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(Article begins on next page)

Factors contributing to volcano lateral collapse

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Abstract

Many factors can lead to volcano lateral collapse, which can produce devastating debris avalanches that travel up to several tens to over 100 km and cover hundreds to more than a thousand km² with debris. Volcanic lateral collapses are severe hazards because of their destructive power and size, and sudden onset. Although their frequency of occurrence is not as high as those of smaller volcanic mass movements, such as rock falls and lahars, globally large collapses ≥ 0.1 km³ have occurred at least five times per century over the last 500 years. A large variety of destabilizing factors such as over-steepened slopes, magma intrusions, hydrothermal activity, climate fluctuations, deformation of the basement, and faulting can create the conditions for volcano collapse. Once a volcano reaches its critical point, a mechanism is necessary to trigger the failure event. We present the state-of-the-art of the knowledge acquired in the last few decades concerning the causes of large-scale volcanic failures to better understand the triggers, preparatory factors, and timing of volcano lateral collapse.

Keywords: volcanic lateral collapse; debris avalanche; instability factors; triggering mechanisms

1 - Introduction

36 Large long-lived composite volcanoes are positive topographic structures
37 characterized mainly by constructive phases. The stability of the edifice is primarily
38 controlled by the characteristics of the outer slope, which, when it reaches a critical
39 configuration, can become unstable. Edifice failure may also be triggered by external
40 processes even before it reaches its critical configuration. Slope instability in volcanoes
41 typically results in a structural failure leading to lateral collapse. Volcano flank movement
42 can be classified as: i) persistent flank motion, typically deep-seated, steady-state
43 movement of large sectors of a volcano edifice due to gravity; ii) transient flank motion (i.e.
44 flank “unrest”), considered the precursor to catastrophic collapses, associated with intrusive
45 processes, co-eruptive deformation, seismic shaking; or iii) surficial slope motion, or shallow
46 motion such as material erosion and remobilization of volcanoclastic and lava deposits
47 (Schaefer et al. 2019). Edifice collapses are not restricted to stratovolcanoes, but have
48 occurred on all types of volcanoes (dome complexes, shield volcanoes, calderas and even
49 within monogenetic fields) (Dufresne et al. 2020; this volume) and may happen in all
50 geodynamic contexts (Ui 1983; Siebert 1984; McGuire 1996; Voight and Elsworth 1997;
51 Voight 2000; Shea and Van Wyk de Vries 2010; Van Wyk de Vries and Davies 2015).

52 Such volcano sector collapses leave behind a scar that could be of a caldera size
53 (kilometers across) and shape (Siebert, 1984; McGuire 2006). This might explain why, before
54 the 1980 Mount St. Helens event, many wide depressions hosted in volcanic summits were
55 considered calderas (Siebert and Roverato 2020 - this volume), which form by vertical
56 subsidence along ring faults due to eruptive events, rather than by a lateral collapse. This
57 erroneous interpretation is understandable because horseshoe-shaped scar can be partly
58 obliterated by volcano regrowth covering the failure area with new eruption deposits.
59 Belousov et al. (1999) wrote: “Evaluation of known cases of large-scale failures on volcanoes
60 throughout the world shows that failure can occur, given the right combination of
61 circumstances, at almost any reasonably high volcanic edifice”. It appears that this behavior
62 is ubiquitous, with evidence for volcanic failures both within the geological record and at
63 many currently active volcanoes. For example, Francis (1994) wrote that 75% of Andean
64 volcanic edifices, with heights greater than 2,500 m, have experienced at least one sector
65 collapse, while in Indonesia, most volcanoes with heights greater than 2,000 m have
66 suffered the same fate (MacLeod 1989). Inokuchi (1988) reported that over 100 volcanic
67 debris avalanche deposits (VDADs) have been identified around Japanese Quaternary

68 volcanoes. Volcanic collapses, and their associated VDAs, are severe potential hazards not
69 only because of their destructive power and size but, as stressed by Siebert and Roverato
70 2020 – this volume), because volcanic edifice failures $\geq 0.1 \text{ km}^3$ have occurred at least five
71 times per century over the last 500 years. Augustine volcano (Alaska, USA) has repeatedly
72 collapsed, once every 150-200 years during the past 2,000 years (Begét and Kienle, 1992),
73 although those were at relatively “small” volumes ($< 0.25 \text{ km}^3$). Mount Taranaki (New
74 Zealand) shows a similar pattern of growth and collapse with 14 VDADs over its 120,000
75 years evolution with a regular 16,000 years return period (Zernack et al. 2012). Colima
76 volcano collapsed at least nine times during the last 20,000 years (Capra et al. 2002). This
77 frequency and cyclic behavior should not be underestimated for hazard assessment.

78 The study of this type of phenomenon strengthened after the collapse and eruption
79 of Mount St. Helens volcano (USA) on 18 May 1980. This event represents the only case
80 where lateral collapse of the volcanic edifice has been observed directly (Lipman and
81 Mullineaux 1981). Although the eruption of Mount St. Helens represented an event of great
82 importance for the understanding of such phenomenon linked to the collapse of a volcanic
83 edifice, the factors that prepare and trigger these events are not simple, but can be multiple
84 and combined. The effects of a volcanic lateral collapse are usually of great magnitude.
85 Mass movement on volcanic slopes can occur at different scales, which can span from
86 relatively minor rock falls consisting of a few hundred to a few thousand cubic meters of
87 material up to the giant landslides that may involve materials more than $1,000 \text{ km}^3$
88 (McGuire 1996, 2003). Events involving small volumes of material may occur with time
89 intervals of weeks or months, whereas larger events are having frequencies of 10s to 100s
90 of thousands of years (Schaefer et al. 2019). The large volumes observed in many volcanic
91 debris avalanches involve the collapse of part of a volcanic edifice and this can in turn
92 influence the runout of a VDAD. These destructive flows can travel at high velocities, cover
93 large areas, overcome topographical barriers and modify the topography in seconds.
94 Generally, these deposits could preserve evidence of the cause of the instability, but it is
95 often difficult to determine the triggering mechanism. Volcanic lateral collapses can be
96 caused by a wide variety of destabilizing factors such as over-steepened slopes, increasing
97 fluid pore pressures (consequence of several combined factors), magma intrusions,
98 hydrothermal alteration, climate fluctuations, deformation of the basement, cataclysmic
99 eruptions, among others (Table 1). It is important to identify the differences between those

100 factors that define the degree of instability of a volcanic edifice from those mechanisms that
101 trigger the collapse. Indeed, the mechanism that triggers a sector collapse could be
102 independent of the causes that prepare its instability. For example, the tectonic setting of a
103 volcano may influence the direction of the collapse and prolonged hydrothermal alteration
104 can deeply weaken the edifice, but the collapse itself could be triggered by an eruption or
105 an earthquake.

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Table 1

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117 **2 - Instability factors**

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Sector collapses can cause tsunamis or lahars increasing the areas and population at risk. In 2018, tsunamigenic lateral collapse of Anak Krakatau, Indonesia (Williams et al. 2019) and lateral collapse of Kilauea, US (Chen et al. 2019) and Etna, Italy (Bonforte et al. 2019) occurred, with resulting loss of life and property. It is therefore important to understand the instability factors and triggers of volcano lateral collapse since that knowledge can provide very important precursory information that can directly improve hazard warnings (Capra 2006).

A wide range of factors may contribute to the instability of a volcanic edifice, and some volcanoes are more prone to lateral failure than others, depending on their structure and the environment in which they form (McGuire 2003). Large-scale structural instability is invariably confined to major composite volcanoes, located either on continental (e.g. Shiveluch in Russia, Taranaki in New Zealand, and Colima in Mexico) or oceanic (e.g. Teide in Canary Islands - Spain, and Piton de la Fournaise on Reunion Island) crust. Destabilization of a volcanic edifice may take thousands or tens of thousands of years, or it may be achieved in only a few months (McGuire 2003). One well-known example of short-lived destabilization-to-collapse event is the Mount St. Helens collapse, where a cryptodome intruded into the edifice only few months before the collapse. That is in contrast to slow, incremental and progressive factors, such as long-term hydrothermal alteration, or long-lasting deformation that can occur individually or often act together as cumulative effects of many small events. Below, we describe the instability factors that favor the weakening of a volcanic edifice and promote sector collapses.

2.1 Basement, tectonics, and faults

Volcanism is linked to tectonically active regions, whether extensional, compressional or along transcurrent faulting, in both transcompressional and transtensional regimes (Tibaldi et al. 2005). Tectonic faults propagate from the substrate and into the volcano edifice, where their geometry and kinematics become more complex. A close relationship exists between fault movement and the stability of the overlying volcanic edifice: basement faults affect volcanoes (Francis and Self 1987; Carracedo 1994; Tibaldi 1995) and vice versa (van Wyk de Vries and Merle 1996; van Wyk de Vries 1998). Volcano edifice morphology, including collapse directions and fracture planes used as pathways for magma, respond to regional stress fields in complex ways (Delaney et al. 1986). Preferential directions of progressive flank destabilization depend not only upon the orientation, kinematics, and dip and inclination of basement faults (Tibaldi et al. 2005), but also on the position of the fault within the edifice (Norini et al. 2008; Wooller et al. 2009; Norini et al. 2020). Once volcanic failure occurs, the collapse depression and the new geometry of the volcanic edifice strongly influence the location and configuration of new craters and dykes.

Nakamura (1977) suggested that cone elongation, strike of dikes and fractures, and alignment of flank vents occur parallel to the regional maximum compressive stress direction (σ_{Hmax}) (Figurea). Moriya (1980) suggested that the elongated growth of volcanic flanks parallel to dike emplacement, steeper slopes normal to the elongation direction of volcanoes, and the dilatational effect of dike intrusion all promote failure perpendicular to σ_{Hmax} , as was observed at 30 Japanese volcanoes. The higher percentage of volcanoes with parallel dike swarms that formed amphitheater-like scars relative to volcanoes without a dominant direction of dike orientation indicates that dike swarms that develop parallel to σ_{Hmax} play an important role in volcanic lateral collapse (Siebert 1984). Francis and Wells (1988) suggested a model of failure scar opening perpendicular to the strike of normal faults and σ_{Hmax} (Figureb).

These models, however, are not consistent with the alignment of volcanoes, dikes, flank vents, and collapse directions in Indonesia, where they tend to be at a high angle to σ_{Hmax} as determined from fault-fold structures in the region (Bahar and Girod 1983). An examination of 39 Japanese volcanoes revealed failure scar openings at an acute angle relative to σ_{Hmax} (Ui et al. 1986). The same is true for the 28 examples of breached cones in the Central Andes (Figurec) (Francis and Wells 1988).

164

165

Figure 1

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167 The inconsistencies in these studies led to the idea that the opening direction may
168 not be influenced by the regional tectonic field, but may be controlled by temporal and local
169 stress fields within the volcanic edifice or the different geodynamic settings where
170 volcanoes are located (Ui et al. 1986).

171 In a study of 1,315 Quaternary volcanoes in different structural settings, Tibaldi
172 (1995) established that failure scar breaching perpendicular to fault strike is prevalent in
173 extensional regions, and collapses are usually in the direction of the downthrown block
174 (Vidal and Merle 2000). This is consistent with the model of Moriya (1980) and Siebert
175 (1984), where plate-shaped conduits and dikes could propagate through extensional faults
176 and thereby promote collapse perpendicular to the regional σ_{Hmax} . Also, in extensional
177 regimes, the regional stress field produces brittle discontinuities that crosscut the edifice by
178 forming rift zones, used as a pathway for the magma to reach the surface. This feeds dyke
179 swarms in a preferred orientation, which then displace the volcano flanks and may induce
180 lateral collapse of the edifice (Elsworth and Voight 1996; Tibaldi 1996).

181 In contrast to extensional regions, breaching in volcanoes along transcurrent faults
182 or transtensional tectonics is at an acute angle relative to σ_{Hmax} (Tibaldi, 1995; Norini et al.,
183 2008). This is true for the volcanoes in Southwest and Southeast Luzon, Philippines including
184 Iriga (Figure 1d; Lagmay et al. 2000; Lagmay and Valdivia 2006; Paguican et al. 2012), as well
185 as Mount St. Helens, USA. Lateral collapses in transtensional tectonic regions underlain by
186 normal faults are more diffused compared to transcompressional and reverse tectonic
187 regions due to tilting and topographic offsets within the basement (Tibaldi 2005). Analogue
188 experiment by Lagmay et al. (2000) and Norini and Lagmay (2005) showed that edifices on
189 top of strike-slip faults form a pair of sigmoids composed of one reverse and one normal
190 fault on the surface, and a flower structure in the interior of the edifice. Two destabilized
191 regions characterized by bulging and landsliding are created on the flanks between the
192 traces of the sigmoidal faults. Sigmoidal summit depressions, fractures, and faults
193 developed within the edifice create space for subsequent magma intrusion, thus causing
194 further instability (van Wyk de Vries and Merle 1996).

195 The variety of fault types and geometries has a general tendency to produce regions
196 of instability parallel to fault strike, irrespective of the fault motion. Analogue experiments
197 by Wooller et al. (2009) showed that in oblique-slip faulting, the instability is always on the
198 downthrown side, and usually in the volcano sector facing the strike-slip sense of motion
199 (Figure 2). The location, type, and magnitude of instability of the edifice changes with the
200 position of the fault beneath the analogue volcano. The further the fault is from the central
201 axis, the larger the destabilized sector. Also, with greater fault offset from the central axis,
202 larger unstable volumes are generated. Such failures are perpendicular to the strike of the
203 underlying fault.

204 These models may not work in volcanoes where an underlying sloping substrate or
205 gravitational spreading have a stronger influence on the collapse direction (van Wyk de
206 Vries and Merle 1996).

207

208

Figure 2

209

210 **2.2 Sloping substrate and gravitational spreading**

211 Many volcanoes grow on sloping ground, and some of them tend to slide downhill
212 and collapse catastrophically on the downslope side (Vallance et al. 1995; Carrasco-Núñez et
213 al. 2006; Murray et al. 2018). Many large stratovolcanoes worldwide are directly in contact
214 with thick sedimentary or igneous sequences that have been affected by different
215 compressional and extensional processes depending on the geotectonic evolution of the
216 area (e.g. the Andean mountain range). These sequences often appear tilted, showing a
217 certain grade of dipping and generating a strength anisotropy. In northern Chile, for
218 example, widespread ignimbrite sheets and thick ductile Cenozoic sedimentary deposits dip
219 toward the west into the Salar de Atacama basin with an angle of 5° (Wooller et al. 2004).
220 The Tumisa volcanic complex, located northward in the same region, is underlain by several
221 hundred meters of salt-rich lake sediments, ignimbrites and evaporite deposits, which
222 together form a westward-dipping ductile package (Kuhn 2002). According to Wooller et al.
223 (2004), the entire Tumisa volcanic complex has undergone a downslope sliding toward the
224 same western direction. Analogue experiments presented by these authors demonstrate

225 that volcano spreading is highly dependent on substrata dip and spreading occurs even at
226 very small substrata tilt ($< 1^\circ$).

227 Many other Central Andean volcanoes experienced slope collapses that left
228 horseshoe-shaped scars oriented toward the direction of the regional slope (Francis and
229 Wells 1988). Vallance et al. (1995) noted a similar pattern with Guatemalan volcanoes,
230 which failed preferentially trenchward to the southwest in the direction of regional slope of
231 the Guatemalan Highlands. At the eastern end of the Mexican Volcanic Belt, the Cofre de
232 Perote-Citlatépetl range is constructed at the margin of the Altiplano. The basement drops
233 dramatically to the Caribbean coastal plain, and all collapse events have been in that
234 direction (Carrasco-Núñez et al. 2006). In Central America, Siebert et al. (2006) noted that
235 with volcanoes at the far ends of the arc, failures tended to occur preferentially
236 perpendicular to the trend of the arc and in a direction toward the Central-American trench.
237 In contrast, volcanoes in El Salvador and Nicaragua in the center of the arc, where the
238 basement topographic effect is less pronounced, have more variable failure directions. This
239 is even more pronounced at an isolated island volcano like Augustine in Alaska, without
240 regional topographic gradients, where failures occurred radially in all directions (Begét and
241 Kienle 1992).

242 The same behavior occurs at volcanoes in Indonesia that are located on dipping
243 ductile substrata and have undergone volcano spreading, or at the volcanoes belonging to
244 the Lesser Antilles, in the Caribbean Sea. Deplus et al. (2001) report that the volcanoes are
245 situated on westward-dipping substrata related to the Tertiary Caribbean arc, the same
246 orientation in all major collapses of the region.

247 Numerous studies (Merle and Borgia 1996; Lundgren et al. 2004; Bonforte et al.
248 2011; Murray et al. 2018) reveal that the eastern and southern flanks of Mt. Etna are sliding
249 eastward towards the Mediterranean Sea, affected by the slope and lithology of the
250 basement, with zones of differential velocities forming several fault-bounded blocks. It is
251 not coincidental that the “Valle del Bove” scar is oriented in the same direction as Mt. Etna
252 creeps. Lateral collapses might also occur for those volcanoes located on steeper dip slopes,
253 compared with low-angle dip basement mentioned above. Due to the migration of an
254 eruptive vent, a consequence of various factors (e.g. changes in the plate subduction angle,
255 variation in the tectonic setting, migration of the magma chamber, among others), a
256 volcano might grow on the flank of an older volcanic edifice with slopes angles $> 20\text{-}25^\circ$. A

257 good example is the Fuego de Colima volcano (Fig. 3), which is located on the southern flank
258 of the older and inactive Nevado de Colima volcano. Whilst the direction of the sector
259 collapses of the Nevado de Colima volcano has been controlled by the Colima Rift
260 confinement (Norini et al. 2019), the inclination of the southward dipping Nevado de Colima
261 flank appears to be the main topographic constraint generating the instability of the Fuego
262 de Colima edifice (Norini et al. 2010; Roverato et al. 2011; Norini et al. 2019).

263

264

Figure 3

265

266 At a lower scale, on 16 November 2006 a lateral collapse affected the unstable
267 eastern slope of the steep South-East Crater (SEC) of Mount Etna, built on one side of the
268 steep central craters of the volcano (Norini et al. 2009). The collapse was triggered by
269 erosion of loose, hydrothermally altered material of the steep south-east sector of SEC from
270 the outpour of lava and produced a landslide of 330,000 - 413,000 m³ that involved both
271 lithic and juvenile material, resulting in a deposit emplaced on the eastern flank of the
272 volcano (Norini et al. 2009). Thus, the substrata dip and the steepness of volcano flanks is
273 important for volcanic edifice instability, as it can control the spreading to downslope
274 sectors and the associated sector-collapse volumes. Differential buttressing at the volcano
275 base is also a major factor controlling flank instability and is an important indication to
276 expect asymmetric activity in any volcano, like flank eruptions and deformations (Norini and
277 Acocella 2011).

278 Volcanic gravitational spreading is one of the most complex factors contributing to
279 instability that can affect a volcanic edifice. This is because numerous variables are involved
280 in the process, although the most important of those is the substratum on which the
281 volcano sits. Gravitational spreading is the result of a set of processes that translate the
282 vertical volcano load into horizontal movement, accommodated by high-angle normal faults
283 on the edifice's slopes and low-angle thrust faults at its boundaries and in the substratum.
284 In a pioneer work, van Bemmelen (1949) reported faults across volcanic edifices and
285 anticlinal ridges and low-angle thrusts along their peripheries for several Indonesian
286 volcanoes (e.g. Soropati, Lawu, Merapi, Bukittinggul, Tjereme). These volcanoes are built
287 on thick sequences of marine clays and fine-grained deposits. The deformation structures of
288 the substrata are attributed to sediment faulting beneath the volcanoes in response to

289 gravitational loading (Borgia et al. 2002). At Parinacota (Chile), the deformation of low-
290 density unconsolidated fluvio-glacial and lacustrine sediments and pyroclastic basin-fill
291 sequences in response to volcano loading is considered as the main causative factor of the
292 ca. 8,000 years old slope collapse that produced a 6 km³ VDAD (Clavero et al. 2002). Other
293 researchers described such typical features of deformation in the substratum sediments,
294 and it appears that they play a fundamental role in controlling the stability of a volcanic
295 edifice and the size of the collapsed landslide. At Mombacho volcano (Nicaragua), two
296 Holocene collapses are well described in the literature, one toward the south (El Crater) and
297 the other to the northeast (Las Isletas) (van Wyk de Vries and Francis 1997; Shea et al. 2008;
298 Shea and van Wyk De Vries 2010). The base of Mombacho volcano in its northeastern side is
299 surrounded by an anticline structure, which was likely formed as Mombacho was slowly
300 spreading outwards (Shea and van Wyk de Vries 2010). A pumice-rich substratum has also
301 been found at the base of the Las Isletas VDAD, much further from the original in-situ
302 ignimbrite deposit. Sedimentary sequences in the deposit indicate extrusion of the
303 substratum, which is eventually ripped up and transported by the VDA resulting from lateral
304 collapse. Socompa volcano, located on the southeastern margin of the Atacama basin
305 (Chile/Argentina), suffered a dominantly westward gravitational spreading (Wooller et al.
306 2004). The collapse was triggered by failure of active thrust-anticlines in sediments
307 underlying the volcano. High amounts of the substratum sediments have been found within
308 the Socompa VDAD (van Wyk de Vries et al. 2001).

309 Two distinct styles of deformation are described: 1) “volcano spreading” if the
310 volcanic edifice spreads with the substrata, and 2) “substrata extrusion” (van Wyk de Vries
311 and Matela 1998) when the volcano and substrata are decoupled and the substrata are
312 extruded from under the edifice. In both cases, compressional structures are produced at
313 the base of the volcano. At Socompa volcano, the gravitational deformation has developed
314 mainly by substrata extrusion expressed by a series of thrust anticlines (van Wyk de Vries et
315 al. 2001). Major structures observed on the summit of other volcanoes like Etna and Kilauea
316 (Borgia et al. 1992; Morgan et al. 2003; Urlaub et al. 2018) suggest that, at the beginning of
317 gravitational spreading, the volcanoes present grabens and rifts on their summits. These
318 features are also reported by Del Camp et al. (2008) for Piton de la Fournaise volcano
319 (Reunion Island), Concepción and Maderas volcanoes (Nicaragua), Merapi volcano
320 (Indonesia), and other Andean Chile/Bolivian volcanoes. Recent analogue models by Shea

321 and van Wyk de Vries (2008), Andrade and van Wyk de Vries (2010) and Paguican et al.
322 (2012) explored the kinematics of a volcanic failure from the early stages of a sector
323 collapse. Andrade and van Wyk de Vries (2010) reproduced in their experiments how
324 grabens and rift-structures develop at the summit of volcanic edifices in response to the
325 compressional forces at the base of the edifice. Furthermore, they report that most of the
326 viscous basal layers used in the experiments had been expelled from underneath the
327 volcano as in natural examples (van Wyk de Vries and Francis, 1997; van Wyk de Vries et al.
328 2001). One important insight from these experiments is that gravitational spreading stops or
329 is reduced after a catastrophic collapse (Andrade and van Wyk de Vries 2010). This is an
330 obvious conclusion as the volcanic loading decreases considerably after part of the edifice is
331 removed and spread onto the surrounding landscape. Alternatively, gravitational spreading
332 enhances catastrophic collapse only during its initial stages. When no catastrophic collapse
333 occurs, the volcanic edifice will continue to spread progressively, gradually reducing the
334 slope of its flanks (Andrade and van Wyk de Vries 2010; Norini et al. 2010).

335

336 **2.3 Hydrothermal alteration**

337 Volcanoes are actively growing and evolving systems characterized by the
338 interaction of fluids, gases and heat that can hydrothermally alter the edifice. Hydrothermal
339 alteration causes rock dissolution and mineral-chemical modification, it can promote high
340 pore pressures, weaken the volcanic edifice's rocks and increase the overall clay-content.
341 Hydrothermal alteration can occur progressively over long periods (>100 yrs), allowing
342 deformation to develop slowly before collapse (Van Wyk de Vries et al. 2000). Each
343 hydrothermal system develops in a unique manner depending on the rate of fluids that
344 circulate within the volcano, their temperature, the porosity and permeability of the
345 material forming the edifice, as well as the amount of external meteoric water infiltrating
346 into the edifice. Consequently, each hydrothermal system has its own distinctive size,
347 volume, chemical, and physical characteristics (Cecchi et al. 2004). Many large volcanic
348 lateral collapses have involved hydrothermally altered rocks (Siebert 1984), which can be a
349 useful indicator of the source of the collapse, and the slope destabilization and collapse
350 triggering mechanisms. Examples of VDADs that display hydrothermal alteration (Figure 4)
351 are reported in the literature and some are listed in Table 2.

352

353 Figure 4
354

355 Clay minerals from hydrothermal alteration are also found in some avalanche-
356 induced lahar deposits at Mount Rainier, including the 3.8 km³ Osceola mudflow (Vallance
357 and Scott 1997), the 0.2 km³ Round Pass mudflow, and the 0.26 km³ Electron mudflow
358 (Scott et al. 1995). The presence of hydrothermally altered rocks in these deposits suggests
359 that this process partly weakened the edifice and likely promoted the slope collapse (Reid et
360 al., 2001). Mount Rainier has large areas of hydrothermally altered rocks (Frank, 1995;
361 Zimbelman, 1996; Reid et al., 2001; John et al., 2008). Reid et al. (2001) noted that the
362 alteration in highly porous rocks such as scoria and pumice is less intense and present
363 mainly as vesicle infillings. On the other hand, the alteration is more severe in dense lavas
364 and dikes, largely obliterating original igneous minerals and textures, and contributing to an
365 overall reduction in rock mass strength.

366
367 Table 2
368

369 The Holocene Teteltzingo deposit, a ~2 km³ clay-rich VDAD formed by the Citlaltépetl
370 (Mexico) volcano's slope collapse, displays a high clay content in the range of 10 to 16%
371 (Carrasco-Núñez et al., 1993). Its features suggest that it had an origin from a water-
372 saturated hydrothermally altered edifice and that it transformed into a cohesive lahar at the
373 very beginning of the failure event. The volume of altered rocks involved in the collapse was
374 on the order of 10⁸ to 10⁹ m³ (Carrasco-Núñez et al. 1993). Currently, the Citlaltépetl
375 volcano is still characterized by a widespread alteration enhanced by the glacier (melted-ice
376 circulation), including massive zones of replacement and vein fillings along numerous
377 fractures (Zimbelman et al. 2004). Much of the replacement alteration is of acid sulfate.
378 During vapor condensation, the SO₂ responsible for the alteration is present in very high
379 concentrations, and a fluid of very low pH forms (Zimbelman et al. 2004). This acid brine
380 reacts with the host rocks, leaching it and creating an increase in permeability. The voids left
381 by the acid dissolution are now available to be filled by fluids such as meteoric water, which
382 may increase the pore pressure into the volcanic edifice if sufficient water exploits the new
383 pathways. Lopez and Williams (1993) stressed the importance of hydrothermal fluids
384 circulation into a volcanic edifice and considered the hot spring compositions, residence
385 times and flow path through faults of fluids to model potential collapse scenarios at Nevado

386 del Ruiz volcano (Colombia). The continual hydrothermal activity reduces the strength of the
387 rock (mainly due to clay mineral formation), and favors the more porous lithologies, which,
388 in turn, can increase pore pressures when fluids fill the voids (Day 1996; Voight and
389 Elsworth 1997; Reid 2004). Day (1996) suggests that the hydrothermal alteration is more
390 effective in the interior of volcanic edifices where higher pressures and temperatures could
391 dissolve the rocks more effectively (Frantz et al. 1981) creating high-permeability zones. On
392 the contrary, in the exterior and cooler parts of the edifice, clay minerals will precipitate
393 from the hydrothermal fluids, decreasing the permeability and promoting for example,
394 meteoric water storage. There is growing evidence that the internal fluid pressure in a
395 volcanic edifice is one of the main factors for potential instability. For example, Borselli et al.
396 (2011) report that, at Colima volcano (Mexico), the fluid pressure likely produces a local
397 reduction of shear strength along a portion or all of a potential sliding surface, promoting a
398 global reduction of the stability factor of the volcanic edifice. This idea is supported by a
399 recent resistivity tomography survey by Rosas-Carbajal et al. (2016) in which they
400 investigated the structure and dynamics of the hydrothermal system at La Soufrière volcano
401 (Guadalupe). The authors suggested that highly altered, fluid-saturated, weak zones and the
402 low basal friction of argilized detachment planes could significantly reduce the factor of
403 safety of the volcano slope. Finally, the recent study of Delmelle and Bernard (2015) showed
404 how slope instabilities, associated with hydrothermal alteration, can increase by the
405 formation of acidic crater lakes on the summits or within the flank topography of active
406 volcanoes, such as the Kawah Ijen hyperacid lake ($\text{pH} < 0.5$) on the flank of Merapi volcano.
407 Leakages through the flanks of the acid lake and dissolution beneath it, weakens the interior
408 of the volcano and build a substantial body of highly altered rock within the upper part of
409 the volcanic edifice that reduces its stability (Delmelle et al. 2015).

410

411 **2.4 Dikes and magma intrusions**

412 Historical volcanic lateral collapses at Bezymianny, Kamchatka in 1957, Mount St.
413 Helens in 1980, and Soufrière Hills, Montserrat in 1997, all generated large eruptions with
414 associated lateral blasts caused by large-scale deformation and steepening of flanks prior to
415 failure (Gorshkov and Bogoyavkenskaya 1965; Gorshkov 1959; Voight et al. 1983; Siebert et
416 al. 1987; Belousov et al. 2004). These volcanoes provide examples of lateral collapse
417 initiated by magma intrusions and stress the importance that this represents for volcano

418 stability and hazard assessment. Instability can be related to the intrusion (dikes and
419 cryptodomes), as well as extrusion (summit domes) of fresh magma, and can trigger failures
420 through overloading and oversteepening, deformation, or mechanical- and temperature-
421 related changes in pore pressures (Elsworth & Voight 1996; Day 1996; McGuire 2003, Norini
422 and Acocella 2011; Samaniego et al. 2015). For example, the high eruption rates of silicic
423 domes and lava flows at Mount St. Augustine (USA) repeatedly produced an oversteepened,
424 unstable volcanic cone which then collapsed (Beget and Kienle 1992). At Ruapehu (New
425 Zealand) a paleomagnetic study (McClelland and Erwin 2003) revealed that a summit dome
426 could have played some contributory part in the destabilization of the edifice. High
427 temperatures (350°C) of some dacite blocks into the 9.5 Ka BP Murimotu formation
428 suggests that a dacite dome was active on Ruapehu immediately prior to the lateral
429 collapse. In Mexico, Macias et al. (2010) report that the late Pleistocene collapse of Tacaná
430 volcano was likely initiated by the extrusion of an andesitic lava dome that weakened the
431 structure of the cone. A recent contribution by Rincon et al. (2018) tried to explain, with
432 analog models, how volcanoes could be destabilized by magma intrusion. The authors noted
433 that a large and deep listric fault crossing the volcano can form when the magma body is
434 still located at the base of the cone. This explains why, in some cases, a volcanic edifice
435 could collapse even if the intrusion does not reach shallower levels. If any trigger induces
436 the slope failure at this stage, no juvenile material will be involved in the collapse. Rincon et
437 al. (2018) also noted that, when the intrusion reaches shallower levels, an elongated
438 subsidence forms at the summit of the cone and a bulge at the middle flank develops. The
439 same deformation was noted at Mount St. Helens before its collapse (Lipman et al. 1981)
440 and described also in other analogue experiments (Donnadieu and Merle, 1998). The
441 intrusion of a cryptodome (Figure 5) into the upper part of a volcanic edifice

442
443 Figure 5
444

445 decreases its stability due to many different factors, for instance lower shear strength of the
446 magmatic body and associated weakening structures, shallow seismicity on the flank caused
447 by edifice fracturing in response to intrusion, outward-dipping normal faults bounding a
448 summit graben, hydrothermal alteration of large portions of the edifice, and bulging (i.e.
449 increase in steepness) of the destabilized flank (Donnadieu et al. 2001). Elsworth and Voight

450 (1996) have drawn attention to the potential role of magmatic intrusions in triggering
451 structural failure as a consequence of their raising pore fluid pressures by means of
452 mechanical or thermal straining (thermal expansion) of the rock-fluid medium and/or
453 hydrothermal fluid circulation. The thermal expansion can develop large uplift forces
454 beneath a potentially unstable block (Elsworth and Voight 1995). The magma has a gas-rich
455 phase at depth. This, coupled with magma degassing at shallower levels, could lead to
456 excess upward-propagating pore pressures (Thomas et al. 2004). In addition, dehydration
457 (boiling) of active hydrothermal systems during magma emplacement can lead to
458 considerably elevated pore pressures (Reid 2004). Once a volcanic system has reached a
459 critical state, collapse may be initiated by any trigger or simply occur because the factor of
460 safety is reduced sufficiently. At Shiveluch volcano, the cause of the destabilization was the
461 ascent of a batch of viscous, water-rich magma to shallow levels (Belousov et al. 1999). The
462 failure occurred before magma intruded into the edifice suggesting that the trigger
463 mechanism was not magma-associated. A more likely destabilizing mechanism could have
464 been related to the increase of pore pressure in the edifice due to pressurized gas released
465 from the ascending magma (Belousov et al. 1999). Evidences from fumarolic gases in the
466 thermal springs at La Soufrière of Guadeloupe volcano (Lesser Antilles) suggest that the
467 current activity is the result of a new magma intrusion which was progressively emplaced at
468 shallow depth in the last decades (Villemant et al. 2014). The assessment of potential
469 hazards associated with the evolution of the current unrest must consider the implications
470 of recurrent intrusions and further pressurization of the hydrothermal system. The role of
471 hydrothermal pressurization, linked with magma intrusion, on the basal friction along low-
472 strength layers within the upper part of the edifice must be evaluated with regards to slope
473 collapses (Borselli et al. 2011; Villemant et al. 2014).

474

475 **2.5 Past and present climate implications**

476 Climate can influence volcanic activity and plays an important role in the stability of
477 a volcanic edifice. Heavy rains or hurricanes, for example, can promote massive landsliding
478 over large areas. During the passage of Hurricane Mitch above Central America in 1998,
479 heavy rains promoted a lateral collapse of Casita volcano (Nicaragua), killing 2,500 people
480 (Kerle and van Wyk de Vries 2001; Scott et al. 2005). The source volume of the landslide
481 (~200,000 m³) was comparatively small to other giant volcanic collapse-related mass flows.

482 However, the landslide entrained older deposits and organic material during the flow,
483 transforming into a devastating debris avalanche (Kerle and van Wyk de Vries 2001) and
484 subsequently into a debris flow (Scott et al. 2005). It is common worldwide that debris
485 avalanches of Casita-size occur every few years (McGuire 1996; Kerle and van Wyk de Vries
486 2001). Climatic hazards rapidly affect widespread areas contributing to volcanic instability.
487 Heavy rains during wet seasons and glacier retreat have become more prominent over the
488 last several decades due to changing climatic conditions as a result of global warming (Capra
489 2006; Roberti et al. 2020 - this volume). Recent examples in periglacial environments of ice-
490 avalanches, rock avalanches, and a mix between both have been considered to be
491 increasing by these changing conditions (Huggel et al. 2007; Huggel et al. 2008; Tuffen 2010;
492 Kos et al., 2016; Coe et al. 2018; Hibert et al. 2019). The 2010 and other large landslides at
493 Mount Meager volcano (Canada), for example, resulted, in part, from glacier retreat and
494 erosion contributing to changes in stress of its already unstable slopes (Grämiger et al. 2017;
495 Roberti et al. 2020 - this volume). With current accelerated rates of glacial melting, glaciated
496 active volcanoes are at an increasing probability of generating landslides, from shallow
497 failures to debris flows to large-volume VDAs. Well-known paleo-climate records from the
498 last deglaciation show that melting of ice triggered a dramatic acceleration in volcanic
499 eruptions (Capra et al. 2015). However, major glaciations (and deglaciations) have been
500 considered not only as a triggering mechanism for large magmatic eruptions but also
501 inducing volcano instability (Capra et al. 2013). Ice-capped volcanoes are sensitive
502 environments to climatic changes, and large-scale glacial melting occurring at the end of the
503 Pleistocene, caused a significant increase in the incidence of large volcanic lateral collapse
504 and debris flows (Tormey 2010). Tormey (2010) reported that Planchon-Peteroa volcano
505 (Chile-Argentina border) had a large slope collapse that produced a highly mobile and
506 erosive debris avalanche 11,000 years BP, and other slope instabilities during the end of the
507 Pleistocene to early Holocene deglaciation. Capra (2006) suggested that most volcanic
508 lateral collapses of ice-capped volcanoes after the last main glacial peaks have been induced
509 by changing climatic conditions. The late Pleistocene and Holocene geological record
510 indicates that several volcanic slope collapses occurred in limited time spans at 1.4–1.3 ka
511 BP, at 1.0–0.9 ka BP, and at ~0.8 ka BP coinciding with several abrupt deglaciations at
512 different latitudes (Clapperton 1998; Capra 2006). Volcanic lateral collapse events in
513 periglacial environments are conditioned by different factors such as 1) shallow

514 hydrothermal alteration driven by snow and ice melting (Carrasco-Nuñez et al. 1993;
515 Waythomas et al. 1999; Roverato et al. 2011; Roverato and Capra 2013), 2) decreasing rock
516 mass-strength by abrasion and deformation due to glacier advancing (Capra et al. 2013), 3)
517 glacial plucking enhancing volcano weaknesses by exploiting preexisting fractures (Bennett
518 and Glasser, 2009) and 4) increasing pore pressures by high water circulation (Roberti et al.
519 2017). The high circulation of water in the collapsing mass could also strongly influence the
520 debris avalanche by transforming it into a debris flow-like rheology (Carrasco et al. 1993;
521 Capra et al. 2002; Roverato et al. 2011). Roverato et al. (2015) reported that at the time of
522 the Pungarehu collapse, during the Last Glacial Maximum, Taranaki volcano (New Zealand)
523 was not glaciated, but it did experience periglacial conditions. The snow/ice water-melting
524 coming from the glacier/snowfield or at least a high humid environment might suggest that
525 the pre-avalanche volcanic edifice was partially water-saturated. The increase in
526 precipitation arising from post-glacial warmer, wetter climatic conditions has been
527 considered possible trigger of catastrophic volcanic slope collapses (Capra 2006). This
528 ice/snow could have been an additional supply of water in the collapsed mass during the
529 Pungarehu event permitting the long runout of the avalanche (Roverato et al. 2015). Similar
530 conditions are reported, for example, by Alvarado et al. (2004) concerning volcanic slope
531 collapses in Costa Rica during the last glaciation, for the Teteltzingo collapse at Citlaltepēt
532 volcano (Carrasco-Nuñez et al. 1993) and for the Tonila event at Colima volcano (Roverato
533 et al. 2011). In light of the potential effects of climate change on landslide hazards, it is
534 crucial to monitor potential landslides locations and establish readily available methods for
535 early debris avalanche prediction (Huggel et al. 2008).

536

537 **3. The instability of volcanic islands**

538 Volcanic lateral collapses have been identified around volcanic islands worldwide
539 and their failure can trigger devastating tsunamis threatening coastal communities and
540 infrastructure (Karstens et al. 2019; Watt et al. 2020 - this volume). The term “volcanic
541 island” is a comprehensive term that includes the main endmembers: oceanic island and arc
542 island. The first ones are gently sloping oceanic shield volcanoes, such the Kilauea (Hawai’i,
543 USA) in the Pacific Ocean or the Piton de La Fournaise (Reunion Island, France) in the Indian
544 Ocean, usually associated with persistent flank movement (Poland et al., 2018). Oceanic
545 island volcanoes, especially in their active shield-building phase, are frequently affected by

546 large-scale lateral collapses (Moore and Fiske 1969; Elsworth and Day 1999). The arc
547 volcanic islands instead are usually steep-sided composite volcanoes such as Fogo (Cabo
548 Verde) in the Atlantic Ocean, Stromboli (Italy) in the Mediterranean Sea, or Augustine (USA)
549 in the Aleutian arc (Reid et al. 2010), associated with minimal ground deformation at the
550 flank-scale (González et al. 2010), sometimes with strong (volcaniclastic) material
551 remobilization (Di Traglia et al. 2018). Failures of volcanic islands include thin-skinned,
552 narrow collapse, and deep-seated instability events, removing much of an edifice and
553 depositing huge amounts (several km³) of debris (Coombs et al., 2007; Montanaro et al.
554 2011). They represent the largest formations which can be created in a single geological
555 moment (Blahůt et al. 2018). In addition to the instability and triggering factors that affect
556 volcanoes in all geographical contexts (Voight and Elsworth 1997; Elsworth and Day 1999;
557 Reid et al. 2010), observing that >10 km³-volume collapses occurred at the onset of glacial
558 to interglacial transitions, it suggested that rapid sea level rise enhanced changing of pore
559 pressure conditions, contributing to slope collapse (Quidelleur et al. 2008).

560 The classification of landslides on volcanic islands usually comprises two
561 endmembers: volcanic debris avalanches, thought to reflect a sudden single catastrophic
562 event, and slumps, thought to reflect progressive slope failure (Blahůt et al. 2018). Slumps
563 affects volcanic flanks with persistent motion and consequent flank deformation, as the
564 Hilina Slumps at Kilauea in Hawai'i (Okubo 2004) and the southeastern flank of the Pico
565 volcanic ridge in the Azores (Hildenbrand et al. 2012). Large VDADs have been observed on
566 the ocean floor surrounding the Hawaiian Islands (Iverson 1995; Morgan et al. 2003), the
567 Canarian Archipelago (Carracedo 1999; Hurlimann et al. 1999), Cape Verde (Elsworth and
568 Day 1999; Ramalho et al., 2015) and Azores (Hildebrand et al. 2018). Some of the debris
569 avalanches may develop by partial or complete collapse of a volcano slope already
570 oversteepened by a slump (Normark et al. 1993). Debris avalanches in arc volcanic islands
571 are typically smaller than those at ocean islands, but they are more frequent (Day et al.
572 2015). The VDADs around ocean islands and island arc volcanoes have similar slope
573 gradients in the deposition area, mobility and lithology (Karstens et al. 2019).

574 In historical times, tsunamigenic volcanic debris avalanches have caused many
575 casualties, two of the most devastating cases were: i) the 1792 Unzen–Mayuyama
576 megaslide (volume of 3.4×10^8 m³ and the maximum depth was 400 m) that killed around
577 15,000 people, and ii) Ritter, Papua New Guinea, in 1888 (mobilized volume of about 4.2 x

578 10^9 m^3 ; Day et al. 2015 or $2.2\text{-}2.6 \times 10^9 \text{ m}^3$, Karstens et al. 2019). The slope collapse of Ritter
579 Island is the largest historically recorded landslide at a volcanic island (Cooke 1981; Day et
580 al. 2015; Karstens et al. 2019). The failure produced a headwall width of $\sim 3 \text{ km}$ and a fall in
581 vent elevation of $\sim 1,500 \text{ m}$, reducing the 780 m high, 1.5-km -wide island to a small remnant
582 (Johnson, 1987). The landslide-induced tsunami affected islands out to $\sim 600 \text{ km}$ from Ritter,
583 killing hundreds of people and producing considerable damage (Cooke 1981; Day et al.
584 2015; Watt et al. 2019). Recent studies (Karstens et al. 2019; Watt et al. 2019) based on
585 high-resolution 2D and 3D seismic and bathymetric data, provided new insights into the
586 collapse sequence and landslide-induced tsunami generation. The collapse of Ritter
587 occurred in two phases: (1) the volcanic edifice was first affected by deep-seated gradual
588 spreading generating compressional deformation within the volcano and the adjacent
589 seafloor sediments; (2) the failure of about 2.4 km^3 of the island and the generation of a
590 highly energetic mass flow. This second phase is inferred to have been deposited up to 80
591 km from Ritter (Watt et al. 2019). The slide triggered an explosive phase that formed a
592 crater-like depression (Karstens et al. 2019).

593 More recently on 22 December 2018 the lateral collapse of the Anak Krakatau
594 occurred (Figure 6). The active volcanic island, situated within the Krakatau caldera in the
595 Sunda Strait between Java and Sumatra, experienced a slope collapse after a period of
596 intense volcanic activity (Williams et al. 2019). This event generated tsunami waves ranging
597 in height from 0.27 to 1.40 m , caused the deaths of 431 people, injured a further $7,200$
598 people, and displaced $46,646$ people (Williams et al. 2019).

599 A good example of a tsunamigenic landslide that occurred in recent times, and is
600 supported by many studies, is the one at Stromboli volcano during its 2002–2003 eruption
601 (Tinti et al., 2006). The volcanic island is located in the Tyrrhenian Sea off the southern coast
602 of Italy, and was affected by several mass-wasting events during its geological history,
603 mainly located in the northwest part of the volcano that produced a depression called
604 "Sciara del Fuoco" (Kokelaar and Romagnoli 1995; Tibaldi 2001). During the 2002–2003 flank
605 eruption, a submarine–subaerial landslide sequence (total volume of $25\text{--}30 \times 10^6 \text{ m}^3$,
606 Bonaccorso et al. 2003; Chiocci et al. 2008) generated two tsunamis. These tsunami waves
607 affected the coastline with a maximum run-up of $6\text{--}7 \text{ m}$ at the Stromboli village (Tinti et al.,
608 2006) and were caused by two initial submarine rotational slides that affected the

609 underwater volcanoclastic apron down to a depth of 350 m (Chiocci et al. 2008), as well as by
610 two second subaerial translational slides (Tommasi et al. 2005).

611

612 Figure 6

613

614 Geological and geotechnical studies (Tommasi et al. 2005; Chiocci et al. 2008; Boldini
615 et al. 2009; Verrucci et al. 2019) suggest that the submarine and sub-aerial slides were
616 caused by large deep-seated gravitational slope deformations (DSGSD) affecting the
617 northern part of the Sciara del Fuoco, triggered by the injection of a lateral intrusion
618 (Bonaccorso et al. 2003). The DSGSD developed under drained conditions, preventing the
619 collapse of the entire block (Chiocci et al. 2008). Contrariwise, during its continuous
620 displacement, excessive pore pressure was likely generated in the submarine volcanoclastic
621 apron (undrained load condition), reducing the effective stresses down to values compatible
622 with the static liquefaction of the material (Tommasi et al. 2005; Boldini et al. 2009).

623 Another volcano showing strong flank instability is Kilauea, on the Big Island of
624 Hawai'i (USA). Kilauea is a persistently active volcano, partially overlapping the larger
625 neighbour Mauna Loa, located in the middle of the Pacific Plate at the leading edge of a
626 ~6,000-km-long chain of islands and seamounts. The south flank of Kilauea volcano,
627 bracketed by two rift zones that extend from Kilauea's summit caldera, is creeping slowly
628 seaward along a weak, low-angle décollement (Poland et al, 2017. Large scale slope collapse
629 (volumes of 5,000 km³) have been recognized. The Hilina Slump is the most notable of these
630 features (Smith et al. 1999). Major rupture and offset on the southern flank have occurred
631 along faults as recently as 1975, during the M7.2 Kalapana earthquake that struck the south
632 coast of Hawai'i (Okubo 2004; Day et al. 2005). Kilauea's south flank experienced also
633 transient aseismic displacements, as occurred in early November 2000, when an event of
634 about 36 hours moved the flank at a maximum slip velocity of about 6 cm per day (Cervelli
635 et al. 2002). The last transient flank motion occurred on early May 2018 that coincided with
636 the lateral propagation of a dike along the East Rift Zone. The flank slip triggered a moment
637 magnitude (M_w) 7.2 earthquake, producing ~5 meters of fault slip (Chen et al. 2019).

638 Other examples of failures of volcanic islands can be found in the Lesser Antilles arc,
639 where 52 volcanic flank-collapse events are recognized, including the active volcanoes of
640 Montserrat, Dominica, Martinique, and St. Lucia (Deplus et al. 2001; Boudon et al. 2013).

641 Large-scale debris avalanche deposits (10-20 km³) lying on the seafloor around the islands,
642 implying that relatively small volcanic islands, as Montserrat, can be affected by repeated
643 large-volume landslides that deposit very large volcanoclastic material into the surrounding
644 basins (Lebas et al. 2011). The largest deposits are related to collapses which include large
645 sectors of the islands, both subaerial and submarine. However, it must be considered that
646 even smaller subaerial landslides, associated with the collapse of lava domes, such as one
647 that occurred on 26 December 1997 (the so-called "Boxing day" collapse; Voight et al.
648 2002), are extremely dangerous. At Soufrière Hills Volcano on Montserrat, a phase of dome
649 growth started in the 1995. The collapse produced a $\sim 50 \times 10^6$ m³ volcanic debris
650 avalanche, and depressurized the interior of the lava dome, which exploded to generate a
651 violent pyroclastic density current (Voight et al. 2002).

652 At Réunion Island spreading and avalanching developed in two different sectors (Le
653 Friant et al. 2011). In particular, Réunion island is a 7,000 m-high volcanic complex with a
654 large submarine part and two emerging volcanoes (Piton des Neiges and Piton de la
655 Fournaise; Lénat et al. 1990; Lebas et al. 2018). Investigations by Le Friant et al. (2011)
656 revealed that slow flank deformation is likely one of the main processes acting on the
657 submarine flanks of Piton des Neiges, causing the continuous flank erosion by landslides,
658 and producing turbidity flows (Bachèlery et al. 2003; Le Friant et al. 2011). Conversely, Piton
659 de la Fournaise volcano collapsed several times in the same seaward direction, generating
660 several voluminous debris avalanches that flowed down the submarine flanks. Le Friant et
661 al. (2011) suggested that the difference in behaviour between the two volcanoes is likely
662 related to the volcano basement. Piton des Neiges was built on marine sediments,
663 promoting radial spreading, whereas Piton de la Fournaise was emplaced on the flank of a
664 pre-existing edifice, favouring repeated collapses.

665

666 **4- Discussion**

667 Several contributing factors can affect the stability of volcanoes. These factors create the
668 conditions that set the volcanic edifice to "ready to collapse". According to classifications
669 proposed by previous authors (Gorshkov, 1965; Moriya 1980; Siebert et al. 1987) three main
670 failure categories exist: Bezymianny, Bandai, and Unzen/Ontake types. The difference
671 between them is mainly based on the presence of a magmatic or phreatic explosive activity
672 (Bezymianny and Bandai) or an earthquake as the trigger (Unzen/Ontake). The Bezymianny-

673 type collapse takes its name from the event that occurred in 1957 at Bezymianny Volcano
674 (Kamchatka), which was mainly related to a magmatic activity (Gorshkov, 1959;
675 Bogoyavlenskaya and Kirnasov 1981). This type of activity corresponds to the destabilization
676 of the volcanic edifice by magmatic intrusions and followed by magmatic eruptions. The
677 edifice undergoes strong deformation due to the emplacement of cryptodomes; the same
678 conditions that occurred at Mount St. Helens. The eruptive phase may contain highly
679 energetic lateral blasts (Belousov et al. 2004) with associated pyroclastic density currents.
680 The lateral blasts of Mount St. Helens and Bezymianny produced major pyroclastic density
681 currents that devastated areas of 500-600 km² around each of the volcanoes (Crandell and
682 Hoblitt 1986; Siebert et al. 1987). The Bezymianny activity is often followed by construction
683 of a lava dome or a new pyroclastic cone within the collapse scarp (Siebert et al. 1987; Leyrit
684 2000). In the case of Bezymianny volcano, the duration and magnitude of deformation, as
685 well as the presence of a flat lava dome in the crater prior to failure and magmatic eruptions
686 suggest that the intrusion was well developed. Other cryptodome-induced deformations
687 have been found at Showa-Shinzan and Usu-Shinzan (Katsui et al. 1985) of Usu volcano
688 (Minakami et al. 1951) in Japan. They did not culminate in a debris avalanche associated
689 with a paroxysmal eruption like at Mount St. Helens or Bezymianny (Donnadieu et al. 2001).
690 Bandai activity (Moriya 1980) refers to the explosion that occurred in 1888 at the Bandai-
691 San volcano (Japan). It is associated with phreatic explosions that favor the collapse of the
692 volcanic edifice without subsequent magmatic activity. It means that no juvenile material is
693 present neither within the debris avalanche deposit nor at the base or top of it. Massive
694 VDADs can also occur without related explosive eruptions, as at the Unzen volcano in 1792
695 or at Ontake volcano in 1984, where earthquake triggered the collapse. The potential of a
696 seismically-triggered volcanic edifice collapse to initiate a volcanic debris avalanche has
697 commonly been recognized (Voight and Elsworth, 1997; McGuire, 2003). Edifice collapse
698 and flank instability-derived deposits (VDADs) are now recognized as ubiquitous events
699 associated with active and growing volcanoes, yet the identification of the triggering
700 mechanisms is poorly correlated due to difficulties that lie in elucidating diagnostic
701 lithological and sedimentological features that record evidence of the trigger. Kataoka et al.
702 (2018) and Fairchild (1987) both highlight the importance of understanding the triggering
703 mechanisms of a landslide to then understand the emplacement mechanisms and resulting
704 hazard. Only in some cases such as Unzen-Mayuyama (1792), Bandai (1888), Bezymianny

705 (1956), Shiveluch (1964), Mount St. Helens (1980), and Ontake (1984) were observations of
706 the events well-constrained enough that linkages could be made between initiation, trigger
707 and flow through to emplacement and deposition. In the cases such as Unzen-Mayuyama
708 (1792) and Ontake (1984) earthquakes were directly attributed to initiating the VDA. The
709 1792 Unzen-Mayuyama VDA is remembered as the trigger for a tsunami that impacted the
710 Ariake Sea, killing ~14,500 people (Watanabe et al. 1993; Miyamoto 2010). Wang et al.
711 (2019) states that from November 1791 to August 1792 earthquakes, eruptions and flows
712 were recorded from various sites around Mt. Unzen. On the 21 May 1792 a magnitude 6.4
713 earthquake triggered a volcanic debris avalanche with an associated tsunami. Usami (1996)
714 estimated that the earthquake which triggered a $3.4\text{--}4.8 \times 10^8 \text{ m}^3$ volcanic debris
715 avalanche (Ui et al. 1986; Inoue 2000) was likely to have been of magnitude = 6.4 ± 0.2 . This
716 was estimated based on record that more than 30% of buildings in the Shimabara area were
717 destroyed and ground ruptures were recorded which requires seismic accelerations of 400
718 cm/s^2 or more (Sassa et al. 2014). Sassa et al. (2014) also reported that ground accelerations
719 of between 370 to 507 cm/s^2 were required from an earthquake to trigger a landslide of
720 large enough volume to generate a tsunami of the size recorded. Although the geological
721 and volcanological relationship between other volcanoes of the Unzen volcanic complex and
722 Mayuyama is difficult to discern, the Mayuyama volcano or lava dome is considered to be
723 5.1 ± 1.5 ka old (Hoshizumi et al., 1999) with the last recorded coeval activity at Unzen being
724 a block-and-ash-flow at 4.2 ± 110 ka B.P. Therefore, eruptive activity or intrusion into the
725 dome was not a contributing factor in triggering the debris avalanche sourced from older
726 brittle and dacitic volcanic rock. Another well-known earthquake-triggered edifice collapse
727 is Ontake type, defined from the 1984 events from Ontake-San volcano. On 14 September
728 1984 the southeast slope of Ontake-San collapsed producing a $32\text{--}36 \times 10^6 \text{ m}^3$ avalanche
729 (Voight and Sousa 1994). The last previous eruptive event at this volcano was in 1979 and
730 since then the volcano had been dormant with the 1984 debris avalanche event being
731 triggered by a magnitude 6.8 earthquake centered at 9-10 km southeast (Endo et al. 1989).
732 The source rock was highly saturated andesite with interbedded or intercalated pyroclasts
733 (Endo et al. 1989). Earthquake activity in this region is strongly aligned with regional stress
734 fields, yet Sano et al. (1998) suggest some relationship with deeper-seated magmatic
735 processes. Whereas Voight and Sousa (1994) focused on modeling the flowing mass and
736 constraining parameters that increase mobility, little attention was paid to modeling the

737 initiation of the collapsing mass or fragmentation of that mass from the dynamic loading of
738 the earthquake.

739 The eruption from Mount St. Helens in 1980 is probably one of the most well-
740 described and documented events prior to, during and post- a VDA. Lipman and Mullineaux
741 (1982) and Christiansen et al. (1981) provide the chronology of the event identifying that
742 the magnitude 5.1 seismic event (Glicken 1996) triggered a series of events that lead to a
743 VDA. Endo et al. (1981) also identified that the magnitude 5+ earthquake had a hypocenter
744 depth of 1.5 km. The series of events (with respect to the VDA) recorded from eyewitness
745 accounts progress from the earthquake with avalanching of rocks from crater walls through
746 to “quivering” of the flank, fracturing and then slope failure. From the analysis of photos, it
747 is estimated that there were 7-20 seconds between the earthquake and the initiation of the
748 collapse (Voight et al., 1981). It is important to note that seismicity increased from 20 March
749 with seismicity being magnitude 4 or less considered to be confined to 2.5 km depth or less.
750 Dzurisin (2018) noted that by 27 March 1980, magma had intruded the edifice with
751 associated deformation being recorded. Voight et al. (1981) further postulated that the
752 failure was due to several cumulative facts, including progressive strength-loss prior to slip,
753 seismicity and gravitational movements, and increased pore pressure within rock units.

754 How earthquakes change the physical properties of the volcanic rock mass is not
755 well defined yet. The complexity of a volcano structure including its lithological layering,
756 variability of rock strengths, constructional boundaries, faulting, water storage,
757 hydrothermal activity, and others make the task of applying simple geotechnical analysis
758 difficult let alone introducing seismicity into the equation. The mechanisms that cause a
759 lateral collapse are in general cumulative or inter-related (Voight and Elsworth 1997) and
760 therefore it is difficult to identify a single initiation mechanism or trigger to edifice collapse
761 which would in turn negate the previous oversimplification of the Bezymianny, Bandai, and
762 Unzen/Ontake type of classifications previously used. It thus requires a more
763 comprehensive classification of edifice collapse based on a range of pre-conditioning and
764 triggering factors.

765 Voight and Elsworth (1997) identified that while the inherent properties of volcanic
766 rocks and the structure of the volcanic edifice are factors that contribute to volcano
767 instability, the driving factors that lead to collapse are generally changes in shear stress on
768 the rock mass. An important associated factor that changes the shear stress is variation in

769 pore fluid pressure. In terms of seismicity, earthquake shaking can increase pore fluid
770 pressure in a saturated material. Modelling or simulating these types of relationships and
771 processes within a volcanic edifice is difficult in particular on large stratovolcanoes where
772 rock layers are massive and varying in thickness and rock strength is juxtaposed in a
773 multitude of positions related to the evolution of that edifice over hundred or even
774 thousands of years. Geotechnical analyses of the volcanic edifices have been undertaken
775 around the world with success in identifying inherent instabilities in a variety of volcano
776 types (e.g. Zimbelman et al. 2004; Moon et al. 2005; del Porto et al. 2013; Apuani et al.
777 2013).

778 Deutschle (2013) provided a comprehensive study of examining a range of triggers
779 that contribute to flank instabilities on Mt. Taranaki in New Zealand. This complex
780 stratovolcano has evolved over ~150,000 years and undergone up to 14 collapse events and
781 regrowth resulting in a structurally and lithologically complex edifice (Zernack et al. 2012).
782 Numerous limit equilibrium and finite element simulations that involved investigating rock
783 strength variability due to changes in basement and edifice conditions, edifice growth, and
784 magmatic intrusion show that while weaker areas of unconsolidated pyroclastic material
785 exist on the current cone, it is still, in general considered stable unless magmatic intrusion
786 or significant seismic shaking (>0.3 g) occurs (Deutschle 2013). Della Pasqua et al. (2016)
787 undertook a similar analysis to understand the hazards posed by VDAs in the Taranaki
788 region and found that susceptibility to failure will increase when there is higher seismic
789 loads on weak, stratified layers. Alloway et al. (2005) identified a number of faults that
790 traverse the Mt. Taranaki edifice and reconstructed earthquakes of magnitude 6.7-7.2 with
791 shaking intensities of magnitude >8 with a recurrence interval of 4 ka. Keefer (1984)
792 suggests that rock and soil avalanches require strong shaking intensities of magnitude >7-8.
793 It is conceivable that a number of Mt. Taranaki VDADs were initiated by ground shaking
794 associated with movements on those faults, but no conclusive evidence is presented in the
795 literature. Earthquakes triggering VDAs is well recorded and while the triggering effects of
796 the ground shaking to induce collapse have been elucidated from landslide investigations
797 through understanding the physical properties of soils and rock, the initiation of the mass
798 that suddenly loses its stability and move downslope is not well constrained or modelled
799 accurately for volcanic edifices.

800 As we can see, most of the information available in the literature regarding
801 eruptions and earthquakes, and their role in sector collapses, arises from historical cases or
802 is inferred by theoretical models and simulations. However, if we take a look at the
803 statistics of real cases, there is not much pre-historic information available about the
804 triggering mechanisms that promote a sector collapse. Dufresne et al. (2020 - this volume)
805 reports that only 16% of VDADs in a new global database of 1001 events have a related
806 triggering mechanism; most of them are associated with magmatic activities. It is obvious
807 that it proves difficult to associate a volcanic collapse with a specific trigger. However, some
808 information on factors and processes involved in past collapses is recorded in the resulting
809 deposits. In Table 3, we summarize key features that could be useful for studying VDADs in
810 the field by providing information on the state of the pre-collapse volcanic edifice and the
811 processes that acted before or during a sector collapse.

812

813 Table 3

814

815 Efforts to couple lateral collapse models and flow models as proposed in Voight and
816 Sousa (1994) need to be further developed. Being able to transfer those simulations and
817 predictions into a hazard and risk analysis framework (e.g. Mead et al. 2018) to provide a
818 more holistic analysis of the event is also required considering that much of the motivation
819 for understanding volcanic debris avalanches arose from better understanding the
820 hazardous events that occurred over the last 300 years.

821

822 **5- Conclusions**

823 Volcanic sector collapses can be prepared by a wide variety of destabilizing factors,
824 which are followed by triggering elements that drive the collapse (Figure 7). Although we
825 group these factors in distinctive sections, it is evident that most of them are inherently
826 related. In fact, factors that prepare and trigger volcanic sector collapses are not simple but
827 can be multiple and cumulative. Over-steepened slopes, magma intrusions, hydrothermal
828 activity, climate fluctuations, deformation of the basement, eruptions, earthquakes, are just
829 a list of elements that could promote a volcanic collapse. Although it is necessary to identify
830 the differences between those factors that define the degree of instability from those
831 mechanisms that trigger the collapse, it is also important to define those factors that, in

832 combination, could lead to different catastrophic scenarios. Indeed, the mechanism that
833 triggers a sector collapse could be independent of the causes that prepare its instability. But
834 it is also true that multiple factors in consort could drastically enhance the power and
835 magnitude of the collapse event and affect its runout.

836

837 **Figure 7**

838

839

840 The intrusion of a magma body (e.g. cryptodome) into a volcanic edifice, for
841 example, has multiple consequences for the stability of the volcano. On one hand, the
842 magma body deforms the edifice structure and, if the magma reaches the surface, its
843 products could overload the volcanic flanks (destabilizing factor). On the other hand, the
844 hydrothermal fluids associated with the magma body could promote all those processes of
845 rock leaching and dissolution that weaken the volcanic material. If the intrusion of a magma
846 body, linked with an earthquake or volcanic tremors, could be considered a factor that
847 induce a volcanic collapse (trigger), clay mineral formation and migration due to the
848 presence of magmatic fluids is an important element both for the loss of rock strength
849 (destabilizing factor) as well as for the increase of impermeability that promote fluids and
850 water (e.g. meteoric) storage. Once the collapse occurred, the presence of fluids and water,
851 and the clay content as well, can play a fundamental role in the transport and mobility of
852 the avalanche body, enhancing its mobility and transforming the moving mass into a
853 cohesive lahar. Thus, considering all the endogenous and exogenous factors, a volcanic
854 sector collapse and debris avalanche motion depend on an intricate system of factors that
855 promote volcano instability, trigger the collapse, and enhance the avalanche mobility. It is
856 therefore important to understand all the antecedent factors and triggers of volcano
857 instability since that knowledge can provide critical precursory information that can directly
858 inform hazard warnings.

859

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861

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864

865

866

867 **References**

868

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1350 Figure captions:

1351 Figure 1 (a) Model showing crater opening perpendicular to dike propagation (Nakamura 1977). (b)
1352 Model off crater opening perpendicular to the strike of normal faults and σ_{Hmax} (Francis and Self
1353 1987). (c) Crater openings in Andean volcanoes (Francis and Wells 1988). (d) Crater opening at an
1354 acute angle to the underlying strike-slip fault (Lagmay et al. 2000).

1355 Figure 2 Analogue experiment results by Wooller et al. 2009 showing the fault patterns created,
1356 areas of instability and the main control of the main geometrical parameters. Figures from Wooller
1357 et al., (2009).

1358 Figure 3 Google Earth image showing the Colima Volcanic Complex formed by the youngest Fuego
1359 de Colima volcano grown into the Paleofuego volcano depression (dashed line). Both volcanoes are
1360 located on the southern flank of the older and inactive Nevado de Colima volcano.

1361
1362 Figure 4 Example of highly hydrothermally altered material in a VDAD at Nevado del Ruiz, Colombia.
1363 *Espeletia* plants for scale at the top of the cliff are ~1.5 m tall (photo: Matteo Roverato).

1364
1365 Figure 5 USGS photograph taken by Peter Lipman on April 27, 1980, showing the impressive bulge on
1366 the northeastern flank of the Mount St. Helens volcano few days before its collapse.

1367
1368 Figure 6 Moderate resolution, SENTINEL-2 multispectral images of the Anak Krakatau A) before
1369 (1st November 2018; true colour) and B) after (2nd February 2019; true color) the lateral collapse
1370 occurred on 22nd December 2018. After the collapse, the intense explosive volcanic activity started
1371 to re-fill the collapse scar, producing a small coastal tuff ring.

1372
1373 Figure 7 Sketch representation of the instability factors that can destabilize a volcano and the
1374 triggering mechanisms, such as explosive activity and earthquakes, which promote a volcanic sector

1375 collapse. The top right rectangle represents schematically how the hydrothermal activity affects a
 1376 volcanic edifice, promoting leaching and dissolution of the rock, favoring the formation of voids and
 1377 the migration and accumulation of clay minerals, and consequently, fluid storage and
 1378 impermeability. This promotes loss of strength and shear reduction of the rock. See the text for
 1379 more details.

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 1381

1382 Tables:

1383

1384 Table 1 References used in the present text regarding factors that lead to the instability of volcanic
 1385 edifices and trigger a volcanic sector collapse. Note that some references are underlined because
 1386 they do not refer directly to volcanic lateral collapse but are still relevant for understanding aspects
 1387 of the processes promoting a volcanic edifice failure.

1388

1389 Table 2 Examples of volcanic debris avalanche deposits that display hydrothermal alteration

1390

1391 Table 3 Key features in a VDAD that could suggest what happened in the volcano edifice before
 1392 and/or during the lateral collapse event

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Table 1

Instability factors	References
Basement tectonics and faults	Nakamura 1977; Moriya 1980; Bahar and Girod 1983; Siebert 1984; <u>Delaney et al. 1986</u> ; Ui et al. 1986; Francis and Self 1987; Francis and Wells 1988; Carracedo 1994; <u>Tibaldi 1995, 1996</u> ; Elsworth and Voight 1996; van Wyk de Vries and Merle 1996, 1998; Lagmay et al. 2000; Vidal and Merle 2000; Norini and Lagmay 2005; Tibaldi et al. 2005; <u>Lagmay and Valdivia 2006</u> ; Norini et al. 2008; Wooller et al. 2009; Norini et al. 2010; Paguican et al. 2012; Norini et al. 2019; Norini et al. 2020
Sloping substrate and gravitational spreading	van Bemmelen 1949; Borgia et al. 1992; Vallance et al. 1995; Merle and Borgia 1996; van Wyk de Vries and Francis 1997; Francis and Wells 1998; van Wyk de Vries and Matela 1998; Borgia et al. 2000; van Wyk de Vries et al. 2001; Deplus et al. 2001; Clavero et al. 2002; <u>Kuhn 2002</u> ; Morgan et al. 2003; Lundgren et al. 2004; Wooller et al. 2004; Carrasco-Núñez et al. 2006; Siebert et al. 2006; Del Camp et al. 2008; Shea et al. 2008; Norini et al. 2009; Andrade and van Wyk de Vries 2010; Norini et al. 2010; Shea and van Wyk De Vrie 2010; Bonforte et al. 2011; Norini and Acocella 2011; Roverato et al. 2011; Paguican et al. 2012; Murray et al. 2018; Norini et al. 2019
Hydrothermal alteration	<u>Frantz et al. 1981</u> ; Siebert 1984; Carrasco--Núñez et al. 1993; Lopez and Williams 1993; <u>Frank 1995</u> ; <u>Scott et al. 1995</u> ; Day 1996; Kerle and Zimbelman 1996; <u>Vallance and Scott 1997</u> ; Voight and Elsworth 1997; Van Wyk de Vrie et al. 2000; van Wyk de Vrie 2001; Reid et al. 2001; Waythomas and Wallace 2002; Merle et al. 2003; Reid 2004; Zimbelman et al. 2004; Alvarado et al. 2004; Cecchi et al. 2005; <u>John et al. 2008</u> ; Shea et al. 2008; Borselli et al. 2011; Roverato et al. 2011; Delmelle et al. 2015; Rosas-Carbajal et al. 2016; Valderrama et al. 2016; Norini et al. 2020

Dikes and magma intrusions	Gorshkov 1959; Gorshkov and Bogoyavkenskaya 1965; Lipman et al. 1981; Voight et al. 1983; Siebert et al. 1987; Beget and Kienle 1992; Elsworth and Voight 1995; Day 1996; Elsworth & Voight 1996; Donnadieu and Merle 1998; Belousov et al. 1999; Donnadieu et al. 2001; McClelland and Erwin 2003; McGuire 2003; Belousov et al. 2004; Reid 2004; Thomas et al. 2004; Macias et al. 2010; Borselli et al. 2011; Norini and Acocella 2011; Villemant et al. 2014; Samaniego et al. 2015; Rincon et al. 2018
Past and present climate implications	Carrasco-Nuñez et al. 1993; McGuire 1996; <u>Clapperton 1998</u> ; Waythomas and Begét 2000; Kerle and van Wyk de Vries 2001; Scott et al. 2005; Capra 2006; Huggel et al. 2007; <u>Huggel et al. 2008</u> ; <u>Bennett and Glasser 2009</u> ; Tormey 2010; Tuffen 2010; Roverato et al. 2011; Capra et al. 2013; Roverato and Capra 2013; Roverato et al. 2015; <u>Kos et al. 2016</u> ; <u>Grämiger et al. 2017</u> ; Roberti et al. 2017; Roberti et al. 2020 (this volume)
Triggering mechanisms	References
Explosive eruptions (Bezymianny/Bandai types)	Gorshkov 1959; Gorshkov 1962; Moriya 1980; Bogoyavlenskaya and Kirnasov 1981; Ui 1985; Siebert et al. 1987; Voight and Elsworth 1997; Leyrit 2000; Donnadieu et al. 2001, Belousov et al. 2004; Dzurisin 2018
Earthquakes (Unzen/Ontake type)	Christiansen et al. 1981; Endo et al. 1981; Lipman and Mullineaux 1981; Voight et al. 1981; Keefer 1984; Ui et al. 1986; Fairchild 1987; Endo et al. 1989; Watanabe et al. 1993; Voight and Sousa 1994; Glicken 1996; Usami, 1996; Voight and Elsworth 1997; Hoshizumi et al. 1998; Sano et al. 1998; Inoue 2000; Donnadieu et al. 2001; McGuire 2003; Miyamoto 2010; Sassa et al. 2014; Della Pasqua et al. 2016; Kataoka et al. 2018; Wang et al. 2019

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Table 2

Deposit	Volcano	Reference
Teteltzingo VDAD	Citlaltepētł volcano, Mexico	Carrasco-Núñez et al. (1993)
1998 VDAD	Casita volcano, Nicaragua	Kerle and van Wyk de Vries (2001)
Chetaslina mass-flow	Mount Wrangell, Alaska, USA	Waythomas and Wallace (2002)
Piton de la Fournaise VDAD	Reunion Island, France	Merle and Lenat (2003)
Fortuna VDAD	Miravalles volcano, Costa Rica	Alvarado et al. (2004)
El Crater VDAD	Mombacho volcano, Nicaragua	Shea et al. (2008)
Tonila VDAD	Colima volcano, Mexico	Roverato et al. (2011)

Tutupaca VDAD	Tutupaca volcano, Peru	Valderrama et al. (2016)
Chimpa Unit	Chimpa volcano, Argentina	Norini et al. (2020)

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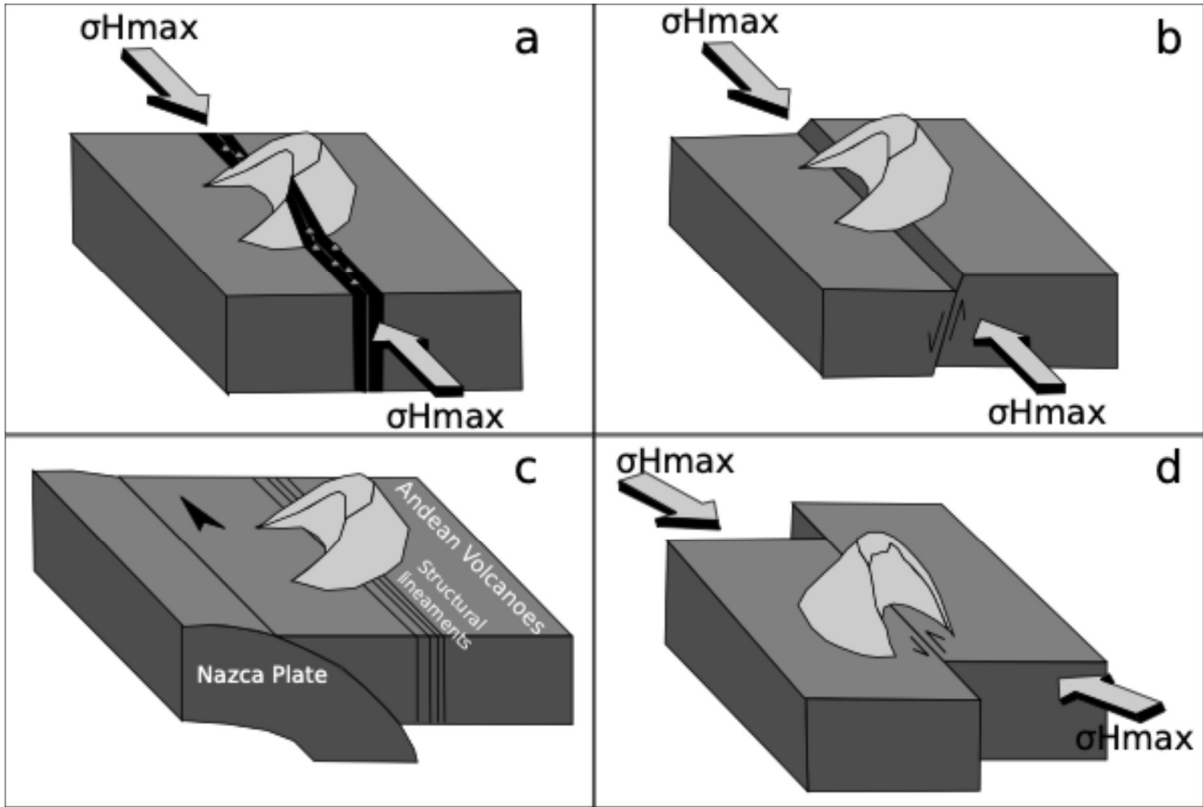
Table 3

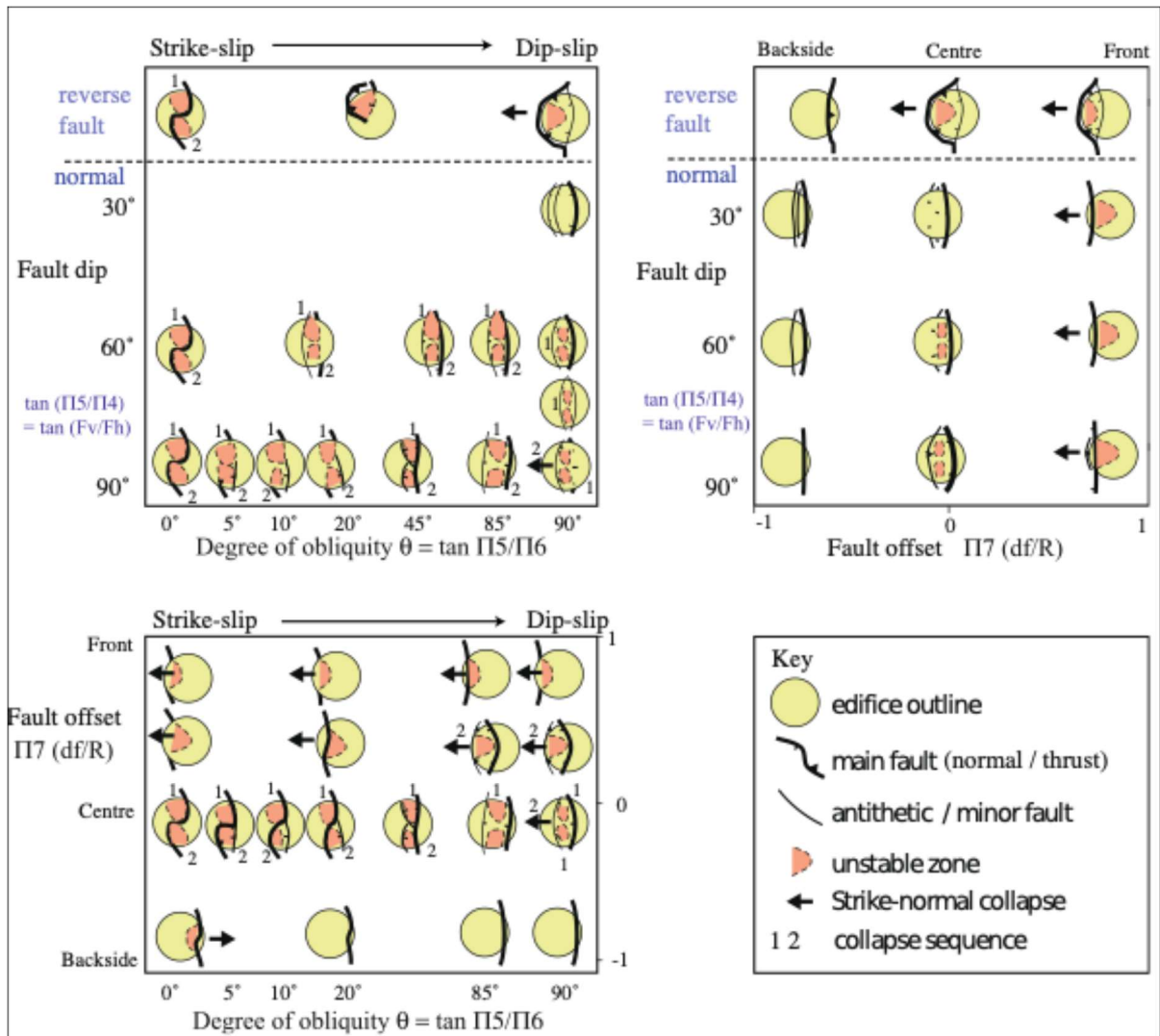
Key features		Pre/syn-collapse setting
Hydrothermal alteration	<p>The VDAD displays:</p> <ul style="list-style-type: none"> - color stains in those reaches (e.g. proximal) where the volcanoclastic material is not mixed with other materials; - homogeneous yellowish matrix in portions (e.g. medial/distal) where the material is partially or completely mixed; - high clay content 	<p>The edifice was affected by hot fluid circulation favoring hydrothermal alteration responsible for the weakening of the strength of the rock and promoting instability of the volcano.</p>
Water content	<p>The VDAD is characterized by a water-saturated (or semi-saturated) aspect displaying facies with characteristics between debris-flow deposits (DFDs) and VDADs</p> <ul style="list-style-type: none"> - homogenous and partially cemented matrix- preservation of millimetric porosity <p>The VDAD is often associated to DFDs that lying directly on top of it</p>	<p>The edifice was affected by high water content due to different climatic conditions such as:</p> <ul style="list-style-type: none"> - high precipitation in tropical settings - presence of snow and/or ice in boreal settings or at high altitudes in tropical settings <p>These factors contribute to the instability of the volcanic edifice and could also increase the area affected by a VDA through flow transformation into more mobile debris flows</p>
Pyroclastic deposits	<p>Pyroclastic deposits such as pyroclastic density current (PDC) or fall-out deposits lying directly in contact on top of the VDAD displaying sharp contacts</p>	<p>Pyroclastic material situated on top of the VDAD suggests that an eruption likely occurred after the collapse event as consequence of the depressurization of the magmatic system. The sharp contact between PDC and VDA deposits suggests that the VDA had already ceased to move when the PDC material emplaced. The magmatic event was likely not responsible for the collapse and other triggering mechanisms are involved.</p>

	<p>PDC deposits such as directed blasts and/or blast-generated PDC deposits are “intimately” associated with the VDAD. Some characteristics are:</p> <ul style="list-style-type: none"> - the contact between PDC and VDA deposits is very irregular - twisted clastic dikes of PDC material penetrate into the VDAD - in distal reached PDC deposits are located below and above the VDAD <p>These features could be associates to the following section (juvenile material).</p>	<p>These characteristics indicate that the VDA was still moving when the PDC material started to emplace on the upper surfaces of the VDA. PDC deposits located below the VDAD in distal portions suggest that the kinetic energy of the PDC material was higher than the VDA and that it emplaced before the arrival of the collapsing mass. This section could be associates with the following one (juvenile material).</p>
<p style="writing-mode: vertical-rl; transform: rotate(180deg);">Juvenile material</p>	<p>Fresh fragments of volcanic material (lava fragments or other pyroclastic material) are immersed into the VDAD</p>	<p>It indicates that the collapse was associated to a magmatic event that triggered the collapse. Fresh lava fragments could indicate the presence of a summit dome or a shallow intrusion. Vesiculated juveniles are the product of the explosive character of the eruption.</p>

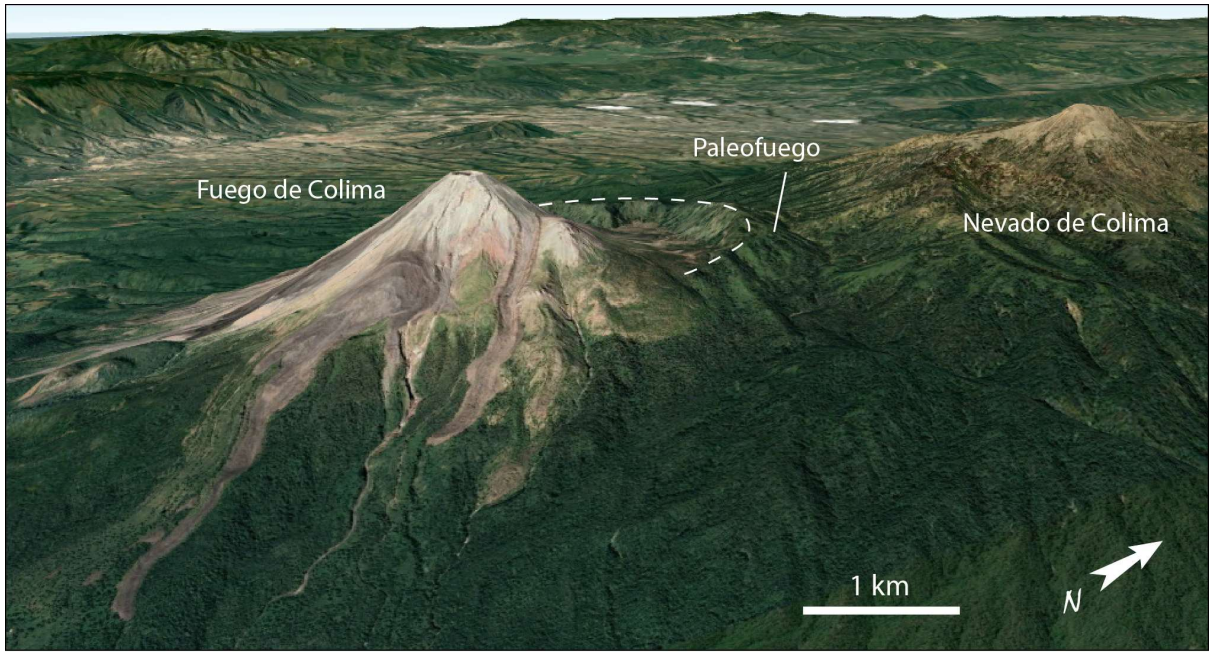
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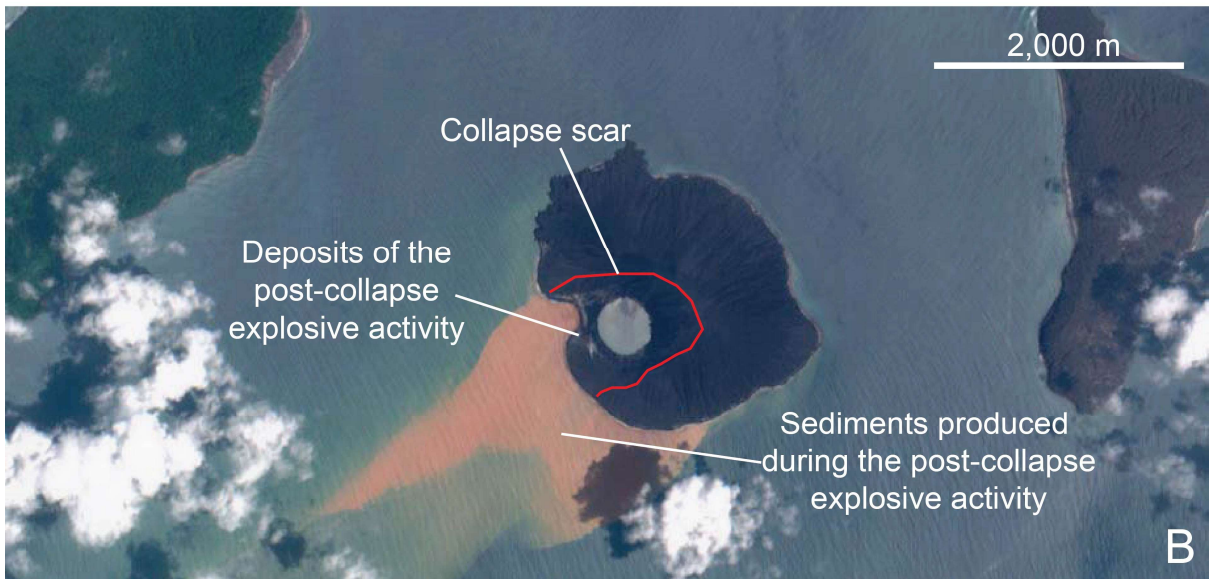
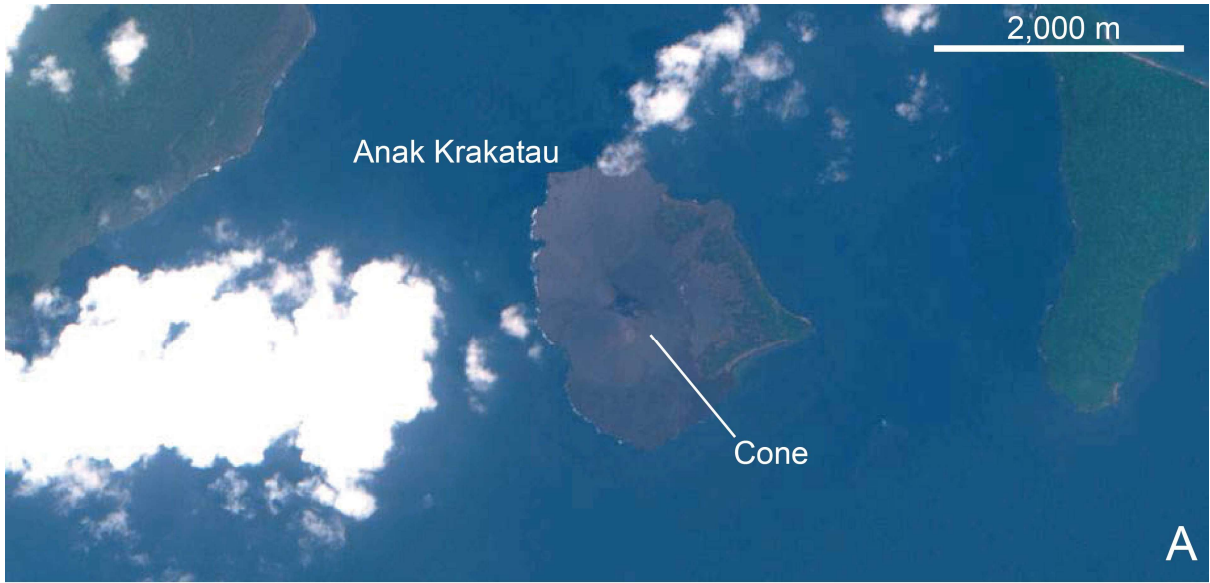
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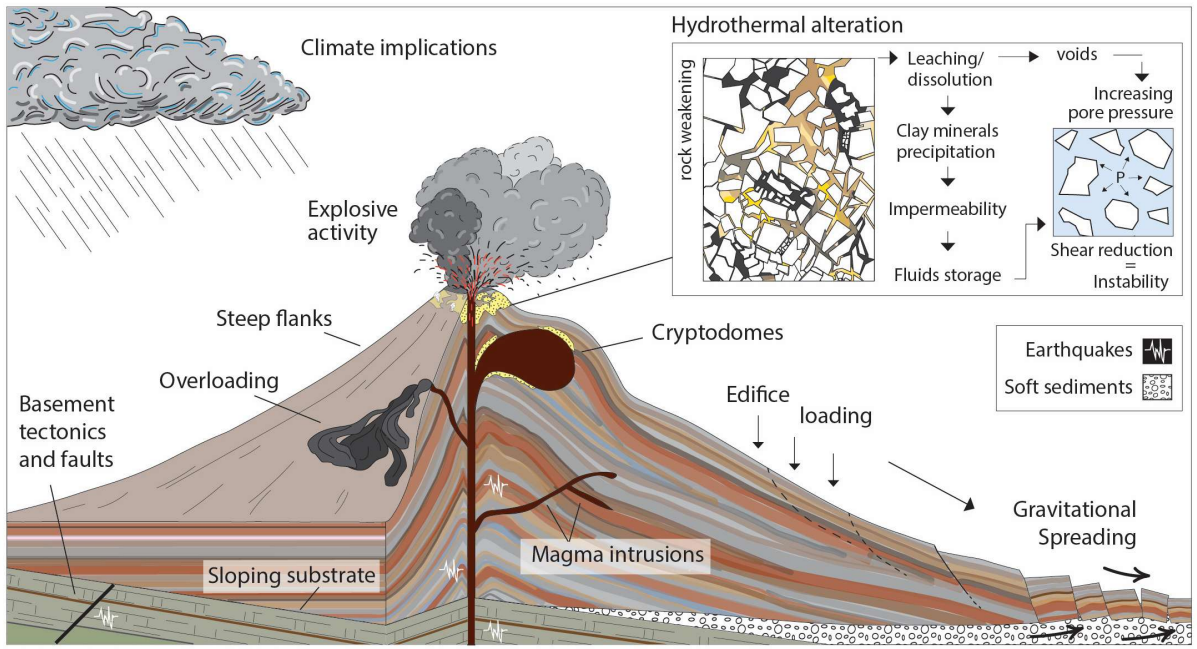
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