Postglacial (after 18 ka) deep-sea sedimentation along the Hikurangi subduction margin (New Zealand): 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 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 ~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 Matakoa re-entrants. The 1500 km^2 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^2 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^2 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 ephemerical 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±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 analyzes of

213   selected samples using a Beckman-Coulter LS 13 320 Lasersizer (size range of 0.38-2000µm). 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 (>53µm) 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).

3.3.   Age Dating

225 Timing and age downcore are provided using tephrochronology and ^{14}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 ^{14}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 ^{14}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 ^{14}C and tephra ages. For this study, we prefer the tephra age

246 and discarded the two AMS ^{14}C samples.

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 volcaniclastic 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 volcaniclastic 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, volcaniclastic 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 volcaniclastic 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 turbidte 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 volcaniclastic grains and angular to

321   rounded light mineral grains of quartz with rare feldspar (Fig. 6B). Volcaniclastic 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   volcaniclastic grains. In the Matakaoa re-entrant, composition is dominated by volcaniclastic 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 μm) (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 volcaniclastic
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 volcaniclastic 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 volcaniclastic 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 volcaniclastic 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 hyperpyncal 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±190, 7,657±137, 11,544±184, 12,863±288, 14,011±347 and 15,681±624 cal.

680   yr BP, in Poverty re-entrant and at 9,266±170, 9,594±218, and 10,882±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±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±184 and 10,882±255 cal. yr BP in the Poverty

703   and Ruatoria re-entrants respectively can be related to large floods of the Waipaoa and Waiapu

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±137,

717   9,266±170, 9,594±218, 12,863±288, 14,011±347 and 15,681±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 volcaniclastic grains, usually
778 pumiceous lapilli, while turbidite tails are mainly quartz grains. Turbidite sand is predominantly
779 composed of quartz and volcaniclastic 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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1120 107.

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 Waiapu

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_W7.8 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 > 1mm/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 differenciating 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 2σ 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 - Terrigeneous 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}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}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}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}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}C ages from foraminifers are in italic; thin dotted line at 11.6 ka is

1205   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