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# Characterization of bedrock mass-wasting at fault-bound abyssal hills

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## Highlights

1. Bedrock mass wasting on abyssal hills shapes scarp morphology from fault inception
2. Earthquakes are a key driver of bedrock mass wasting on abyssal hills
3. Ratio of mass removal efficiency to uplift rate is similar across spreading rates
4. Global earthquake-driven mass flux for abyssal hills between 24–1428 M m<sup>3</sup> yr<sup>-1</sup>

## Abstract

Fault-bound abyssal hills form at mid-ocean ridges and cover ~65% of Earth's surface, but few studies have characterized the extent to which bedrock erosion controls their morphology. Here, we use bathymetry data to characterize the morphology of fault-bound abyssal hills on a global scale, and employ numerical modelling and seismicity catalogues to quantify how simultaneous rock uplift and bedrock erosion sculpt deep-ocean landscapes. By generating a global database on abyssal hill morphology, we show that most large abyssal-hill scarps (>100 m in height) within the near-axis zone of seismicity (i.e., <30 km from axis) have slopes between 10 and 30°, well below the expected range of underlying normal fault dips of 45–60°. We interpret this as a manifestation of efficient bedrock mass wasting on near-axis growing faults, a process that operates from fault inception. Using a non-linear topographic diffusion model to parameterise the effects of erosion, we find a balance between erosion and rock uplift that is similar for slow, intermediate, and fast spreading rates. We express the ratio of erosion to uplift as an inverse Peclet number that ranges between 0.06 and 0.82 for abyssal hills. We also calculate a global bedrock diffusivity for abyssal hills in the range 0.01–1.51 m<sup>2</sup> yr<sup>-1</sup>. These results imply that bedrock erosion is a significant process that sculpts abyssal hill morphology and reshapes the oceanic crust. Overall, this study provides a framework to incorporate bedrock mass wasting into future models of ocean-floor evolution and, more generally, to active extensional settings on Earth and beyond.

## 37 1. Introduction

38 Abyssal hills cover approximately 65% of the Earth's seafloor. These elongated, regularly-spaced  
39 topographic highs form at mid-ocean ridges (MORs, Figure 1a) where normal faults uplift and tilt  
40 freshly-accreted basaltic seafloor (e.g., Macdonald et al., 1996). While the mechanisms responsible for  
41 fault formation and tectonic uplift at MORs are reasonably well understood (Goff, 1991; Buck et al.,  
42 2005; Olive et al., 2015), relatively little is known about the erosive processes sculpting these submarine  
43 landscapes. It is often assumed that erosion only has a second-order impact on the morphology of  
44 uplifting abyssal hills, and that bathymetry faithfully records tectono-magmatic processes (e.g., Cowie  
45 et al., 1993; Bohnenstiehl & Carbotte, 2001; Paulatto et al., 2015; Goff, 2020). This assumption is  
46 challenged by high-resolution seafloor surveys that reveal widespread rockslides on the active fault  
47 scarps that delineate MOR axes (Allerton et al., 1995; Goff and Tucholke, 1997; Searle et al., 1998;  
48 Mitchell et al., 2000; Cannat et al., 2013). These high-resolution studies are conducted primarily at local  
49 scales, near MOR axes, and we lack a global assessment of bedrock mass wasting. Researchers have  
50 previously attempted to quantify erosion and transport of pelagic sediments on the abyssal plain (Webb  
51 & Jordan, 1993; 2001; Mitchell, 1995; 1996). More recent endeavors have started to quantify active  
52 bedrock erosion on the ocean-floor more generally (Hilley et al., 2020; Hughes et al., 2021; 2023).  
53 Despite these efforts, a quantitative framework to describe the evolution of tectonically-active bedrock  
54 landscapes at the scale of the ocean floor has proven difficult to establish because the distribution,  
55 frequency, and magnitude of gravitational mass wasting remain largely unknown.

56 On land, landslides are recognized as key drivers of landscape evolution (e.g., Densmore et al.,  
57 1997; Roering et al., 1999). The difficulty in subaerial studies lies in teasing out the influence of rainfall,  
58 soil strength, topographic stresses, vegetation cover, and seismic activity on the time-averaged erosion  
59 rates that result from repeated landslides (e.g., Rodríguez et al., 1999; Malamud et al., 2004; Hovius et  
60 al., 2011). Farther afield on the icy moons of Jupiter and Saturn, recent work identified ground  
61 accelerations from moonquakes as the principal driver of mass wasting along tectonics ridges (Mills et  
62 al., 2023). Back on Earth, fault scarps and fault-bound abyssal hills that formed along MORs (herein  
63 collectively referred to as abyssal hills) experience no precipitation-driven or vegetation-modulated  
64 erosion processes. The absence of these key variables, combined with rapid seismogenic uplift of  
65 lithologically-uniform basaltic bedrock make abyssal hills an ideal endmember setting to explore how  
66 mass wasting shapes topography on Earth and beyond.

67 Earthquakes can be important drivers of submarine erosion (Hughes et al., 2021; 2023).  
68 Observations from bathymetry, side-scan sonar, and seismicity data show that abyssal hill faults grow  
69 within a 2 to 30 km wide window from the ridge axis (Alexander & Macdonald, 1996; McAllister &  
70 Cann, 1996; Searle et al., 1998; Bohnenstiehl & Carbotte, 2001; Smith et al., 2003). In areas with  
71 pronounced ridge shoulders, abyssal hills may even experience a reduction in amplitude between ~10  
72 and ~30 km off-axis, due to reverse slip along hill-bounding faults associated with reverse earthquakes  
73 (Deffeyse, 1970; Olive et al., 2024). The presence of thin fingers of fresh bedrock talus at the base of  
74 growing abyssal hills has been taken to indicate active faulting under the assumption that mass wasting

75 is mostly triggered by earthquakes (Figure 2a; Searle et al., 1998; Escartín et al., 1999). As fault throw  
76 evolves, mass wasting erodes the fault scarp and buries its base before eventual blanketing by pelagic  
77 sedimentation (Figure 2a–b; Allerton et al., 1995; Goff & Tucholke, 1997). Prior studies have used  
78 seafloor slopes to calculate diffusivity values that describe the redistribution of pelagic sediments away  
79 from the ridge axis at rates of  $0.007\text{--}10\text{ m}^2\text{ yr}^{-1}$  (Webb & Jordan, 1993; 2001; Mitchell, 1995; 1996).  
80 However, without information on the spatial and temporal evolution of abyssal hill bedrock erosion, we  
81 have insufficient data to assess the impact of mass wasting on the morphology of growing abyssal hills.  
82 Ultimately, such estimates are required to use bathymetry as a reliable recorder of tectonic and  
83 magmatic processes at MORs (Cowie et al., 1993; Bohnenstiehl & Carbotte, 2001; Paulatto et al., 2015).

84 Abyssal hill morphology varies with spreading rate and embeds, to first order, the modulation  
85 of fault activity by MOR magmatism (Figure 1c; Goff, 1991; Buck et al., 2005; Roth et al., 2019;  
86 Tucholke et al., 2023). At fast-spreading ridges ( $>75\text{ mm yr}^{-1}$  full spreading rate), magmatic  
87 emplacement in the brittle lithosphere accounts for greater plate separation than at slow spreading ridges  
88 ( $<55\text{ mm yr}^{-1}$ ). Faults at fast ridges rapidly migrate off-axis where they are abandoned (Buck et al.,  
89 2005; Behn and Ito, 2008; Olive et al., 2015). As a result, abyssal hills at fast ridges have less relief  
90 ( $\leq 100\text{ m}$ ) and are more closely spaced than those at slow ridges, where they grow taller ( $\sim 1000\text{ m}$ ) and,  
91 eventually, more widely-spaced (e.g., Goff, 1991; Goff et al., 1997; Roth et al., 2019; Goff, 2020). As  
92 a corollary, faults at fast and slow spreading ridges should potentially slip at similar rates (e.g., a slip  
93 rate that is 10% of a fast spreading rate like  $10\text{ cm/yr}$  is the same as 50% of a slow rate like  $2\text{ cm/yr}$ ).  
94 The seismogenic behavior of MOR faults also appears modulated by spreading rate, with a lesser  
95 seismic moment release rate measured at faster ridges (Cowie et al., 1993; Frohlich and Wetzel, 2007;  
96 Mark et al., 2018). To untangle the respective roles that the total duration of fault slip and the associated  
97 seismic moment release play in shaping abyssal hill morphology through gravitational mass wasting,  
98 we assembled an extensive dataset of fault scarp parameters, across a wide range of MOR spreading  
99 rates. We focused on the entire erosional history of abyssal hills starting at the ridge axis (Figure 1a,  
100 2c) and measured the height and slope of global abyssal hill scarps to characterize spatiotemporal  
101 variations in morphology. We then used our global database to constrain a numerical model for abyssal  
102 hill erosion using non-linear topographic diffusion laws, as commonly applied to subaerial hillslopes  
103 (e.g., Roering et al., 1999) and to the redistribution of pelagic sediment on abyssal hills (e.g., Webb &  
104 Jordan, 2001). Outputs from the models were used to evaluate the ratio of rock uplift to erosion for  
105 abyssal hills globally.

## 106 2. Geomorphic Analysis

### 107 2.1 Quantification of scarp morphology

108 We compiled scarp parameters from abyssal hills within 2500 km of the global mid-ocean ridge  
109 system to sample a broad range of erosional and tectonic histories across all spreading rates. We focused  
110 on areas with well-lineated abyssal terrain, and avoided transects across transform valleys. We also  
111 avoided complex tectonic areas such as those with prominent detachment faulting linked to reduced

112 melt supply (e.g., Escartín et al., 2008; Smith et al., 2014) or non-transform offsets. We extracted 124  
113 spreading-parallel transects (Figure 1a) in areas with high-resolution, publicly available, shipboard  
114 multibeam bathymetry data (Global Multi-Resolution Topography Data Synthesis:  
115 <https://www.gmrt.org/index.php>; Ryan et al., 2009). The transect length (horizontal distance between  
116 first and last scarp crests), is limited by bathymetric data availability at each location and ranges from  
117 45 to 490 km. We measured scarp morphology from digital elevation models (Figure 1b–f) and  
118 resampled all data at the highest common resolution of 100 m to avoid resolution-dependent slope  
119 artifacts.

120 We measured the height, slope, and distance to the ridge axis for 5803 abyssal hill faults from the  
121 124 transects following the method of Howell et al., (2016) illustrated in Figure 1, and described in the  
122 Supplementary Materials. We also recorded the crustal age and spreading rate for each fault, based on  
123 global grids of age and spreading rate (Seton et al., 2020). For our analysis, we binned the data based  
124 on spreading rate as slow ( $<55$  mm yr<sup>-1</sup>), intermediate (55–75 mm yr<sup>-1</sup>), and fast ( $\geq 75$  mm yr<sup>-1</sup>), with  
125 3136, 1127, and 1540 hills in each bin, respectively. We combined ultra-slow with slow ridges and  
126 ultra-fast with fast ridges (e.g., Dick et al., 2003) to maximize the data in each bin.

## 127 **2.2 Slope measurements in shipboard bathymetry**

128 Due to the limited resolution of shipboard multibeam bathymetry (typically 50–100 m resolution),  
129 small scarps ( $<100$  m height) are best analyzed using near-bottom bathymetry data acquired with  
130 underwater vehicles at horizontal resolutions of a few meters (Le Saout et al., 2018; Hughes et al.,  
131 2021). However, these datasets are scarce at MORs and focus primarily on the axial area (Cannat et al.,  
132 2013; Smith et al., 2014; Le Saout et al., 2018). Studies of tall abyssal hill scarps ( $>100$  m) at meter-  
133 scale resolution are restricted to very few sites (Cannat et al., 2013). Therefore, variations in abyssal  
134 hill morphology can only be studied globally using shipboard multibeam bathymetry (Figure 1).

135 For small faults from 2D shipboard bathymetric profiles, slope can be underestimated due to the  
136 data resolution (Figure 1g). Bathymetry gridded with a resolution of 100 m pixel<sup>-1</sup> cannot resolve slopes  
137  $>45^\circ$  for scarps shorter than 100 m. Even for scarps  $>100$  m tall, shipboard bathymetry can  
138 underestimate the true local slope for stepped scarps on a scale not resolvable at that resolution (e.g.,  
139 several adjacent faults; Figure 1g). When scarps grow taller than  $\sim 100$ –200 m, the slope measured from  
140 shipboard bathymetry approaches that measured in near-bottom, high-resolution bathymetry (Figure  
141 1h). To minimize spurious slopes due to data resolution (see 4.1), we focused our analysis on faults  
142  $>100$  m in height (3516 out of 5804 faults) and on time-averaged mass wasting (as opposed to individual  
143 rocksliding events) on the largest abyssal hills. The full database of global abyssal hill morphology is  
144 included in the supplementary materials.

## 145 **2.3 Trends in abyssal hill morphology**

146 From the database, we documented an overall decrease in scarp height with increasing spreading  
147 rate, consistent with independent studies (Figure 2c, 3a; e.g., Goff, 1991; Goff et al., 1997; Roth et al.,  
148 2019; Goff, 2020). This agreement confirms that our scarp picks constitute a representative abyssal hill

149 population. Crucially, almost all measured slopes (3506 out of 3516) are below the 45–60° range of  
150 fault dips suggested for mid-ocean ridge faults (Figure 3b; Thatcher and Hill, 1995).

151 To characterize scarp height and slope evolution with increasing age and distance from the ridge  
152 axis (Figures 2b, 3c–f, S5–S7), we used the 95<sup>th</sup> percentile. Because the data show significant and  
153 expected scatter, trends in the median value are subdued and the 95<sup>th</sup> percentile best records key overall  
154 trends in fault morphological evolution. There is a rapid increase in the height and slope of the largest  
155 scarps within 10 km of the ridge axis, with maximum values of 1750 m and 55° by ~7 km and ~6 km  
156 from the axis, respectively (Figures 3c–d, 4b–c). The 95<sup>th</sup> percentiles of height and slope decrease  
157 rapidly from a peak of 700 m and 37° at 5–10 km from the axis, before levelling off between 300–400  
158 m and 20–25° at distances beyond 500 km (Figure 3c–d). The abrupt end to scarp height increase at ~10  
159 km off-axis is consistent with the unfauling process described by Olive et al. (2024). Farther off-axis  
160 over increasingly old seafloor, there is a subtle but gradual decrease in the 95<sup>th</sup> percentile of scarp height  
161 from 700 m at 20 Ma and 37° at 5 Ma down to ~500 m at ~60 Ma and ~30° at 40 Ma (Figure 3e). This  
162 slow decay, over hundreds of km, is likely related to pelagic sedimentation filling topographic lows. In  
163 fact, typical sedimentation rates are consistent with the observed rate of apparent relief reduction in old  
164 seafloor (Figure 3e). Plots of morphological data for slow, intermediate, and fast spreading rates are  
165 included in the supplementary materials (Figures S5–S7).

#### 2.4 Erosion during scarp growth

166 The geomorphological data documents the simultaneous growth and erosion of abyssal hills by  
167 fault slip and bedrock mass wasting. Abyssal hills grow within at least 10 km of the ridge axis, and  
168 possibly as far as 30 km (Figure 4b, Table S1; Alexander & Macdonald, 1996; McAllister & Cann,  
169 1996; Searle et al., 1998; Bohnenstiehl & Carbotte, 2001; Smith et al., 2003). We herein refer to the  
170 area of abyssal-hill growth within ~30 km of the ridge axis as the “fault growth window”, but fault  
171 activity within the ridge shoulder may combine reverse and normal slip (Deffeyse, 1970; Olive et al.  
172 2024). In this window, many abyssal hills exhibit slopes below 30° and almost all abyssal hills >100 m  
173 in height have slopes well below the assumed dip of underlying faults (45–60°; Figure 4c; Thatcher and  
174 Hill, 1995). We suggest this discrepancy indicates erosion via bedrock mass wasting during scarp  
175 growth. Slope degradation results from the combined removal of material from the scarp face and talus  
176 deposition at its base (Figure 2). In some cases, small, closely spaced scarps may be identified as a  
177 single unit in lower resolution data (Figure 2g). Removing scarps <100 m in height partially mitigates  
178 their inclusion in our analysis.

179 The evolution of scarp slope as a function of scarp height thus captures the early tectonically driven  
180 scarp growth and decay with the increasing importance of mass wasting as a control on scarp  
181 morphology (Figure 5). We analyzed the smoothed median of abyssal hills within the fault growth  
182 window assuming that the upper and lower limits of the data envelope record minimum and maximum  
183 amounts of scarp degradation, respectively (Figure 5a). Overall, the smoothed median of slope for  
184 abyssal hill increases non-linearly with height (Figure 5a). The upper envelope of slope vs. scarp-height  
185 (upper black dashed line in Figure 5a) reaches a maximum value of ~55° for ~300 m high hills before  
186

187 gradually decreasing to  $\sim 45^\circ$  for the tallest hills. Conversely, the lower envelope increases gradually to  
 188 a maximum slope of  $\sim 40^\circ$  for the tallest scarps ( $\sim 1750$  m high, Figure 5a). A boundary for the theoretical  
 189 maximum slope resolvable at a given horizontal resolution can be obtained using  
 190  $\arctan(\text{height}/\text{resolution})$  (black long-dashed line in Figure 5a for a horizontal resolution of 100 m).  
 191 Notably, the upper envelope lies beneath the resolution envelope for scarps below  $\sim 300$  m in height.  
 192 This data gap could suggest efficient erosion lowering the scarp slope away from the fault plane dip  
 193 angle during growth. Alternatively, this gap could be an artifact of the coarse resolution of the data.

194 The evolution of slope against height can be framed as a hysteresis loop. The raising limb is defined  
 195 by younger abyssal hills ( $< 20$  Ma), which have median slopes  $3\text{--}5^\circ$  higher than abyssal hills  $> 20$  Ma  
 196 (Figure 5b). Abyssal hills  $> 20$  Ma define the falling limb where the 95<sup>th</sup> percentile of slope for a given  
 197 height is between  $5\text{--}10^\circ$  lower than abyssal hills  $< 20$  Ma (Figure 5b). We interpret lowered slopes for  
 198 older abyssal hills to represent the gentle smoothing of bathymetry by pelagic sediment deposition and  
 199 redistribution (Figure 2b; e.g., Webb & Jordan, 2001; Tominaga et al., 2011).

### 200 3. Quantifying oceanic basement erosion

201 We designed a numerical model to approximate how mass wasting reworks abyssal hills and infer  
 202 global rates of seafloor bedrock mass wasting from the abyssal hill dataset.

#### 203 3.1 Scarp evolution as a diffusive process

204 We simulated 1-D (across-strike) scarp morphology based on realistic erosion parameters for  
 205 comparison with the quantitative data from abyssal hills presented in section 2. At the 100 m resolution  
 206 used here, one cell may encompass a wide range of erosional and depositional processes. The  
 207 bathymetric data therefore reflects scarp evolution with mass transfer from steep erosive to gentle  
 208 depositional sections of the scarp. We chose to parameterize mass wasting as a non-linear diffusive  
 209 process, with local downslope mass flux increasing non-linearly with local topographic gradient and  
 210 increasing rapidly for slopes beyond a critical failure angle ( $\theta_c$ ). This framework is commonly applied  
 211 to model subaerial hillslopes (e.g., Roering et al., 1999), and similar diffusion equations have been used  
 212 to study redistribution of pelagic sediment on abyssal hills (Webb & Jordan, 1993, 2001; Mitchell,  
 213 1995; 1996; Tominaga et al., 2011). To convey the efficiency of scarp diffusion as a competition  
 214 between erosion and rock uplift, we non-dimensionalized our model inputs as an inverse Peclet number  
 215 ( $Pe^{-1}$ ).  $Pe^{-1}$  is the ratio of diffusion ( $K$ ) relative to advection (in this case fault throw rate) and is reflected  
 216 in the evolution of abyssal hill slope with increasing height (Figure 5a). This approach also allowed us  
 217 to quantify  $K$  for a given rate of rock uplift at abyssal hills.

218 Our simplified approach includes a number of assumptions. We assumed that in situ bedrock is  
 219 pervasively fractured as often observed in submersible dives from axial valley floors (e.g., Wright et  
 220 al., 1995). We accounted for variation in the degree of disaggregation by running models using a critical  
 221 angle of  $35^\circ$  and  $40^\circ$  (Table S2). The model also does not account for other processes that control  
 222 effective strength, like cohesion or compressive strength. Additionally, sediment transport distance is a  
 223 product of momentum as well as slope, but the model does not incorporate varying transport distance

224 resulting from different failure methods (e.g., rockfall, debris flow etc.). In the model, deposited volume  
 225 roughly equals sediment flux (volume conservation), so we assume that porosity and, therefore, volume  
 226 of intact but pervasively fractured bedrock is similar to that of bedrock talus. The non-uniform  
 227 distribution of talus on hills of similar heights and at similar distance from the ridge axis (Figure 2;  
 228 Allerton et al., 1995) suggests some spatial variability in effective rock strength and, therefore, critical  
 229 failure angle. Conversely, the relatively flat geometry of the median slope for a given height in abyssal  
 230 hills >20 Ma (Figure 5b) supports a threshold behavior inherent in our non-linear diffusion model.  
 231 Nevertheless, our approach provides a first-order approximation of the time-averaged effect of bedrock  
 232 mass wasting along abyssal hills. Alternative methods, like advection based on progressive rather than  
 233 critical failure, or a lattice grain model for hillslope evolution (Tucker et al., 2018), may also be  
 234 appropriate but were not considered further.

### 235 3.2 Model setup

236 In the model, 1-D seafloor topography along a spreading-parallel direction,  $h(x,t)$ , follows the  
 237 following equation:

$$238 \quad \frac{Dh}{Dt} = \frac{\partial}{\partial x} \left( K(\theta) \frac{\partial h}{\partial x} \right), \quad (1)$$

239 where  $\frac{Dh}{Dt}$  denotes the Lagrangian derivative of topography as it is advected both horizontally and  
 240 vertically by a steady tectonic velocity field ( $v_x(x)$ ,  $v_z(x)$ ). We only considered relief growth on the  
 241 footwall side, i.e.,  $x \geq 0$ , with  $x = 0$  corresponding to the seafloor trace of the normal fault that shapes  
 242 the abyssal hill. For simplicity, we assumed that down-warping of the hanging wall produces a  
 243 topographic low that is instantaneously filled through volcanic extrusion such that  $h(x < 0, t) = 0$ . In  
 244 equation (1),  $K$  is the effective diffusivity of topography which depends on topographic gradient:

$$245 \quad \theta = \tan^{-1} \left( \left| \frac{\partial h}{\partial x} \right| \right), \quad (2)$$

246 according to the following equation:

$$247 \quad K = 10^{(\log_{10} K_{\min} + \log_{10} \left( \frac{K_{\max}}{K_{\min}} \right) (0.5 + \frac{1}{\pi}) \tan^{-1} \left( \frac{\theta - \theta_c}{w} \right))} \quad (3)$$

248 In equation (3),  $K_{\min}$  is the minimum diffusivity at very low gradient and  $K_{\max}$  is a maximum value,  
 249 typically three orders of magnitude greater than  $K_{\min}$ . This approach keeps diffusivity finite for all  
 250 gradients and ensures a sharp increase in diffusivity over a narrow range of gradients around the critical  
 251 angle ( $\theta_c$ ).  $w$  acts as a regularization parameter that controls how fast modeled gradients approach  $\theta_c$   
 252 where a larger  $w$  means a faster increase towards higher diffusivity at gradients approaching  $\theta_c$ . In our  
 253 models, we set  $\theta_c = 35\text{--}40^\circ$  and  $w = 0.035$  and  $0.042$ , respectively.

254 Tectonic uplift and horizontal advection were parameterized in relation to the horizontal slip  
 255 rate,  $U$ , on the hill-bounding fault. Specifically, we used piece-wise linear functions  $v_x(x)$  and  $v_z(x)$  that  
 256 equal zero for all  $x \leq 0$ , and reach maximum values respectively equal to  $U/2$  and  $\tan(\theta_F)U/2$  at  $x =$   
 257  $Ut/2$  ( $\theta_F$  denotes the dip of the fault). Beyond  $x = Ut/2$ ,  $v_x(x)$  is kept constant at  $U/2$ , and  $v_z(x)$  decreases  
 258 linearly to 0 until  $x$  reaches  $Ut/2 + S$ , where  $S$  represents the width of the gently sloping side of the  
 259 abyssal hill (non-axis-facing). We set  $S = 10$  km in our calculations, but note that  $S$  has virtually no



260 effect on the morphology of the steep side of the hill (the fault scarp), which was our primary interest.  
 261 This definition of  $v_x(x)$  and  $v_z(x)$  captures the uplift, advection, and tilting of abyssal hill topography as  
 262 the normal fault slips without the complexities related to footwall flexure (e.g., Weissel and Karner,  
 263 1989). We also assumed a fault dip of  $\theta_F = 45^\circ$  (after Thatcher and Hill, 1995), for simplicity:  $U$  can  
 264 thus be viewed as a proxy for full heave and throw rate on the fault. Because the model does not account  
 265 for the subsiding motion of the hanging wall, we assumed the representative rock uplift rate in the hill  
 266 is  $U/2$ , which amounts to assuming symmetric uplift and subsidence in the footwall and hanging wall  
 267 of a normal fault. To estimate  $U$ , we measured the fraction of horizontal extension taken up by faults  
 268 for each transect. This fraction,  $F$ , is classically estimated as the slope of a cumulative fault heave vs.  
 269 distance to the axis plot (e.g., Searle and Laughton, 1977). Multiplying  $F$  by the full spreading rate  
 270 along the transect yields a representative horizontal fault slip rate or uplift rate (adopting a fault dip of  
 271  $45^\circ$ ), assuming only one fault is active at any given time (e.g., Buck et al., 2005).

272 Equation (1) was discretized with a backward-Euler finite difference scheme and solved with  
 273 Newton iterations to handle the non-linearity of the diffusion law. Advection of topography within the  
 274 ( $v_x(x)$ ,  $v_z(x)$ ) field is achieved through a particle-in-cell method. We solved Equation 1 starting from a  
 275 flat topography to calculate how predefined values of diffusivity ( $K$ ) and  $\theta_c$  degrade the interpolated  
 276 scarp face during fault growth.

### 277 3.3 Model results and scarp diffusivity

278 Our numerical model simulates the evolution of a scarp subjected to diffusive erosion and  
 279 deposition. Examples are shown in Supplementary Figure 6e–h. Early in the simulations, topographic  
 280 scarps with very small throws are almost instantaneously leveled by diffusion. As a result, they initially  
 281 develop very gentle slopes even though their underlying fault is steep. When tectonic throw exceeds a  
 282 characteristic advection-diffusion length ( $K / U \sim 10$  m), diffusion no longer keeps up with uplift, and  
 283 scarp slope increases. This, in turn, increases diffusivity non-linearly, and caps slope at values near the  
 284 critical angle. Overall, our model reproduces observed trends of increasing scarp slope vs. height which  
 285 eventually plateau near  $30\text{--}40^\circ$  (Figure 6a).

286 We compared our simulation results (using the maximum slope measured along modeled scarps) to  
 287 the moving median of measured slopes with increasing height for scarps within the  $\sim 30$  km-wide fault  
 288 growth window (Figure 6a). Cropping the dataset at  $<30$  km from the axis also partly mitigates any  
 289 reduction in height or lowering of slopes by pelagic sediment deposition (Figure 5b). Goodness of fit  
 290 was assessed through a root-mean-squared (RMS) residual relative to the smoothed median from the  
 291 slope/height plots (thick red line in Figure 6a–d). The lower and upper bounds were calculated using  
 292 the RMS residuals relative to the smoothed 5<sup>th</sup> and 95<sup>th</sup> percentiles, respectively (thin red lines in Figure  
 293 6a–d). RMS residuals for each model run are shown in Figure S4.

294 Each model run was characterized by an inverse Peclet number  $Pe^{-1}$  defined as:

$$295 Pe^{-1} = \frac{K}{lU}, \quad (4)$$

296 where  $l$  is a characteristic median scarp height for faults  $>100$  m in height ( $l = 199$  m). This is akin  
 297 to dimensionless "uplift-erosion numbers" used in a variety of studies that couple tectonics and surface

298 processes (e.g., Olive et al., 2022; Wolf et al., 2022): a large  $Pe^{-l}$  means erosion dominates uplift and  
 299 relief is subdued. To estimate the  $Pe^{-l}$  of natural systems, we ran 40 simulations with a broad range of  
 300  $K$  (0.0003–3 m<sup>2</sup> yr<sup>-1</sup>, all other parameters kept constant) and report the  $Pe^{-l}$  value of the run yielding the  
 301 lowest RMS misfit of median slope vs. height trends. We note that changing  $U$ ,  $w$ ,  $\theta_c$ , and the subsurface  
 302 fault dip also affects the final values of  $Pe^{-l}$  (Figures S2, S3). Our chosen parameters are justified in the  
 303 supplementary materials.

304 Models indicate that scarp diffusion is relatively inefficient at leveling tectonically induced seafloor  
 305 uplift, but still represents a significant process that modifies the axis-facing slopes of abyssal hills during  
 306 fault growth.  $Pe^{-l}$  is 0.06–0.82 for faults >100 m in height within 30 km of the ridge axis (range is  
 307 defined by the 95<sup>th</sup> and 5<sup>th</sup> percentile of  $Pe^{-l}$ , respectively), with  $U = 2.57$  mm yr<sup>-1</sup> (average  $U$  for all  
 308 faults >100 m in height) and  $\theta_c = 40^\circ$  (Figure 6a; Table S2). When the same data is binned by spreading  
 309 rate,  $Pe^{-l}$  was 0.06–0.55 for slow (Figure 6b), 0.10–0.62 for intermediate (Figure 6c), and 0.14–0.82 for  
 310 fast spreading rates (Figure 6d), respectively. Hence, we did not resolve a systematic variation of  $Pe^{-l}$   
 311 with spreading rate. Results using  $\theta_c = 35^\circ$  broadly agree with results using  $\theta_c = 40^\circ$  and are presented  
 312 in the supplementary materials (Table S2).

## 313 4. Discussion

### 314 4.1 The balance between erosion and uplift

315 There are possible differences in the tectono-magmatic conditions or seismogenic behavior of  
 316 abyssal hill-bounding normal faults at different spreading rates (e.g., Frohlich & Wetzell, 2007).  
 317 Nevertheless, to the first order,  $Pe^{-l}$  values indicate a similar ratio of erosion to uplift for fast,  
 318 intermediate, and slow spreading rates (Figure 7a). This implies that, based on bathymetry derived from  
 319 shipboard multibeam systems, potential variations in tectono-magmatic processes between spreading  
 320 rates do not perturb the ratio of erosion to uplift. The results facilitate quantification of bedrock erosion  
 321 rates for abyssal hills in terms of diffusivity,  $K$ , by rearranging equation 4:

$$322 \quad K = \frac{lU}{Pe} \quad (5).$$

323 For  $U = 2.57$  mm yr<sup>-1</sup>,  $l = 199$  m (average height for all faults >100 m in height), and the overall  
 324  $Pe^{-l}$  values of 0.06–0.82, we obtained a  $K$  of 0.03–0.42 m<sup>2</sup> yr<sup>-1</sup> (Table S2). In our model,  $K$  scales  
 325 linearly with  $U$  (Equation 5) so that individual  $K$  values only apply to a specific fault slip rate.  
 326 Accordingly, we can use the full range of  $U$  from the transects measured here (0.72–9.25 mm yr<sup>-1</sup>;  
 327 Figure 8a) to scale the smallest and largest  $K$  values derived from the median  $U$ . The resulting overall  
 328  $K$  using the full range of  $U$  is 0.01–1.51 m<sup>2</sup> yr<sup>-1</sup>. This range is comparable to gravity driven failure and  
 329 turbidity flows of pelagic sediment from the MAR, which range from 0.04–1.00 m<sup>2</sup> yr<sup>-1</sup> (Figure 7b;  
 330 Mitchell, 1995; Webb and Jordan, 2001).

331 The values of  $U$  for each transect do not support a systematic variation in fault slip rate, and  
 332 therefore  $K$ , for different MOR spreading rates (Figure 8a).  $U$  ranges from 0.72–9.25 mm yr<sup>-1</sup> with a  
 333 large spread encompassing most of this range for all spreading rates (Figure 8a). Previous estimates of  
 334 growth rates for abyssal hill faults based on the height and spacing of faults are 1–2 mm yr<sup>-1</sup> from the

335 slow spreading MAR (McAlister and Cann, 1996) and 6–8 mm yr<sup>-1</sup> from the EPR (Cowie et al., 1993).  
 336 Our dataset shows a data gap for slow spreading faults with high fault growth rates (Figure 8a). This  
 337 gap could be filled by accretion dominated by detachment faulting, which we excluded from our  
 338 analysis. In such terrains, detachments may accommodate most of the spreading resulting in elevated  
 339 uplift rates like the >7 mm yr<sup>-1</sup> suggested at the Atlantis Massif on the slow-spreading Mid-Atlantic  
 340 ridge (Figure 8a; Escartin et al., 2022). We note that  $U$  values calculated here and fault growth rates  
 341 from the literature do not account for processes such as partial or complete scarp burial by sedimentation  
 342 or by volcanic deposits (e.g., Webb and Jordan, 2001; Escartín et al., 2007). Moreover, these growth  
 343 rates are a high end-member because we assume only one fault is active at a time (Buck et al., 2005).  
 344 The fault growth window is on the order of 10 to a few 10s of km, so several faults may be active  
 345 simultaneously for fault spacings <10 km. Hence, we cannot argue for systematic variation in uplift rate  
 346 with spreading rates.

347 Our analysis has certain limitations. We note that the morphological evolution documented in  
 348 Figure 3 is heavily weighted towards slow spreading ridges, because scarps >100 m high are most  
 349 abundant at slow rates and scarce at fast rates (n=2258, 584, and 674 for slow, intermediate, and fast,  
 350 respectively). Where possible, our analysis focused on areas with higher data density. For example, we  
 351 only compared abyssal hills <1100 m tall at slow spreading rates with our model results to avoid the  
 352 sparse data coverage above 1100 m height (Figure 6b). Due to the low scarp heights at fast spreading  
 353 ridges (n = 180 >100 m in height and <30 km from the axis), the trend of the smoothed median line in  
 354 Figure 6d is poorly constrained beyond ~300 m in height. Therefore, the calibration of  $Pe^{-1}$  against data  
 355 is limited at fast-spreading sites and the range of  $Pe^{-1}$  of 0.14–0.82 for fast faults is less robust than for  
 356 faults developing at slower-spreading ridges (Figure 6b–c).

## 357 **4.2 Processes driving mid-ocean ridge mass wasting**

### 358 **4.2.1 Earthquake controls on erosion during scarp growth**

359 Like terrestrial settings, if earthquakes are an important driver of abyssal hill bedrock erosion, then  
 360 spatial variation in earthquake ground accelerations should be an important control on patterns in mass  
 361 wasting (e.g., Croissant et al., 2019; Medwedeff et al., 2020). In simplest terms, ground acceleration  
 362 depends on earthquake magnitude, depth, hypocentral distance, and the shear wave velocity of shallow  
 363 bedrock (e.g., Abrahamson and Silva, 1997). In theory, differences in seismic behavior at different  
 364 spreading rates should modify erosion parameters. To understand mass wasting behavior at different  
 365 spreading rates requires an examination of how seismic parameters vary for different spreading rates.

366 Seismic moment release and coupling are both thought to decrease with increasing spreading rate  
 367 (Cowie et al., 1993; Frohlich & Wetzell, 2007; Mark et al., 2018). However, the normalization of the  
 368 total seismic energy by distance along the ridge axis varies greatly at slow spreading ridges, covering  
 369 the overall trend across spreading rates (Figure 8b). A similar trend is observed for the depth of micro-  
 370 seismicity, although these experiments have limited spatial and temporal extents (Figure 8c). Shallow  
 371 micro-seismicity is observed at all ridges, but deep micro-seismicity (e.g., >5–6 km) is only recorded  
 372 at slow spreading rates (Figure 8c). Consequently, we speculate that while seismicity may be less

373 frequent with slightly lower energy release at fast spreading ridges, the seismicity may be concentrated  
374 at shallow depths and result in similar ground shaking compared to slow spreading ridges. Similar  
375 average ground shaking for different spreading rates, despite possible variations in seismic parameters,  
376 could be one factor that accounts for the similarity in  $Pe^{-1}$  across spreading rates.

377 Seismicity at MORs occurs mostly within a few tens of kilometers of the ridge axis. The overall  
378 frequency of earthquakes measured with hydrophone arrays from the Mid-Atlantic Ridge (MAR) and  
379 the Southwest Indian Ridge (SWIR) is greatest within 10–20 kilometers of the ridge axis and gradually  
380 decreases away from the axis with infrequent earthquakes occurring up to 200 km from the axis (Figure  
381 4a; Smith et al., 2003; Tsang-hin-sun et al., 2016). These seismicity observations are consistent with  
382 previous estimations for the width of the fault growth window (Table S1). Previous estimates are  
383 derived from the distribution of seismicity or indirectly from the evolution of fault parameters (spacing,  
384 height) and the distance from the axis at which these stabilize (Table S1). A 20–30 km limit for the fault  
385 growth window is broadly consistent with the pronounced increase in height of the largest scarps within  
386 10 km of the ridge axis (Figure 4b). Moreover, only ~15 % of all abyssal hills in our dataset are within  
387 40 km of a transform fault, so transform seismicity is not considered to be a major contributor to abyssal-  
388 hill slope degradation here.

389 The documentation of fresh bedrock talus at the ridge axis in sidescan sonar data (Figure 2a) and  
390 near-bottom bathymetry surveys (Figure S1) supports scarp average slopes being degraded via bedrock  
391 mass wasting within the fault growth window. Shipboard bathymetry and AUV sidescan sonar data  
392 from the Southeast Indian Ocean collected as part of the search for Malaysian Airlines flight MH370  
393 (AusSeabed Marine Data Portal at: <https://portal.ga.gov.au/persona/marine>) constitute a substantial  
394 dataset to evaluate bedrock mass wasting outside the fault growth window. From data between 350–  
395 2500 km from the axis, we identified possible thin fingers of fresh bedrock talus at the foot of abyssal  
396 hill-bounding faults in just two out of 66 abyssal hills examined from a study area of ~98,000 km<sup>2</sup>  
397 (Figure 9; methods in the supplementary materials).

398 The few observations of potential fresh talus at 350–2500 km from the axis indicates that bedrock  
399 mass wasting is extremely rare far off axis. Other seafloor geomorphic studies also report little evidence  
400 for fresh talus >30 km off axis at fast and slow spreading rates (e.g., Macdonald and Luyendyk, 1985;  
401 Escartin et al., 1999). The sidescan sonar data from the Southeast Indian Ocean records widespread low  
402 reflectivity seafloor, which indicates widespread sediment cover (Figure 9). Limited bedrock talus and  
403 widespread sediment deposition are consistent with the levelling off in the 95<sup>th</sup> percentile of height and  
404 slope beyond ~500 km off-axis (Figure 3). Seismicity is also limited away from the ridge axis. Only  
405 two earthquakes were recorded in the deep-tow sonar survey area for the Southeast Indian Ocean  
406 between 1950–2021 (ANSS ComCat catalog, <https://earthquake.usgs.gov/data/comcat/>; Figure 9a).  
407 These combined off-axis morphological and seismological observations indicate that bedrock mass  
408 wasting is not significant beyond ~30 km off axis and highlight the importance of coseismic shaking as  
409 a trigger for mass wasting near-axis.

#### 410 4.2.2 Alteration processes

411 Alteration may also facilitate mass wasting during scarp growth. Bedrock alteration (e.g.,  
 412 serpentinization, clay formation) and possibly fluid flow (changes in pore fluid pressure), can reduce  
 413 the internal friction and shear strength of the bedrock and, theoretically, promote erosion (Tucholke et  
 414 al., 1997). Bedrock alteration is thought to occur primarily at and in the vicinity of the ridge axis during  
 415 exhumation (e.g., Cannat et al., 2010; Merdith et al., 2020). The degree of off-axis, shallow, bedrock  
 416 alteration depends on sediment cover. Fluid ingress and egress, and accompanying bedrock alteration  
 417 may preferentially occur at un-sedimented bathymetric highs with seafloor bedrock exposures (Figure  
 418 9; Coogan and Gillis, 2018). If alteration gradually weakens bedrock and promotes mass wasting, then  
 419 an increase in the frequency of mass wasting with distance from the axis would be expected along with  
 420 the corresponding evolution in abyssal hill morphology. The limited bedrock mass wasting documented  
 421 outside of the fault growth window (Figure 9), coupled with a leveling off in scarp height and slope  
 422 >500 km off axis (Figure 3), indicate that any potential long-term bedrock alteration does not noticeably  
 423 enhance mass wasting.

424 Conversely, alteration processes like mineral precipitation, may stabilize bedrock and reduce the  
 425 potential for mass wasting. Palagonitization, for example, can increase the density and decrease the  
 426 porosity and permeability of basalt (e.g., Weaver et al., 2020). We speculate that crack closure or surface  
 427 precipitation may locally increase the internal friction of oceanic bedrock with a stabilizing effect acting  
 428 against any potential alteration weakening. This is supported by subsurface geophysical observations.  
 429 A rapid increase in seismic velocity at the top of layer 2 from 3.0 km/s at 0 Ma to 4.6 km/s at 10.5 Ma  
 430 followed by a more gradual increase to 5.0 km/s at 170 Ma has been attributed to precipitation of  
 431 hydrothermal alteration products within cracks (Christenson et al., 2019). Consequently, any potential  
 432 weakening from alteration may be partially offset by strengthening from mineralization. Either way,  
 433 based on the observations above, we suggest the primary trigger for bedrock mass wasting is coseismic  
 434 shaking during scarp growth near-axis.

#### 435 4.3 Global mass flux on abyssal hill

436 To assess the global significance of abyssal hill mass wasting in term of detrital sediment fluxes,  
 437 we quantified the erosional mass flux of bedrock associated with the growth of the largest abyssal hills.  
 438 We calculated the average annual mass flux for an abyssal hill ( $Q_{ave}$ ) using the average area eroded and  
 439 corresponding growth time for a representative large abyssal hill (i.e., 200 m tall; Figure S8). Next, we  
 440 multiplied  $Q_{ave}$  by the total length of the global MOR system of 60,000–70,000 km assuming only one  
 441 abyssal hill is active per cross-axis transect. For a typical hill spacing of 3 km (from our global dataset)  
 442 and a cross-axis width of 60 km for the fault growth window, we calculate that 20 abyssal hills are  
 443 active in one cross-axis transect. Therefore, by multiplying the number of active abyssal hills by  $Q_{ave}$   
 444 and considering all uncertainties, we calculate an annual earthquake-induced abyssal hill mass flux of  
 445 24–1428 million  $m^3 yr^{-1}$ . We verified our calculation based on model outputs with an alternative  
 446 approach using simple trigonometry (full details on all calculations in the supplementary materials).  
 447 This value records the combined effects of near-axis fault growth and erosion, with the associated

448 uncertainties, and excludes other mass wasting environments (e.g., transform- and non-transform  
449 scarps, detachment-dominated seafloor, flanks of seamounts and volcanic islands, etc.).

450 For comparison, the total global mass flux from earthquake-induced terrestrial landslides is  
451 estimated at 16.8–22.4 billion  $\text{m}^3 \text{yr}^{-1}$  (Broeckx et al., 2020). While up to two orders of magnitude less  
452 than subaerial sites, the total earthquake-induced, abyssal-hill, mass wasting is still a significant surface  
453 process shaping the oceanic crust, especially given the process is concentrated along a narrow strip of  
454 seafloor. Mass wasting denuding the oceanic lithosphere may impact alteration and modulate chemical  
455 fluxes between the lithosphere and the ocean (e.g., Merdith et al., 2020). Further investigations into  
456 submarine mass wasting are required to study its impact on weathering fluxes (Isson et al., 2019), on  
457 seafloor ecosystems (Stefanoudis et al., 2016), and on the safety of submarine infrastructure and  
458 resource extraction (Smith et al., 2018). Finally, complete quantification of ocean seafloor mass wasting  
459 should include other sites. Transform faults have cumulative lengths on the order of a third of the global  
460 MOR system, and are typically more seismically-active than their adjacent ridge segments. Transforms  
461 may thus increase the global mass flux of bedrock erosion by an amount commensurable to ridges.  
462 Intraplate volcanoes and seamounts are also ubiquitous and show pervasive, albeit unquantified, flank  
463 erosion (e.g., Mitchell et al., 2003).

#### 464 **4.4 Growth and decay of abyssal hills**

465 Based on the morphological and seismological data presented here, we posit that the overall  
466 morphology of abyssal hills is established by simultaneous tectonic rock uplift and coseismic bedrock  
467 mass wasting during fault growth (Figure 10). A rapid increase in height for young scarps within the  
468 fault growth window is accompanied by a simultaneous reduction in slope below the assumed fault dip  
469 of 45–60° (Figure 4c). As faults are rafted off-axis, the draping and redistribution of pelagic sediments  
470 by turbidity currents and sediment mass wasting smooths topography (Webb & Jordan, 2001; Tominaga  
471 et al., 2011). This is reflected in gentler slopes for a given height for abyssal hills >20 Ma compared to  
472 abyssal hills <20 Ma (Figure 5b). Sediment accumulation at the foot of scarps can potentially explain  
473 the gradual decrease in the 95<sup>th</sup> percentile for scarp height and slope away from the ridge axis, although  
474 the data is somewhat noisy (Figure 3c–d). Isolated earthquakes and, possibly, isolated bedrock mass  
475 wasting, may occasionally occur outside of the fault growth window (e.g., Bergman & Solomon, 1984;  
476 Figure 4a, 10). However, the overall decreasing trend in the 95<sup>th</sup> percentile of height and slope off axis  
477 indicates that these isolated events do not promote scarp growth, at least at the resolution of our data  
478 (Figure 5b, 10).

### 479 **5. Conclusion**

480 Our global database on abyssal hill morphology supports a model in which coseismic shaking is  
481 the primary trigger for mass wasting events that reshape abyssal hills as they grow at or near the ridge  
482 axis. For the tallest abyssal hills (>100 m), height increases rapidly, but scarp slopes are generally  
483 degraded to <30° within 10 km of the ridge axis compared to an original underlying fault dip of 45–  
484 60°. The rapid decrease in scarp slope during scarp growth records the cumulative effect of discrete

485 coseismic mass wasting events. This result demonstrates that abyssal hill scarp degradation is active  
 486 from fault inception at the ridge axis. Beyond 20–30 km from the ridge axis, seismogenic fault growth  
 487 ceases and a gradual decrease in both the height and slope of the tallest scarps is attributed to the gradual  
 488 smoothing of topography via the blanketing and redistribution of pelagic sedimentation with minimal  
 489 bedrock erosion.

490 We used numerical modelling and a quantitative database of abyssal hill fault scarp morphology  
 491 from across the global mid-ocean ridge (MOR) system to investigate how cumulative coseismic erosion  
 492 controls the morphology of abyssal hills. The results show that erosion is relatively inefficient at  
 493 leveling the hills, but still profoundly alters the slopes of abyssal hill scarps. The balance between  
 494 erosion and uplift is similar for slow, intermediate, and fast spreading rates, as expressed with the uplift-  
 495 erosion number  $Pe^{-1}$  ranging between 0.06 and 0.82 with no discernable trend across spreading rates.  
 496 Our results also provide the first constraint on bedrock scarp diffusivity with rates ranging from 0.01–  
 497 1.51 m<sup>2</sup> yr<sup>-1</sup>. This study demonstrates that mass wasting during abyssal hill growth is a significant  
 498 surface process operating along the global MOR system and provides a framework to incorporate  
 499 bedrock mass wasting into future models of landscape evolution in actively uplifting extensional  
 500 settings.

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 503 Topography Data Synthesis [GMRT] available at: <https://www.gmrt.org/index.php> (Ryan et al., 2009).  
 504 Near-bottom bathymetry used in Figures 1 and S1 were downloaded from the Marine Geosciences Data  
 505 System available at: <https://www.marine-geo.org/index.php>. Shipboard bathymetry and sidescan sonar  
 506 data used for Figure 9 were collected as part of the search for Malaysian Airlines flight MH370 and are  
 507 publicly available from the AusSeabed Marine Data Portal at: <https://portal.ga.gov.au/persona/marine>.  
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47 732  
 48 733 **Figure 1.** Summary of datasets and methods. a) Global spreading rate map [Seaton et al., 2020] and  
 49 734 location of the transects measured in this study. Plate boundaries are from Bird et al. [2003]. The green  
 50 735 star locates example profile in Figure 1e. Thin black lines are non-ridge plate boundaries. b) Shaded  
 51 736 relief bathymetry, c) slope, and d) curvature maps used to identify faults at the Southwest Indian Ridge.  
 52 737 UTM = Universal Transverse Mercator. Maps in UTM zone 34S. e) Example profile extracted from  
 53 738 black line in b–d) showing picks for fault scarp crests (blue lines) and bases (red lines) used to constrain

739 scarp height. f) Schematic diagram showing how we measured height and slope. Upper: slope profile  
 740 (blue dots); lower: bathymetry (blue line). Here, slope is the average form a slope profile section (from  
 741 the red dots) along the steepest scarp section (red line). g) Comparison of scarp morphology between  
 742 shipboard bathymetry (90 m resolution) and near-bottom bathymetry (1 m resolution) data showing the  
 743 potential underestimation of slope by shipboard data. h) Comparison of scarp morphology between  
 744 shipboard bathymetry and near-bottom bathymetry data showing general agreement of slopes for scarps  
 745 >200 m height. See Figure S1 for line and location of profiles in parts f–h.

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 747 **Figure 2.** a) Deep-towed sonar image from Searle et al. [1998] showing recent active bedrock mass  
 748 wasting at 29° N on the Mid-Atlantic ridge. Strong acoustic backscatter is white and indicates bedrock  
 749 or bedrock talus while weak acoustic reflections are dark gray and represent sediment cover. Black  
 750 areas are acoustic shadows. In the interpretation sketch, bedrock or fresh bedrock talus on the scarp face  
 751 is highlighted in white and sediment in grey. Location is shown in Figure 1a. b) Schematic diagram  
 752 showing scarp evolution with increasing distance from the ridge axis after Allerton et al. [1995]. Rates  
 753 for sediment diffusivity are from Webb and Jordan [2001]. Range of fault growth rates are from  
 754 McAllister and Cann [1996] and Cowie et al. [1993]. c) Density plot of height as a function of spreading  
 755 rate for scarps analyzed here. Density is the number of hills within each bin of width  $\sim 3.33 \text{ mm yr}^{-1}$  and  
 756 height  $\sim 35 \text{ m}$ . Average values for each transect (white circles) are for comparison with the root-mean  
 757 squared heights of abyssal hills from Roth et al. [2019] (purple diamonds).

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 759 **Figure 3.** Density plots showing the morphological data for abyssal hills >100 m in height. Red lines  
 760 are smoothed 95<sup>th</sup> percentiles and the dashed lines are areas with no data. a–b) Height and slope as a  
 761 function of increasing full spreading rate. The bin width for spreading rate is  $\sim 3.3 \text{ mm/yr}$ , height is  
 762 binned in  $\sim 41 \text{ m}$  increments, and slope in  $\sim 1.3^\circ$  increments. The 95<sup>th</sup> percentile is smoothed using a  
 763 window of  $\pm 5 \text{ mm/yr}$ . c–d) Height and slope as a function of distance from the ridge axis. The bin  
 764 width for distance is  $\sim 50 \text{ km}$ , height is binned in  $\sim 36 \text{ m}$  increments, and slope in  $\sim 1.0^\circ$  increments. The  
 765 95<sup>th</sup> percentile is smoothed using a window of  $\pm 25 \text{ km}$ . e–f) Height and slope as a function of age. The  
 766 bin width for age is 2 Ma, height is binned in  $\sim 52 \text{ m}$  increments, and slope in  $\sim 1.6^\circ$  increments. The  
 767 95<sup>th</sup> percentile is smoothed using a window of  $\pm 10 \text{ Ma}$ . The sedimentation rate in e) shows the  
 768 reduction in fault scarp height (initially 700m), assuming infill of adjacent topographic lows.  
 769 Sedimentation rates are based on a value of  $6.5 \pm 2.3 \text{ m/Myr}$  (Webb and Jordan, 2001).

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 771 **Figure 4.** a) Decreasing earthquake frequency with increasing distance from the ridge axis in 10 km  
 772 bins recorded with a hydrophone array from the Mid-Atlantic ridge for the period 1999–2003  
 773 (Bohnenstiehl et al., 2002; Smith et al., 2002, 2003) and the Southwest Indian Ridge for the period  
 774 2012–2013 (Tsang-Hin-Sun et al., 2016). Both datasets have a magnitude range between  $\sim 205$ – $245 \text{ dB}$   
 775 ( $\sim M_w 2.5$ – $6.5$ ) and include some transform seismicity. b) Density plot for abyssal hills >100 m in height  
 776 showing a rapid increase in scarp height within 10 km of the ridge axis for the largest scarps defined by

777 the 95<sup>th</sup> percentile. Sedimentation rates are based on 6.5 +/- 2.3 m/Myr (Webb and Jordan, 2001). c)  
 778 Density plot showing evolution of slope with distance to the ridge axis for abyssal hills > 100 m in  
 779 height. Measured slopes are well below the assumed minimum fault dip of 45° for all but 10 data points  
 780 and the moving median value of slope is generally 20–30°. The red box denotes the 20–30 km outer  
 781 edge of the fault growth window. The median and 95th percentiles (red lines) in parts b and c were  
 782 smoothed using a moving average filter with a smoothing window of 50 km.

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 784 **Figure 5.** a) Density plot of slope as a function of height for global abyssal hills taller than 100 m and  
 785 within 30 km of ridge axis overlain with interpretation. Maximum and minimum erosion envelopes are  
 786 hand drawn. The thick red line is the smoothed moving median from 10 m bins of height and the thinner  
 787 red lines are the 5<sup>th</sup> and 95<sup>th</sup> percentiles. b) Comparison of height and slope for abyssal hills younger  
 788 and older than 20 Ma. The >20 Ma bracket includes hills up to 67 Ma. The yellow (<= 20 Ma) and blue  
 789 (> 20 Ma) shaded areas show the range of the data for each age bracket defined by the 5<sup>th</sup> (lower) and  
 790 95<sup>th</sup> (upper) percentiles. The inset shows schematically how abyssal hill morphology evolves similar to  
 791 a hysteresis loop due to a temporal evolution in erosional and depositional process.

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 793 **Figure 6.** a–d) Model results plotted on scarp data for all hills (a), slow (b), intermediate (c), and fast  
 794 (d) spreading rates. Scarp data comprises abyssal hills <30 km from the ridge axis. Red lines represent  
 795 the best-fit model runs relative to the median (middle line), 5th (lower line), and 95th (upper line)  
 796 percentile of the scarp data (black lines). Goodness of fit is based on a root-mean-squared residual  
 797 relative to the smoothed percentiles from the scarp data. To avoid comparing model results with areas  
 798 with very few hills, the upper height limit of the red domain is capped at 1100 m for all hills and slow  
 799 spreading rates, 600 m for intermediate spreading rates, and 400 m for fast ridges. Model parameters  
 800 are fault dip = 45°,  $w = 0.042$ , and critical angle ( $\theta_c$ ) = 40°. e–f) Time steps in the modelled evolution  
 801 of an uplifting abyssal hill for hill height (e) and slope (f) using the same input parameters as parts a–d.  
 802 g–h) Morphology of a modeled abyssal hill at ~1100 m height (i.e., limit of red lines in part a) for hill  
 803 height (g) and slope (h) using different diffusivity. Models use the same input parameters as in a–d) and  
 804 a horizontal slip rate of 2.57 mm yr<sup>-1</sup>.

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 807 **Figure 7.** a) Plot of inverse Peclet number ( $Pe^{-1}$ ) and diffusivity ( $K$ ) with increasing spreading rate. The  
 808 colored boxes represent the spreading rates bounds (slow = <55 mm yr<sup>-1</sup>, intermediate = 55–75 mm yr<sup>-1</sup>  
 809 <sup>1</sup>, and fast = >75 mm yr<sup>-1</sup>) and the 5th and 95th percentiles of  $Pe^{-1}$  or  $K$  for a given uplift or horizontal  
 810 slip rate ( $U$ ) (assuming a fault dip of 45°). The boxes assume  $U = 2.57$  mm yr<sup>-1</sup> based on the median  $U$   
 811 for faults >100 m in height. b) Comparison of bedrock  $K$  calculated in this study with  $K$  for erosion of  
 812 pelagic sediment over different length scales (amended from Webb and Jordan [2001]). The black boxes  
 813 represent ranges of uncertainty whereas stars have no uncertainty. The red box is bounded by the

814 minimum and maximum  $K$  values (using  $U = 0.72$  &  $9.25$  mm yr<sup>-1</sup>), and we define the length scale  
815 using fault heave with a minimum of 100 m and a maximum of 1750 m.

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1 817 **Figure 8.** a) Minimum fault growth rates versus spreading rate for every transect measured here. The  
2 grey boxes are ranges of previously published estimates for fault growth rates based on the references  
3 shown here. b) Seismic moment release rates from Frohlich and Wetzler [2007] normalized by ridge  
4 axis length. Data provided in Table S3. c) Depth range of microseismicity across spreading rates.  
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9 821 References and data for part c are included in Table S4.

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12 823 **Figure 9.** Side scan sonar (5 m horizontal resolution) and bathymetry data (40–60 m horizontal  
13 resolution) from the Southeastern Indian Ocean. a) Map locating the data relative to the Southeast Indian  
14 Ridge axis (area of (a) is shown in Figure 1a). Earthquakes are from the USGS Advanced National  
15 Seismic System Comprehensive Earthquake Catalog (USGS, 2017) for earthquakes  $> M_b$  2.5 from 1950  
16 to the present. b–d) AUV side scan sonar data (top) above corresponding shaded relief, ship-board,  
17 bathymetry (bottom) covering the same area. Roman numerals represent features of interest described  
18 here. In the sonar data sedimented seafloor is darker brown or orange whereas exposed bedrock is pale  
19 brown or yellow. b) Representative abyssal hill. i: smooth, linear contact between sediment and bedrock  
20 indicates no recent mass wasting. c) Isolated abyssal hill with possible recent bedrock mass wasting. ii:  
21 Possible fingers of fresh talus. iii: Sedimented talus. d) A relay zone. iv: Possible recent bedrock mass  
22 wasting which could also be sedimented talus. v: Chutes or fingers of talus on the relay scarp. e)  
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24 830 Bathymetric profiles across X-X' and Y-Y' from b) and c). Pale grey domain sketches extent of pelagic  
25 sediment and dark grey is schematic talus. Bathymetry and sonar data are from Geoscience Australia.  
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29 833 Sonar maps show artifacts indicated by green arrows and that generally run WSW-ENE (nadir of sonar  
30 swath, differences in absolute acoustic backscatter across adjacent tracks, mosaicking seams).  
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39 839 **Figure 10.** Summary of the morphological evolution of mid-ocean ridge fault scarps and abyssal hills  
40 with increasing distance from the ridge axis. The geomorphic data and model results presented in this  
41 study allow us to add quantitative bedrock erosion parameters to previously existing tectonic models of  
42 abyssal hill evolution. All values are from this study except the diffusivity value for pelagic sediment,  
43 which is from Webb and Jordan [2001]. The block models are amended from Allerton et al. [1995].  
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48 844 Faults in red are potentially active and faults in black are inactive.

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