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1 **Advances in the study of mega-tsunamis in the geological record**

2

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9

10 **Abstract**

11 Extreme geophysical events such as asteroid impacts and giant landslides can generate mega-
12 tsunamis with wave heights considerably higher than those observed for other forms of
13 tsunamis. In this paper, we review recent advances in the study of mega-tsunamis in the
14 geological record, focusing on well-documented examples that have captured particular
15 attention over the past decade. We provide up-to-date background on the source mechanisms
16 of tsunami generation during asteroid impacts and ocean-island landslides, which are the
17 largest landslides on Earth. We also discuss the main sources of uncertainty for modelling
18 such mega-tsunamis, and for addressing associated hazards.

19

20 **Introduction**

21 The 2011 Tohoku-oki and 2004 Indian Ocean tsunamis were occasionally described as
22 “mega-tsunamis”, not only in the media but also in peer-reviewed literature (e.g. Lekkas et al.,
23 2013), probably because they were the largest ones observed in the modern era (Goff et al.,
24 2014). However, there is evidence in the geological record for rare but extreme events, with
25 wave heights considerably larger than those observed during the 2011 and 2004 tsunamis (e.g.
26 Moore and Moore, 1984; Bourgeois et al., 1988).

27 As a background to this discrepancy, it is important to recognize the fundamental difference in
28 viewpoints between geologists, other scientists and the public regarding hazards. For instance,

29 as commonly observed in risk assessment (Fig. 1), event probability decreases with an
30 increase in hazard intensity (e.g., Kawata, 2003). Engineers and government agencies have to
31 set thresholds of design force to enable infrastructure to be built for protection against
32 hazards. However, although the frequency is low, some events are larger than this design force
33 and as such, the protection that has been constructed can fail. For such overwhelming hazards,
34 engineers and government agencies have to estimate a maximum size that they can reasonably
35 prepare for reducing damage. Sometimes, people misunderstand this and assume that this
36 maximum hazard is actually the real maximum, but larger hazards can still occur although at
37 such a low frequency that we consider them negligible in a human lifetime. These hazards are
38 beyond the limit of modern-style disaster prevention countermeasures so that when they do
39 occur they are most definitely “unexpected (or “*souteigai*” in Japanese) hazards” (Goff et al.,
40 2014). The 2011 Tohoku-oki and the 2004 Indian Ocean tsunamis were examples of these
41 kinds of unexpectedly large events. Depending on the people who study these larger hazards,
42 they use various terms such as “maximum”, “unexpected” and “mega” events.

43 In contrast, geologists who study the Earth’s history usually investigate extreme hazards such
44 as asteroid impacts, large igneous provinces, snowball earth, etc. Without doubt, these events
45 are hazards with global impacts and are far beyond the normal-style risk assessment (Fig. 1);
46 these hazards can lead even to mass extinctions. Compared with these extreme hazards,
47 “normal” hazards which occur over intervals of 100s to 1000s of years (e.g., large earthquakes
48 and tsunamis) are quite frequent and relatively small events from a geological point of view.
49 Sometimes these events are below the resolution and detection limits of geological research.
50 Indeed, even recent large events such as the 2011 Tohoku-oki and the 2004 Indian Ocean
51 tsunamis are probably only just large enough that sedimentary evidence for them will be
52 preserved in strata over geological timescales. Also, geologically speaking, the occurrence of
53 these events is not unexpected.

54 The discrepancy is in terms of the human perspective. While risk assessment is carried out
55 from the human perspective (datum on the left-hand side of the horizontal axis in Figure 1),
56 geologists study it from the point-of-view of deep time (datum on right-hand side of the
57 horizontal axis in Figure 1). This is essentially why geologists generally consider that neither
58 the 2011 Tohoku-oki nor the 2004 Indian Ocean tsunamis should be called “mega-tsunamis”
59 since they know that there have been far larger events (or they consider that evidence for
60 larger events is still undiscovered) that deserve to be called as such; “Mega” defines the
61 largest events we know about at the moment.

62 Goff et al. (2014) proposed that the term “mega-tsunami” should be reserved for tsunamis
63 with an initial wave amplitude exceeding 50 m at their sources, thus excluding all tsunamis
64 generated by historical earthquakes. Paris et al. (2018) adopted the following definition:
65 mega-tsunamis have a magnitude exceeding all published tsunami magnitude scales.
66 Whatever the definition, among all the possible source mechanisms of tsunamis, only large
67 landslides and asteroid impacts have the potential to generate mega-tsunamis.

68 With a maximum runup of 524 m, the 1958 tsunami in Lituya Bay (Miller, 1960) could be
69 considered as the only historical example of a mega-tsunami. However, this exceptional runup
70 was caused by the restriction due to the slope just opposite the source landslide (30.6×10^6
71 m^3) and values of runup rapidly decreased down to 10 m at 12 km from the source. By
72 comparison, the December 2018 Anak Krakatau volcano flank collapse had a volume three
73 times larger ($> 0.1 \text{ km}^3$: Gouhier and Paris, 2019) than the Lituya Bay landslide, and an initial
74 leading positive water displacement of 50 to 80 m (Grilli et al., 2019; Paris et al., 2019).
75 Following the definition proposed by Goff et al. (2014), the 2018 Anak Krakatau tsunami
76 could thus enter the mega-tsunami category, but the wave heights observed on the coasts of
77 Sumatra and Java (40-60 km away from the volcano) were lower than 7 m (Takabatake et al.,
78 2019; Muhari et al., 2019), which falls far from the maximum wave heights of the 2004
79 Indian Ocean (Lavigne et al., 2009) and 2011 Tohoku-oki tsunamis (Mori et al., 2012).
80 Volumes of the 1958 Lituya Bay and 2018 Anak Krakatau landslides are at least one order of
81 magnitude lower than the largest historical volcano flank failures that generated tsunamis, i.e.
82 1888 Ritter Island (5 km^3 : Cooke, 1981; Johnson, 1987), and 1741 Oshima-Oshima (2.4 km^3 :
83 Satake and Kato, 2001). Mass transport deposits on the seafloor and giant collapse scars on
84 the flanks of ocean islands with volumes of tens to hundreds of km^3 support the existence of
85 even larger events in the geological record (e.g. Moore et al., 1989; Carracedo et al., 1999;
86 Oehler et al., 2004; Masson et al., 2002, 2008).

87 The sizes of the clasts moved upward by a tsunami can be compared between modern,
88 historical, and geological events. Coastal boulders with a mass of ~50 to 700 tonnes have
89 been transported by modern and historical events such as the tsunamis in the Indian Ocean in
90 2004 (85 tonne, Sumatra: Paris et al., 2009), Tohoku-oki in 2011 (690 tonne, Tanohata: Iwai et
91 al., 2019), Krakatau in 1883 (200 tonne, Java: Verbeek, 1886), and Meiwa in 1771 (216 tonne,
92 Ishigaki Island: Goto et al., 2010), but all were deposited on the coastal lowland. Tsunami
93 waves caused by ocean island flank collapses have the potential to carry clasts that are
94 considerably heavier and higher above sea level. In the Cape Verde Islands, Ramalho et al.

95 (2015) reported boulders up to 700 tonnes, which were quarried from the edge of a palaeocliff
96 (presently at 160-190 m above present sea level (a.p.s.l.)) and transported upwards to 220 m
97 a.p.s.l. The likely cause was a mega-tsunami generated by a massive flank collapse of Fogo
98 volcano at ca. 70 ka (Day et al., 1999; Masson et al., 2008; Paris et al., 2011; Ramalho et al.,
99 2015). However, the largest known tsunami boulders have been found on Tongatapu Island,
100 Tonga (1600 tonnes: Frohlich et al., 2009) and Shimoji Island, Japan (2500 tonnes: Goto et
101 al., 2010).

102 In this paper, we briefly provide some background on pioneering studies of the sedimentary
103 signatures of mega-tsunamis generated by ocean island flank collapse and asteroid impact,
104 and we discuss the main conclusions and perspectives that must be drawn from studies of the
105 last decade. For a detailed review of the characteristics of mega-tsunami conglomerates on
106 ocean islands, we refer to Paris et al. (2018).

107

108 **Marine conglomerates at high elevation on the flanks of ocean islands**

109

110 *Debated evidence of mega-tsunamis in Hawaii...*

111 The origin of elevated marine conglomerates on the flanks of the Hawaiian Islands has long
112 been debated since Moore and Moore (1984) revisited the Hulope Gravel (southern LānaʻI, up
113 to 326 m a.p.s.l.), previously presented as an ancient littoral deposit (Stearns, 1938), and
114 concluded that it was actually laid down by a tsunami. Similar fossiliferous conglomerates in
115 MolokaʻI (Moore et al., 1994) and Big Island (McMurtry et al., 2004) later contributed to the
116 controversy. The different arguments in favour or against the tsunami hypothesis have been
117 widely discussed (e.g. Grigg and Jones, 1997; Felton et al., 2006; Crook and Felton, 2008)
118 and we refer to Paris et al. (2018) for a comprehensive review. The interpretation of the
119 conglomerates also needs to account for the history of vertical motion of the Hawaiian
120 Islands, and the dating of coral clasts in their deposits. In short, voluminous landslide and
121 slump deposits on the seafloor around Hawaii (e.g. Moore et al., 1989; Normark et al., 1993),
122 reefs drowned by long-term subsidence (e.g. Moore and Fornari, 1984; Webster et al., 2007),
123 and coeval ages of coral clasts from different islands (Rubin et al., 2000; McMurtry et al.,
124 2004) strongly support the hypothesis of mega-tsunamis related to flank failures.

125

126

127 *... later supported by unequivocal evidence in the Canary and Cape Verde Islands*

128 In parallel to studies in Hawaii, another “unusual” marine deposit was interpreted as being
129 caused by a mega-tsunami in the Canary Islands. In the Agaete Valley (western Gran
130 Canaria), Perez-Torrado et al. (2006) described a fossiliferous conglomerate attached to the
131 slopes at elevations ranging between 41 and 188 m a.p.s.l., a stratigraphical position that does
132 not fit into the framework of relative sea-level changes and uplift of Gran Canaria Island
133 (Meco et al., 2007), i.e., it cannot easily be interpreted as a paleo-beach deposit. Perez-
134 Torrado et al. (2006) thus concluded that it was most likely created by a tsunami caused by a
135 massive collapse, with the Güímar event on the eastern flank of Tenerife Island being the best
136 candidate.

137 During the last decade, mega-tsunami deposits were also found in the Cape Verde Islands
138 (Paris et al., 2011; Ramalho et al., 2015; Madeira et al., 2020), which is not surprising
139 considering the number of massive flank collapses that have affected these islands (Masson et
140 al., 2008), including the Late Pleistocene Fogo collapse (Day et al., 1999). On northern
141 Santiago Island, mega-tsunami evidence is provided by fossiliferous conglomerates and
142 boulders at elevations up to 100 m and 220 m a.s.l. respectively (Paris et al., 2011; Ramalho et
143 al., 2015). The Santiago conglomerates share many sedimentological characteristics with their
144 Hawaiian and Canarian counterparts: complex but diffuse internal organisation with a poor
145 lateral continuity of the subunits, poor sorting, landward and seaward imbrication of the
146 clasts, heterogeneous composition of locally derived volcanic rocks and mixed taxa of
147 shallow and deep-water fossils (never in life position, often fragmented) (Fig. 2), an erosive
148 base with rip-up clasts of the underlying substratum, and downward-injected veins of
149 conglomerates (clastic dykes) inside the palaeosol (Paris et al., 2011). The megaclasts
150 reported by Ramalho et al. (2015) were transported upwards from a cliff edge presently at
151 160-190 m a.p.s.l. (Fig. 2) to elevations up to 220 m a.p.s.l. (equivalent to a runup of 270 m
152 above coeval sea-level) and 650 m from their source. The largest megaclast measured
153 9.4×6.8×3.8 m and had an estimated mass of 700 tonnes. ³He exposure ages of the megaclasts
154 range between 65 and 84 ka, with an arithmetic mean age of 73 ka (Ramalho et al., 2015),
155 which is concordant with the age of the last massive flank collapse of Fogo Island volcano ca.

156 70 ka (Foeken et al., 2009; Cornu et al., 2017), and with the age of another tsunami
157 conglomerate recently identified on Maio Island (Madeira et al., 2020).

158 Quaternary deposits are particularly well preserved in the Cape Verde Islands, making them
159 useful for identifying tsunami deposits within a complex framework of elevated marine
160 terraces, littoral deposits, and alluvial fans (e.g. Madeira et al., 2020). For a detailed review of
161 the sedimentological criteria used for distinguishing mega-tsunami conglomerates from other
162 coastal deposits, we refer to Perez-Torrado et al. (2006) and Paris et al. (2011, 2018), but
163 many of the criteria listed by Paris et al. (2018) in their Table 1 are not specific to tsunamis.
164 Paris et al. (2018) identified three specific criteria: (1) the succession of landward and
165 seaward clast imbrication; (2) the increasing abundance of terrestrial material upward and
166 landward; (3) and the mixed and unusually rich fauna, ranging from terrestrial to circa-littoral
167 species (Fig. 2). Following these criteria, a number of mega-tsunami conglomerates have been
168 identified, not only in the Canary Islands (Meco, 2008; Paris et al., 2017) or the Cape Verde
169 Islands (Madeira et al., 2020), but also on Mauritius (Paris et al., 2013). In Maio, Madeira et
170 al. (2020) reported four distinct tsunamis occurring over the last 500 kyrs, including the Fogo
171 tsunami previously identified on Santiago by Paris et al. (2011) and Ramalho et al. (2015).
172 The runup on Maio was in excess of 60 m above coeval sea level at 120 km from Fogo
173 volcano, compared to a runup of 270 m on Santiago at 70 km from Fogo.

174

175 *Other candidate islands?*

176 With more than 40 flank failures identified over the last 2 Ma (Labazuy, 1996; Oehler et al.,
177 2004), Réunion Island represent a significant source of mega-tsunamis in the Indian Ocean.
178 However, there is so far only one published study on mega-tsunami deposits related to the
179 impact of a Réunion Island flank collapse on the southern coast of Mauritius Island, where
180 Paris et al. (2013) reported reef megaclasts and a tsunami conglomerate at elevations up to 40
181 m. The maximum age of the tsunami is given by a ^{14}C age of 4425 ± 35 BP on a coral branch,
182 which is concordant with that of the last flank collapse on the eastern flank of Piton de la
183 Fournaise volcano on Réunion Island (Labazuy, 1996; Oehler et al., 2004). The humid climate
184 of Mauritius and Réunion Islands does not allow tsunami deposits to be as well preserved as
185 in the Canary and Cape Verde Islands, but other mega-tsunami evidence might yet be
186 discovered in Rodrigues Island or Madagascar.

187 In the Atlantic Ocean, studies in the Cape Verde and Canary Islands should be extended to the
188 Azores (Andrade et al., 2006), where large-scale mass wasting has been documented,
189 especially on the southern flank of Pico Island (e.g. Mitchell et al., 2011; Costa et al., 2015).
190 However, there is so far no report of mega-tsunami deposits that could be related to flank
191 collapses in the Azores. The same is true for the southern Pacific Islands. Quaternary flank
192 collapses in the Society and Marquesas Islands (e.g. Clouard et al., 2001; Clément et al.,
193 2002; Hildenbrand et al., 2006) might have generated mega-tsunamis whose traces are yet to
194 be identified. Blahūt et al. (2019) recently created a worldwide database of “giant landslides
195 on volcanic islands” (<https://www.irms.cas.cz/ext/giantlandslides/index.php>) which can
196 provide a general framework for completing the catalogue of mega-tsunamis.

197

198 *Multistage flank collapses*

199 Following the seminal work of Perez-Torrado et al. (2006), Madeira et al. (2011) described
200 new sections in the Agaete Valley (Gran Canaria) and they found two other tsunami
201 conglomerates below the one described by Perez-Torrado et al. (2006). This succession of
202 three tsunami deposits in the same valley not only illustrates the recurrence of massive flank
203 collapses in the Canary Islands, but it also raises the following question: do ocean island
204 mega-failures collapse in one-go or retrogressively as multistage events separated by short
205 periods of time? Answering this question is of fundamental importance for evaluating the
206 mega-tsunami threat.

207 Giachetti et al. (2011) demonstrated that a multistage scenario of collapse (rather than in one
208 go) generates a tsunami large enough to explain the maximum inundation distance inferred
209 from the spatial distribution of tsunami deposits in the Agaete valley. Hunt et al. (2010, 2013)
210 used the sequences of distal turbidites to illustrate the multistage nature of the majority of the
211 Canary Islands flank collapses, including the Güímar collapse that produced the Agaete mega-
212 tsunami. On the northern flank of Tenerife, the Icod collapse is recorded offshore by three
213 debris lobes that were successively emplaced (Watts and Masson, 2001) and a stacked
214 sequence of seven turbidites (Hunt et al., 2011). The compositions of the successive turbidites
215 and their subunits vary from basaltic lavas of the submarine flank to phonolitic-trachytic lavas
216 of the subaerial edifice, recording how successive failures removed different parts of the
217 edifice (Hunt et al., 2011).

218 Based on the structure of the tsunami deposits in Santiago (Cape Verde) and available
219 geophysical data offshore, whether the Fogo collapse was multistage or not is far from evident
220 (Masson et al., 2008; Paris et al., 2011). However, as for the Agaete Valley in Gran Canaria
221 (Perez-Torrado et al., 2006; Madeira et al., 2011), numerical simulations show that tsunami
222 runup resulting from a multistage failure can reproduce the spatial distribution of mega-
223 tsunami deposits reported by Paris et al. (2011) and Ramalho et al. (2015), whereas massive
224 collapses (in one-go) may over-estimate the runup (Paris et al., 2018).

225

226 *Links between volcanic activity, flank instability, and mega-tsunami*

227 Causal links between some ocean island flank collapses and major explosive eruptions was
228 unclear until new evidence of a mega-tsunami combined with explosive activity was
229 demonstrated by Paris et al. (2017) in Tenerife. Indeed, the Icod collapse on the northern flank
230 of Tenerife was highly debated because its genesis was coeval with a major caldera-forming
231 volcanic eruption (El Abrigo eruption and the Las Cañadas caldera: Marti et al., 1994;
232 Ancochea et al., 1999; Edgar et al., 2007). The Icod collapse was a multistage retrogressive
233 event that mobilised a volume of ~200 km³ from both the submarine and subaerial flanks of
234 the island (Watts and Masson, 1995, 2001; Hunt et al., 2011). Paris et al. (2017) recently
235 proposed a scenario that links the successive flank collapses with the Abrigo eruption and two
236 major tsunamis at ca. 170 ka. Paris et al. (2017) used the composition of tsunami deposits on
237 the northwestern coast of Tenerife (at altitudes up to 132 m a.p.s.l.) to propose the following
238 scenario: an initial tsunami was generated during the submarine stage of the retrogressive
239 failure and before the onset of the Abrigo eruption, whereas a second and larger tsunami
240 immediately followed the debris avalanche of the subaerial edifice and formation of the
241 caldera. This original scenario of coupled flank collapse and explosive eruption seems to be
242 recurrent for the central volcanic edifice of Tenerife (Hunt et al., 2018), and it represents a
243 new type of volcano-tectonic event on ocean islands.

244

245 *Numerical modelling of ocean island mega-tsunamis: learning from historical examples of* 246 *landslide tsunamis*

247 Landslides and their resultant tsunamis could be an extreme hazard with potentially several
248 tens to hundreds of meters of local tsunami run-up height. Recent developments in numerical

249 simulation can help us to estimate the impact and affected areas of such events, although it is
250 a challenging exercise (e.g. Løvholt et al., 2015; Yavari-Ramshe & Ataie-Ashtiani, 2016). The
251 choice of parameters for numerical simulations of mega-tsunamis generated by large-scale
252 flank collapses of ocean islands is difficult because there is no instrumental or observational
253 data available. This is particularly well illustrated by the debate on the potential impact of a
254 tsunami generated by a large-scale collapse of the western flank of La Palma, Canary Islands
255 (e.g. Ward and Day, 2001; Løvholt et al., 2008; Abadie et al., 2012; Tehranirad et al., 2015).

256 The main factors used to determine the initial size of a landslide-induced tsunami are the
257 volume of landslide mass, water depth, velocity of movement, landslide scenario (in one-go
258 or multistage) and the types of landslide that are classified by movements such as rotational
259 landslide, debris flow or avalanche (USGS, 2006), and conditions (submarine or subaerial).
260 Compared to earthquake-induced tsunamis, landslide tsunamis can cause locally extreme run-
261 ups with a high wave energy caused by the relatively short wave periods that are generated
262 (Muhari et al., 2018). However, short period waves are likely to be significantly attenuated
263 over a short distance (Friz et al., 2004; Heidarzadeh & Satake, 2015; Heidarzadeh & Satake,
264 2017). To simulate landslide-induced tsunamis, it is necessary to develop models of both
265 landslide movement and the short-period tsunami waves. To explain the physics of landslide-
266 induced tsunamis, three types of models have been proposed.

267 The first type of model generates the tsunami waves by inputting an initial waveform
268 estimated by analytical, empirical or observational landform changes (e.g. Tappin et al., 2014;
269 Grilli and Watts 2005; Satake and Kato, 2001). This method is commonly used for
270 preliminary assessments because the model is easy and simple to use (e.g. Okamura et al.,
271 2018). The model can also be used for the calculation of tsunamis caused by landform
272 deformation such as caldera collapse (e.g. Ulvrova et al., 2016). However, it is difficult to
273 estimate the interactive process between landslide movement and the resulting tsunami wave.

274 The second type is a two-layer model of depth-averaged equations representing the tsunami
275 (water mass) and the landslide (soil mass) in upper and lower layers, respectively (e.g.
276 Imamura et al., 2001; Kawamata et al., 2005). This approach was applied to simulating
277 massive flank collapses and their resulting mega-tsunamis in Reunion Island (Kelfoun et al.,
278 2010), Tenerife (Giachetti et al., 2012; Paris et al., 2017), and Fogo (Paris et al., 2011). Since
279 this type of model considers the landslide as a fluid, it is adequate for simulating the tsunami
280 caused by a debris flow or debris avalanche. As examples, Ioki et al. (2019) simulated the

281 1741 tsunami caused by the sector collapse of the Oshima-oshima volcano in Japan using the
282 two-layer model with a Coulomb viscosity model to reproduce both the distribution of
283 landslide deposits and historical tsunami wave heights. Yanagisawa et al. (2018) simulated the
284 1792 tsunami caused by sector collapse of Unzen Mayuyama volcano in Japan using a two-
285 layer model. Their model predicted that a wave amplitude greater than 50 m would occur
286 soon after tsunami generation. These modelling results were well validated by historical
287 information of tsunami sediments (Imamura & Matsumoto, 1998; Tsuji et al., 2002; 2017).
288 We suggest that the two-layer approach should accurately model extremely large landslide
289 tsunamis.

290 The third type is a 3-dimensional (3D) model used to simulate landslide-generated impulse
291 waves. Mader (2002) simulated the 1958 Lituya Bay tsunami using the Navier-Stokes
292 equations for compressible fluid flow. The calculated maximum wave height in the bay was
293 about 250 m above sea level, which ran up to a height of 580 m, agreeing well with the
294 observed runup of 524 m (Miller, 1960). Franco et al. (2020) simulated this tsunami using the
295 Computational Fluid Dynamics (CFD) software Flow-3D, which computes the movement of
296 two fluids having a different density. Recently, methods that do not use gridding, such as
297 Smoothed Particle Hydrodynamics (SPH), perform well in simulating the complex multi-
298 phase flow of a landslide-generated tsunami (Xenakis et al., 2017). There is also a hybrid
299 using 2D and 3D models for landslide and water movement, respectively, which can
300 accurately simulate waves with short periods with a reducing computational load for the
301 landslide movement calculation (Ma et al., 2015; Grilli et al., 2019).

302 Although modelling is indeed useful, accuracy of the initial parameters of the landslide
303 remains controversial because these are usually difficult to observe directly or to infer from
304 geological data (e.g. geometry of the landslide scar, characteristics of the landslide deposits).
305 For the 1741, 1792 and 1958 events, observations and inundation data are used to validate the
306 model including, for example, the topography before and after sector collapses, the extent of
307 the landslide deposit and erosional area, historical tsunami heights, and tsunami deposits; and
308 it was therefore possible to infer the landslide parameters using these data. Since mega-
309 tsunamis are a low frequency hazard we only have rare opportunities to collect sufficient
310 evidence to determine most of the necessary parameters. With this in mind, Sassa et al. (2016)
311 enhanced the landslide parameters for the 1792 sector collapse of Unzen Mayuyama volcano
312 by conducting ring shear stress tests on soil from the landslide material. Although these results
313 are important as a fundamental dataset, the scaling of experimental results to in situ prototype

314 remains an issue. Therefore, it is important to collect as much accurate information as
315 possible, not only from experimental data but also from studies of recent events, such as the
316 2018 Anak Krakatau tsunami (Grilli et al., 2019; Muhari et al., 2019; Paris et al., 2019).
317 Although this recent landslide tsunami was not mega, the simulation of such events is useful
318 for developing reliable models that can be later applied to the study of potential
319 megatsunamis.

320

321 **Asteroid impacts and mega-tsunamis**

322 Asteroid impacts are extreme phenomena that infrequently punctuate Earth's history. They
323 can drastically change Earth's environment and climate and can lead to mass extinctions
324 (Alvarez et al., 1980; Schulte et al., 2010). Since major parts of the Earth's surface are
325 covered by oceans (~70%), many past (and future) impact events will have occurred in the
326 oceans and potentially generated mega-tsunamis. However, oceanic impact craters are
327 difficult to identify, so there are few examples recorded (they represent only ~20% of all
328 known impact craters, e.g., Jansa, 1993; Ormö and Lindström, 2000; Dypvik and Jansa,
329 2003). Many of them are undoubtedly waiting to be discovered (e.g., Nozaki et al., 2019),
330 although some oceanic impact craters may well have been eroded away or subducted into the
331 mantle over time. When seeking evidence for oceanic events, tsunami deposits with
332 associated traces of the impact are extremely useful (e.g., Gersonde et al., 1997).

333 Impact-induced mega-tsunamis can also occur on other planets and satellites if water or other
334 liquids exist on their surfaces. Within our solar system, Mars is of the greatest interest since it
335 has been suggested to have had a putative ocean in its past (Ormö et al., 2004; Iijima et al.,
336 2014; Costard et al., 2017, 2019). While the presence of an ocean on Mars is still under debate
337 (e.g., Parker et al., 1993; Tanaka, 1997), the identification of Martian oceanic impact craters
338 and tsunami deposits has the potential to assist further discussion about the ocean's possible
339 existence, its depth and lifespan. On this point, the understanding of the oceanic impact
340 process on Earth can provide crucial information for the identification of similar features on
341 other planets and satellites.

342 Among oceanic impacts on Earth, that caused by the Chicxulub asteroid, which struck the
343 Yucatàn Peninsula, Mexico, ca. 66 Ma, is currently the largest known impact. The
344 paleontological evidence of it defines the Cretaceous/Paleogene (K/Pg) boundary and the

345 impact is now widely accepted as the major trigger of a mass extinction at this time (Schulte
346 et al., 2010). The K/Pg boundary tsunami deposits have been studied for more than 30 years
347 since Bourgeois et al. (1988). Also, the Eltanin asteroid at 2.51 Ma (Kyte et al., 1988; Goff et
348 al., 2012) is another interesting example of an oceanic impact event that was recognized in the
349 absence of an impact crater. The research histories of the Chicxulub and Eltanin events and
350 the resultant tsunami deposits are interesting examples that may serve to guide future work in
351 identifying undiscovered oceanic impacts and resultant generation of mega-tsunamis.

352

353 *The K/Pg boundary tsunami deposits*

354 Since it was first proposed by Alvarez et al. (1980), there have been numerous discussions
355 about the K/Pg impact and extinction event. The most serious weaknesses of the impact
356 hypothesis in the 1980s was the lack of conclusive evidence of an impact crater. The K/Pg
357 boundary deposits, which were reported during the 1980s mostly around Europe and North
358 America, are usually a few mms to cms in thickness (e.g., Alvarez et al., 1980; Smit et al.,
359 1980). During the late 1980s, Bourgeois et al. (1988) reported meter thick tsunami deposits on
360 the K/Pg boundary at the Brazos River in Texas, USA. It is interesting to note that this impact
361 tsunami work was being reported at almost exactly the same time as the first scientific papers
362 about Holocene paleotsunami deposits were being reported in the Pacific Northwest, USA
363 (Atwater, 1987), Japan (Minoura et al., 1987) and Europe (Dawson et al., 1988). The presence
364 of impact-induced tsunami deposits around the Gulf of Mexico indicated that the impact must
365 have occurred close to the Gulf coast (e.g., Hildebrand and Boynton, 1990; Maurrasse and
366 Sen, 1991). Finally in 1991, Hildebrand et al. (1991) identified the Chicxulub crater on the
367 northern flank of the Yucatàn Peninsula. The impact site was on the Yucatàn carbonate
368 platform and was interpreted as having occurred in relatively shallow water (~200 m, Matsui
369 et al., 2002).

370 After the discovery of the Chicxulub crater, during the 1990s and 2000s, numerous K/Pg
371 boundary offshore deposits were reported around the Gulf of Mexico (Maurrasse and Sen,
372 1991; Alvarez et al., 1992; Smit et al., 1992, 1996; Smit, 1999) and the proto-Caribbean Sea
373 (e.g., Takayama et al., 2000). These offshore deposits have similar sedimentary features that
374 can be characterized by some of following: 1) basal erosional contact, 2) large rip-up clasts or
375 coarse basal spherule layer, 3) multiple sets of bi-directional cross laminations that indicate
376 repeated reversal of flow directions, 4) massive but monotonous upward fining units with

377 slight variations in grain size and/or composition, 4) they are a meter to few hundred meters
378 thick, and 5) there is an abundance of organic matter such as wood (e.g., Smit et al., 1992,
379 1996; Smit, 1999; Tada et al., 2003; Schulte et al., 2012). Based primarily on these
380 characteristics, these offshore deposits have been interpreted as having a tsunami origin (e.g.,
381 Smit, 1999). On the other hand, other researchers have suggested that these deposits can
382 alternatively be interpreted as impact-induced turbidites or submarine landslide deposits
383 (Bohor et al., 1996; Bralower et al., 1998) or the combination of lower gravity flow deposit
384 and upper thick tsunami deposit (Kiyokawa et al., 2002; Tada et al., 2003; Goto et al., 2008).

385 Since there are very few reports of modern tsunami deposits in the deep ocean, but they seem
386 likely to have been emplaced, the differentiation between tsunami deposits and turbidites
387 remains an important issue for paleotsunami researchers. Indeed, submarine tsunami deposits
388 formed by recent events have only been reported since the 2000s (e.g., Noda et al., 2007),
389 although some possible Holocene ones were reported in the Mediterranean Sea (Kastens and
390 Cita, 1981; Cita et al., 1996). Note that the potential offshore K/Pg tsunami deposits were
391 studied and interpreted without any modern analogues. Following the recent 2011 Tohoku-oki
392 tsunami, there were some reports concerning deposits formed in 100-6000 m water depth
393 (Ikehara et al., 2014; McHugh et al., 2016). Interestingly, some of these deposits are
394 interpreted as being a combination of a lower turbidite triggered by the earthquake or tsunami
395 and an upper tsunami deposit (Ikehara et al., 2014; Usami et al., 2017) with peaks of Cs
396 concentration, like the iridium anomaly of the K/Pg boundary deposits, released from the
397 damaged Fukushima Daiichi Nuclear Power Plant (Ikehara et al., 2014). This interpretation is
398 similar to those of the K/Pg boundary reported in the proto-Caribbean sea (Takayama et al.,
399 2000) although the 2011 deposits are significantly thinner.

400 More recently, DePalma et al. (2019) reported an onshore K/Pg surge deposit in North
401 Dakota, USA. The site is located ~3000 km away from the impact site near the narrow seaway
402 that connected to the Gulf of Mexico during the Cretaceous (DePalma et al., 2019). They
403 proposed that the deposit was formed not by the tsunami generated by the impact but by an
404 impact-induced seismic seiche similar to those observed in Norway following the 2011
405 Tohoku-oki event (Bondevik et al., 2013).

406

407 *K/Pg tsunami generation mechanisms*

408 Another important discussion concerns how the tsunami was generated by the Chicxulub
409 impact. Bralower et al. (1998) suggested that the tsunami could have been generated by an
410 impact induced landslide. Matsui et al. (2002) suggested mainly three generation mechanisms
411 of the K/Pg impact-induced tsunami: 1) a rim wave formed from the collapse of the large
412 splash caused by the impact, 2) impact-induced landslide, and 3) ocean water flowing into
413 (resurge) and flowing out from the crater. Tsunamis could have been generated by all of these
414 processes simultaneously with differing intensities and waveforms. For instance, Matsui et al.
415 (2002) suggested that the rim wave might have had only a small impact along the Gulf coast.
416 This is because a rim wave is characterized by a high wave amplitude with a short wave
417 period at the impact site. The tsunami may be very large around the proximal impact site but it
418 is unstable and dissipates quickly. An impact-induced landslide also has the potential to
419 generate a large tsunami that is unidirectional. Namely, if the submarine landslide were
420 generated off the northern flank of the impact crater (Bralower et al., 1998), a large tsunami
421 would have been preferentially propagated northwards.

422 The resurge process may induce the movement of a significant volume of ocean water
423 (Matsui et al., 2002). If this was indeed generated during the Chicxulub impact, wave periods
424 tend to be long (10 h according to Matsui et al., 2002) and stable, propagating through the
425 ocean and becoming very large along the coast. Matsui et al. (2002) estimated that the
426 maximum run-up height around the Gulf of Mexico could have reached as high as 300 m and
427 this could explain the ocean-wide distribution of tsunami deposits in the Gulf of Mexico and
428 proto-Caribbean Sea.

429 The resurge should rework fallback ejecta inside the crater and so the process is testable using
430 core samples from the site. Goto et al. (2004) and Smit et al. (2004) studied the ICDP YAX-1
431 cores recovered 65 km from the crater centre (near the crater rim) and found sedimentary
432 features with influence of flow such as climbing ripples and repeated upward fining and
433 coarsening units; they suggested that resurge did indeed occur following the Chicxulub
434 impact. However, others have suggested that a large portion of the inferred resurge deposit
435 can be interpreted as fallback ejecta (e.g., Wittmann et al., 2007). Also, if the crater rim rose
436 high above sea level then it may have prevented ocean water flowing into the crater (Bahlburg
437 et al., 2010). Since the water depth of the impact site was considered to be shallow, numerical
438 modeling results indicate that only a low energy resurge might have occurred locally
439 (Bahlburg et al., 2010). On the other hand, the resurge process might have been stronger than
440 previously thought according to the recent studies. Gulick et al. (2008, 2019) suggested that

441 water depths around the impact site would have been much deeper than previously thought. In
442 2016, IODP Expedition 364 recovered continuous cores at the peak ring inside the crater (45.6
443 km from the crater centre). Gulick et al. (2019) studied the sedimentology of these cores and
444 based on the presence of laminations and variations in grain size, suggested that resurge
445 deposits were also present. The presence of inferred resurge deposits at the two separate sites
446 near the crater centre and rim suggests that this process was large and occurred at a crater-
447 wide scale.

448 The tsunami generation process of the Chicxulub impact is still under discussion and more
449 research including numerical modeling for sediment transport should be carried out in the
450 future.

451

452 *The Eltanin impact event*

453 While it was originally thought that the Eltanin asteroid impact occurred around 2.15 Ma, its
454 date has subsequently moved back to 2.51 ± 0.07 Ma, contemporaneous with the Pliocene–
455 Pleistocene boundary at 2.58 Ma (Goff et al., 2012). Unlike Chicxulub, this was a deep ocean
456 (4–5 km) impact that struck in the Southern Ocean some 1500 km SSW of Chile.
457 Furthermore, unlike Chicxulub there is no seafloor crater with the impact being identified
458 through traces of intense erosion and the presence of meteoritic material in sedimentary rocks
459 collected 500 km apart (Gersonde et al., 1997). The absence of a physical crater and the
460 distribution of meteoric material have allowed researchers to estimate the asteroid diameter at
461 between 1 and 4 km. However, there is considerable debate concerning this diameter, which
462 has significant implications for numerical modelling of any resultant tsunami (e.g.
463 Korycansky & Lynett, 2005; Ward & Asphaug, 2002; Weiss et al., 2006; 2015). There are
464 significant disparities between most of the model outputs, but recently Weiss et al. (2015)
465 have produced the most comprehensive study using an assumed diameter of a mere 750 m.
466 This resulted in wave amplitudes between 8 and 10 m in southern Chile to less than a meter in
467 northern Chile. Needless to say, if a larger diameter is used, amplitudes are larger. These
468 results have been used to suggest that much of the proposed physical evidence for a circum-
469 South Pacific tsunami generated by the impact is incorrect. The size of the modelled wave
470 means that it could not transport much of the material found in the proposed coarse-grained
471 tsunami deposits.

472 The geological evidence is compelling, however, so the problem lies mainly with the
473 modelling of asteroid impact tsunamis. Modelling waves produced by large sized impactors
474 becomes increasingly complex and as such it is simpler to work with a smaller diameter (Goff
475 et al., 2012). Numerous arguments have been made for the inferred size of the Eltanin asteroid
476 but ultimately it seems most reasonable to place it somewhere between 1 and 4 km and as
477 such we are still unclear about the possible maximum size of the tsunami wave amplitude
478 along the Chilean and other South Pacific coastlines.

479 Possible Eltanin impact related deposits have been reported around the Pacific Ocean from as
480 far north as Japan and as far south as Antarctica, with the densest concentration being along
481 the Chilean coast, largely associated with either the Ranquil or Pisco Formations (Goff et al.,
482 2012). Some of these deposits have been dismissed as tsunamites through both numerical
483 modelling and geological reassessment, the most notable being those at Hornitos where the
484 tsunamite was reinterpreted as a debrite (Spiske et al., 2014). Most recently though, Le Roux
485 (2015) has questioned this reinterperatation stating that it may well still be a tsunami deposit
486 but that it could be 5.3-7.2 Ma old and therefore not related to the Eltanin asteroid impact.

487 Geological evidence for the Eltanin tsunami from other sites range from boulder deposits and
488 associated bonebeds of marine fauna, to coarse bioclastic sands, soft sediment deformation
489 features, rip-up clasts, unconformable contacts and deeply channelled erosional features (Goff
490 et al., 2012). This evidence generally mirrors that used to positively identify modern tsunami
491 deposits, but of particular note are the unusual bonebed deposits noted by Gersonde et al.
492 (1997) from Peru and Chile. These bonebeds contain the skeletons of rorqual whales, sperm
493 whales, seals, aquatic sloths, walrus-whales, predatory bony fish and many other species
494 (Pyenson et al., 2014). The end of the Pliocene was marked by an extinction event among
495 marine megafauna (mammals, seabirds, turtles and sharks) with 36% of Pliocene genera
496 disappearing including apex predators such as *Carcharocles megalodon* (the megashark) the
497 largest shark that ever lived (Pimienta and Clements, 2014; Pimienta et al., 2017). It also
498 marks the last appearance date of numerous marine microfossils (Berger, 2011). This
499 prominent die-off of marine species may well be linked to severe acoustic trauma or shock
500 waves associated with a deep ocean asteroid impact (e.g. Ketten, 1995). The
501 contemporaneous combination of bonebeds and sedimentary evidence make a compelling
502 case for an Eltanin tsunami that was larger than current modelling scenarios suggest.

503 Like Chicxulub, the tsunami generation process of the Eltanin impact is still under discussion
504 and more research, particularly with respect to combined bonebed-possible tsunami deposit
505 sequences, should be carried out in the future.

506

507 **Hazard assessment**

508 Assessing hazards related to mega-tsunamis generated by volcano flank collapse or asteroid
509 impact amounts to a reconciliation between the human and geological perspectives.
510 Considering the low frequency of such events and the high cost of monitoring, one could
511 conclude that the implementation of any form of prevention strategy is not cost-effective.
512 However, whether or not this is the case, it is the role of geologists to improve society's
513 awareness of such extreme hazards.

514 Addressing mega-tsunami hazards related to volcano flank instability could start with an
515 adequate monitoring strategy and a regional collaboration between both the volcano
516 observatories and the tsunami warning systems. The monitoring of active volcanoes such as
517 Kilauea in Hawaii and the Piton de la Fournaise in Reunion Island now offers the possibility
518 to detect small flank displacements (e.g. using GNSS and InSAR), but the scientific
519 community lacks milestones to anticipate large-scale flank collapses of ocean islands. The
520 major eruption of Piton de la Fournaise volcano in April 2007 was associated with a large (up
521 to 1.4 m horizontally) seaward motion of its eastern flank along a detachment fault (Froger et
522 al., 2015). Although no collapse occurred, it illustrates the relevancy of InSAR for an early
523 detection of ground motion on volcanoes. Earth observation programs such as Copernicus
524 now give access to worldwide data from satellites, including radar imaging that can be rapidly
525 post-processed to produce interferograms. The second step is the definition of scenarios to be
526 used in probabilistic tsunami hazard assessment (PTHA) and Tsunami Early Warning Systems
527 (TEWS). Given the uncertainty of the collapse mechanisms, numerical models could yield
528 unrealistic results and any conclusions concerning hazard assessment should be viewed with
529 caution.

530 Similar uncertainties weigh on the assessment of tsunamis generated by asteroid impacts. For
531 instance, Ward and Asphaug (2003) assumed an oceanic impact of a 1950DA asteroid with 1.1
532 km diameter and found that the east coast of the United States would be inundated by a
533 tsunami over 100 m high if it struck the Atlantic Ocean. Wünnemann et al. (2007) performed

534 numerical modelling of the tsunami generation process from an small asteroid impact, which
535 predicted that the rim wave decays quickly so that it can't be treated as a long wave. Gisler et
536 al. (2011) suggested that even a relatively large asteroid with <500 m diameter hit the ocean at
537 a depth of <5 km depth, tsunamis would quickly dissipate and around adjacent shorelines the
538 wave height would be smaller than the 2004 Indian Ocean tsunami. These results suggest that
539 large tsunamis with long wavelengths may not be generated when the impact site is deep
540 and/or the impactor is small. Also, if the impact site is shallow and a high crater rim is
541 formed, then large tsunamis may not be generated as well. Future investigations of the most
542 effective conditions for maximizing the size of impact tsunamis need to be carried out (Goto
543 et al., 2013).

544 For hazard assessment, probabilistic analysis with information on potential impact size and
545 frequency is important (Ward and Asphaug, 2000). According to Chapman and Morrison
546 (1994), the recurrence interval over which an asteroid with a 100 m diameter hits the Earth is
547 about 10^3 - 10^4 years and with a 1000 m diameter it is about 10^5 - 10^6 years. While an asteroid
548 with a diameter of a few hundred meters may break up in the atmosphere so that the impact
549 frequency of small asteroids may well be lower, this effect is less likely for large asteroids
550 (Bland and Artemieva, 2003, 2006). Bland and Artemieva (2006) estimated that an asteroid
551 with a 200 m diameter, which may generate a hazardous tsunami, will hit the Earth's surface
552 every $\sim 10^5$ years, which is a far lower frequency than Ward and Asphaug (2000)'s estimated
553 (3000-4000 year interval for a 220 m diameter impactor).

554 Goto et al. (2013) carried out a preliminary evaluation of the tsunami hazard for the Japanese
555 coast from the impact of a 500 m diameter asteroid hitting 5000-m-deep ocean. In this case,
556 the impact tsunami was characterized by short wavelengths with high amplitudes. The
557 amplitude reached nearly 80 m in the deep ocean, decaying quickly to only a few meters when
558 it reached 50 m water depth. They proposed that the tsunami broke on the continental shelf far
559 away from the shoreline so that it was considerably attenuated prior to striking the coast.
560 Importance of wave breaking is known as the "Van Dorn effect" (Melosh, 2003) and is
561 important for tsunami hazard assessment of coastal communities.

562

563 **Conclusions: the future hazard of mega-tsunamis**

564 There is a risk of mega-tsunamis generated by ocean island flank collapses or asteroid impacts
565 in the future. Although they occur with a very low frequency compared to earthquake-induced
566 tsunamis, they can potentially be a significantly large hazard. It is extremely difficult to assess
567 and mitigate the risks associated with such high-magnitude but low-frequency hazards. Mega-
568 tsunamis are treated as geophysical curiosities rather than true threats, probably because their
569 recurrence interval far surpasses that of political and economic development strategies.
570 However, both the 2004 Indian Ocean and 2011 Tohoku-oki earthquakes raised awareness of
571 global disasters and their associated cascading risks. In this context, addressing future mega-
572 tsunami hazards requires international collaboration.

573 To end on a positive note, García-Olivares et al. (2017) used a combination of geological and
574 mtDNA data to demonstrate that ocean island flank collapses may be drivers of island
575 colonization, and subsequent lineages of diversification. Extreme geophysical events can thus
576 have positive biotic consequences that balance out their catastrophic repercussions.

577

578

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- 1066

1067 **Figure captions**

1068

1069 Fig. 1 – Tsunami size versus event probability (modified after Kawata, 2003), as seen from
1070 two different perspectives: disaster management, and geology.

1071

1072 Fig. 2 – Offshore to inland sediment transport during a mega-tsunami (here the source of the
1073 tsunami is a massive flank collapse from an oceanic island). A: the flank collapse from
1074 the source island produces a submarine debris avalanche with a volume of tens to
1075 hundreds of km³, and the mega-tsunami starts propagating in the ocean until it strikes
1076 another island (runup island); B: tsunami starts mobilising sediment on the shelf,
1077 including boulders; C: as the inundation progresses, sediment from different sources are
1078 mixed until a maximum altitude is reached (runup); D: tsunami backwash implies a re-
1079 mixing of the sediments deposited during the uprush inundation; E: idealized cross-
1080 sections of sedimentary deposits resulting from a mega-tsunami generated by an ocean
1081 island flank collapse. Colours correspond to successive sediment sources from left
1082 (offshore) to right (inland).

1083

1084 Fig. 3 – Sediment transport processes acting during an impact tsunami. A: The impact of the
1085 asteroid with the water body (ocean, lake, etc.) normally generates a crater that
1086 corresponds to the initial stage of the tsunami, as noted for explosion-generated
1087 tsunamis. B: The water crater then collapses and waves start propagating radially
1088 around the impact, while a large plume of ejecta is formed. C: While tsunami is
1089 propagating, seismic wave associated with the impact may trigger debris flows on
1090 submarine slopes. Ejecta are transported by density currents (collapsing plume) over the
1091 surface of the ocean, and then reworked as submarine suspension clouds. Wave ripples
1092 are formed in deep-water by the passage of the tsunami (e.g. Hassler and Simonson,
1093 2001). D: Tsunami breaks on the shelf and erodes the substrate. Sediments deposited
1094 during inundation have a mixed marine – continental composition, including ejecta from
1095 the impact. E: A second generation of submarine debris flows is generated during the
1096 backwash of the tsunami, and oscillations produces compositional variations of the

1097 deep-sea sedimentary sequences (e.g. Goto et al., 2008). For more details on stages D
1098 and E we refer to figure 1.