationship between hyperpycnite occurrence in deep basins and large onshore flood events. This is in good agreement with the presence of the 25-30 km wide shelf, which reduces connectivity between river mouths and the upper slope. However, the 2,930 cal. yr BP hyperpycnite occurs during the ENSO-dominated climatic regime described by Gomez et al. (2004), which began ~4 ka ago. This period corresponds to an increase in storminess and the transition from fluvial incision to landsliding as the dominant mode of sediment production onland. This particular climatic regime coupled with an exceptional heavy rainfall could possibly have increased the sediment load over the minimum threshold to produce a hyperpycnite in the deep sea.

We believe that the two hyperpycnites dated at 11,544±184 and 10,882±255 cal. yr BP in the Poverty and Ruatoria re-entrants respectively can be related to large floods of the Waipaoa and Waiapu rivers. There is a temporal link between these floods and the warm climatic period (11.6-10.8 ka) defined by Alloway et al. (2007), which occurs just after the Late Glacial Cold Reversal, known as a cooler climatic period with temporary expansion of grassland and shrubland in northern North Island. Following the river incision model developed by Litchfield and Berryman (2005), the Late Glacial Climate Reversal (at 13.5-11.6 ka; Alloway et al., 2007) may have created high erosion and aggradation of fluvial terraces due to low stream power. The warm period increases the stream power which incises the newly formed terrace and consequently generate high river loading and delivery. The coupling with a large storm such as Cyclone Bola or greater, would have generated suitable conditions to deposit hyperpycnite in deep basins, as for the 2,930 cal. yr BP hyperpycnite. There is also a possibility with the age uncertainties that the two hyperpycnites in the Poverty and Ruatoria re-entrants were synchronous and record a single catastrophic event. A better age model is needed to confirm this synchronicity and the occurrence of a large storm at that time.

The timing of other hyperpycnites recognised during the transgression period (7,657±137, 9,266±170, 9,594±218, 12,863±288, 14,011±347 and 15,681±624 cal. yr BP) closely correspond to five stillstand periods (c. 7.5 , 9.5 , 12.5 , 13.7 , 15.7 ka) identified regionally (Carter and Carter, 1986; Carter et al., 2002). Similarly to previously, these hyperpycnites may be the record of extremely large storms and catastrophic floods occurring during stillstands resulting in periods of enhanced sediment flux reaching the deep ocean (Carter et al., 2002).

5.2.2. *Large earthquakes*

Earthquakes have been identified as the dominant triggering mechanisms in numerous active margin settings during the Late Holocene (e.g. Adams, 1990; Goldfinger et al., 2003; Blumberg et al., 2008; Noda et al., 2008). Evidences of prehistoric large earthquakes $M_w > 7$ are derived from the 9 ka record of uplifted or subsided marine terraces (Cochran et al., 2006; Hayward et al., 2006; Wilson et al., 2006; Wilson et al., 2007), which provide an average return time of 670 years (150-1500 years). However, this earthquake record may be incomplete and underestimated since uplift and subsidence episodes are mainly driven by near-shore upper plate fault ruptures (Wilson et al., 2007).

Sedimentological evidences have confirmed that volcanism and catastrophic floods are minor triggering mechanisms of turbidites since 18 ka i.e. a small amount of identified primary monomagmatic turbidites and flood-induced turbidites and low proportion of rock fragments in turbidites (Fig. 6B-D). Most of the turbidites contain material from environments deeper than the shelf break as confirmed by the dominance of deep water foraminiferal assemblages since 18 ka, despite sea-level fluctuations (Fig. 7A and D). This is particularly marked during the highstand (last 7 ka; Fig. 7D) when foraminiferal assemblages show a majority of deep water species (70% of Association 4) and a negligible amount of shelf species (<5% of Association 1). The 20% of shelf and upper slope species (Association 2) are attributable to local conditions in the Matakaoa re-entrant (Fig. 7B). Considering the intense tectonic activity of the Hikurangi margin and according to studies undertaken in similar settings (e.g. Adams, 1990; Blumberg et al., 2008; Goldfinger et al., 2003; Noda et al., 2008), turbidites deposited during the present day highstand, unlikely to be flood or volcanism related, are supposed to be triggered by large earthquakes. Oceanographic processes such as storms, which are able to cannibalize upper slope material (Piper and Normak, 2009), are assumed to be efficient triggering mechanisms only during early marine transgression and lowstand.

The calculated mean return times of turbidites for the last 7 ka in the Poverty (MD06-3003) and Ruatoria re-entrant (MD06-3008) and in the MTS (Tan0810-11) are 270, 410 and 430 years, respectively (Figs. 10 and 12). The return time in the three re-entrants is smaller than estimates of near shore upper plate fault ruptures from onland records (670 years; Cochran et al., 2006; Hayward et al., 2006; Wilson et al., 2006; Wilson et al., 2007). This is consistent with the high seismic activity of the Hikurangi Margin (Reyners, 1998; Reyners and McGinty, 1999; Wallace et al., 2009).

Similar return times in the Ruatoria and Matakaoa re-entrants suggest that both basins share the same tectonic regime with a large earthquake every 420 year in average. The difference with the Poverty re-entrant may reflect a variation in tectonic activity. Poverty re-entrant is located at the boundary of two rupture segments of the subduction interface (Wallace et al., 2009) and is also the area where most of coastal paleo-earthquake evidences are reported. Consequently, the 270 year return time estimated for large earthquakes in the Poverty re-entrant very likely includes the fourteen near-shore faults ruptures identified and dated onland (Cochran et al., 2006; Hayward et al., 2006; Wilson et al., 2006; Wilson et al., 2007), as well as offshore upper plate faults and subduction interplate ruptures.

Core MD06-3009, collected on the Ruatoria Debris Avalanche 250 m above the main sediment pathway, shows the longest turbidite return time (850 years) over the last 7 ka (Fig. 11). Since core MD06-3009 is located on a topographic high, we infer that this return time represents only the large to very large earthquakes record (interplate?), which trigger extremely large turbidity currents able to deposit sediments onto the Debris Avalanche. Similar conclusions have been drawn along the Chile active margin (Blumberg et al., 2008). Furthermore, most turbidites identified in the core are stacked turbidites, which are interpreted elsewhere to be associated with very large subduction earthquakes (Goldfinger et al., 2003; Nakajima and Kanai, 2000).

6. Conclusion

This study presents a detailed history of turbidite sedimentation captured in a series of cores collected from the Poverty, Ruatoria and Matakaoa re-entrants that indent the active northern Hikurangi Margin, eastern New Zealand. Sedimentological analyses combined with strong chronological control afforded by numerous radiocarbon dates and tephra identifications, enabled us to identify and characterise more than a thousand turbidites in the late Quaternary basin sequence.

The last postglacial sequence is overwhelmingly terrigenous, and composed of alternating cm-thick turbidites and hemipelagites, with sparse tephra layers and extremely rare mass transport deposits. Colour is the key parameter in the distinction between hemipelagite and turbidites. The composition of the silt fraction determines colour: hemipelagites mostly contain volcaniclastic grains, usually pumiceous lapilli, while turbidite tails are mainly quartz grains. Turbidite sand is predominantly composed of quartz and volcaniclastic grains indicative of a remobilisation of material supplied by the adjacent muddy rivers. Benthic foraminifers within the turbidites suggest an upper slope origin. Five facies of turbidites are recognised : muddy turbidites, silt laminae turbidites, silty turbidites, sandy turbidites and basal reverse-graded turbidites, which include flood-induced hyperpycnites.

Turbidites are deposited continuously throughout the c. 18 ky period captured in the cores. Glacio-eustatic variations strongly control turbidite accumulation. During the late lowstand – early transgressive period, closely-coupled fluvial sources directly fed submarine canyon heads for the

Poverty Canyon and Matakaoa Turbidite Systems, bypassing intraslope basins of the Paritu and Ruatoria re-entrants. In contrast, during highstand conditions, continuous and stable turbidite generation occurs within the intraslope basins but point source fed systems are inactive. Also, in deep portions of the margin, the Holocene/Pleistocene climatic boundary is imprinted as a sharp sedimentary boundary at c.12 ka, separating coarse silty-sandy turbidites (18-12 ka) from fine muddy-silty turbidites (12-0 ka) with no impact on the overall accumulation rate. Seabed morphology affects turbidites emplacement and frequency. A change in turbidite sedimentation provides new evidence of the emplacement of the Poverty Debris Avalanche in the Paritu Trough at 13.6-12 ka.

Catastrophic floods are recognized as a rare triggering mechanism for turbidite generation, with only nine hyperpycnites recognized since 18 ka representing 3 to 4 % of the total turbidites. The most recent hyperpycnites, dated at c. 2,930±190 cal. yr BP is contemporaneous with a storm-bed from lacustrine records. Despite the high annual flood frequency for the muddy rivers draining the Raukumara ranges, hyperpycnites might only be generated during the most extreme climatic events.

Because of the deep seated source of sediment located well beyond the shelf edge, large earthquakes are the most plausible triggering mechanism for the turbidite sequence described from the northern Hikurangi Margin over the last 7 ka. During that period, the average return time of turbidites is shorter than the coastal records of large earthquakes, since coastal records reflect only proximal near-shore fault ruptures and the preservation potential is reduced due to terrestrial erosion. The Matakaoa and Ruatoria re-entrants suggest similar return times, implying a similar tectonic regime, with an average return time for large earthquakes of 420 years. The Poverty re-entrant shows many more turbidite-triggered earthquakes, with a mean return time of 270 years. One core, specifically collected on a topographic high, contains evidence of very large earthquakes with an average return time of 850 years.

The 7 ka storm record of Lake Tutira indicates that storm periods associated to large river discharges show an average recurrence of 290 years. This recurrence time is close to the return time of turbidites caused by earthquake in Poverty re-entrant pointing to a possible link between periods of sediment flushing by earthquakes at the shelf edge and periods of recharge by storms and floods. These kind of interrelationships between climate and tectonic triggering on turbidite deposition still need to be explored and will certainly require very comprehensive datasets.

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Figure captions

Figure 1 - The Raukumara Peninsula and northern Hikurangi Margin, NE New Zealand along the Pacific-Australia subduction front (teeth line). Onland, the Waipaoa, Uawa and Waiapu river catchments are highlighted in grey. Contour interval is 100 m. The 120 m contour (blue dashed line) provides an approximate position of the last glacial shoreline. Frames indicate location of subsequent figures. Historical earthquakes M>6 (yellow stars) are from the Geonet database (geonet.co.nz), including the M_W7.8 1931 Napier earthquake (orange star). East Auckland Current (EAC), Wairarapa Coastal Current (WCC), East Cape Current (ECC) and Deep Water Bottom Current (DWBC) are from Chiswell (2000), McCave and Carter (1997), Stanton (1998), and Stanton et al. (1997). The main active fault earthquake sources identified so far, either normal, transverse or reverse, are reported onland and offshore after Stirling et al. (in press). Their average slip rate (in mm/a) is reported when > 1mm/a. **Insert** shows the New Zealand region with the PAC-AUS plate boundary, the Hikurangi Trough (Hik T), the Kermadec Trench (K T), the back-arc Havre Trough (Hav T) and the Central Volcanic Region (CVR) from which all tephra identified in the cores originate. The relative plate motion of 50 mm/a at the PAC-AUS boundary is from de Mets et al. (1994). Black arrows in the CVR indicate the average extension rate of 6-8 mm/a from Villamor and Berryman (2001). Black half arrows indicate the dextral strike slip of < 1 mm/a from Lamarche et al. (2006).

Figure 2 - The Poverty re-entrant seafloor morphology from EM300 multibeam echo-sounder data. Red dots indicate the location of sediment cores used in this study. Contour lines every 100 m; the 120 m isobaths (blue dashed line) indicates the approximate shoreline during the last lowstand. Recent landslides and debris avalanches (grey

shade) are from Pedley et al. (2010). Arrows show pathways of the main gravity-flows supplying the basin. Location on Fig. 1.

Figure 3 - The Ruatoria re-entrant seafloor morphology. See Fig. 2 for full caption. Location on Fig. 1.

Figure 4 - The Matakaoa re-entrant seafloor morphology. See Fig. 2 for full caption. Location on Fig. 1.

Figure 5 - Characterisation of the four lithofacies and the five turbidite facies identified in cores from sediment color, internal structure from X-radiograph and grain-size (mean or median (D50) and distribution). Red arrows show grain size trend (normal and reverse grading). The two zooms in MD06-3008 0-50 cm and MD06-3009 610-680 cm show the detailed grain size trend differenciating stacked turbidites and silt laminae turbidites : no decantation phase (turbidite tail in pink) is observed between grain-size peaks in silt laminae turbidites conversely to stacked turbidites which present decantation after each pulse, characterized by sorting and skewness index. Horizontal black arrows in MD06-3008 0-50 cm indicate silt laminae.

Figure 6 - Sediment composition. *n* is the number of samples. The error bars are 2σ error bar. a) characteristic composition of hemipelagite silty-clays compared to turbidite silty-clays. Analysis were undertaken on 21 samples from the Ruatoria basin. b-d show turbidite sand grain composition from individual sedimentary systems (b, core locations in Figs. 1 to 4); classified by turbidite facies (c); and during the last lowstand, marine transgression and actual highstand. *c* is the number of cores in which samples have been taken.

Figure 7 - Foraminiferal assemblages from turbidite sand-size material. a) Identified species and corresponding living water depth for the four associations. Foraminiferal assemblages and percentage of planktic foraminifers grouped (b) by cores and sedimentary systems; (c) by turbidite facies ; and (d) in the last lowstand, marine transgression and actual highstand periods. *c* is the number of cores in which samples have been taken.

Figure 8 - Age model generated from dates obtained on samples collected along the core (see text) *vs*. depth for the six longest cores. Total sediment depths provide uncorrected sedimentation rate (a), whereas hemipelagite thickness provides corrected

sedimentation rate (b). The TAR - Terrigeneous Accumulation Rate - (c) is provided by the cumulated turbidite thickness. (d) The two components of the uncorrected sedimentation rate (bold lines), namely the corrected (lines) and the TAR (dashed lines) are plotted for two long cores to illustrate that fluctuations in the TAR control variations in the uncorrected sedimentation rate, as the corrected sedimentation rate remains roughly constant since 18 ka. Short cores with less than three ages are not plotted. Circle: ^{14}C age; square: Tephra. Insert shows sedimentation rate for 25, 50, 100 and 200 cm/ka for comparison. When available, tephra ages are preferred to ^{14}C age with a specific calibration (MD06-3002; see table 3).

Figure 9 - Sedimentological logs of the cores collected in the Poverty re-entrant. Thick dotted lines are time correlations between cores made from tephra identification with their age in bold; ^{14}C ages from foraminifers are in italic; thin dotted line at 11.6 ka is the Holocene-Pleistocene boundary. Legend in Fig. 11.

Figure 10 - Turbidite records in Poverty re-entrant from MD06-3002 and MD06-3003 since ~18 ka compared to climate boundaries and sea-level fluctuations. From top to base : (1) Relative sea level (after Gibbs, 1986; Pillans et al., 1998), (2) Lithofacies distribution expressed as cumulative turbidite lithofacies I to IV (Fig. 6) from a 9-turbidite rolling mean (key and color from Fig. 12), (3) turbidite return time calculated as the time difference between 2 consecutive events, (4) Thickness of isolated and stacked turbidite layers.

Figure 11 - Sedimentological logs of the cores collected over the Ruatoria re-entrant. Thick dotted lines are time correlations between cores made from tephra identification with their age in bold; ^{14}C ages from foraminifers are in italic; thin dotted line at 11.6 ka is the Holocene-Pleistocene boundary. Important note: Vertical scale for Tan cores indicated on the left is double that of MD cores indicated on right.

Figure 12 - Turbidite records in Ruatoria re-entrant for the last 18 ka. Full caption in Fig. 11.

Figure 13 - Sedimentological logs from the cores collected over the Matakaoa re-entrant. Thick dotted lines are time correlations between cores made from tephra identification with their age in bold; ^{14}C ages from foraminifers are in italic; thin dotted line at 11.6 ka is the Holocene-Pleistocene boundary. Legend in Fig. 11.

Tables

Table 1 – Location and main analysis results for MD06 long cores and Tan0810 short cores. T: gravity flow deposits (turbidites); H: hemipelagites; ST: Stacked turbidites; IT: Isolated Turbidites. *: full recovered length; when core deformation is too high, the used core length is given between brackets; ** total number of turbidite layers identified in the core.

Table 2 – Tephra stratigraphic position, uncorrected depth, corrected depth, identification, and calibrated ages after Lowe et al. (2008).

Table 3 – Radiocarbon ^{14}C ages from mixed planktonic foraminifers. A reservoir age of 395 years is used expect for * where reservoir age is 800 years.

Table 4 – Poverty re-entrant turbidite sedimentation

Table 5 - Ruatoria re-entrant turbidite sedimentation

Table 6 - Matakaoa re-entrant turbidite sedimentation