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# Marine Geology

## Cliff-top boulder morphodynamics on the high–energy volcanic rocky coast of the Reykjanes Peninsula (SW Iceland)

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<b>Abstract:</b>	<p>Cliff-top boulder deposits (CBDs) are morphological indicators of high-energy conditions. Since 2014, a monitoring of CBDs dynamics has been undertaken on the south-western coast of Iceland (Reykjanes Peninsula) to monitor their long-term activation (quarrying, transport and deposition) as a proxy of the inter-annual winter storminess variations and basaltic cliff erosion processes in a context of rocky coast progradation. Annual topomorphological surveys of four study sites were conducted and Structure-from-Motion photogrammetry was performed to quantify CBDs displacements. Hydrodynamic conditions were analyzed based on offshore waves and water level. Results show that CBDs activation occurs every winter, regardless of the variability of hydrodynamic conditions. Depending on the site and the year, more than 2% and 17% of the CBDs accumulated above 8 m to 10 m asl at the top of the cliffs are regularly mobilized. While inland movements represent the main mode of transport of blocks (between 50% to 60%), seaward and longshore movements are also well represented (10% to 20%). Longshore displacement is favored by the wide tabular morpho-structural setting of the wave-scour cliff-top platforms, which is explained by the structure of pāhoehoe lava flows. The activation of CBDs –measured from the volumes of displaced boulders–, shows a good correspondence with the frequency and duration of storms. However, as was the case during the winter of 2018-2019, it was rather the intensity of two highly morphogenic episodes combining storm waves and especially very high spring tide water levels, that generated the largest boulders displacements. Substantial interannual activation of the CBDs confirms that they constitute an important and still understudied proxy of the morphogenic system of high-energy rocky coasts, whose the dynamic in terms of carrying, transport, and deposition, could significantly increase with rising sea level.</p>
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# Cliff-top boulder morphodynamics on the high-energy volcanic rocky coast of the Reykjanes Peninsula (SW Iceland)

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## Abstract:

Cliff-top boulder deposits (CBDs) are morphological indicators of high-energy conditions. Since 2014, a monitoring of CBDs dynamics has been undertaken on the southwestern coast of Iceland (Reykjanes Peninsula) to monitor their long-term activation (quarrying, transport and deposition) as a proxy of the inter-annual winter storminess variations and basaltic cliff erosion processes in a context of rocky coast progradation. Annual topomorphological surveys of four study sites were conducted and Structure-from-Motion photogrammetry was performed to quantify CBDs displacements. Hydrodynamic conditions were analyzed based on offshore waves and water level. Results show that CBDs activation occurs every winter, regardless of the variability of hydrodynamic conditions. Depending on the site and the year, more than 2% and 17% of the CBDs accumulated above 8 m to 10 m asl at the top of the cliffs are regularly mobilized. While inland movements represent the main mode of transport of blocks (between 50% to 60%), seaward and longshore movements are also well represented (10% to 20%). Longshore displacement is favored by the wide tabular morpho-structural setting of the wave-scour cliff-top platforms, which is explained by the structure of pāhoehoe lava flows. The activation of CBDs –measured from the volumes of displaced boulders–, shows a good correspondence with the frequency and duration of storms. However, as was the case during the winter of 2018-2019, it was rather the intensity of two highly morphogenic episodes combining storm waves and especially very high spring tide water levels, that generated the largest boulders displacements. Substantial interannual



54 activation of the CBDs confirms that they constitute an important and still understudied proxy  
55 of the morphogenic system of high-energy rocky coasts, whose the dynamic in terms of  
56 carrying, transport, and deposition, could significantly increase with rising sea level.  
57

58 **Keywords:** boulder, cliff-top, storm, basaltic coast, survey, Iceland.  
59

## 60 **1. Introduction**

61  
62 Coastal boulder deposits (CBDs) are supratidal coarse-clastic sediments dislodged and  
63 transported by high-energy waves to the coast. They are widespread geomorphological features  
64 of high-energy rocky coasts such as the Atlantic coasts of western Europe (Williams and Hall,  
65 2004; Hall et al., 2006; Suanez et al., 2009; Oliveira et al., 2020b; Cox et al., 2012) and NW  
66 Africa (Mhammdi et al., 2019), the Pacific coasts –including Tasman Sea coast–, of eastern  
67 Asia (Goto et al., 2011, Kennedy et al., 2016, 2017), and Australia and New-Zeland (Susmilch,  
68 1912; Bishop and Hughes, 1989; Shelley, 1968), or the coasts of remote oceanic islands fully  
69 exposed to powerful oceanic swells (Etienne and Paris, 2010; Richmond et al., 2011b). CBDs  
70 are also found on the coasts of semi-enclosed maritime basins exposed to intense  
71 meteorological disturbances such as along the Mediterranean coasts (Deguara and Gauci, 2017;  
72 Delle Rose et al., 2020; Pepe et al., 2018; Piscitelli et al., 2017; Scicchitano et al., 2020; Terry  
73 et al., 2016).

74 CBDs are probably one of the most controversial types of subaerial coastal deposits  
75 among coastal sedimentary assemblages (Etienne et al., 2011; Kennedy et al., 2021; Lau and  
76 Autret, 2020; Paris et al., 2011). Their topographic positions on supratidal cliff-top platforms  
77 can reach between 10 to 20 m above the high spring water level, up to 30 m in some cases, and  
78 weigh up to 620 tons for the most massive (Cox et al., 2018a). A large part of the dedicated  
79 literature on CBDs focuses on whether they are tsunami or storm deposits (Nott, 2003; Saintilan  
80 and Rogers, 2005; Williams and Hall, 2004; Switzer and Burston, 2010; Richmond et al.,  
81 2011a; Prizomwala et al., 2015; Kennedy et al., 2021). In recent years, numerous studies have  
82 demonstrated the critical involvement of giant storm waves in the quarrying, transport, and  
83 deposition processes of CBDs, through (i) their morphosedimentary characteristics (Cox et al.,  
84 2018b; Etienne and Paris, 2010; Foster et al., 1991; Goto et al., 2010; Nanayama et al., 2000),  
85 (ii) the surveys carried out before and after high-energy storm events (Autret et al., 2016a;  
86 Fichaut and Suanez, 2010, 2011; Goto et al., 2012; May et al., 2015; Richmond et al., 2011a;  
87 Spiske and Bahlburg, 2011; Cox et al., 2018a), (iii) the critical analysis of theoretical boulder  
88 transport equations based on hydrological forcing (Dewey and Ryan, 2017; Lau et al., 2016;  
89 Prizomwala et al., 2015; Salzmann and Green, 2012; Shah-Hosseini et al., 2016; Soria et al.,  
90 2018; Kennedy et al., 2017, 2021), and (iv) the modelling and/or laboratory experiments  
91 (Bressan et al., 2018; Cox et al., 2019; Hansom et al., 2008; Watanabe et al., 2019; Weiss and  
92 Sheremet, 2017; Zainali and Weiss, 2015; Kennedy et al., 2016).

93 Cliff-top boulder deposits are highly persistent landforms in coastal landscapes. They  
94 constitute sedimentary archives capable of providing essential information on past wave  
95 climates extreme nearshore wave events (Ávila et al., 2020; Oliveira et al., 2020a). This is a  
96 key issue for future research on coastal boulder deposits (Cox et al., 2019), but we still have  
97 limited knowledge on their preservation potential. Therefore, CBDs may be considered as  
98 relevant geological records in coastal environments especially from their genesis to the moment  
99 they are considered as inert deposits (Cooper et al., 2019; Spiske et al., 2019). As indicated by  
100 Cox et al. (2019), they represent an effective proxy for long-term storminess analysis. The high  
101 frequency monitoring of CBDs carrying, transport and deposition may provide valuable  
102 information on the significant long-term variations of meteo-oceanic conditions, especially for  
103 the winter storm-periods. Coastal monitoring on sandy coast has been conducted in many

104 multi-decade research programs across the world and has proven its effectiveness to build up  
105 knowledge on the functioning of coastal environments driven by regular storms (Turner et al.,  
106 2016; Jaud et al., 2019; Banno et al., 2020; Castelle et al., 2020; Bertin et al., 2022; Nicolae  
107 Lerma et al., 2022). Some experiments of CBD monitoring based on at least one annual  
108 measurement have been undertaken (Nagle-McNaughton and Cox, 2020), notably on the  
109 Brittany coast (Autret et al., 2018), but they remain very anecdotal compared to the monitoring  
110 carried out on sandy beaches.

111 In this paper, we present the first results of a topo-morphological monitoring of CBDs  
112 started in 2014 on the coast of the Reykjanes peninsula, located in SW Iceland. This paper aims  
113 to study the impact of winter storms on the erosion of basaltic cliffs, resulting in the  
114 accumulation of CBDs at their top. We present annual monitoring data between 2014 and 2019  
115 from 4 study sites corresponding to rocky cliff areas where CBDs are accumulated. These data  
116 are then analyzed considering winter weather and ocean conditions, specifically swell and  
117 water levels. Through this work, the correspondence between winter forcing conditions and  
118 CBD activation is analyzed.

119

## 120 2. Study area

121

### 122 2.1 Geological and geomorphological settings

123

124 The study area corresponds to a low rocky-cliff (with cliff-top elevations between 4 and  
125 15 m above mean sea level - asl) located on the south-western coast of the Reykjanes Peninsula,  
126 in the south-west of Iceland (Fig. 1). The Reykjanes Peninsula corresponds to the southern part  
127 of the Icelandic Western Volcanic Zone, which is the transition area between the Reykjanes  
128 Ridge to the West and the Western Rift zone to the East (Einarsson, 1991). In this area, the  
129 volcanic activity has been quasi-continuous since the Mid Pleistocene, with thirteen historic  
130 eruptions (post 1200 BP), and at least nine prehistoric eruptions (Einarsson et al., 1991;  
131 Einarsson and Jóhannesson, 1989; Jóhannesson, 1989; Jóhannesson and Einarsson, 1988b,  
132 1988a; Jónsson, 1983); the last eruption in Fagradalsfjall started in March 2021. This  
133 geological setting attests that this SW part of Iceland was never glaciated during the cold phases  
134 of the Quaternary period, and that as such, the boulders (e.g. CBDs) accumulated at the top of  
135 the cliffs were not eroded nor deposited by the glacier ice processes. Therefore, the Reykjanes  
136 Peninsula is indeed an active prograding lava delta, which corresponds to a particular case  
137 among rocky coast types, showing exacerbated morphosedimentary dynamics as a result of  
138 strong hydro-morphosedimentary disequilibrium (Marie, 2007; Ramalho et al., 2013).

139 Etienne and Paris (2010), and Autret et al. (2016b) described the lithostructural setting  
140 of this Reykjanes Peninsula rocky coast, focusing on the diversity of CBDs accumulated along  
141 more than 30 km of coastline.

142

143 **Fig. 1.** Location map. (a) Regional map of the North-East Atlantic showing boxed inset location of map (b). (b)  
144 Zoom on the study area. Squares refer to the four study sites. The tide gauges and the wave buoy for observational  
145 datasets are displayed as black triangles and polygon, respectively. The polar plot shows the wave directions at  
146 the WW3 grid output from January 1948 to April 2018 (located at W22.5°; N63.5°, 35 km south of Grindavík,  
147 ~250 m depth). (c) and (d) show the monthly statistics of modeled significant wave height and peak period  
148 between 1948 and 2018, respectively. Bars indicate monthly averaged wave statistical parameters with the  
149 corresponding standard deviation (vertical error bars).

150

151 The stack of lava flows through time is a determining factor in the morphology of the  
152 cliffs, the coastline shape and the cliff-top boulder deposits. The cliff-top boulder deposits stand  
153 on supratidal platforms that always correspond to the structural surface of pāhoehoe lava flows  
154 (Fig. 2). Basaltic pāhoehoe is generally characterized by a smooth or folded continuous crust

155 without (or few) rough topographies (Gregg, 2017). Typically, the structure of pāhoehoe lava  
156 surface is characterized by circular wrinkles which look like ropes and therefore is called ropy  
157 pāhoehoe. These surface features are due to the movement of very fluid lava under a congealing  
158 surface crust inducing a behavioral difference between the center and the top-skin of the lava  
159 flow. However, the surface texture of pāhoehoe surface varies widely with increasing distance  
160 from the lava source, displaying in some cases all kinds of major rough topographies of various  
161 shapes often referred to as lava sculpture (e.g., collapsed tumuli, unusual craters, craters with  
162 raised rims, lava ponds, lava-rise –inflation– pits, lava-rise, inflation plateaus, etc.) (Detay  
163 and Hróarsson, 2018).

164 In cross-section, the vertical structure of pāhoehoe flows presents 3 distinct zones  
165 (Aubele et al., 1988; Autret et al., 2016b; Gregg, 2017): (i) an upper crust, typically vesicular  
166 and glassy or microcrystalline; (ii) an interior massive zone; and (iii) a relatively thinner basal  
167 crust that is similar in morphology to the upper crust. The top of the upper vesicular zone is  
168 often a repetitive sequence of glassy and red or maroon scoria, or slab-like layers a few  
169 centimeters thick, with fluid surface textures (Fig. 2). The erosion of this zone by waves provides  
170 sand, rubbles, pebbles-size particles which constitute the major part of the CTSD's  
171 accumulations. The central non-vesicular (dense) zone is characterized by distinct vertical  
172 columns consisting of clustered large vesicles with internal platy structures favoring the  
173 quarrying of large boulders by waves along the cliff sections. These few larger boulders are  
174 generally deposited front of the CTSD's accumulations (Fig. 2).

175 This lithostructural context exerts a strong control on the clast size at a local scale.  
176 Pieces of basalt columns coming from flow interiors form the largest boulders (typically, b-  
177 axis > 50 cm). They sometimes break into smaller boulders, but most of the small boulders (b-  
178 axis < 50 cm) are typically brought by the erosion of superficial and interflow zones. In some  
179 cases (see Katlahraun site in Fig. 3c), few collapsed and well-rounded boulders are also  
180 projected from the tidal zone of the base of the cliff to the cliff-top (Autret et al., 2016b) (Fig.  
181 2).

182  
183

184 **Figure 2.** Lithostructural setting of the basaltic coastal bedrock of the Reykjanestá Peninsula. (a) Selatangar site;  
185 (b) Katlahraun site; (c) Reykjanestá site.

186

187 The four study sites (Kerling, Reykjanestá, Katlahraun and Selatangar) are presented in  
188 Fig. 3. They are located on post-glacial basaltic outcrops aged from ~1900 to ~7000 BP  
189 resulting from subaerial volcanism (Sæmundsson et al., 2016). As indicated above, because of  
190 their ages and locations, these CBDs have a modern origin and are only related to high-energy  
191 wind-generated wave erosion processes (Etienne and Paris, 2010).

192 Kerling (Fig. 3a) is a rocky promontory facing west. It is part of the Kerling volcanic  
193 complex. The basaltic substratum is aged ~1900–~2400 BP. The site is composed of three main  
194 geomorphological units: (i) bluffs of 2 m to 4 m high and oriented N-S; (ii) a barren and rough  
195 supratidal platform of 2 m to 4 m elevations asl where a few very coarse boulders are trapped;  
196 (iii) a boulder ridge of 165 m long, 70 m wide and 3 m tall on average. The boulder ridge is  
197 oriented N-S and faces West at a minimum inland distance ( $D_{\min}$ ) of 20 m from the edge of the  
198 supratidal platform and a maximum inland distance ( $D_{\max}$ ) of 75 m from the cliff edge.

199 Reykjanestá (Fig. 3b) is the southwestern tip of the Reykjanes Peninsula, exposed South  
200 to SW. The basaltic substratum is older than ~2400 BP and younger than ~7000 BP. It also  
201 presents three main geomorphological units: (1) vertical cliffs of 4.5 m to 5 m high, oriented  
202 N-S in the west part and NW-SE in the south part; (2) a cliff-top platform at an average  
203 elevation of 4.8 m asl, barren and relatively smooth; (3) a boulder ridge located close to the  
204 cliff edge ( $D_{\min} = 1$  m,  $D_{\max} = 14$  m), oriented N-S, 85 m long, 25 m to 30 m wide and 2 m tall  
205 on average.

206 The cliff-top boulder deposits of Katlahraun (Fig. 3c) stands on a basaltic substratum  
207 aged ~1900~2400 BP. The geomorphological setting of this site is more complex with two  
208 additional geomorphological units: (1) bluffs topped between 2 m and 3 m asl and oriented E-  
209 W; (2) an intermediate supratidal platform barren and moderately rough at an average elevation  
210 of 3 m asl; (3) steps separating the intermediate supratidal platform and the top platform (e.g.,  
211 P2 and P3 in Fig. 3c). The top of this step is considered as the cliff-top edge of this site; (4) a  
212 top platform at an average elevation of 4 m asl, barren and relatively smooth; (5) a boulder  
213 ridge of 150 m long, 30 m width and 2 m tall on average. The boulder ridge is oriented E-W  
214 and faces South. It is located at distances ranging from 15 m to 30 m from the cliff edge.

215 Selatangar (Fig. 3d) is part of the same lava flow unit as Katlahraun (aged ~1900~2400  
216 BP). It also presents three main geomorphological units: (1) vertical cliffs of 5 m high, oriented  
217 NW-SE; (2) a cliff-top platform barren and relatively smooth at an average elevation of 8 m  
218 asl; (3) a boulder ridge located at distances ranging from 8 m to 30 m from the cliff-edge,  
219 oriented NW-SE, 200 m long, 20 m to 40 m wide and 2 m to 3 m tall on average.

220  
221 **Fig. 3.** Presentation of the four study sites. Kerling (a), Reykjanestá (b), Katlahraun (c), Selatangar (d), showing  
222 for each site an oblique aerial photography, a detailed topography and a 2D view of surface profiles extracted from  
223 the DEMs.

224

## 225 2.2 Oceanographic settings

226

227 The Reykjanes Peninsula is located north of the North Atlantic storm track and therefore  
228 experiences frequent storm waves (Davies, 1972; Einarsson, 1976). The wave climate is  
229 characterized by a strong seasonal modulation of the monthly-averaged significant wave height  
230 ( $H_s$ ), ranging from 1.5 m in July to 3.6 m in January. The four winter months (December to  
231 March) are particularly energetic with  $H_s$  reaching 3.5 m in December ( $\sigma = 1.7$  m), 3.6 m in  
232 January ( $\sigma = 1.7$  m), 3.6 m in February ( $\sigma = 1.8$  m) and 3.2 m in March ( $\sigma = 1.7$  m) (Fig. 1c).  
233 The significant wave height ( $H_s$ ) frequently reaches 8 m during storms (Sigbjarnarson, 1986),  
234 with a probable maximum  $H_s$  of 16.7 m recorded by a buoy in January 1990 (Tomasson et al.,  
235 1997). As a result of the narrow continental shelf, wave energy poorly dissipates into nearshore  
236 waters. Ocean waves are a major driver of coastal change in Iceland, but the impacts of storms  
237 on coastal settlements (mostly overland flooding resulting from overwash and overtopping)  
238 have been clearly linked to the concurrent forcing during peak events, i.e., when peak storm  
239 waves coincide with high tides. For instance, damages observed on buildings, roads, and  
240 coastal structures during a major coastal flood on January 9, 1990, in Stokkseyri were among  
241 the most important ever observed in south Iceland since 1910 (Geirsdóttir et al., 2014). This  
242 event occurred under an extremely low-pressure system (928 hPa), contributing to an unusually  
243 high tide (above 4 m in Reykjavik in the bay of Faxaflói), and powerful waves reaching 23 m  
244 near Surtsey (Viggósson et al., 2016). In Grindavík, the tidal range is slightly lower (mean  
245 spring tide of 3.2 m compared to 3.8 m in Faxaflói) with a mean sea level of 1.88 m (Larusson,  
246 2010). Finally, there is no sea ice formation in this area of Iceland because of the warmer water  
247 discharge from the Irminger current. Wave energy is not attenuated by coastal sea ice during  
248 the winter months. Sea ice brought by the East Greenland Current from the polar basin is also  
249 too scarce to bring significant protection against winter storms (Etienne and Paris, 2010).

250

## 251 3. Methods

252

253 We quantitatively investigated cliff-top boulders and boulder ridges'  
254 morphosedimentary dynamics on the four study sites. Field surveys started in 2014 and are still  
255 ongoing. In this paper, the data and surveys conducted over the period 2014-2019 are presented  
256 (Table 1). The field measurement campaigns were carried out once a year in spring (usually in



257 May) after the winter storm period. The measurements consisted of topomorphological surveys  
 258 based on ground measurements and very high-resolution topographic monitoring using *Kite*  
 259 *Aerial Photography* (KAP) and *Unmanned Aerial Vehicle* (UAV).  
 260

261 **Table 1.**

262 Inventory of the topomorphological surveys achieved between 2014 and 2019.  
 263

	Field campaign activities	Methods	Sites
2014, May 11 to 18	Field reconnaissance, identification of CBDs, block size measurements, morphological measurements and field photographs	GPS-RTK, decameters, cameras	30 km of coastline from Selatangar to Kerling
2015, May 4 to 15	Topo-morphological survey	KAP, GPS-RTK	Selatangar, Katlahraun, Reykjanestá, Kerling
2016, May 9 to 19	Topo-morphological survey	KAP, GPS-RTK	Selatangar, Katlahraun, Reykjanestá, Kerling
2017, May 14 to 24, and August 27 to September 3	Topo-morphological survey	KAP, UAV (Phantom2 - DJI®), GPS-RTK	Selatangar, Katlahraun, Reykjanestá, Kerling
2018, May 16 to 26	Topo-morphological survey	UAV (Phantom4 Pro - DJI®), GPS-RTK	Selatangar, Katlahraun, Reykjanestá, Kerling
2019, May 12 to 19	Topo-morphological survey	UAV (Phantom4 Pro - DJI®), GPS-RTK	Selatangar, Katlahraun, Reykjanestá

264

### 265 3.1 Low-altitude aerial imagery and topography

266

267 Low-altitude aerial surveys (using KAP and UAV) were carried out, providing a  
 268 detailed orthorectified aerial imagery of the boulder deposition and mobility areas. Digital  
 269 Elevation Models (DEM) were created with Structure -from-Motion and Multi-View Stereo  
 270 (SfM-MVS) processing (Fonstad et al., 2013; Westoby et al., 2012; Nagle-McNaughton and  
 271 Cox, 2020; Jaud et al., 2019) (Fig. 4).

272

273 The KAP surveys were performed in May 2015, 2016 and 2017 with the following  
 274 settings (Table 1): (1) two kites were alternatively used as lifters, a Delta Trooper (Dan Leigh®)  
 275 for strong winds and a HQ Kites® Flowform 4.0 for light winds; (2) aerial photographs were  
 276 taken with a Ricoh® GR camera (16.2 MP resolution, fixed focal length of 28 mm in the 35  
 277 mm format) connected to the kite line by a BROOXES® Gent-X picavet set; (3) ground control  
 278 points (GCPs) and ground validation points (GVPs) were surveyed by a Topcon Hyper V  
 279 Differential GPS in Real Time Kinematics mode (Fig. 4). X, Y and Z coordinates were  
 280 referenced to EPSG 3057 (ISN93). The same KAP settings or a UAV (DJI®, Phantom 2 and 4  
 281 Pro equipped with a 20 MP camera with an equivalent focal length of 24 mm and stabilized on  
 282 the 3-axis) were used to collect aerial images in August 2017, May 2018 and May 2019 (note  
 283 that there is no data for Kerling in 2019 due to bad weather). KAP images were collected at  
 284 multiple angles and altitudes depending on weather conditions. The variability of the shooting  
 285 angles and heights did not affect the final data quality. Drone images were collected at a  
 286 constant altitude of ~50 m and angle of -88° (subnadir) with flight paths set up to provide 80%  
 287 overlap between consecutive images and 70% sidelap between contiguous paths. These settings  
 288 were selected according to previous studies' recommendations and performed very well  
 289 (Dandois et al., 2015; Matthews, 2008; Mosbrucker et al., 2017; Torres-Sánchez et al., 2018).

290

291 **Fig. 4.** SfM-MVS methodological workflow. (a) Raw images are first collected by KAP or UAV with a high rate  
 292 of overlap along parallel transects. Ground markers corresponding to control (GCP) and validation (GVP) points  
 293 have previously been positioned and measured on the study area. (b) step by step framework from data collection  
 294 to work files generation. (c) shows an example of a volumetric differential between two surveys in the white-  
 295 dotted box shown on (a). The left panel shows a surface rendering, and the right panel shows the corresponding  
 diagrams.

296

297 We used SfM-MVS algorithms to reconstruct the georeferenced dense point clouds and  
298 the orthorectified aerial images from the aerial images sets. For each survey, a set of about 200  
299 to 300 images were processed separately. The resulting dense point clouds were classified using  
300 a Cloth Simulation Filter (Zhang et al., 2016) in order to remove the vegetation and the storm  
301 debris (especially driftwood and many fishing buoys and ropes). We then generated DSMs at  
302 a 3 cm resolution by an Inverse Distance Weighting interpolation with a power of 2 (in this  
303 way, known values (the generated points) have more weight and the interpolation is closer to  
304 the ground truth). The DSMs accuracy in X, Y and Z were estimated by calculating the RMSE  
305 from GVPs independent of the GCPs network; the margins of error are between 0.044 and  
306 0.025 m.

307

### 308 *3.2 Geomorphic changes*

309

310 Once data collected and postprocessed, interannual cliff-top boulder displacements and  
311 related morphosedimentary dynamics were investigated in two steps: (1) by computing the  
312 digital elevation model of differences (DoDs) from one survey to another in order to quantify  
313 the overall geomorphic changes, and (2) by tracking the displaced boulders that we were able  
314 to identify with certainty in order investigate boulder transport.

315 The DoDs were computed using the quantitative differencing method originally  
316 implemented by Wheaton et al. (2009), which improves the way to consider the total  
317 uncertainty of several topographic surveys in the DoDs, including uncertainties in the  
318 topographic survey data, the propagated error into the DoD and the significance of DoD  
319 uncertainty. A vertical detection threshold was set to 10 centimeters, minimizing the volumetric  
320 computation errors associated with SfM photogrammetry for the reconstruction of rough 3D  
321 surfaces. Nevertheless, this underestimates the net changes and the volumes given hereafter  
322 should therefore be considered as an order of magnitude rather than exact values.

323 When it was possible, individual transported and recognizable boulders were tracked  
324 using GIS geospatial tools. According to the topographic changes highlighted by the DoDs and  
325 field or high-resolution orthomosaic validations, pre- and post-transport boulders positions  
326 were measured. For each transported boulder found into its new position, we measured: (i) the  
327 length of the *a*-, *b*- and *c*-axis (*A*, *B*, *C*); (ii) the pre- and post-transport location (start\_x, start\_y,  
328 start\_z, end\_x, end\_y, end\_z); (iii) the closest pre- and post-transport distance from the seaward  
329 edge of the supratidal platform (start\_d, end\_d). The direction and the length of transport were  
330 then computed (transport\_length and transport\_bearing, respectively). Finally, the *a*-, *b*- and *c*-  
331 axis measurements were used to estimate the volume of each boulder, from which 20% were  
332 subtracted before computing the mass. One-way analysis of variance (ANOVA) was used to  
333 compare groups of displaced boulders in terms of masses, elevation and distance. A post-hoc  
334 Tukey test was performed to identify group pairing discrepancies. However, as noted above,  
335 most of the boulder movements associated with the morphological changes in the boulder  
336 ridges and/or clusters as a whole, could not be measured because it was not possible to identify  
337 the start and/or end zone of the boulder movement.

338

### 339 *3.3 Analysis of the meteo-oceanic conditions*

340

#### 341 *3.3.1 Wave conditions and water levels during the survey period (2014-2019)*

342

343 Hourly observed water levels covering the period between June 2014 and April 2019  
344 have been retrieved from the tide gauge in Sandgerdi Harbor, ~25 km from the study area (Fig.  
345 1b). This gauge was selected in order to overcome large data gaps at the Grindavik tide gauge

346 in 2017. Following [Bernier and Thompson \(2015\)](#), the water level have been decomposed into  
 347 tidal components and surge residual using the Matlab tidal analysis package T\_TIDE  
 348 ([Pawlowicz et al., 2002](#)). Note that local effects induced by basin morphology and barometric  
 349 effects (e.g., seiches) can impact the overall surge signal ([Donn and Wolf, 1972](#)). Therefore, it  
 350 is expected that the surge signal recorded in Sandgerði Harbor may not represent accurately  
 351 the conditions in Grindavík. For this reason, less emphasis has been laid on this aspect in the  
 352 present study.

353 Hourly offshore wave characteristics between December 1995 and April 2019,  
 354 including significant wave height ( $H_s$ ) and average wave period ( $T_z$ ) were provided by the  
 355 *Icelandic Road and Coastal Administration* for the Grindavík buoy (63°48.80'N; 22°27.63'E;  
 356 62 m depth) ([Fig. 5b](#)). This 24-yr continuous timeseries was used for the wave model  
 357 validation ([Fig. 5c](#)). Only the 2014-2019 timeframe was used for the current analysis of  
 358 hydrodynamic conditions related to morphological changes.

359 Storminess was analyzed using a Peak Over Threshold method (POT) with a two-step  
 360 threshold. Considering the high-energy wave climate near the study sites, the 99<sup>th</sup> percentile of  
 361 the modeled significant wave height time series was used as a primary threshold for the  
 362 strongest storms. The start and end time of the storm have been set according to a lower up-  
 363 and down-crossing of the 95<sup>th</sup> percentile of the  $H_s$  time series, following the storm definition  
 364 of [Masselink et al. \(2015\)](#) applied on the southwest coast of England. Ensuring that these storm  
 365 events are all independent ( $H_s$  maxima are separated by at least a 24h independence criterion),  
 366 observed water levels ( $WL$ ) both at each maximum and during the events were extracted using  
 367 the Sandgerði gauge as the reference to still water level.

368  
 369 **Fig. 5.** Modelled (a) and observed (b) offshore significant wave height ( $H_s$ ) between 1948 and 2019. The WW3  
 370 validation period covers the period between 1995 and 2017. Debiased modelled  $H_s$  are plotted against  
 371 observations at the Grindavík buoy (c). Example of surge waves overtopping cliffs at Reykjanestá on May 19,  
 372 2018 under stormy but not extreme wave conditions ( $H_s = 4$  m) (d).

373  
 374 Joint probability occurrence of storms was quantified with copulas for the available  
 375 observations period (2011-2019) using the MhAST toolbox ([Sadegh et al., 2018](#)). A time series  
 376 of pairs of interest of the two hazard drivers  $H_s/WL$  was constructed using the  $H_s$  and  $WL$   
 377 maxima during each storm. The joint time series is short (8 years) but can be useful to  
 378 understand the combinations of variables that could trigger boulder movement during this  
 379 period. Following Sklar's theorem ([Nelsen, 2006](#)), if  $F_H(H_s)$  and  $F_W(WL)$  are the marginal  
 380 distributions of significant wave height ( $H_s$ ) and water levels ( $WL$ ), a copula  $C$  combining these  
 381 two marginal distributions represents the joint distribution function following:

382  
 383 
$$F(H_s, WL) = C(F_H(H_s), F_W(WL)) = C(u_1, u_2)$$

384  
 385 If marginal distributions are continuous,  $C$  is unique and the joint probability density  
 386 function  $c()$  can become:

387 
$$c(u_1, u_2) = \frac{\partial^2 C(u_1, u_2)}{\partial u_1, u_2}$$

388  
 389 MhAST is based on the Multivariate Copula Analysis Toolbox (MvCAT) ([Sadegh et al.,](#)  
 390 [2017](#)) and ranks each copula performance based on Likelihood, Akaike Information Criterion  
 391 (AIC), Bayesian Information Criterion (BIC), Nash-Sutcliffe Efficiency (NSE), and Root Mean  
 392 Squared Error (RMSE).

393  
 394 **3.3.2 Long-term assessment of the potential activation frequency (1948-2019)**

395

396 Offshore wave statistical parameters from January 1948 to March 2017 were retrieved  
397 from a Wavewatch III regional model hindcast covering the North Atlantic Ocean (0°-80°W;  
398 0°-70°N) at a 0.5° resolution, described and validated in [Masselink et al. \(2016\)](#), at location  
399 W22.5°; N63.5° (35 km south of Grindavík, ~250 m depth) ([Fig. 5a](#)). To validate the model  
400 results, simulated and measured  $H_s$  and  $T_p$  ( $T_p = 1.4 T_z$ ) at the Grindavík buoy located 3 km  
401 south of Grindavík at 62 m depth ([Fig. 1](#)) were compared for the period covering December  
402 1995 to March 2017 (the buoy time series extends up to April 2019 in [Fig. 5b](#)). Comparisons  
403 of modeled  $H_s$  ( $H_{s_{model}}$ ) with measured data show an averaged bias of 0.5 m, a squared  
404 correlation coefficient ( $\rho^2$ ) of 0.67, and a normalized root-mean square error of 0.34. Modeled  
405  $H_s$  were then linearly corrected following  $H_{s_{deb}} = H_{s_{model}}/0.96 - 0.58$ , resulting in a significantly  
406 lower bias (0.02 m for the validation period). No correction was applied to the wave period  
407 time series considering the bias under 0.2 s between simulated peak wave periods and  
408 observations during storm conditions (i.e., when  $H_{s_{deb}} > 7.5$  m).

409 To better contextualize the observed cliff-top boulders' morphodynamics between 2014  
410 and 2019, we analyzed the periodicity of annual offshore wave statistics and storm activity in  
411 SW Iceland over the historical period (1948-2019). Linear regressions were first computed to  
412 verify the presence of trends using the square correlation coefficient  $\rho^2$  at a 95% significance  
413 level. A wavelet analysis was performed using the Analytic Wavelet Transform toolbox in  
414 Matlab ([Aguilar-Conraria and Soares, 2014](#)). The Morlet wavelet function was applied ( $\omega 0 = 6$ )  
415 over the annual time series of storm occurrence and annual wave characteristics with a lower  
416 and upper period of 2 and 50 years (pad = 0).

417

## 418 **4. Results**

419

### 420 *4.1 Morphological changes and boulder transport*

421

422 Morphological changes were quantified by DoDs calculation using low-aerial  
423 topographic data surveyed between 2015 and 2019 (except for Kerling, not surveyed in 2019).  
424 Additional qualitative analysis was also based on the 2014 field reconnaissance. The results  
425 presented hereafter are expressed in cubic meters per site and per winter, and then expressed  
426 as the relative volume of the total volume of each cliff-top boulder accumulation (percentage  
427 per winter). Note that these values are orders of magnitude, not fully accurate values.

428 Between May 2015 and May 2016, ~522 m<sup>3</sup> of boulders were transported at Kerling  
429 (corresponding in proportion to ~2.3% of the total number of boulders), ~17 m<sup>3</sup> at Reykjanestá  
430 (~0.8%), ~68 m<sup>3</sup> at Katlahraun (~1.1%) and ~41 m<sup>3</sup> at Selatangar (~0.5%). The volume of  
431 transported boulders was higher between 2016 and 2017 than the previous year, with ~1081 m<sup>3</sup>  
432 at Kerling (~4.8%), ~100 m<sup>3</sup> at Reykjanestá (~4.8%), ~153 m<sup>3</sup> at Katlahraun (~ 2.4%) and  
433 ~103 m<sup>3</sup> at Selatangar (~1.3%). Between 2017 and 2018, ~665 m<sup>3</sup> of boulders were transported  
434 at Kerling, ~54 m<sup>3</sup> at Reykjanestá, ~69 m<sup>3</sup> at Katlahraun and ~33 m<sup>3</sup> at Selatangar. This  
435 corresponds to ~3% of the total number of boulders in the boulder ridge at Kerling, ~2.6% at  
436 Reykjanestá, 1.1% at Katlahraun and 0.4% at Selatangar. Finally, the greatest changes occurred  
437 during the last monitoring period (2018-2019) with ~369 m<sup>3</sup> of moved boulders at Reykjanestá  
438 (~17.6%), ~354 m<sup>3</sup> at Katlahraun (~5.6%) and ~141 m<sup>3</sup> at Selatangar (~1.7%). ([Fig. 6a](#) and  
439 [6b](#)).

440

441 The successive DEMs and orthophotos also helped to identify the start and end positions  
442 of  $n = 988$  recognizable boulders from one survey to the next, hereafter referred to as  
443 transported boulders. Between 2015 and 2016,  $n = 123$  boulders were transported (i.e.,  $n = 31$   
444 at Kerling,  $n = 35$  at Reykjanestá,  $n = 33$  at Katlahraun, and  $n = 24$  at Selatangar),  $n = 360$   
445 boulders between 2016 and 2017 (i.e.,  $n = 115$  at Kerling,  $n = 73$  at Reykjanestá,  $n = 83$  at



445 Katlahraun, and  $n = 89$  at Selatangar),  $n = 281$  boulders between 2017 and 2018 (i.e.,  $n = 83$  at  
446 Kerling,  $n = 50$  at Reykjanestá,  $n = 50$  at Katlahraun,  $n = 98$  at Selatangar), and  $n = 224$  boulders  
447 between 2018 and 2019 (i.e.,  $n = 39$  at Reykjanestá,  $n = 108$  at Katlahraun,  $n = 77$  at Selatangar).  
448 We consider the distance between these two positions as a marker of boulders' movement, but  
449 these vectors should not be interpreted as the true trajectories of the CBDs. Boulders can indeed  
450 experience multiple movements during the same winter period, especially if it is characterized  
451 by several storm events (Autret et al., 2016a).

452  
453 **Fig. 6.** Summary of transported boulders at each site between 2015 and 2019 expressed in absolute volume ( $\text{m}^3$ )  
454 (a) and relative volume of the total accumulation (%) (b). Boulder mass (kg) (c), elevation above sea level (m) (d)  
455 and inland distance from the cliff-top before transport (m) (e) are also shown.

456  
457 The mass, the elevation and the inland distance of the transported boulders are  
458 summarized in Fig. 6c and 6e. There is a statistically significant difference between mean  
459 transported boulder masses (mean = 1.47 tons,  $\sigma = 2.66$  tons) between sites, with significantly  
460 heavier transported boulders at Kerling compared to the other sites (one-way ANOVA  $p < 0.001$ ;  
461 post hoc Tukey HSD tested significance at the  $p < 0.001$  level). The heaviest boulders were all  
462 moved at Kerling (i.e., 29 tons between 2017 and 2018, 24 tons between 2015 and 2016, 22  
463 tons between 2016 and 2017), but regardless of the year, no significant interannual difference  
464 in averaged transported masses is observed at this site ( $p > 0.3$ ). Eastwards of Kerling, more  
465 variability in transported CBDs is observed. The movement of the heaviest, and statistically  
466 heavier boulders in average than other years ( $p < 0.001$ ), were recorded between 2016-2017 in  
467 Katlahraun or 2018-2019 in Selatangar and Reykjanestá.

468 In average, CBDs are found at 7.28 m asl ( $\sigma = 1.97$  m) (Fig. 6d), but in contrast to  
469 boulder mass, the averaged elevation of the transported boulders shows a statistically  
470 significant difference between all sites (one-way ANOVA  $p < 0.001$ ; post hoc Tukey HSD tested  
471 significance at the  $p < 0.001$  level). This is due to major topographic variabilities between all  
472 sites. Moreover, interannual variability in transported CBDs elevations varies greatly from one  
473 site to another. While transported CBDs elevations between 2015 and 2019 remain similar at  
474 Selatangar ( $p > 0.19$ ), transported boulders between 2016-2017 and 2017-2018 were higher than  
475 between 2015-2016 at Kerling ( $p < 0.05$ ). Statistically, averaged transported boulder elevations  
476 were higher (and at similar elevations) between 2016-2017 and 2018-2019 at Reykjanestá.  
477 Compared to these two periods at this southernmost cliff, transported boulder elevations were  
478 significantly lower between 2015-2016 and 2017-2018 ( $p < 0.001$ ). At Katlahraun, CBDs have  
479 been moved at higher elevations between 2018-2019 ( $p < 0.001$ ), and similarly to Reykjanestá,  
480 the averaged elevation of the moved boulders increased between the first two years of the  
481 surveys, followed by a decrease between 2016-2017 and 2017-2018, while an increase in  
482 transported boulder elevations is observed during the last year.

483 At Kerling, Katlahraun, and Reykjanestá (at this later site, the averaged boulder distance  
484 relative to the cliff edge are significantly shorter than other sites ( $p < 0.001$ )), boulders were  
485 moved from the supratidal platforms to the boulder fields (including the ridges). It was not the  
486 case at Selatangar, the higher site, where transport was more limited to the cliff-top platform  
487 and the face of the boulder ridges. The inland distance of the boulders before transport was of  
488 the same order of magnitudes between 2015 and 2019 at Selatangar and Kerling, except at  
489 Katlahraun and Reykjanestá where significant increases of ~41% ( $p < 0.001$ ) and ~20% ( $p < 0.05$ )  
490 are observed, respectively, at the end of the monitoring period (Fig. 6e).

491 Fig. 7 summarizes the raw transport directions. A large directional spreading in  
492 transported boulders is observed, both on a single site during a single period but also between  
493 each period. A majority of boulders was transported inland, which is consistent with the  
494 theoretical direction of wave-generated flows to the coast (60.7% at Kerling, 59.4% at  
495 Reykjanestá, 59.2% at Katlahraun and 59.4% at Selatangar). Seaward (22.3% at Kerling, 24.4%

496 at Reykjanestá, 27% at Katlahraun and 25.7% at Selatangar) and longshore transports (17% at  
497 Kerling, 16.2% at Reykjanestá, 13.8% at Katlahraun and 14.9% at Selatangar) are less  
498 important, suggesting deviated or reflected flows just as competent (Fig. 8a).

499 There is a slight difference in mass—several tens of kilos for all sites—between the three  
500 transport directions (Fig. 8b). The same applies for the inland distance of the boulders before  
501 transport. Boulders transported seaward were generally located further inland than inland and  
502 longshore transported boulders (Fig. 8d). On the other hand, there is an inverse relationship  
503 between the length of transport and the normalized transport direction (Fig. 8d). Inland  
504 transport is longer than longshore transport by factors 1.5 to 3 depending on the site, and  
505 longshore transport is slightly longer than seaward transport.

506  
507 **Fig. 7.** Boulder transport directions relative to their initial location along the cliff-top at each site (a to d) between  
508 2015 and 2019. The bin color scale indicates boulder masses (in tons), and the blue/green background shades in  
509 the roses represent seaward and inland transport, respectively.

510  
511 **Fig. 8.** Normalized CTSD transport orientation analysis between 2015 and 2019 (a). Panels (d) and (d) represent  
512 transported boulder mass (kg.), distances from cliff-top (m) and length of boulder transport (m) relative to their  
513 initial location (inland, longshore, seaward).

#### 514 4.2 Storm analysis over the survey period (2014-2019)

515  
516  
517 The wave climate along the southwestern coast of Iceland follows a strong seasonal  
518 cycle punctuated by lows (summer) and highs (winter) (Fig. 9a and 9b). Between June 2014  
519 and April 2019, the monthly averaged significant wave height ( $H_s$ ) was 2.2 m ( $\sigma = 0.9$  m)  
520 offshore Grindavík (monthly  $H_{s98\%} = 4$  m), with a mean peak period of 10.3 s ( $\sigma = 1.1$  s). During  
521 winter months (Dec., Jan., Feb., March), the monthly-averaged values of  $H_s$  and  $T_p$  increases  
522 to 3.15 m ( $\sigma = 0.6$  m) and 11.4 s ( $\sigma = 0.7$  s), respectively. The highest  $H_s$  reached 11.4 m ( $T_p$   
523 = 16.8 s) on March 16, 2015. March 2015 was one of the most energetic months with monthly  
524 averaged  $H_s$  reaching above 4m (monthly averaged  $T_p = 12.1$  s), but globally mean  $H_s$  values  
525 have been slightly higher in February 2018 (4.1 m,  $T_p = 12.04$  s). These two months were  
526 therefore characterized by waves reaching more than a meter above the maximum monthly-  
527 mean value observed during the winter of 2015-2016 ( $H_s = 3.1$  m in March).

528  
529 **Fig. 9.** Time series of corrected 6-hourly  $H_s$  modelled offshore Grindavík (a) (grey lines) and hourly observed  $H_s$   
530 at the Grindavík buoy from 2014 to 2019 (b). Monthly moving averages of observed (green line) and modelled  
531 (blue line) are also shown in (a) and (b). Storms are detected when  $H_s > 99^{\text{th}}$  percentile  $H_s$  (thick dashed black  
532 line, a-b). Their duration is proportionally illustrated with bubbles in panels (a) and (b), and is calculated as the  
533 time spent above the 95<sup>th</sup> percentile  $H_s$  (thin dashed black line) on either side of the peak storm. Water levels ( $WL$ )  
534 at Sandgerði (hourly) are shown in (c) for 2014-2019.  $WL$  at the peak  $H_s$  during each storm are plotted as diamond,  
535 whereas the most extreme concurrent event is indicated by circles at the maximum  $WL$  during each storm event.  
536 Residual surge is indicated as a blue line.

537  
538 Using the 1% exceedance  $H_s$  as a critical threshold for storm occurrence ( $H_{s99\%} = 7.6$   
539 m) and the lower secondary threshold  $H_{s95\%}$  as the trigger and end time of each event,  
540 observations show a total of 85 storms in southwest Iceland, offshore Grindavík, between  
541 December 1994 and May 2019. These storms (~3.5 storms per year) lasted 28h on average with  
542 mean significant wave heights ( $H_s$ ) of 6.9 m  $\pm$  0.7m. During the survey period (2014-2019), 22  
543 distinct storms have been detected (~4.4 storms per year). Averaging 31 hours  $\pm$  18h and lasting  
544 at least 17h, these events therefore lasted more than a tidal cycle and potentially had varying  
545 coastal impact intensities over the course of up to ~4 consecutive days (e.g., March 5, 2015).  
546 Most storms (8 events) occurred in winter 2014-2015 with a mean duration of 36 hours (mean  
547  $H_s = 7.1$  m and  $T_p = 14.2$  s, mean peak  $H_s = 9.3$  m and  $T_p = 16.7$  s) (Table 2). Unfortunately,  
548 no detailed topographic measurements were carried out in 2014 and therefore we cannot

549 quantitatively address the morphogenic impacts of the 2014-2015 winter storms on the boulder  
550 activation. In winter 2015-2016, 3 storms occurred under slightly less energetic conditions  
551 (mean peak  $H_s = 8.7$  m and  $T_p = 15.7$  s). Earlier events are observed in the following storm  
552 seasons. A total of 6 storms occurred between October 2016 and February 2017 with similar  
553 storm characteristics (mean peak  $H_s = 8.3$  m and peak  $T_p = 15.5$  s). Only 3 storms impacted  
554 the coast in winter 2017-2018 (mean peak  $H_s = 8.8$  m and  $T_p = 15.9$  s). The following period  
555 (2018-2019), the only 2 storms recorded peaked at low tide ( $<-0.8$  m), even though a  
556 particularly high tide (2.2 m) occurred during the February event.

557

558 **Table 2**

559 Storm characteristics during the survey period (2014–2019) on the southwest coast of the Reykjanes peninsula.  
560 Wave statistics ( $H_s$  and  $T_p$ ) are derived from offshore observations at the Grindavík buoy. Observed water levels  
561 (referenced to mean sea level) were acquired at the Sandgerði tide gauge.

562

Storm #	date/time at peak event	Max. $H_s$ (m)	Max. $T_p$ (s)	Mean $H_s$ (m)	Mean $T_p$ (s)	Duration (h)	Water level (m)	Max. water level (m)	Time interval between peak storm and max. water level (h)
1	2014-12-09 12:00	7.9	18.5	6.9	14.9	25	-0.9	1.6	8
2	2015-01-08 09:00	8.4	17.2	6.8	14.1	48	1.4	1.8	1
3	2015-01-19 09:00	9.3	15.9	6.9	13.8	32	-0.2	1.9	21
4	2015-02-18 16:00	9.2	16.8	7.1	14.2	19	0.6	1.7	2
5	2015-03-05 07:00	8.6	16.7	6.5	14.0	94	1.5	1.7	1
6	2015-03-10 21:00	9.1	15.3	6.8	14.1	25	1.7	1.7	0
7	2015-03-14 11:00	10.7	16.3	7.8	14.3	35	0.9	1.0	11.5
8	2015-03-16 08:00	11.4	17.1	8.1	14.5	20	-0.7	1.2	5
9	2015-12-31 09:00	7.9	15.5	6.3	14.0	26	1.1	1.3	0.5
10	2016-03-12 14:00	8.5	15.5	7.0	13.1	17	-1.9	1.9	6
11	2016-03-14 02:00	9.5	16.0	7.2	14.6	23	-0.9	1.7	4
12	2016-10-03 14:00	7.8	16.0	6.3	13.9	28	-1.4	1.8	7
13	2016-10-20 01:00	8.6	15.1	6.6	13.3	31	-1.0	1.9	8
14	2016-12-13 19:00	8.8	15.4	6.4	13.4	19	1.4	2.1	11
15	2016-12-27 20:00	8.7	15.6	7.3	13.9	46	0.3	1.7	10
16	2017-01-21 00:00	7.5	14.0	6.4	12.7	20	0.7	0.8	2
17	2017-02-07 07:00	8.4	16.7	6.8	13.3	66	-0.4	1.6	46
18	2017-11-05 20:00	8.2	14.9	6.5	13.4	29	1.9	2.2	11
19	2018-01-09 13:00	9.0	15.7	6.9	14.0	21	1.0	1.4	1
20	2018-02-24 07:00	9.2	17.0	7.6	13.9	28	-0.7	1.2	7
21	2018-12-11 16:00	7,6	13.5	6,5	13	21	-0.8	1.3	8
22	2019-02-22 01:00	9,2	16.6	7,6	15	18	-1.7	2.2	7

563

564 A statistical analysis based on the joint occurrence of storms –defined as energetic wave  
565 events–, and water levels at Sandgerði was conducted to determine favorable conditions  
566 associated to extreme water levels capable of overtopping the cliff top. During the joint time  
567 series of storm events (2011-2019), 34 storms have been detected. Maximal tidal range during  
568 spring tides is ~4.5 m to 5 m. The maximum surge level recorded is about 0.66 m (99% surge  
569 = 0.21 m). Peak water levels reached the 99<sup>th</sup> percentile value (99%  $WL = 1.81$  m) at least  
570 during one storm each year since January 2015, but the timing of the storm peak was not always  
571 synchronized with a high tide level (Table 2). Most peak storms occurred within  $\pm 6$  hours of  
572 the highest water level (53%), and about 21% of events occurred within  $\pm 1$ h. In the last decade,

573 critical extreme combination of both high  $H_s$  and high water levels at the same time occurred  
574 only once, on March 10, 2015 (Fig. 9c).

575 Multivariate analysis and copulas were used to calculate the joint probability of  $H_s$  and  
576 water levels. Despite the short duration of the joint time series (8 years) –and acknowledging  
577 that copula model parameters depends on the length of observations (Sadegh et al., 2018)–, the  
578 best copula (Fig. 10a) describing the correlation structure between the peak storm (max  $H_s$ )  
579 and the highest water level is the Raftery copula (RMSE = 0.1098) (Nelsen, 2006). Fig. 10a  
580 shows the absence of correlation between both variables, but the joint probability contours  
581 indicate some variability in storm events characteristics. For values of  $H_s$  slightly above the  
582 detection threshold (7.5–8 m), joint probability contours are nearly vertical, indicating that these  
583 storm wave conditions can be observed under varying water levels. Moreover, storms during  
584 which water levels overpass the 99<sup>th</sup> percentile (1.82 m) are not uncommon, and they are mostly  
585 associated with  $H_s \lesssim 9.5$  m. With increasing significant wave height above 9.5 m, lower rates  
586 of increases in joint probability (larger gaps between isolines) are associated with a lower  
587 variability in water levels (e.g., for a joint density probability of 0.8, storms associated with  
588  $H_s > 9.5$  m would occur above the 99<sup>th</sup> percentile of  $WL$ ). Therefore, different configurations of  
589 events ( $H_s/WL$ ) can have similar non-exceedance probability. For instance, a storm with high  
590 waves combined with low water levels (e.g., storm #11) can have a similar probability of  
591 occurrence than a less energetic storm event with almost twice the water level (e.g., storm #18).  
592 It is interesting to note that the last storm recorded, with  $H_s = 9.2$  m and  $WL = 2.2$  m, has the  
593 highest joint probability ( $p = 0.74$ ).

594 These results show that extreme events can yield the same return period (here shown in  
595 probability space only) while potentially inducing different morphogenic impacts on the coast.  
596 Despite the displacement of a significant volume of boulders during the survey period, the  
597 storms and the wave climate observed during the 2014–2019 period were not extreme compared  
598 to the whole storm history along the southwestern coast of Reykjanes Peninsula since 1948  
599 (Fig. 10b). Storms with energetic waves ( $H_s > 10$  m) lasted up to about a week (e.g., in January  
600 1993 and 2002). Major coastal hazards have been observed under shorter but more intense  
601 storms (storm duration  $< 3$  days and  $H_s > 12$  m) in 1990 and 1967 (Fig. 10b), during which  
602 coastal floods, damaged ships and also moving boulders have been observed (Geirsdóttir et al.,  
603 2014).

604  
605 **Fig. 10.** The Raftery copula (a) shows the bivariate dependence structure of maximum significant wave height  
606 ( $H_s$ ) and highest water levels observed during a single storm in southwest Iceland between 2011 and 2019. The  
607 copula isolines in (a) are plotted over data space, and all points represent observations. In (b), the density plot  
608 shows the duration (hours) and max.  $H_s$  of each detected storms between 1948 and 2019. Two of the most extreme  
609 storms in southwest Iceland during this period (Geirsdóttir et al., 2014) are indicated by dark dots (storms of 1967  
610 and 1990). Green dots and colored diamonds represent the storms detected between 2014 and 2019. Diamonds (a)  
611 and (b) show the storms with the highest  $H_s$  observed in each season during the survey period (2014–2019). The  
612 colorbar represents point density in (b).

#### 614 4.3 Long-term storm events analysis (1948–2019)

615  
616 Between 1948 and 2019, there has been important variability both in terms of storm  
617 occurrence and storm duration offshore Grindavík. No storm occurred in 2007 and 2009, while  
618 the highest number of storms occurred in 1993 with 11 storms cumulating over  $\sim 22$  days,  
619 followed by 1962 (9 storms,  $\sim 19$  days) and 2015 (8 storms,  $\sim 13$  days) (Fig. 11). Linear  
620 regression performed on both the total number of storms and cumulative days of storms  
621 annually do not show any trend ( $R^2 < 0.02$ ) (Fig. 11a and 11b). However, the storm behavior  
622 exhibits a strong periodicity, as shown by the 3–year averages in Fig. 11a and 11b. Years with  
623 3 storms or less occurred about 48% of the time but some periods with more, longer storms  
624 annually can be identified, such as between 1987 and 1995. Significant wave heights ( $H_s$ )



625 during storms also exhibits no particular trend (Fig. 11c), and similarly to the storm behavior,  
626 rather appears to have undergone periods of lows and highs. Overall, Fig. 11 clearly shows that  
627 the monitoring carried out in this study (2014-2019) were not done during the most energetic  
628 period, but rather in a period of diminishing storm occurrence and duration.

629  
630 **Fig. 11.** Storm characteristics between 1948 and 2019 offshore Grindavík. The annual total number of storms (a)  
631 and duration (in days) (b) are shown with a 3-points moving average (solid lines). The linear regressions are  
632 plotted as dashed lines in panels (a) to (c). Wave characteristics during storms are shown in (c): mean  $H_s$ , mean  
633 peak  $H_s$  and maximum  $H_s$  during storms over a year are shown by khaki, green, and blue solid lines, respectively  
634 (circles represent the values in 2008).

635  
636 The wavelet spectral analysis performed on the total storm duration (cumulative days  
637 over a year), mean storm duration and 99.8<sup>th</sup> percentile of  $H_s$  allows a more in-depth  
638 understanding of the storm variability, periodicity and transient dynamics over time. The  
639 wavelet power spectrums (left panels on Fig. 12a to 12c) show the frequency domain ( $y$  axis in  
640 period cycles) and the time dimension with areas of higher (pink) and lower (blue) power. The  
641 average spectral power exhibits a strong peak between 2 and 5 years for all three variables.  
642 This indicates that storm climate alternates between periods characterized by longer storms  
643 associated with more extreme wave heights, and somewhat calm periods with shorter storms  
644 on a timescale of 2 to 5 years. The alternating presence of warmer areas in the wavelet power  
645 spectrum (left panels on Fig. 12a to 12c) roughly between 1950 and 1965, 1975 and 1990, and  
646 2005 and 2015 (>95% significant) also suggests that this periodicity is not stationary, which  
647 implies that the storm climate also alternates between periods of stability and periods of strong  
648 variability on a timescale of about 10 to 15 years.

649  
650 **Figure 12.** Wavelet power spectrums (scalograms on the left) and averaged wavelet power over each frequency  
651 (right column), for the total number of days of storms (a), the mean storm duration annually (b) and the extreme  
652 wave height annually (99.8<sup>th</sup> percentile  $H_s$ ) (c). The color bars on the right of the scalograms indicate power  
653 intensity. Black contours on the scalograms indicate the 5% and 10% significance levels, and the black vertical  
654 curves indicate the border of the cone of influence.

## 655 656 **5. Discussion**

### 657 658 *5.1 Temporal frequency of cliff-top boulders movements in SW Iceland*

659  
660 The first finding of this 6-years monitoring is that cliff-top boulders were annually and  
661 substantially reworked. In comparison, between 2012 and 2017, pluriannual cliff-top boulder  
662 activation in Brittany was very limited, with no boulder displacements for certain years (Autret  
663 et al., 2018). In SW Iceland since 2014, hydrodynamic processes inducing wave overtopping  
664 the cliffs occurred each year on every study sites. These processes are usually described to be  
665 related to extreme storm wave events (Bressan et al., 2018; Hansom et al., 2008; Kennedy et  
666 al., 2017; Watanabe et al., 2019), generally combined with high spring tide in a macrotidal  
667 environment (Fichaut and Suanez, 2011; Autret et al., 2016a, Cox et al., 2018a). In the case of  
668 SW Iceland, such extreme meteo-oceanic conditions are not required to generate boulder  
669 displacements at the top of the cliffs. Indeed, the long-term analysis of morphogenic events  
670 combining extreme storm waves and high tide levels, showed that the 2014-2019 period was  
671 rather less impacted compared to other years, such as winter 1990 and 1967. Only one event  
672 that perfectly paired extreme waves and high spring tide levels was observed in November 5,  
673 2017 (storm #18). In details, our results highlight a significant interannual variability in boulder  
674 movement is observed. While in 2015-2016, 2016-2017, and 2017-2018, there is a good  
675 correspondence between storm frequency/duration (Fig. 9b), and the global volume of boulder  
676 displacements at each site (Fig. 6a and 6b), this is not the case for the winter of 2018-2019.

677 During this period, there was only one storm inventoried in February with high spring tide (Fig.  
678 10a), while the volumes of displaced boulders were larger over the entire survey period (Fig.  
679 6). In this case, it is not the frequency of winter storms that seems to have been the determining  
680 parameter for the movement of the boulders, but the intensity of an extreme event combining  
681 energetic waves and very high spring tide levels. Therefore, the interannual variability in  
682 boulder activation, or the correspondence between the morphogenic events frequency/intensity  
683 (i.e., occurrence of joint events combining storm wave and high spring tide level) and the  
684 volume of displaced boulders, would need more detailed analysis of nearshore and coastal  
685 hydrodynamic processes. As it stands, this monitoring based on annual survey does not allow  
686 us to answer this question, which also refers to the question of storm wave thresholds (Almeida  
687 et al., 2012; Grieco and DeGaetano, 2019), the efficiency of processes combining storm waves  
688 and a spring tide (Harley, 2017), or the impact of repeated low-energy storms vs. a single  
689 extreme storm event (Ferreira, 2005). As seen on Fig. 3c, cliff overtopping can occur with 4  
690 m-high waves. Increasing the monitoring frequency of CBDs to a monthly basis, or even at the  
691 event scale, would help to better understand the various hydrodynamic conditions that generate  
692 boulder movement.

693 In addition, this survey shows that the interannual displacements of boulders at the four  
694 monitored sites show the same morphological signature both in terms of the total volume of  
695 boulders displaced, as well as in terms of the volume as a percentage of the global volume of  
696 the ridges (Fig. 6a and 6b). We can also note that globally, the mass of the transported boulders,  
697 as well as the altitude and the inland distance of these displacements, is proportionally the same  
698 for the four sites from year to year of monitoring (despite the lack of data for the Kerling site  
699 in 2018-2019) (Fig. 6c to 6e).

700 Globally, these results confirm the extremely stormy and morphogenic context of the  
701 meteo-oceanic forcing that impacts SW Iceland at these sub-polar latitudes. This is consistent  
702 with the fact that this coast faces the most energetic wave climate in the North Atlantic Basin  
703 (Betts et al. 2004; Ardhuin et al., 2019). Indeed, based on atmospheric condition analysis  
704 between 1931 and 1991, Schinke (1993) showed this zone of the north Atlantic basin (up to  
705 60°-65° latitude) was 5 times more impacted by deep cyclones ( $\leq 990$  hPa) than temperate  
706 latitudes (45°-50° latitude).

## 707 708 *5.2 Specificities of the boulder displacements in Reykjanes Peninsula*

709  
710 A prominent characteristic of the cliff-top boulder deposits in the four study sites is their  
711 abundance and disposition of swash-aligned elongated ridges in a context of backstopped  
712 platforms (Autret et al., 2016b; Etienne and Paris, 2010). As indicated by many authors, this  
713 type of morphological setting concerns imbricated boulders subject to frequent activation  
714 (Chen et al., 2011; Cox et al., 2012; Etienne and Paris, 2010; Lorang, 2011) which is in  
715 agreement with the results obtained by our surveys. We also indicated that the cliff-top boulder  
716 deposits of the Reykjanes Peninsula experience significant changes over small areas (few  
717 square meters) spread over the seaward side of the ridges without apparent pattern of mobility.  
718 Some coarse boulders (>10 tons and up to 29 tons during our monitoring) are moved while  
719 adjacent finer boulders (<1 ton) are unaffected.

720 The main dominant boulder transport direction (>50% of boulders displaced whatever  
721 the site and whatever the year) (Fig. 6) is oriented inland. This is in accordance with the fact  
722 that inland-directed transport is generally the most common for boulder displacements due to  
723 the cross-shore dynamic of storm wave facing cliffs. A significant part of the cliff-top boulders  
724 is transported toward the sea (22% to 27% of the transported boulders, depending on the site)  
725 and may be attributed to backwash processes. As shown by Cox et al. (2019), even if inland  
726 transport is always dominant, seaward boulder transport by backwash also occurred on the

727 context of backstopped platforms where boulder accumulations take the form of strongly fixed  
728 imbricated linear ridges. In that context, the energetic bore overtopping the cliff is reflected by  
729 the boulder ridge.

730 Longshore boulder transports are also observed on our studied sites as shown in [Fig. 6](#)  
731 (between 10% and 30% of the displacements depending on the site and the year). This is mainly  
732 due to the oblique direction of the swell at break-up in relation to the orientation of the cliff  
733 edge characterized by a certain angle to the coastline. However, these movements are facilitated  
734 by the litho-morphostructural context of the wave-scour cliff-top platforms which are particu-  
735 larly tabular ([Autret et al., 2016b](#); [Etienne and Paris, 2010](#)). This morphology is explained by  
736 the surface characteristics of pāhoehoe lava flows whose high fluidity produces smooth and  
737 flat surfaces as the lava solidifies ([Fig. 2](#)). In contrast, on the very chaotic and rough granite  
738 wave-scour cliff-top platforms of the old folded and fractured Armorican Hercynian massif  
739 (e.g., Banneg Island in Brittany), the longitudinal transport of boulders is made very difficult  
740 by the presence of multiple trapping microforms (i.e., steps, gullies, diaclasses, hollow faults,  
741 etc.) ([Fichaut and Suanez, 2011](#); [Autret et al., 2016a](#)).

742

### 743 *5.3 Boulder activation as a proxy of long-term morphogenetic/catastrophic event analysis*

744

745 The cliff-top boulders, through the processes of their activation/displacement, appears  
746 to be an excellent proxy for the study of extreme events in terms of frequency and intensity.  
747 Systematic monitoring is crucial to study their morphodynamics. In addition to the proxy as  
748 geological archive they represent, it is also essential to better consider their movements and  
749 geomorphic interactions in the global high-energy rocky coast system. As shown for other  
750 coastal boulder deposits, repeated boulder activation is an important factor in platforms erosion  
751 ([Cullen et al., 2018](#); [Cullen and Bourke, 2018](#); [Etienne, 2007](#); [McKenna, 1990](#); [Swirad et al.,](#)  
752 [2019](#); [Naylor and Stephenson, 2010](#); [Naylor et al., 2016](#)). From a geomorphological  
753 perspective, further theoretical and field-based developments coupled to continuous boulder  
754 transport and hydrodynamic observations in their natural contexts are needed to help modelers  
755 to adapt existing morphodynamic models –or develop new ones– to cliff-top boulders transport  
756 and related morphological changes. From a coastal risk perspective, they are often used as a  
757 marker of inundation extents. But when activated, they also are a risk factor on its own. For  
758 instance, coastal hazards in southwest Iceland have clearly been associated to concomitant high  
759 waves and high water levels, and the timing of the peak storm (in terms of wave energy), spring  
760 tide and surge is considered as an important risk to coastal population ([Geirsdóttir et al., 2014](#)).  
761 Such extreme event can activate boulder movement or man-made CBDs-like structure (such as  
762 breakwaters) and turn them into projectiles against infrastructure. For instance, the 1990 flood  
763 event in Stokkseryri (67 km east of Grindavík), which occurred at the highest tide level, is  
764 deeply rooted in the collective memory, especially because of the large rocks that were plucked  
765 from the breakwater and spread around the village. A similar event happened during storm  
766 Dennis (February 11 to 18, 2020) on the coast of Reykjanesbær, 19 km north of our study area  
767 ([Fig. 13](#)).

768

769 **Figure 13.** Boulder displacements during the storm event Dennis (14-15 February, 2020) in Reykjanesbær.

770

## 771 **6. Conclusion**

772

773 The southwest coast of Iceland, especially the Reykjanes Peninsula characterized by basaltic  
774 rocky-cliffs, is subject to an intense annual storm activity because it is located in the main track  
775 of the rail of winter atmospheric low-pressure systems that circulate in the Northern Atlantic  
776 basin. The monitoring that has been undertaken since 2014 shows that the CBDs accumulated

777 at the top of the cliffs are an excellent proxy to analyze the variability of this North-Atlantic  
778 cyclogenesis. However, the annual frequency of the survey does not allow us to study  
779 morphogenic processes on an episode scale; it nevertheless shows a good correspondence  
780 between the volumes of CBDs displaced and the frequency and/or intensity of storms over the  
781 entire winter period. In the longer term, this monitoring will also allow to evaluate the impact  
782 of storms on the coastal morphological dynamics of rocky coasts in a context of very high  
783 meteo-oceanic energy. In a context of climate change and the implications that this has on the  
784 modification of hydrodynamic conditions, particularly at lower temperate latitudes, this work  
785 is of great interest.

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