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1 **The relationship between eruptive activity, flank collapse and sea-level at volcanic**
2 **islands: a long-term (>1 Ma) record offshore Montserrat, Lesser Antilles**

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93 **Key Points**

94 Heightened volcanic activity on Montserrat at 120-190 ka, 760-810 ka, and 900-930 ka

95 Large landslides coincide with rapid sea-level rise at island arc volcanoes

96

97

Abstract

98

99 Hole U1395B, drilled southeast of Montserrat during Integrated Ocean Drilling Program
100 Expedition 340, provides a long (>1 Ma) and detailed record of eruptive and mass-
101 wasting events (>130 discrete events). This record can be used to explore the temporal
102 evolution in volcanic activity and landslides at an arc volcano. Analysis of tephra fall and
103 volcanoclastic turbidite deposits in the drill cores reveals three heightened periods of
104 volcanic activity on the island of Montserrat (~930 ka to ~900 ka, ~810 ka to ~760 ka,
105 and ~190 ka to ~120 ka) that coincide with periods of increased volcano instability and
106 mass-wasting. The youngest of these periods marks the peak in activity at the Soufrière
107 Hills volcano. The largest flank collapse of this volcano (~130 ka) occurred towards the
108 end of this period, and two younger landslides also occurred during a period of relatively
109 elevated volcanism. These three landslides represent the only large (>0.3 km³) flank
110 collapses of the Soufrière Hills edifice, and their timing also coincides with periods of
111 rapid sea-level rise (>5 m/ka). Available age data from other island arc volcanoes
112 suggests a general correlation between the timing of large landslides and periods of rapid
113 sea-level rise, but this is not observed for volcanoes in intra-plate ocean settings. We thus
114 infer that rapid sea-level rise may modulate the timing of collapse at island arc volcanoes,
115 but not in larger ocean-island settings.

116

117 **Key words**

118 Landslide, volcanism, sea-level, IODP, Expedition 340

119

120 **1. Introduction**

121

122 Volcanic islands in arc settings grow and decay through eruptive and mass-wasting
123 processes. The rate of island growth reflects the balance of these processes, including the

124 style, composition and magnitude of individual eruptions [Houghton et al., 1995; Singer
125 et al., 2008; Germa et al., 2010], and variations in the magnitude and frequency of mass-
126 wasting events (e.g. lava dome collapses, flank landslides) [Ablay and Marti, 2000; Cole
127 et al., 2002; Trofimovs et al., 2013]. As a volcanic island grows, the edifice becomes
128 increasingly unstable, resulting in partial flank collapses, which are a ubiquitous feature
129 of composite volcanoes (e.g. Siebert, 1984). Such collapse events are potential geohazards
130 through the generation of landslides [Siebert, 1984; Watt, et al., 2012a, 2012b], and their
131 potential to generate tsunamis [Ward and Day, 2003]. There are few individual volcanic
132 records that span the life cycle of individual composite volcanoes (on the order of 10^5 to
133 10^6 years), but such records can be used to investigate temporal patterns in total volcanic
134 output, and the relative timing of flank collapses. The instabilities that drive collapses
135 may relate both to internal (e.g. total volcanic output) and external (e.g. eustatic sea-level
136 change) processes. To obtain a more complete understanding of the overall controls on
137 volcano growth and destruction, we use a marine stratigraphic record offshore Montserrat
138 to explore patterns of volcanism, sea-level change and flank collapse timing over a 10^6
139 year period.

140

141 Here, we use the term flank collapse to refer to gravity-driven failures of volcanic
142 material from subaerial and/or submarine volcanic island flanks, potentially also
143 involving carbonate shelf material. Collapses that only involve material from the
144 carbonate shelf (i.e. generating bioclastic deposits) are referred to as shelf collapses. Flank
145 collapses can involve several cubic kilometres of material, and may or may not be
146 associated with volcanic eruptions. They are distinct from the generally smaller dome
147 collapses, which involve juvenile lava and are a common mass wasting process during the
148 lava-dome forming eruptions that typify volcanism on Montserrat. *Mass-wasting* or

149 *landslides* are used as collective terms here for the different types of collapse events.
150 Landslides may also generate a range of density currents that are represented in the
151 geological record as a variety of density current deposits. In this study we refer to all
152 submarine density current deposits as *turbidites*, regardless of whether the deposits were
153 generated from fully turbulent or non-turbulent flows.
154
155 Triggers of flank or shelf collapses on volcanic islands are poorly understood, but it has
156 been proposed that their frequency is related to eustatic sea-level changes, edifice growth,
157 tectonic activity, and rainfall [McGuire et al., 1997; Masson et al., 2006; Sato et al., 2007;
158 Marques et al., 2008; Quidelleur et al., 2008; Hunt et al., 2013, 2014]. Reconstructing pre-
159 historical earthquake and rainfall records are challenging, resulting in difficulties when
160 comparing these potential triggers to landslide occurrence. Previous studies of eustatic
161 sea-level change and landslides have been hampered due to difficulties in acquiring
162 sufficient and accurate dates for landslide events. Such studies have typically relied on
163 incomplete information from on-land observations of collapse structures, and/or been
164 restricted to relatively recent events recorded in shallow (<6 m) marine sediment cores
165 and seismic profiles [McMurtry et al., 2004a, 2004b; Boudon et al., 2007; Trofimovs et
166 al., 2013]. Consequently, very few volcanic island landslide events have been dated
167 precisely over time periods that are long enough to include multiple climatic, volcanic,
168 and tectonic cycles [Longpré et al., 2011; Trofimovs et al., 2013; Hunt et al., 2013].
169
170 We analysed Hole U1395B from IODP Expedition 340, drilled in 2012. This Hole is
171 >120 m in length and contains an unusually long (>1 Ma) and detailed marine record of
172 island arc volcanism and mass-wasting activity. Hole U1395B was drilled ~25 km
173 southeast of Montserrat at ~1200 m below sea-level (Figure 1), and provides an excellent

174 opportunity to study a long and detailed record of sedimentological processes around an
175 island arc volcano.

176

177 The record of marine volcanoclastic deposits (both turbidites and tephra fall deposits) is
178 used as a proxy for volcanic activity at Montserrat. This approach is based on studies of
179 marine deposits associated with the 1995-2010 eruption on Montserrat [Kokelaar, 2002;
180 Trofimovs et al., 2006], which showed that individual pyroclastic flows formed via dome-
181 collapse and explosive eruption events produced widespread volcanoclastic turbidites. We
182 thus expect most volcanoclastic turbidites to represent individual eruptions, but note that
183 some volcanoclastic turbidites may be produced by non-eruptive flank collapses, or by
184 reworking of older volcanoclastic deposits. Turbidites from large flank collapses can be
185 identified by their thickness and correlation with debris avalanche deposits (and there are
186 only a few of these around Montserrat; cf. Lebas et al., 2011), while resedimented
187 turbidites may be more mixed, including bioclastic material. The latter may also still
188 provide a broad proxy for volcanic activity, since they are more likely to be generated
189 during periods of volcanism, when flanks may be destabilised by eruptive activity and
190 deposition of volcanic products [Collins and Dune et al., 1986; Major et al., 2000; Hunt et
191 al., 2014].

192

193 By identifying and dating volcanoclastic deposits from Hole U1395B, we aim to
194 investigate the temporal relationships between mass-wasting, volcanic activity, and sea-
195 level over an interval of $\sim 10^6$ years. This period encompasses a sufficient number of
196 volcanic cycles and eustatic sea-level changes to allow for statistically robust hypothesis
197 testing. It is also shown that all of the well-dated major collapse events around Montserrat
198 occurred during periods of rapid sea-level rise. We explore whether other volcanic islands

199 (island arcs and intraplate ocean islands) show similar relationships between landslide
200 ages and sea level or volcanism, whilst noting that many of these landslide ages have
201 considerable uncertainties.

202

203

204 **2. Study Area: Introduction to Montserrat**

205

206 Montserrat is an island arc volcano located in the Lesser Antilles (Figure 1), which
207 erupted between 1995-2010, devastating the city of Plymouth and affecting large parts of
208 the island economy [Kokelaar et al., 2002; Wadge et al., 2014]. The eruptions included
209 collapse of the active lava dome generating pyroclastic flows and block and ash flows.

210 The largest dome collapse (0.21 km³) occurred in 2003 and generated a 0.5-1 m high
211 tsunami on Guadeloupe [Herd et al., 2006; Trofimovs et al., 2008]. Bathymetric mapping
212 has since identified much larger landslide deposits offshore Montserrat with volumes of
213 0.3-20 km³ [Deplus et al., 2001; Boudon et al., 2007; Lebas et al., 2011; Watt et al., 2012a,
214 2012b]. Such large landslides have a much higher tsunamigenic potential, and thus
215 represent a more significant hazard than events associated with the recent eruptions.

216

217 Since ~290 ka (based on subaerial Ar-Ar dates; Harford et al., [2002]), activity at
218 Montserrat has been focussed on the andesitic Soufrière Hills volcano, except for a brief
219 interlude of basaltic volcanism at ~130 ka, forming the South Soufrière Hills (Figure 1).
220 The previously active volcanic centre, at Centre Hills, is dated at 990-550 ka [Harford et
221 al., 2002]. Between these two periods of activity (550 ka to 290 ka) the subaerial record
222 suggests a period of quiescence. The apparent gap in volcanism from 550 ka to 290 ka is
223 not clear in the marine sediment cores, and the existing subaerial ages may thus reflect a

224 limited stratigraphy represented by on-land exposures, or incomplete study and dating of
225 subaerial outcrops.

226

227 The offshore eruption and landslide record around Montserrat has been studied in detail
228 [Deplus et al., 2001; Trofimovs et al., 2006, 2010, 2012, 2013; Boudon et al., 2007; Lebas
229 et al., 2011; Le Friant et al., 2010, 2015; Cassidy et al., 2012b, 2013]. To the southeast
230 and southwest of Montserrat is an extensive shallow core (<6 m) data set (Figure 1).

231 These >80 cores contain hemipelagic sediment and numerous volcanoclastic and bioclastic
232 turbidites from eruption and landslide deposits emplaced during the last ~110 ka [Le
233 Friant et al., 2009; Trofimovs et al., 2010, 2012, 2013; Cassidy et al., 2012b, 2013].

234 Swath bathymetry and 2D and 3D seismic data reveal seven large landslide deposits,
235 dating back to the time of Centre Hills volcanism (Figure 1). These large landslide
236 deposits likely represent flank collapse events as opposed to dome or localised shelf
237 collapses.

238

239 It is important to recognise that any study based on a single sampling site will introduce
240 some sampling bias in the record of eruptive and collapse events. The activity at

241 Montserrat is typified by dome-forming eruptions with durations of months to years,

242 interspersed by moderate-sized explosive pulses such as the 1995-2010 eruption

243 [Kokelaar, 2002; Wadge et al., 2014]. Such eruptions form complex volcanoclastic

244 intervals generated by multiple pyroclastic density currents and tephra fall events

245 distributed radially from the volcanic edifice [Trofimovs et al., 2006, 2013; Le Friant et

246 al., 2015]. Hole U1395B is located ~25 km southeast of Montserrat, hence it is only likely

247 to sample mass flows that entered the ocean to the east and south of Montserrat. While

248 some individual events may therefore not be preserved within Hole U1395B, periods of

249 eruption are still likely to be represented, because any one eruption is likely to produce
250 multiple mass flow and tephra fall deposits, with the mass flows, in particular, travelling
251 in a range of directions. For example, the 2003 collapse event (0.1 km^3) deposited a ~20
252 cm thick volcanoclastic turbidite at site U1395B [Trofimovs et al., 2008], but small
253 vulcanian eruptions that occurred throughout the 1995-2010 eruption are not represented
254 within Hole U1395B [Kokelaar, 2002; Wadge et al., 2014].

255

256 Hole U1395B is only likely to sample a small proportion of tephra fall deposits, as
257 dispersal depends on wind direction and the magnitude of the explosive event. Prevailing
258 wind directions in the troposphere and upper stratosphere are predominantly from the east,
259 and prevailing wind directions in the lower stratosphere are from the west, based on
260 historical data from 1956 to present [Radiosonde wind dataset on Guadeloupe]. Tephra
261 fall deposits may therefore be subject to a spatial sampling bias, with many eruption
262 plumes transported to the west leaving no record within U1395B. However, large
263 explosive eruptions, which can involve multiple phases and be associated with pyroclastic
264 density currents and the generation of offshore turbidites, are still likely to have some
265 representation at the core site as an eruptive event. Because we are interested simply in
266 event timing, rather than magnitude or style, this single core is likely to provide a
267 relatively comprehensive record of eruptions at Montserrat.

268

269

270 **3. Methods**

271

272 *3.1. Event deposit identification*

273 We defined five facies at Site U1395: hemipelagic mud, bioclastic turbidites, mixed
274 bioclastic-volcaniclastic turbidites, volcanoclastic turbidites, and tephra fall deposits
275 (Figure 2). Hemipelagic mud is mostly composed of carbonate and detrital clay with
276 abundant interspersed foraminifera. The volcanoclastic turbidite and tephra fall deposits
277 are less easily distinguished from each other, as both can comprise normally-graded sand
278 and silt [Trofimovs et al., 2013]. Their discrimination requires grain size and component
279 analysis to identify the type of event deposit [Cassidy et al., 2014, 2015].

280

281 Tephra fall layers were defined in this study as having <30% bioclasts, a Folk and Ward
282 [1957] sorting coefficient of <0.5 phi (where grainsize (phi) = $-\log_2(\text{grainsize (mm)})$)
283 (analyses conducted on particles from -1 to 9 phi), and a thickness of <20 cm. Tephra fall
284 deposits are well sorted by density and dominated by volcanic clasts. Hole U1395B is ~25
285 km away from Montserrat, where tephra fall deposits from moderate-sized explosive
286 eruptions, even downwind, are likely to be a few centimetres thick; consequently, thick
287 (>20 cm) deposits are unlikely to represent tephra fall layers. Large magnitude eruptions
288 are rare within the Lesser Antilles [Palmer et al., 2016]. Thin tephra fall layers are
289 commonly mixed with surrounding hemipelagic material, through bioturbation, bottom
290 current reworking, or disturbance during coring, thus artificially increasing their apparent
291 bioclast content. We thus examined the core for any evidence of post-depositional
292 processes (Supplementary figure S1) and have defined tephra fall layers as comprising <
293 30% bioclast content to allow for post-depositional mixing. Volcanoclastic turbidites are
294 deposits from high-energy, erosive flows that may entrain bioclastic material and pre-
295 existing volcanoclastic sediments. Therefore, although some parts of a volcanoclastic
296 turbidite can be well-sorted, they are likely to be less well-sorted than tephra fall deposits
297 [Cassidy et al., 2014]. Volcanoclastic turbidites are defined here as comprising <30%

298 bioclasts, with a Folk and Ward [1957] sorting coefficient >0.5 (ϕ). Mixed turbidites are
299 defined as having 30-70% bioclasts. Bioclastic turbidites are defined as containing $>70\%$
300 bioclasts (see Supplementary Figures S1-S3).

301

302 Grain size measurements using laser-diffraction analyses were carried out using a
303 Malvern Master-sizer 2000 particle size analyser, which can measure grain sizes between
304 0.2–2000 μm . To disperse grains, 25 ml of reverse osmosis water with 0.05% sodium
305 hexametaphosphate dispersant was added to 1 cm^3 of sample and left overnight on a
306 shaking table. Samples were analysed in triplicate and accuracy was monitored using
307 standard size particles (32 and 125 μm) (see Hunt et al., [2013] for details). Componentry
308 analysis was conducted on sieved fractions >63 μm and <250 μm material from all
309 volcanic-rich units and some bioclastic-rich units. For each sample, approximately 400
310 grains were point-counted using an area counting method. Componentry classes follow Le
311 Friant et al., [2008] and Cassidy et al., [2014]: 1) vesicular pumice clasts; 2) non-vesicular
312 andesite; 3) altered lithic clasts; 4) crystal and glass fragments; 5) mafic scoria clasts; and
313 6) bioclasts. Grainsize and componentry data are summarized in Table S1 and Table S2,
314 respectively (see Supplementary Figures S1-S3 for photos).

315

316 *3.2. Dating Hole U1395B*

317 The core from U1395B core was dated using a combination of oxygen isotope
318 stratigraphy, biostratigraphy, AMS radiocarbon dating, and the shipboard paleomagnetic
319 reversal records. Higher resolution dating was carried out on the upper 40 m of Hole
320 U1395B using oxygen isotope stratigraphy of the hemipelagic mud (Figure 2). Twenty
321 *Globigerinoides ruber* specimens between 250-355 μm in size were picked and analysed
322 from each hemipelagic sample. Samples were 7 cm apart and analysed at Plymouth

323 University on an Isoprime Instruments continuous flow mass spectrometer with a Gilson
324 Multiflow carbonate auto-sampler. Oxygen Isotope values are given as deviations in the
325 isotope ratios ($^{18}\text{O}/^{16}\text{O}$) per mil (‰), using the VPDB scale (Table S3).

326

327 To limit the ambiguity of identifying marine isotope stages, biostratigraphic boundaries
328 and AMS radiocarbon dates (for sediments <50 ka) were used (Table 1).

329 Calcareous nannofossils in the <63 μm material from hemipelagic samples were analysed
330 using scanning electron microscopy (SEM), employing the calcareous nannofossil
331 zonation of Kameo and Bralower [2000] for the Caribbean Sea. Sediment was fixed to
332 metal stubs using a thin layer of spray adhesive, then sputter-coated with gold. The first
333 occurrence of *Emiliana huxleyi* (250 ka) was found at 36.74 m, close to the MIS 7/8
334 boundary (243 ka). The first occurrence of *E. huxleyi* has also been identified across the
335 MIS 7/8 transition in Hole U1396C [Wall-Palmer et al., 2014] and CAR-MON 2 [Le
336 Friant et al., 2008] (Figure 3).

337

338 New AMS dates were obtained in this study from four samples in the upper 4 m of Hole
339 U1395B (aged <57 ka), in addition to the radiocarbon AMS dates reported by Trofimovs
340 et al., [2013]. Approximately 1000 pristine tests of white *Globgerinoides ruber* >150 μm
341 in size were picked (~17 mg) and then sonically cleaned. The new samples were located
342 beneath two of the largest turbidites (Figure 3). Radiocarbon dates were measured at
343 Scottish Universities Environmental Research Council (SUERC) using their in-house
344 protocol [see Trofimovs et al., 2013].

345

346 Paleomagnetic reversals were determined on board during Expedition 340 as 180°
347 changes in declination (after azimuthal correction). Associated changes in inclination after

348 demagnetization of the natural remnant magnetization (NRM) in a field of 20 mT (to
349 remove the coring overprint) were also recorded (see Hatfield et al., [2013] for details).
350 Here we report ages based on the geomagnetic polarity timescale (GPTS) of Ogg et al.,
351 [2012] instead of the GPTS of Cande and Kent [1995] as was reported on board the ship.
352 Two paleomagnetic reversals occur at 781 ka (6.3% error) and 988 ka (11.3% error) in
353 core U1395B [Cande and Kent 1992a, 1992b; Ogg et al., 2012], with a possible third
354 reversal at the base of the core at 1072 ka (11.3% error) [Cande and Kent 1992; Ogg et al.,
355 2012]. The 781 ka reversal is obscured by a volcanoclastic turbidite and coring
356 disturbance at 68.5-71 m (Figure 2). The coring disturbance occurs primarily within the
357 volcanoclastic turbidite. The reversal is likely to have occurred shortly prior to the
358 emplacement of the volcanoclastic turbidite, thus obscuring the reversal through erosion of
359 hemipleagic mud. Here we take the base of the volcanoclastic turbidite as 781 ka. The
360 MIS 8/9 (300 ka) boundary is interpreted to occur at ~44 m, suggesting that sedimentation
361 rates decrease with depth in the core. The 988 ka reversal occurred between 88.3-89.9m
362 with a mid-point at 89.1 m (Figure 2). The 1072 ka reversal may be present at the base of
363 Hole U1395B (Figure 2) but this is less certain due to poor sample recovery at the base of
364 the core.

365

366 Data from Hole U1395B were also compared and correlated to previously studied cores
367 around Montserrat (Figure 3). These cores include JR123-5V, JR123- 6V, CAR-MON 2,
368 and Hole U1396C (Figure 3). JR123-5V and JR123-6V are part of an extensive vibrocore
369 data set that has been used to compile a comprehensive stratigraphy of Montserrat over
370 the past 110 ka [Trofimovs et al., 2013]. The vibrocore data set includes over 80 <6 m
371 long cores with 40 accelerator mass spectrometry (AMS) radiocarbon dates (Figure 3).
372 JR123-5V and JR123-6V are located 1-2 km north of Site U1395, and some units within

373 core U1395B can be correlated to units found in JR123-5V and JR123-6V by age (Figure
374 3). CAR-MON 2 is a piston core collected in 2002, which extends 5.75 m and was taken
375 ~55 km to the southwest of Montserrat (Figure 3) [Le Friant et al., 2008]. Site U1396C is
376 situated ~33 km southwest of Montserrat (Figure 3) and was collected during IODP
377 Expedition 340 [Wall-Palmer et al., 2014].

378

379

380 **4. Results**

381

382 Hole U1395B is 127.51 m long and it is composed of 62.6% hemipelagic mud, 24.4%
383 volcanoclastic deposits, 9% mixed turbidites and 4% bioclastic deposits, by deposit
384 thickness. The core comprises 18 bioclastic turbidites, 48 tephra fall deposits, 26 mixed
385 turbidites, and 41 volcanoclastic turbidites. Core recovery is good (>90%), with only one
386 occurrence of basal flow-in coring disturbance over the studied core length at the bottom
387 of Core U1395B-2H, followed by probable fall-in at the top of Core Section U1395B-3H
388 [Jutzeler et al., 2014].

389

390 *4.1. Age Models*

391 Unit ages have been estimated by calculating hemipelagic sedimentation rates between
392 dated horizons, assuming constant sedimentation rates between dated horizons.

393 Developing accurate age models for marine cores is difficult due to the effects of erosion
394 (commonly at the base of turbidites), short-term fluctuations in sediment supply, and the
395 sometimes ambiguous identification of marine isotope stages. We therefore include three
396 age models to help capture these uncertainties, and their implications.

397

398 Age models can be affected by erosion. In the upper 5 m of U1395B ~88 cm of
399 hemipelagic sediment may have been eroded by the 12-14 ka turbidite [Trofimovs et al.,
400 2013], indicating that the effects of erosion at Site U1395B may be significant. Erosion
401 rates in the upper 10 m of U1395B are well constrained by a combination of AMS dates
402 from this study, and correlation with well-dated units (40 AMS dates) in the shallow
403 vibrocore dataset [Trofimovs et al., 2013]. Below the threshold for AMS radiocarbon
404 dating (>10 m in Hole U1395B) the effects of erosion by turbidites cannot be accurately
405 constrained at site U1395B, potentially leading to inaccuracies in age models. It is likely
406 that most turbidity currents are erosive, removing underlying hemipelagic mud and event
407 deposits from the stratigraphy resulting in the underestimation of the true sedimentation
408 rate. We expect this to be a systematic error affecting the whole core, however, rather
409 than something that introduces bias to specific time periods.

410

411 Dating cores using oxygen isotopes may also lead to age model inaccuracies due to
412 difficulties in identifying marine isotope stages (MIS). Oxygen isotopes from Hole
413 U1395B have been compared to the Lisiecki and Raymo, [2005] curve; but, local factors
414 may affect the magnitude of isotope fluctuations and erosion may remove parts of the
415 isotope record, resulting in the misidentification of MIS boundaries. Any inaccuracies in
416 the age models used will affect the reliability of unit ages assigned to individual events,
417 and thus affect subsequent analysis of the dataset.

418

419 Assuming that the average total erosion beneath turbidites is on the order of centimetres,
420 and assuming that MIS boundaries have been correctly constrained within a few
421 centimetres, the event dating errors are likely to be on the order of 10^2 - 10^5 years. This is
422 of a similar magnitude to stratigraphic gaps identified in other cores around Montserrat,

423 including instances where >30 ka of stratigraphy was removed by turbidite erosion (JR
424 123-5-V and Hole U1396C) (Trofimovs et al., 2013; Wall-Palmer et al., 2014). In order to
425 assess how sensitive our event-frequency analysis is to dating errors of this magnitude, we
426 conduct the same analysis with three different age models. Age Model 1, described below,
427 uses all age constraints (i.e. MIS boundaries, palaeomagnetic reversals etc.) but is also
428 potentially more susceptible to the effects of erosion, producing a record with apparent
429 fluctuations in sedimentation rate. Age Models 2 and 3 use fewer age constraints,
430 resulting in a smoother estimate of long-term sedimentation rate that may be more
431 geologically realistic. Individual unit ages derived from the three age models may vary by
432 up to 10^5 years. By using all three age models within subsequent analyses, we can test
433 how robust our results are to these uncertainties.

434

435 **Age Model 1:** This model uses 11 dated horizons, including horizons derived from
436 correlations to units described in Trofimovs et al. [2013], identification of MIS
437 boundaries, and identification of paleomagnetic reversals. The 2-1.5 ka, 6 ka, 14 ka, 74-59
438 ka, 110-103 ka, and 130 ka deposits from Trofimovs et al. [2013] and Cassidy et al.
439 [2013] have been correlated to units in U1395B. Unit ages of 1.75 ka, 6 ka, 14 ka, 66.5 ka,
440 107 ka, and 130 ka in Hole U1395B at core depths of 0.49 m, 2.52 m, 3.98 m, 6.26 m,
441 10.47 m, and 18.49 m respectively were used in age model 1. Using oxygen isotope
442 analysis 3 MIS boundaries were identified and used. These are MIS boundaries 6/7 (191
443 ka), 7/8 (243 ka) and 8/9 (300 ka) at core depths of 28.31 m, 35.93 m, and 43.84 m
444 respectively [Lisiecki and Raymo. 2005]. Finally, two paleomagnetic reversal dates of
445 781 ka, and 988 ka, at depths of 63.06 m, and 89.08 m are used in age model 1.

446

447 Using Age Model 1, the calculated sedimentation rates in the upper 300 ka of Hole
448 U1395B fluctuate between 2 to 17 cm/ka, and encompass the range of hemipelagic
449 sedimentation rates (5 to 10 cm/ka) previously determined in the area [Reid et al., 1996;
450 Watt et al., 2012b]. The greater variability in apparent accumulation rates determined
451 from Age Model 1 (particularly for the lower sedimentation rates) likely reflects the
452 effects of turbidite erosion. Sedimentation rates in age model 1 between 300-988 ka are
453 relatively constant at 3.2-9 cm/ka. The reduction in variability of sedimentation rates
454 further down the core likely reflects the fact that there are fewer dated horizons deeper in
455 Hole U1395B, resulting in a smoother apparent sedimentation rate.

456

457 **Age Models 2 and 3:** Age Model 2 uses Deposit 2 as a boundary at 130 ka (Figure 2;
458 Cassidy et al., [2015]) at 18.49 m, and the paleomagnetic dates of 781 ka, and 988 ka, at
459 depths of 63.06 m, and 89.08 cm respectively. Age Model 3 uses only the two
460 paleomagnetic dates. Using Age Models 2 and 3, estimated sedimentation rates are less
461 variable throughout the core with long-term rates of 4-8 cm/ka.

462

463 *4.2. Event frequency offshore Montserrat*

464 Events are not distributed regularly through time in core U1395B, but appear to cluster
465 within specific time periods (Figure 4). To better understand this long-term variation in
466 event timing we conduct a moving sum, where the number of events are summed every
467 50 kyr, at 10 kyr increments. Observations do not change when using bins of 30 kyr and
468 100 kyr (see Figures S4 and S5). Conservatively, we classify periods of increased
469 frequency as periods where all three age models show that the frequency of event deposits
470 is above the mean 90% confidence interval of the moving sum data (Figure 5). Periods of
471 reduced activity are defined as periods where all three age models show that event

472 frequency is below 1 standard deviation from the mean. In all age models, these periods
473 of relative quiescence last on the order of 10^4 - 10^5 kyr. In Hole U1395B, all age models
474 show an increased frequency of all event deposits (above the 90% confidence interval of
475 8.44 events in 50 ka) between ~170 ka to ~120 ka. This peak in Montserrat's volcanism
476 lies within the period of Soufrière Hills activity (constrained from subaerial dates as 290
477 ka to present, but see discussion in Section 5.1), and suggests that the largest flank
478 collapse of Soufrière Hills (Deposit 2, at 130 ka) and the South Soufrière Hills basaltic
479 episode (~130 ka) both occurred towards the end of the most active phase of Soufrière
480 Hills' eruptive history.

481

482 When examining volcanoclastic deposits only (volcanic turbidites and tephra fall), there
483 are three periods where frequency is greater than the mean 90% confidence interval,
484 between ~930 ka to ~900 ka, ~810 ka to ~760 ka, and ~190 ka to ~120 ka. We interpret
485 periods with a higher frequency of both tephra fall and volcanoclastic turbidites as
486 representing episodes of elevated eruptive activity at Montserrat (Figure 5). The older two
487 periods coincide with activity at Centre Hills volcano, and the younger period (which is
488 similar to that derived from the whole dataset) to Soufrière Hills.

489

490 The periods of heightened volcanic activity show a general correlation with changes in the
491 style of eruptions at Montserrat (Figures 4 and 5). The ~810 to ~760 ka increase in event
492 frequency coincides with the appearance of scoria within Hole U1395B. The ~190 ka to
493 ~120 ka increase in event frequency coincides broadly with the onset of activity at South
494 Soufrière Hills [Harford et al., 2002].

495

496 To test if there is any correlation between sea-level and turbidite frequency we use a
497 linear model (LM), generalised linear model (GLM), and proportional hazards model
498 (PHM) after binning the data into 10 kyr intervals. For further details on the statistical
499 methods see Hunt et al. [2014] and Clare et al. [2016]. Unit ages are given in Table S4
500 and test results are given in Table 2.

501

502 Statistical analyses are conducted on all three age models in order to test the sensitivity of
503 results to dating errors and to determine if the results are artefacts of the specific age
504 model used. By comparing results from the three age models, the effects of age
505 uncertainty can be better understood. We pose the null hypothesis that turbidite frequency
506 is not correlated with sea-level, and P-values < 0.05 allow us to reject this hypothesis.

507 Table 2 shows that using age model 1 does not show significance in any statistical tests,
508 but using age models 2 and 3 shows significance in all three statistical tests. It should be
509 noted, however, that the correlation coefficient for LM is very low (0.002, 0.085, and
510 0.077 for age models 1-3 respectively). Low correlation coefficients for age models 2 and
511 3 indicate that while there may be a broad correlation between sea-level and turbidite
512 frequency, the variation means that the scatter about that trend is very wide. Grouping
513 together of turbidites from eruptive and collapse activity, effects of data binning, or
514 genuine natural noise, may cause this broad scatter. We therefore conclude that all three
515 age models are in agreement and show that there is little to no correlation between sea-
516 level and turbidite frequency. Dating errors are unlikely to affect this observation due to
517 the consistency of this result across all three age models.

518

519 Although there appears to be no correlation between turbidite frequency and sea-level, all
520 three of the dated larger landslide deposits from Soufrière Hills volcano (Deposits 1 and 5,

521 8-14 ka and Deposit 2, 130 ka occur during periods of rapid sea-level rise (>5 m/ka)
522 (Figure 5). Note that older landslides of Centre Hills age are not well dated. Larger
523 landslide deposits, associated with volcanic flank collapse, are easily distinguishable from
524 other eruption-related deposits, unlike turbidites, and there is a better degree of
525 confidence in the dating of these events. We investigate this apparent correlation in a later
526 section using a wider global dataset (*section 5.5*).

527

528 **5. Discussion**

529

530 *5.1. How does the submarine record relate to subaerial activity?*

531 Montserrat is part of an active island arc with numerous active volcanoes on other islands,
532 which may be the source of some deposits in the U1395B stratigraphy. For example, site
533 U1395B is situated ~ 25 km southeast of Montserrat and ~ 35 km downwind (northwest) of
534 Guadeloupe [Radiosonde wind data], which is thus also a plausible potential source for
535 deposits in the core. Eruptive products from Guadeloupe can be distinguished using lead
536 isotopes because Guadeloupe has a higher radiogenic lead component [Cassidy et al.,
537 2012a; Palmer et al., 2016], and Coussens et al. [2016] use Pb isotopes to show that most
538 tephra layers recorded within Hole U1395 B are likely to be from Montserrat. Over the
539 past 1 Ma [Samper et al., 2007], volcanic activity occurred in the southeast of Guadeloupe
540 at Grande Découverte Soufrière, producing landslides that have mostly travelled
541 westwards into a canyon system [Samper et al., 2007]. The canyon system has
542 accumulated deposits towards the west or southwest of Guadeloupe: thus, landslides and
543 turbidites from Guadeloupe are unlikely to produce deposits within Hole U1395B, and we
544 assume that the majority of visible turbidite layers are from Montserrat.

545

546 Limited Ar-Ar dating from on-land samples on Montserrat suggests that volcanism
547 occurred at Centre Hills between ~990–550 ka [Harford et al., 2002] followed by a long
548 hiatus in activity until ~290 ka, when activity commenced at Soufrière Hills. At ~130 ka
549 there was a brief period (~10 kyr) of eruptive activity at the new volcanic centre, South
550 Soufrière Hills, after which activity migrated back to the Soufrière Hills. The apparent
551 gap in subaerial volcanism from ~500-290 ka is not replicated in any of the age models
552 for Hole U1395B (Figures 4 and 5), where the periods of relative quiescence are much
553 shorter, lasting 10^4 - 10^5 kyr. This suggests that activity migrated relatively rapidly between
554 Centre Hills and Soufrière Hills, and may even have overlapped. This difference between
555 the subaerial and marine records is likely an artefact of limited dating, exposure and
556 erosion of the subaerial record. As such, the offshore stratigraphy provides a higher
557 resolution record and a better means of investigating temporal patterns in volcanic activity
558 in greater detail.

559

560 Trofimovs et al. [2006] showed that 90% of the eruptive products from the 1995-2010
561 eruption of Montserrat had been deposited offshore, supporting our inference that the
562 submarine record is likely to provide a more complete archive of events than the subaerial
563 record. However, deposition in marine settings involves complex transitions between
564 subaerial and submarine transport processes. Thus, although the offshore stratigraphy
565 records more volcanic events than the subaerial record, offshore facies may not provide a
566 truly representative record of a particular type of eruption.

567

568 *5.2. How does the frequency of collapses relate to volcanism?*

569 Increases in the frequency of tephra fall deposits above the mean 90% confidence interval
570 (3.64 events in 50 ka) indicate that there was more explosive volcanic activity between

571 ~800 ka to ~760 ka (Centre Hills period), and ~230 ka to ~140 ka (Soufrière Hills period)
572 (Figure 5). Turbidite frequency also increases during these periods, but not all observed
573 increases are above the mean 90% confidence interval (Figure 5). These observations are
574 consistent for all age models used, suggesting that these periods of increased volcanism
575 and turbidite frequency are unlikely to be age model artefacts. During periods of effusive
576 volcanic activity (i.e. the lava-dome forming eruptions that typify much of Montserrat's
577 volcanism), there is a greater rate of edifice growth and an increase in the number of
578 volcano-tectonic earthquakes and explosive eruptions, all of which are likely to enhance
579 the likelihood of mass wasting. The largest flank collapse from Soufriere Hills (130 ka)
580 occurred near the end of the peak activity at the centre (whether derived from all core
581 deposits, or tephra fall deposits alone), suggesting that elevated output at this time led to
582 major flank instability.

583

584 Bioclastic turbidites show another increase in frequency at ~560 ka to ~500 ka suggesting
585 that other factors may be important in triggering or priming carbonate shelf collapse
586 (Figure 5), such as sea-level change or large earthquakes.

587

588 *5.3. How does the frequency of collapses at Montserrat relate to sea-level?*

589 Turbidite frequency does not show a statistically robust correlation with eustatic sea-level
590 changes (in all age models), but all three of the well-dated large-scale landslides (>0.3
591 km^3) from Montserrat (Deposits 1, 2 and 5) occurred during periods of rapid (>5 m/ka)
592 global sea-level rise (Figure 5) [Miller et al., 2005]. Deposits 1 and 5 (dated between 14-8
593 ka) coincide with rapid sea-level rise at 15 ka to 5 ka, from 60 m to 1.6 m below present
594 day sea-level (Figure 5). Deposit 2 is dated at 130 ka, which coincides with a second
595 period of rapid sea-level rise at 130 ka to 120 ka (Figure 5). Although this is a small

596 number of events, the record suggests that rapid sea-level rise may trigger or precondition
597 the flank collapses on Montserrat.

598

599 *5.4. Large-scale flank collapse: constructing a global landslide record.*

600 To investigate if large landslide occurrences during rapid sea-level rise are replicated
601 more widely, we have compiled a global dataset of landslide ages at volcanic islands over
602 the past 1 Ma, building on the compilation by Quidelleur et al., [2008]. The compiled
603 global data set has 25 ages from volcanic islands (errors $<\pm 22.5$ ka) that are constrained
604 within errors of ± 22.5 ka, comprising 9 from ocean islands and 16 from island arcs
605 (Tables 3 and 4).

606

607 Many landslides have large age uncertainties, and if age errors are too large (i.e., on the
608 order of sea-level cycles) then relationships between sea level, volcanism, and landslides
609 cannot be determined [Pope et al., 2015]. Our age-uncertainty criterion of $<\pm 22.5$ ka has
610 been selected on this basis, because it provides a sufficiently large dataset to observe
611 potential patterns, with age errors that are a similar order of magnitude to the periodicity
612 of sea-level cycles (10^3 - 10^4 yrs).

613

614 *5.4.1 Ocean islands*

615 Multiple large-scale landslide deposits dating back to at least 22 Ma lie offshore the
616 Canary Islands [Stillman, 1999]. The oldest well-dated landslide from the island of El
617 Hierro is El Julan, to the SW (130 km^3) [Gee et al., 2001]. Onshore there is an 8 km wide
618 scar, with infilling material dated at $<158 \pm 4$ ka [Guillou et al., 1996]. Turbidite P, in the
619 Madeira Abyssal Plain, is dated at 540 ± 10 ka using biostratigraphy, and is correlated to
620 the El Julan landslide based on geochemistry and subaerial date constraints [Hunt et al.,

621 2013], and is our best age estimate of the El Julan collapse. North of El Hierro lies the El
622 Golfo deposit (150 km³) [Urgeles et al., 1997], which is correlated with a large (14.5 km
623 wide) subaerial crater [Urgeles et al., 1997]. This scar is dated between 87± 8 ka and 11 ±
624 7 ka using Ar-Ar dating of material that are incised by and infill the scar [Guillou et al.,
625 1996]. In the Madeira Abyssal Plain turbidite B is correlated with the El Golfo collapse
626 [Hunt et al., 2013] and is dated at 15 ± 5 ka [Weaver et al., 1992; Hunt et al., 2013] using
627 biostratigraphy of surrounding hemipelagic material. We use the age of turbidite B for the
628 date of the El Golfo collapse in this study.

629

630 The Icod, Orotava and Guimar landslides are the best-dated landslides from Tenerife. The
631 Guimar landslide (120 km³) [Krastel et al., 2001] correlates with turbidite Z in the
632 Madeira Abyssal Basin [Hunt et al., 2013], which places the events via biostratigraphic
633 ages at 850±10 ka. The Orotava and Icod landslides, north of Tenerife, correlate with
634 large amphitheatre-shaped depressions. The minimum age of the Orotava landslide is
635 deduced from the oldest lava flow (PS4) infilling the Orotava Valley (534±9 ka)
636 [Boulesteix et al., 2013], and the maximum age from lavas intersected by the landslide
637 scar (558±8 ka). The landslide also correlates with turbidite MO in the Madeira Abyssal
638 Plain, dated at 540±5 ka using biostratigraphy [Hunt et al., 2013]. Subaerial dating
639 suggests the Icod landslide occurred between 175±3 ka and 161±5 ka by K-Ar dating
640 material that are incised by and infill the landslide scar respectively [Boulesteix et al.,
641 2011] but more recently Hunt et al., [2013] used turbidite G in the Madeira Abyssal Plain
642 to obtain a biostratigraphic age of 165±5 ka, which we use here.

643

644 Offshore La Palma, the Cumbre Nueva landslide is well-dated. The headwall of the
645 landslide has been identified as the prominent Cumbre Nueva Ridge that encloses a

646 horseshoe-shaped depression [Urgeles et al., 1999]. The Cumbre Nueva landslide is
647 thought to have occurred after the emplacement of the Upper Taburiente volcanic deposits
648 (566 ± 8 ka) and was followed by the emplacement of the Bejenado volcano (537 ± 8 ka) in
649 the landslide scar [Guillou et al., 1998, 2001; Carracedo et al., 2001]. Hunt et al [2013]
650 date the Cumbre Nueva landslide at 480 ± 5 ka by correlating the landslide with turbidite N
651 in the Madeira Abyssal Plain. This contrasts with subaerial dating, and we preferentially
652 use the subaerial dates, because turbidite N may correlate with a younger landslide from
653 La Palma.

654

655 To the east of Pico del Fogo in the Cape Verde Islands is a $130\text{-}160$ km³ landslide deposit
656 [Lebas et al., 2007; Masson et al., 2008], which has been correlated with fields of
657 boulders and chaotic conglomerates (interpreted as tsunami deposits) on the nearby island
658 of Santiago [Ramalho et al., 2015]. ³He dating of these tsunami deposits suggests a
659 landslide age of 73 ± 7 ka [Ramalho et al., 2015]. This is close in age to a post collapse
660 lava flow on Pico del Fogo dated at 86 ± 3 ka [Paris et al., 2011].

661

662 In the Hawaiian Archipelago at least 68 large volume landslides (up to 5000 km³) have
663 been identified [Moore et al., 1989, 1994]. However, only the Alika 2 deposit, offshore
664 Mauna Loa, is well dated. Cores F88 B-13, F91 39B1 and F91 4082 sample the Alika 2
665 deposit. Planktonic forams within the Alika 2 deposit date the landslide at 127 ± 5 ka using
666 oxygen isotope stratigraphy [McMurty et al., 1999].

667

668 The island of Tahiti has two horseshoe shaped depressions to the north and south of the
669 island that correlate with large hummocky landslide deposits offshore with respective
670 volumes of 800 km³ and 1150 km³ [Hildenbrand et al., 2004]. The Tahitian shield was

671 constructed between 1.4-8.7 Ma [Hildenbrand et al., 2004], with vent locations
672 concentrated along an E-W rift down the centre of the island. Late activity along these
673 vents (~870 ka) is thought to have triggered the two landslides [Hildenbrand et al., 2004].
674 After the landslides, activity moved to the north, infilling the northern crater (851±11 ka)
675 [Hildenbrand et al., 2004]. We therefore assume an age range for the northern landslide
676 between 870 to 851±11 ka. The southern crater is dated at ~500 ka, but is too poorly
677 constrained to be included in the database.

678

679 *5.4.2 Island Arc Landslides*

680 Evidence of large landslides has been identified at several islands in the Lesser Antilles
681 (in addition to Montserrat). At Guadeloupe, subaerial exposures allow two events, the
682 Carmichael and Carbet landslides, to be well dated [Boudon et al., 1984, 1987, 2007]. The
683 Carmichael landslide covers 17 km² onshore, and large blocky deposits are observed
684 offshore [Boudon et al., 1984]. The landslide is thought to be associated with the
685 Carmichael crater that has a volume of 0.3 km³. Uncarbonised wood found within the
686 deposit yields ¹⁴C ages of 8400 ±1500 to 11270±185 BP [Boudon et al., 1984]. The
687 Carbet landslide (estimated volume of 0.5 km³) is dated by subaerial outcrops along the
688 Rivière du Carbet, where uncarbonised wood within the deposits yielded ¹⁴C dates
689 between 2800-3450 BP [Boudon et al., 1984, 2007; Daigun et al., 1981].

690

691 On Dominica four large landslides are observed, but only the Soufriere event is well dated.
692 The Soufriere event (estimated volume 6-7 km³) formed from the collapse of the Plat
693 Pays volcanic edifice [Le Friant et al., 2002]. The lower bound age constraint is from ¹⁴C
694 dating of carbonaceous material found within a pyroclastic flow deposit from the Plat-
695 Pays volcano dated at 6600±50 ka. An upper bound comes from ¹⁴C dating material

696 (2380± 75 ka) from a tephra fall that overlies a mega block from the landslide (Scotts
697 Head) [Le Friant et al., 2002].

698

699 On Martinique, three of the four large landslide deposits are well dated (D1, D2 and D3).
700 The D1 event is associated with a large west-facing landslide scar and has an estimated
701 volume of 25 km³. The northern rim of the crater is well exposed, but the southern rim
702 may have been destroyed and/or covered by later landslide events [Boudon et al., 2007].
703 K-Ar dating of material infilling and incised by the landslide scar are dated at 126±2 ka
704 and 127±2 ka respectively. The D2, or St Pierre event formed a large west-facing
705 amphitheatre-shaped structure and produced an offshore deposit with a volume of >13
706 km³. The upper bound date is given by U-Th disequilibrium dating of material
707 intersected/outside the D2 structure, yielding ages of 37±4 ka (Mont Calebasse lava
708 dome) and 25±1 ka (Mourne Plumé lava dome) [Le Friant et al., 2003]. The lower bound
709 age, of 25 000±1040 years, is given by a ¹⁴C charcoal age from a scoria flow deposit that
710 in-fills the D2 structure [Vincent et al 1989]. D3, or the Rivière Sèche event, has an
711 estimated volume of 2 km³. The Aileron lava dome (summit part of Montagne Pelée) is
712 intersected by the D3 structure and is dated at 9.7±0.5 ka, providing an upper age limit
713 [Le Friant et al., 2003]. The “Sans Nom” lava dome in-fills the D3 structure and is dated
714 at 9±1 ka [Le Friant et al., 2003].

715

716 Further afield, multiple large landslide deposits have been identified on and offshore
717 Ischia, Italy [Chiocci and de Alteriis 2006; De Alteriis and Violante, 2009], five of which
718 are well dated. The South landslide (1.5 km³) is correlated with the offshore DF1 deposit
719 and is dated at 2.4-3 ka (¹⁴C on planktonic forams within hemipelagic material above and
720 below DF1) [De Alteriis et al., 2010]. The Falanga and Pietre Rosse landslides outcrop to

721 the east of Ischia, with a combined offshore volume of 11 km³. The Falanga deposit is
722 dated at 400 BC based on historical chronicles [Buchner 1986]. The Pietre Rosse
723 landslide has a maximum age constraint as it overlies the Citara Formation 40 ka [Vezzoli
724 1988], and the upper bound age is given by the overlying Falanga landslide.

725

726 To the north of Ischia two large landslide deposits (NDDe and NDDw) are seen [De
727 Alteriis and Violante 2009]. The NDDe correlates onshore with the Casamicciola and
728 Lacco Ameno landslide deposits [Seta et al., 2012]. The Lacco Ameno deposit is poorly
729 dated and is not included within the global database. The Casamicciola deposit overlies
730 the Punta La Scrofa Tuff, dated at <800 BC [De Vita et al., 2010]. Within seismic
731 reflection profiles, a horizon beneath NDDw has been correlated with the onshore Zaro
732 lava flow, dated at 6±2 ka and 9±1 ka using K-Ar dating [Vezzoli et al., de Aleriis and
733 Violante, 2009]. Although the Casamicciola and NDDw have no upper bounds, the ages
734 are within the error of ± 22.5 kyr, and so are included within the global dataset.

735

736 The Japanese volcano of Usu is situated in the South of the Japanese island of Hokkaido.
737 Usu formed ~20 ka [Soya et al., 2007] and after the edifice reached 1000 m there was a
738 large summit collapse (1-2 km³). This collapse, named the Zenkojii collapse, occurred at
739 ~7-8 ka based on subaerial dating of infilling and cross-cut deposits [Moriya, 2003 (in
740 Japanese); Soya et al., 2007].

741

742 A large 5 km³ collapse from Ritter island (offshore New Guinea), which occurred 13th
743 March 1888, is also included in the global database [Cooke 1981; Ward and Day 2003].

744

745

746 *5.5. Large-scale flank collapse and sea-level: comparison to global records*

747 At Montserrat, the three well-dated large landslides from Soufrière Hills coincide with
748 periods of rapid sea level rise. Using the global compilation, box and whisker plots
749 (Figure 6) show that landslide occurrences at other island volcanoes are skewed towards
750 periods of more rapid sea-level rise. In contrast, at ocean islands the distribution of event
751 timing shows no clear correlation with specific regimes of sea level (Figure 6). This is
752 supported by preliminary statistical analyses (Mann-Whitney and Kolmogorov-Smirnov
753 tests, see supplementary information).

754

755 The global data set of landslide ages is restricted (9 events from ocean islands, 16 events
756 from island arcs). Dating landslide deposits is challenging, because most material is
757 emplaced offshore in poorly accessible, deep marine settings, typically comprising
758 chaotic mixtures of material. Direct dating of material within landslides only provides
759 maximum age constraints of landslide emplacement, unless material such as vegetation
760 incorporated into the landslide deposits can be identified and dated. Indirect dating of
761 material overlying and underlying landslide deposits or scars can result in large age gaps.
762 Consequently the global landslide database is sparse, and dominated by a very few well-
763 studied regions (e.g. Ischia, Lesser Antilles). Even in other regions where numerous
764 events have been identified (e.g. Aleutian Islands; Coombs et al. [2007]), very few events
765 are dated, and so are excluded from our dataset.

766

767 Given the limitations of a restricted dataset, it is thus difficult to test the strength of the
768 observed correlation between event timing and sea level change. Furthermore the island
769 arc database is biased towards recent landslide events (<20 ka), which includes a period
770 of rapid sea-level rise (Figure 6). Thus the apparent prevalence of island arc landslides

771 during periods of rapid sea-level rise could also be affected by a temporal record bias.
772 Further investigation (by substantially increasing the dated landslide database) is required
773 to test this initial result, which suggests a difference between how sea level change
774 influences flank stability in island arc and ocean island settings.
775
776 Nevertheless, it is worth briefly considering the processes by which sea level rise may
777 increase the likelihood of flank collapse at island-arc volcanoes. A ~100 m rise in sea
778 level potentially enhances both dyke ascent (and hence volcanism) and fault movement in
779 island-arc settings [cf. Nakada and Yokose, 1992], although the relationship between
780 volcanism and sea level change may be complex [McGuire et al., 1997] and is
781 incompletely understood. Overpressure in magma chambers, which decreases by
782 viscoelastic relaxation, is more sensitive to loading rate than the absolute load [Jellinek et
783 al., 2004] consistent with the observed correlation of collapse rapid sea-level rise.
784 Furthermore, sealevel rise may result in greater magma- water interaction encouraging
785 explosive, phreatomagmatic behaviour. Such explosive behaviour may reduce flank
786 stability. An incipient instability within a volcanic edifice may also move closer to a
787 critical failure point due either to direct stress increases associated with sea-level rise,
788 possibly combined with other factors. These may include enhanced erosion and
789 modification of the groundwater system associated with rapid sea-level change. Although
790 the consequence of sea-level rise on groundwater may depend on local factors, pore-
791 pressure increases within hydrothermal systems have been suggested as a cause of edifice
792 collapse [Elsworth and Voight, 1995; Day, 1996; Reid, 2004].
793
794 Our results suggest that these factors may be less significant in ocean-island settings, but
795 the reason for this is as yet unclear. Structurally, the more basaltic composition of ocean

796 islands means they are likely to have more stable flanks than island arcs, with a broader
797 profile and lower slope gradients [Watt et al., 2014], yet the presence of widespread
798 landslide deposits around many ocean islands indicates that they are still prone to
799 catastrophic flank failure.

800

801

802 **6. Conclusions**

803

804 In contrast to the subaerial stratigraphic record, volcanoclastic layers within IODP Hole
805 U1395B provide a more complete record of volcanism on Montserrat. Periods of reduced
806 event frequency (interpreted as periods of volcanic quiescence) last of the order 10^4 – 10^5
807 ka, substantially less than apparent gaps in the subaerial record. The pauses in activity
808 suggested by the subaerial record are likely to be a manifestation of poor preservation and
809 burial of onshore stratigraphy, coupled with the predominantly offshore deposition of
810 volcanic products. The marine record reveals that three periods of heightened volcanic
811 activity at Montserrat occurred at ~930 ka to ~900 ka, ~810ka to ~760 ka (both during the
812 Centre Hills period), and ~190 ka to ~120 ka (during the Soufrière Hills period). A
813 notable increase in event frequency at ~810 ka to ~760 ka coincides with the occurrence
814 of mafic scoria in Hole U1395B, while the peak in Soufrière Hills volcanism coincides
815 with the largest flank collapse from the volcano (130 ka) and a brief switch to mafic
816 volcanism.

817

818 Periods of heightened volcanic activity coincide with periods of increased turbidite
819 occurrence, suggesting that intensified volcanic activity may facilitate mass-wasting
820 processes. No statistically significant correlation of turbidite occurrence (collapse

821 frequency) and sea level were observed, but all of the well-dated and large-scale
822 landslides on Montserrat occurred during periods of rapid sea-level rise. While the global
823 dataset is small and there are uncertainties associated with dating landslide events, this
824 pattern of large-scale landslides coinciding with periods of rapid sea-level rise is
825 replicated within a global data set for island-arc volcanoes, but not at ocean islands. The
826 reasons for this difference between island arcs and ocean islands remain unclear, but may
827 reflect a greater susceptibility of the steeper, more lithologically diverse flanks of island-
828 arc volcanoes to rapid sea-level changes, relative to ocean islands.

829

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843 maya.pratt@mail.com) and IODP LIMS reports (<http://web.iodp.tamu.edu/UWQ/>).

844

845 **8. References**

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1250

1251 Figure 1. Topographic and bathymetric map of Montserrat showing site of U1395B and
1252 large debris avalanche flows [Le Friant et al., 2004; Boudon et al., 2007; Lebas et al.,
1253 2011; Watt et al., 2012a, 2012b] .

1254

1255 Figure 2: Upper 44 m of Hole U1395B including core log and photos; calculated
1256 sedimentation rates for all age models; NRM declination data from shipboard
1257 measurements; section of stratigraphic core photos and log; sample of componentry data;
1258 oxygen isotopes from this study and the global oxygen isotope curve from Lisiecki and

1259 Raymo [2005]. For full componentry data and stratigraphic log of Hole U1395B see
1260 supplementary data.

1261

1262 Figure 3: Correlations of the top of core U1395B with shallow cores based on
1263 componentry and oxygen isotope data [Le Friant et al., 2008; Cassidy et al., 2012b, 2013;
1264 Ogg et al., 2012; Trofimovs et al., 2013; Wall-Palmer et al., 2014].

1265

1266 Figure 4: Timing of (a) all events, (b) volcanoclastic events (tephra fall and turbidite
1267 deposits; including mixed composition turbidites), (c) volcanoclastic events (tephra fall
1268 and volcanoclastic turbidite), (d) tephra fall deposits, and (e) all turbidites at Montserrat.
1269 In each case, three different age models are shown. Age model 1 (green) uses oxygen
1270 isotope data, biostratigraphic data, AMS dates, and paleomagnetic dates. Age model 2
1271 (red) uses Deposit 2 as a boundary at 130 ka (Figure 2), and the paleomagnetic intervals.
1272 Age model 3 (blue) uses only the paleomagnetic data. Clustering of events in all age
1273 models can be seen at ~950-860 ka, 810-750 ka, 200-100 ka and from ~50 ka, on
1274 Montserrat. There is an absence of prolonged pauses in activity. Thin grey lines show
1275 occurrence of scoria within core U1395B, the thick grey line shows known extents of
1276 activity at volcanic centres on Montserrat based on limited subaerial dating [Harford et al.,
1277 2002].

1278

1279 Figure 5: Number of events within 50 kyr (centred on window, at 10 kyr increments).
1280 Peaks correspond to clusters in Figure 4. Age model 1 (green) uses oxygen isotope data,
1281 biostratigraphic data, AMS dates, and paleomagnetic dates. Age model 2 (red) uses
1282 Deposit 2 as a boundary at 130 ka (Figure 2), and the paleomagnetic intervals. Age model
1283 3 (blue) uses only the paleomagnetic data. Peaks in event frequency occur at ~930 ka to

1284 ~900 ka, ~810 ka to ~760 ka, and ~190 ka to ~120 ka. Periods when event frequency is
1285 above the 90% confidence interval in all age models are shaded in grey. The black line is
1286 the Miller et al., [2005] sea-level curve in meters from present day sea-level. Thin grey
1287 lines show occurrence of mafic scoria within Hole U1395B, the thick grey line shows
1288 known extents of activity at volcanic centres on Montserrat based on limited subaerial
1289 dating [Harford et al., 2002]. There is a peak in event frequency that coincides with the
1290 appearance of mafic scoria in Hole U1395B and all large flank collapses occur during
1291 periods of rapid sea-level rise.

1292

1293 Figure 6: A. Diagram summarising the age of large ($>0.3 \text{ km}^3$) landslides around
1294 volcanic islands and sea-level. The sea-level curve used is Miller et al., [2005]. Events
1295 considered have date errors of $\leq \pm 22.5 \text{ ka}$. Periods of rapid sea-level rise ($>5 \text{ m/ka}$)
1296 are highlighted in blue. Vertical bars show errors of individual landslide ages. Ocean
1297 island landslides are shown in green, and island arc landslides are shown in red.
1298 Letters identify individual landslides and correspond to letters in Table 3. B.

1299 Comparison of sea level conditions with those at the time of documented landslides
1300 (Table 3) using data in 5 ka bins. Grey solid circles indicate the full range of conditions
1301 over the past 1 Ma at which no landslides were recorded. Annotated box and whisker
1302 plots indicate the spread of each data set, with boxes showing 25%, 50% and 75% of the
1303 data, whiskers showing the minimum and maximum. Numbers next to data points show
1304 the number of overlapping datum. A general skew of landslides occurring at rapid sea-
1305 level rises can be seen in the box and whisker plots.

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Allocation number	Publication code	Sample identifier	$\delta^{13}\text{C}_{\text{VPDB}}\text{‰}$ ± 0.1	Carbon content (% by wt.)	^{14}C Enrichment (% modern)	+/- 1σ (% modern)	Conventional Radiocarbon Age (years BP)	+/- 1σ (radiocarbon yrs BP)	Stratigraphic position (cm)
1721.0513	SUERC-46961	1H2W-108-110	1.1	10.9	28.82	0.13	9993	37	258-260
1721.0513	SUERC-46962	1H2W-115-117	1.1	10.8	23.93	0.12	11489	40	265-267
1721.0513	SUERC-46965	1H3W-109-111	1.2	11.4	0.33	0.06	45806	1463	409-411
1721.0513	SUERC-46966	1H3W-116-118	1.0	11.3	0.40	0.06	44294	1208	416-418

1310 Table 1: AMS carbon dates from core U1395.

1311

	Age Model 1	Age Model 2	Age Model 3
Linear model	0.0898	<i>0.00176</i>	<i>0.00285</i>
Generalised linear model	0.083407	<i>0.00138</i>	<i>0.00696</i>
Proportional hazards model	0.106	<i>0.00219</i>	<i>0.00942</i>

1312 Table 2: Results of linear statistical tests for turbidite frequency and sea-level correlations. Italicised values show when statistical tests are

1313 significant, (ie when the null hypothesis can be rejected).

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1315 Table 3: Global landslide database for ocean island landslides and island arc landslides
1316 with volumes $>0.3 \text{ km}^3$ and with age errors of $< \pm 22.5 \text{ kyr}$.

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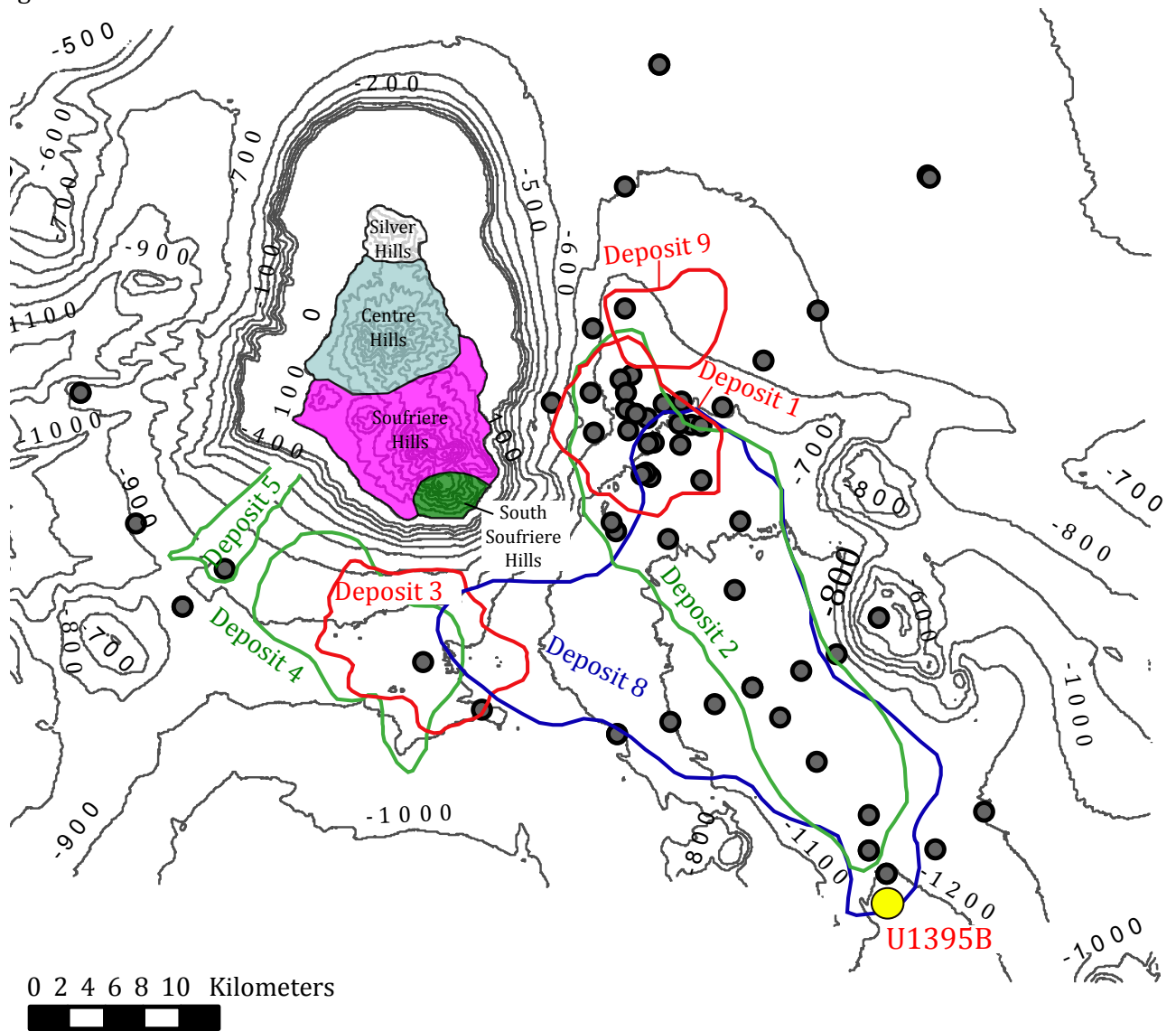
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Flank collapse name	Country/ Island	Age (ka)	Volume (km ³)	References	Letter in Supplementary Figure S4
<i>Ocean Islands</i>					
El Golfo	Canary Islands, El Heirro	20-10	150	Weaver et al., [1992]; Longpre et al., [2011]; Hunt et al., [2013]	a
El Julan	Canary Islands, El Heirro	550-530	130	Hunt et al., [2013]	b
Cumbre Nueva	Canary Islands, La Palma	574-529	95	Costa et al. [2014]	c
Icod	Canary Islands, Tenerife	170-160	320	Ancochea et al., [1990]; Wynn et al., [2002]; Carracedo et al., [2007]; Hunt et al., [2013, 2014]	d
Orotava	Canary Islands, Tenerife	566-525	1000	Watts and Masson [1995]; Masson et al., [2002]; Boulestreix et al., [2013]; Hunt et al., [2013]	e
Guimar	Canary Islands, Tenerife	860-840	120	Krastel et al., [2001]	f
Fogo	Cape Verde	80-66	160	Ramalho et al., [2015]	g
Alika 2	Hawaii, Mauna Loa	132-122	>0.5	McMurty et al. [1999]	h
Tahiti North	Tahitian Archipelago, Tahiti	870-850	800	Clouard et al., [2000]; Hildenbrand et al., [2004, 2006].	i
<i>Island Arcs</i>					
Carmichaël	Lesser Antilles, Guadeloupe	11.5-6.9	0.3	Boudo et al., [1984, 2007]	j
Carbet	Lesser Antilles, Guadeloupe	3.5-2.8	0.5	Boudon et al., [1984, 2007]; Jerime [1979]; Paterne [1980]; Daigun [1981]; Daigun et al., [1981]	k
La Soufriere	Lesser Antilles, Dominica	6.7-2.3	6 to 7	Le Friant et al., [2002]; Boudon et al., [2007]	l
South	Italy, Ischia	3-2.4	1.5	de Alteriis et al., [2010, 2014]	m
Pietre Rosse	Italy, Ischia	40-2.4	>0.3	Vezzoli [1988]; Buchner [1986]; Seta et al., [2012]	n
Falanga	Italy, Ischia	2.4	>0.3	Buchner [1986]; Seta et al., [2012]	o
Casamicciola (NDDe)	Italy, Ischia	2.8	>0.3	de Vita et al., [2010]; Seta et al., [2012]	p
NDDw	Italy, Ischia	<10 ka		Vezzoli et al., [1988]; de Aleriis and Violante, [2009]; Seta et al., [2012]	q
D1	Lesser Antilles, Martinique	129-124	>25	Germa et al., [2011]	r
D2 or the St Pierre event	Lesser Antilles, Martinique	23.6-26.4	13	Vincent et al., [1989]; Le Friant et al., [2003]	s
D3 or Rivière Sèche event	Lesser Antilles, Martinique	9.8-8	2	Le Friant et al., [2003]; Boudon et al., [2007]	t
Usu	Japan, Hokkaido	8-7	1 to 2	Moriya, [2003] (in Japanese); Soya et al., [2007]; Yoshida et al., [2012]	u
Ritter	Papua New Guinea, Ritter Island	0.1	5	Cooke [1981]; Ward and Day [2003]	v
Deposit 1	Lesser Antilles, Montserrat	14-11	1.8	Trofimovs et al. [2013]; Watt et al. [2012a], [2012b]; Le Friant et al. [2004]; Lebas et al. [2011]	w
Deposit 2	Lesser Antilles, Montserrat	146-112	9	Harford et al. [2002]; Le Friant et al. [2004]; Lebas et al. [2011]; Watt et al. [2012a], [2012b]; Crutchley et al. [2013]	x
Deposit 5	Lesser Antilles, Montserrat	12-8	0.3	Cassidy et al. [2013]; Le Friant et al. [2004]; Lebas et al. [2011]	y

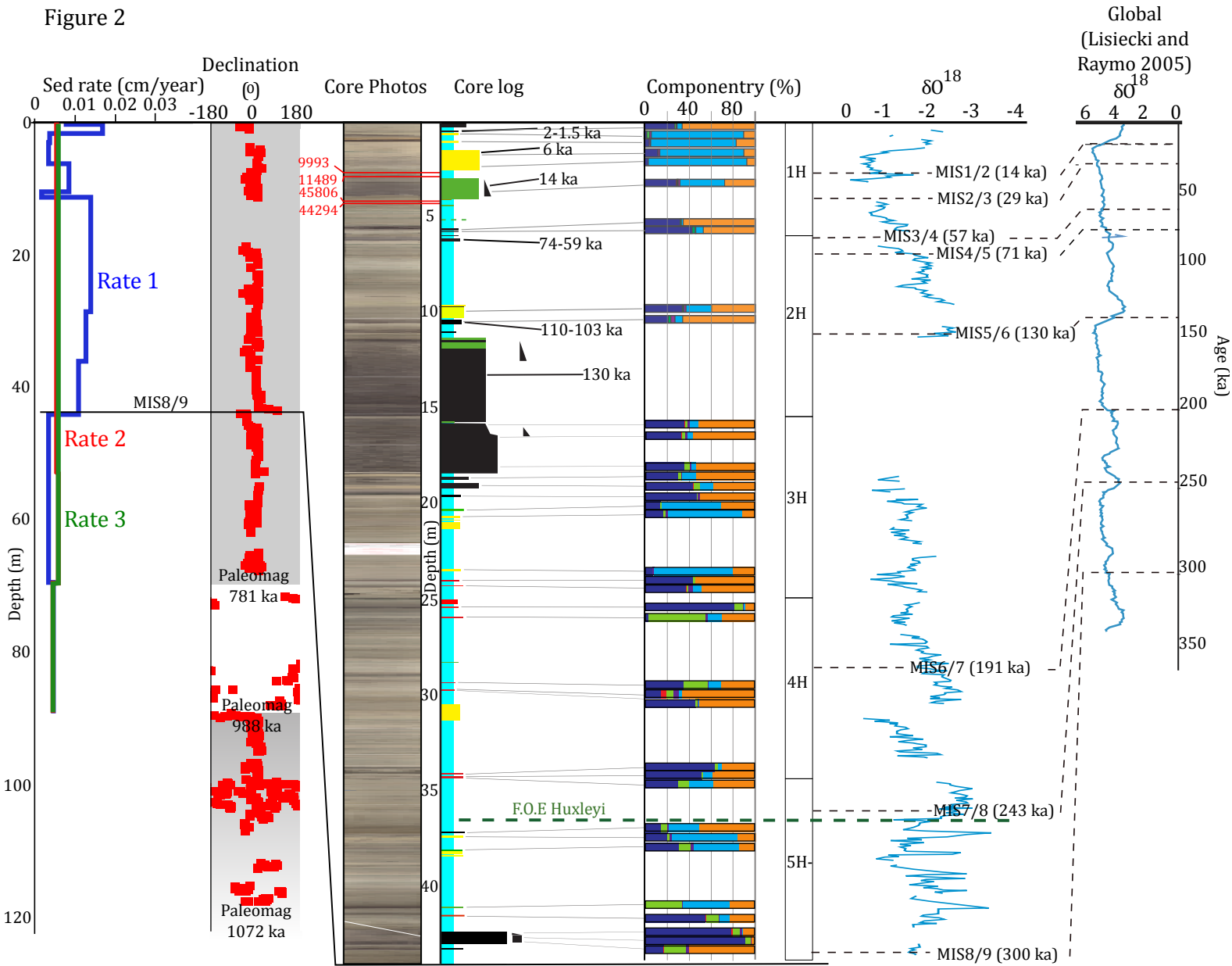
Figure 1



KEY

- contours every 100 m above sea level
- large flank collapse deposit outline (near surface)
- large flank collapse deposit outline (buried)
- large flank collapse deposit outline (deeply buried)
- shallow vibrocore locations (Trofimovs et al., 2013)
- site of IODP core U1395B

Figure 2



KEY

Stratigraphic log

- hemipelagic mud
- bioclastic turbidites (>70% Bioclasts)
- mixed turbidites (30-70% Bioclasts)
- volcanoclastic turbidites (<30% bioclasts, Folk and Ward sorting coefficient >0.5)
- tephra (<30% bioclasts, Folk and Ward sorting coefficient <0.5)

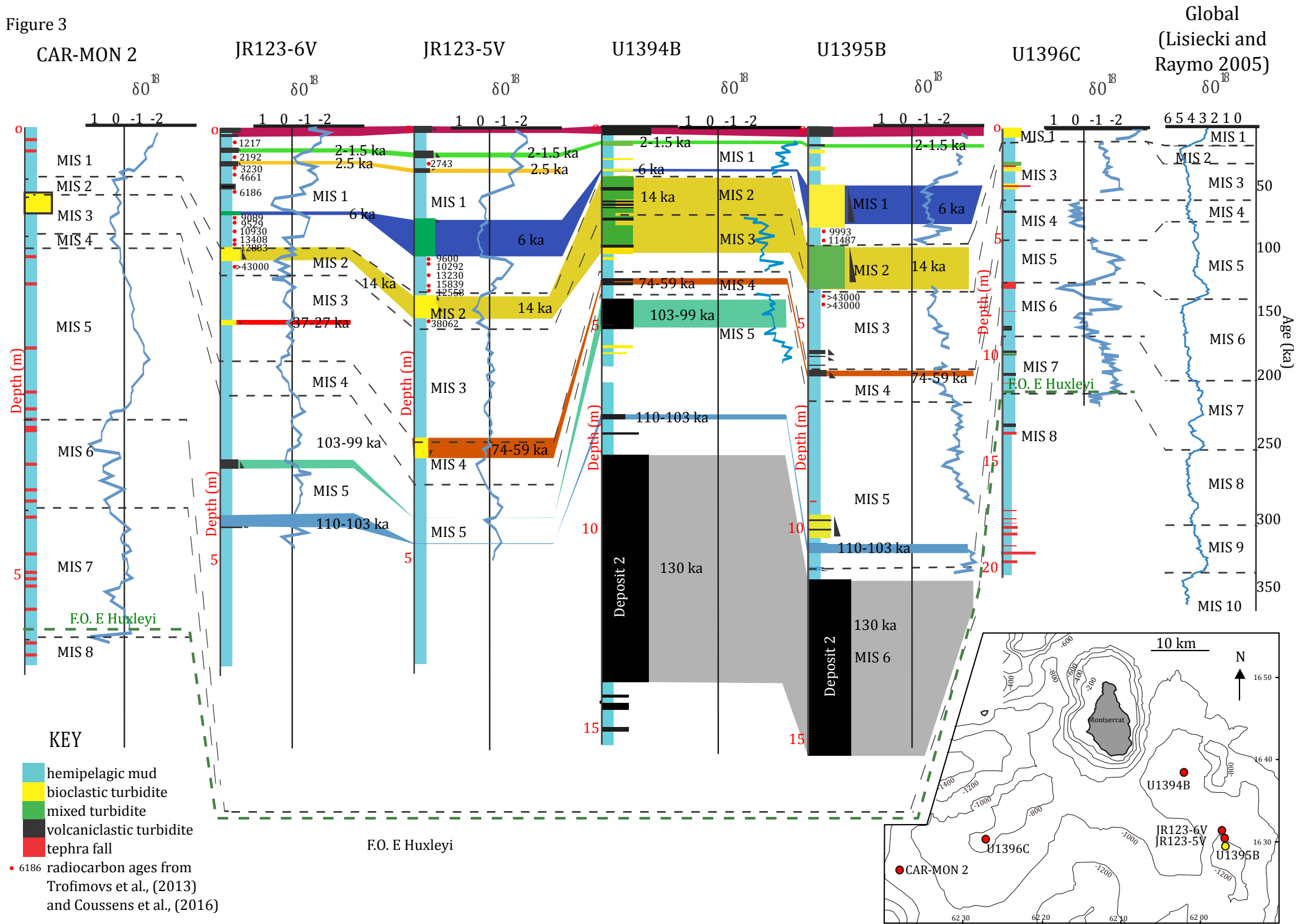
Componentry

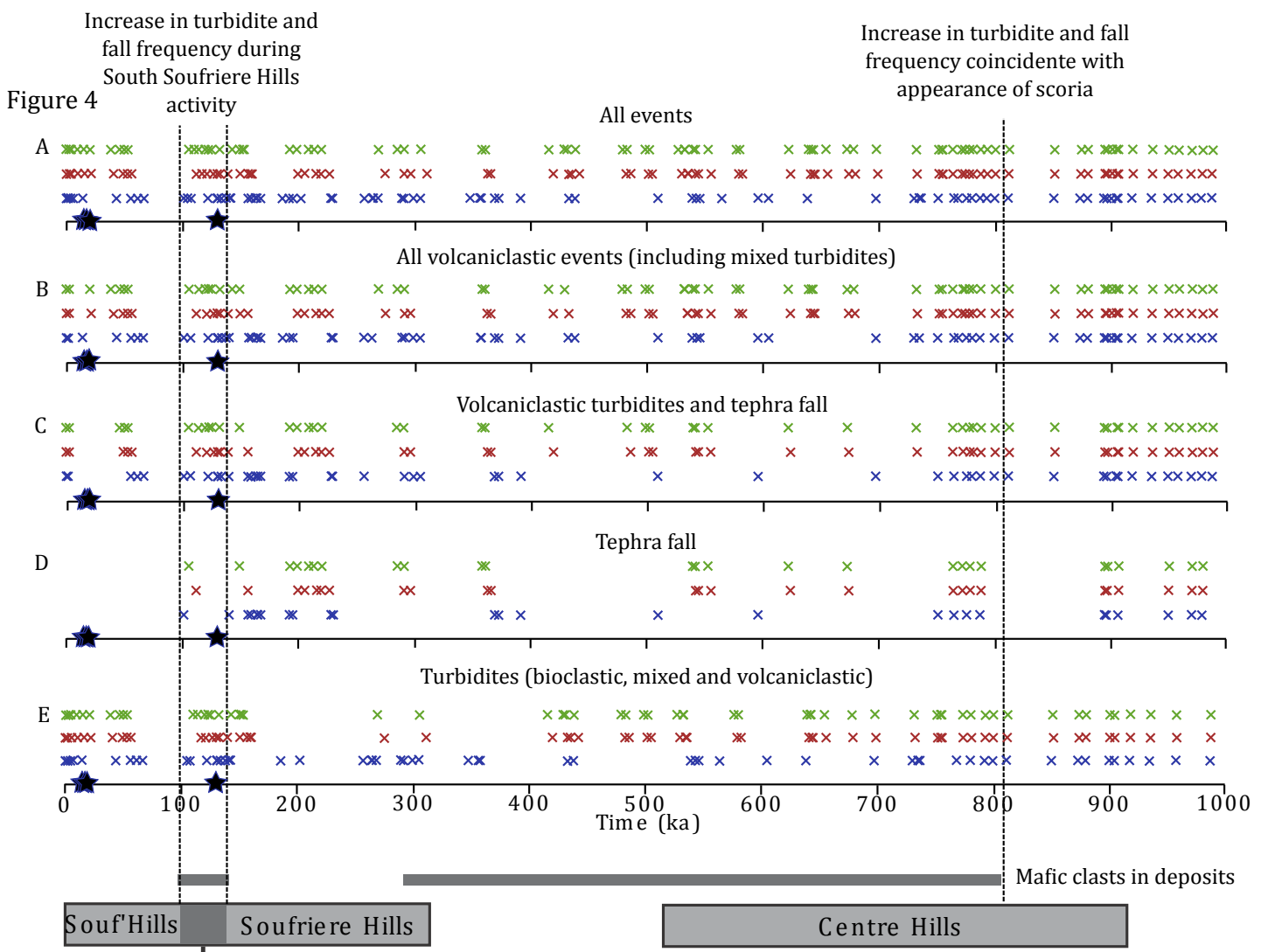
- glass and crystal
- bioclast
- altered lithic
- mafic
- dense andesite
- andesitic pumice

Other

- first occurrence (F. O) of E. Huxleyi
- 9993 AMS radiocarbon dates (in uncalibrated years)
- Grading
- 74-59 kyr units correlated with vibrocoring (Trofimovs et al. 2013)
- sedimentation rates, age model 1
- sedimentation rates, age model 2
- sedimentation rates, age model 3

Figure 3





KEY

- x age model 1
- x age model 2
- x age model 3
- ★ age of large-scale flank collapses.

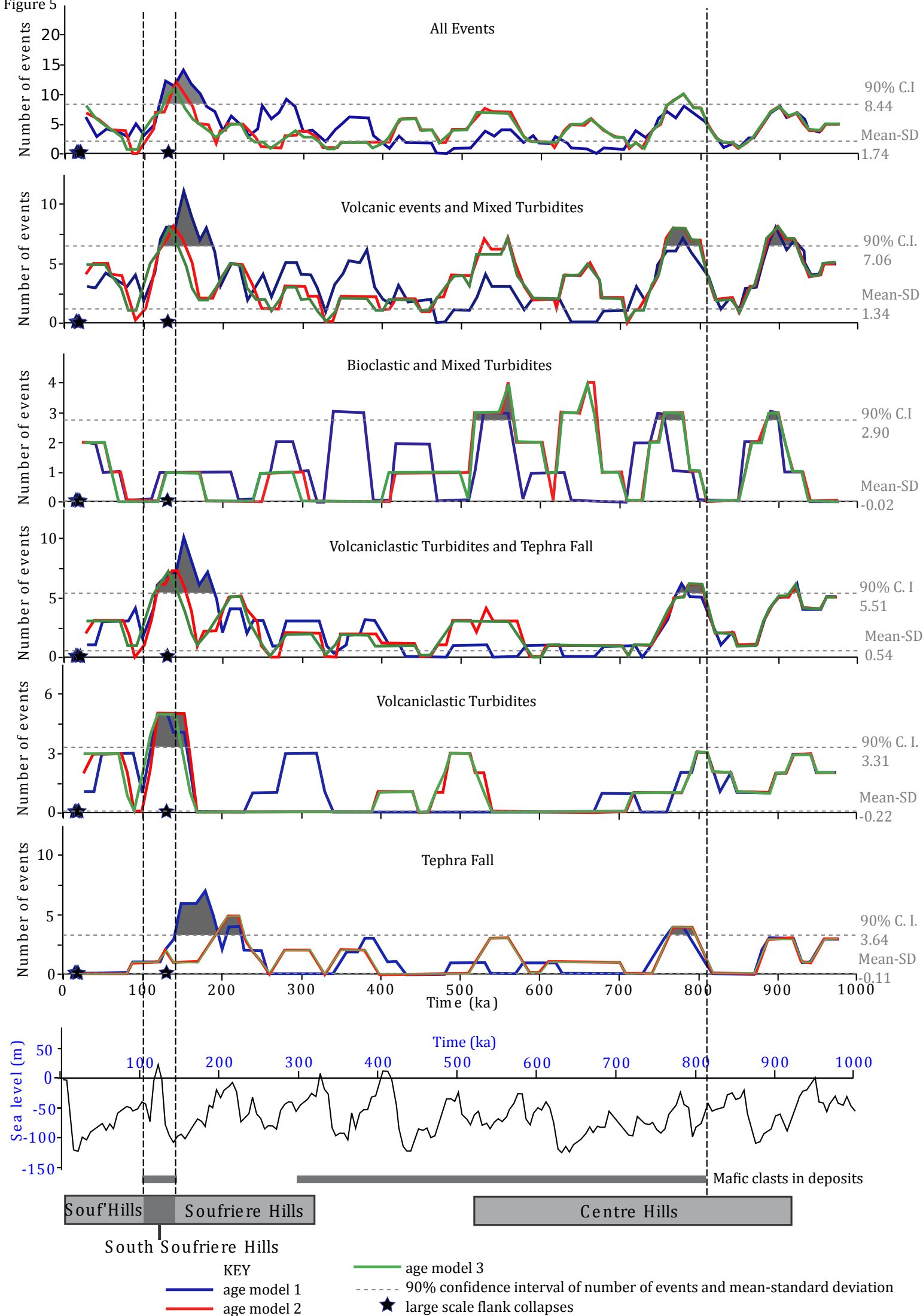


Figure 6

