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The build-up and triggers of volcanic eruptions

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1	The build-up to and triggers of volcanic eruptions
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10	Abstract
11	Volcanic eruptions can directly impact more than 800 million people living in proximity to
12	active volcanoes. Thus, anticipating the future behaviour of volcanic systems serves to
13	mitigate the effects of eruptions on our society. Essential to this target is an understanding of
14	the fundamental processes driving volcanic activity. Here, we review the processes leading to
15	magma accumulation in the Earth's crust and the temporal evolution of the thermal and
16	physical properties of magma storage regions. We discuss mechanisms to initiate magma
17	reservoir failure and the ascent of magma from crustal storage regions, including the
18	factors that control whether magma reaches the surface or stalls at depth. We show that the
19	evolution of temperature and the physical properties of volcanic plumbing systems favour
20	volcanic activity after a period of thermal priming, while storage becomes more likely for
21	mature volcanic systems with large reservoirs (hundreds of km^3) when the crust is relatively
22	warm. Anticipating volcanic activity requires a multidisciplinary approach as monitoring
23	and geophysics provide information on the current state of system while petrology and the
24	eruptive history are essential to trace the temporal evolution of volcanic systems over longer
25	timescales. Modelling serves to link these different observational timescales, and the
26	inversion of datasets using physics-based statistical approaches is a promising way forward

to advance our understanding of the processes controlling recurrence rate and magnitude of
volcanic eruptions.

29

30 **1. Introduction**

31 Volcanic eruptions occur when magma reaches or, for some phreatic eruptions, approaches the surface^{1,2}. In order to anticipate the timing, size, and style of volcanic 32 33 eruptions, we need a scientific understanding of magma plumbing systems and the processes that govern the transfer of magma to the surface³. Volcanic eruptions are the culmination of a 34 35 long series of processes that occur in disparate regions of the earth's lithosphere, starting with 36 the generation and supply of melt from the mantle, the accumulation of magma in crustal 37 reservoirs, and the transport of magma between and within storage regions and to the surface 38 (Fig.1). Each of these processes represents a crucial step in the journey of magma to the 39 surface, and in the last few decades we have made considerable progress in understanding 40 particularly the storage and transport of magma through the crust. From these advances, a 41 counterintuitive truth has emerged: it is surprisingly difficult for magma to reach the surface. 42 Volcanic eruptions require a sufficiently large volume of magma with adequately low viscosity and density to reach the surface without totally solidifying en-route $^{4-10}$. However, 43 44 the assembly and growth of magma reservoirs in the crust requires a supply of heat sufficient 45 to overcome freezing against the host rocks. While this may be easier in the lower crust 46 where more primitive magma reservoirs are thought to form, the development of more silicic 47 reservoirs in the shallow crust likely requires a protracted period of thermal maturation^{11,12}. 48 Even if a magma reservoir can be thermally sustained, in order to feed a potential eruption it must build up pressure, although the range of overpressure required is debated^{13,14}. External 49 triggers such as earthquakes¹⁵, landslides^{16–18}, and tides¹⁹ can provide the "final kick" at 50 51 different time scales to destabilize a magma reservoir, but the small pressure changes

52 associated with some of these processes require the reservoir to be close to failure to trigger 53 magma ascent. Once magma starts ascending from a reservoir, its ability to reach the surface 54 depends on a wide variety of factors, including: the evolution of physical properties of the magma, which is largely influenced by the behaviour of volatiles^{5,20,21} and heat loss to the 55 surrounding rocks²²¹⁸: the physical properties of the surrounding rocks; and the local stress 56 57 field, which is influenced by local tectonics and topographic loading of the volcanic edifice^{8,23–25}. The reservoir must supply sufficient energy and volume to drive magma to the 58 surface without it becoming thermally or mechanically arrested²⁶⁻²⁸. Many episodes of 59 60 volcanic unrest that do not culminate in an eruption are thought to represent "failed eruptions" or dike intrusions that became arrested at shallow depths²⁹. Considering all the 61 62 barriers to magma reaching the surface, volcanic eruptions may seem improbable; however, an average of 50 volcanoes erupt every year on Earth³⁰. The prevalence of eruptions despite 63 64 thermal and mechanical obstacles is consistent with a much larger amount of magma emplaced at a depth that never reaches the surface 12,31-36. 65

66 In this review, we discuss the processes behind volcanic eruptions following the 67 journey of magma to the surface. Although we focus on eruptions from polygenetic 68 volcanoes with crustal sources, many of the general principles we discuss are applicable to 69 monogenetic volcanoes. We begin with a short summary of observations about volcanic 70 eruption triggers gathered from analyses of erupted products and volcano monitoring signals. 71 Next, we review the processes responsible for magma accumulation, reservoir pressurisation, 72 and the factors that might promote or hinder the propagation of magma to the surface. We 73 close this review by summarising the current challenges to understanding what controls 74 volcanic eruptions and highlighting how multidisciplinary research is key to our endeavour.

75

76 **2. Magma storage**

77 **2.1 Pre- and post-eruption record**

78 We cannot directly observe the long-term (hundreds to hundreds of thousands of 79 years) processes preceding volcanic eruptions both because of their timespan and the 80 inaccessible depths at which these processes occur. Therefore, we rely on monitoring data 81 along with data from the textures and chemistry of erupted materials, as well as field and 82 structural geology to build models for the sequence of events culminating in an eruption. 83 Volcano monitoring is essential to determine the status of volcanic systems and identify the 84 potential signs of an impending eruption. The number of volcanoes for which data are 85 available has recently increased, thanks to the advent of satellite technology and deployment of state-of-the-art ground-based instrumentation^{37–45}. Such data availability⁴⁶ has opened the 86 87 monitoring and wav to multi-parametric stimulated the comparison between volcanoes^{37,38,41,44,47-50}. 88

89 The chemistry and textures of volcanic rocks contain information about pre- and syneruptive processes that are decrypted using experiments (e.g. Refs.^{20,51–59}). Combining the 90 91 investigation of eruptive products and intrusive and eruptive geometries/patterns in the field 92 with the multiparametric monitoring record of recent volcanic eruptions is a promising tool to shed light on the sequence of events leading up to volcanic eruptions^{38,43,60–87}. Even 93 94 considering a highly idealized magmatic plumbing system (Fig. 1), similar monitoring signals 95 and petrologic phenomena may be produced by fundamentally different magmatic processes. 96 For instance, an increase in CO_2 -rich or H_2O -rich emissions at the surface could be explained equally well by the release of fluids during magma ascent from depth (e.g. Ref.⁸⁸; Fig. 1a), or 97 98 by fluid flushing from depth that triggers the release of gases to the surface 89,90 (Fig. 1d). The 99 presence of partially reacted minerals, or groups of minerals of distinct chemistry in volcanic 100 products (Fig. 1b), could be the result of interaction between a hotter, more mafic magma and 101 a colder, more felsic magma, or it could be the product of the interaction between magma and 102 fluids ascending from depth (e.g. Ref.⁵⁸). Both magma injection and fluid flushing/fluxing 103 (the percolation and chemical interaction between externally sourced magmatic fluids and 104 magma) could drive inflation of the volcanic edifice detected geodetically at the surface and 105 produce similar petrologic signals^{91–94}. These considerations highlight the need for a 106 multidisciplinary approach to determine the sequence of events that finally lead to a volcanic 107 eruption.

108

109 2.2. Assembly of magma reservoirs

110 Geologic and geodetic data together with modelling show that the transfer and accumulation of magma in the Earth's crust is not continuous^{8,26,95–97}. However, thermal 111 112 modelling shows that the long-term (hundreds of thousands to millions of years) thermal and 113 physical evolution of magma within crustal plumbing systems can be described by the average rate of magma input^{11,12,36,98}. The thermal evolution of volcanic plumbing systems 114 115 directly impacts the capacity of magmatic systems to feed volcanic activity by controlling the 116 rate of accumulation of eruptible magma and the temporal evolution of the physical 117 properties of magma and rocks surrounding the plumbing system. To illustrate quantitatively these effects we use the thermal modelling results of Ref.³⁶ simulating the episodic and 118 119 prolonged input of magma in the crust.

The calculations show that the presence of even a minor amount of exsolved (or excess) volatiles strongly increases magma compressibility^{99,100} (Fig. 2a). Magma input into the crust results initially in rapid cooling below its solidus temperature, after which the model reservoir starts growing at a rate proportional to the rate of magma input (Fig. 2b, c). This initial "incubation period" is longer for lower rates of magma input. For the greatest input rate presented here, eruptible magma (defined here as T>750 °C, melt fraction>0.4) starts accumulating after a few hundred thousand years, while for the lowest magma input rate, 127 eruptible magma is sporadically present throughout the 1.5 million years of magma injection 128 in the crust (Fig. 2b, c). For both rates of magma input, compressibility increases rapidly and 129 then reaches a relatively constant value over time (Fig. 2d, e). Once the relatively constant 130 values are achieved, magma compressibility is greater in the reservoir assembled with higher 131 rates of magma input, as it contains a higher fraction of magma above solidus (Fig. 2d, e). To 132 summarise, after a period of time over which eruptible magma is present in extremely small fractions, i.e. the "incubation period¹¹", eruptible magma starts to accumulate for relatively 133 134 high rates of magma input, while is present sporadically for lower rates (Fig. 2b, c). With increasing time, the wall rock temperature increases (i.e. viscosity decreases¹²), as does 135 136 magma compressibility, both of which decrease the capacity of a magma reservoir to 137 pressurise to values sufficiently high to initiate magma ascent. Moreover, the release of 138 exsolved H₂O-rich fluids results in the variation of the temperature-melt fraction 139 relationships, which, in turn, decreases the capacity of heat (i.e. magma) input to alter the properties of the resident magma¹⁰¹. All together, these results suggest that the likelihood of 140 141 an eruption to occur is lower during the incubation period because of the small amount of 142 eruptible magma available, higher after substantial eruptible volumes accumulate, but then lower again over time once volatiles exsolve and the magma becomes highly compressible¹⁰². 143 144 However, on the long-term, the accumulation of super-solidus volatile-rich magma leads to a 145 progressive decrease in magma density, and an associated increase of magma buoyancy, 146 which could bring the magmatic system to critical conditions and initiate the release of magma from the reservoir 102-104. Somewhat counterintuitively, the time window over which 147 148 eruptions are most likely is shorter for systems assembled at the highest rates of magma input 149 because the temperature of the wall rock and the compressibility of magma increase more 150 rapidly in systems assembled at high rates of magma input. We note that these calculations do 151 not include the possibility of volatile outgassing, which would decrease compressibility over

time^{105–107}, nor do they include the effects of magma withdrawal from a reservoir (e.g. during an eruption); the removal of mass and heat associated with frequent withdrawal events could inhibit the growth of large magma reservoirs and potentially prolong both the incubation period as well as the period over which a plumbing system feeds volcanic activity¹⁰³.

156 Internal magma reservoir dynamics such as magma mixing, convective overturn, and 157 crystal-melt-volatile phase separations can also influence the rheology, and thus the 158 eruptibility, of magma. If the Rayleigh number is sufficiently high, magma mixing and convective processes may act to homogenize a magma reservoir¹⁰⁸⁻¹¹⁰, which has been 159 thought to stimulate the exsolution of volatiles if an intruding magma is volatile-rich¹¹¹, but 160 161 could also lead to resorption of volatiles if an intruding magma supplies sufficient heat and pressure¹¹². Volatile exsolution can be a driver of mixing processes by altering magma 162 density¹¹³⁻¹¹⁵, or if volatiles migrate more easily they can act to suppress mixing and melt 163 migration^{116,117}. Magma mixing and convective overturn have been invoked as eruption 164 triggers^{110,111,118–120}, but we note that most physical models for internal magma reservoir 165 166 dynamics typically do not include mechanical interaction with the surrounding crust, and that 167 this link would be an important target for future work to better understand how mixing 168 processes contribute to reservoir overpressure.

169

170 2.3. The "critical overpressure" for magma reservoir failure

Magma transport out of crustal storage zones takes place primarily by fluid-driven fracturing into the surrounding brittle crust, and potentially to the surface^{22,121-123}. Depending on the geometry and orientation of these magma-filled fractures, they are referred to as dikes, sills, inclined sheets, cone sheets, or ring dikes¹²⁴⁻¹²⁶. Although in detail magma-filled fractures display structural complexities and may even intrude as viscous fingers through poorly-consolidated sediments in the uppermost part of the crust¹²⁷, the vast majority of these

intrusive structures are approximately planar and dominantly opening-mode fractures^{128–130}, 177 178 which we will hereafter refer to as dikes for convenience. The details surrounding the 179 initiation of dikes and their connection to magma chambers are poorly understood, which has 180 led to competing theories about the criteria to initiate a diking event. Because dikes are 181 dominantly opening-mode structures, most magma transport models adopt a criterion based on one of two conditions: (1) *tensile failure* of wall rocks^{13,131,132} or (2) the *dilation and* 182 propagation in pre-existing fractures driven by magma pressure^{102,133–136}. Less commonly, 183 shear failure criteria are included^{137,138}. 184

185 *Tensile failure* in otherwise unfractured wall rocks can occur if the stress tangential to 186 the chamber walls $\sigma_{\theta\theta}$, sometimes referred to as the "hoop stress", exceeds the tensile 187 strength of the wall rocks T_s minus the pore-fluid pressure p_w (Box1), using a tension-188 positive convention:

189

$$\sigma_{\theta\theta} \ge T_{s-} p_w \tag{1}$$

190 In hydrostatic conditions, $p_w = \rho_w gh$, where ρ_w is the fluid density, g is gravitational acceleration, and h is depth 13,131,139,140 . Laboratory measurements on rocks suggest an average 191 tensile strength of ~0-10 MPa (e.g. Ref.¹⁴¹). Elastic models for the stress around spherical 192 magma chambers indicate that the magma overpressure ΔP_{crit} at which the hoop stress 193 becomes tensile is approximately twice the lithostatic pressure^{13,142,143}. In other words, for a 194 195 spherical magma chamber at ~8 km depth or ~200 MPa lithostatic pressure in a homogeneous 196 elastic crust, ΔP_{crit} would be ~400 MPa. Evidence that magmas in shallow crustal reservoirs 197 reach such extreme overpressures is lacking, which suggests that either 1) irregularities along 198 the chamber margins or deviations from a spherical shape act to concentrate stress, leading to tensile fracture at substantially lower overpressures^{144,145}; or 2) the host crust is not a 199 200 homogeneous elastic solid but contains pre-existing weaknesses.

201 *Pre-existing fractures* along magma chamber walls may be dilated if the magma 202 pressure exceeds the fracture-normal compressive stress^{133,134}. If magma initially invades and 203 pressurizes a pre-existing fracture, it may lead to further fracturing and the creation of new 204 fractures near the tip region that magma can flow into. In other words, these pre-existing 205 fractures may help to weaken the host rock and get the magma initially moving out of the 206 reservoir, but ultimately the magma may still create its own path. In order to propagate a pre-207 existing fracture of initial length l (Box 1), the magma overpressure must reach:

208
$$\Delta P_{crit} \ge \frac{\kappa_c}{\sqrt{l}} \tag{2}$$

where K_c is the fracture toughness of the rock, which may be on the order of ~1-10 MPa m^{1/2} 209 for centimetre- to meter-scale fractures^{22,102}. Magma will more easily propagate in longer, 210 211 suitably oriented pre-existing fractures. The critical overpressure to propagate a fracture that 212 is 1 meter long is only $\sim 1 - 10$ MPa (Fig. 3), which is significantly less than that required to 213 create new fractures altogether and therefore a more likely scenario. If wall rocks are weak or 214 fracture toughness is minimal, the critical overpressure may instead be that required to overcome magma freezing against the fracture walls²² or the viscous resistance of magma to 215 flow through a narrow fracture 8,146 . 216

Viscous deformation of the wall rocks also may influence whether magma reservoir failure can occur (Fig. 3), especially over longer timescales as magma heats up the surrounding rocks or develops a layer of "crystal mush" around the edges of the reservoir (Box 1). This idea has led to the development of magma chamber models that consider a viscoelastic crust^{147–149}. Within this framework, we can consider the viscous strain rate in the crust $\dot{\gamma}$ due to a differential stress σ associated with reservoir overpressurisation

223
$$\dot{\gamma} = A\sigma^n e^{-(Q+PV_a)/(R_g T)} \tag{3}$$

where *T* is temperature, *P* is the mechanical pressure in the wall rocks, *Q* is the activation energy, V_a is activation volume, R_g is the gas constant, and *A* depends on the mineralogy of the wall rock, the minerals' grain size, and water fugacity¹⁵⁰. Under cold upper crustal conditions and mm-size grains, the stress exponent n>2 is appropriate, which indicates flow in dislocation creep regime¹⁵⁰. With increasing strain rates, brittle failure becomes more likely, and we can recast the criterion for magma reservoir failure and diking in terms of a critical strain rate $\dot{\gamma}_{crit}$ associated with a critical differential stress σ_{crit} corresponding to the critical magma overpressure ΔP_{crit} :

232
$$\dot{\gamma}_{crit} = \frac{\sigma_{crit}}{\eta_r} \sim \frac{\Delta P_{crit}}{\eta_r} \equiv \frac{1}{\tau_{relax}}$$
(4)

 η_r is the effective host-rock viscosity around a magma chamber, and we can see the critical 233 234 strain rate for diking increases as the host-rock viscosity decreases. This suggests, for 235 example, that magma chamber failure is easier at shallower depths where the ambient host-236 rock temperature is colder and η_r is greater. The specific relationship between the critical 237 stress σ_{crit} and critical overpressure ΔP_{crit} will be a function of the host rock properties such 238 as temperature, pressure and mineralogy, as well as the magma reservoir properties such as 239 the size and shape of the magma body, and the presence or absence of pre-existing fractures as discussed above. In this formulation, the critical strain rate $\dot{\gamma}_{crit}$ is the inverse of the 240 viscous relaxation timescale τ_{relax} as defined in Ref.¹⁵¹. Refs.^{100,152–154} use the ratio $\eta_r/$ 241 242 ΔP_{crit} to estimate τ_{relax} for magma chambers at various depths, using a calculation of η_r 243 based on the steady-state temperature profile around a magma chamber and assuming that ΔP_{crit} has to be at least that required to propagate pre-existing fractures (Equation 2). For 244 example, if we apply the calculation of η_r from Ref.¹⁵¹, we find that a depth range between 245 $\sim 10-6$ km corresponds to η_r of $\sim 10^{19}-10^{21}$ Pa·s for the crust around a magma chamber at 246 850 °C. Using ΔP_{crit} of 20 MPa, $\dot{\gamma}_{crit} \sim \frac{1}{\tau_{relax}}$ ranges from 60 myr⁻¹ – 0.6 myr⁻¹. In the next 247 248 section, we review processes that give rise to magma overpressure and strain in the wall 249 rocks, and we evaluate how they compare to $\dot{\gamma}_{crit}$.

250

251 **3. Magma pressurization/eruption triggering**

3.1 Internal triggers

253 **3**.1.1 Magma recharge

254 Geophysical, geochemical, and petrological evidence support the hypothesis that 255 magma injection from deeper sources to shallow crustal chambers can lead to volcanic eruptions (e.g. Refs.^{44,86,111}). For example, magma injection has been invoked as a trigger for 256 the eruption of Pinatubi in 1991⁶⁷ and the 3.6 ka Bronze Age eruption of Santorini¹⁵⁵. The 257 258 accumulation of magma in crustal reservoirs can cause ground inflation that can be detected geodetically^{38,156,157}. Ground inflation usually precedes volcanic eruptions, although in some 259 260 cases eruptions have occurred without any preceding measured ground deformation, and 261 episodes of inflation often occur in active volcanic settings without culminating in eruption¹⁵⁸. In some cases, uplift over longer (1-10 kyr) timescales is recorded in geomorphic 262 features (e.g. Refs.^{159,160}) indicating that magma recharge and accumulation may occur 263 264 episodically over thousands of years, incrementally increasing the magma reservoir volume 265 and potentially the overpressure. The addition of new magma or volatiles to a shallow 266 reservoir also may stimulate gas emissions and increase surface heat flow, another sign of potential unrest^{161,162}. In the rock record, the interaction of hotter mafic magmas with resident 267 268 silicic magmas may be recorded in plutonic settings as microgranular enclaves and magma mingling structures that form in layered intrusions^{120,163–165}, as well as in the geochemistry 269 270 and textures of volcanic products (Fig. 1b). Crystals record chemical variations from core to 271 rim that can be deconvolved to identify the addition of more mafic and hotter magma into a colder, more silicic reservoir^{166–169}. Additional information about the timescales of recharge 272 273 and mixing events can be extracted from chemical profiles in minerals providing the

opportunity to link pre-eruptive monitoring signals to the processes that ultimately led to an
 eruption^{76,82,170}.

276 Physical models can provide a theoretical framework to determine the conditions under which magma recharge may trigger an eruption^{102,136,151,171–173}. Whether or not magma 277 278 recharge leads to the critical overpressure required for reservoir failure depends on the 279 magnitude, rate, and style of magma recharge, the current pressure in the reservoir, magma 280 compressibility (Fig. 2d, e), and the rate at which pressure may be relaxed by viscous or 281 plastic deformation of the wall rocks. If the host rock is purely elastic, i.e. if the magma 282 recharge occurs faster than the timescale for viscous deformation, the increase in pressure in the reservoir ΔP due to an increase in magma volume ΔV is approximately $\Delta P = \frac{\Delta V}{V} \beta^{-1}$ (Box 283 1), where β is the sum of magma and reservoir compressibility¹⁴⁸. Hence, in a purely elastic 284 285 crust, the potential for recharge to trigger an eruption decreases for increasing reservoir volume and compressibility¹⁵³. Over timescales approaching and exceeding the timescale for 286 287 viscous deformation of the crust (Eq. 4), pressure can be relaxed by deformation of the wall rocks (e.g. Ref.¹⁴⁷). To explore whether magma recharge can build up critical overpressure 288 over these longer timescales, we can considered the strain rate $\dot{\gamma}_{in}$ associated with the rate of 289 magma injection \dot{V}_{in} ^{102,103,136,151} 290

291
$$\dot{\gamma}_{in} = \frac{\dot{v}_{in}}{v} \equiv \frac{1}{\tau_{in}}$$
(5)

We note that $\dot{\gamma}_{in}$ is the inverse of the injection timescale τ_{in} of Ref.¹⁵¹. Thus, on timescales over which viscous relaxation is expected to be important, recharge may trigger the release of magma from a reservoir if $\dot{\gamma}_{in} > \dot{\gamma}_{crit}$ (Eq. 4).

To estimate $\dot{\gamma}_{in}$ for natural systems we need constraints on the magma injection rate and the total volume of the magma chamber being recharged, neither of which can be measured directly. Eruptive rates are commonly used as a proxy for magma supply rates (e.g.,

Refs.^{81,100}), although this requires making assumptions about the ratio of magma intruded to 298 erupted^{35,81}. The volume of a magma chamber can be estimated from geophysical imaging 299 such as gravity or magnetotellurics^{174,175}, or from eruptive volumes, as models for effusive 300 eruptions suggest that the volume erupted scales as $\sim 1 - 10\%$ of the chamber volume^{27,176}. 301 302 One exceptional example where both the magma chamber volume and supply rate have been 303 relatively well constrained is Laguna del Maule in the southern Andes. Based on the 304 combination of eruptive flux and inflation of the volcano over the Holocene, the magma supply rate is estimated at ~0.0023 km³/yr¹⁵⁹. A gravity survey by Ref.¹⁷⁴ indicated a melt-305 rich magma body $\sim 30 \text{ km}^3$ in volume located under the caldera. Together, this leads to 306 $\dot{\gamma}_{in} \sim 76 \text{ myr}^{-1}$, which is faster than our estimates of $\dot{\gamma}_{crit}$ in Section 2.3, which suggests that 307 308 the Holocene phase of eruptive activity at Laguna del Maule likely was triggered by magma 309 injection to a shallow crustal chamber.

310

311 3.1.2 The role of volatile exsolution

Magma cooling and crystallisation can increase the concentration of magmatic H₂O 312 313 and CO_2 in the residual melt, which may eventually trigger the exsolution of a low-density, relatively compressible magmatic volatile phase e.g. Ref.⁹⁰. This process, commonly referred 314 315 to as "second boiling," can increase pressure in a magma reservoir and has been invoked as a potential eruption trigger (e.g. Calbuco 2015¹⁷⁷ and Kelud 2014¹⁷⁸ eruptions). Another 316 mechanism that may stimulate volatile exsolution is the "flushing" or addition and chemical 317 interaction of deeply sourced magmatic fluids with magma stored at shallower depths⁸⁹. 318 319 Erupted products might provide sufficient information to identify whether crystallization-320 induced degassing or volatile flushing ultimately initiated the propagation of magma toward 321 the surface. Melt inclusions are used to trace the evolution of volatiles in magmatic systems 322 and can in principle be used to discern crystallisation-induced degassing and

flushing^{89,93,179,180}; however, potential issues linked to post-entrapment processes should be 323 carefully considered before applying this method^{179,181,182}. Apatite is a promising proxy to 324 325 trace the processes responsible for the presence of excess fluids and their chemistry in magmatic systems^{52,71,183,184}. While both crystallisation-induced degassing and flushing are 326 327 associated with magma crystallisation, in the first case the water activity in the melt 328 increases, and in the second it decreases, which implies that the phase equilibria and the chemical evolution of the residual melt fraction will differ (e.g. Refs.^{58,59}). Thus, these two 329 330 processes could be discerned also using major and trace element chemistry of major mineral 331 phases.

332 Geophysical observations of volatile exsolution may be cryptic because the high compressibility of the volatile phase may suppress surface deformation^{43,99}. In addition, 333 334 because the rate of volatile exsolution is linked to magma's cooling rate, this process may 335 result in the increase of magma volume over timescales that are longer than the viscous 336 relaxation of the crust and therefore do not generate overpressure (Fig. 4). Furthermore, even 337 if overpressure builds up, the rate of increase may be too slow to be detected without monitoring over extended periods of time^{171,185}. The rate of volumetric increase associated 338 339 with flushing depends on the rate of fluid supply and thus may be faster than second boiling. 340 On the other hand, the decrease of H_2O activity caused by the increase of CO_2 activity forces 341 crystallisation, which increases magma viscosity and hinder the ability of magma to 342 propagate to the surface and erupt⁸⁹.

To assess whether volatile exsolution can initiate magma chamber failure, we can compare the strain rate imposed to the wall rocks by exsolution $\dot{\gamma}_{ve}$ to the critical strain rate for reservoir failure, $\dot{\gamma}_{crit}$ (Eq. 4). If volatile exsolution is caused by flushing of CO₂-rich fluids from depth, $\dot{\gamma}_{ve}$ is function of the supply rate of CO₂⁸⁹ relative to the size and volatile 347 content of the resident magma chamber. If volatile exsolution is caused by crystallization and 348 second boiling, $\dot{\gamma}_{ve}$ is linked to the cooling timescale of a magma reservoir¹⁵¹:

349
$$\dot{\gamma}_{cool} \sim \frac{1}{\tau_{cool}} \equiv \frac{\kappa}{V^{2/3}} \tag{6}$$

350 where κ is the thermal diffusivity of the host crust. The faster a volatile-saturated magma 351 cools, the faster volatiles exsolve and potentially pressurize a magma chamber. On the other hand, rapid cooling could lead to thermal death of a magma reservoir^{102,136,151}. In Figure 4a 352 we show an example of model results from Ref.¹⁵¹, who consider the combined effects of 353 354 magma recharge, crystallization and volatile exsolution, and viscoelastic behaviour of the 355 crust. We frame their results in terms of the strain rates due to magma injection ($\dot{\gamma}_{in}$) and cooling/second boiling ($\dot{\gamma}_{cool}$) and how these compare to the critical strain rate for magma 356 reservoir failure ($\dot{\gamma}_{crit}$) using a critical overpressure of $\Delta P_{crit} = 20$ MPa and effective crust 357 viscosity η_r of ~10¹⁹ Pa·s. 358

From the thermo-mechanical model of Ref.¹⁵¹, we can compare the efficiency of 359 360 magma recharge and second boiling as eruption triggers. When magma cooling dominates the strain rate ($\dot{\gamma}_{cool} > \dot{\gamma}_{in}$ and $\dot{\gamma}_{cool} > \dot{\gamma}_{crit}$), eruptions can be triggered by second boiling for 361 362 volatile-saturated magmas. In addition, the fast cooling rate leads to a smaller number of 363 eruptions that can occur before the magma reservoir freezes to some rheological lockup 364 threshold⁶. This cooling-dominated regime would be favoured in smaller reservoirs subjected 365 to lower recharge rates and embedded in relatively cold and more elastic crust (smaller $\dot{\gamma}_{crit}$). 366 These conditions may prevail in immature systems that have not yet had the time and magma 367 input to build up a large plumbing system in a warmer crust. Magma reservoirs that feed 368 polygenetic volcanoes are likely already large enough that volatile exsolution is not an 369 effective eruption trigger, or at least not as effective as magma injection. Using Laguna del Maule as an example, the size of the current subcaldera magma reservoir of $\sim 30 \text{ km}^3$ implies 370 $\dot{\gamma}_{cool}$ ~3 myr⁻¹, which is only fast enough to compete with $\dot{\gamma}_{crit}$ for the coldest/shallowest 371

372 range of crustal conditions (~6 km depth or less). When magma recharge dominates the strain rate $(\dot{\gamma}_{in} > \dot{\gamma}_{cool} \text{ and } \dot{\gamma}_{in} > \dot{\gamma}_{crit})$, many more eruptions can occur before the reservoir freezes 373 374 (Fig. 4a), which suggests that eruptions are more likely triggered by magma recharge than 375 second boiling. As of yet, we are not aware of natural systems that can definitively be 376 categorized in the cooling- and exsolution-dominated regime. While this may partly be due to 377 a bias of studies focused on larger and more active volcanoes, another hypothesis is that 378 systems in the cooling regime are less likely to contribute to the volcanic record because they 379 are short-lived.

As magma reservoirs grow, both the cooling- and recharge-induced strain rate decrease, the wall rocks become hotter and less viscous (BOX 1), and the compressibility of magma within the reservoir increases (Fig. 2). All together, these factors may eventually make eruptions more difficult to trigger, which implies that magma accumulation and reservoir growth is favoured in mature plumbing systems if they are still being fed by magma injections (Fig. 4).

386

387 3. 2 External eruption triggers

If a reservoir is already at pressure conditions close to those required for failure, small stress changes produced externally to the volcanic system may be sufficient to initiate magma ascent. As discussed in the previous section, under crustal conditions with deviatoric stress (from the lithostatic), fractures may not necessarily be purely tensile (Ref.⁹) so that shear strength can regulate the initiation of fractures. Crustal rocks usually have pre-existing cracks, so frictional strength is a lower limit for shear strength¹⁸⁶. The summation of frictional strength and cohesion σ_0 is known as Coulomb failure stress

$$|\sigma| = \sigma_0 + f_s(\sigma_n - p_w) \tag{7}$$

396 where σ_n is normal force acting on the fault surface, f_s is the friction coefficient.

397 External triggering mechanisms may act by changing the stress field and the strength 398 of the host rock. Several phenomena can affect the strength: volcanic gas emissions associated with earthquakes may lubricate pre-existing cracks¹⁸⁷; hot and acid volcanic gases 399 sourced from magma may alter the host rocks around dikes, which lose strength^{188,189} or are 400 transformed in clay minerals that have lower friction coefficient^{190,191}. Such modifications are 401 402 important as even a reduction of the friction coefficient by 0.1 lowers the strength on the 403 order of 1 MPa (Fig. 3). We thus consider that, in some cases, small external stress perturbations may reduce the strength of the surrounding rocks together with pore pressure¹⁹². 404

405

406 *3.2.1 Loading or unloading*

407 A variety of surface loading processes can produce stress perturbations sufficient to 408 trigger the release of magma from reservoirs at critical conditions. Climate change causes variations in the gravitational load of glaciers and ice sheets on land¹⁹³⁻¹⁹⁶ and water masses 409 at sea¹⁹⁷⁻¹⁹⁹. The correspondence between Milankovitch cycles and patterns of global 410 volcanic activity suggests a link between climate change and volcanic eruptions²⁰⁰. Large-411 scale deglaciation can increase mantle melting at great depths (>50 km; Refs.^{195,201,202}), while 412 413 smaller-scale deglaciation can modulate lithospheric stress and promote dike formation^{196,203,204}. Increased erosion associated with deglaciation can further enhance these 414 effects¹⁹⁹. Whether deglaciation encourages or discourages eruptions depends on the 415 416 geometry of the surface load redistribution and the initial location and orientation of the dikes²⁰⁵. Similarly to glaciers, variations of surface loading associated with sea-level 417 variations can also influence magma productivity 206-208. 418

The gravitational forces exerted by the sun and moon modulate stress in the Earth's crust on a variety of timescale from diurnal to seasonal, and although the stress perturbations associated with tides are small (~1 kPa), the hypothesis that tides can trigger eruptions dates

back almost a century^{19,209-214}. Small stress perturbations could unclog pre-existing cracks 422 423 and mobilise bubbles in a low-viscosity magma, thus modulating seismicity, outgassing, and potentially stimulating unrest^{215,216}. Observations from persistently degassing volcanoes have 424 shown an association especially between fortnight tides and degassing^{217–220}, although this 425 correlation is not seen at every volcano^{212,221}, and even the same volcanic system may not 426 427 show a consistent sensitivity to tides^{215,222}. Additionally, the period of diurnal tides is close to 428 that of daily fluctuation of atmospheric temperature and pressure, which for some volcanoes correlate with volcanic activity²²³ and make the direct association between volcanic activity 429 430 and tides less clear. However, for longer time scales, a statistically significant correlations is observed between seasonal variations of sea level and volcanic eruptions^{224,225}. 431

432 Volcanic activity itself can be a source of unloading. The climactic eruption of Mount 433 St. Helens, 1980, was preceded by unloading of the summit area. A magma intrusion bulged 434 the north flank causing its failure, which decompressed the shallow gaseous magma, triggering a laterally directed blast and Plinian column of volcanic ash^{16,226-228}. Strength 435 reduction of previously shattered dome rock due to crypto dome intrusion also might provoke 436 collapse²²⁹. Similarly, a sector collapse event at Anak Krakatau volcano in 2018, which 437 438 sourced a deadly tsunami, marked the onset of elevated volcanic activity, increased SO_2 emissions, and local earthquakes¹⁷. Degassing during quiescence can also cause unloading as 439 440 the release of large amounts of gas from magma in the shallowest portions of the plumbing 441 system may increase the pressure difference between the shallow and deep magma reservoirs triggering magma ascent from depth²³⁰. Such a top-down triggering mechanism is reflected 442 443 by seismicity starting shallower than the estimated depth of the magma reservoir and migrating deeper 231 . 444

445

447 Historically, some volcanoes have erupted after large earthquakes, and a causal relationship between earthquakes and eruptions was first proposed 50 years ago^{232} . Statistical 448 449 analysis shows a significant increase of the likelihood of eruption in a period of days to years following an earthquake^{233,234}. Large eruptions rarely occur immediately after large 450 451 earthquakes, but the sudden onset of volcanic unrest after the 1992 Landers earthquake at distances as great as 1200 km from the mainshock epicentre²³⁵ renewed discussion on the 452 mechanisms that might link earthquakes, volcanic unrest and eruptions²³⁶. Changes of 453 454 volcanic monitoring parameters occurring immediately after large earthquakes are increasingly reported, including enhanced seismicity^{237,238}, deformation by dike 455 intrusion^{239,240} and subsidence^{241,242}. Following earthquakes, volcanic degassing is enhanced 456 457 in more mafic, open system volcanoes but tends to decrease in closed system volcanoes erupting more chemically evolved magmas¹⁸⁷. This suggests that magma properties and the 458 459 orientation of faults around a magma reservoir affect the response of volcanic systems to earthquakes^{187,239,240,243}. Both volcano types including magma viscosity and characteristics of 460 stress perturbation induced by earthquakes affect triggering efficiency¹⁵. The hydrothermal 461 system is more sensitive to seismic triggering¹⁵. Importantly, not all volcanoes react to large 462 earthquakes²⁴⁴, suggesting that only volcanic systems that were already in a critical state react 463 to earthquakes^{234,245,246}. 464

A fault rupture causes static and dynamic stress changes. The static stress is associated with permanent deformation of the crustal rock, and the dynamic stress change is caused by radiation of the seismic waves. Static stress quickly decays outside the near field²⁴⁷, while dynamic stress further traveling as seismic waves^{248,249}. These stress perturbations are small, but if the host rock is close to failure, they can help to initiate crack propagation by exceeding the Coulomb failure stress²⁵⁰ (Eq. 7). Static stress change may modulate the permeability of the host rock, enhancing the magma ascent²⁵¹. The surface

topography and its resonance can amplify the dynamic stress change²⁵². Dynamic stress 472 473 change can force otherwise static bubbles to ascend and coalesce, resulting in shear deformation, or sloshing^{239,253–255}. The enhanced gas mobility could stimulate volcanic 474 475 activity in a variety of ways. High-temperature gas could change the friction coefficient along open fractures surrounding the reservoir (Eq. 7) 256 , and ease dike propagation. As an 476 477 example, the enhanced extrusion rate observed in the 2006 eruption of Merapi Volcano, Indonesia, following the M_w 6.4 earthquake 50 km to the south^{257,258}, has interpreted to 478 479 results from the enhanced circulation in the crustal rocks of the CO_2 produced by the decarbonation of the limestone bedrock^{94,259,260}. 480

481 While large distant earthquakes have been found to be weakly, but significantly 482 correlated with volcanic unrest and more rarely eruption, the effect of smaller local 483 earthquakes is less clear. As discussed later, propagating dikes often induce seismic swarms whose cumulative seismic moment is correlated with the dike volume²⁶¹. Thus, large dikes 484 485 may induce large earthquakes. The 2000 Miyakejima intrusion, for example, induced six magnitude M >6 earthquakes (e.g. Ref.²⁶²). The seismicity induced by magma propagation 486 487 itself has been shown both on the base of observations and numerical models to tend to arrest dikes, rather than promote their further propagation^{23,263}, and thus could contribute to prevent 488 489 an eruption. The same has been found for diffuse seismicity, as it relieves elastic energy and 490 increases the effective fracture toughness of rock, making it harder for dikes to reach the surface²⁶⁴. Thus, local earthquakes may decrease or increase the likelihood of eruptions 491 492 depending on the specific case.

493 *3.2.3 Rainfall*

The strength of the host rock regulates gas and magma transport and depends on pore pressure (Equations 1 and 7), suggesting that rainfall can plausibly influence volcanic 496 activities by changing the pore pressure. As an example, pore pressure perturbations of the 497 order of 0.01-0.1 MPa can cause some earthquakes²⁶⁵⁻²⁶⁷.

498 Indeed, enhanced volcanic activities after heavy rain have been reported for basaltic magma^{268,269}. Eruption durations and explosivity at volcanoes such as Stromboli (Italy) has 499 been observed to increase after rainstorms²⁷⁰. Additionally, the record-breaking levels of 500 501 rainfall in early 2018 have been suggested to have facilitated the creation of a pathways for 502 magma ascent of the 2018 rift eruption at Kilauea Volcano, which devastated the south-503 eastern part of the island^{269,271}. Heavy rain events have also been suggested to have influenced seismicity rates at Mt. Merapi volcano²⁷² and Soufrière Hills Volcano, 504 Montserrat²⁷³, contributing to further destabilize their domes. Heavy rainfall also contributes 505 to the hydrothermal pressurisation of domes¹⁸, which has been deemed responsible for the 506 collapses of domes at Soufrière Hills Volcano, Montserrat^{274,275}, Unzen, Japan²⁷⁶, Merapi 507 Volcano, Central Java, Indonesia²⁷⁷, and Mount St. Helens²⁷⁸ and for phreatic explosions²⁷⁹. 508 509 The hydrothermal alteration also weakens the minerals to promote dome collapse 280 . 510 Unloading by dome collapse decompresses the dome-core or shallow conduit lava, causing explosion and promoting further dome growth^{275,281}. 511

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513

3 4. **Magma propagation to the surface**

514 Considering a scenario in which a reservoir containing eruptible magma has been pressurised 515 to critical values, for an eruption to occur, magma has still to ascend for several kilometers 516 before reaching the Earth's surface. In the following we will discuss the factors that can 517 facilitate the ascent of magma to the surface and those that act to arrest magma propagation 518 leading to an aborted eruption for both closed systems and frequently erupting volcanoes.

519

520 4.1 Magma ascent through conduits

521 Considering the presence of a previously established conduit to the surface, with no 522 mechanical obstacles to eruption such as a plug of lava or layers of solid rock, the likelihood 523 of the magma to erupt at the surface will be mostly determined by the properties of the 524 magma. In particular the variations of viscosity and density upon ascent and decompression 525 will determine the fate of the rising magma. Magma composition and initial volatile content will dictate the ascent dynamics, volume expansion and eruptive style (e.g., Ref.²⁸²). 526 527 Parameters describing the host rock response to pressurization, such as conduit geometry, 528 rock elastic parameters and coupling to a draining magma reservoir will contribute less to the 529 decompression rate. Exceptions to this rationale are "thin", compressible conduit geometries: 530 the thinner the conduit, the larger the role played by elasticity in regulating magma pressure (e.g., Ref.²⁸³). These dynamics have been described in many models that have progressed 531 532 much in recent years and are now reaching the stage where they can be used to interpret 533 quantitatively a variety of field observations such as the distribution of ejecta, lithic 534 fragments, crustal deformation, and link them to conduit and magma parameters (e.g., Refs.^{284,285}). 535

While many eruptions occur by magma flowing along pre-existing conduits, the lithostatic pressure increases with depth so that below ~1 km depth such conduits will rapidly collapse once drained. Moreover, even in "open conduit" volcanoes such as Etna or Stromboli the distribution of active vents changes dramatically over time scales of years or even months^{286–288}. These considerations and observations demonstrate that the existence of pathways persistently used for degassing does not guarantee that magma will use them rather than opening new ones.

543

544 *4.2 Magma propagation by diking: driving forces*

What pathway magma will take to erupt is not only a central problem for hazard assessment²⁸⁹, but also ultimately determines if magma will actually be able to reach the surface and erupt. The details of the path geometry, together with magma properties such as viscosity and density, dictate in many non-intuitive ways what could abort an eruption while magma is already on its way to the surface. Thus, the identification of the pathway the magma will take during ascent and the quantification of the stresses the magma pocket will experience are essential to determine if magma will eventually reach the surface.

552 Magma propagates through brittle rock by diking, a mechanism similar to hydraulic fracturing^{8,9}. Dikes get arrested when the energy released during propagation is less than the 553 energy required to create new fracture surface for the dike to advance²⁹⁰, or, equivalently, 554 555 when the stress intensity at their tip, K, becomes smaller than the rock's fracture toughness, K_c^{291} . K is determined by the combined contribution of internal and external stresses or 556 557 pressures and their variations along the dike plane, which is responsible for shaping the dike 558 and its tip. Pressure is sometimes provided by hydraulic connection to a magma chamber, but 559 dikes can also achieve enough pressure at their tip by being subject to "pressure gradients" 560 (stress difference between dike tip and dike tail, over the length of the inflated region of the dike)²⁹². That is because stresses may vary along the dike in a way to squeeze its tail and 561 562 inflate its nose enough to achieve $K > K_c$, forcing dike propagation. This illustrates how the 563 problem of "premature" dike arrest can be formulated in terms of dike size and the total stress 564 gradient acting on the dike, that we indicate as $\Delta \gamma$. Both magma properties and external 565 factors contribute to the total gradient $\Delta \gamma$ affecting a dike. One important such gradient 566 originates from the difference between lithostatic and "magma-static" pressure along the dike, which varies as $\Delta \rho g \cos \theta$ (often called "buoyancy" pressure), where $\Delta \rho = \rho_r - \rho_m$ is 567 the difference between host rock density, ρ_r , and magma density, ρ_m , g is the acceleration 568 569 due to gravity, and θ is the dike's dip angle. In contrast, the viscous dissipation due to magma flow tends to swell more the dike tail than its nose, and to slow down the dike. Additional contributions need to be quantified case by case, and include stresses arising from uneven overburden load distributions²⁹³, differential stress accumulation linked to host rock temperature gradients (e.g., Ref.²⁹⁴), regional stress gradients, previous intrusions and earthquakes²⁸⁹.

575

576 4.3 Critical magma volumes for dike propagation

577 Until recently, analytical equations for the 'critical volume for propagation', V_c , of 578 buoyancy-driven dikes filled with inviscid magma were only available in two 579 dimensions^{295,296}. Ref.²⁸ extended the model to three dimensions considering only the 580 buoyancy gradient, but it is straightforward to rewrite the equation so to account for the total 581 stress gradient acting on the dike:

582
$$V_c = 0.75 \frac{(1-\nu)}{16\mu} \left(\frac{9\pi^4 K_c^8}{\Delta \gamma^5}\right)^{1/3}$$
 (8)

583 where ν is Poisson's ratio, μ is shear modulus, K_c is the rock fracture toughness and $\Delta \gamma$ is the 584 total driving pressure gradient.

Since $V > V_c$ is required for a dike to carry on propagating, it follows that 585 propagation-hindering processes are those that either tend to decrease V, or, alternatively, 586 increase V_c (e.g. by increasing K_c or decreasing $\Delta \gamma$) during propagation. Dikes leave some 587 588 magma behind when they propagate because they cannot pinch perfectly closed at their back. 589 The higher the magma viscosity, the thicker their tail and the more abundant the magma left on the way during propagation^{8,22}, which decreases the dike ability to move further. 590 Propagation-hindering processes that work by increasing K_c include seismicity, plasticity, 591 faulting^{23,297} and approaching a more competent layer²⁹⁸. Factors contributing to decrease $\Delta \gamma$ 592 593 include the increase of viscosity, resulting from decompression-induced crystallisation or cooling⁹, or a transition from vertical propagation to lateral when approaching a strong load 594

(e.g., Ref.²⁹⁹). The dike's dip angle θ is a rarely discussed, but important, factor for dike 595 596 arrest. A horizontal dike or sill (θ =0) will lack pressure due "buoyancy" and likely stall, 597 while shallow dipping dikes will require large volumes to propagate. Since uneven surface 598 loads have a large effect on the orientation of the principal stresses and thus on the dip of 599 dikes, a closer look at the shape of volcanic edifices will offer more clues on the chances of 600 dykes to reach the surface and feed an eruption. All this can be compensated, at least in part, 601 by other processes enhancing V, such as volatile exsolution and vesiculation or the dike 602 approaching the free surface 300 .

603

604 *4.4 The pathway of dikes*

605 It is often assumed that magma is channeled by pre-existing weaknesses, such as 606 faults or fractures. However, the orientation of most faults is optimised for shear rather than for opening³⁰¹, so that in most cases magma emplacement through faults requires more work 607 608 than opening a new path in a more convenient direction. This is why the vast majority of dikes fail to occupy pre-existing faults and create their own pathways (e.g., Refs.^{133,134,302}). 609 Seminal fieldwork and theoretical studies on the stress controls on dike propagation^{124,130} has 610 611 shown that dikes tend to open perpendicular to the least compressive stress axis, σ_3 . Such 612 "least resistance to opening" pathways are accurately determined by calculating the elastic energy released during propagation^{303–305}. Provided abundant magma pathways are observed 613 614 in an area, an accurate model of the stress field can be calibrated, allowing to forecast future 615 dike pathways²⁸⁹. Non-flat topography, heterogeneities of elastic parameters, land 616 movements, active faults, pressurized reservoirs, high pore pressure in hydrothermal systems, 617 previous intrusions may all contribute to stress heterogeneities and complex rotations of the principal stresses^{289,306}. Eruptive fissures' patterns on volcanoes are often attributed mainly to 618 stresses due to the pressurization of a magma reservoir of appropriate shape^{126,307}). However, 619

620 Ref.²⁵ demonstrated that stresses from the growth of a volcanic edifice together with regional 621 stresses are often much larger than the stresses induced by pressurization of a magma 622 reservoir, and control the curvature of dikes in the field.

Many studies have confirmed that the shape of the volcanic edifice both exerts a strong control on the orientation and dip of magma pathways and provides a driving force to propagation, thereby controlling the time scales of magma migration and storage below the volcanoes and the likelihood of dikes getting trapped or erupting.

627 Large volcanic edifices (stratocones and shield volcanoes) compress the underlying 628 rock, which results in both attracting dikes from offset magma reservoirs, and efficiently trapping them at depth^{264,299,303,304,308,309} (Fig. 5a); only dikes with a large buoyancy manage 629 to avert such trapping effect and erupt (e.g. Refs.^{303,310}). If magma manages to intrude into the 630 edifice, topographic load gradients drive the dikes radially away from the summit^{311–317} (Fig. 631 632 5b-c). The dikes may erupt or remain trapped due to the relation between magma density and the density profile of the host rock^{318,319}, or by inducing graben faulting (e.g., Refs.^{130,263,320}). 633 634 The propagation is usually accompanied by seismicity, which can also trigger dike arrest by 635 releasing elastic energy 264 .

636 Calderas are another example showing that the modulated stress field by topography determines the direction, and influences the rate, of magma propagation (e.g., Ref.^{289,321,322}). 637 638 Large-scale excavations such as a caldera cause a vertical σ_3 below the caldera floor, with 639 topographic load gradients from the caldera being surrounded by a rim trapping the dikes and causing them to accumulate as stacked sills (e.g. Ref.^{289,322}; Figs. 5d,e). Dikes may nucleate 640 641 as sub-horizontal intrusions and initially lack buoyancy, so that only large dikes may be able 642 to escape the stress trap. The gradual accumulation of caldera infill and growth of resurgent domes may change the stress balance over the caldera cycle³²³ and modify dike pathways and 643 vent patterns²⁸⁹. 644

In summary, a magma propagation perspective on storage regions is that they represent a "bottleneck" where stresses slow down propagation or entirely trap the magma. If dikes achieve to escape from this trap, they still have great chances to get arrested on their way in a number of ways that can only be evaluated by combining concepts from petrology, structural geology and geophysics.

The issue with these concepts is that elastic stresses are notoriously very difficult to 650 both measure directly and model (e.g., Ref.³²⁴). Elastic stresses result from many overlapping 651 652 factors, some of which vary at the time scale of monitoring (e.g. magma reservoir stresses) 653 and can be "sensed" through the deformation they are linked to, while others vary on much 654 longer time scales (e.g. topographic loading) and "act in the background". In order to 655 correctly model dike pathways we need to account for all stress-generating mechanisms, 656 including both those that are linked to "visible" and "invisible" deformation, keeping in mind 657 that "invisible" stresses might be the dominant ones, e.g. a 4-km-tall edifice generates ~ 100 658 MPa of compression on the underlying rock. The relative size of the individual contributions 659 is challenging to estimate as they all depend on distinct poorly constrained factors such as the 660 crustal profiles of rock density and rheology and various tectonic processes, to name just a 661 few, and thus are difficult to bring together in a well-calibrated model. At the same time, magma trajectories are very sensitive to the ratios of the relative contributions^{25,325}, as such 662 663 ratios determine the orientation of principal stress axes. This brings much confusion and 664 uncertainty to stress models and has so far hindered accurate forecasting of dike paths^{289,312,326}. 665

666

667 5. Summary and future perspectives

668 One of the main goals of volcanology is to anticipate the future behaviour of 669 volcanoes, an endeavour that requires a scientific understanding of the processes that lead to 670 the accumulation and transport of magma through the lithosphere, and the mechanisms that 671 trigger eruptions. Physical models for magma reservoir assembly and growth demonstrate 672 that the rate of magma supply to a reservoir is one of the key parameters that governs the rate 673 of accumulation of eruptible magma, the pressurization of magma reservoirs required to 674 initiate magma ascent to the surface, and the evolution of physical properties of both the 675 magma and the surrounding crust (Fig.1). In general, the rate of eruptible magma 676 accumulation in a reservoir increases with greater magma supply rates; however, greater 677 magma supply rates can also pressurize and destabilize reservoirs, leading to heat and mass 678 loss through magma withdrawal. As magma reservoirs grow, crystallizing magma becomes 679 volatile saturated and host rocks become warmer and weaker (Fig.4). This changes the 680 response of magma reservoirs to recharge, eruptions, and external perturbations such as 681 earthquakes and changes in surface loading, leading to slower pressurization but larger 682 volumes of magma withdrawal following reservoir failure. Similarly, the evolution of magma 683 properties may impact the ability of magma to ascend through dikes. As magmas become 684 more evolved and water-rich, the increased buoyancy could help to drive magmas to the 685 surface; however, the increased viscosity could counteract these effects. In addition, 686 structural changes to the volcano such as edifice growth or caldera collapse could alter the 687 external stress field and hence the pathway of ascending dikes, in many cases trapping dikes 688 in the shallow crust.

In light of the many factors that govern magma transport, storage, and eruption triggers, it becomes clear that in order to forecast future eruptions we need to be able to 1) characterize the current state of the magmatic system, including the distribution of magma volumes, pressure and temperature conditions, volatile content and saturation state; 2) characterize the "boundary conditions" that influence the magmatic system, such as the flux of magma from the mantle, and the rheology and stress field of the crust. Below we list the

- major outstanding challenges to this endeavour that have emerged from our review, alongwith some recommended avenues for future work to address these challenges.
- 697

698 1. The flux of primitive basaltic magmas from the mantle (and the proportion of 699 this flux that directly supplies crustal reservoirs) is perhaps the most 700 influential yet least constrained parameter that governs the growth and 701 evolution of crustal magmatic systems and the transport of magma to the 702 surface (Section 2.2, 3.1). Estimates based on eruptive volumes likely are 703 inaccurate unless the intrusive extrusive ratio is well constrained; however, the 704 intrusive:extrusive ratio likely is not a fixed parameter but varies as a function 705 of magma supply rate, the size of the magma reservoir, and the thermal maturity of the system.^{11,12,36,100} Geodetic monitoring of recharge events 706 cannot uniquely constrain the mass of magma intruded³²⁷, and the signal is 707 708 impacted by the presence of volatiles and viscoelastic response of the crust^{43,99}. Similarly, petrologic data, thermo-barometry and zircon 709 geochronology on erupted products provide snapshots of the magmatic 710 711 system, but only at discrete moments in time corresponding to past eruptions. 712 We suggest that although none of these datasets independently can constrain the mantle flux and reservoir recharge rates, we can tighten our estimates 713 714 through the joint inversion of these data with numerical modeling of the 715 coupled thermal, mechanical, and chemical evolution of the plumbing system (e.g., Refs.^{36,75,284}). 716

717
2. The distribution of magma in the subsurface beneath volcanoes is key to
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718
719
(Section 3) and places a lower bound on the potential size of an eruption, yet

720 our ability to "see" the magmatic system remains limited. The combination of 721 data on eruptive volumes, co-eruptive deformation, and volatile content from 722 melt inclusions may be used to place bounds on the size and depth of a reservoir feeding a particular eruption (e.g., Mount St. Helens 2004-2008²⁸⁴; 723 Kīlauea Volcano 2018³²⁸), but this does not necessarily provide information 724 725 about the distribution of magma throughout the rest of the crust, only the 726 reservoir(s) being tapped by that eruption. Advances in seismic tomography, 727 gravimetry, and magnetotellurics would be required to potentially resolve melt fractions at the resolutions needed to image melt-rich magma bodies³²⁹. 728

729 3. Compared to their small abundance by mass, volatiles play an outsized role in 730 both magma storage and transport by influencing magma properties such as 731 compressibility, density, and viscosity; however, we are usually not able to 732 directly measure the volatile content, saturation state, and distribution of 733 volatiles in present-day reservoirs or during magma transport. In addition, the 734 multiphase nature of volatile-saturated magmas makes it challenging to 735 understand the dynamics of volatiles, melt, and crystals in a reservoir and their 736 evolution over time, which has implications for both magma rheology 737 (influencing viscosity and compressibility) and host-rock rheology (outgassing 738 of hot, acid fluids can alter the strength and fracture distribution). More work 739 to develop physical models of volatile behavior in reservoirs and to link these 740 models to observations such as gas monitoring data may shed light on how 741 volatiles migrate through reservoirs and how we might use monitoring data to 742 constrain the current distribution of volatiles in subvolcanic reservoirs.

7434. The pathway of magmatic dikes, which ultimately dictates the fate of magma744 ascending from depth, can be highly challenging to understand and anticipate.

745 Dike propagation is a complex multiphysics process for which numerical 746 models are still oversimplified. Dikes are also extremely sensitive to both local- and regional-scale heterogeneities in crustal density and stress (e.g., 747 Ref.³³⁰), which are usually only crudely characterized at most 748 volcanoes^{289,293,310}, as well as the distribution of internal magma pressure. 749 750 which evolves as a function of dike growth, connection to a reservoir, and 751 internal magmatic processes such as volatile exsolution which are rarely considered in dike propagation models³³¹. Dike propagation cannot be directly 752 753 observed in real time, but only indirectly through geophysical monitoring data 754 or geologic studies of eroded systems. Improvements to multiphysics models 755 for diking could come from engaging with other scientific communities (e.g. 756 engineers) who work on hydraulic fractures, but the application of these 757 models to dikes needs to integrate observations from the complementary 758 perspectives of geophysics and geology.

759 5. Once a magmatic system reaches a critical state, the timing of the eruption can 760 be modulated by external factors, such as earthquakes, dome collapses, tides, 761 and rainfall. Whether such mechanisms actually trigger an eruption depends 762 on both the magma conditions and the characteristics of the external force. In 763 the case of earthquakes, for instance, these could include the peak ground velocity, frequency, and static stress change amplitude¹⁵. The strength of the 764 altered host rock also regulates the behavior of magma³³². Accumulation of 765 766 such knowledge would help to anticipate eruptions. Another challenge in 767 understanding both internal and external eruption triggers is in establishing a 768 clear causal link; our confidence in identifying a particular trigger increases if 769 the triggering event occurs close to the volcano, produces a large change in

stress, and occurs immediately prior to the eruption (e.g., the landslide
preceding the 1980 Mount St. Helens eruption¹⁶).

772 In summary, while we still need to understand many of the basic mechanisms and timescales 773 involved in the storage and transport of magmas, the scientific community is making rapid 774 progress. Simultaneously, new monitoring technology is being developed, instrumental networks are expanding, and global volcano databases are being established^{30,46,333}. In order 775 776 to take advantage of these developments to achieve effective eruption forecasting, we need to 777 increase our efforts to reshape our theoretical understanding into forecast models. This 778 requires merging simple (having a small number of independent parameters) deterministic 779 physical models with data-driven approaches and statistical methods to help us estimate "in 780 situ" rock and magma parameters that are crucial for determining how the system will evolve. 781 Initiatives for independent model testing, which is standard practice in many fields, such as in 782 seismic hazard analysis, may also help identifying the best strategies to move forward.

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786 Figure Captions

787 Figure 1: Schematic illustration of a volcanic plumbing system with selected elements 788 relevant for the assembly and trigger of volcanic eruptions. a) Fluids from the deepest 789 portion of the crust. The panel shows how the chemistry of the degassed fluids at depth 790 (expressed as H₂O and CO₂ molar fractions) change with magma cooling and crystallisation. 791 b) Calibrated elemental map showing the distribution of anorthite content in plagioclase (left 792 hand side) and aluminium (atoms per formula unit [apfu] - 23 oxygens) in clinopyroxene 793 and amphibole (right hand side). The white rectangle shows selected crystals of plagioclase 794 and amphibole with distinct chemical characteristics. These two groups of plagioclase and

amphiboles crystallised from magma of different chemical composition. c) illustration of the potential evolution with time of the wall rock properties showing the increase of fracturing and veins produced by hydrothermal activity. d) Chemistry of degassed fluids from a subvolcanic reservoir exposed to the flushing of increasing amounts of CO_2 .

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800 Figure 2: Relationships between temperature and magma properties and their temporal 801 evolution as calculated from thermal modelling. The simulations were performed 802 considering the periodic input of magma in the crust (initial temperature 400 °C) at a depth 803 corresponding to a confining pressure of 200 MPa. The injected magma is a H₂O-saturated (CO₂-free) andesite, with temperature-melt fraction determined experimentally by Ref.³³⁴. 804 805 The volume of excess fluids is calculated assuming full incompatibility in the crystallising minerals and ideal behaviour. Magma compressibility was calculated following Ref.¹⁰⁰ at 806 807 pressure of 200 MPa and considering an overpressure of 1MPa (the results are virtually 808 identical for overpressures of 10 MPa). We consider that excess fluids leave the system once 809 magma cools to solidus temperature. a) Variation of melt and excess fluid fraction, and 810 compressibility as function of temperature. b, c) Total volume of injected magma (black line), 811 reservoir (blue line) and eruptible magma (red line) as function of time for two rates of 812 magma input in the crust (expressed as vertical accretion rates). The coloured curves shows 813 spikes corresponding to the injection of a new sill into the system, while the black line is 814 smooth because it is calculated using the average rate of magma input. d, e) Average crystal 815 fraction, eruptible/injected magma volume and compressibility as function of time calculate 816 from the thermal modelling results presented in panel b and c, respectively.

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818 Box 1: Schematic illustration reporting physical properties of magma³³⁵ and wall 819 rock^{9,141,150} relevant for the pressurization of magma reservoirs. Processes leading to 820 variation of volume (ΔV), such as magma injection or volatile exsolution, can lead to the 821 pressurisation of the reservoir. The host rock surrounding the magma chamber can behave 822 both viscously and elastically. When strain rate produced by magma supply or volatile 823 exsolution is sufficiently slow, the host rock deforms viscously, which may inhibit nucleation 824 and propagation of a crack in the wall rock. In contrast, at high strain rates the wall rock 825 behave elastically and overpressure increases in the magma reservoir until it exceeds a 826 critical value for failure and cracks propagate into the wall rock, initiating magma transport. Typical magma input rates \dot{V}_{in} vary between 10⁻⁴ and 10⁻² km³/year (Refs.^{336,337}). Considering 827 volume of spherical magma reservoir (V) between 1 and 100 km³, the range of strain rate 828 produced by such rates of magma input can be calculated as $\dot{\gamma}_{in} = \frac{\dot{v}_{in}}{V} = 10^{-15} - 10^{-9} s^{-1}$ or 829 $0.03 - 30,000 \text{ myr}^{-1}$. 830

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832 Figure 3: A summary of stress and pressure scales. Continuous blue and pink curves show 833 the stress required for viscous deformation of wet quartz and wet plagioclase at a shear rate of 2.5x10⁻¹² s⁻¹ calculated by Equation 1 (Refs.^{338,339}). This shear rate corresponds to a magma 834 supply of 0.001 km³/year into a magma chamber with a radius of 1km. We also calculated the 835 836 required stress to deform quartz at other strain rates, as shown in the legend. We consider a 837 temperature gradient of 30 °C/km with a surface temperature of 20 °C. The thick blue curve 838 refers to a temperature of 500 °C higher than the geotherm and shows the effect of 839 temperature increase of the wall rock due to the presence of magma in the crust. Red and 840 black-gray curves provide the dry tensile failure condition and friction strength without 841 cohesion calculated by Equations 1 and 7, respectively, where f_s is friction coefficient. The 842 boxes of earthquake, rain, the arrow with the note of landslide and degassing, purple region 843 labelled glacier/sea level, and the green-dashed line labelled tide indicates the range of stress 844 perturbations caused by each respective phenomenon.
846 Figure 4: a) Regime diagram of the number of diking events ("eruptions") from a magma 847 chamber before the chamber freezes to 50% crystal volume fraction, based on the model of 848 Degruyter and Huber (2014). Number of eruptions is shown as a function of the strain rates due to magma injection $\dot{\gamma}_{in}$, cooling and volatile exsolution $\dot{\gamma}_{cool}$, and the critical strain rate 849 for diking to occur $\dot{\gamma}_{crit}$. In the upper left region, eruptions are triggered by second boiling. In 850 851 the upper right region, eruptions are triggered by magma injection. In the lower third of the 852 diagram, no eruptions occur because the critical strain rate is not met. Contours of number of 853 eruptions are shown by the coloured lines. The six triangles are examples of where systems 854 would plot in regime space for the hypothetical conditions shown in panel (b). b) 855 Hypothetical values for magma chamber volumes, crust viscosity and magma injection rates 856 over time. Initially we imagine a small chamber in a cold, higher-viscosity crust. As magma 857 injection rates increase, the crust warms up and viscosity drops, and eventually the magma 858 chamber grows to larger sizes. In the last example (triangle 6), magma injection wanes, the 859 crust starts to cool again (viscosity increases) and the chamber loses volume as it freezes.

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861 Figure 5: Shallow pathways of dikes from the geological record and fluid injection experiments in gelatine. a) Gravitational loading attracts and focuses deep dikes, before 862 causing their arrest at depth (from Ref.³⁰⁸, with permission). **b**) Gravitational loading causes 863 864 dikes within a stratocone or shield volcano to propagate radially away from the centre of the 865 edifice. Red bodies are injected dyed water, the yellow mass is solidified gelatine shaped to 866 model a gravitationally loaded volcanic edifice. Unpublished experiment. c) Map of Summer Coon volcanic center, showing radial dykes as black segments. From Ref.³⁴⁰, with 867 868 permission. d) Three-dimensional model of cone-sheets (blue ribbons) at the Ardnamurchan igneous complex projected on the basis of their surface expression. From Ref.³⁴¹, with 869

870	permission. e) Asymmetric excavation simulating a caldera or rift system. The black curves
871	are pathways of injected dyed water. Unpublished experiment.

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