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Characterisation, timing and origin of turbidites**

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

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10 **ABSTRACT**

11 Recent sedimentation along the Hikurangi subduction margin off northeastern New Zealand is  
12 investigated using a series of piston cores collected between 2003 and 2008. The active Hikurangi  
13 Margin lies along the Pacific-Australia subduction plate boundary and contains a diverse range of  
14 geomorphologic settings. Slope basin stratigraphy is thick and complex, resulting from sustained high  
15 rates of sedimentation from adjacent muddy rivers throughout the Quaternary. Turbidites deposited  
16 since c. 18 ka in the Poverty, Ruatoria and Matakaoa re-entrants are central to this study in that they  
17 provide a detailed record of the past climatic conditions and tectonic activity. Here, alternating  
18 hemipelagite, turbidite, debrite and tephra layers reflect distinctive depositional modes of marine  
19 sedimentation, turbidity current, debris flow and volcanic eruption, respectively. Turbidites dominate  
20 the record, ranging in lithofacies from muddy to sandy turbidites, and include some basal-reverse  
21 graded turbidites inferred to be derived from hyperpycnal flows. Stacked turbidites are common and  
22 indicate multiple gravity-flows over short time periods. The chronology of turbidites is determined by  
23 collating an extremely dense set of radiocarbon ages and dated tephra, which facilitate  
24 sedimentation rate calculation and identification of the origin of turbidites. Sedimentation rates  
25 range from 285 cm/ka during late glacial time (18.5-17 ka) to 15 to 109 cm/ka during postglacial time  
26 (17-0 ka). Turbidite deposition is controlled by: (1) the emplacement of slope avalanches reorganizing  
27 sediment pathways; (2) the postglacial marine transgression leading to a five-fold reduction in  
28 sediment supply to the slope due to disconnection of river mouths from the shelf edge, and (3) the  
29 Holocene/Pleistocene boundary climate warming resulting in a drastic decrease in the average  
30 turbidite grain-size. Flood-induced turbidites are scarce: nine hyperpycnites are recognized since 18

31 ka and the youngest is correlated to the largest ENSO-related storm event recorded onland (Lake  
32 Tutira). Other turbidites contain a benthic foraminiferal assemblage which is strictly reworked from  
33 the upper slope and which relate to large earthquakes over the last c. 7 ka. They yield a shorter  
34 return time (270-430 years) than the published coastal records for large earthquakes (c.670 years),  
35 but the offshore record is likely to be more complete. The deep-sea sedimentation along the New  
36 Zealand active margin illustrates the complex interaction of tectonic and climate in turbidite  
37 generation. Climate warming and glacio-eustatic fluctuations are well recorded at a millennial  
38 timescale (18 ka), while tectonic deformation and earthquakes appear predominant in fostering  
39 turbidite production at a centennial timescale (270-430 years).

40

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

## 43 **1. INTRODUCTION**

44 Gravity-driven flows are ubiquitous and fundamental process that control sediment dispersal where  
45 steep bathymetric gradients, enhanced tectonic activity and voluminous terrigenous sediment supply  
46 prevail such as at active margins. They range from submarine avalanches, cohesive debris or grain  
47 flows, liquefied and fluidized flows and turbidity currents (Stow and Mayall, 2000; Stow et al., 1996).  
48 Such processes can generate complex sets of sedimentary structures from a variety of triggering  
49 mechanisms and scales including giant avalanches consisting of  $>100 \text{ km}^3$  of lithified sediment (e.g.  
50 Collot et al., 2001; Canals et al., 2004), thick successions of density-variable turbidites (Bouma, 1962;  
51 Stow and Shanmugam, 1980; Lowe, 1982), to centimetre-thick hyperpycnites that can be linked to  
52 individual flood events (Mulder et al., 2003). As such, gravity flow deposits contain invaluable  
53 information about past stratigraphic, climatic and tectonic history (Adams, 1990; Goldfinger et al.,  
54 2003; St Onge et al., 2004; Blumberg et al., 2008; Noda et al., 2008; Nakajima et al., 2009). However,  
55 due to the geomorphologic complexity of active margins lateral correlation of events is often  
56 problematic, both in terms of dealing with the spatial variability of gravity events and recognising  
57 synchronous “event assemblages”.

58 The active Hikurangi convergent margin, New Zealand is an excellent locality for the study of gravity-  
59 driven events because of the diversity of geomorphological settings, the intense tectonic activity  
60 (e.g., Lewis and Pettinga, 1993; Collot et al., 1996) and the high rates of sedimentation that produce  
61 an expanded stratigraphic record at an exceptional resolution. As the Hikurangi Margin lies along the  
62 Pacific-Australia subduction plate boundary, it is subjected to intense seismic activity. Here, a well

63 documented upper-plate earthquake record exists for magnitude  $M_w < 7.8$  (Reyners, 1998; Webb  
64 and Anderson, 1998) but only a poorly documented record of inferred plate interface ruptures  
65 capable of generating  $\sim M_w 8.8$  earthquakes (Reyners, 1998; Reyners and McGinty, 1999; Wallace et  
66 al., 2009; Cochran et al., 2006). At the northern extent of margin, intense mass wasting and margin-  
67 collapses activity is manifested as large morphological re-entrants in the continental slope (Collot et  
68 al., 2001; Lamarche et al., 2008a; Pedley et al., 2010). Due to the vigorous maritime climate, floods  
69 are a common feature of northeastern New Zealand (Hicks et al., 2004). Some prehistoric  
70 catastrophic floods have been inferred from river flood-plains and continental shelf sediments  
71 (Brown, 1995; Gomez et al., 2007; Brackley et al., 2010) which might be capable of rapidly  
72 transporting sediment directly from the coast to slope basins via hyperpycnal flows. The occurrence  
73 of numerous tephra originating from the Central Volcanic Zone (Fig. 1) provide excellent  
74 chronological control in the offshore stratigraphic record (e.g. Carter et al., 2002). The northern  
75 Hikurangi Margin was intensely studied over the last 20 years, and contributed to a robust  
76 understanding of long- and short-time scale tectonic deformation (Collot et al., 1996; Reyners, 1998;  
77 Reyners and McGinty, 1999), sedimentary processes and stratigraphy (Foster and Carter, 1997;  
78 Joanne et al., 2010; Orpin, 2004, Gomez et al., 2007; Paquet et al., 2009; Kniskern et al., 2010) and  
79 Holocene sediment budgets (Orpin et al., 2006; Alexander et al., 2010; Gerber et al., 2010; Paquet et  
80 al., 2011). But the thick and complex suite of Quaternary turbidites that infill the slope basins remain  
81 largely understudied and their event stratigraphy underutilised.

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

## 94 **2. GEOLOGICAL AND SEDIMENTOLOGICAL SETTINGS**

## 95 **2.1. Geomorphology**

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

105 The northern Hikurangi Margin includes a flat, 20-30 km-wide continental shelf, a steep sediment-  
106 starved slope, and a 3500 m-deep subduction trough (Fig. 1). Tectonic erosion has produced three  
107 large slope avalanches: the 30-50 km-wide, Poverty re-entrant (Pedley et al., 2010); the 30-40 km  
108 Ruatoria re-entrant (Collot et al., 2001); and landward of the trench wall and immediately north of  
109 the Raukumara Peninsula, the Matakaoa passive margin contains the 50 km-wide Matakaoa re-  
110 entrant (Lamarche et al., 2008a). Elsewhere smaller debris slides, slumps and head-wall scarps are  
111 abundant, indicating ongoing slope instability (e.g. Lewis et al., 1998). The current study focuses on  
112 sediment cores within the Poverty, Ruatoria, and Matakaoa re-entrants. The 1500 km<sup>2</sup> Poverty re-  
113 entrant is a major continental margin depression resulting from successive margin collapses since  
114 1,500±500 ka (Pedley et al., 2010). The bathymetry of the Poverty re-entrant is complex and  
115 comprises several basic morphologic components (Orpin, 2004) including: a heavily gullied upper  
116 slope; the beheaded Poverty Canyon System; the gently sloping mid-slope Paritu Trough; margin-  
117 parallel North and South Paritu Ridges that are cross-cut by a small canyon feeding into the Lower  
118 Paritu Basin (Fig. 2). The Paritu Trough is filled with the Poverty Debris Avalanche (PDA), which is  
119 blanketed by sediments. Although the PDA is undated, the surface of the avalanche is rough and  
120 hummocky suggesting a recent event. The 3300 km<sup>2</sup> Ruatoria re-entrant formed following a giant  
121 debris avalanche 170±40 ka ago (Collot et al., 2001). The re-entrant consists of a gullied upper slope,  
122 a vast highly chaotic debris avalanche composed of individual blocks of several cubic kilometres in  
123 size, and the subduction trough (Fig. 3). The 1000 km<sup>2</sup> Matakaoa re-entrant resulted from multiple  
124 mass transports events, which occurred between 1,300 and 35 ka ago (Carter, 2001; Lamarche et al.,  
125 2008a; Joanne et al., 2010) (Fig. 4). The eastern half of the re-entrant is infilled by the Matakaoa  
126 Turbidite System (MTS), which developed subsequently to the Matakaoa Debris Avalanche, 600±150  
127 ka ago (Joanne et al., 2010). The MTS is a classical channelized turbidite system with a canyon

128 incising into the shelf break, a well-developed channel/levee turbidite plain and a fan growing in the  
129 Raukumara Plain.

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

## 137 **2.2. Sedimentology**

138 Up to a kilometre of Quaternary sediment fill accumulates in ponded basins along the northern  
139 Hikurangi Margin continental shelf (Lewis et al., 2004) and slope (Orpin, 2004; Orpin et al., 2006;  
140 Paquet et al., 2009) as well as in the Hikurangi Trough (Lewis and Pettinga, 1993) and Raukumara  
141 basin (Kohn and Glasby, 1978) (Fig. 1). The mass accumulation rate along the margin is generally high  
142 over the last 1 My (4 Mt/a in Hawkes Bay), with millennial variations over glacio-eustatic cycles  
143 (Carter and Manighetti, 2006; Paquet et al., 2009). Over the last 30 ka, the highest rates were  
144 recorded during last-glacial lowstand through to the early highstand stage (30-7 ka). The Holocene  
145 highstand period (7-0 ka) shows a declining flux to the lower continental slope as more sediment is  
146 retained in subsiding shelf basins (Gerber et al., 2010) and baffled in intra-slope basins bounded by  
147 imbricate thrust ridges (Lewis et al., 1998; Paquet et al., 2011), where the hemipelagic flux is around  
148 60 cm/ka since the mid-late Holocene (Orpin, 2004).

149 Driven by the vigorous maritime climate across the Raukumara Ranges, the present day sediment  
150 flux delivered to the adjacent shelf and slope basins is 70 Mt/a. Forest clearing by early Polynesian  
151 settlers 500-700 y BP and then by European colonisation in the mid-eighteenth century resulted in  
152 present day river sediment fluxes an order of magnitude greater than pre-human colonization (e.g.  
153 McGlone et al., 1994; McGlone and Wilmshurst, 1999). Paquet et al. (2009) estimated an increase of  
154 110-250% in Hawkes Bay whereas Kettner et al. (2007) calculated a rise of 660% for the Waipaoa  
155 river alone (Fig. 1).

156 Regional oceanography plays a major role in the offshore dispersal of sediments from Raukumara  
157 rivers (Fig. 1). On the continental shelf, swell waves, wind direction, the northward-flowing  
158 Wairarapa Coastal Current (WCC) and large ephemeral gyres affect current direction (Foster and  
159 Carter, 1997; Chiswell, 2000). Beyond the shelf break, the southward-flowing East Cape Current (ECC)

160 is the dominant current affecting the region during the Holocene (Stanton, 1998; Stanton et al.,  
161 1997; Carter et al., 2002). During the Last Glacial Maximum (LGM), the ECC strength decreased while  
162 the proto-WCC, flowing northward near the shelf break, increased (Carter and Manighetti, 2006).  
163 Deep circulation in the Hikurangi Trough is influenced by the Southwest Pacific Deep Western  
164 Boundary Current (DWBC). The main flow of the DWBC is confined by the northeast scarp of the  
165 Hikurangi Plateau, but a shallower westward-flowing branch reaches the Hikurangi Trough at Poverty  
166 Bay where it deviates northward and joins with the main DWBC over the Kermadec trench (McCave  
167 and Carter, 1997).

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

### 181 **3. DATA AND METHODS**

#### 182 **3.1. Collection of sediment cores**

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

189 High-resolution 3.5 kHz seismic reflection data and multibeam bathymetry were systematically  
190 acquired prior to coring in order to ascertain the suitability of the sampling sites, providing sub-

191 surface stratigraphic information up to 20 m below the seafloor with a vertical resolution of <1 m.  
192 The bathymetry is compiled from data acquired during the Geodynz survey using the 12 kHz EM12  
193 echo-sounder of R/V L'Atalante (Collot et al., 1996) and a large number of surveys using the 30 kHz  
194 Kongsberg EM300 echo-sounder of R/V Tangaroa with an optimal accuracy of ~0.2 % of the water  
195 depth. The margin morphology is provided by Digital Terrain Models (DTM) generated from the  
196 multibeam bathymetry database maintained at NIWA (CANZ, 2008).

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

### 206 **3.2. Sedimentological analyses**

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

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

222 500µm), to determine the source of the sediments deduced from the distribution of modern benthic  
223 foraminifers in New Zealand (Hayward et al., 2010; Camp, 2009).

### 224 **3.3. Age Dating**

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

233 Radiocarbon dating was performed on handpicked mixed planktonic foraminifers at the Rafter  
234 Radiocarbon Laboratory, GNS Science. The 0.7-1.0 cm-thick samples were collected in hemipelagite  
235 layers, 0.7-1.0 cm below gravity-flow deposits to prevent any contamination and mixing by  
236 bioturbation. AMS <sup>14</sup>C were calibrated to calendar years by using the MARINE09.14 calibration curve  
237 (Reimer et al., 2009) in CALIB Rev 6.0 program (Stuiver and Reimer, 1993), applying an average  
238 regional reservoir age of 395±57 years calculated from published East Cape reservoir age (Higham  
239 and Hogg, 1995; Kalish, 1993; Calib database at <http://calib.qub.ac.uk/marine/>). A reservoir age of  
240 800±110 years has been applied for the time of the Waiohau tephra deposition (13,635 cal. yr BP;  
241 Table 3), as defined by Sikes et al. (2000) and Carter et al. (2008). The <sup>14</sup>C radiochronology calibration  
242 is adequate for most of the samples from the Marion Dufresne cores as shown by the good  
243 correlation with tephrochronology. Two samples on MD06-3002 located less than 10 cm above the  
244 Waiohau Tephra suggest a stratigraphic reversal. Reservoir age modification during that period may  
245 explain the inconsistency between AMS <sup>14</sup>C and tephra ages. For this study, we prefer the tephra age  
246 and discarded the two AMS <sup>14</sup>C samples.

### 247 **3.4. Sedimentation Rates**

248 Based on lithofacies identification, we distinguish uncorrected and corrected sedimentation rates.  
249 Uncorrected sedimentation rates includes the total sediment thickness from all lithofacies, whereas  
250 corrected sedimentation rate includes only the hemipelagite. Corrected sedimentation rate is  
251 calculated by subtracting the thickness of the turbidites and tephra layers from the total sediment  
252 thickness, and assumes limited erosion at the base of the turbidite layers. Corrected rate is used here

253 to estimate the age of turbidites. Hemipelagites represent a continuous and steady mode of  
254 deposition whereas gravity-driven depositional events are emplaced instantaneously. The  
255 Terrigenous Accumulation Rate (TAR) is the difference between uncorrected and corrected  
256 sedimentation rates, representing the cumulated thickness of gravity-driven deposits (mostly  
257 turbidites) through time. Because of the high density of dated samples in each core, deformation in  
258 the piston core does not significantly influence our results and interpretations. Deformation is  
259 localised and easily identified in the age model by a change in the slope of the age curve.

## 260 **4. RESULTS**

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

### 265 **4.1. End-members facies**

#### 266 4.1.1. *Tephra*

267 All cores contained several tephra composed of 1 to 2 cm-thick (rarely up to 6 cm-thick), pinkish,  
268 normally graded silts, capped by a clay-rich bioturbated horizon (Fig. 5). They are exclusively  
269 composed of volcanoclastic debris (glass shards and pumiceous lapilli mostly) and identified by their  
270 typical colour and high values of magnetic susceptibility (>40 SI). In places, tephra are thoroughly  
271 reworked by intense bioturbation, which suggests that the original depositional layer was less than 1  
272 cm-thick, preventing asphyxia of the benthic fauna (Hess and Kuhnt, 1996). In this case, corrected  
273 sedimentation rates are calculated using a 1 cm thickness for highly bioturbated tephra. In this study,  
274 we assume that all tephra originate from ash-fall coincident with volcanic eruptions (Wiesner et al.,  
275 1995; Carter et al., 1995).

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

282 4.1.2. *Debrites*

283 This end-member facies consist of < 35 cm thick chaotic intervals of dark olive-grey silty-clay with  
284 sand, granules, pebbles and occasionally deformed stratified lithoclasts (Fig. 5). The sand to granule  
285 size material shows weak reverse grading. Debrites are composed of quartz, volcanoclastic clasts,  
286 bivalve and gastropod shells and 2-3 cm-large clasts of poorly laminated silty clays and laminated  
287 fine- to medium-clayey silts. This facies is rare, representing only five events in two cores (Tan0810-5  
288 and MD06-3003). The chaotic facies, the absence of well-defined basal erosion and the matrix  
289 supported texture, suggest a mass transport deposit from a debris flow (Mulder and Alexander,  
290 2001).

291 4.1.3. *Hemipelagite*

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

298 4.1.4. *Turbidites*

299 Together with hemipelagites, turbidites dominate the sedimentary record. There are a maximum of  
300 101, 89 and 20 single turbidites per core in the Poverty, Ruatoria and Matakaoa re-entrants,  
301 respectively. Turbidites are recognized by their coarser grain size and a typical fining upward trend  
302 (Fig. 5). Thickness ranges from 1–75 cm. Turbidites are usually interbedded with hemipelagites. The  
303 basal boundaries are easily identified from a change to coarser-grain size, darker color and increase  
304 in density, magnetic susceptibility and P-wave velocity. The top boundary is progressive with  
305 bioturbated contact from the turbidite tail, grading into the hemipelagite background. Compositional  
306 analysis shows a doubling in quartz grain concentrations in turbidite tails (> 50%) and slightly higher  
307 values of rock fragments and micas than in hemipelagite (Fig. 6A). All turbidites in this study are  
308 interpreted as deposited by low to medium density turbidity current as defined by Stow and  
309 Shanmugam (1980) and Bouma (1962).

310 Several turbidites can be stacked in sequences over up to 75 cm thick bounded by hemipelagites.  
311 These are termed herein “stacked turbidites”, as opposed to “isolated turbidites”, which consist of a  
312 single gravity-flow deposit under- and overlain by hemipelagite (Fig. 5). The small thickness of

313 individual turbidites, the lack of thick coarse grain basal unit (<20cm) and the thick uppermost silty-  
 314 clay unit in stacked turbidites suggests very low erosion at the base of individual gravity events.  
 315 Hence we infer that the lack of intervening hemipelagite in stacked turbidites is due to non-  
 316 deposition rather than erosion. These conditions suggest only a short duration of time between  
 317 successive turbidites.

## 318 **4.2. Turbidite composition and facies**

### 319 4.2.1. Sand Composition

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

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

338

### 339 4.2.2. Foraminiferal assemblages

340 We identified 28 benthic foraminifera species, of which *Uvigerina peregrina*, *Bulimina marginata f.*  
 341 *aculeata*, *Evolvocassidulina orientalis*, *Notorotalia depressa*, *Bolinita quadrilatera*, *Globobulimina*  
 342 *pacifica* and *Quinqueloculina auberina* largely dominate. These species are indicative of a variety of

343 environments from the inner shelf to the abyssal plain. However, most of them are characteristic of  
 344 environments seaward of the shelf break (>150m).

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

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

#### 357 4.2.3. *Turbidites facies*

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

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

374 *Silt laminae turbidites (T II)* consist of irregular, <40 cm-thick sequences of interbedded, thinning and  
375 fining upward clay and silt laminae (Fig. 5). Silt laminae are usually <1 cm-thick and stacked in sets of  
376 2 to 10. Weak cross-stratification in the basal silt laminae merge up section to planar lamination,  
377 followed by poorly laminated, then homogeneous silty-clay. Sand grains are predominantly light  
378 minerals (59%) and volcanoclastic grains (30%). Foraminiferal assemblages show a majority of planktic  
379 species (80%) and a benthic assemblage dominated by Association 4 (81%) with rare species from  
380 Associations 2 (11%) and 3 (8%) (Fig. 6C). Silt laminae turbidites are interpreted as fine-grained  
381 turbidites deposited by a low density turbidity current (T2 to T8 subdivisions of Stow and  
382 Shanmugam (1980)). They differ from stacked turbidites by their specific grain-size showing a single  
383 sequence (Fig. 5).

384 *Silty turbidites (T III)* are composed of 0.5–55 cm-thick, fining upward clayey silt sequences (Fig. 5).  
385 The basal contact is usually sharp, with little evidence of erosion. A complete graded sequence shows  
386 from base to top: (1) a massive coarse clayey silt base; (2) laminated coarse clayey silt; (3) cross-  
387 stratified coarse clayey silt; (4) laminated fine to medium clayey silt; and in places, (5) a laminated silt  
388 and clay and (6) poorly laminated to homogeneous silty-clay top. Sand grains within silty turbidites  
389 are composed of volcanoclastic grains and light minerals (33% each) and a relatively high proportion  
390 of foraminifers (9%). In core MD06-3008, some mica-rich silty turbidites (>90% of mica) artificially  
391 increase the average value of the class “other detritic grains” (Fig. 6C). Silty turbidites are also  
392 characterised by 69% of planktic foraminifers and benthic assemblage is composed of Association 4  
393 (73%) with a minor component from other associations (from 6 to 12% each). These clayey silt  
394 sequences are interpreted as the upper subdivisions Tc to Te of medium density turbidites (Bouma,  
395 1962) deposited by low density turbidity currents.

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

407 6C). These sequences are interpreted as the Ta to Te subdivisions of medium density turbidites  
 408 (Bouma, 1962).

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

#### 423 **4.3. Age controls and sedimentation rates**

424 Ages recovered from core material are compiled in Table 2 and 3 and summarised in Fig. 8. In the  
 425 Poverty re-entrant core MD06-3003 offers a continuous chronology from c.1 ka to c.16.5 ka, but in  
 426 MD06-3002 a truncated range from c.6 to c.17 ka was recovered. In the Ruatoria re-entrant MD06-  
 427 3009 presents the longest record from c.1 ka to c.18 ka, whereas ages in core MD06-3008 range from  
 428 c. 0.5 ka to c.16.5 ka. All short cores in the Ruatoria and Matakaoa re-entrants exhibited dates  
 429 spanning a shorter period from c.8 ka to present day except for core Tan0314-8, which shows a  
 430 truncated range from c.5 to c.17 ka.

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

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

#### 448 **4.4. Main patterns of turbidite activity in the different re-entrants**

##### 449 4.4.1. Poverty re-entrant

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

457 Downslope, during the truncated period sampled c. 17 – 6 ka, the Lower Paritu Basin shows a high  
 458 and constant TAR (69 cm/ka) with 30% of stacked turbidites. As in the Paritu Trough, a change at c.  
 459 12 ka is discernible with moderate turbidite frequency (4.4 turb/ka) and coarse grained turbidites  
 460 from 17 to 12 ka as opposed to high turbidite frequency (6.7 turb/ka) and fine grained turbidites  
 461 from c. 12 to 6 ka (Figs. 9 and 10, Table 4).

462 These results shows two distinct periods of sedimentation in the Poverty re-entrant (Table 4): (1) the  
 463 17-12 ka period characterised by coarse grained, silty to sandy turbidites and (2) the 12-0 ka period  
 464 characterised by fine grained, silt laminae to silty turbidites (Figs. 9 and 10). This change in turbidite  
 465 grain size is punctuated by a sharp spatial migration of high turbidite frequencies and TAR from the  
 466 Paritu Trough during the 17-12 ka period to the Lower Paritu Basin through the 12-0 ka one.

##### 467 4.4.2. Ruatoria re-entrant Turbidite Sedimentation

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

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

480 (2) Small troughs on the Ruatoria Debris Avalanche (MD06-3009) record strong variations in turbidite  
481 sedimentation over the last c.18.5 ka with a sharp boundary at c. 17 ka (Figs. 11 and 12, Table 5). The  
482 period 18.5-17 ka is characterised by extremely high TAR (261 cm/ka) and turbidite frequency (9.1  
483 turb/ka), and coarse grained turbidites while the period 17-0 ka shows moderate TAR, low turbidite  
484 frequencies and fine grained turbidites. A minor change is supposed at c. 7 ka with slightly lower TAR  
485 (56 cm/ka) and turbidite frequency (1.5 turb/ka) from 17 to 7 ka than during the 7-0 period (62cm/ka  
486 and 1.9 turb/ka).

487 (3) The Hikurangi Trough shows temporal variations in turbidite sedimentation over the last c. 17 ka  
488 as well as spatial fluctuations depending on geomorphology. At the mouth of the Ruatoria channel  
489 (MD06-3008, Fig. 3), a boundary at c. 7 ka separates the 17-7 ka period characterised by high TAR (91  
490 cm/ka) and turbidite frequency (5 turb/ka), variable turbidite grain size with 30% of stacked  
491 turbidites, and the 7-0 ka period characterised by moderate TAR (56 cm/ka) and turbidite frequency  
492 (2.4 turb/ka), fine grained turbidites with only 5% of stacked turbidites. During the first 17-7 ka  
493 period, minor changes in turbidite grain size are recorded at c. 12 ka with coarse grained turbidites  
494 from 17 to 12 ka and fine grained turbidites occurring from 12 to 7 ka. In isolated areas not fed by  
495 large channels (Tan0810-6), turbidite sedimentation during the period 7-0 ka is different,  
496 characterised by low TAR (28 cm/ka), low turbidite frequency (1.4 turb/ka), coarse grained turbidites  
497 and 50% of stacked turbidites.

498 Consequently, the turbidite sequence in the re-entrant can be divided into three periods (Table 5):  
499 (1) 18.5-17 ka characterised by high TAR and turbidite frequencies, a majority of stacked turbidites

500 and coarse grained, silty to sandy turbidites; (2) 17-7 ka characterised by a decrease in the overall  
 501 TAR and turbidite frequencies, fewer stacked turbidites and a progressive fining upcore; and, (3) 7-0  
 502 ka characterised by a moderate TAR and low turbidite frequencies, generally high proportions of  
 503 stacked turbidites except in the Hikurangi Trough and a dominance of fine grained turbidites.

#### 504 4.4.3. *Matakaoa Turbidite System Sedimentation*

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

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

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

515 (3) The deep-sea fan (Tan0314-8) which records turbidites in a truncated period starting before 17 ka  
 516 to 5 ka. A sharp boundary at c. 17 ka is identified. The period pre-17 ka contains only a stack of fine  
 517 grained turbidites. The lack of datable material below 16,677 cal. yr BP (Fig. 13, Table 3), prevents an  
 518 estimate of sedimentation rate and turbidite frequencies. The period 17-5 ka is characterised by  
 519 extremely low TAR (7 cm/ka) and turbidite frequency (0.9 turb/ka), with rare stacked turbidites  
 520 (10%). Turbidite grain size and occurrence show a period from 12 to 9.5 ka free of turbidites, that  
 521 separates silty turbidites from 12 to 17 ka from muddy turbidites from 9.5 to 5 ka. A thin  
 522 hemipelagite drape is recorded younger than 5 ka, suggesting that no turbidite deposition occurred  
 523 since that age.

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

## 531 **5. DISCUSSION**

532

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

### 539 **5.1. Control parameters on turbidite sedimentation**

#### 540 *5.1.1. Local changes in slope morphology*

541 Large continental slope failures, such as the one that formed the Matakaoa re-entrant, are known to  
542 reorganize down-slope sediment pathways controlling basin sediment supply (e.g. Joanne et al.,  
543 2010). In the Poverty re-entrant, the successive margin collapses have created two distinct  
544 sedimentary systems : the Poverty Canyon System and the Paritu mid-slope basins (Orpin, 2004;  
545 Pedley et al., 2010). In the Paritu Trough, the Poverty Debris Avalanche (PDA) has also most likely  
546 impacted the sedimentary dynamics of the mid-slope basins. Our data provide evidence of  
547 reorganisation of the sedimentary system over the last 17 ka and incidentally allow a first age  
548 estimation of the PDA.

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

562 period corresponding to a low-gradient submarine ramp system, when the Paritu Trough is the main  
563 collecting basin, (2) the emplacement of the PDA to a large submarine slide, and (3) the post-PDA  
564 period to a fine-grained sand-rich submarine fans system, when the two basins are connected.

#### 565 5.1.2. *Influence of glacio-eustatic sea-level variations*

566 During the late lowstand – early transgressive period (LLET; 18.5-17 ka), the coastline along the  
567 northern Hikurangi margin was c.120 m below present day sea level (Gibb, 1986; Pillans et al.,  
568 1998)(Fig.1). The very high sedimentation rates recorded in MD06-3009 (261 cm/ka and 9.1 turb/ka  
569 being minimum values since the core is located on a perched basin; Table 5) and the high proportion  
570 of rock fragments contained in turbidites are consistent with strong connectivity between rivers and  
571 continental slope (Fig. 6D). Although no sedimentation rates could be derived for the Matakaoa re-  
572 entrant (Fig. 13), the thick sequence of stacked turbidites indicates sustained high sediment supply  
573 from the nearby rivers. Here, the paleo-Waiapu River was very likely connecting to the Matakaoa  
574 Turbidite System (MTS) through the Matakaoa Canyon (Joanne et al., 2010) (Figs. 1 and 4) and to the  
575 Ruatoria re-entrant through the dense network of upper slope gullies (Fig. 3). Similarly, in the  
576 Poverty re-entrant, the Waipaoa River likely directly fed the Poverty Canyon System bypassing the  
577 Paritu Trough and Lower Paritu Basin (Lewis et al., 1998; Orpin, 2004).

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

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

594 Poverty shelf, but consistent with observations in the Ruatoria outer shelf. There, Addington et al.  
595 (2007) and Kniskern et al. (2010) showed that part of Holocene river sediments by-pass locally the  
596 shelf and reach the upper slope. This is confirmed in this study by sedimentological and  
597 morphological evidence showing differential upper slope channelized activity: the active northern  
598 channel (Tan0810-5 and -2) connects upslope to incised V-shape gullies where sediment bypassing is  
599 observed while the inactive southern channel (Tan0810-3) connects upslope to smooth gullies where  
600 shelf depocenters trap shelf sediments. In the MTS, the thin layer of hemipelagite draping the deep-  
601 sea fan indicates that no turbidites reach the fan. In the channel-levee complex, low TAR (3 to 8  
602 cm/ka) and low turbidite frequency (1.8 turb/ka), together with the dominance of hemipelagites,  
603 indicate that the MTS is mainly inactive.

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

### 621 5.1.3. *Impact of changes in sediment supply to the slope*

622 Coarse sandy and silty turbidites deposited older than c. 12 ka contrast with the concentration of fine  
623 silty to muddy turbidites during 12-0 ka, with no variations in the TAR (Figs. 10, 12 and 13). The  
624 change in turbidite facies at c.12 ka, coeval of the Pleistocene/Holocene climatic boundary, is very

625 likely a consequence of the combined effect of abrupt climatic control on sediment supply and  
626 glacio-eustatic fluctuations.

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

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

## 644 **5.2. Turbidite generation and triggering mechanisms**

645 Known triggering mechanisms of turbidites generally involve large earthquakes ( Goldfinger et al.,  
646 2003; St Onge et al., 2004; Blumberg et al., 2008; Noda et al., 2008; Beck, 2009), tsunamis  
647 (Shanmugam, 2006), storm waves (Mulder et al., 2001; Puig et al., 2004), volcanism (Schneider et al.,  
648 2001) and catastrophic floods (Mulder et al., 2003; St Onge et al., 2004; Beck, 2009). Other possible  
649 mechanisms such as sediment overloading or gas hydrate destabilization are assumed to be indirect  
650 effects of regional changes like increase of sediment delivery or glacio-eustatic sea-level variations.  
651 Because large earthquakes, tsunamis and storm waves are all able to originate turbidity currents on  
652 the upper slope, the distinction by using their subsequent deposits is extremely hazardous. However,  
653 large storm and tsunami waves affect the seafloor up to a maximum depth of 80-120 mbsl (Mulder et  
654 al., 2001; Puig et al., 2004; Shanmugam, 2006). Since the present-day and highstand shelf edge is  
655 lying between 150 and 200 mbsl, these processes are unlikely to generate turbidites over the last 7  
656 ka at the difference of earthquakes. Only six large volcanic eruptions are directly associated with

657 primary monomagmatic turbidites in the last c. 18 ka (Figs. 9, 11 and 13), demonstrating that  
658 volcanism is a minor process for turbidite generation and will not be discussed fully in the following.  
659 Finally, catastrophic floods generating particular turbidites easily distinguishable from others are  
660 discussed below.

#### 661 5.2.1. *Catastrophic floods*

662 Flood-induced turbidites, recorded as hyperpycnites (Mulder et al., 2003), have been recognized in  
663 deep sea basins as far as 700 km away of a river mouth when there is close spatial coupling to deeply  
664 incised canyon heads (Nakajima, 2006; Nakajima et al., 2009). In contrast, on the Hikurangi Margin,  
665 rivers were connected to canyon heads only during the last sea-level lowstand and disconnected  
666 during the present day highstand. Over the last 150 years of catchment deforestation on the  
667 Raukumara Peninsula, river discharges attain the threshold for hyperpycnal flows around once a year  
668 for the Waiapu River and every 40 years for the Waipaoa River (Hicks et al., 2004). The 1-in-100 year  
669 flood caused by Cyclone Bola in 1988 led to thick deposits on the Poverty shelf speculated to have  
670 been the result of hyperpycnal flows (Foster and Cater, 1997; Brackley et al., 2010). The 7,200 year  
671 sedimentary record of Lake Tutira (Orpin et al., 2010; Page et al., 2010) suggested seven catastrophic  
672 storms greater in magnitude than Cyclone Bola (Sinclair, 1993), and twenty-five large storm periods  
673 (40-400 year-long) similar to Cyclone Bola representing an average frequency of one storm period  
674 every 290 years. Such periods generate intense erosion of the catchment and increase sediment  
675 delivery to the shelf. Consequently, there is compelling circumstantial evidence that several  
676 hyperpycnal flows could have occurred in the Waipaoa and Waiapu Rivers since 18 ka, even under  
677 forest cover.

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

687 Only one hyperpycnite is recorded during the present day highstand. Compared to onland climate  
688 proxies, this  $2,930\pm 190$  cal. yr BP hyperpycnite (MD06-3003, Fig. 9) is contemporaneous of a thick

689 heavy rainfall-related bed recorded in Lake Tutira at c. 2,950 cal. yr BP. This bed is the thickest one  
690 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  
691 thick storm beds interpreted as the result of catastrophic rainfall events over 500 mm/day (Orpin et  
692 al., 2010). These seven events exceeded in magnitude the 1988 Cyclone Bola (300mm/day; Sinclair,  
693 1993). The deep-sea record of only one of these seven events shows that there is no systematic  
694 relationship between hyperpycnite occurrence in deep basins and large onshore flood events. This is  
695 in good agreement with the presence of the 25-30 km wide shelf, which reduces connectivity  
696 between river mouths and the upper slope. However, the 2,930 cal. yr BP hyperpycnite occurs during  
697 the ENSO-dominated climatic regime described by Gomez et al. (2004), which began ~4 ka ago. This  
698 period corresponds to an increase in storminess and the transition from fluvial incision to landsliding  
699 as the dominant mode of sediment production onland. This particular climatic regime coupled with  
700 an exceptional heavy rainfall could possibly have increased the sediment load over the minimum  
701 threshold to produce a hyperpycnite in the deep sea.

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

716 The timing of other hyperpycnites recognised during the transgression period ( $7,657 \pm 137$ ,  
717  $9,266 \pm 170$ ,  $9,594 \pm 218$ ,  $12,863 \pm 288$ ,  $14,011 \pm 347$  and  $15,681 \pm 624$  cal. yr BP) closely correspond to  
718 five stillstand periods (c. 7.5 , 9.5 , 12.5 , 13.7 , 15.7 ka) identified regionally (Carter and Carter, 1986;  
719 Carter et al., 2002). Similarly to previously, these hyperpycnites may be the record of extremely large  
720 storms and catastrophic floods occurring during stillstands resulting in periods of enhanced sediment  
721 flux reaching the deep ocean (Carter et al., 2002).

722 5.2.2. *Large earthquakes*

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

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

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

751 Similar return times in the Ruatoria and Matakaoa re-entrants suggest that both basins share the  
752 same tectonic regime with a large earthquake every 420 year in average. The difference with the  
753 Poverty re-entrant may reflect a variation in tectonic activity. Poverty re-entrant is located at the

754 boundary of two rupture segments of the subduction interface (Wallace et al., 2009) and is also the  
755 area where most of coastal paleo-earthquake evidences are reported. Consequently, the 270 year  
756 return time estimated for large earthquakes in the Poverty re-entrant very likely includes the  
757 fourteen near-shore faults ruptures identified and dated onland (Cochran et al., 2006; Hayward et al.,  
758 2006; Wilson et al., 2006; Wilson et al., 2007), as well as offshore upper plate faults and subduction  
759 interplate ruptures.

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

## 768 **6. CONCLUSION**

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

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

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

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

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

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

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

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1121

## 1122 **Figure captions**

1123 **Figure 1** - The Raukumara Peninsula and northern Hikurangi Margin, NE New Zealand along the  
 1124 Pacific-Australia subduction front (teeth line). Onland, the Waipaoa, Uawa and Waipau  
 1125 river catchments are highlighted in grey. Contour interval is 100 m. The 120 m contour  
 1126 (blue dashed line) provides an approximate position of the last glacial shoreline.  
 1127 Frames indicate location of subsequent figures. Historical earthquakes  $M > 6$  (yellow  
 1128 stars) are from the Geonet database (geonet.co.nz), including the  $M_w 7.8$  1931 Napier  
 1129 earthquake (orange star). East Auckland Current (EAC), Wairarapa Coastal Current  
 1130 (WCC), East Cape Current (ECC) and Deep Water Bottom Current (DWBC) are from  
 1131 Chiswell (2000), McCave and Carter (1997), Stanton (1998), and Stanton et al. (1997).  
 1132 The main active fault earthquake sources identified so far, either normal, transverse or  
 1133 reverse, are reported onland and offshore after Stirling et al. (in press). Their average  
 1134 slip rate (in mm/a) is reported when  $> 1$  mm/a. **Insert** shows the New Zealand region  
 1135 with the PAC-AUS plate boundary, the Hikurangi Trough (Hik T), the Kermadec Trench  
 1136 (K T), the back-arc Havre Trough (Hav T) and the Central Volcanic Region (CVR) from  
 1137 which all tephra identified in the cores originate. The relative plate motion of 50 mm/a  
 1138 at the PAC-AUS boundary is from de Mets et al. (1994). Black arrows in the CVR  
 1139 indicate the average extension rate of 6-8 mm/a from Villamor and Berryman (2001).  
 1140 Black half arrows indicate the dextral strike slip of  $< 1$  mm/a from Lamarche et al.  
 1141 (2006).

1142 **Figure 2** - The Poverty re-entrant seafloor morphology from EM300 multibeam echo-sounder  
 1143 data. Red dots indicate the location of sediment cores used in this study. Contour lines  
 1144 every 100 m; the 120 m isobaths (blue dashed line) indicates the approximate  
 1145 shoreline during the last lowstand. Recent landslides and debris avalanches (grey

1146 shade) are from Pedley et al. (2010). Arrows show pathways of the main gravity-flows  
 1147 supplying the basin. Location on Fig. 1.

1148 **Figure 3** - The Ruatoria re-entrant seafloor morphology. See Fig. 2 for full caption. Location on  
 1149 Fig. 1.

1150 **Figure 4** - The Matakaoa re-entrant seafloor morphology. See Fig. 2 for full caption. Location on  
 1151 Fig. 1.

1152 **Figure 5** - Characterisation of the four lithofacies and the five turbidite facies identified in cores  
 1153 from sediment color, internal structure from X-radiograph and grain-size (mean or  
 1154 median (D50) and distribution). Red arrows show grain size trend (normal and reverse  
 1155 grading). The two zooms in MD06-3008 0-50 cm and MD06-3009 610-680 cm show the  
 1156 detailed grain size trend differentiating stacked turbidites and silt laminae turbidites :  
 1157 no decantation phase (turbidite tail in pink) is observed between grain-size peaks in silt  
 1158 laminae turbidites conversely to stacked turbidites which present decantation after  
 1159 each pulse, characterized by sorting and skewness index. Horizontal black arrows in  
 1160 MD06-3008 0-50 cm indicate silt laminae.

1161 **Figure 6** - Sediment composition.  $n$  is the number of samples. The error bars are  $2\sigma$  error bar. a)  
 1162 characteristic composition of hemipelagite silty-clays compared to turbidite silty-clays.  
 1163 Analysis were undertaken on 21 samples from the Ruatoria basin. b-d show turbidite  
 1164 sand grain composition from individual sedimentary systems (b, core locations in Figs.  
 1165 1 to 4); classified by turbidite facies (c); and during the last lowstand, marine  
 1166 transgression and actual highstand.  $c$  is the number of cores in which samples have  
 1167 been taken.

1168 **Figure 7** - Foraminiferal assemblages from turbidite sand-size material. a) Identified species and  
 1169 corresponding living water depth for the four associations. Foraminiferal assemblages  
 1170 and percentage of planktic foraminifers grouped (b) by cores and sedimentary  
 1171 systems; (c) by turbidite facies ; and (d) in the last lowstand, marine transgression and  
 1172 actual highstand periods.  $c$  is the number of cores in which samples have been taken.

1173 **Figure 8** - Age model generated from dates obtained on samples collected along the core (see text)  
 1174 vs. depth for the six longest cores. Total sediment depths provide uncorrected  
 1175 sedimentation rate (a), whereas hemipelagite thickness provides corrected

1176 sedimentation rate (b). The TAR - Terrigenous Accumulation Rate - (c) is provided by  
 1177 the cumulated turbidite thickness. (d) The two components of the uncorrected  
 1178 sedimentation rate (bold lines), namely the corrected (lines) and the TAR (dashed  
 1179 lines) are plotted for two long cores to illustrate that fluctuations in the TAR control  
 1180 variations in the uncorrected sedimentation rate, as the corrected sedimentation rate  
 1181 remains roughly constant since 18 ka. Short cores with less than three ages are not  
 1182 plotted. Circle:  $^{14}\text{C}$  age; square: Tephra. Insert shows sedimentation rate for 25, 50, 100  
 1183 and 200 cm/ka for comparison. When available, tephra ages are preferred to  $^{14}\text{C}$  age  
 1184 with a specific calibration (MD06-3002; see table 3).

1185 **Figure 9** - Sedimentological logs of the cores collected in the Poverty re-entrant. Thick dotted lines  
 1186 are time correlations between cores made from tephra identification with their age in  
 1187 bold;  $^{14}\text{C}$  ages from foraminifers are in italic; thin dotted line at 11.6 ka is the  
 1188 Holocene-Pleistocene boundary. Legend in Fig. 11.

1189 **Figure 10** - Turbidite records in Poverty re-entrant from MD06-3002 and MD06-3003 since ~18 ka  
 1190 compared to climate boundaries and sea-level fluctuations. From top to base : (1)  
 1191 Relative sea level (after Gibbs, 1986; Pillans et al., 1998), (2) Lithofacies distribution  
 1192 expressed as cumulative turbidite lithofacies I to IV (Fig. 6) from a 9-turbidite rolling  
 1193 mean (key and color from Fig. 12), (3) turbidite return time calculated as the time  
 1194 difference between 2 consecutive events, (4) Thickness of isolated and stacked  
 1195 turbidite layers.

1196 **Figure 11** - Sedimentological logs of the cores collected over the Ruatoria re-entrant. Thick dotted  
 1197 lines are time correlations between cores made from tephra identification with their  
 1198 age in bold;  $^{14}\text{C}$  ages from foraminifers are in italic; thin dotted line at 11.6 ka is the  
 1199 Holocene-Pleistocene boundary. Important note: Vertical scale for Tan cores indicated  
 1200 on the left is double that of MD cores indicated on right.

1201 **Figure 12** - Turbidite records in Ruatoria re-entrant for the last 18 ka. Full caption in Fig. 11.

1202 **Figure 13** - Sedimentological logs from the cores collected over the Matakaoa re-entrant. Thick  
 1203 dotted lines are time correlations between cores made from tephra identification with  
 1204 their age in bold;  $^{14}\text{C}$  ages from foraminifers are in italic; thin dotted line at 11.6 ka is  
 1205 the Holocene-Pleistocene boundary. Legend in Fig. 11.

1206 **Tables**

1207 **Table 1** – Location and main analysis results for MD06 long cores and Tan0810 short cores. T: gravity  
1208 flow deposits (turbidites); H: hemipelagites; ST: Stacked turbidites; IT: Isolated Turbidites. \*:  
1209 full recovered length; when core deformation is too high, the used core length is given  
1210 between brackets; \*\* total number of turbidite layers identified in the core.

1211 **Table 2** – Tephra stratigraphic position, uncorrected depth, corrected depth, identification, and  
1212 calibrated ages after Lowe et al. (2008).

1213 **Table 3** – Radiocarbon  $^{14}\text{C}$  ages from mixed planktonic foraminifers. A reservoir age of 395 years is  
1214 used expect for \* where reservoir age is 800 years.

1215 **Table 4** – Poverty re-entrant turbidite sedimentation

1216 **Table 5** - Ruatoria re-entrant turbidite sedimentation

1217 **Table 6** - Matakaoa re-entrant turbidite sedimentation