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1	Controls on the magmatic fraction of extension at mid-ocean ridges
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10	
11	Keywords
12	Mid-ocean ridge; seafloor spreading; normal faults; diking; submarine volcanism.
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15	
16	Highlights
17	• Magmatic component of seafloor spreading (<i>M</i>) increases non-linearly with spreading rate.
18	• Stress interactions between fault slip events and magma intrusions provide physical basis for
19	М.
20	• Axial lithosphere thickness and rate of pressure build-up in sills are primary controls on <i>M</i> .
21	• Transitions between modes of seafloor spreading reflect thresholds in axial lithosphere
22	thickness.

23 Abstract

24 The fraction M of plate separation accommodated by magma emplacement at mid-ocean ridges has been recognized as the main control on seafloor spreading modes, yet the factors that control 25 26 *M* itself are poorly understood. Here we put forward a simple theoretical framework explaining 27 M in terms of short-term cycles of earthquakes and dike intrusions interacting with one another 28 by modulating the stress state of the ridge axis. Axial lithospheric thickness and the rate of 29 pressure build-up in shallow, replenishing magma sills are the two main parameters controlling 30 M in our simulations. Combined with plausible scenarios for the increase of pressure build-up 31 rate and the decrease of lithospheric thickness with increasing spreading rate, our model 32 appropriately brackets available measurements of M from slow, intermediate, and fast-spreading 33 ridges. Our model further suggests that the transitions between major modes of seafloor 34 spreading (detachment faulting, symmetric faulting, and fully-magmatic) correspond to 35 thresholds in axial lithospheric thickness, and that the great variability in M at slow and ultraslow 36 ridges directly reflects along-axis variability in thermal structure. More generally, this implies 37 that the balance between bottom-up magmatic heating and top-down hydrothermal cooling fully 38 determines the time-averaged rate of magmatic intrusions in the brittle lithosphere, and thus the 39 modes of mid-ocean ridge faulting which shape $\sim 2/3$ of Earth's surface.

40

41 **1. Introduction**

Seafloor spreading at mid-ocean ridges (MORs) typically involves a combination of tectonic and magmatic processes. Adiabatic decompression melting in the asthenosphere produces magma, which ascends to the ridge axis and crystallizes in sills, dikes or lava flows, forming new oceanic crust. Magmatic emplacement is accompanied by normal faulting, which dissects the brittle upper crust into discrete blocks that undergo uplift and tilting as they are advected off-axis
(MacDonald et al., 1996). These processes interact over geological time to shape elongated,
regularly spaced abyssal hills: the most common landform on the surface of the Earth (e.g.,
Kappel and Ryan, 1986).

50 The long-term (>100 kyr) partitioning of tectonic and magmatic extension at MORs is 51 straightforwardly quantified by summing the heaves of normal faults located within a certain 52 distance to the axis (e.g., Escartín et al., 1999; Schouten et al., 2010). The ratio of magmatically accommodated extension to total plate separation —commonly referred to as M (Fig. 1a)— 53 54 exceeds ~0.9 at fast-spreading ridges (full spreading rate $U \ge 8$ cm/yr) such as the East Pacific 55 Rise (Cowie et al., 1993; Buck et al., 2005; Escartín et al., 2007). At intermediate-spreading 56 MORs ($5 \le U < 8$ cm/yr), M typically ranges between ~0.7 and ~0.95 (Ito and Behn, 2008), 57 although values in excess of 0.95 have recently been reported locally along the Chile Ridge and 58 the Southeast Indian Ridge (Howell et al., 2016). Variability in the M fraction is maximized at slow ($1.5 \le U < 5$ cm/yr) and ultraslow (U < 1.5 cm/yr) MORs. These ridges are characterized by 59 60 the along-axis juxtaposition of sections with symmetrically faulted (staircase-like) topography, 61 and sections populated by large-offset detachment faults shaping oceanic core complexes and 62 other irregular bathymetric highs (Escartín et al., 2008; Sauter et al., 2011). M ranges between ~0.6 and ~0.8 at symmetrically faulted sections of the Mid-Atlantic ridge (Cowie et al., 1993; Ito 63 64 and Behn, 2008; Olive and Escartín, 2016). At detachment-bearing sections of the Mid-Atlantic 65 Ridge, M typically varies between ~ 0.2 and ~ 0.6 (Behn and Ito, 2008; MacLeod et al., 2009; 66 Schouten et al., 2010). It can reach zero in amagmatic sections of the ultraslow Southwest Indian 67 Ridge, where the seafloor consists of serpentinized peridotite exhumed by detachment faults (Dick et al., 2003; Sauter et al., 2011; Cannat et al., 2019). 68

69 Mechanical models suggest that M is the primary control on MOR tectonics (Buck et al., 70 2005). This includes the average spacing and maximum offset of normal faults, the presence of 71 an axial valley vs. an axial high, and the development of symmetric faults vs. detachments (Behn 72 and Ito, 2008; Ito and Behn, 2008; Tucholke et al., 2008; Olive et al., 2010; 2015; Tian and Choi, 73 2017; Liu and Buck, 2018; 2020; Howell et al., 2019). The primary effect of the M parameter is 74 to modulate the rate at which an active fault formed in thin axial lithosphere will migrate off-75 axis, encountering progressively thicker and stronger lithosphere. At some point, breaking a new 76 fault in thin axial lithosphere becomes more energetically favorable than continuing to slip on a 77 fault that has moved off-axis. The greater the M, the sooner this point is reached, meaning that 78 faults form in closer succession on either side of the axis, and accumulate less total slip. This 79 leads to regularly spaced, symmetric abyssal hills. When M is close to 0.5, faults can remain at 80 the ridge axis for extended periods of time and accumulate very large offsets, making detachment faulting possible (Buck et al., 2005). Values of M < 0.5 enable faults to migrate 81 82 towards the axis, which in numerical models results in complex inward- and outward-dipping 83 faults that cross-cut each other (Tucholke et al., 2008; Olive et al., 2010; Bickert et al., 2020).

84 A common limitation of all the models listed above is that they treat M as an empirical input 85 parameter that cannot be determined from other observables that characterize MOR axes. 86 Measurements show that M increases non-linearly with spreading rate (Olive et al., 2015) in a manner that resembles the increase in magma flux ($\Phi_C = H_C \times U$, with H_C the thickness of the 87 88 magmatic crust) from ultraslow to fast MOR (e.g., Cannat et al., 2008) (Fig. 1a). Even though 89 the *M* parameter is often used interchangeably with "magma supply", it has never been quantitatively related to $\Phi_{\mathcal{C}}$, let alone spreading rate. It is also important to note that M only 90 91 refers to magma intruded in the brittle lithosphere, and ignores melt emplaced in the ductile

portion of the ridge axis, which Olive et al. (2010) argued does not impact fault evolution. One
might further expect that the strength and thickness of axial lithosphere should affect the modes
of shallow melt emplacement, and therefore the *M* value (e.g., Keller et al., 2013).

95 Here we propose a quantitative framework that identifies first-order controls on M, and 96 explains its non-linear increase with spreading rate (Fig. 1a). Our approach is rooted in the fact 97 that on human time scales, seafloor spreading primarily manifests as normal faulting 98 earthquakes, which can reach magnitudes up to ~6.5 (e.g., Solomon et al., 1988; Cowie et al., 99 1993; Olive and Escartín, 2016), as well as dike intrusion events that are typically ~1-m wide 100 and sometimes lead to a seafloor eruption (e.g., Qin and Buck, 2008; Tan et al., 2016). We 101 design the simplest possible analytical model of tectono-volcanic cycles, in which plate 102 separation on decadal time scales is either taken up by a dike or an increment of fault slip, 103 depending on the ever-evolving thresholds for magmatic intrusion and fault slip. Averaging the 104 model behavior over many cycles yields an effective M value that reflects the influence of 105 spreading rate, lithospheric thickness and melt flux to the ridge axis.

106

107 2. A simple model of tectono-magmatic interactions

108 <u>2.1. Governing equations</u>

109 We consider an idealized mid-ocean ridge axis (**Fig. 1b**) subjected to a full plate separation rate 110 *U*, which deforms over a characteristic cross-axis length scale $L \sim 10$ km. The brittle on-axis 111 lithosphere has a thickness *H* and overlies a sill-like axial melt lens (AML) located immediately 112 below the brittle-ductile transition. Horizontal stresses at the axis (at depth *z* below seafloor) 113 consist of a lithostatic component partially relieved by pore pressure, and a horizontal tectonic 114 component $\Delta \sigma_{xx}$ such that

115
$$\sigma_{xx}(z,t) = -\rho g(1-\lambda)z + \Delta \sigma_{xx}(t), \quad (1)$$

116 where ρ is the density of the crust (2900 kg/m³), and *g* is the acceleration of gravity (9.8 m/s²). λ 117 denotes fluid pressure normalized by lithostatic pressure ρgz and will be assumed hydrostatic, 118 i.e., $\lambda = \rho_w / \rho = 0.34$ (with ρ_w the density of water) except near the roof of the AML where 119 $\lambda = 0$. The horizontal tectonic stress (positive in tension) is assumed to continuously increase 120 over time at a rate prescribed by the far-field plate separation rate:

121
$$\frac{\partial \Delta \sigma_{xx}}{\partial t} = \frac{EU}{L}, \qquad (2)$$

where *E* is the Young's modulus of axial lithosphere. A plausible range for *E* is 5–30 GPa (Heap et al., 2020). In the following we will use E = 10 GPa unless otherwise specified, and a Poisson's ratio $\nu = 0.25$. $\Delta \sigma_{xx}$ however cannot increase to infinity and can reach either a tectonic threshold $\Delta \sigma_{xx}^{F}$, which triggers a seismic (or aseismic) slip event on a fault, or a magmatic threshold $\Delta \sigma_{xx}^{M}$, which triggers the intrusion of a vertical dike.

127 The tectonic threshold is determined by assuming the ridge axis is bounded by an 128 optimally-oriented, cohesionless normal fault of friction coefficient $\mu = 0.6$. The static strength 129 of the fault is thus reached when

130
$$\Delta \sigma_{xx}(t=t_F) = \Delta \sigma_{xx}^F = \frac{\mu \rho g (1-\lambda)H}{\mu + \sqrt{1+\mu^2}}.$$
 (3)

Equation (3) corresponds to the strength of the fault averaged over the thickness of the brittle lithosphere (e.g., Behn and Ito, 2008). For simplicity, we assume that a tectonic event occurring at time t_F corresponds to an increment in fault slip δ_F in the horizontal direction (e.g., $\delta_F = 1$ m for a magnitude ~6 earthquake with a standard stress drop of ~1 MPa). Such events induce compression at the axis, which instantaneously relieves a small portion of the built-up horizontal 136 tectonic stress. We model this effect by subtracting a stress increment from $\Delta \sigma_{xx}$ after each 137 event:

138
$$\Delta \sigma_{xx}(t = t_F^+) = \Delta \sigma_{xx}(t = t_F^-) - E \frac{\delta_F}{L}.$$
 (4)

139 At certain times, the total horizontal stress at the base of the brittle lithosphere may reach the rocks' tensile strength T_0 (~1 MPa), triggering the intrusion of a dike of width $\delta_F = 1$ m 140 141 (Kelemen and Aharonov, 1998; Qin and Buck, 2008). This is because magmatic overpressure p_M 142 in the AML (or a sill-like region within the AML) imparts an increase in total horizontal stress 143 near the edges of the area undergoing replenishment and pressurization (Wilcock et al., 2009). 144 This is illustrated in Supplementary Figure S1. The additional tensile stress above the sill is 145 proportional to magmatic overpressure, and may be exacerbated by stress concentrations around 146 sharp edges or irregularities along the top of the AML. The zero-dimensional nature of our 147 approach prevents us from incorporating these effects in a systematic manner. We therefore 148 assume that the increase in horizontal tension near a pressurized sill is equal to magmatic 149 overpressure and thus fluctuates at a rate imposed by the dynamics of magmatic replenishment. 150 The area above the AML is known to be impermeable to fluid circulation (Gillis 2008). It therefore experiences a confining pressure equal to $p_w + \rho g H$, where p_w is water pressure at the 151 152 seafloor (19.6 MPa for 2000 m depth). With these assumptions, the threshold for diking can be 153 expressed as

$$-p_w - \rho g H + \Delta \sigma_{xx} + p_M = T_0 , \qquad (5)$$

155 which can be recast as:

156
$$\Delta \sigma_{xx}(t = t_M) = \Delta \sigma_{xx}^M(t = t_M) = T_0 - p_M(t = t_M) + p_W + \rho g H.$$
(6)

157 When the threshold described by equation (6) is reached at time t_M , the instantaneous 158 emplacement of a dike imparts an increment of compression on the axis, which we model as:

159
$$\Delta \sigma_{xx}(t = t_M^+) = \Delta \sigma_{xx}(t = t_M^-) - E \frac{\delta_M}{L}.$$
 (7)

Equation (6) shows that the diking threshold decreases through time as magma overpressure builds up. We assume that this occurs at a constant rate \dot{p}_M , which depends on the volume flux of magma that continuously replenishes the AML. As the intrusion of a dike temporarily connects the AML to the seafloor, it likely lowers magma pressure to a static state, meaning that magmatic overpressure drops to zero following a diking event:

165
$$p_M(t = t_M^+) = 0.$$
 (8)

166 When this occurs, the magma intrusion threshold is maximized:

167
$$\Delta \sigma_{xx}^{M,MAX} = T_0 + p_w + \rho g H . \tag{9}$$

Finally, we assume that slip events on ridge-bounding faults, which relax the built-up axial stress $\Delta \sigma_{xx}$, do not otherwise influence the ever-changing threshold for dike initiation $\Delta \sigma_{xx}^{M}(t)$.

171 <u>2.2. Choice of parameter values</u>

172 <u>2.2.1. Axial lithospheric thickness</u>

173 The thickness of axial lithosphere is a critical parameter in our model. It corresponds to the depth 174 of the brittle-ductile transition, where the shallowest melts can pool, and below which no 175 seismicity occurs. At sections of MORs that are magmatically robust enough for melt to be 176 seismically imaged, the depth to the shallowest AML reflector (circles in Fig. 2a) constitutes a 177 good proxy for the base of the brittle lithosphere (Sinton and Detrick, 1992; Phipps-Morgan and 178 Chen, 1993). This dataset is however very scarce at slow MORs. There, the seismogenic depth as 179 measured from OBS-based micro-earthquake surveys (e.g., Grevemeyer et al., 2019) constitutes 180 a reasonable estimate of H (squares in Fig. 2a). To fully account for the scatter in the data, we 181 empirically bracket our estimates between two end-member trends of lithospheric thickness vs.182 spreading rate, using the mathematical form:

$$H(U) = H_{MIN} + (H_{MAX} - H_{MIN})e^{-\frac{U}{\Delta U}}.$$
 (10)

184 The "thinnest" and "thickest" trends (red and blue curves in **Fig. 2a**) correspond to H_{MIN} , H_{MAX} , 185 and ΔU of 1.2 km, 6 km, 1.8 cm/yr; and 2 km, 26 km, and 2.7 cm/yr, respectively.

186

187 <u>2.2.2. Magmatic pressure rate</u>

188 The rate at which magma pressure increases with time in a continuously replenishing reservoir is 189 by far the most difficult parameter to constrain in our model. Indirect estimates can be obtained 190 by measuring and modeling seafloor inflation that sometimes precedes eruptions. To date, this 191 has only been achieved at Axial seamount, a volcano located on a hot-spot influenced section of 192 the Juan de Fuca ridge (Nooner and Chadwick, 2016), and at 9°50' N on the East Pacific Rise 193 (Nooner et al., 2014). We focus on the latter example, which —being more representative of a typical fast MOR segment— can be extrapolated to construct a trend of p_M with spreading rate 194 195 (Fig. 2b). Nooner et al. (2014) deployed pressure sensors along an axis-perpendicular profile to 196 measure seafloor uplift over several years. Sensors located near the axial trough uplifted at ~ 7 197 cm/yr while sensors located ~10 km off-axis did not exhibit resolvable vertical motion. Upon 198 fitting this deflection profile with two end-member models of magma pocket inflation (See Supplementary Material), we obtain ranges of plausible pressure rates: between 1.9×10^{-1} and 199 1.9×10^3 Pa/s for a Mogi model, and between 4.0×10^{-5} and 2.3×10^{-3} Pa/s for a laccolith model 200 201 (Fig. 2b)

The large discrepancy between these estimates highlights the difficulty of quantifying magma pressure build-up through direct observation. We therefore complement this approach with an order-of-magnitude assessment of \dot{p}_M vs. spreading rate largely following the model of Aharonov and Kelemen (1998). We treat a magmatic sill as a horizontal, elliptical pressurized crack of width *w*, and infinite along-axis extent. The maximum thickness *h* of the sill is linked to magma overpressure (Rubin, 1995a) according to:

208
$$h = 2(1-\nu)(1+\nu)\frac{\Delta P}{E}w.$$
 (11)

From equation (11) we can relate the pressure rate to the flux of magma inflating the sill (Φ ,

210 volume flux into a single sill, per unit length along axis):

211
$$\Phi = \pi (1 - \nu) (1 + \nu) \frac{p_M}{2E} w^2 .$$
 (12)

The total magma flux leading to a mean crustal thickness H_C (= 6 km) is simply $\Phi_C = H_C U$. Assuming this flux gets equally partitioned within *N* magma sills, we can derive an order-ofmagnitude estimate for pressure rate as a linear function of spreading rate:

215
$$\dot{p}_M = \frac{2E}{\pi(1+\nu)(1-\nu)} \frac{H_C U}{Nw^2}$$
. (13)

A high estimate of p_M vs. U can be obtained using a high estimate of E (30 GPa, Heap et al., 2020) and low estimates for w (50 m) and N (10) (Aharonov and Kelemen, 1998). Conversely, E= 5 GPa, w = 500 m (Wilcock et al., 2009), and N = 100 provide a low estimate of p_M . A geometric average of these two end-members yields a reasonable intermediate estimate of $\partial p_M / \partial U = 6.312 \times 10^7$ Pa.m⁻¹. The full range of estimates is plotted in **Fig. 2b**, and is consistent with inferences of p_M based on seafloor deflection.

222

3. Results

224 The co-evolution of $\Delta \sigma_{xx}$ and p_M over time is solved numerically through a forward Euler 225 discretization of equation (2), with instantaneous stress changes prescribed when tectonic or

226 magmatic thresholds are reached (equations 4, 7 and 8). The corresponding codes are provided as 227 part of the Supplementary Material. Typical model behavior is illustrated in Fig. 3, which 228 represents 4 characteristic types of MOR section. Fig. 3a corresponds to a slow-spreading 229 segment with an 8-km thick axial lithosphere. It shows the continuous decrease of the diking 230 threshold from its peak (post-eruption) value of ~250 MPa, in response to increasing magma 231 pressure. For a new eruption to become possible, the diking threshold has to fall below the faulting threshold (~63 MPa). Assuming a value of p_M appropriate for this spreading rate (5×10⁻² 232 233 Pa/s at U = 2.5 cm/yr according to our intermediate estimate, Fig. 2b), this occurs within ~130 234 years, which sets the characteristic recurrence time of magmatic intrusions. About 2 earthquakes 235 typically occur in 130 years, because far-field extension increases the axial stress fast enough to 236 overcome the characteristic earthquake stress drop twice in this time frame. Two out of three 237 "events" of seafloor spreading are thus tectonic. When averaged over time scales of kyrs to tens 238 of kyrs, this translates to an *M* value of 0.3. In **Fig. 3b**, the spreading and pressure rates are the 239 same, but on-axis lithospheric thickness is lowered to 3 km. This lowers both the diking and 240 faulting thresholds, as well as the difference between them. The recurrence time of eruptions is 241 thus reduced to ~50 years, which sometimes allows for an earthquake to occur, and sometimes 242 does not. The time-averaged M is consequently increased to 0.7. Similar mechanisms are at play 243 in Fig. 3c, which corresponds to a faster spreading rate, thinner lithosphere, and greater pressure 244 build-up rate. Finally, in Fig. 3d, the pressure rate is so fast that earthquakes cannot occur within 245 the short eruption recurrence time (~7 yrs). This leads to fully-magmatic seafloor spreading (M =246 1).

To further explore the primary controls on *M*, we carried out 80000 simulations spanning a wide range of pressure rates and spreading rates, with lithospheric thickness indexed on spreading rate following the "thinnest" and "thickest" end-members introduced in **Fig. 2a**. Contour lines in **Fig. 4** shows the time-averaged *M* values yielded by each of these simulations. In **Fig. 2c**, we show the trends of *M* vs. spreading rate obtained upon combining the intermediate scenario for \dot{p}_M vs. spreading rate (solid curve in **Fig. 2b**, red lines in **Fig. 4**) with the two endmember scenarios for lithospheric thickness vs. spreading rate (red and blue curves in **Fig. 2a**). These two trends successfully bracket our dataset of *M* vs. spreading rate.

255

256 **4. Discussion**

257 <u>4.1. Primary controls on M</u>

The results shown in **Fig. 2** suggest that the non-linear increase in *M* with spreading rate may reflect both the (possibly linear) increase in the rate of magma pressure build-up (**Fig. 4**), and the change in lithospheric thickness, which is most pronounced at slow rates. The key effect of increasing lithospheric thickness is to increase the difference between the peak magmatic threshold stress $\Delta \sigma_{xx}^{M,MAX}$ and the tectonic threshold $\Delta \sigma_{xx}^{F}$ (**Fig. 3**). This in turn increases the characteristic recurrence time of magmatic intrusions, which can be approximated as:

264
$$\tau_M = \frac{\Delta \sigma_{xx}^{M,MAX} - \Delta \sigma_{xx}^F}{\dot{p}_M} . \quad (14)$$

This approximation is valid as long as an earthquake stress drop remains small compared to the difference between peak magmatic and tectonic thresholds. This ensures that an eruption closely follows the moment when $\Delta \sigma_{xx}^{M}$ becomes smaller than $\Delta \sigma_{xx}^{F}$. Once the timing of intrusions is known, an expression for *M* straightforwardly follows:

269
$$M = \min\left(\frac{\delta_M}{U\tau_M}, 1\right). \quad (15)$$

The predictions from this analytical approximation closely match the values of *M* estimated from our numerical simulations over a wide range of pressure and spreading rates (**Fig. 4**).

By combining the above equations with our linear model for \dot{p}_M vs. U (equation 13), we can cancel out spreading rate from the expression of M (valid for M < 1), and obtain:

274
$$M = \frac{\delta_M \frac{\partial p_M}{\partial U}}{P_0 + \rho g \gamma H'} \quad (16)$$

275 where γ is a dimensionless factor introduced for concision: $\gamma = 1 - \mu'(1 - \lambda)$, with $\mu' = \frac{\mu}{\mu + \sqrt{1 + \mu^2}}$. P_0 denotes a characteristic pressure: $P_0 = T_0 + p_w$, which is typically small (~10 MPa)

277 compared to the $\rho g\gamma H$ term (~100 MPa). Equation (16) is plotted in Fig. 5a for several values of $\partial \dot{P}_M / \partial U$, which represent different scenarios for the increase in magma volume flux (expressed 278 279 as pressure increase in a shallow sill) with spreading rate. This graph clearly shows that a key 280 influence of spreading rate on M actually occurs through changes in lithospheric thickness H. In 281 fact, of all the parameters controlling M in our model, lithospheric thickness is the one that 282 changes most significantly from slow to fast spreading rates (Fig. 2a). The fact that using end-283 member estimates of H vs. U appropriately brackets the available dataset of M vs. U (Fig. 2c) 284 further supports the idea that variability in lithospheric thickness is a strong source of variability 285 in M at MORs.

Bracketing the *M* dataset in **Fig. 2c** works best when using our intermediate estimate for $\partial \dot{p}_M / \partial U$ (6.312×10⁷ Pa.m⁻¹), which corresponds to the geometric average of high and low estimates that span several orders of magnitude (Section 2.2.2, **Figs. 2b, 5b**). The value of $\partial \dot{p}_M / \partial U$ depends on parameters such as crustal thickness and the number (*N*) and characteristic cross-axis extent (*w*) of inflating sills (equation 13). These parameters are likely to vary significantly from one ridge section to another, but may not show a systematic trend with respect

292 to spreading rate. Crustal thickness is for example known to remain roughly constant at ~6 km 293 for most MORs, except along ultraslow ridges where it can oscillate between ~ 0 and ~ 10 km 294 from segment to segment, because of extreme along-axis melt focusing (e.g., Dick et al., 2003; 295 Cannat et al., 2008; Sauter et al., 2011; Niu et al., 2015) (See Section 4.2). Unlike crustal 296 thickness, the cross-axis extent of shallow magma reservoirs is not well characterized. 297 Seismically-imaged AMLs generally appear narrower at faster ridges (e.g., Lowell et al., 2020), 298 but magma pressure may only build up in a small portion of the AML prior to an eruption (Wilcock et al., 2009; Nooner and Chadwick, 2016), making w (of order $10^2 - 10^3$ m) difficult to 299 constrain. Finally, intrusion width δ_M is also a strong control on M in equation (16). Qin and 300 301 Buck (2008) pointed out that the typical width of dikes observed at ophiolites and MORs is ~1 302 m, and does not clearly correlate with spreading rate. These authors however noted that greater 303 tectonic stresses and a strong magmatic input could theoretically enable intrusions as wide as a 304 few meters.

305

306 <u>4.2. Implications for seafloor spreading regimes</u>

307 <u>4.2.1. Symmetric vs. detachment faulting at slow spreading rates</u>

Our model provides a straightforward explanation for the increased variability in M at slower spreading rates (**Fig. 2c**) by directly attributing it to the increased variability in axial lithospheric thickness (**Fig. 2a**). This variability is largely due to focusing of melt towards segment centers, which makes the lithosphere warmer and thinner there compared to segment ends (Cannat, 1996). Variability in magma supply also occurs on a larger scale at slow and ultraslow MORs, and shapes 50–100-km long MOR sections with axial volcanic ridges, symmetrically faulted abyssal hill morphology ($M \sim 0.6$ –0.8), and shallow seismicity. These sections alternate with areas characterized by a more complex, "asymmetric" morphology associated with widespread detachment faulting ($M \sim 0.2$ –0.6), deeper and more active seismicity, deeper melt fractionation, a greater occurrence of active hydrothermal sites, and in some extreme cases at ultraslow spreading ridges: a much thinner or virtually absent igneous crust (Dick et al., 2003; Escartín et al., 2008; Tucholke et al., 2008; Sauter et al., 2011; Olive and Escartín, 2016; Cannat et al., 2019).

321 In our theoretical framework, M values of ~0.5 and below, which enable detachment 322 faulting (Buck et al., 2005), are a direct consequence of the lithosphere being locally thicker than 323 a threshold value of ~5 km (Fig. 5a; Fig. 6) for a given magmatic influx. This is fully consistent 324 with the fact that microseismicity appears confined within 5-6 km below seafloor at abyssal hill-325 bearing sections of the Northern Mid-Atlantic Ridge, while detachment-bearing sections produce 326 micro-earthquakes down to ~12 km (e.g., Barclay et al., 2001; Parnell-Turner et al., 2017). 327 Detachment bearing sections may therefore receive an influx of melt comparable to 328 symmetrically-faulted sections, but allow a lesser fraction of this melt to intrude their colder, 329 thicker brittle lithosphere. The exact value of the threshold thickness for detachment faulting however strongly depends on $\partial p_M / \partial U$ (dashed green curve in Fig. 5b), which itself depends on 330 331 crustal thickness and a number of other poorly-constrained parameters. This could explain why 332 symmetrically faulted sections of ultraslow MORs such as Segment 27 of the Southwest Indian 333 Ridge can host micro-earthquakes ~10 km below seafloor (Yu et al., 2018). This segment is 334 indeed characterized by ~10-km thick crust (Niu et al., 2015), a likely result of extreme along-335 axis magma focusing (Dick et al., 2003). Such a high volume flux of magma could plausibly 336 cause anomalously high pressure build-up rates in magma reservoirs (compared to the global

trend of \dot{p}_M vs. spreading rate shown in **Fig. 2b**), which would enable *M* values in excess of 0.5 even in 10-km thick lithosphere (**Fig. 5**).

339 If for a given volume flux of magma, the time-averaged M is mainly set by lithospheric 340 thickness, it is important to keep in mind that lithospheric thickness itself is strongly controlled 341 by magmatic input. In a broad sense, the thermal structure of a MOR reflects a balance between 342 heat advected upwards from the asthenosphere —through magma ascent and mantle flow— and 343 heat transferred through the lithosphere by hydrothermal circulation (e.g., Chen and Morgan, 344 1990; Phipps-Morgan and Chen, 1993). Models that balance magmatic input and hydrothermal 345 output are generally successful at explaining the thermal state of MORs receiving a steady 346 magma supply, as they accurately replicate the depth to the shallowest AML at intermediate to 347 fast rates (Fig. 2a) (e.g., Phipps-Morgan and Chen, 1993). In these models, the heat input stems 348 from magmatic emplacement below the lithosphere (e.g., Maclennan et al., 2004), yet repeated 349 dike intrusions in the brittle domain are also likely to influence the equilibrium thermal structure 350 (e.g., Behn and Ito, 2008). Feedbacks are also likely to exist between hydrothermal and tectonic 351 processes. For example, fracturing associated with the damage zones of large-offset faults could 352 increase crustal permeability, and thus the vigor of hydrothermal convection in detachment-353 bearing ridge sections (Escartín et al., 2008). Enhanced cooling would further thicken the 354 lithosphere, which in turn would promote lower M values that favor detachment faulting, in a 355 self-reinforcing feedback loop that would stabilize the detachment forming regime.

356

357 <u>4.2.2. Transition to fully-magmatic spreading at faster spreading rates</u>

Another important feature of our model is that it predicts conditions under which seafloor spreading becomes fully magmatic (M = 1). This specifically occurs when axial lithosphere 360 becomes thinner than a critical thickness that is straightforwardly obtained by setting M = 1 in equation (16), and strongly depends on $\partial p_M / \partial U$ (Fig. 5). In the framework of Buck et al. (2005) 361 362 and Liu and Buck (2018; 2020), reaching M = 1 marks the transition between a mode of 363 spreading dominated by tectonic stretching, where M directly controls the characteristics of 364 faults that bound an axial valley (Behn and Ito, 2008; Olive et al., 2015), and a buoyancy-365 dominated regime where an axial high forms due to the upward load exerted on the lithosphere 366 by low-density melts (Fig. 6). In this regime, only small inward- and outward-dipping faults 367 form to accommodate near-axis lithospheric flexure (e.g., Escartín et al., 2007). In real systems, 368 the transition from axial valley to axial rise occurs at intermediate rates, between 5.5 and 7.5 369 cm/yr (Dick et al., 2003), which corresponds to axial lithospheric thicknesses between ~ 2 and ~ 5 km (Fig. 2a). This is well accounted for by our model when using values $\partial \dot{p}_M / \partial U$ close to 10^8 370 371 Pa.m⁻¹ (Fig. 5b), which are well within the plausible range estimated in Section 2.2.2 and plotted 372 in Fig. 2b. This said, the extent to which the transition from axial valley to axial high truly 373 represents a transition to fully-magmatic spreading remains debatable. Ito and Behn (2008) for 374 example showed that alternating between periods of M = 0 and M = 1 lasting kyrs to tens of kyrs 375 could allow the growth of an axial high with a time-averaged M < 1, as is seen at some 376 intermediate and fast-spreading MORs. In numerical models, this requires the axis to be weak 377 enough to rapidly develop positive relief during intermittent, fully-magmatic phases.

A major transition in seafloor spreading modes was also observed in recent analog experiments conducted by Sibrant et al. (2018) using Ludox. This viscous colloidal dispersion behaves as an elastic-brittle material when mixed with salt. By pouring a saline solution from the top down into a tank filled with Ludox, Sibrant et al. (2018) were able to model a brittle lithosphere that diffusively thickens off-axis and overlies a fluid asthenosphere. Upon subjecting 383 this system to increasing extension rates, they observed a transition from fault-accommodated 384 plate separation to a regime dominated by vertical Ludox "intrusions". They proposed that this 385 occurs when the axial thickness of the lithosphere (controlled by extension rate) falls below the 386 characteristic size of the process zone of mode-I cracks. They further showed that this reasoning 387 could be upscaled to real MORs to predict the switch to fully-magmatic extension at fast-388 spreading rates. An interesting analogy can be drawn with our approach, in which intrusion-389 dominated spreading occurs when (1) magma pressure builds up at a rapid rate in shallow 390 reservoirs and (2) axial lithosphere is thin enough that the difference between peak magmatic and 391 tectonic thresholds is low. Both of these conditions should be met at fast-spreading MORs. In the 392 experiments of Sibrant et al. (2018), it is unclear whether fluid pressure builds up at the base of 393 the lithosphere prior to intrusion events, and whether the rate of pressure increase is affected by 394 spreading rate. It may instead be controlled solely by the buoyancy difference between pure and 395 saline Ludox. If the pressure build-up rate is sufficiently fast, Ludox intrusions may become so 396 frequent that the differential stress required for faulting is never reached at the axis. In any case, 397 our respective modeling approaches agree that the transition to fully magmatic spreading 398 involves a threshold of axial lithospheric thickness (Fig. 5b; Fig. 6).

399

400 <u>4.3. Model limitations and possible improvements</u>

We have attempted to construct the simplest possible model that explains global trends of M vs. spreading rate in terms of stress interactions between magmatic and tectonic events. By doing so, we have made a number of simplifying assumptions that we briefly discuss here. First, we model the seismic cycle as a time-predictable process (e.g., Shimazaki and Nakata, 1980) with a constant stress build-up rate, triggering stress, and earthquake stress drop, all of which are likely

406 to fluctuate through time in a real system. In particular, we have only considered magnitude ~ 6 407 earthquakes, each accounting for 1 m of horizontal slip with total slip and 1 MPa of stress drop. 408 In reality, normal faulting earthquakes at MORs follow a Gutenberg-Richter distribution with 409 relatively high b-values (Olive and Escartín, 2016), which means that most of the seismically-410 accommodated extension is accounted for by the largest earthquakes. The significant deficit of 411 MOR earthquakes compared to their expected moment release however suggests that a sizable 412 portion of tectonic extension may occur aseismically, particularly at faster-spreading ridges 413 (Cowie et al., 1993; Frohlich and Wetzel, 2007; Olive and Escartín, 2016). This idea is backed 414 up by recent numerical models of MOR seismic cycles (Mark et al., 2018).

415 Second, we rely on a simplified model for intrusion triggering which involves reaching the 416 tensile strength of the lithosphere at the roof of a magma pocket under the combined influence of 417 magmatic overpressure and tectonic stress. This is consistent with recent seismological 418 investigations of the early stage of an East Pacific Rise eruption, which began with synchronous 419 ruptures distributed over a distance spanning multiple distinct magma reservoirs (Tan et al., 420 2016). This was interpreted as multiple, independently replenished AMLs --potentially out-of-421 sync with respect to magma pressure— being brought to failure synchronously by tectonic 422 loading. Our model obviously does not capture the complexity of such events, warranting further 423 investigations of the 3-D stress field around replenishing sills (e.g., Wilcock et al., 2009; 424 Supplementary Fig. S1), and the consideration of alternative triggering criteria for example based 425 on fracture mechanics (e.g., Sibrant et al., 2018). Another limitation of our model is its 426 simplified treatment of magmatic pressure build-up, which we assume to occur at a constant, 427 poorly constrained rate, lumping together a number of complex processes related to sill 428 emplacement and replenishment in the lower oceanic crust. On the one hand, whether sill

429 replenishment happens continuously or in pulses depends on the way magma ascends to the base 430 of the lithosphere and through the axial mush zone, which could involve wave-like transport 431 (Parnell-Turner et al., 2020). On the other hand, the link between sill inflation rate and pressure 432 build-up potentially involves damping of pressure changes by viscous relaxation of hot rocks 433 above the magma (e.g., Kelemen and Aharonov, 1998) and poro-elastic effects in the 434 surrounding mush zone. Better characterizing the stress changes imparted by replenishing 435 magma reservoirs at MOR axes before and after intrusions will be key to improving our theory. 436 This will involve continuous monitoring of MOR axes through seafloor geodesy and seismology 437 (e.g., Nooner and Chadwick, 2016), with the caveat that extrapolating short-term measurements 438 to geological time scales will remain difficult until we at least capture a full seismic / volcanic 439 cycle, particularly at a slow ridge section not influenced by a hot spot (Tolstoy et al., 2001; 440 Dziak et al., 2004). Another difficulty in bridging time scales of seafloor spreading is the 441 increasing amount of evidence for temporal fluctuations in MOR activity on periods ranging 442 from 10s of kyrs to millions of years (e.g., Parnell-Turner et al., 2020). This has led Tolstoy 443 (2015) to question whether the behavior of MORs that we observe today is representative of past 444 or longer-term activity.

The last important limitation of our approach is the assumption that dikes and faults compete to release the elastic strain energy stored in axial lithosphere, but have no possibility of triggering each other. Such interactions have however been observed several times at rifts and ridges. Calais et al. (2008) for example documented a 2007 slow slip event on a normal fault that likely promoted the intrusion of a dike in the eastern branch of the East African Rift. Tolstoy et al. (2001) reported the reciprocal phenomenon of a migrating dike triggering a dozen magnitude ~5 earthquakes along the normal faults that bound the axial valley of the Gakkel ridge (85°N), 452 throughout most of the year 1999. This raises the question of whether tectonic extension at 453 MORs primarily occurs during or in response to magmatic events. We note that the 20+ km 454 length of the valley-bounding faults at Gakkel makes them susceptible to nucleating magnitude 455 ~6–6.5 earthquakes (Wells and Coppersmith, 1994). Such events could occur separately from 456 intrusions and would account for ~3 times more slip over the same along-axis distance compared 457 to ~10 magnitude-5 earthquakes. Again, a longer record of observations is needed to assess 458 whether the tectonic extension that accompanies diking is systematically small compared to the 459 associated magmatic extension, and/or to purely tectonic spreading events. Dziak et al (2004) for 460 example estimated that all the seismicity associated with the 2001 dike intrusion event at the 461 Lucky Strike segment of the Mid-Atlantic Ridge amounted to centimeters of horizontal 462 extension, which would be negligible if the dike alone accommodated ~ 1 m of extension. In any 463 case, available observations warrant the development of more detailed seismic cycle models of 464 MOR faults (e.g., Mark et al., 2018) that interact with an inflating sill and episodic dike 465 intrusions. This will require detailed constraints on the geometry of the fault(s) relative to that of 466 the magmatic plumbing system, as static stress transfers often display sharp spatial variations 467 (Supplementary Figure S1). The modes of volcano-tectonic or tectono-volcanic triggering may 468 thus be strongly dependent on the local geometry of the ridge, e.g., a deeper magmatic system at 469 a slower MOR would influence faults within a broader horizontal area compared to a shallow 470 sill-dike system at a fast MOR (e.g., Tolstoy et al., 2001).

471

472 **5.** Conclusions

We have designed a simple theoretical framework explaining the long-term partitioning ofmagmatic and tectonic extension at MORs in terms of short-term cycles of earthquakes and dike

intrusions that interact with one another by modulating the stress state of the ridge axis (**Figs. 1**, **3**). Axial lithospheric thickness and the rate of pressure build-up in shallow, replenishing magma sills are the two main parameters controlling M in our model (**Fig. 4**). We constructed plausible scenarios for the increase of pressure rate with spreading rate, as well as end-member trends for lithospheric thickness across MOR spreading rates (**Fig. 2a, b**). This allowed us to bracket the observed non-linear increase in M and relate its reduced variability with increasing spreading rate (**Fig. 2c**) to reduced along-axis variability in lithosphere thickness.

482 To first-order, spreading rate controls the flux of magma that reaches the ridge axis. The 483 thermal component of this flux fuels hydrothermal circulation, which modulates the strength and 484 thickness of axial lithosphere. The mass component of this flux contributes to setting the pace of 485 pressure build-up in shallow sills, which together with lithosphere thickness, sets the value of M. 486 This in turn controls the intensity of tectonic deformation, which could affect the vigor of 487 hydrothermal circulation through pervasive fracturing of the crust, exerting a feedback on 488 lithosphere thickness. The dichotomy between symmetrically faulted and detachment-bearing 489 sections of slow MORs could thus reflect thresholds in lithosphere thickness under a near 490 constant along-axis magma flux (Figs. 5 & 6). Extreme focusing of melt along the axis of 491 ultraslow MORs however points to the combined action of a spatially-varying magma flux and 492 lithosphere thickness in shaping symmetrically-faulted segments and quasi-amagmatic sections. 493 On the other end of the spectrum, the large, spatially uniform influx of magma to fast MORs 494 ensures a uniformly thin lithosphere with frequent intrusions, which makes fully magmatic 495 spreading possible. Overall, modes of seafloor spreading reflect a balance between magmatic, 496 tectonic, and hydrothermal processes, raising the question of whether temporal fluctuations in 497 magma supply can alter the stability of these equilibria and trigger regime transitions.

498

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510 Appendix A.

511 Supplementary Material related to this article consists of:

512 • Supplementary Information (.pdf file). Additional details regarding the modeling of seafloor 513 inflation at the East Pacific Rise to infer pressure increase rates in a shallow magma reservoir. 514 Instructions on how to use the supplementary codes. This file includes 4 Supplementary Figures. 515 • Supplementary Code: M across rates.m; a MATLAB script producing Figs. 1a, 2, 3, 4 and 5. 516 • Supplementary Code: tectono magmatic fcn.m; MATLAB implementation of the toy model of 517 tectono-volcanic cycles. 518 • Supplementary Text files: .txt files for use by the MATLAB scripts, containing datasets of M 519 (M data.txt), micro-earthquake depths (depth of microEQs.txt), and AML depths

520 (depth_of_AMLs.txt).

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Figure 1 – a. Fraction of magmatic extension across spreading rates. MAR: Mid-Atlantic Ridge;
GSC: Galapagos Spreading Center; JdF: Juan de Fuca Ridge; CR: Chile Ridge; SEIR: Southeast
Indian Ridge; ELSC: Eastern Lau Spreading Center; EPR: East Pacific Rise. Data from Cowie et
al. (1993); Ito and Behn (2008); Olive and Escartín (2016) and Howell et al. (2016). b. Model
setup.

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692 Figure 2 - a. Estimates of axial lithospheric thickness from seismically imaged AML depths 693 (circles, as compiled by Buck et al., 1997; Singh et al., 2006) and depth of micro-earthquakes 694 (squares, as compiled by Grevemeyer et al., 2019). Colored lines correspond to thick and thin 695 end-members as defined by equation (10). Light-green stars correspond to example cases shown 696 in Fig. 3. Sd: Slow, detachment faulting regime; Ss: Slow, symmetric faulting regime; I: 697 Intermediate; F: Fast. b. Modeled pressure rate vs. spreading rate (equation 13), with 698 intermediate estimate as continuous line and high and low estimates as dashed lines. Estimates 699 from seafloor deflection at the EPR (See Supplementary Material) are shown as vertical bars 700 (Mo: Mogi; La: Laccolith models). c. Same dataset of M vs. U as in Fig. 1, with curves 701 corresponding to the thin (red) and thick (blue) end-members models.

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Figure 3 – Modeled tectono-volcanic cycles (axial stress $\Delta \sigma_{xx}$, in black, and diking threshold $\Delta \sigma_{xx}^{M}$, in red, vs. time) in 4 example configurations. **a.** slow, detachment-bearing ridge section (Sd in Fig. 2): H = 8 km; U = 2.5 cm/yr. **b.** slow, symmetrically faulted ridge section (Ss): H = 3km; U = 2.5 cm/yr. **c.** Intermediate-spreading MOR (I): H = 2.5 km; U = 6 cm/yr. **d.** Fastspreading MOR (F): H = 1.5 km; U = 15 cm/yr. In each example the pressure rate is 3.1560×10^7 Pa.m⁻¹.s⁻² times the spreading rate (intermediate estimate) from Fig. 2b. 709

Figure 4 – Model predictions of *M* as a function of spreading rate and magma pressure rate. The color code corresponds to the analytical approximation from equations (14) and (15) while black contour lines (spanning M = 0.1 to 1, with increments of 0.1) show results from the simulations. The red line is the average estimate of pressure rate vs. *U*, and the dashed red line is the high estimate. Panels a and b respectively correspond to the thin and thick lithosphere end-members of *H* vs. *U* shown in Fig. 2a.

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Figure 5 – **a.** Lithospheric thickness control on *M* for different values of $\partial p_M / \partial U$ corresponding to the vertical bars in panel b. The solid curves correspond to the low (blue), intermediate (black) and high (red) estimates of pressure rate vs. spreading rate from Fig. 2b. **b.** Green dashed curve: critical lithospheric thickness above which extension becomes detachment-dominated (M = 0.5, dashed green circle in panel a), plotted against $\partial p_M / \partial U$. Green solid curve: critical lithospheric thickness below which extension becomes fully magmatic (M = 1, solid green circle in panel a). These critical thicknesses are obtained by setting M = 0.5 or 1 in equation (16).

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Figure 6 – Schematic diagram of major seafloor spreading modes as a function of spreading rate
and axial lithosphere thickness. Vertical orange arrow represents volume flux of magma to
shallowest reservoir (orange ellipse), which is assumed to scale linearly with spreading rate.
Horizontal dashed lines represent lithosphere thickness thresholds identified in Fig. 5. Each
spreading mode corresponds to a case shown in Fig. 3: Sd: Slow, detachment faulting regime;
Ss: Slow, symmetric faulting regime; I: Intermediate; F: Fast.



Figure 2 Click here to download Figure: Figure_2_M_



b.



C.

magmatic fraction











SPREADING RATE

Supplementary Information for "Controls on the magmatic fraction of extension at mid-ocean ridges"

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Text S1: Magmatic pressure rates at the East Pacific Rise

Here we provide two different estimates of the magmatic pressure rate characterizing the inflation of a shallow melt lens at the axis of a mid-ocean ridge. We specifically rely on the seafloor uplift rates recorded by Nooner, Webb, Buck, and Cormier (2014), and fit the measurements with two end-member models of magma reservoir.

Corresponding author: J.-A. Olive, Laboratoire de Géologie, École Normale Supérieure / CNRS UMR 8538, PSL Research University, Paris 75005, France (olive@geologie.ens.fr) The first one is a Mogi point source within an elastic half-space (Mogi, 1958). In this case, the surface vertical displacement rates v are given by:

$$v(r) = \frac{(1-\nu)V_0}{\pi} \frac{d}{(r^2+d^2)^{3/2}},\tag{1}$$

where d is the depth of the source, r is the radial distance from the source at the seafloor, ν is the Poisson's ratio of the elastic half-space and V₀ an equivalent volumetric flux, defined as:

$$V_0 = \frac{\pi \dot{p}_m a^3}{G},\tag{2}$$

 \dot{p}_m , *a* and *G* being the magmatic pressure rate, the source radius and the shear modulus of the elastic medium respectively. Here, the Poisson's ratio ν is considered constant $\nu = 0.25$ and we invert for the depth *d* and the volumetric flux V_0 using a grid search algorithm. Confidence intervals on *d* and V_0 are presented in figure S2a., and the comparison between acceptable models and the data are shown in figure S2b. V_0 ranges from 1.2×10^6 to 4.8×10^6 m³.yr⁻¹ at the 60% confidence level (figure S2a.). End-member values for the source radius (a = 50 m and a = 500 m) and the shear modulus (G = 2 GPa and G = 12GPa, Heap et al. (2020)) lead to plausible pressure rates \dot{p}_m ranging from 1.9×10^{-1} to 4.6×10^3 Pa.s⁻¹.

We then inverted the same dataset using the laccolith model of (Turcotte & Schubert, 2002). In this setup, the lithosphere is treated as a thin elastic plate of rigidity D uplifted above a magma reservoir (of full width L and overpressure p_m). The surface vertical displacement are given by:

$$v(r) = \begin{cases} \frac{\dot{p}_m}{24D} \left(r^4 - \frac{L^2 r^2}{2} + \frac{L^4}{16} \right) & \text{if } r \le L/2 \\ 0 & \text{if } r > L/2 \end{cases}$$
(3)

D is given by:

$$D = \frac{EH^3}{1 - \nu^2},\tag{4}$$

where E, ν and H are the Young's modulus, the Poisson's ratio and the thickness of the elastic plate. The Poisson's ratio and the thickness are considered fixed here at $\nu = 0.25$ and H = 1.5 km (Fig. 2a, see main text). We invert for the magmatic pressure rate \dot{p}_m and the laccolith width L assuming two extreme scenarios for the Young's modulus (E = 5 GPa and E = 30 GPa, corresponding to G = 2 GPa and G = 12 GPa, Heap et al. (2020)). Results are shown in figures S3 (E = 5 GPa) and S4 (E = 30 GPa). To the 60% confidence level, \dot{p}_m ranges from 4.0×10^{-5} to 2.3×10^{-3} Pa.s⁻¹.

Text S2: Reproducing results and figures with supplementary codes

We provide supplementary codes to fully reproduce the results and figures from the main text. MATLAB script $M_{across_rates.m}$ plots the analytical model, and uses function $tectono_magmatic_fcn.m$ to run toy model simulations. To generate Figs. 1b; 2; 3; 4 and 5, one can run $M_{across_rates.m}$ by setting variable regime (line 9) to 'thin' or 'thick' to select an end-member scenario for lithospheric thickness vs. spreading rate (red and blue curves in Fig. 2a). All scripts must be run from the same folder, which must include the three .txt files containing the M ($M_{ata.txt}$), micro-earthquake depth ($depth_{of_microEQs.txt}$), and AML depth ($depth_{of_AMLs.txt}$) datasets, respectively.

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Figure S1. Change in total horizontal stress induced by an over-pressured sill of half-width *a* (red line), normalized by overpressure (Pollard & Segall, 1987). Yellow lobes show areas where tensile stress builds up and a dike is likely to initiate. Tensile stress perturbation in this region is of same order of magnitude as magmatic overpressure. White dashed lines show contours of zero stress perturbation.



Figure S2. a. Source depths and volumetric fluxes explaining seafloor vertical displacements at the East Pacific Rise following the model of Mogi (1958). Labels on the contours indicate the confidence level. Colored dots indicate the parameters used for the model predictions shown in b. b. Data (red dots) vs. model predictions (colored lines) of equation (1).



Figure S3. Same figure as figure S2 for the laccolith model of equation (3), assuming G = 2 GPa.



Figure S4. Same figure as figure S2 for the laccolith model of equation (3), assuming G = 12 GPa.