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1 **SUBMARINE PALEOSEISMOLOGY OF THE NORTHERN HIKURANGI SUBDUCTION MARGIN OF NEW**
2 **ZEALAND AS DEDUCED FROM TURBIDITE RECORD SINCE 16KA**

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10
11 **KEYWORDS**

12 active margin; Poverty Bay; paleoearthquake; turbidite paleoseismology; synchronous slope failures;
13 earthquake hazard assessment; peak ground acceleration

14 **ABSTRACT**

15 Paleoseismic studies seek to characterise the signature of pre-historical earthquakes by deriving
16 quantitative information from the geological record such as the source, magnitude and recurrence of
17 moderate to large earthquakes. In this study, we provide a ~16,000 yr-long paleo-earthquake record
18 of the 200 km-long northern Hikurangi Margin, New Zealand, using cm-thick deep-sea turbidites
19 identified in sediment cores. Cores were collected in strategic locations across the margin within
20 three distinct morphological re-entrants – the Poverty, Ruatoria and Matakaoa re-entrants. The
21 turbidite facies vary from muddy to sandy with evidence for rare hyperpycnites interbedded with
22 hemipelagites and tephra. We use the Oxal probabilistic software to model the age of each turbidite,
23 using the sedimentation rate of hemipelagite deduced from well-dated tephra layers and
24 radiocarbon ages measurements on planktonic foraminifera.

25 Turbidites are correlated from one core to the other using similarity in sedimentary facies,
26 petrophysical properties and ages. Results show that 46 turbidites are synchronous along the entire
27 margin. Amongst them 41 are interpreted as originating from the upper continental slope in
28 response to earthquake-triggered slope failures between 390 ± 170 to $16,450 \pm 310$ yr BP. Using well-
29 established empirical relationships that combine peak ground acceleration, magnitude and location
30 of earthquakes, we calculate that synchronous slope failures were triggered by the rupture of 3 of
31 the 26 known active faults in the region, each capable of generating M_w 7.3 to 8.4 earthquakes – two
32 are crustal reverse faults and one is the subduction interface. The 41 $M_w \geq 7.3$ earthquakes occurred
33 at an average recurrence interval of ~400 yr over the last ~16,000 yr. Among them, twenty are
34 interpreted as subduction interface earthquakes that occurred at a median recurrence interval of
35 ~800 yr, with alternating periods of high activity and low return times (305 – 610 yr) and quiescence
36 periods with high return times (1480 – 2650 yr). Based on turbidite paleoseismology, we propose
37 that subduction interface earthquakes were of lower magnitude during active periods ($M_w > 7.5$)
38 than during quiescence periods ($M_w \geq 8.2$).

39 **1. INTRODUCTION**

40 Submarine paleoseismology provides the means to derive reliable information on the spatial
41 distribution and recurrence of prehistoric earthquakes in the marine environment from the

42 sedimentary record. The method has been successfully applied at active margins (Adams, 1990;
43 Goldfinger et al., 2003; 2007; Huh et al., 2004; Noda et al., 2008; Gracia et al., 2010; McHugh et al.,
44 2011), in lakes (Strasser et al., 2006; Beck, 2009), and in intraplate domains (St-Onge et al., 2004).
45 However, identification of earthquakes as the triggering mechanism of turbidites often remains
46 equivocal despite careful sedimentological characterization (e.g. Gorsline et al., 2000; Nakajima and
47 Kanai, 2000). Convincing interpretations have been obtained by demonstrating a synchronicity of
48 trigger over long distances and across distinct sedimentary systems (Goldfinger et al., 2003; 2007;
49 Gracia et al., 2010). In some cases, correlations with historical and instrumental records of
50 earthquakes substantiate the interpretations, but in regions where written history does not exceed
51 200 years like in New Zealand, specific methodologies need to be developed to ascertain the
52 earthquake origin of turbidites (Pouderoux et al., 2012a; 2012b; Barnes et al., 2013).

53 The northern Hikurangi subduction margin, east of New Zealand's North Island contains well-
54 developed series of Quaternary turbidites (Lewis, 1973, 1980; Lewis et al., 2004; Pouderoux et al.,
55 2012a). It is characterised by high sediment delivery (Hicks and Shankar, 2003; Hicks et al., 2004), and
56 intense tectonic activity (Reyners and McGinty, 1999; Doser and Webb, 2003), which makes it an
57 excellent location for marine turbidite paleoseismology studies. Three large morphological re-
58 entrants in the margin, the Poverty, Ruatoria and Matakaoa re-entrants (Fig. 1), concentrate gravity
59 flow sedimentation (Lewis et al., 2004; Joanne et al., 2010; Pedley et al., 2010) and record a
60 succession of turbidites emplaced since the Last Glacial Maximum (Pouderoux et al., 2012a). In
61 Poverty re-entrant, Pouderoux et al. (2012b) demonstrated that most of the turbidites were
62 triggered by paleo-earthquakes. However, their study focused on two cores recovered from a single
63 sedimentary system and lacks the margin-scale embrace which may afford the characterisation of
64 earthquakes of great magnitudes ($M_w > 8$) such as those generated along a subduction interface.

65 The present paper aims at establishing a calendar record of earthquakes that occurred along the
66 northern Hikurangi Margin for the last ~16,000 yr, at identifying the fault sources and estimating the
67 magnitude of earthquakes at the origin of the turbidites. To do so, we use sedimentological and
68 geomorphological observations, chronostratigraphic correlations and peak ground acceleration
69 attenuation models.

70 **2. BACKGROUND AND GEOLOGICAL SETTINGS**

71 ***2.1. Morphological and tectonic settings***

72 The northern Hikurangi Margin (Fig. 1) comprises from east to west the 3500 m-deep Hikurangi
73 Trough, a narrow sediment-starved accretionary prism, an unstable continental slope and a
74 continental shelf supplied in sediments by the rivers of the Raukumara Peninsula (Lewis, 1980; Lewis
75 and Pettinga, 1993; Collot et al., 1996). The margin extends northwards into the Kermadec
76 subduction margin. Inland, west of the peninsula lies the rhyolitic Central Volcanic Region, which is
77 the main source of geochemically distinct ash layers (tephra) that punctuate the terrestrial and
78 subaqueous Quaternary sedimentary record of the North Island (Lowe et al., 2008). Three
79 morphological re-entrants scar the continental slope (Fig. 1): (1) the 1500 km² Poverty re-entrant
80 which formed after successive margin collapses since 1500 ±500 ka (Pedley et al., 2010); (2) the 3300
81 km² Ruatoria re-entrant and associated Ruatoria Debris Avalanche formed 170 ±40 ka ago (Collot et
82 al., 2001); and (3) the 1000 km² Matakaoa re-entrant located ~100 km landward of the subduction
83 margin and resulting from multiple mass transport events deposited between 1300 and 35 ka
84 (Lamarche et al., 2008a; Joanne et al., 2010). The three re-entrants (thereafter named Poverty,
85 Ruatoria or Matakaoa for simplicity) include gullies and canyons on the upper slope, mid-slope basins
86 and trench basins, and represent independent Quaternary sedimentary systems.

87 The subduction of the Pacific Plate beneath the Australian Plate, at a present rate of 5 cm/yr
88 (DeMets et al., 1994; Beavan et al., 2002) is resulting in uplifting of the inland axial range of the

89 Raukumara Peninsula at an estimated maximum rate of 3 mm/yr (Reyners and McGinty, 1999;
90 Wilson et al., 2007), and intense seismic activity along the margin (e.g. Webb and Anderson, 1998;
91 Reyners and McGinty, 1999). Quantitative estimate of the interseismic coupling (Wallace et al., 2004;
92 2009), seismological studies (Reyners, 1993; 1998), tectonic investigations (Nicol and Wallace, 2009)
93 and geomorphological characteristics (Collot et al., 1996) all contribute to the interpretation that the
94 600 km-long Hikurangi Margin divides into three subduction interface segments, namely from south
95 to north, the Wairarapa, Hawke's Bay and Raukumara segments. The change from accretion to
96 erosion-dominated margin, North of Gisborne marks the transition from Hawke's Bay to Raukumara
97 segments, where the study area is located. Empirical modelling suggests that rupture of the
98 Raukumara or Hawke's Bay segment would result in M_w 8.2 – 8.4 earthquakes, whereas a
99 simultaneous rupture of both segments would produce a M_w 8.6 earthquake, and that rupture of the
100 entire Hikurangi Margin would result in a M_w 8.8 earthquake (Wallace et al., 2009; Stirling et al.,
101 2012). Northwards, the Kermadec Margin may trigger $M_w \geq 8.5$ earthquakes that can affect the study
102 area (Power et al., 2011). Seismic reflection surveys on the continental shelf have helped to identify a
103 series of upper plate active faults (Barnes et al., 2002; Barnes and Nicol, 2004; Lewis et al., 2004; Fig.
104 1) and empirical relationships suggest some of these faults are capable of generating $M_w \geq 6.5$
105 earthquakes (Stirling et al., 2012; see also SM1).

106 The instrumental record of earthquakes in New Zealand since 1940 includes 298 earthquakes of $M_w >$
107 5 in the study area, with only six of $M_w \geq 6.5$ (<http://www.geonet.co.nz>, as in December 2011). Two
108 subduction interface earthquakes of M_w 6.9 – 7.1 were recorded in 1947 in the Poverty Bay area
109 (Downes et al., 2000; Doser and Webb, 2003), and the 1931 M_w 7.8 Napier upper plate earthquake is
110 the largest and most damaging historic earthquake recorded in the area (Downes, 1995). Prehistoric
111 $M_w \geq 7$ earthquakes over the last 9 kyr are evidenced by uplifted marine terraces (Berryman, 1993;
112 Wilson et al., 2006; 2007), subsided swamps (Cochran et al., 2006) and tsunami deposits (Goff and
113 Dominey-Howes, 2009). Marine terraces uplifts at Pakarae river mouth and Mahia Peninsula are
114 thought to be the result of near-shore fault ruptures (Wilson et al., 2007; Litchfield et al., 2010),
115 namely the Gable End and Lachlan 3 faults (Fig. 1), that might have ruptured coevally. Sudden
116 subsidence episodes at the origin of swamps flooding in Hawke's Bay were interpreted as the result
117 of large offshore earthquakes either from ruptures of the Lachlan 3 fault or the subduction interface
118 (Fig 1; Cochran et al., 2006). In Poverty, a 18 kyr paleo-earthquake record based on turbidite
119 deposition reveal a mean return time of ~ 230 years for moderate to large earthquakes and a 90%
120 probability of occurrence ranging from 10 to 570 years (Pouderoux et al., 2012b). Probabilistic
121 seismic hazard assessment suggests that earthquakes in the region with an associated Peak Ground
122 Acceleration (PGA) of 0.3-0.5 g have a 500-year return time in the coastal area and that earthquakes
123 with a PGA of 0.8-1.4 g have a 2500 year average return time (Stirling et al., 2012).

124 **2.2. Sedimentation patterns and turbidites deposition**

125 The ubiquitous gravity flow sedimentation of the northern Hikurangi Margin ranges from fine
126 turbidites deposited on mid-slope basins to margin-scale debris avalanches (Lewis, 1973; Collot et al.,
127 2001; Lamarche et al., 2008a; Mountjoy and Micallef, 2011; Pouderoux et al., 2012a). Quaternary
128 sedimentation is characterised by interlayering of turbidites and hemipelagites infilling slope and
129 trench basins (Lewis, 1980; Lewis and Pettinga, 1993; Lamarche et al., 2008b; Paquet et al., 2011).
130 Over the last ~ 18 kyr, accumulation rates in mid-slope basins and in the Hikurangi Trough range from
131 ~ 15 to ~ 110 cm/ka (Orpin, 2004; Orpin et al., 2006; Pouderoux et al., 2012a). The Waipaoa and
132 Waiapu rivers (Fig. 1) provide most of the turbidite material, contributing to an annual sediment
133 delivery of 70 Mt/yr off the Raukumara Peninsula (Hicks and Shankar, 2003). The present day
134 sedimentation rate estimated from ^{210}Pb activity decreases from $\sim 1,000$ cm/ka on the continental
135 shelf to ~ 100 cm/ka in mid-slope basins (Alexander et al., 2010; Kniskern et al., 2010), but forest
136 clearing during human settlement in New Zealand 500 – 700 yr ago resulted in river sediment fluxes

137 110 to 660% greater than during the Holocene interglacial period (McGlone and Wilmshurst, 1999;
138 Kettner et al., 2007; Paquet et al., 2009).

139 Turbidites along the northern Hikurangi Margin are recognised as cm-thick fining up sandy to silty
140 beds rich in volcanoclastic debris, quartz and bioclasts (Pouderoux et al., 2012a; Fig. 2). Few debrites,
141 characterised by <35 cm-thick chaotic silty-clay beds with sand to pebble size clasts are also
142 recognised. These two lithotypes alternate with hemipelagites, which consists of 1 – 90 cm-thick
143 layers of olive-grey silty-clay, and tephra layers, composed of volcanoclastic debris, mainly glass
144 shards and pumiceous lapilli, and arranged in <10 cm-thick silty beds (Table 1). Boundaries between
145 lithotypes are usually sharp, except between turbidites tails and hemipelagites.

146 Turbidites tails and hemipelagites have similar grain-size and texture but differ by their colours, with
147 hemipelagites being lighter than turbidites. Hemipelagites are better sorted than turbidites, with a
148 characteristic single grain-size peak < 10 μm . The composition of the silty fraction in hemipelagites
149 explains the colour variations with abundant volcanoclastic debris and low quartz content when
150 compared to turbidites, which show abundant quartz and low volcanoclastic debris. The differences
151 are confirmed by $\delta^{13}\text{C}$ and C/N values, with higher $\delta^{13}\text{C}$ and lower C/N in hemipelagites than in
152 turbidites (Pouderoux et al., 2012b). Erosion at the base of turbidites is considered negligible for
153 those deposited during the Holocene but common during the Late Pleistocene (Pouderoux et al.,
154 2012a). Floods and volcanic eruptions may have occasionally triggered turbidity currents in the area
155 and resulted in the deposition of hyperincites and primary monomagmatic turbidites (Pouderoux et
156 al., 2012a). The majority of turbidites contains benthic foraminifers indicative of environments ≥ 150
157 m, suggesting an upper slope origin of the reworked material. In Poverty, 67 synchronous turbidites
158 are interpreted as being originated from earthquake-triggered slope failures on the upper
159 continental slope between 820 ± 190 and $17,730 \pm 700$ yr BP (Pouderoux et al., 2012b).

160 **3. METHODS**

161 **3.1. Sediment core analyses**

162 The present study is based on detailed stratigraphic correlation of sediment cores collected in
163 Poverty, Ruatoria and Matakaoa (Fig. 1; Table 2). Cores were recovered from strategic locations to
164 sample turbidites deposition since the Last Glacial Maximum (Proust et al., 2006). Four giant piston
165 cores, 12 to 20 m long, were collected during the MD152 MATACORE voyage of *R.V. Marion-Dufresne*
166 (Proust et al., 2006; 2008) in Poverty and Ruatoria in water depth ranging from 1400 to 3500 m. In
167 Poverty, MD06-3003 and MD06-3002 targeted two mid-slope basins in water depths of 1390 m and
168 2300 m respectively. In Ruatoria, MD06-3009 was collected in 2940 m water depth on top of the
169 Ruatoria Debris Avalanche and ~ 250 m above main sediment pathways. MD06-3008 was collected in
170 3520 m water depth in the Hikurangi Trough. Nine short piston cores were acquired onboard *R.V.*
171 *Tangaroa* during research voyages TAN0314 (Carter et al., 2003) and TAN0810 (Lamarche et al.,
172 2008b) in Ruatoria and Matakaoa in water depth ranging from 650 to 3400 m. In Ruatoria, Tan0810-2
173 and -3 were collected on the upper continental slope in water depth of ca. 1080 m, and Tan0810-6
174 was retrieved from the floor of the 3400 m-deep Hikurangi Trough. In Matakaoa, Tan0810-9 to -13
175 were collected within the channel/levee complex of the Matakaoa Turbidite System in water depth
176 ranging from 1090 to 1260 m (Fig. 1). More specifically, Tan0810-9 and -12 were collected within the
177 Matakaoa channel, Tan0810-10 on the left hand levee and Tan0810-11 and -13 on the right hand
178 levee, all less than 2 km away from the channel. Tan0314-8 was collected within the deep-sea fan in
179 2030 m water depth.

180 Petrophysical analyses, including continuous gamma density, magnetic susceptibility and P-wave
181 velocity, were obtained on split cores at 1 cm intervals using a Geotek Multi-Sensor Track. These
182 measurements tied to the main lithofacies and lithofacies successions proved to be essential for
183 core-to-core correlation. P-wave velocities tends to be underestimated (1200-1500 m/s) but the

184 downcore fluctuations, which were used for core-to-core correlations are similar to those in the
 185 density and magnetic susceptibility measurements as well as facies variations so that they are
 186 considered usable.

187 **3.2. Age model**

188 The stratigraphic framework is based on tephrochronology and AMS ^{14}C radiochronology with an
 189 average frequency of 0.7 to 2 ages per meter of core (Pouderoux et al., 2012a). Tephra were
 190 characterised by glass chemistry, mineralogy and their stratigraphic position and tied to the well-
 191 established regional charts of volcanic eruptions to get their precise calibrated ages (Shane, 2000;
 192 Lowe et al., 2008). The cores contain one to six tephra layers correlated to the nine large volcanic
 193 eruptions that occurred in the Central Volcanic Region. Radiocarbon dating was performed on hand-
 194 picked mixed planktonic foraminifers from hemipelagite samples collected 0.7-1.0 cm below
 195 turbidite layers. Raw radiocarbon ages are calibrated using the Oxcal 4.1 software (Bronk Ramsey,
 196 2008) with a regional reservoir age of $395 \pm 57\text{yr}$ ($\Delta R = -5 \pm 57\text{yr}$) calculated from the published East
 197 Cape reservoir ages (Kalish, 1993; Higham and Hogg, 1995). A reservoir age of $800 \pm 110\text{yr}$ ($\Delta R = 400$
 198 $\pm 110\text{yr}$) was applied for ^{14}C dates between 12,400 and 12,900 ^{14}C yr BP, as temporal variations of the
 199 reservoir age were identified during this period (Carter et al., 2008; Sikes et al., 2000).

200 We derive the age model for each core by interpolating hemipelagite sedimentation rates between
 201 time markers (tephra and ^{14}C ages) according to the *P_Sequence* deposition model of the Oxcal 4.1
 202 software, following the approach developed by Pouderoux et al. (2012b). The model provides the
 203 68% and 95% probability age ranges (1σ and 2σ) of each turbidite. In the following sections, ages are
 204 reported with 2σ uncertainties. The 2σ ages are considered reliable for Holocene turbidites as basal
 205 erosion is negligible, but the age may be overestimated during the Late Pleistocene when basal
 206 erosion is likely to have occurred (Pouderoux et al., 2012a).

207 **3.3. Correlation criteria**

208 We use four criteria to correlate turbidites from core to core (see SM2). (1) Tephra layers form
 209 absolute time-lines and robust stratigraphic markers. Their identification in distinct cores constitutes
 210 the first criteria for correlating events across the margin. (2) The estimated age of each turbidite is
 211 used as second criteria to make stratigraphic correlations between two tephra layers. However,
 212 because of the high number of turbidites and their age uncertainties, correlation from one core to
 213 another can be ambiguous at time. (3) We use the similarity in petrophysical properties (gamma
 214 density, magnetic susceptibility and P-wave velocity) and facies as third criteria, and (4) the thickness
 215 of turbidites and hemipelagite as fourth criteria to ascertain the correlations. Petrophysical
 216 properties are good correlation criteria (Patton et al., 2013; Goldfinger et al., 2012; Gracia et al.,
 217 2010) because turbidite coarse grain-size or volcanoclastic compositions typically result in high
 218 densities, magnetic susceptibilities and P-wave velocities and correlate well from core to core (Fig. 2).

219 **3.4. Terminology**

220 In this study, the term “*turbidite event*” (Tx) refers to a single, well-dated depositional episode under-
 221 and overlain by hemipelagite. Stacked turbidites with no intertwined hemipelagite are considered as
 222 a single depositional event, as only the presence of hemipelagite guarantees the occurrence of
 223 significant time between two successive events. Two turbidites separated by a tephra layer represent
 224 two distinct events as usually tephra settle down within days to months after the volcanic eruptions
 225 (Wiesner et al., 1995).

226 “*Basin events*” represent synchronous turbidite events recorded in at least two cores in a single re-
 227 entrant. Basin events are labelled Px in Poverty, Rx in Ruatoria and Mx in Matakaoa, x being the
 228 event sequential number in the re-entrant from the youngest to the oldest.

229 Hikurangi “margin events” (*Hx*) are synchronous *basin events* recorded in at least two re-entrants
 230 along the northern Hikurangi Margin. The age of a *margin event* is determined by the intersection of
 231 the age ranges shared by the synchronous *basin events*. Non-correlative events are called “*isolated*
 232 *events*”.

233 4. RESULTS

234 4.1. Petrophysical properties of sediments

235 The downcore variability in gamma density, magnetic susceptibility (MS) and P-wave velocity (V_p)
 236 depends on sediment lithology and the alternation of hemipelagites, turbidites and tephra (Fig. 2;
 237 Table 3). Hemipelagites usually have constant low density (1.1 to 1.8 g/cm²), MS (10 to 60 SI) and V_p
 238 (1225 to 1775 m/s). Turbidites have systematically higher density (1.2 to 2.2 g/cm²), MS (10 to 120 SI)
 239 and V_p (1225 to 1950 m/s) with a decreasing trend from base to top. When preserved, tephra show
 240 values similar to turbidites, but higher than hemipelagites, with characteristic sharp variations at
 241 their base and top boundaries.

242 The boundary between turbidites and hemipelagites is progressive and difficult to identify using
 243 petrophysical properties alone. Turbidites show generally a progressive decrease while hemipelagites
 244 have stable values. Usually the decreasing trend of turbidite sequence is in good agreement with
 245 grain-size measurements, and coarser beds are noticed by a sharp increase of the density, MS and V_p
 246 (Fig. 2).

247 4.2. Age model and time-lines

248 Cores in Ruatoria cover a continuous age range from 630 ±10 yr BP to 15,360 ±70 ¹⁴C yr, with the
 249 oldest sediments contained in cores MD06-3008 and MD06-3009 (Pouderoux et al., 2012a). In
 250 Matakaoa, cores from the turbidite plain (Tan0810-9 to -13) cover a continuous age range from 630
 251 ±10 yr BP to 4,710 ±40 ¹⁴C yr. In core Tan0314-8, two basal ¹⁴C ages complement the tephra
 252 identification of Joanne et al. (2010), and show that the core contains a truncated record from 5,530
 253 ±60 yr BP to 13,910 ±70 ¹⁴C year.

254 Age models provide an age estimate for every single turbidite event (Fig. 3). Overall the average 2σ
 255 age range is 410 years (13 – 1141) in Ruatoria and 327 years (13 – 970) in Matakaoa, comparable to
 256 the 300 year (25 – 757) 2σ age range calculated in Poverty (Pouderoux et al., 2012b). The ages of
 257 turbidite events in Ruatoria and Matakaoa correspond to the ages calculated at each corrected depth
 258 and ranges from 170 ±140 to 18,150 ±150 yr BP, similar to what was determined in Poverty (820
 259 ±190 to 17,730 ±700 yr BP). From 0 to 6 kyr, 11 cores (all but MD06-3002 and Tan0314-8) are usable
 260 to correlate turbidite events along the margin, while from 6 to 17 kyr, only five cores provide
 261 potential for turbidite events correlations.

262 4.3. Core-to-core correlations

263 4.3.1. At basin scale

264 In Ruatoria, turbidite events are thick, commonly >10cm, and correlate well from the upper slope to
 265 the top of the debris avalanche and the Hikurangi Trough (Fig. 4; see also SM2 Fig. SM2.01). Cores on
 266 the upper slope (Tan0810-2 and -3) and on the top of the Ruatoria Debris Avalanche (MD06-3009)
 267 contain only basin events whereas the cores in the Hikurangi Trough (MD06-3008 and Tan0810-6)
 268 contain basin events and scattered isolated events. Two of these isolated events are primary
 269 monomagmatic turbidites deposited directly after the Taupo and Waimihia eruptions (Pouderoux et
 270 al., 2012a). We identified 30 basin events in Ruatoria, from 390 ±250 to 15,940 ±580yr BP. Basin
 271 events younger than 6 ka (R1 to R14) correlate in all five cores, whereas events from 6 to 17 ka (R15
 272 to R30) correlate only in cores MD06-3008 and MD06-3009 (R15 to R30). The mean recurrence

273 intervals of these 30 basin events is 520 years. Only MD06-3009 contains material older than ~17 ka
 274 with a recurrence intervals of turbidite events of 97 years. This important difference in return time
 275 suggests that these turbidite events cannot be used as proxy for basin events for the period older
 276 than ~17 kyr.

277 In Matakaoa, turbidite events are commonly <5cm-thick, and often homogenised in the hemipelagite
 278 background because of severe bioturbation, hence a higher uncertainty in our interpretation of thin
 279 events as isolated or basin events than for Ruatoria and Poverty. We identify 19 basin events (M1 to
 280 M19) deposited in Matakaoa between 170 ± 140 and $16,400 \pm 780$ yr BP (Fig. 5; see also SM2 Fig.
 281 SM2.02). Basin events younger than 5 ka are only recognised in the turbidite plain (M1 to M9). Cores
 282 Tan0810-9, -10 and -13 mostly record basin events while cores Tan0810-11 and -12 are scattered
 283 with isolated events. Basin event M10 dated at $5,130 \pm 290$ yr BP is the only event that correlates in
 284 the turbidite plain (Tan0810-10) and the deep-sea fan (Tan0314-8). No other events in the deep-sea
 285 fan correlate with upslope events because of the lack of age overlap in cores. Although this
 286 characterises them as isolated events, they likely represent basin events as the core was collected in
 287 the deep-sea fan at the outlet of the Matakaoa Turbidite System.

288 4.3.2. *At margin-scale*

289 Twenty-eight basin events are synchronous in two or more re-entrants over the last ~16 kyr, and
 290 therefore fulfil the criteria of margin events (Fig. 6; see also SM2 Fig. SM2.03). Ten margin events are
 291 documented in the three re-entrants (H3, H5, H8, H12, H23, H30, H33, H35, H41, and H43; Table 4).
 292 H4 and H6 are identified from correlative basin events in Poverty and Ruatoria but also correlate to
 293 isolated events in Matakaoa. The remaining 16 margin events are only correlative in Poverty and
 294 Ruatoria (H7, H9-11, H13-14, H16, H18-19, H22, H26-28, H31-32, and H34; Table 4). All basin events
 295 in Ruatoria, except for the two youngest R1 and R2 dated at 390 ± 250 and 790 ± 200 yr BP, fulfil the
 296 criteria of margin events. Originally, R1 and R2 were not recognised as margin events since they did
 297 not correlate to basin events in Poverty. However, because R1 and R2 correlate to isolated events in
 298 Matakaoa and because material younger than 820 yr BP was not retrieved in Poverty cores, we
 299 interpret them as margin events as all basin events in Ruatoria are margin events (H1 and H2 in Table
 300 4).

301 A further 15 basin events in Poverty, dated from 6 to 16 kyr (P24, P26, P30, P33, P38, P40, P49, P57-
 302 61, P63, P65 and P66), correlate to 16 isolated events in core MD06-3008, and are interpreted as
 303 margin events (H15, H17, H20, H21, H24, H25, H29, H36-40, H42 and H44-46; Table 4). That specific
 304 period recorded in Ruatoria only by cores MD06-3008 and MD06-3009 is characterised by an
 305 unknown number of MD06-3008 isolated events not recognised as Ruatoria basin events due to a
 306 lack of data, as core MD06-3009 is located on a topographic high. In addition, during the period 0-6
 307 kyr recorded in all cores, seven basin events not recorded in MD06-3009 (R1-2, R8-9, R11-12 and
 308 R14) fulfil the criteria of margin events. These observations confirm that the 15 correlated basin
 309 events in Poverty with isolated events in MD06-3008 are margin events.

310 Overall, the margin-scale correlation results in the identification of 46 margin events deposited from
 311 390 ± 170 to $16,450 \pm 310$ yr BP, with an average age uncertainty of ~170 years (ranging from 6 to 400
 312 years; Table 4).

313 5. DISCUSSION

314 5.1. *Earthquake's origin of turbidites*

315 The associations of benthic foraminifers contained in basin and isolated events suggest a shelf edge
 316 origin of the reworked material within 150-200 m of water depth in Ruatoria and Poverty, and 150-
 317 600 m of water depth in Matakaoa (Pouderoux et al., 2012a; 2012b). Isolated events in the trench
 318 interpreted as margin events also rework material from the upper slope, which corroborates that

319 turbidity currents originate from the upper slope and suggests a slope failure origin of most gravity
320 flows. This suggestion is supported by the present day high state of instability of the Hikurangi
321 Margin's upper slope associated with gas and fluids (Lewis and Marshall, 1996; Orpin, 2004; Faure et
322 al., 2006), and high sediment supply during the Holocene (Hicks and Shankar, 2003; Addington et al.,
323 2007; Lewis et al., 2004). At margin-scale, five margin events are interpreted as triggered by
324 mechanisms other than slope failure: H9, H22, H23 and H28 which contain at least one hyperpycnite,
325 and H5 which is a primary monomagmatic turbidite. The remaining 41 margin events are related to
326 synchronous slope failures along the 100 km-long northern Hikurangi Margin and all originated
327 within 150-600 m of water depth since ~16ka (Table 4; see also SM2 Fig. SM2.03).

328 Synchronicity of trigger over such wide areas is recognised as the most likely signature of large
329 earthquakes in other regions of the world (Gracia et al., 2010; Goldfinger et al., 2003; 2012),
330 although storms and tsunami waves may occasionally trigger gravity flows and slope failures (Mulder
331 et al., 2001; Puig et al., 2004). Repeated storms have occurred along the northern Hikurangi Margin
332 over the Late Holocene (Page et al., 2010), and are likely to have affected the seafloor by
333 remobilising surficial sediments of the shelf or to have triggered sediment liquefaction (Lee and
334 Edwards, 1986; Ma et al., 2010; Goldfinger et al., 2012). Puig et al. (2004) characterized sediment
335 gravity flows directed down-canyon during storms on the California margin, and Mulder et al. (2001)
336 described a storm-generated turbidite at 650 m water depth in a canyon head few months after a
337 large historical storm in the Bay of Biscay. Furthermore, turbidity currents generated during storms
338 are typically less voluminous and spread over smaller geographic areas than those triggered by
339 earthquakes (Gorsline et al., 2000; Goldfinger et al., 2007; Blumberg et al., 2008). They also usually
340 settle in water depth < 1000 m (Puig et al., 2004). The Matakaoa canyon's head contains a stack of
341 recent cm-thick turbidites recorded in 650 m water depth, sedimentologically similar to the storm-
342 induced turbidites found in the Bay of Biscay (Pouderoux, 2012; Pouderoux et al., 2012a). Even if
343 confirmed along the Hikurangi Margin, these storm-related turbidites are likely to be restricted to
344 canyon heads and not observed at water depth > 1000 m where we found deep-sea synchronous
345 turbidites. Tsunami waves may also generate slope failures or turbidity currents directed down-slope
346 and initiated on the continental shelf (Shanmugam, 2006). The largest historical tsunami affecting
347 the region produced a run-up of ~10 m north of Gisborne and was triggered by the local M_w 7.1
348 Gisborne earthquake of 25 March 1947 (Downes et al., 2000; Doser and Webb, 2003). Other tsunami
349 run-ups of 10 m interpreted as likely generated by local earthquakes appear in the New Zealand
350 paleo-tsunamis record (Goff and Dominey-Howes, 2009). The tsunami generated by the 23 May 1960
351 M_w 9.5 Chilean earthquake, the largest earthquake ever recorded worldwide, resulted in a 3 m-high
352 run-up at Gisborne and 4.5 m in Hawke's Bay (geonet.org.nz). This suggests that tsunamis potentially
353 able to trigger slope failures are more likely to be generated by local earthquakes, which ground-
354 shaking is far more likely to trigger slope failures than the tsunamis wave itself.

355 Earthquakes are therefore the most likely triggering mechanism during the last sea-level highstand
356 for the generation of synchronous slope failures at the origin of margin events, as inferred in other
357 active margins (e.g. Noda et al., 2008; Gracia et al., 2010; Goldfinger et al., 2012). Storms and
358 tsunamis are possibly secondary players during the marine transgression. Even if the average
359 recurrence interval of margin events varies slightly between the Late Holocene highstand (~440 yr
360 over the last 7.5 kyr) and the Late Pleistocene – Early Holocene marine transgression (~375 yr from
361 7.5 to 16ka), it is possible that earthquake ground shaking was not the sole triggering mechanism of
362 synchronous slope failures before 7.5 ka. It is extremely difficult to estimate the impact of storms
363 and tsunamis waves on slope stability during the marine transgression. Consequently, if the turbidite
364 record of the northern Hikurangi Margin is considered a good proxy for paleo-earthquakes during the
365 Late Holocene and provides a calendar of 17 large earthquakes that occurred in the region between
366 390 ± 170 and 7480 ± 120 yr BP with an average return time of ~440 years, the Late Pleistocene –
367 Early Holocene part of this record incurs an increased uncertainty. Nevertheless, the turbidite record
368 during the marine transgression could be use complementary to the Late Holocene highstand to

369 constraint the earthquake hazard on the area, knowing that the recurrence intervals may be
 370 underestimated and the hazard overestimated.

371 **5.2. Source and magnitude estimation of paleo-earthquakes**

372 The 9 kyr-long coastal paleo-earthquake record is time correlative to margin events (Fig. 7). The 2σ
 373 age range of margin events overlies the age of all marine terraces uplifts except one at Pakarae river
 374 mouth and one at Mahia peninsula. These uplifts were interpreted as the result of the rupture of
 375 near-shore Gable End and Lachlan 3 faults (Wilson et al., 2007; Litchfield et al., 2010). Conclusions
 376 are however sometimes equivocal: H4 and H10 are time correlative to the rupture of these two faults
 377 suggesting a simultaneous rupture of the two faults or a rupture of the subduction interface; H1
 378 correlates with uplifts at the Pakarae river mouth and Mahia Peninsula and with a paleo-tsunami.

379 Over the last 7.5 kyr, ten margin events correlate to the paleo-earthquake record onland. These
 380 margin events are part of a group of 20 particularly large margin events identified over the last 16 ka
 381 (Table 4; see also SM2 Fig. SM2.03), characterized by a ~ 40 cm-thick turbidite layer deposited on a
 382 topographic high in Ruatoria, ~ 250 m above the main sediment pathways (core MD06-3009); such
 383 thickness is twice that of turbidites in other cores. They correspond to exceptionally voluminous
 384 turbidity currents triggered simultaneously in the three re-entrants. Such sedimentological evidences
 385 coupled with the systematic correlation with onland record are consistent with a triggering by
 386 subduction earthquakes initiated on the Raukumara segment of the subduction interface.

387 Slope failure triggering and turbidite deposition depend more on the shaking intensity felt on the
 388 slope than on the magnitude of the earthquake itself. The shaking intensity can be evaluated by
 389 calculating the peak ground acceleration (PGA) (Lee and Edwards, 1986; Douglas, 2000). The PGA
 390 threshold for slope failure and turbidity current generation in conditions similar to the upper slope of
 391 the northern Hikurangi Margin ranges from 0.08 to 0.6 g and more likely from 0.08 to 0.15 g (Lee et
 392 al., 1999; Lykousis et al., 2002; Strasser et al., 2007; Noda et al., 2008; Dan et al., 2009; see also SM3).
 393 By using the PGA empirical attenuation relationships of Cousins et al. (1999) and Si and Midoriwaka
 394 (1999), which are best suited for the region, we estimated the earthquake magnitude (M_w) and the
 395 location of the hypocentre (depth and distance from the upper slope) of paleoearthquakes
 396 responsible for the triggering of upper slope failures. This enabled us to create isomagnitude maps
 397 and to identify areas where an earthquake of a given magnitude and origin has to occur to trigger
 398 synchronous slope failures and turbidity currents (Fig. 8; see also SM3 Fig. SM3). Superposing these
 399 isomagnitude maps over the known active faults (Stirling et al., 2012; Litchfield et al., in press)
 400 provides the means to infer the sources of paleo-earthquakes capable of generating our deep-sea
 401 turbidite record. (Table 5; see also SM3 Fig. SM3).

402 We disregard using a PGA threshold of 0.15 g for the triggering of turbidity currents as this would
 403 suggests that the 17 Late Holocene margin events were all subduction interface earthquakes, which
 404 is very unlikely (Table 5). A more reasonable PGA threshold between 0.08 and 0.1 g suggests that
 405 these events were associated with ruptures of upper plate faults or the subduction interface, all
 406 capable of generating $M_w \geq 7.3$. This latter interpretation is more consistent with our findings that
 407 only 10 out of these 17 margin events correspond to large subduction interface earthquakes. These
 408 values of PGA are also extremely closed to that determined along the Japan (Noda et al., 2008) and
 409 Algerian margins (Dan et al., 2009).

410 The 17 margin events provide a meaningful calendar of $M_w \geq 7.3$ paleo-earthquakes that have
 411 affected the region between 390 ± 170 and $7,480 \pm 120$ yr BP : 10 subduction interface earthquakes
 412 and 7 upper plate earthquakes (Tables 4 and 5). The results show that only three out of the 26
 413 known active fault sources recognized by Stirling et al. (2012) in the offshore northern Hikurangi
 414 Margin (faults 2, 6 and 7, namely the Raukumara subduction interface segment, Ruatoria South 1 and
 415 Areil Bank; Table 5) are responsible for this deep sea turbidite record.

416 **5.3. Recurrence intervals of large earthquakes**

417 Our turbidite paleoseismology approach reveals that the recurrence interval (RI) of $M_w \geq 7.3$
 418 earthquakes ranges from 150 to 1240 years with an average of 440 years during the Late Holocene
 419 (Figs. 9 and 10). Seismological modelling shows that the three active faults identified as the potential
 420 sources of our paleo-earthquakes have an empirical RI of rupture of 1300-1670, 3340, and 720 years
 421 (Table 5; see also SM1; Stirling et al., 2012). Assuming that upper plate faults rupture independently
 422 from the subduction interface (Stirling et al., 2012), the average RI of fault ruptures would be
 423 comprised between 390 and 460 years, which fits well with the 440 years RI of margin events. During
 424 the Late Pleistocene – Early Holocene, margin events had an average RI of 350 years, ranging from a
 425 few years to up to 1100 years, which is in good agreement with the estimated average RI of fault
 426 ruptures from Stirling et al. (2012).

427 The RI of the 10 large margin events triggered by subduction interface earthquakes varies from 370
 428 to 2090 years during the Late Holocene with alternating periods of high and low RI, and an average RI
 429 of ~800 years. A similar pattern is recorded for the Late Pleistocene – Early Holocene period.
 430 Assuming that subduction interface earthquakes are the sole triggering mechanism of these large
 431 margin events over the last 16 kyr, this pattern of RI suggests two different tectonic regimes with
 432 periods of intense activity separated by periods of relative quiescence (Fig. 9). Such scenarios have
 433 been suggested by Berryman et al. (1989) for the Hikurangi Margin, Patton et al. (2009) for the
 434 Sumatra margin and Goldfinger et al. (2009; 2013) for the Cascadia margin. Active periods exhibit
 435 shorter durations (0.6-3 kyr-long) and drastically shorter RI (305-610 years) than quiescence ones
 436 (1.5-3.2 kyr-long with RI range of 1480-2650 years). These RIs differ from the predicted 1300-1670
 437 years calculated using seismological modelling by Stirling et al. (2012). The latter RIs are closer to that
 438 observed during periods of quiescence than that during active periods. The RI and M_w of Stirling et al.
 439 (2012) are maximum values determined from empirical relationships and represent the time needed
 440 for the subduction interface to accumulate enough strain to rupture the full length of the Raukumara
 441 segment. Active seismic periods recorded in the sedimentary record suggest that in a stable tectonic
 442 regime (convergence rate, slip rate, etc...), the deformation and energy released by the subduction
 443 interface may be partitioned with multiple ruptures generating earthquakes less than M_w 8.2 – 8.4.

444 Isomagnitude maps suggest that a M_w 7.5 – 8 earthquake on the subduction interface segment
 445 would be enough to generate a PGA of 0.08 – 0.1 g on the upper slope and therefore trigger
 446 simultaneous turbidity currents along the northern Hikurangi Margin (see SM3 Fig. SM3). Considering
 447 the constraint given by core MD06-3009 and the presence of turbidites on topographic highs, we
 448 propose that during active periods the Raukumara segment of the subduction interface produced
 449 regular large M_w 7.5 – 8 earthquakes. These earthquakes may not rupture the full length of the
 450 Raukumara segment or release the total accumulated strain. This is in good agreement with the two
 451 inferred moderate subduction interface earthquake M_w 7 and 7.1 which affected the Gisborne
 452 district in 1947 (Doser and Webb, 2003).

453 **6. CONCLUSION**

454 This study presents the first chrono-stratigraphic correlation of deep-sea turbidites along the
 455 northern Hikurangi Margin of New Zealand, from the detailed description and age dating of sixteen
 456 sediment cores collected in the Poverty, Ruatoria and Matakaoa re-entrants. Age models provide a
 457 precise age estimate for every single turbidite deposited since the Last Glacial Maximum. The age of
 458 turbidites ranges from 170 ± 140 to $18,150 \pm 150$ yr BP.

459 Basin-scale correlations of turbidites result in the identification of 30 basin events (synchronous
 460 turbidites) deposited in Ruatoria between 390 ± 250 and $15,940 \pm 580$ yr BP, and 19 basin events
 461 deposited in Matakaoa between 170 ± 140 and $16,400 \pm 780$ yr BP. Previous studies have recognised
 462 73 basin events in Poverty, deposited between 820 ± 190 and $17,730 \pm 700$ yr BP.

463 Margin-scale correlations result in the identification of 46 margin events (synchronous basin events)
 464 deposited from 390 ± 170 to $16,450 \pm 310$ yr BP, among which four are recognised as catastrophic
 465 floods deposits (hyperpycnites) and one a primary monomagmatic turbidite. The remaining 41
 466 margin events are related to synchronous slope failures along the margin.

467 Earthquakes are the triggering mechanism of these slope failures during the sea-level highstand, and
 468 are also likely to be the main triggering mechanism during the marine transgression. The turbidite
 469 record of the northern Hikurangi Margin is therefore a good proxy for paleo-earthquakes,
 470 corroborated by the time correlation with onland paleo-earthquake evidences.

471 The use of empirical relationships evaluating the upper slope stability allows us to estimate
 472 magnitude M_w and location (depth and distance from the upper slope) of paleo-earthquake. The 41
 473 margin events correspond to $M_w \geq 7.3$ earthquakes that have affected the region from 390 ± 170 to
 474 $16,450 \pm 310$ yr BP, involving 3 of the 26 known active faults in the region (10%). Recurrence interval
 475 deduced from turbidite chronology is similar to the estimated activity of these three active faults.
 476 Twenty margin events are interpreted as subduction interface earthquakes of $M_w > 7.5$ and up to 8.4
 477 affecting the three re-entrants and able to trigger coeval voluminous turbidity currents together with
 478 onland paleoseismic evidences.

479 Our study shows that large earthquake of $M_w \geq 7.3$ occurring on the northern Hikurangi Margin are
 480 more likely to occur with a RI of 200 ± 100 years, while large to great subduction interface
 481 earthquakes of $M_w > 7.5$ to occur every 550 ± 50 years. RI of subduction interface earthquakes
 482 suggests alternating periods of intense activity with frequent but smaller earthquakes separated by
 483 periods of relative quiescence characterized by rare but more powerful earthquakes.

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769

769 **9. FIGURE CAPTIONS**

770 **Figure 1** - Morpho-tectonic settings of the northern Hikurangi Margin, New Zealand. Red dots show
 771 sediment cores used in this study. The present day average sediment deliveries of the three
 772 main rivers, which catchments are highlighted in grey shade, are indicated in dark grey (from
 773 Hicks and Shankar et al., 2003). White squares are locations of coastal paleo-earthquakes
 774 evidences in (A) Pakarae river mouth, (B) Mahia Peninsula, and (C) northern Hawke's Bay.
 775 Bold lines indicate the main earthquake sources identified by Stirling et al. (2012) and
 776 Litchfield et al. (in press) (see also SM1 Table SM1). Bold teeth-line indicates the subduction
 777 front. Insert: the Kermadec-Hikurangi margin with the Pacific (PAC) and the Australian (AUS)
 778 plates, the Wairarapa (W), Hawke Bay (HB) and Raukumara (R) regions, and the Central
 779 Volcanic Region (CVR). Dark arrow is relative plate motion at the plate boundary from
 780 Beavan et al. (2002).

781 **Figure 2** - Characterisation of turbidites and hemipelagite from visual (photo), internal structures (X
 782 Ray), Geotek petrophysical properties (gamma density, magnetic susceptibility and P-wave
 783 velocity) and grain size (sand, silt and clay percentage). Peaks in grain size correspond to
 784 peaks in the Geotek; the turbidites' fining up trend is illustrated by the Geotek
 785 measurements.

786 **Figure 3** - Example of the Oxcal age model from a giant piston core (MD06-3009) and two short
 787 piston cores (Tan0810-2 and -6), showing the sedimentation rate of hemipelagite through
 788 time. The k parameter is used to define the regularity of the sedimentation rate along the
 789 core: the higher the k parameter, the more linear the sedimentation rate and the smaller the
 790 turbidite age uncertainties. Since hemipelagite settles at an assumed roughly constant rate,
 791 the highest k parameter was chosen for each core. All ages are plotted with their 2σ age
 792 range.

793 **Figure 4** - Simplified correlation diagram of turbidites between cores in Ruatoria (upper slope,
 794 Ruatoria Debris Avalanche and Hikurangi Trough). Detailed correlations in SM2 Fig. SM2.01.
 795 Dashed lines are tephra. Basin events are labelled Rx. The c. 6 ka boundary (thin black line) is
 796 the average basal boundary of short cores. Black stars on isolated events show events
 797 correlated to basin events in Poverty and indicated unrecognised basin events in Ruatoria
 798 (see text for details). Number on the left of logs are turbidite facies as in Table 1.

799 **Figure 5** - Simplified correlation diagram of turbidites between cores in Matakaoa (channel and
 800 levees in the turbidite plain and fan). Detailed correlations in SM2 Fig. SM2.02. Cores are
 801 arranged upstream to downstream from left to right; see caption and legend in figure 4.

802 **Figure 6** - Example of margin-scale correlation of turbidites using one core from Poverty, two from
 803 Ruatoria and two from Matakaoa. Each re-entrant is c. 100 km apart. Detailed correlations in
 804 SM2 Fig. SM2.03. Turbidites are indicated by basic core log, facies (number from 1 to 5) and
 805 Geotek petrophysical properties (density and magnetic susceptibility). See legend and
 806 caption in figure 4. Margin events are labelled Hx. The age of tephra layers (purple) is
 807 reported in cal. yr BP.

808 **Figure 7** - Temporal correlation between margin events and onland evidences of paleo-earthquakes
 809 from the Pakarae river mouth (A; Wilson et al., 2006; 2007), Mahia Peninsula (B; Berryman et
 810 al., 1993), and northern Hawke's Bay (C; Cochran et al., 2006), and paleo-tsunamis identified
 811 along the east coast of the North Island East Coast (Goff and Dominey-Howes, 2009). Purple
 812 bands are age range of correlated margin events. Large margin events interpreted as the
 813 record of subduction interface earthquakes are framed in red.

814 **Figure 8** - Examples of isomagnitude maps defining the area shaken by an earthquake of a given
 815 magnitude, which can trigger synchronous slope failures in the source areas of the turbidites
 816 (blue area in the upper slope). Grey dots show the location of sediment cores. Each map is
 817 generated from an empirical relationship that defines the Peak Ground Acceleration (PGA)
 818 from the magnitude, depth, distance and type of earthquakes (see SM3 for details). Faults
 819 capable of producing earthquake shaking large enough to generate turbidity currents in the
 820 source area are shown in bold red (epicentre shown by red dots) and summarized in Table 5.
 821 Blue dots are epicentres of earthquakes that cannot generate ground-shaking capable of
 822 generating turbidites in the source area. In each map, the isomagnitude M_w 7.5 is shown as
 823 an example of the methodology used to define the M_w 7.5 area. (A) Isomagnitude map built
 824 from the relationship of Cousins et al. (1999), labelled Eq. (1). $PGA = 0.08g$ is reached
 825 simultaneously in the upper slope of Poverty and Ruatoria for a $M_w \geq 6.5$ earthquake located
 826 between both re-entrants. This corresponds to the activity of only two upper plate faults
 827 (red bold lines). See SM3 Fig. SM3 for subduction interface earthquake isomagnitude maps.
 828 (B) Isomagnitude map built from the relationship of Si and Midoriwaka (1999), labelled Eq.
 829 (2). $PGA = 0.1 g$ is reached simultaneously in the three re-entrants for $M_w \geq 7.2$ earthquakes
 830 located between Poverty and Ruatoria. This corresponds to the rupture of the subduction
 831 interface. No upper plate faults can trigger synchronous slope failures in the three re-
 832 entrants.

833 **Figure 9** - Recurrence intervals (RI) of large paleo-earthquakes at the origin of margin events vs
 834 time. (A) RI of $M_w \geq 7.3$ earthquakes at the origin of the 41 margin events identified over the
 835 last c. 16.5 kyr (grey line). Dots correspond to the average age of the margin event Hx vs the
 836 average time span since the last one. The recurrence intervals of $M_w > 7.5$ earthquake at the
 837 origin of the 20 large margin events is indicated by dashed blue line. (B) Distribution of RI of
 838 $M_w > 7.5$ earthquakes showing an alternation of active periods in light yellow with low RI and
 839 numerous earthquakes, and quiescence periods in light blue during which larger RI are noted
 840 (for each period, average RI are noted in italic and their duration noted in bold).

841 **Figure 10** - Frequency diagrams of paleo-earthquakes during (A) the Late Holocene and (B) the Late
 842 Pleistocene – Early Holocene. The latter is likely to be underestimated due to uncertainties
 843 discussed in the paper. $M_w > 7.5$ earthquakes correspond to large margin events and $M_w \geq$
 844 7.3 earthquakes to all margin events (see text for details). Histograms (grey bars) show the
 845 frequency of recurrence interval (RI) within each bin of 100 years, with their statistical
 846 distribution estimated for each centile (black line). The median RI of each plot is noted in
 847 white for histograms and in black for the statistical distribution (cross).

848

849 10. TABLES

850 **Table 1** – Lithotype characteristics of deep-sea sediments along the northern Hikurangi Margin
 851 (summarized from Pouderoux et al., 2012a).

852 **Table 2** – Location and characteristics of sediment cores used in this study. T: gravity flow deposits
 853 (turbidites); H: hemipelagite.

854 **Table 3** – Range of petrophysical properties of hemipelagite and turbidites calculated from the cores.

855 **Table 4** – Margin events (Hx) modelled age deduced from basin events correlation in Poverty,
 856 Ruatoria and Matakaoa.

857 **Table 5** – Earthquake sources and estimated magnitude deduced from the overlap of isomagnitude
858 maps (Fig. 8; see also SM3 Fig. SM3) and the known active faults compiled by Stirling et al (2012).
859 Different scenarios are considered according to the two type of margin events, the three PGA
860 thresholds for slope failures triggering and the two empirical relationships used to build
861 isomagnitude maps .

862

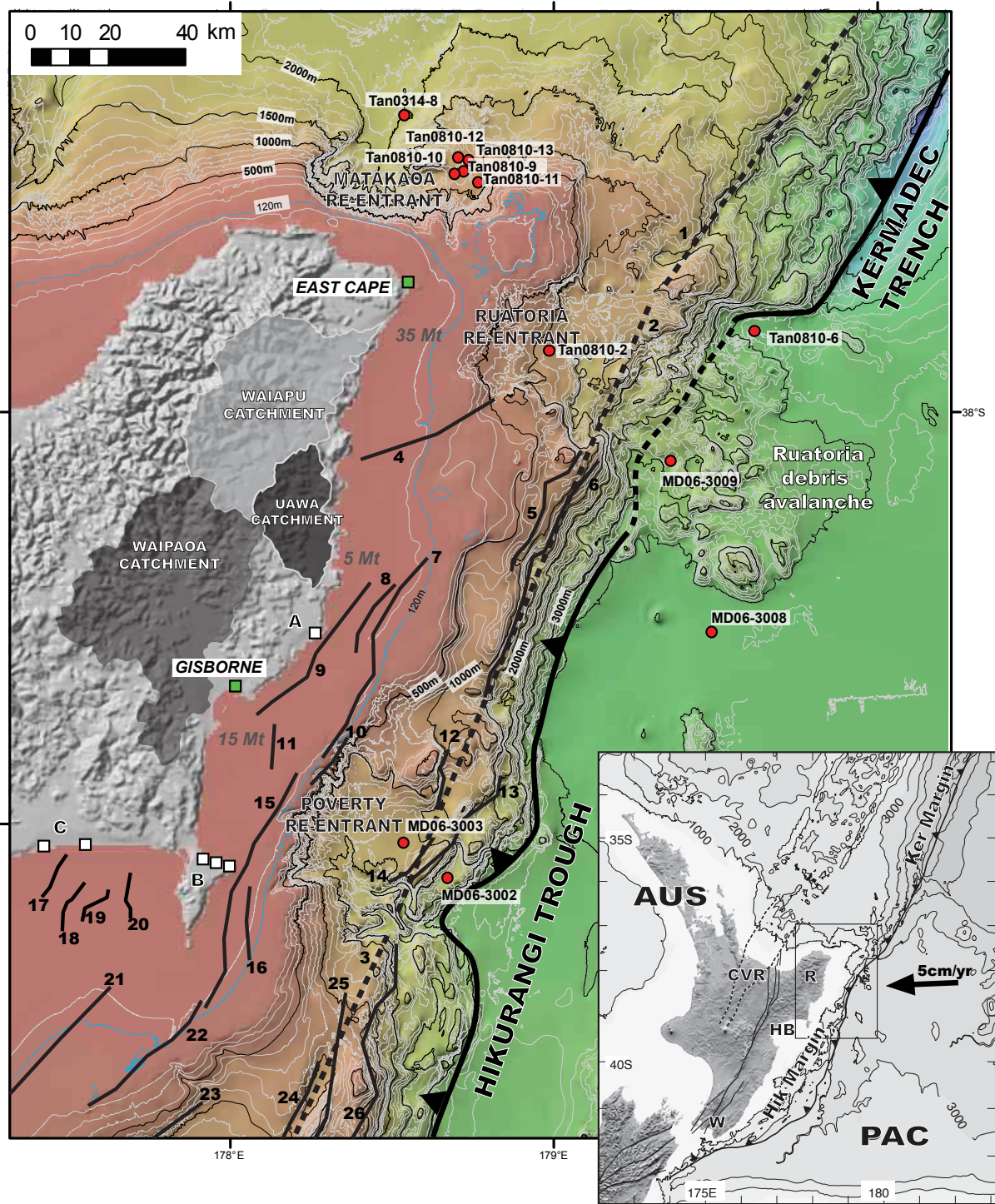


Figure 1

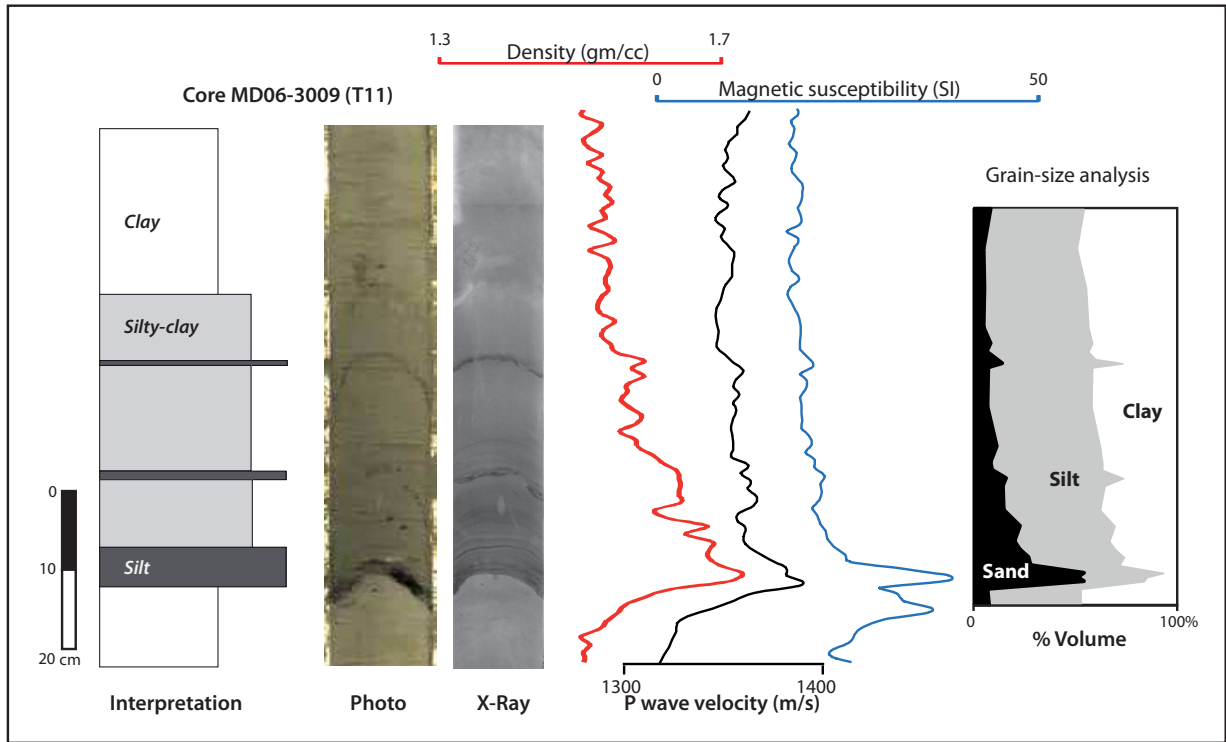
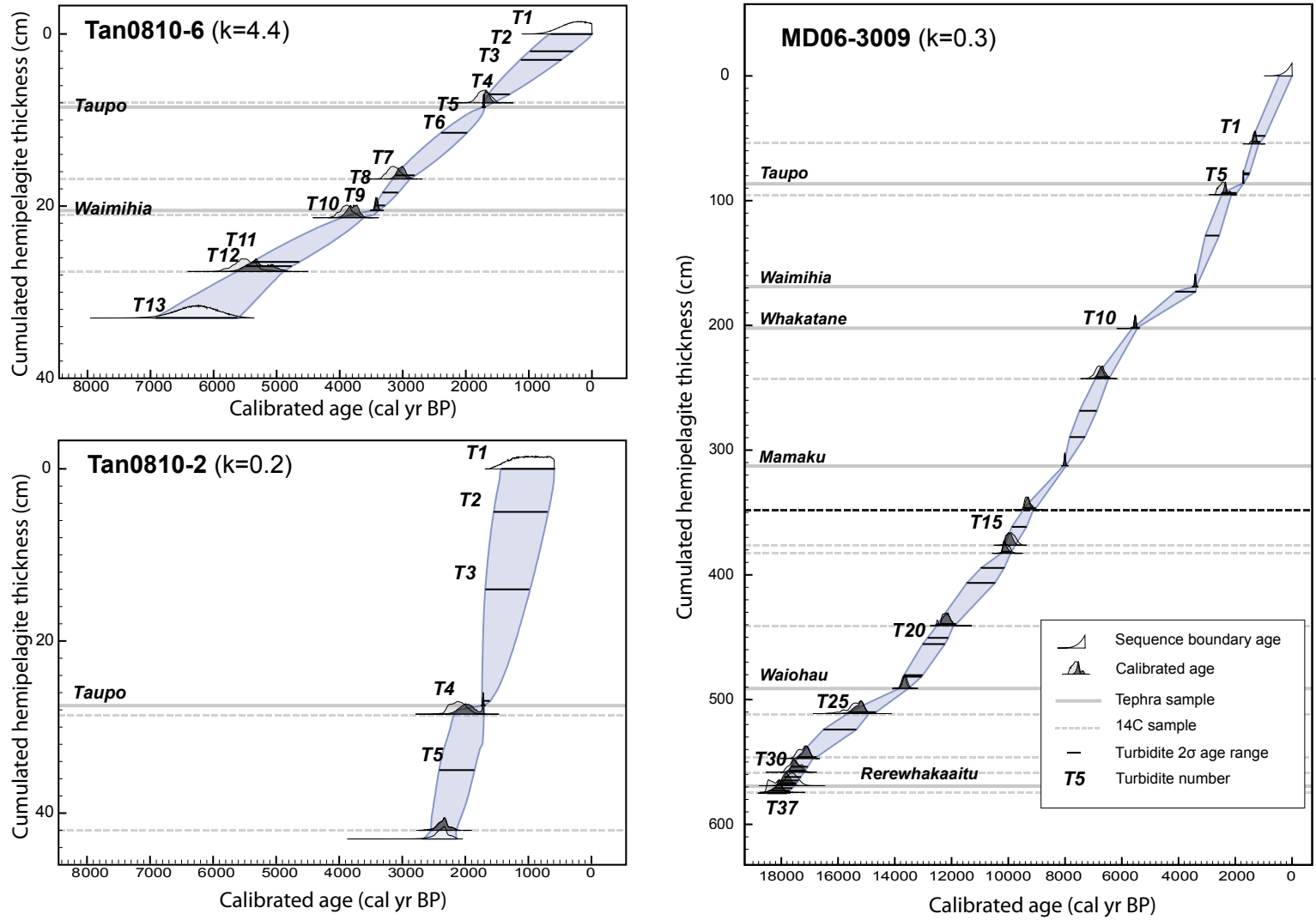


Figure 2

Figure 3



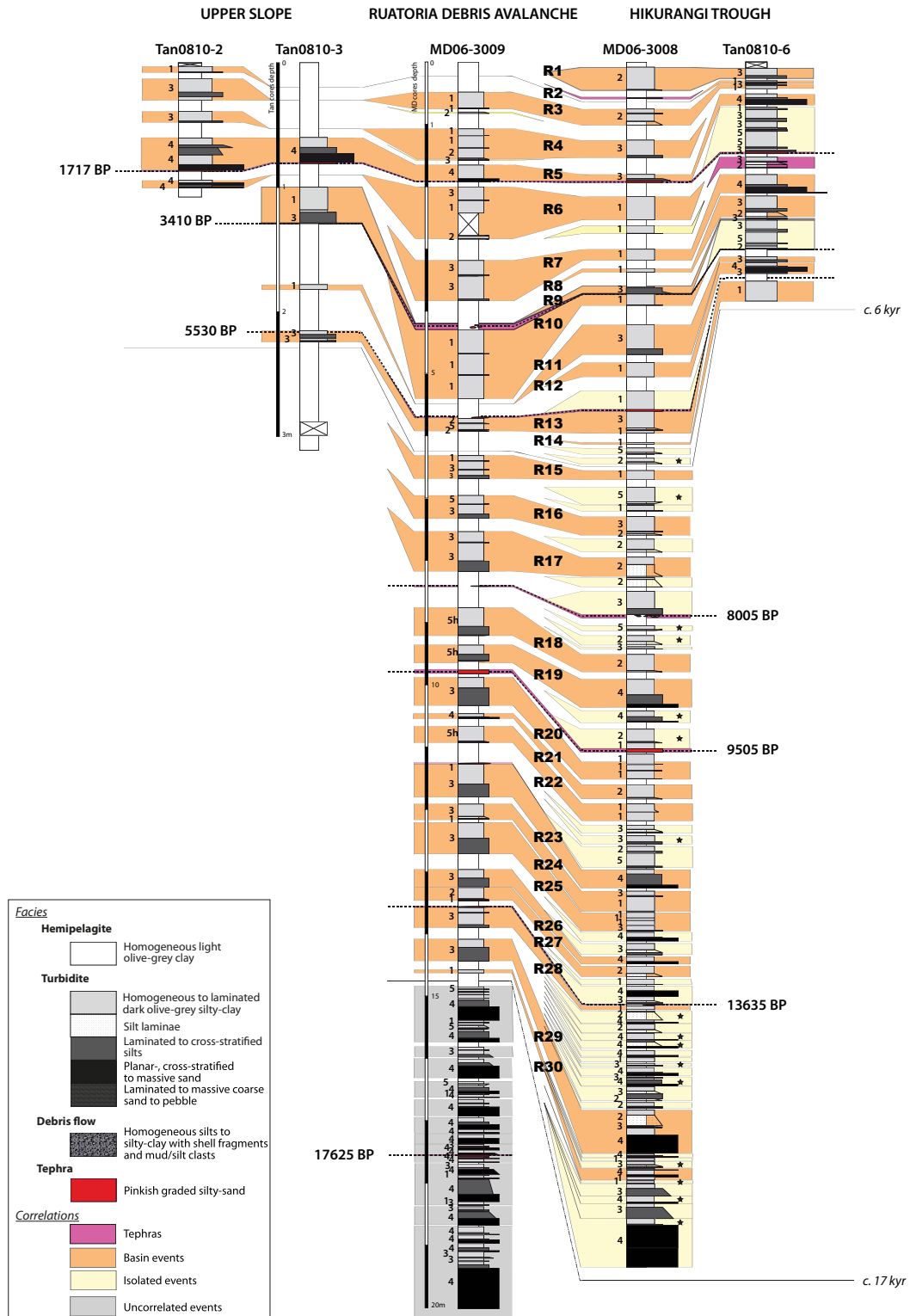


Figure 4

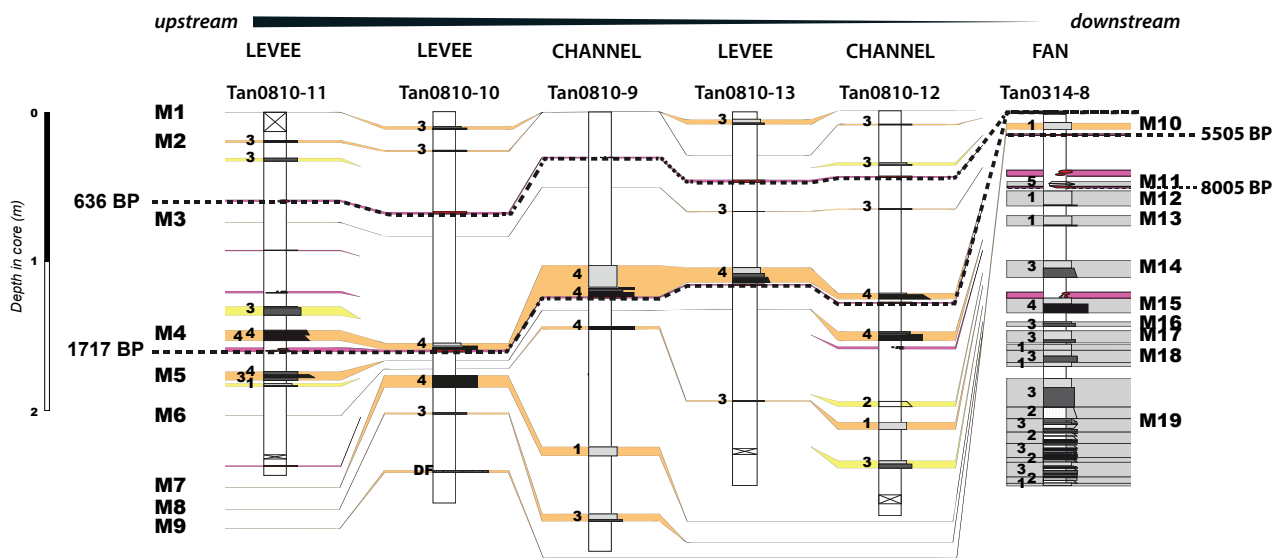


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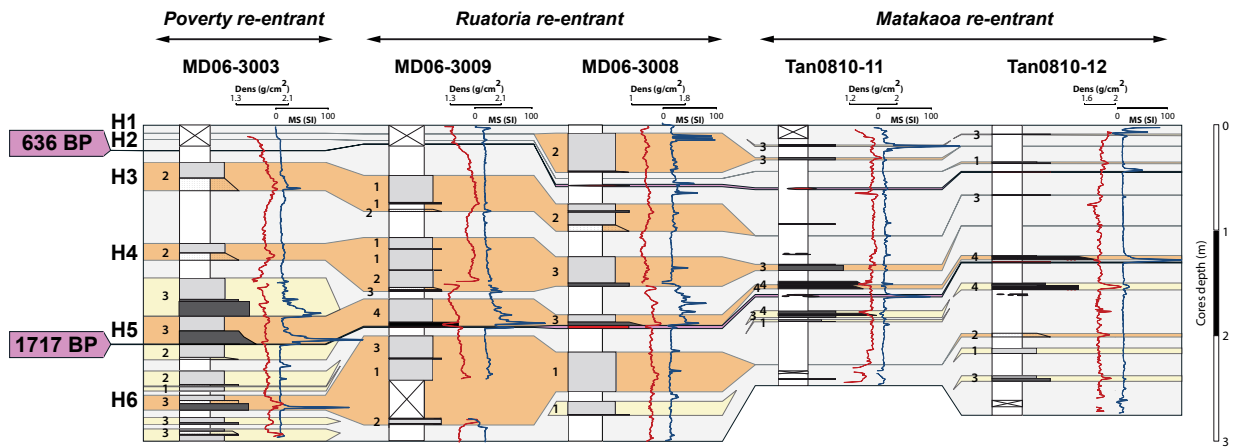


Figure 6

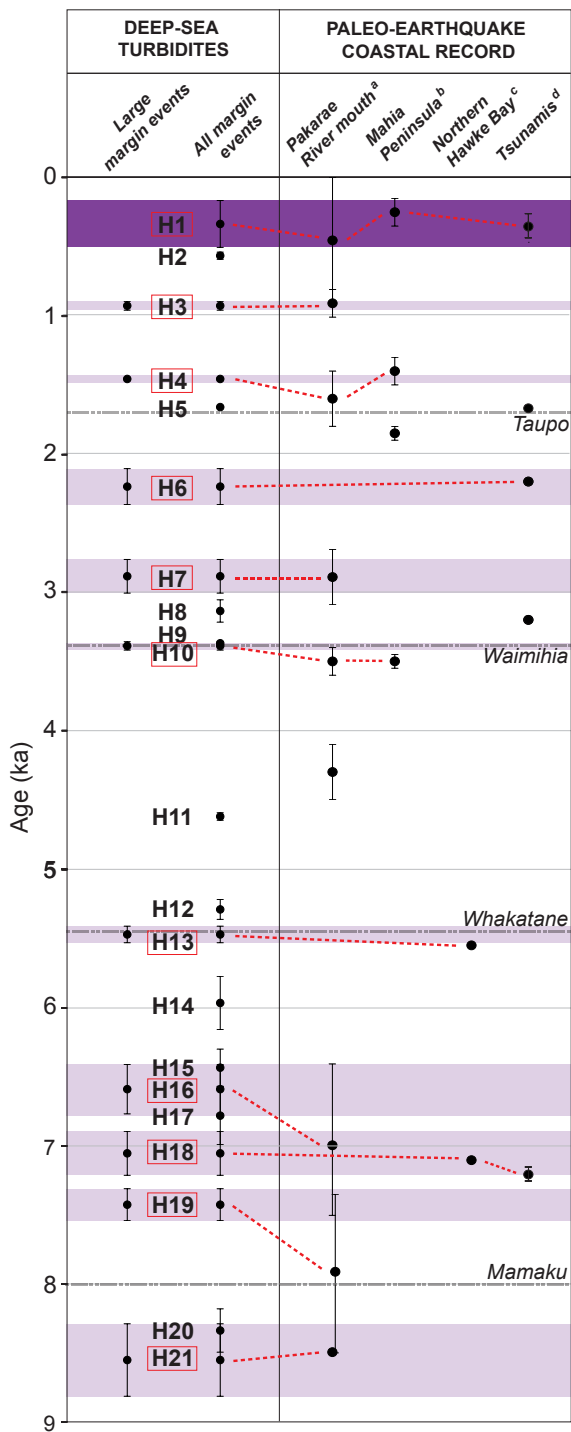


Figure 7

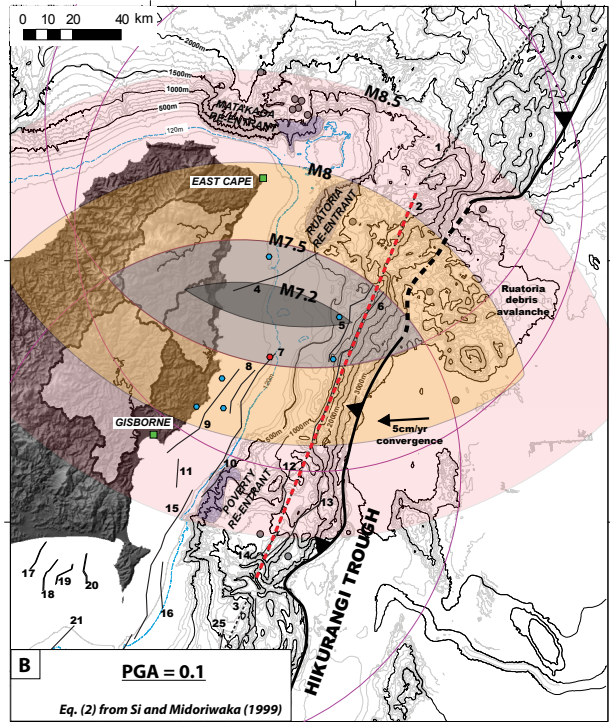
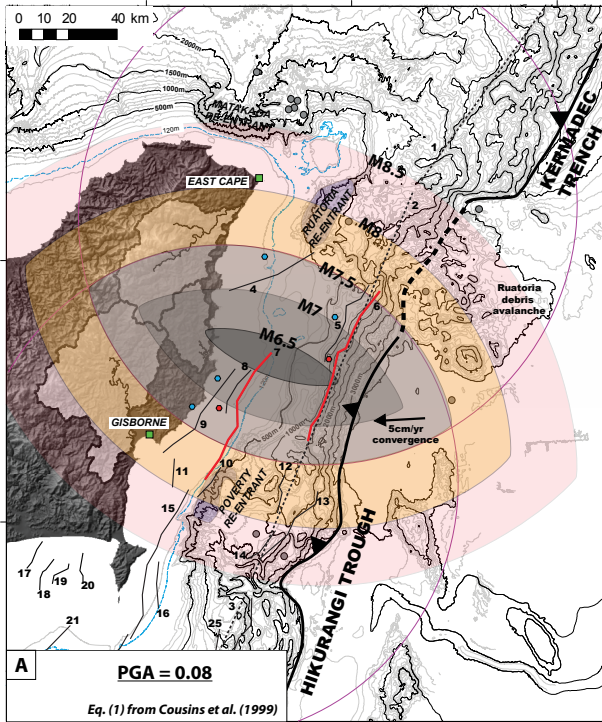
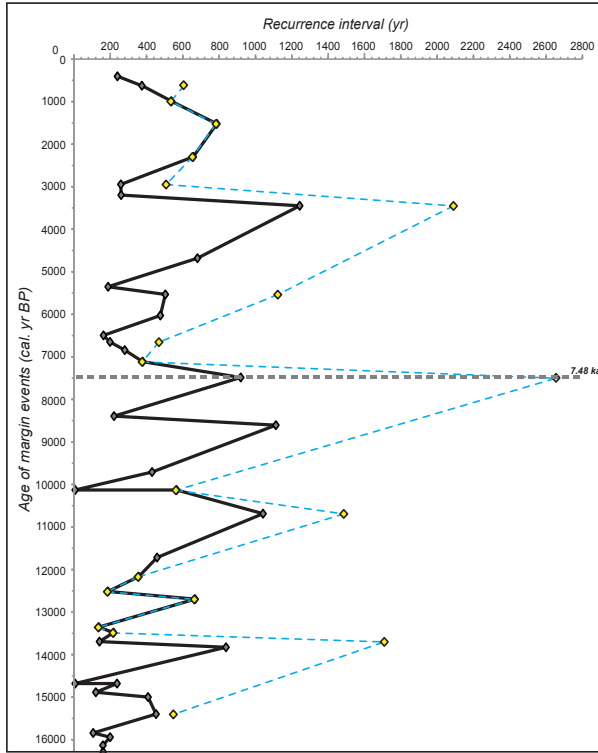


Figure 8

A - Recurrence interval of earthquakes at the origin of all margin events (n=41)



B - Recurrence interval of earthquakes at the origin of largemargin events (n=20)

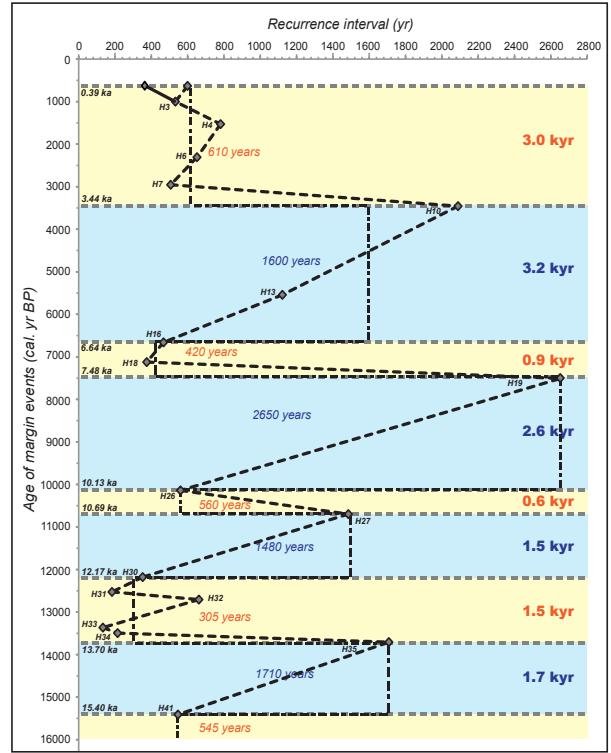
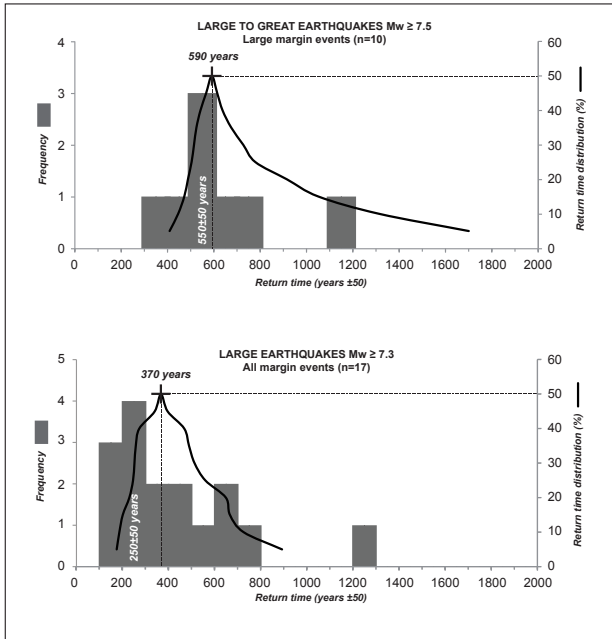


Figure 9

A - Recurrence interval of earthquakes during the Late Holocene



B - Recurrence interval of earthquakes during the Late Pleistocene - Early Holocene

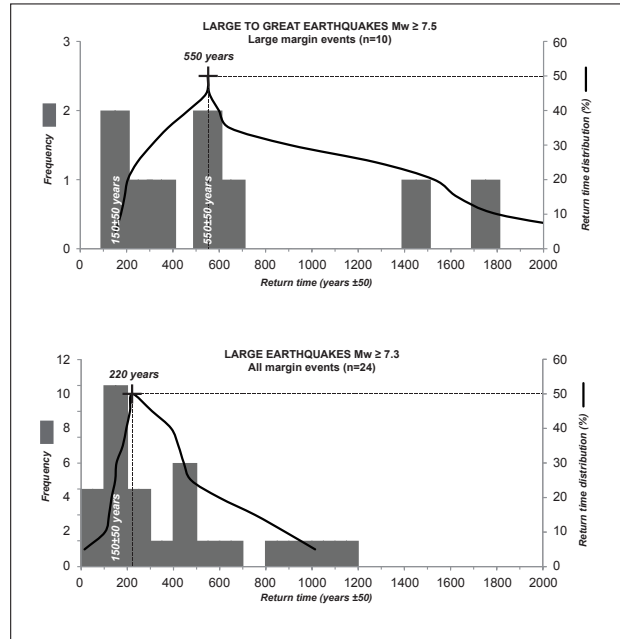


Figure 10

Lithotype	Texture	Colour	Grain size (microns)	Thickness (cm)	Composition (sand fraction)	Depositional process
Hemipelagite	Homogenised, heavily bioturbated silty-clay	Olive-grey	<10	0.5-> 50 cm	Volcaniclastic grains, quartz, planktonic organisms	Marine sedimentation
Tephra	Graded silts	Pinkish	10-20	<10 cm	Volcaniclastic debris (glass shards and pumiceous lapilli)	Airfall volcanic ash
Debrite	Chaotic silty-clay with sand to pebble size particules or deformed stratified lithoclasts	Dark olive-grey	NA	<35 cm	Quartz, volcaniclastic debris, bivalve and gasteropod shells, clasts of laminated silty-clays	Debris flow
Turbidite	Sandy to silty base grading up to silty-clay	Dark olive-grey	10-> 200	0.5-75 cm	Volcaniclastic debris, light minerals, rock fragments, foraminifers, shell fragments	Turbidity current
<i>1 Muddy turbidite</i>	<i>Thin silt lamina overlain by massive silty-clay</i>	<i>Dark olive-grey</i>	<i>10-20</i>	<i>1-40 cm</i>	<i>Quartz, foraminifers</i>	<i>Very low density turbidity current</i>
<i>2 Silt laminae turbidite</i>	<i>Interbedded, thinning and fining up clay and silt laminae</i>	<i>Dark olive-grey</i>	<i>10-20</i>	<i>1-40 cm</i>	<i>Quartz, volcanoclastic debris</i>	<i>Low density turbidity current</i>
<i>3 Silty turbidites</i>	<i>Fining upward clayey silt sequence</i>	<i>Dark olive-grey</i>	<i>10-< 100</i>	<i>0.5-55 cm</i>	<i>Volcaniclastic debris, quartz, foraminifers, micas</i>	<i>low to medium density turbidity current</i>
<i>4 Sandy turbidites</i>	<i>Sand base grading up to silty-clay</i>	<i>Dark olive-grey</i>	<i>10-> 100</i>	<i>1-75 cm</i>	<i>Quartz, volcaniclastic debris, rock fragments, micas, heavy minerals</i>	<i>Medium density turbidity current</i>
<i>5 Hyperpynite</i>	<i>Basal reverse graded turbidite</i>	<i>Olive-grey</i>	<i>10-< 100</i>	<i>5-45 cm</i>	<i>Foraminifers, wood fragments, quartz, volcaniclastic debris</i>	<i>Hyperpynal flow</i>

Table 1

Core	Longitude		Latitude		Water depth (m)	Core length (m) *	Composition		Number of gravity-flow deposits	Thickness of gravity-flow deposits (cm)			
	deg.	min.	deg.	min.			T	H		Min	Max	Mean	s.d.
MD 06-3002	39	7.83	178	40.31	2305	20 (12)	75%	25%	100	1.5	22	7.6	4.4
MD 06-3003	39	2.79	178	32.17	1398	12.88	77%	23%	101	1	36	9.8	6.8
MD 06-3008	38	32.12	179	32.04	3520	25.4 (19)	75%	25%	89	2	48	16	10.8
MD 06-3009	38	7.02	177	21.69	2940	20	70%	30%	77	1	76	17.9	14.9
TAN0810-2	37	50.973	178	59.201	1078	1.65	60%	40%	8	2	18	8	4.8
TAN0810-3	37	52.691	178	57.289	1090	3.2	22%	78%	6	3	20	10.3	6.6
TAN0810-6	37	48.105	179	37.228	3400	1.8	82%	18%	25	1	18.5	6.1	4.1
TAN0314-8	37	16.088	178	32.29	2034	2.51	63%	37%	20	2	11.5	7	2.9
TAN0810-9	37	24.5876	178	43.108	1180	3.2	12%	88%	5	2	16	6.8	4.8
TAN0810-10	37	24.77	178	41.799	1159	2.2	8%	92%	6	1	8	3	2.5
TAN0810-11	37	26.1094	178	45.8894	1089	2.6	12%	88%	8	1.5	6	3.1	1.4
TAN0810-12	37	23.382	178	42.851	1255	2.75	11%	89%	8	0.5	6	3.3	2
TAN0810-13	37	23.55	178	44.054	1167	2.5	7%	93%	4	1	10	3.9	3.7

*: 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

Re-entrant	Core	Core depth (m)	Gamma density (g/cm ²)		Magnetic susceptibility (SI)		P-Wave velocity (m/s)	
			H	T	H	T	H	T
Ruatoria	MD06-3009	2940	1.4 - 1.6	1.4 - 2	10 - 25	15 - 100	1300 - 1500	1350 - 1700
	MD06-3008	3520	1.1 - 1.4	1.2-1.7	10 - 25	15 - 100	1650 - 1700	1775 - 1950
	Tan0810-6	3400	1.7 - 1.8	1.7 - 2	10 - 25	15 - 100	1650 - 1700	1700 - 1800
Matakaoa	Tan0810-10 to -	1168±82	1.6 - 1.8	1.6 - 2	10	10 - 100	1225 - 1425	1400 - 1550
	Tan0314-8	2034	1.8	1.8 - 2.1	15	15 - 50	1225	1225 - 1375
Poverty	MD06-3003	1398	1.8	1.8 - 2.2	10	10 - 100	1300	1300 - 1500
	MD06-3002	2305	1.8	1.8 - 2.2	60	60 - 120	1400	1400 - 1600

H: hemipelgite; T: turbidites

Table 3

Table 4

MARGIN EVENTS			<i>Poverty re-entrant</i>				<i>Ruatoria re-entrant</i>						<i>Matakoa re-entrant</i>												
EVENT	mean age	2σ range	EVENT	mean age	2σ range	CORE		EVENT	mean age	2σ range	CORE				EVENT	mean age	2σ range	CORE							
name			name			MD3003	MD3002	name			Tan0810-2	Tan0810-3	MD06-3009	MD06-3008	Tan0810-6	name			Tan0810-9	Tan0810-10	Tan0810-11	Tan0810-12	Tan0810-13	Tan0314-8	
H1	387	170						R1	387	253					T1	T1							T2		
H2	616	26						R2	789	199	T1				T2	T2	M3	941	212					T3	T3
H3	978	32	P1	819	191	T1		R3	1040	94	T2			T1	T2	T3								T3	T3
H4	1508	13	P2	1388	132	T2		R4	1566	71	T3				T3	T3							T3		
H5 **	1711	6	P3	1699	38	T3		R5	1711	6	T4	T1		T5	T4	T4	M4	1716	11	T1	T3		T4	T4	T3
H6	2286	130	P7	2426	270	T6		R6	2185	232	T5			T6	T5			2255	245						T6
H7	2935	122	P9	2880	212	T8		R7	2935	122				T7	T7	T7	M7	3657	552	T3	T4				
H8	3186	81	P11	3060	206	T10		R8	3201	125				T8	T8	T8									
H9 **	3419	11	P13	3438	30	T12		R9	3391	39		T2			T9	T9									
H10	3438	30	P14	3438	30	T12		R10	3462	69				T8	T9										
H11	4672	27	P17	4357	342	T15		R11	4821	176		T3			T10	T11									
H12	5344	72	P18	5409	137	T16		R12	5132	326					T11	T12	M9	5127	289		T6				T1
H13	5525	60	P19	5535	77	T17		R13	5525	60		T4	T9-10		T12										
H14	6021	193	P23	6021	193	T22	T1	R14	6072	445					T13	T13									
H15	6489	135	P24	6489	135	T23	T2		<i>6303</i>	<i>583</i>					<i>T15</i>										
H16	6644	179	P25	6644	179	T24	T3	R15	6668	248				T11	T16										
H17	6836	210	P26	6836	210	T25	T4		<i>6926</i>	<i>311</i>					<i>T17</i>										
H18	7108	159	P27	7039	228	T26	T5	R16	7221	272				T12	T19										
H19	7480	116	P28	7480	116	T27	T8	R17	7570	269				T13	T21										
H20	8390	157	P30	8390	157	T29	T19		<i>8624</i>	<i>612</i>					<i>T24</i>										
H21	8603	262	P33	8604	263	T32	T22		<i>8800</i>	<i>654</i>					<i>T25</i>										
H22 *	9128	101	P34	9067	161	T33	T24	R18	9240	213				T14	T27										
H23 *	9505	25	P37	9505	25	T36	T27	R19	9617	260				T15	T28	M12	9325	750							T4
H24	9706	143	P38	9706	143	T37	T28		<i>9839</i>	<i>393</i>					<i>T29</i>										
H25	10129	103	P40	10155	129	T39	T31		<i>10010</i>	<i>221</i>					<i>T30</i>										
H26	10129	103	P42	10169	143	T39	T32	R20	10073	159				T16	T31										
H27	10686	113	P45	10579	220	T40	T35	R21	10767	194				T17	T32										
H28 *	11340	118	P48	11532	310	T41	T39	R22	11256	201				T18	T33										
H29	11719	289	P49	11659	348	T42	T40		<i>11896</i>	<i>466</i>					<i>T35</i>										
H30	12169	291	P50	12081	378	T43	T42	R23	12189	311				T19	T37	M13	12290	970							T5
H31	12518	242	P51	12518	242	T44	T43	R24	12481	365				T20	T38										
H32	12698	264	P52	12698	264	T45	T44	R25	12628	391				T21	T40										
H33	13357	238	P54	13357	238	T48	T46	R26	13348	330				T22	T42	M14	13564	814							T6
H34	13487	210	P55	13490	213	T49	T47	R27	13377	320				T23	T43										
H35	13698	143	P56	13736	181	T50	T53	R28	13678	162				T24	T47	M15	14338	594							T7
H36	13831	187	P57	13831	187	T51	T54		<i>14233</i>	<i>717</i>					<i>T48</i>										
H37	14661	271	P58	14685	295	T56	T55		<i>14299</i>	<i>783</i>					<i>T49</i>										
H38	14661	271	P59	14685	295	T56	T56		<i>14365</i>	<i>823</i>					<i>T50</i>										
H39	14890	295	P60	14890	295	T58	T57		<i>14728</i>	<i>901</i>					<i>T53</i>										
H40	15004	307	P61	14993	317	T60	T58		<i>15163</i>	<i>466</i>					<i>T55</i>										
H41	15404	224	P62	15549	369	T62	T59	R29	15350	278				T25	T58	M17	15016	698							T9
H42	15849	388	P63	15849	388	T63	T60		<i>15865</i>	<i>653</i>					<i>T61</i>										
H43	15983	357	P64	15948	393	T64	T60	R30	15935	583				T26	T62	M18	16404	778							T10
H44	16140	398	P65	16140	398	T65	T60		<i>16050</i>	<i>698</i>					<i>T63</i>										
H45	16292	295	P66	16292	295	T66	T61		<i>16056</i>	<i>704</i>					<i>T64</i>										
H46	16446	314	P66	16451	319	T67	T62		<i>16056</i>	<i>704</i>					<i>T64</i>										

*, ** are margin events related to catastrophic floods or volcanism (after Poudroux et al., 2012a)
 Isolated events are in italic grey

All margin events (n=41)

PGA	<i>Cousins et al. (1999) - Eq. (1)</i>			<i>Si and Midoriwaka (1999) - Eq. (2)</i>		
	Faults *	RI **	Mw range	Faults *	RI **	Mw range
0.08	2; 6; 7	390-460 years	7.3-8.4	2; 6; 7	390-460 years	7.3-8.4
0.1	2; 6	890-1235 years	7.3-8.4	2; 6; 7	390-460 years	7.3-8.4
0.15	2	1300-1670 years	8.2-8.4	2	1300-1670 years	8.2-8.4

*: Numbers correspond to faults detailed in Table SM1; ** Recurrence intervals calculated for the duration of our turbidite record (16,060 years, from 390 to 16,450 yr BP)
Average recurrence interval deduced from the turbidite record : 400 years

Large margin events (n=20)

PGA	<i>Cousins et al. (1999) - Eq. (1)</i>			<i>Si and Midoriwaka (1999) - Eq. (2)</i>		
	Faults *	RI **	Mw range	Faults *	RI **	Mw range
0.08	2	1300-1670 years	8.2-8.4	2; 6	890-1235 years	7.3-8.4
0.1	∅	∅	∅	2	1300-1670 years	8.2-8.4
0.15	∅	∅	∅	2	1300-1670 years	8.2-8.4

*: Numbers correspond to faults detailed in Table SM; ** Recurrence intervals calculated for the duration of our turbidite record (16,060 years, from 390 to 16,450 yr BP)
Average recurrence interval deduced from the turbidite record : 800 years

Table 5