Magma chamber formation by dike accretion and crustal melting: 2D thermal model with emphasis on zircon record

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Abstract

Rapid progress in investigation of zircon records for U-Th-Pb ages and O and Hf isotopes in igneous rocks require understanding how magma bodies are formed and evolve in the crust. We here present a 2D model of magma bodies formation in granitic crust by injection of rhyolitic or andesitic dikes and sills. We combine this model with our zircon crystallization/dissolution software and compute zircon survival histories in individual batches of magma and country rocks.

Simulations reproduces incremental accumulation of intruded magma into magma chambers generating eruptible and interconnected magma batches with melt fraction >50 vol% that form in clusters. The rate of melt production is highly variable in space and time. The volume of eruptible melt strongly depends on the input rates of magma Q and the width W of the dike injection region. For example, dikes injection with Q=0.25 m^3/s with W=500 m during 4 ka generate 20 km^3 of melt while no significant melt forms if W=4 km. Injection of andesitic dikes produces only slightly more melt than rhyolite to granite injection despite of much larger thermal input.

Due to rock melting most of zircons loose significant portion of their old cores and, thus, average age. Magmatic zircons in the periphery of the intrusion form very quickly while in its central part crystals contain old cores and young rims and can grow during several hundreds of ka. We observe diverse proportions of crustal melt/newly intruded magma, which translates into diverse O and Hf isotope distribution in zircons.

1 **Magma chamber formation by dike accretion and crustal melting: 2D thermal model**

2 **with emphasis on zircon record ***

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8 Paper is dedicated to late Alexander R. McBirney (1924-2019), renown volcanologist and a

9 founder of the Volcanology Program in the University of Oregon

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11 **Key Points:**

- A 2D model of magma chamber growth by dike injections shows that a vertically extended system of magma batches can be formed in a few ka

- Behavior of zircon crystals is linked to thermal evolution of magma-host rock system by diffusion growth model in > 100,000 points.

- Isotopically diverse magma batches are generated naturally in an interconnected network of eruptible magma bodies with >50% melt

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26 igneous rocks require understanding how magma bodies are formed and evolve in the crust. We

27 here present a 2D model of magma bodies formation in granitic crust by injection of rhyolitic or

28 andesitic dikes and sills. We combine this model with our zircon crystallization/dissolution

29 software and compute zircon survival histories in individual batches of magma and country

30 rocks.

31 Simulations reproduces incremental accumulation of intruded magma into magma chambers

32 generating eruptible and interconnected magma batches with melt fraction >50 vol% that form in

33 clusters. The rate of melt production is highly variable in space and time. The volume of

34 eruptible melt strongly depends on the input rates of magma Q and the width W of the dike

35 injection region. For example, dikes injection with Q=0.25 m^3/s with W=500 m during 4 ka

36 generate 20 km^3 of melt while no significant melt forms if W=4 km. Injection of andesitic dikes

37 produces only slightly more melt than rhyolite to granite injection despite of much larger thermal

38 input.

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43 O and Hf isotope distribution in zircons.

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45 **Plain Language Summary**

46 Magma bodies formation in the crust via its melting is important for understanding how the

47 continental crust is created and melted and for volcanic hazard assessment. In particular, it is

48 important to understand how much magma can be produced and stored in the crust and what is

49 the architecture of such magma systems. As numerical methods that investigate these phenomena

50   are developing on all scales, rapidly improving dating methods of mineral zircon and its isotopic

51   and chemical content serve the purpose to recognize magma processes and their timing in real

52   geologic systems. In this paper we present a new model of magma chamber formation based on

53   intrusion of dikes and sills and offer the 2D software to do that for user-specific situations. We

54   further combine this software that produces temperature-time histories in more than 100

55   thousand points with zircon crystallization software previously published by us. We are now able

56   to describe in detail magma chamber growth and formation, and the history of zircon dissolution

57   and growth.

58

1 Introduction

60   Magma bodies formation in the crust via its melting is important for understanding how the

61   continental crust is created and melted and for volcanic hazard assessment. In particular, it is

62   important to understand how much magma can be produced and stored in the crust under active

63   volcanoes and caldera systems, what is its state, and what is the architecture of such magma

64   systems. The main mechanism of magma transport in the Earth's crust is via formation of cracks

65   along which magma rises to the surface in the form of dikes and sills (Rubin, 1995). Basaltic

66   magmas typically rise from mantle depths of several tens of kilometers carrying heat that enables

67   a multitude of melting phenomena in the crust (Huppert and Sparks, 1988; Annen and Sparks,

68   2002; Annen, 2009; Dufek and Bergantz, 2005). A common place of their ponding revealed by

69   geological and isotopic arguments (DePaolo et al, 2019), and thermomechanical modeling is in

70   the lower crust or Moho (Colon et al. 2018) for a variety of tectonic environment ranging from

71   plume to subduction zones. Their differentiation products that include basaltic andesites and

72   andesites rise from these magmatic "hot zones" (e.g. Annen, 2009) in the lower crust and intrude

73   the middle and upper crust in continental arc systems, or areas in intracontinental magmatism.

74   Dike widths can vary from centimeters to tens of meters, and in horizontal extent - from meters

75   to up to several hundred kilometers (Krumbholz et al., 2014). Magma may flow in the dikes with

76   rates of meters per second and erupt many cubic kilometers of basalts in less than a year as is

77   evidenced by historically observed eruptions such as Laki (15 km^3, Thordarsson and Self, 2003)

78   and Bardarbunga (Gudmundsson et al. 2014). Estimated magma flux rates range from 0.0001

79   km^3 in monogenetic magma fields in subduction-related environments to 0.1-1 km^3/yr in hot-

80   spot volcanoes such as Hawaii (Robinson & Eakins, 2006), or during outbursts of magmatism

81   such as those happening in Large Igneous Provinces (Costa et al. 2008; Bryan and Ernst, 2007).

82   In nature, high rates of magma injection generate significant overpressure in the upper crustal

83   magma bodies, leading to volcanic eruptions (Caricchi et al., 2014). This gives a relatively

84   narrow window of the influx rates and supply durations that can generate large eruptible volume

85   of magma suitable for super eruptions. The volcanic/plutonic ratio is an important and poorly

86   constrained parameters in many magmatic systems; higher magma rates are thought to result in

87   greater proportion of erupted materials, while low magma flux rate results in formation of

88   plutons without a connection to the surface (e.g. Glazner et al. 2004).

89   Whether mantle-derived magma will erupt on the surface without much interaction with the

90   crust, or stall in the crust, differentiate and cause crustal melting is a century-old debate of

91   petrologists, geophysicists and modelers (Becerril et al. 2013; Edmonds et al., 2019 and

92   references therein). In particular, magma ascent in dikes is controlled by the buoyancy forces

93   and the tectonic stress field among the main factors (Simakin and Talbott, 2001; Townsend et al.

94   2017). Most dikes do not reach the surface but are blocked at the level of neutral buoyancy

95   (Walker, 1989), or with structural barriers in the form of stronger rock layers or melt zones

96   (Edmonds et al. 2019). Repeated introduction of dikes into the near-surface (first kilometers)

97   region of the Earth's crust leads to mechanical dike stalling and their solidification, or upon

98   repeated flow, formation of magma chambers with or without associated crustal melting

99   (McBirney, 2006). As is evidenced from Large Igneous Provinces or supervolcanic eruptions,

100 volumes of liquid magma can reach thousands of cubic kilometers, although usually the volumes

101 are much smaller (kilometers-tens of kilometers) and globally magmatic volumes vs frequency

102 follow typical log-linear relationships (Papale, 2018). Magma chambers can be detected with

103 some success by seismic tomography using shear wave attenuation (e.g. Heath et al., 2018).

104 They may have an irregular shape, but most often appear to be flattened bodies with vertical or

105 horizontal strike. The current paradigm is that there can be several chambers located at different

106 depths under active volcanoes or caldera systems (Elsworth et al., 2014; Cashman et al., 2017),

107 not very far from a textbook example of lower (hot zone) and upper crustal magma bodies in the

108 crust (McBirney, 2006).

109   The numerical modeling of formation of magma chambers by injection of magma in the crust

110   started with a pioneering work of Huppert and Sparks (1988) who used 1D heat transfer model

111   with a realistic phase diagrams. Further development of these models into pseudo-2D and

112   axisymmetric cylindrical coordinates (Annen et al. 2006, Annen, 2009, Dufek and Bergantz,

113   2005). Current modeling efforts proceed on all scales. For example, using a large scale

114   thermechanical program of Gerya and Yuen (2003) that takes into consideration fully

115   mechanically coupled crustal rheology, Colón et al. (2018, 2019) simulated crust-wide multi-

116   level magma system formation under the hot-spot environment of Yellowstone on million-year

117   timescales. In more detailed local models where penetration and heat transfer between individual

118   dikes and sills and host rocks are considered (Annen, 2009, Dufek & Bergantz, 2005) modeling

119   of individual magmatic outputs are possible. Models of the first type consider regions with a

120   characteristic size of tens of kilometers and a grid spacing of several hundred meters. They

121   cannot resolve the sub-grid heat exchange processes that occur during the real transport of

122   individual portions of magma, but they allow to estimate the size and position of magma

123   chambers based on the crust-upper mantle-wide distribution of temperatures, rheological

124   properties of rocks and stresses, as well as the distribution of magma between individual

125   chambers. In the models of the second type, the region into which magma is introduced is set in

126   advance based on the known geological structure of the crustal block, and available

127   geochronological estimates of the time of formation of magmatic bodies. An example of the

128   reconstruction of a real magmatic system of Jurassic Yerington batholith in Nevada, is presented

129   in (Schöpa et al., 2017). The model assumes horizontal sill emplacement with sinking of the

130   underlying rock layers to accommodate the added mass, as in Annen et al., (2006). The heat

131   equation is solved taking into account the latent heat of melting for the rocks and the real

132   temperature dependence of the concentration of crystals in rocks and magma. An explicit scheme

133   for solving the heat equation is used, which imposes a significant limitation on the time step and

134   the total computational time. Biggs and Annen (2019) considered merging of two magmatic

135   centers located at the same depth and corresponding ground deformation. They found that

136   closely spaced intrusions can develop combined viscoelastic shells over time scales of 10s ka and

137   form laterally extensive mush zones over time scales of 10–100 ka for the high melt supply rates

138   and deep chambers where thermal conditions are suitable. Melt separation and migration is not

139   considered in the model but should play significant role for lower magma fluxes.

140   In the model (Dufek & Bergantz, 2005), the introduction of dikes can appear in random

141   directions within a certain crust volume. To determine the field of displacements, the rocks are

142   considered as a viscous fluid and the Naiver-Stokes equations are solved. However, this

143   approach does not apply for low temperatures, at which the behavior of the rocks is elastic. In

144   (Karakas et al., 2017), the introduction of dikes is considered vertical. Each dike ends up with a

145   sill at the top. Dike emplacements are stochastic in position and length. Rock movement is

146   determined solely by kinematic relationships. Initially dikes are placed in the lower crust and

147   lead to thermal maturation of the whole crustal body. Later the upper crustal reservoir is formed

148   by dike injection into the upper crust. Simulations reveal that igneous provinces of less than a

149   few hundred thousand years old are unlikely to support large upper crustal reservoirs, whereas

150   longer-lived systems (active for longer than 1 million years) can accumulate magma and build

151   reservoirs capable of producing super-eruptions.

152   In all these models the efficiency of crustal melting defined as the volume of crustal melt

153   produced by volume of deep magma injected is strongly dependent on the ambient temperature,

154   (depth of the process), and the geothermal gradient (e.g., Colón et al. 2019). Melting starts after a

155   period of heat incubation which is dependent on the flux rate. Furthermore, the efficiency of

156   melting varies for different sill emplacement modes: underplating (when each subsequent sill is

157   intruded under the previous), overplating, intraplating, and random intrusion into the sill-affected

158   areas. Expectedly, efficiency of crustal melting and crustal melt productivity is decreasing in the

159   listed order, because overplating delivers heat directly to the melting boundary and maintains the

160   steep temperature gradient.

161   In the parallel effort by geologists, the lifetimes of individual magmatic systems, large and small,

162   arc- or hotspot related, can be understood by the ever-improving U-Th-Pb geochronological

163   methods involving single crystal zircon studies and most recently CA-IDTIMS methods (e.g.,

164   Schaltegger et al. 2019). When considering recent (<1-2 Ma) magmatic system, resolution of

165   individual magmatic episodes is possible on century timescales by considering zircon age spectra

166   (Crowley et al. 2007). These recent studies demonstrated for example that silicic supereruptions

167   with volume exceeding 1000 km^3 of eruptible magma can be generated in less than 5,000-10,000

168   years (which we call "Yellowstone" type), to longer than 500,000 years (which we call "Fish

169   Canyon" type) based on concurrent CAIDTIMS efforts (Wotzlaw et al 2013, 2014, 2015). The

170   eruptible magma in the first case is near- liquidus, crystal poor rhyolites, while in the second case

171   erupted magma is represented by well-mixed near-solidus mushes with 45% crystals (Bachmann

172   et al. 2002). It is important to state that "cold and wet" arc magmatic system, the most abundant

173   silicic magmatic factory on earth, exhibit a full transition of these behaviors, between these end-

174   members. Glazner et al. (2004) argued that given 10 Myr spread in ages of zircons in granites of

175   the Tuolumne Intrusive Suite of Yosemite National Park, a major batholith in the Sierra Nevada

176   Mountains (California) are formed by accumulation of silicic magmatic dikes and intrusions over

177   similarly long and less robust magmatic episodes (also cf. Miller et al. 2011). These may never

178   have an active volcanic connection to the surface, and late-plutonic processes after such dike

179   accretion, have fused them together and created relatively homogenous batholiths (Bartley et al.

180   2020). In other cases, there is undeniable evidence of the existence of near-all liquid magma

181   bodies that exist in the crust in >1000 km^3 volume and erupt to the surface in the form of a single

182   supereruption (Cashman et al. 2017; Wotzlaw et al. 2014). It can be generalized without a doubt

183   that magmatic arcs, continental and oceanic, demonstrate a great variety of sizes and styles of the

184   magmatic systems (Costa, 2008, Annen, 2009). We thus explore these possibilities in our 2D

185   modeling by varying intruded magma composition (andesite or rhyolite), flux per km^2, and

186   demonstrate below that flux and distribution of flux are the two parameters that can explain

187   multitudes of these observations.

188   Zircon is extremely resilient igneous mineral that can record multiple magmatic episodes. Its

189   crystallization is related to saturation and crystallization in T-X(compositional) space (Watson

190   and Harrison, 1983; Bindeman and Melnik, 2016). When combined with a phase diagram of a

191   magma, zircons can help understand temperature history of magmas and rocks that undergo

192   melting-crystallization phenomena (e.g., Claiborne et al. 2010b 2018; Melnik and Bindeman,

193   2018). Zircon also record isotopic values of its host magma batch, and, in particular, crustal and

194   mantle magma proportions via O and Hf isotopes, and % remelted hydrothermally altered rocks

195   (low-d^{18}O zircons). Studies of example of recent magmatism demonstrate that zircons commonly

196   show extreme O and Hf isotope diversity but have an unresolvable U-Pb eruption age (e.g.,

197   Wotzlaw et al. 2014; 2015). This demonstrates the need of mixing between isotopically diverse

198   magma bodies that are simultaneously present in the crust (Bindeman and Simakin, 2014; Colon

199   et al. 2019).

200　Links between zircon evolution with thermal maturation of the magma-rock system is studied by

201　(Caricchi, Simpson, et al., 2014) and (Caricchi et al., 2016) with a 2D axi-symmetric model of

202　magma injection. These authors considered generic magma flux and temperature evolution of the

203　system, coupled with randomized sampling of eruptible magma regions (where proportion of

204　melt exceeds 0.5) to explain formation and eruption of large volume silicic ignimbrites. A simple

205　criterion is used for zircon age distribution based on the difference between time when magma

206　becomes zircon saturated and cools below the solidus. Kinetic effects of zircon growth are not

207　considered.

208　In this paper we are motivated to further interrogate questions of crustal melting without

209　prescribing a geometry and displacements to the intrusions, in this sense we get a detailed view

210　of melting and magma accumulation processes across the growing magma body. We further

211　investigate how does the rate, distribution, and composition of magma flux into the crust

212　influence the distribution of zircon populations.

213　Therefore, this paper pursues two main goals. We first present a new generic 2D model for heat

214　and mass transfer during periodic injection of andesitic and rhyolitic dikes and sills into the

215　granitic host rocks that allows variable direction of dike emplacement, dike to sill transition,

216　country rock melting, and elastic rock displacement. Such 2D model, and software, have not yet

217　been presented to our knowledge. The model uses simplified T-X phase diagram representing

218　melt fraction, X as a function of temperature, T. Second goal is the integration of this model with

219　zircon growth-dissolution software of (Bindeman & Melnik, 2016) that reads temperature-time-

220　melt fraction properties in a Lagrangian particle. Thermal histories recorded by rock and

221　individual magma parcels are thus used for estimation of zircon crystals evolution and their

222　survivability and average growth ages in the entire magma body. We further monitor the fraction

223　of eruptible magma in different parts of the model and estimate the proportion of crustal rock

224　melt produced by newly injected magma. We next estimate O and Hf isotopic values in different

225　parts of the intrusion at assumed and typical O and Hf isotopic end-member compositions of the

226　crust and the mantle.

227   **2 Materials and Methods**

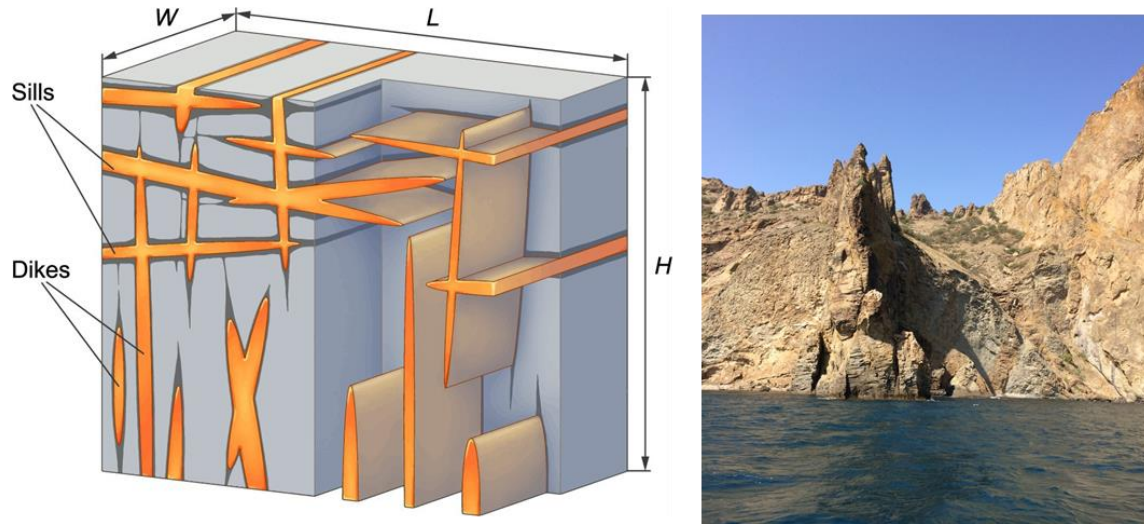228   2.1 Mathematical model

230   Figure 1. A) Set-up of the model showing intersected network of dikes and sills causing melting

231   of country rocks. B) mid-Jurassic Karadag pluton (coastal Crimea) showing similar ellipsoidal

232   dike structure, with columnar jointing marking each individual dike. (photo by A. Tugolesova)

233   Set-up of the model is presented in Fig. 1. We assume that injection occurs in 2D plain geometry,

234   model individual dike as an ellipsoid with semi-axes *a* and *b* and use analytical solution

235   (Muskhelishvili, 1977) in order to calculate host rock displacement. Each dike contains either

236   andesitic or rhyolitic magma with a specified temperature, while host rocks are granitic in

237   composition and are elastic with fixed properties. Initial temperature distribution is linear with a

238   specified geothermal gradient. Injection of individual dikes leads to displacement of elastic host

239   rocks and initiates heat transfer that causes country rock melting and intruded magma

240   solidification. Volume of the individual dike and the frequency of emplacement is controlled by

241   the specified influx rate of the magma Q_{in} (km^3/y). The flux is laterally varied per different

242   simulations from defocused to more centrally focused by specification of a normal or uniform

243   distribution of the coordinates of the dike centers. In order to calculate the volume of the

244   individual injection, the third spatial dimension is specified at *L*= 2 km and kept constant. This

245   situation is possible in the extensional tectonic environment, where the local stress field leads to

246   preferentially parallel dike orientation. We allow random angle for the individual dike

247   emplacement or the change in the dike orientation from sub-vertical at depth to sub-horizontal

248   near the surface reflecting dike to sill transition (Barnett & Gudmundsson, 2014).

$$\rho C \left(\frac{\partial T}{\partial t} + \vec{V} grad\left(T\right) \right) = div\left(k\; grad\left(T\right)\right) + \rho L \frac{d\beta}{dt}$$

$$\frac{\partial \alpha}{\partial t} + \vec{V} grad\left(\alpha\right) = 0$$

249     $$\rho C = \rho_r C_r \left(1-\alpha\right) + \rho_m C_m \alpha \qquad (1)$$

$$k = k_r \left(1-\alpha\right) + k_m \alpha$$

$$\beta_r = \beta_r \left(T\right), \beta_m = \beta_m \left(T\right).$$

250   Rock and magma temperature evolution *T* is governed by heat conduction equation (1) that

251   accounts for advection due to rock and magma displacement, latent heat of crystallization and

252   heat conduction. Here ρ is the density (indexes "r" and "m" reflect rock and magma

253   respectfully), *C* is the heat capacity, $\vec{V}$ is the advective velocity, *k* is the conductivity, *L* is the

254   latent heat of crystallization, β is the mass fraction of crystals that depends on temperature

255   according and magma type (see Supplementary Fig. S1 for $T-\beta$ diagrams), α is the volume

256   fraction of magma in the rock-magma system.

257   The emplacement of a dike is considered instantaneous, therefore, the velocity field $\vec{V}$ is

258   determined based on the displacement field $\vec{U}$ arising from the loading of an elliptical crack.

259   The values of the semi-axes *a* and b, as well as the angle of dike injection are set randomly at

260   certain intervals. The number of dikes that are emplaced during a time step is calculated in such a

261   way that the average increase in magma volume is equal to a given magma influx rate for the

262   area. Rock displacements depends on dike overpressure, that can be determined from the width

263   of the dike *b* at a given length *a*, based on the solution (Muskhelishvili, 1977):

264     $$\Delta p = \frac{bE}{a\left(1-\nu\right)} \qquad (2)$$

265   Here, *E* is the Young's modulus and ν is the Poisson's ratio of the host rocks, which are assumed
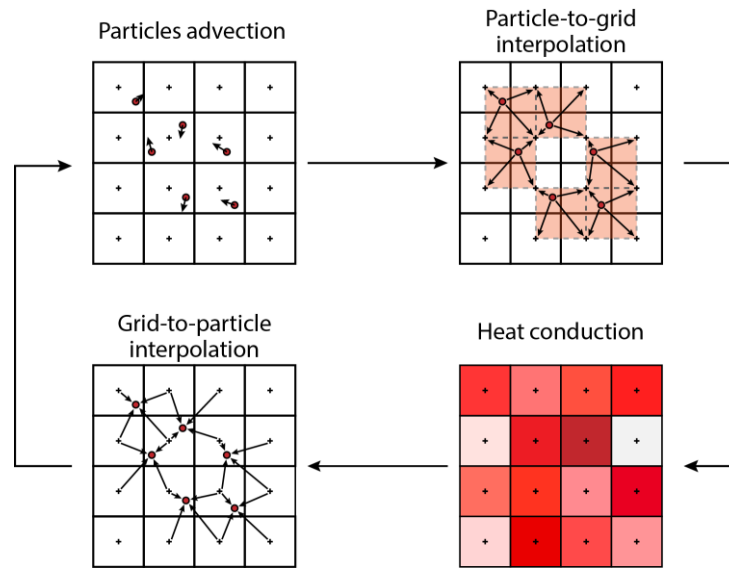
266   constant in the present study.

267   We postulate elastic rock response in our model to keep mechanical problem relatively simple

268   and separated from the thermal model in this study. Account for viscoelastic properties of rocks

269 and solidifying magma, heterogeneities of rheological properties within the domain, volatile

270 exsolution and transport, local tectonic stresses and other complexities will require solution of a

271 coupled thermo-mechanical-chemical problem that is extremely difficult for wide ranges of

272 physical properties of the system. Clearly, the model captures well the initial stage of magma

273 chamber formation when the melt volume fraction is relatively small, but we expect deviations in

274 velocity field calculations as the system matures.

275   2.2 Numerical method

276   Discretization of equations (1) is carried out by the control volume method with operator

277 splitting in spatial coordinates (Patankar, 1980). For each coordinate direction, the system of

278 linear equations resulting from discretization is solved by the tridiagonal algorithm, which makes

279 computations stable at larger time steps unlike explicit modeling schemes previously employed.

280   Since the dependence of the melt fraction on temperature (T-X, diagram, Fig. S1), as well

281 as the thermophysical parameters of the rocks can differ for magma and surrounding rocks, the

282 volume fraction α of the introduced magma is stored in each cell of the computational grid.

283 Injection of a new dike leads to advection of the volume fraction in accordance with the

284 displacement field around the dike. Grid-based methods, such as method of characteristics, suffer

285 from significant numerical diffusion, which leads to unphysical mixing between magma and

286 surrounding rocks.

287

288   **Figure 2**. Outline of the PIC method used for advection of temperature and volume fraction.

289

290   To reduce numerical diffusion, the hybrid PIC / FLIP method is used to determine the

291   volume fraction of magma and for advection of the temperature field. As shown on Fig. 2, in

292   addition to the grid, the continuous medium is also discretized by a set of marker particles that

293   store a flag indicating whether a particle belongs to magma or surrounding rocks, as well as the

294   temperature at a point in space corresponding to the coordinates of the particle. Since the

295   displacement field is determined from the analytical solution, for each particle its displacement

296   can be calculated precisely. After that, the volume fraction of magma at grid cell centers is

297   calculated by bilinear interpolation from each particle to the four cells closest to the particle.

298   Thus, the numerical diffusion for the transfer of the volume fraction of magma is significantly

299   reduced.

300   Since the heat conduction equation is additionally solved for the temperature field on the

301   grid, according to the original PIC method, it is necessary to interpolate the temperature from the

302   grid back to the particles. It can also lead to numerical diffusion of the temperature field. The

303   FLIP modification to the PIC method, developed by Brackbrill and Ruppel (1986), is designed to

304   interpolate not the temperature field itself, but only the difference between the temperature

305   obtained by interpolating from the particles onto the grid and the temperature after solving the

306   heat conduction equation. This significantly reduces numerical diffusion but leads to the

307   unphysical noise in the regions with large temperature gradients, e.g., on the boundary between

308   the dike and rocks. A common solution to this problem is to add a small fraction of PIC

309   interpolation to the result, usually about 5%, which makes it possible to smooth out these

310   oscillations by adding a small amount of scheme diffusion, which is negligible compared to

311   physical heat diffusion due to thermal conductivity.

312   When a new dike is introduced, new particles are generated. If any of the grid cells

313   contains too many particles as a result of elastic compression of the medium, excess particles are

314   destroyed. We stop tracking a particle when it reaches the boundary of the computational

315   domain. If there are too few particles in the cell due to the stretching of the rocks, new particles

316   are created, and the values of temperature and volume fraction in them are interpolated from the

317   grid.

318   We use fixed grid spacing of 5 m for temperature calculations. This results in 2000x2500

319   grid points in the computational domain. The code was tested on analytical solutions of 2D heat

320   exchange problems.

321   **3. Results**

322   The model presented above allows users to study a wide range of magmatic phenomena. The

323   code is written in Julia programming language (julialang.org) and individual simulation setup is

324   governed by a run file that utilizes YAML markup language capabilities to specify physical

325   properties of the rock and magma, depth, distribution, angles and sizes of injected dikes,

326   proportion of dikes to sill and dike to sill transition and distribution vs depth, magma

327   emplacement schedule with different influx rates and durations for each injection episodes.

328   Results of the simulations are stored in hdf5 files for mesh related variables and .csv files for

329   individual Lagrangian particles that are used as input conditions for zircon growth/dissolution

330   simulations. The program and algorithm are accessible to anyone with minimal programming

331   skills and is designed to be modified by the users to their specific situations, for example for

332   studies of individual magmatic and caldera systems.

333   Below we present examples of application of our approach to typical island arc situations.

334   The results of this modeling show capabilities of the model and mimic basic scenario for the

335   formation of magma chamber by rhyolitic and andesitic magma intrusions.

336

337   3.0.

338   **3.1. Rhyolitic and andesitic dike injection into granitic crust**

339   We first present simulation results for magma (0% crystals, T=900 °C) of rhyolitic composition

340   injected sequentially as dikes and sills into the crust of the same, fully crystalline granitic

341   composition. Depths of injections varies between 5 and 12 km, with an initial temperature

342   gradient of 20°C per km, with a given temperature at the top of 200 °C and 500 °C at the bottom

343   of the injection zone, typical depths of crustal magma chambers and batholiths. This setup

344   simulates formation of upper crustal plutons and magma bodies containing eruptible magmas.

345   Parameters that are used in simulations are listed in Table 1.

346

Parameter	Description	Value
ρ	Density	2650 kg/m^3
λ_r	Thermal conductivity of rock	1.5 W/m/K
λ_m	Thermal conductivity of magma	1.2 W/m/K
C_p	Specific heat capacity	1350 J/K/kg
L	Latent heat of melting	3.5×10^5 J/K/kg
T_m	Magma intrusion temperature	Andesite: 1100 C Rhyolite: 900 °C
T_{top}	Temperature at depth $z = 5$km	200 °C
ΔT	Temperature geothermal gradient	20 °C/km
ν	Possion's ratio of rock	0.3
E	Young's modulus of rock	15.6 GPa
Q_{in}	Intrusion rate	0.25 – 1 m^3/s
$[z_{min}, z_{max}]$	Dike center depth	6 – 14 km
w	Half-width of dike intrusion region	250 – 2000 m
$[zs_{min}, zs_{max}]$	Depth range of sills formation	6 – 8 km
θ	Dike rotation angle	85° – 95°
a	Dike length	200 – 3000 m
b	Dike thickness	10 – 20 m
h	Dike transverse width	2000 m

347   Table 1. Parameters used in the simulations

348

349   Fig. 3 presents the graphical results of calculations of the melt fraction and temperature evolution

350   during formation of a magma chamber at different stages of its growth and cooling. Magma

351   injection rates are taken as Q = 0.5 and 1 m^3/s (0.015-0.03 km^3/year), and few simulations below

352   consider lower 0.25 m^3/s rates. Dikes are introduced sub-vertically in the range of depths from

353   12 to 5 km. The injection zone grows in vertical and horizontal as the system swells by vertical

354   and lateral extension (Fig. 3). Given that dike intrusions are changing orientation to sills as is

355   abundantly seen in nature (Fig. 1b, McBirney, 2006; Thomson, 2007) and experiments (Menand

356   et al. 2010), we adopt such transition in the model. We assume that if the center of the dike is at a

357   depth between 6 to 8 km, then the emplacement of the magma becomes sub-horizontal (sills).

358   Such situation is common in the areas of active volcanism and may be associated with the

359   distribution of the density of the host rocks (level of neutral buoyancy for the magma, (Walker,

360   1989) or the presence of layers with a larger Young's modulus (Barnett & Gudmundsson, 2014).

361   The analytical solution that is used to calculate the displacements of host rocks in current model

362   cannot be applied for the latter case as it assumes homogeneous elastic rock properties.

363   In current set of simulations injection takes place during first 4000 years (4ka), followed

364   by a subsequent cooling for 750,000 yr (750 ka). No volcanic eruption (removal of melt and

365   heat) is allowed; however, we track percent and distribution of areas with >50% (eruptible

366   magma), and >1% melt (solidus), and % of crustal melt vs intruded melt throughout the

367   intrusion. For low injection rate ($Q = 0.5$ m^3/s) emplacement of dikes into W=4 km width region

368   of the crust forms patchy distribution of molten rocks with observed low connectivity in between

369   during the intrusion period (Fig. 3). Due to heat transfer after the injection and melting of

370   country rocks the connectivity and sizes of individual batches of magma progressively increase

371   while their melt fractions decrease. After ~300 ka temperature drops below the solidus at every

372   point and the magmatic system solidifies completely into a mid-crustal pluton consisting of a

373   series of closely spaced and accreted dikes.

374   Larger influx rate ($Q = 1$ m^3/s) leads to a formation of fully molten magma body,

375   maintaining the 100 % melt fraction in its center even after 100 ka of cooling if magma injected

376   in a narrower region (W=2 km). In the case of wider distributed magma flow W=4 km, a

377   magmatic system is formed with high melt fraction and small batches of fully molten magma. In

378   the latter case, the overall lateral proportion of dikes is larger than the injection area due to

379   displacement of the host rocks and previously injected magma.

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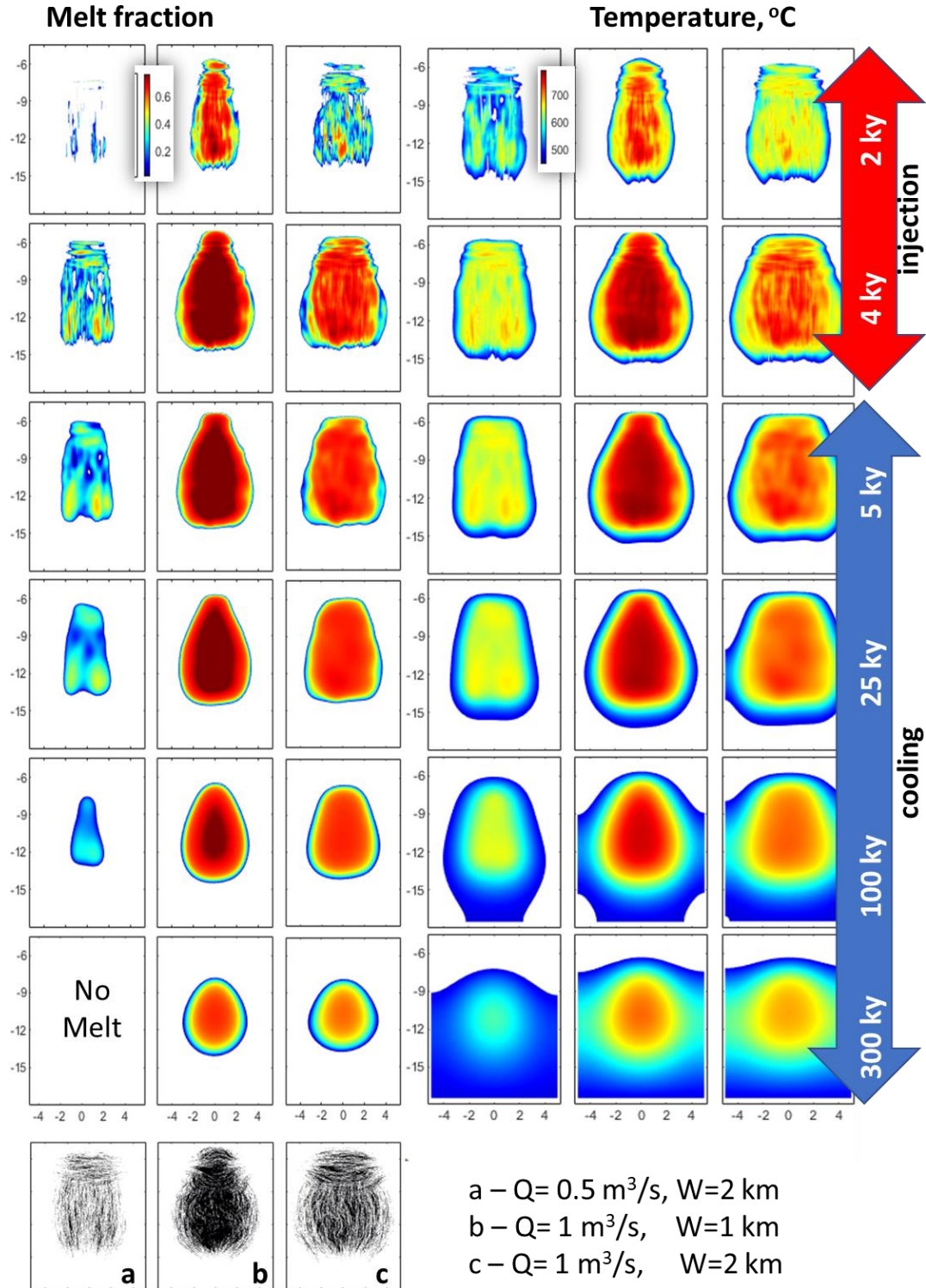
Figure 3. Evolution of melt fraction and temperature in the magma body grown due to sill and dike injection. Each vertical column represents a separate case with a different rate Q, and injection zone W. Magma injection time is 4 ka followed by magma cooling without injection.

387   This case only considers rhyolitic dikes intruded into a granitic crust of similar composition. For

388   andesitic magma intruded into the granitic crust, see Fig. S2 in the supplementary material.

389   Fig. 4 shows kinematics of Lagrangian particles associated with rocks and magma. In the

390   dike injection region displacement is preferentially horizontal and downward, while intrusion of

391   the sills leads to a vertical displacement (uplift) of several hundred meters. Particles near the

392   central line of the intrusion exhibit small horizontal shift, dikes injected at later stages stay within

393   the injection region, while the early injected material is pushed to the periphery of the intrusion,

394   in a fashion similar to spreading. The total displacement during active diking period lasting 4 ka

395   exceeds 2 km at depth and a few hundred meters near the surface, while thermal equilibration

396   length-scale on this timescale is of order of 300 m. It means that during the formation of the

397   magma chamber heat transfer and balance is dominated by advection and latent heat

398   release/consumption. This is a novel result of our 2D simulation with elastic displacement, not

399   seen in previous 1D models.

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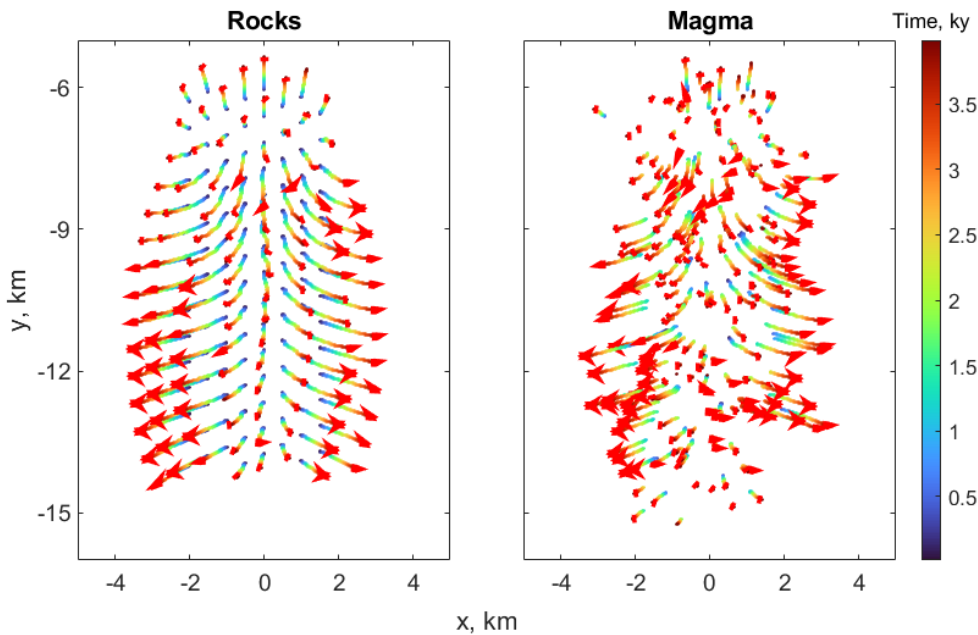
402   **Figure 4.** Displacement of Lagrangian particles associated with magma (right) and rocks (left).

403   The time of dike emplacement is shown by color. Note the "Christmas Tree" sagging pattern in

404   the lower part of the model and an uplift in the upper part.

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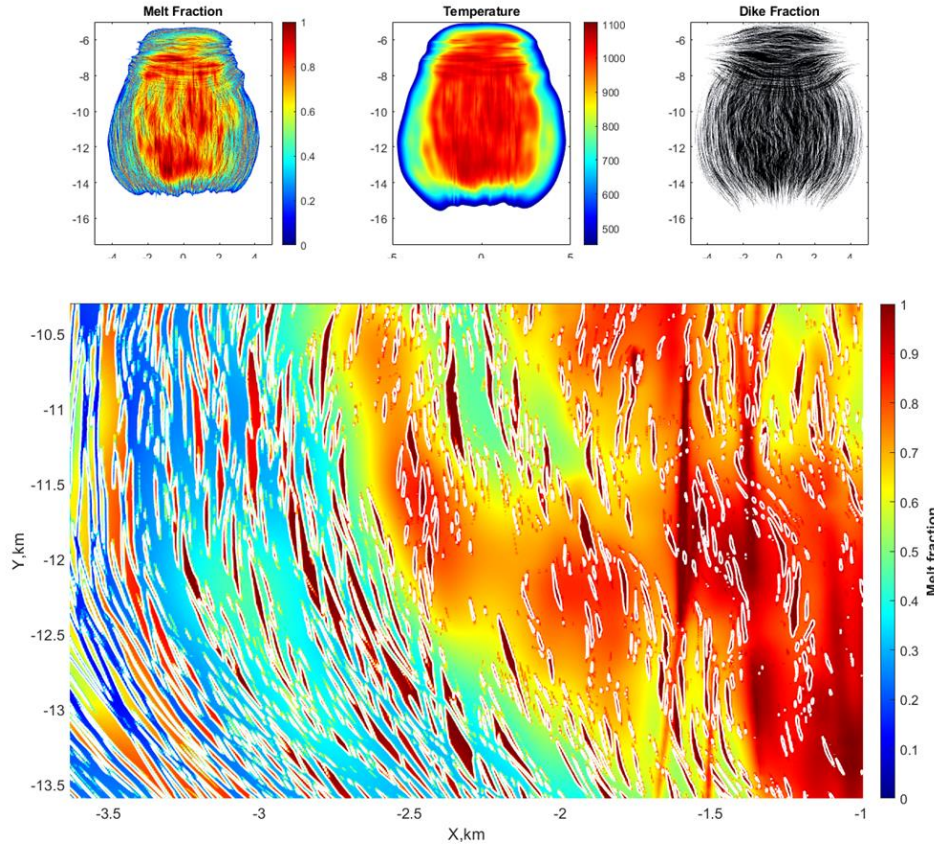
Figure 5. Distributions of the melt fraction, the temperature, and dikes in the case of injection of the hotter andesitic magma into granitic crustal rocks with Q = 1 m^3/s. Note high melt fractions and connectivity in the right-hand side of the model. The shown region can be treated as largely molten interior and partially molten contacts.

410

We next consider a case when hotter andesitic magmas are intruded into the granitic crust (Fig. S2). Like is the case with the rhyolitic dike injection, first andesitic dikes freeze without melting, and an incubation period of several hundred years is required before the system is able to sustain sizable and growing magma volumes. The behavior of andesitic -granite system is also different because of different composition of the intruded dikes and greater liquidus-solidus separation, which affects composition of the partial melt that is stored in andesitic dikes at different time and place. However, and perhaps counterintuitively, hotter andesitic dikes do not produce more melt in the system. Fig. 5 shows distributions of the melt fraction, the temperature, and dikes in the case of injection of the hotter andesitic magma into granitic crustal rocks with a high flux rate Q = 1 m^3/s after 4 ky of magma influx. Distribution of the melt fraction is patchier than in the case

421   of injection of rhyolitic magma. Zoom out of the lower portion of the intrusion is shown on the

422   bottom panel where fully molten regions are associated with the remaining host rocks and newly

423   injected dikes, meanwhile magma injected earlier shows variable degree of remaining partial

424   melt. Intrusion of andesitic dikes generates rhyolitic partial melt in the country rocks and a

425   residual melt of variable SiO_2 as interstitial in andesites. At large melt fraction, melts would

426   segregate and higher temperature restites would sink, but this is not modeled in this work.

427

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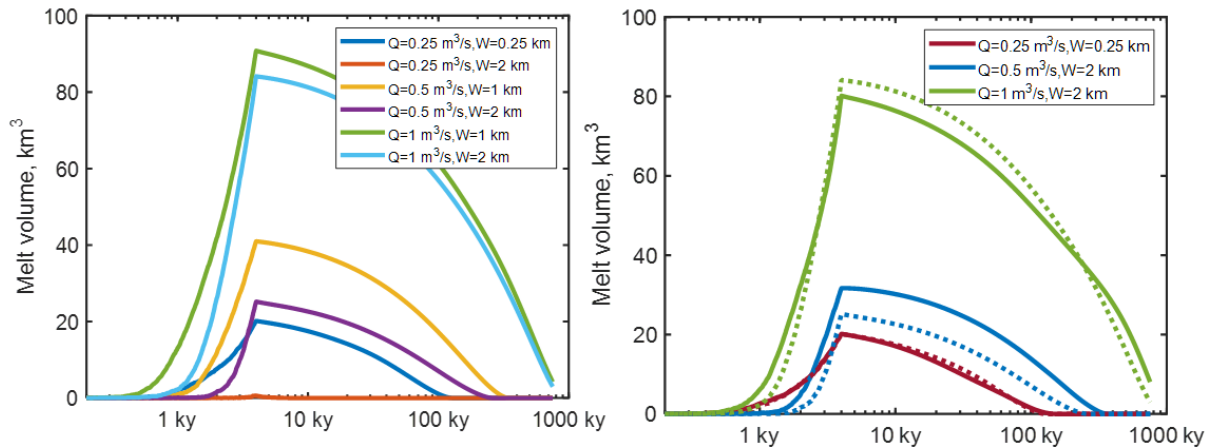
430   **Figure 6**. Melt volume fraction for different injection rates and width of the domain (a) –

431   rhyolite to granite. (b) – andesite to granite. Intrusions continue to 4 ka after which they stop and

432   the system cools. Note the incubation period needed for preheating of the crust on the order of

433   500 yrs, followed by a rapid onset of melting. Andesites intruding granite generates the same

434   amount of total melt because at lower temperatures andesitic crystallize more (see phase

435   diagrams on Fig. S1).

436

437   **3.2. Variation in the Magma influx rate**

438   Efficiency of melt production and formation of molten magma bodies strongly depends

439   on the magma influx rate into the upper crust. Fig. 6.a presents the dependence of melt volume in

440   time for four-fold change in the magma influx rates Q from 0.25 to 1 m^3/s. At these flux rates,

441   over 4 ka the total injected magma volume varies from 31.5 to 126.3 km^3, respectively. The

442   effect of this magma flux is however strongly dependent on the total widths of the injection

443   region (magma focusing). Variations in the flux rate and the width results in dramatically

444   different outcomes for magma productivity and the efficiency of crustal melting (Fig. 6). Low

445   influx rate injection of rhyolitic magma ($Q = 0.25$ m^3/s) over a wide region W=4 km, produced

446   only 0.2 km^3 of melt at 4000 years, and nearly 20 km^3 if dikes are focused and W= 0.5 km. At

447   higher influx rate there is still a significant influence of the dike focusing on the produced total

448   melt volume. At high overall magma intrusion rates of $Q = 1$ m^3/s the dike focusing influence is

449   rather small.

450   Change in the magma type from rhyolite to andesite again surprisingly results in a

451   comparable overall melt production, due to rapidly changing melt% with a decreasing

452   temperature in the andesitic phase diagram (Fig. S1). Although the temperature of andesitic

453   magma (solid lines on Fig. 6.b) is 200 ^{o}C higher than for the rhyolitic magma (dashed lines),

454   melt volumes produced by both magmas are similar and are controlled by the injection rate and

455   the region width. Melt production for the case of andesitic magma intrusion starts slightly earlier

456   than for rhyolitic due to more rapid melting of the host rocks by hotter magma.

457

458   **3.3. Percent country rocks vs magma differentiate and its variation with time**

459   Fig.7 represents exploration of the extents of crustal melting by rhyolitic magma

460   intrusion by specifically tracking the regions with country rock melting (>1%), and for eruptible

461   magma (in which total melt fraction exceeds 50%). This number is taken as a simple measure or

462   rheological extractability of magma from mushy zones (e.g. Marsh, 1981; Huber and Parmigiani,

463   2018). We explore different magma flux rates identical to Fig. 6, resulting in dramatically

464   different total magma production. The ratio of country rock melt to intruded magma melt in the

465   eruptible magma (Fig 7a) has slightly different patterns. First, proportion of crustal melt is

466   increasing with time but reaches its peak before the end of magma injection period at 4 ka, as the

467   injected magma dilutes the melt zones with newly intruded magma. This is a combination of

468   cooling and melting as well as spreading of the melt zones. Regardless of the robustness of

469   magma production (greater flux Q at narrower zone W) amount of crustal melt and the pattern of

470   its increase remain approximately the same with slightly different rise time. Less robust

471   magmatic systems (e.g. Q=0.5 m^3/s, W=4 km) exhibit expectedly slower increase in total melt

472   (Fig. 6) and crustal melt production (Fig. 7a). Upon cessation of intrusions, for most cases except

473   one for least robust magma system (Q=0.5 m^3/s, W=4 km), percent of crustal melt relative to

474   intruded magma in dikes drops quickly, likely because the crustal melt in our model is located on

475 contacts of intruded dikes and thus cools quicker by conduction. In real magmatic system where

476 melts are given a chance to segregate, this may not be the case. Nonetheless, the total amount of

477 eruptible crustal melt never exceeds 15-20%, suggesting that in all cases it is cooling and

478 crystallization of intruded magma that dominates the system. Additionally, because our model

479 allows for the spreading of particles upon magma injection, country rocks are advected away

480 from the heat source faster than the thermal diffusion is capable of melt them. This leads to lesser

481 efficiency of crustal melting than it is possible to achieve in psudo-1D models. For example,

482 overplating in such models always keeps the upper melting boundary hot (Annen et al. 2006)

483 leading to significant up to 50% efficiency of crustal melting.

484   The melting efficiency defined as a total melt volume divided by the volume of injected

485 magma (Fig. 7b) shows patterns comparable to total melt production (Fig. 6). Expectedly, least

486 robust magma system ($Q=0.5$ m^3/s, $W=4$ km) have least efficiency peaking at 15%. Most other

487 flux (Q) and width (W) combinations operate with 60 to 70% efficiency, with melt dominated by

488 the intruded hot magma. Melting efficiency drops with cessation of new magma intrusion as the

489 whole system crystallizes reducing melt volume (Fig. 3).

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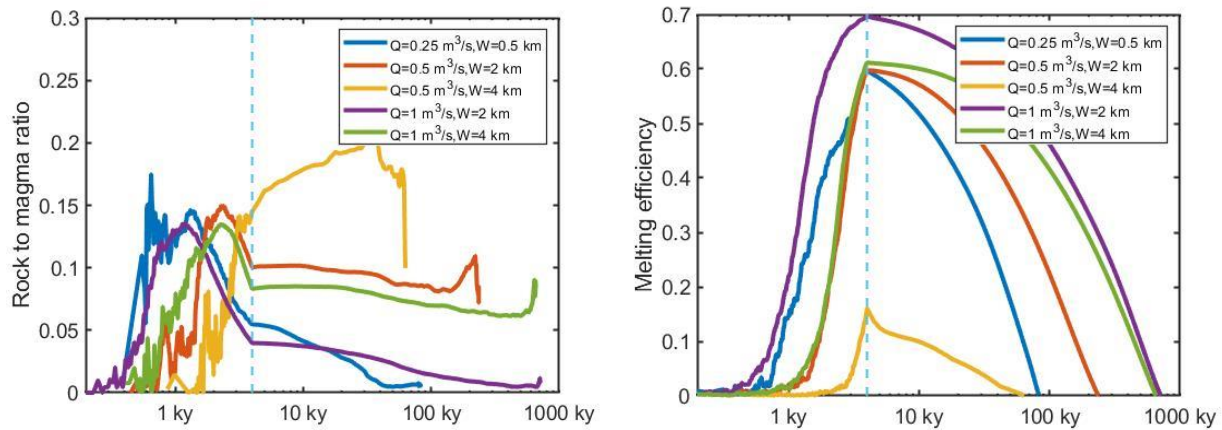
493 **Figure 7.** Crustal melting and its efficiency in the model runs with indicated magma flux Q and

494 width of intruded area W. Vertical dashed line represent a point when magma intrusion stops at

495 4ka, and the cooling of the whole system begins. a) Rock to magma ratio in cumulative areas

496 within the model where melt fraction is greater than 50%, meaning to represent potentially

497 rheologically-extractable and eruptible magma. b) Temporal variation of melting efficiency

498   defined as total melt volume (melt fraction >1%) divided by the volume of injected magma.

499   Efficiency peaks at the end of magma intrusion at 4 ka and the decreases as magma system cools

500   and crystallizes.

501

502   Rock to magma ratio is drastically different in the case of andesitic magma injection in granitic

503   crust (Fig. S3C). In the case of wide injection area or low magma flux magma solidifies very

504   quickly and melts host rocks resulting to close to one rock to magma ratio. At higher or more

505   confined magma flux up to 30 % magma remains molten end eruptible.

506

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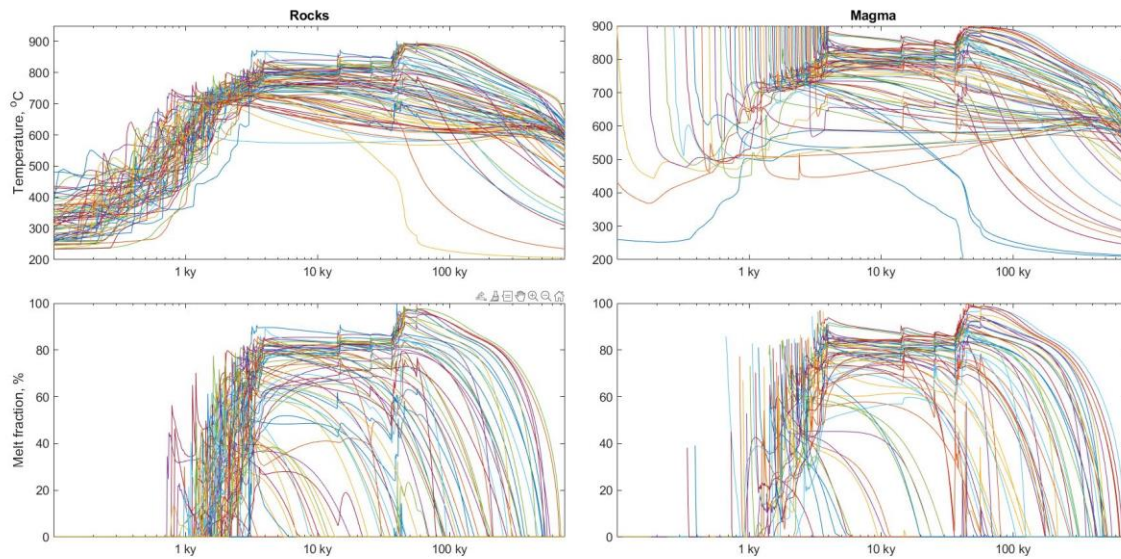
508   **Fugure 8.** Temperature and melt fraction variation in granitic rocks (a-b) and rhyolitic magma

509   (c-d) parcels with time for Q = 0.5 m^3/s and W=0.5 km. Main dike intrusion period lasts for 4 ka,

510   three 1000-year episodes of dike injection occur each 10 ka resulting in spikes in temperature

511   during cooling of the magma chamber.

512

513

514   **3.3. Magma and host rock thermal histories**

515   The 2D model presented in this work not only allows investigation of the cumulative

516   thermal (and compositional, % crystals) evolution of the whole system as was done in the

517   previous models but permits investigation of the behavior experienced by multiple areas within

518   the growing magma body and country rocks, as well as inside of each dike or a sill (Fig. 1).

519   Country rocks, and previously intruded and now solidified dikes can experience a non-trivial,

520   spatially- and temporally independent series of heating, melting and crystallization episodes.

521   Thus, the real 2D model presents a more realistic scenario for natural situations, and these

522   changes will be reflected in their crystals including zircons. Furthermore, consideration of elastic

523   displacement of rocks after intrusion in our model (spreading), provides an additional, previously

524   underexplored feedback between mass and heat advection, melting and crystallization in

525   different parts of the system.

526   Fig. 8 shows the dependences of the melt fraction and temperature on time for magma

527   particles and host rocks for injection of rhyolitic dikes in granitic crust for $Q = 0.5$ m^3/s and $W =$

528   0.5 km. Each magma injection cools down quickly and completely solidifies during first 1-2 ky,

529   but later intruded, and country rock melt fractions persists for a long time, until the temperature

530   of the whole magma bodies drops down below the solidus. Occasional, late-stage intrusions

531   (three spikes in Fig. 8) results in slight temperature and melt fraction increases. Rocks are not

532   significantly melted by the magma at the first stage of the injection, but the degree of melting

533   reaches more than 80% at later stages of magma chamber formation. Variation of temperature

534   and degree of melting in an individual particle is non-monotonic and experience several heating

535   and cooling episodes depending on its location within the area of dike injection. For example,

536   early dikes cool down quickly and are later displaced to the periphery of the forming magma

537   chamber. At the same time, dikes in the center remain hot and are later remelted and their melt is

538   then reincorporated into the large body of the magma chamber. With progressive dike addition,

539   subsequent dikes deliver heat and mass to a maturing reservoir those temperature increases

540   smoothly. Upon cessation of new magma addition, its temperature decreases monotonically with

541   time, while reservoir attains more spherical shape with respect to its thermal halo and melt

542   distribution (Fig. 3). Later parcels of magma experience more complicated t-T path. The

543   temperature and melt-fraction-time history that we observe in our 2D model both confirm "spike,

544   then exponential decay" schemes often shown in modeling and petrologic literature dealing with

545   magma intrusions (Caricchi et al.,2014; Szymanowski et al., 2017), used for example o model

546   solution-reprecipitation of crystals with complex zoning. However, they are different in several

547   important aspects. Initial stage of magma chamber growth is characterized my highly

548   inhomogeneous temperature distribution. As magma chamber matures newly injected dikes do

549 not change its thermal state significantly because the volume of individual intrusion is much less

550 than the volume of accumulated melt. Periodic unrests of the system do can locally result in large

551 temperature increases but the average cooling trend continues.

552

553 Temperature histories for andesitic magma injection into granitic crust are presented and

554 discussed in Supplementary material, Fig. S4. In comparison with Fig. 8 many of rock parcels

555 become completely molten and stay molten for a long time, especially at high Q values. Hardly

556 any magma parcels stay completely molten after injection except in the case of consolidated

557 magma flux (Fig. S4b). This is very different from prescribed fixed position injection of the

558 magma simulated in previous models where newly injected magma is always separated from

559 host rocks by previous injections.

560

561 Overall temperature and melt fraction evolution in our 2D model inform us on new

562 developments as a result of proper accounting for whole system spreading at rates faster than

563 heat conduction. In some central parts the system exhibits behavior consistent with "overplating"

564 models in which newly injected melt occupies the magma/rock boundary, and maintain rapid

565 temperature rise and melting, most other parts of the model exhibit less melting that even in the

566 random dikes and sills models of Dufek and Bergantz (2005) or Annen et al. (2006). Overall, T-

567 Xmelt-time trajectories of the different parts of the modeled magma system experience extreme

568 diversity, that span prescribed overplating to random dike regimes. Overall, such melting is less

569 efficient and reflect processes in the real growing magma system.

570   The model further informs us on realistic Time-Xmelt-time trajectories that can be used

571 in modeling of crystal destinies. For example, zircon (and other mineral) zoning patterns, can be

572 now modeled and interpreted more realistically than previously prescribed spike-then decay

573 schemes.

574

575 **3.4. Zircon evolution**

576 An important method of investigation of plutonic and volcanic system to understand their

577 formation is evolution is provided by examining their crystal record, among which zircon

578 petrochronology is playing increasingly more important role. Not only domains within zircons

579   can be dated with ever improving precision informing on the duration of zircon growth and

580   inheritance, but also the trace elemental zoning pattern within zircons can provide information on

581   T-X(%melt)-time changes (Claiborne et al. 2010; 2018; Melnik and Bindeman, 2018). This is in

582   addition to the use of the O and Hf isotopic values within zircons that informs on whether they

583   represent products of crustal melt, newly intruded magma, or product of remelting of

584   hydrothermally altered rocks (e.g., Bindeman and Simakin, 2014; Wotzlaw et al. 2015; Colón et

585   al. 2019).

586   This paper adds to these tasks by examining zircon behavior in the course of magma

587   body construction and cooling. For this purpose, we use temperature and compositional history

588   recorded in each marker points (50,000 total markers) related to the spatio-temporal evolution in

589   host rocks and magma and integrate these with zircon crystallization-dissolution software of

590   (Bindeman & Melnik, 2016). This relatively straightforward coupling enables reconstruction of

591   dissolution and growth histories of great multitude of individual zircons of the entire system. For

592   example, depending on the time, composition, and the position of the injected magma zircons in

593   it can 1) crystallize rapidly (years-decades) during complete solidification of magma short after

594   its emplacement or 2) crystallize over a prolonged period of time if magma is injected into the

595   hot interior and thus remains molten in a long time and cools slowly with the whole system, 3)

596   crystallize from a dike or a country rock melt inheriting their respective O and Hf isotopes, or 4)

597   first crystallize then dissolve multiple times. Zircons in the country rocks can undergo variable

598   amounts of partial dissolution depending on the temperature-time history of each particular

599   marker and preserve or not preserve the inherited core and completely or partially reset of their

600   ages and O and Hf isotopic values. Given the new 2D thermal approach presented in this paper,

601   similar modeling can be implemented for any other accessory mineral for which there is

602   diffusion and saturation conditions, or a major mineral for which the phase diagram relations can

603   be specified.

604

605   We have performed a series of such computations of zircon destinies for 50,000 of markers

606   located inside of the growing magma body. Fig. 9 shows the evolution of the melt fraction (up),

607   temperature (middle) and zircon radius (bottom panels) recorded in two magma parcels. Solid

608   line corresponds to an early injected dike, dashed – to the dike injected at the end of the

609   magmatism. In the first case magma solidifies completely and is getting remelted several times.

610   We start with zircon evolution simulation from the last complete solidification episode assuming

611   that the crystal is has initial size R_z governed by the bulk concentration of zirconium C_{bulk} in

612   magma calculated as:

$$R_Z = \left(\frac{C_{bulk}}{C_Z} \right)^{1/3} L_{cell}; \ C_Z = 490,000 \, \text{ppm} \quad (3)$$

614   Here C_z is the zirconium concentration in the zircon, L_{cell} is the size of the cell from which the

615   crystal is growing or dissolving. This parameter is specified by the number density of zircon

616   crystal N_z per unit volume $L_{cell} = \left(\frac{1}{N_Z} \right)^{1/3}$. Smaller number density of the crystals will result in

617   larger cell sizes, and larger initial zircon radius (see Bindeman and Melnik, 2016 for more

618   details). We assume total Zr concentration in the granitic crust is 150 ppm and in intruded

619   magma has 50-150 ppm. We use Boehnke et al. (2013) zircon saturation conditions, although

620   Watson and Harrison (1983) conditions can be easily implemented if preferred. Crystal sizes

621   range from 49 to 60 μm and M-factor of the melt is assumed to be a function of the temperature

622   per Bindeman and Melnik (2016) formulation. As was shown in the latter paper, consideration of

623   compositional effects (M-factor) typically results in faster zircon dissolution and growth during

624   temperature oscillations. Furthermore, we allow for change in zircon cell size caused by

625   crystallization of other minerals on the zircon cell's boundary. This promotes faster zircon

626   growth during cooling, and faster zircon dissolution during country rock melting.

627   We additionally consider cells with different bulk Zr concentration (Fig. 9) ranging from

628   50 to 150 ppm. While concentrations of <~80 ppm are least realistic for igneous rocks, as is

629   described in Bindeman and Melnik (2016) metamorphic and intraplutonic recrystallization may

630   lead to areas within a rock with strongly different Zr concentration. For example, zircon may be

631   surrounded by Zr-poor quartz-feldspar assemblage that will melt as cotectic during temperature

632   increase. Survivability of zircons is, thus, strongly depends on the zircon size and its cell size

633   defining total Zr concentration within a cell. All the crystals in current model will grow to the

634   same size governed by bulk zirconium concentration if the temperature returns to the same value.

635

636   Below we consider several scenarios of zircon behavior. For crustal rock particle close to the

637   center of the intrusion, as the temperature increases due to continues magma supply, zircon

638　　dissolution begins. In the case of a large initial crystal (60 μm sitting in a cell of with and large

639　　C_{bulk}= 150 ppm Zr) only small portion of the crystal dissolves (red solid curve), while for C_{bulk}=

640　　25 zircon dissolves completely and restarts the growth when the temperature drops down to

641　　saturation conditions. In this case the crystal will be significantly younger.

642

643　　Inside of the intruded magma, zircon crystallize as soon as the temperature drops below

644　　saturation temperature and crystallizes until the solidus (initially) or until the thermal

645　　equilibration with the main reservoir. In the case of late injection, the initial temperatures in the

646　　beginning are too high and magma remains Zr undersaturated until the whole magma body

647　　temperature drops below zircon saturation after cessation of new magma addition. Expectedly,

648　　for smaller bulk Zr concentrations (C_{bulk}), magma saturation with zircon occurs at lower

649　　temperatures and zircon age is much younger than for large C_{bulk}. For Zr-richer magmas zircon

650　　growth will start immediately after the injection.

651

652　　As our 2D models permits simultaneous computation of destinies in 50000 of magmatic zircons,

653　　we can describe evolution of zircon age patterns in the entire magma body. Fig. 10 shows the

654　　distribution of growth durations (final time - starting time) of these 50000 magmatic zircons,

655　　which demonstrate important spatial and temporal complexity. As dikes cool and crystallize

656　　quickly without a subsequent episode of remelting in the periphery of the intrusion, zircons grow

657　　for years to tens of years and the age distribution between the core and the rim is very restricted

658　　and is identical to that of the country rock or that of the intruded dike.

659　　　　Spatially, in the central part of the intrusion, newly accumulated melts crystallize and

660　　continue to cool down much slower and zircons record continuous growth of hundreds ka, with

661　　rims of these crystals younger than their cores by this exact amount of time. Zircon core age-

662　　relationships can be a diagnostic criterion of the magma parcels location within the intrusion, and

663　　the mechanism of the intrusion formation. A great number of zircons record periods of

664　　dissolution, then growth, the intensity of which depends on the proximity to the intruded dikes.

665　　In studying plutonic rocks, this may be a tool to investigate spatial distribution. However, in the

666　　case when the boundaries between individual zones inside of the plutons are blended (cf Bartley

667　　et al. 2020), random sampling of an accreted intrusion may just record a particular episode of

668　　local zircon crystallization.

669

670   Temporally, early eruptions from the growing magmatic system will expectedly sample ages of

671   these individual intrusions and not the country rocks, which have not yet melted. At intermediate

672   stages of intrusion development (Fig. 3), formed magmas will sample zircons from country rocks

673   providing a great deal of inherited zircons diversity ranging in age from country rocks to time of

674   the intrusion. Some magma system at low flux rates will end up in this situation.

675

676   Among 12800 markers related to host rock parcels only half (~ 6390) experienced certain degree

677   of melting. The rest of the rocks were displaced by dike injection far from high temperature

678   region and were not altered by the magma intrusion. Fig. 11 shows the distribution of the zircons

679   that experienced dissolution and therefore rejuvenation of the average zircon age reduction (a)

680   and the histogram of the volume fraction of preserved cores (b) for the initial zircon radius of

681   100 μm. In the central area of the intrusion ~ 5% zircons associated with host rocks were totally

682   melted and later grew again, therefore, the age of zircons erupted from this region will be totally

683   controlled by magma intrusion time. Due to high special heterogeneity of the central part of the

684   intrusion some of the host rocks zircons remain unaltered in the central part of the intrusion.

685   There is a spectrum of average zircon ages in a single magma blob that can complicate the

686   interpretation of the history of particular magmatic chamber. The proportion of magmatic to

687   host-rock zircons that can be erupted depends on the degree of melting and temperature-time

688   histories for particular locations, and these are explored more below by assigning specific O and

689   Hf isotopic values to magma and areas in the crust.

690   Zircon destinies will strongly depend on the overall magma flux Q and magma focusing

691   W. Defocused systems with low magma flux will generate and preserve great deal of zircon age

692   diversity and inherited cores. Notice that the majority of cores will not be from country rocks but

693   rather from earlier crystallized intrusions (80-90%, Fig. 7), thus recording the earliest episodes of

694   magma intrusion. However, at large magma flux rate, most initially crystallized and most

695   country-rock zircons (except the ones at the periphery) will dissolve.

696   The main conclusion of the current 2D model are that high flux and focused systems lead

697   to rejuvenation of most zircon ages and maintaining strong core-rim heterogeneity between the

698   core and the rim of the intrusion as a whole. If only the magma stored in the center of intrusion is

699   allowed to erupt (thus sampling the main body of the growing magma) only "eruption age"

700   zircons will be observed. These are predicted to have a long tail of ages going back to the

701   beginning of zircon crystallization, which as we demonstrate here may be as long as a few tens-

702   hundred ka. The fact that many erupted zircons from large supervolcanic systems such as Bishop

703   Tuff and Yellowstone do not have such long tails (Crowley et al. 2007; Wotzlaw et al. 2015) and

704   most zircons are of eruption age (±5-10ky) signals that magma system was not allowed to age

705   before eruption, and magmas were erupted soon after formation. Such relationships signify an

706   important difference between such volcanic and mush-dominated (Fish Canyon tuff, with zircon

707   age tails of 300 ka, Wotzlaw et al. 2013) and plutonic environments, where zircons can span 10

708   My (Glazner et al. 2004).

709

710   It is also important to note that sporadic magma injection during the cooling stage of the

711   already formed magma body leads to small amplitude thermal spikes due to the overall high

712   thermal inertia of the system, resulting in short-lived episodes of zircon dissolution. We earlier

713   modelled that oscillatory zoning in zircons can be formed this way (Melnik and Bindeman,

714   2018).

715
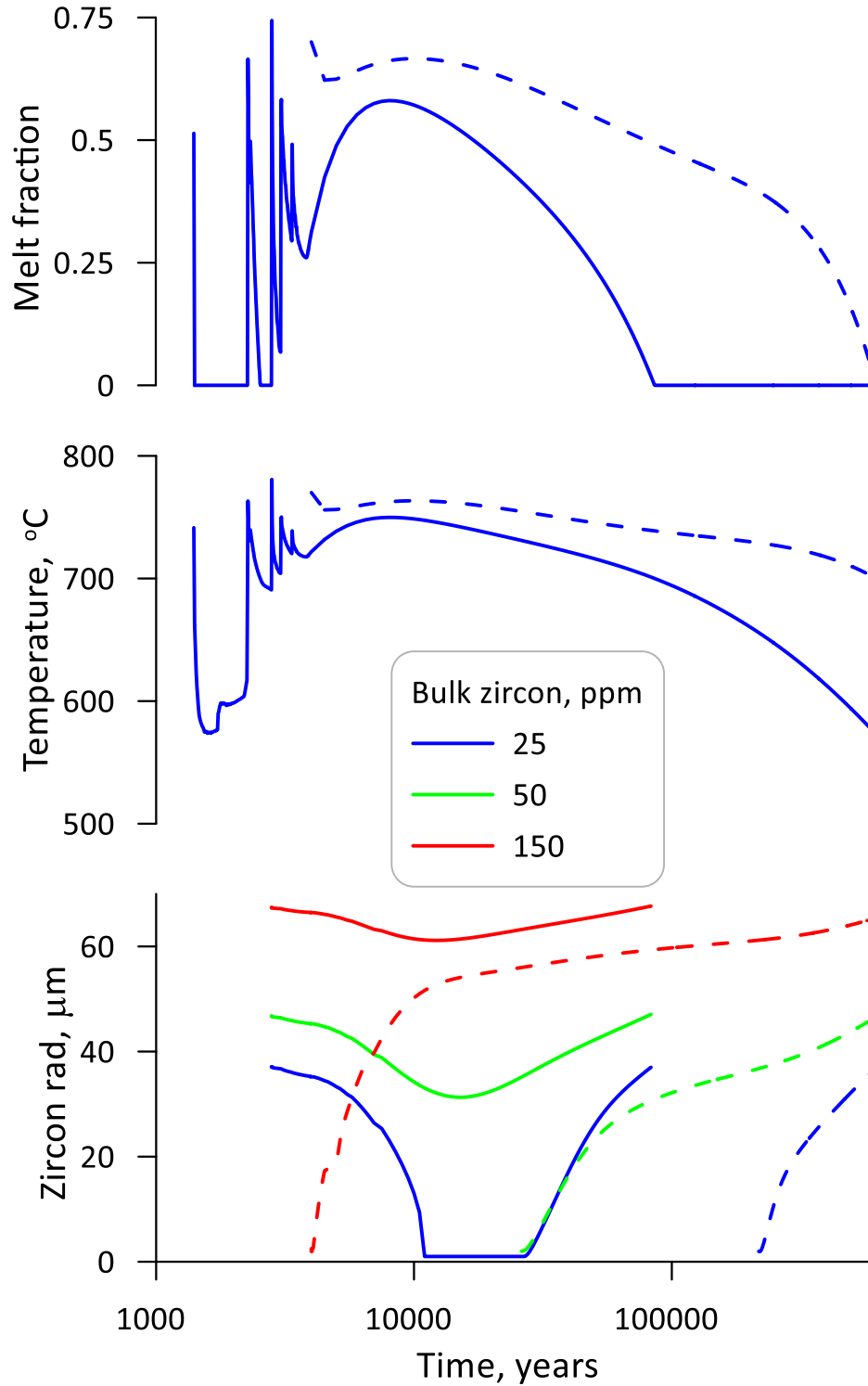
716
717 **Figure 9**. Evolution of the melt fraction (up), temperature (middle) and zircon radius (bottom)
718 recorded in two magma parcels, surrounded by rock with Different bulk zirconium contents in
719 magma are shown by colors. Notice that zircons in Zr-poor rocks will get completely dissolved,

720   then regrown, while zircons in Zr-richer rocks will undergo only partial dissolution, developing

721   structure of inherited core surrounded by a rim

722
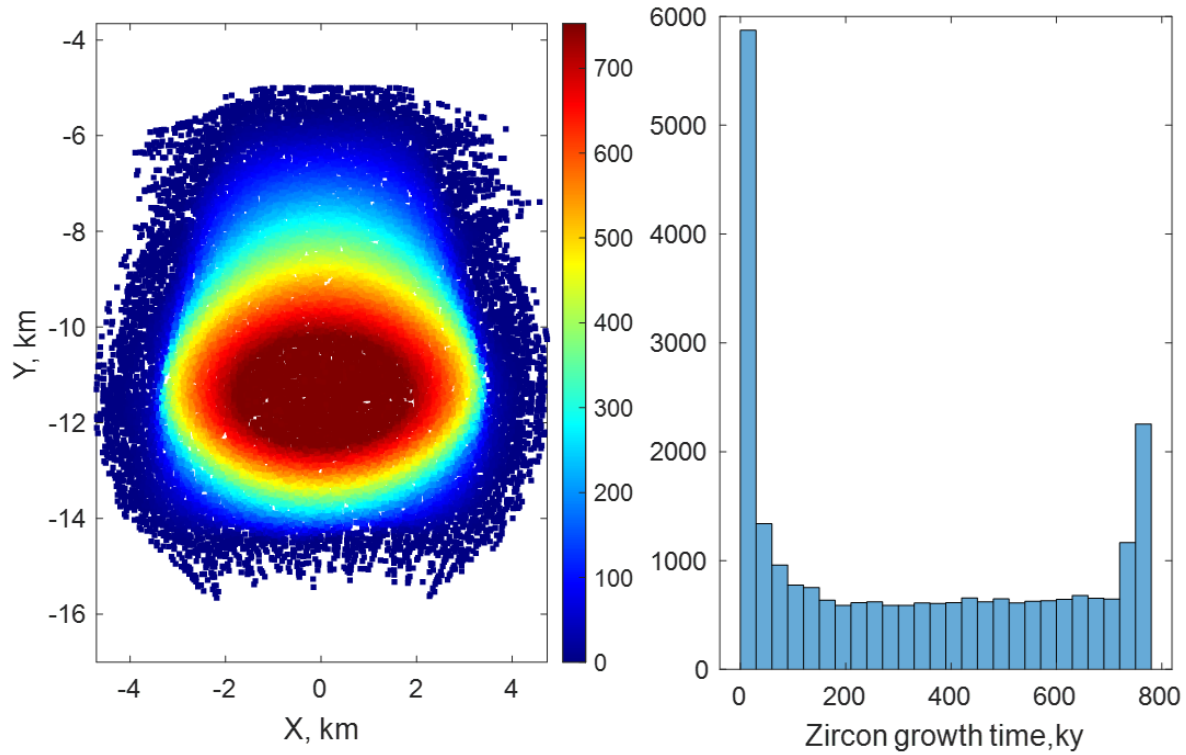
723   **Figure10.** Distribution of the growth time of magmatic zircons within the intrusion,

724   corresponding to case Q=1 m^3/s L=4 km in Fig. 3. Notice distinctly concentric structure due to

725   completely molten core of the intrusion. Zircons near intrusion edges retain great diversity of

726   ages shown in B.

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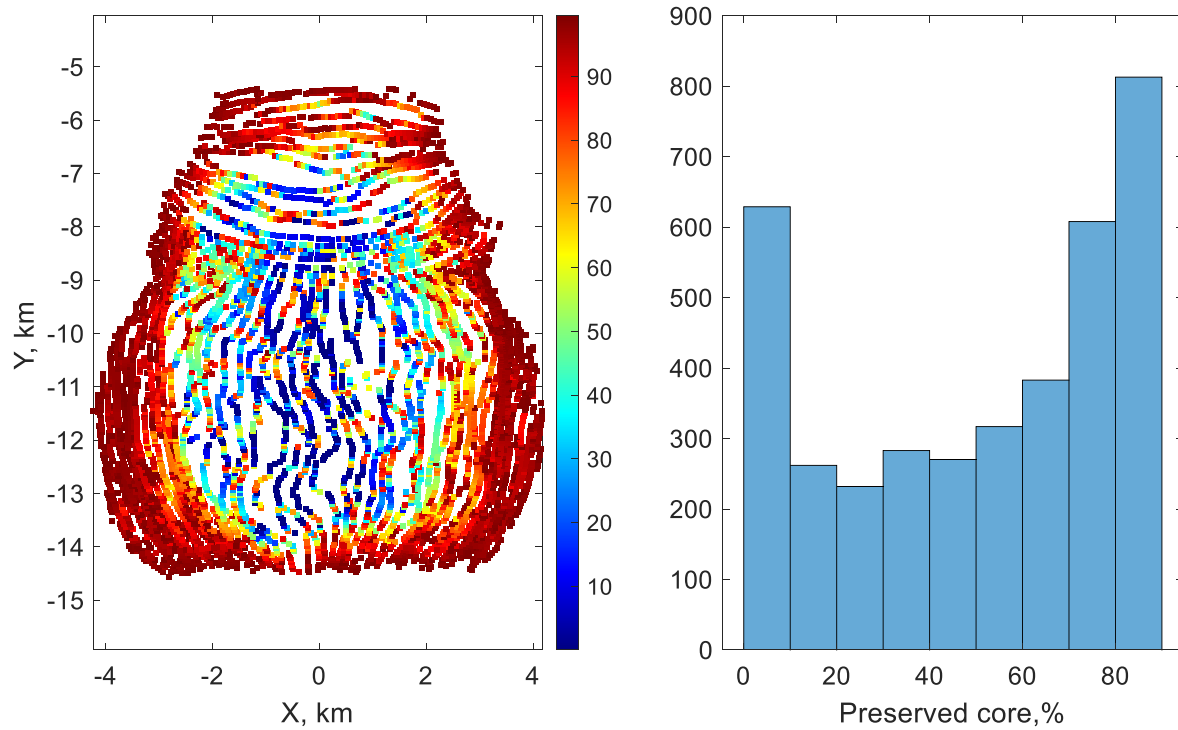
Figure 11. Distribution of zircon core preservation in country rocks affected by dike intrusions.
(a) Notice that rind of the intrusions preserves the original zircons without much dissolution
because these formed early in thermal history, when country rocks were cold; these areas were
displaced to the sides by subsequent intrusions. The core of the intrusion preserves diverse cores.
(b) Number of zircon cores plotted for % preservation.

3.5. Hf and O isotope distribution

Fig 12 considers distribution of oxygen and Hf isotope inside of the growing magma
body, which tracks down percentage of crustal and mantle-magma (Hf isotopes) and % of
crustal, mantle, and hydrothermally altered country rocks involved in intrusion formation. As our
model does not allow magma movement, this modeling signifies potential for naturally
generating isotopic heterogeneities in different areas of the growing magma body. In this sense
each 500x500 m zone will have parcels (batches of melt) that can locally mix (and crystallize
zircons with distinct O and Hf isotopes) in proportion to melts from the above mentioned crustal
and mantle-derived rock parcels. Should these magma batches allowed to further mix along the
laterally and vertically interconnected network of magmatic column of eruptible magma (e.g.,

750 Cashman et al. 2017), such eruption will likely preserve these diversity in the eruptive products,

751 similar to what is observed in many supervolcanic eruptions of the Yellowstone plume

752 (Bindeman and Simakin, 2014; Wotzlaw et al 2014; 2015).

753 However, if such system were not allowed to erupt but are allowed to internally convect

754 over many hundreds of thousands of years in a semi-mushy state, periodically activated by

755 intrusions, this should generate a more homogenous mix of diverse zircons within a single hand

756 specimen.

757 Aging inside of plutons may further anneal isotopic differences, resulting in relatively

758 homogenous (with respect to O and Hf isotopes) magma parcels, which still record long zircon

759 crystallization tails.

760
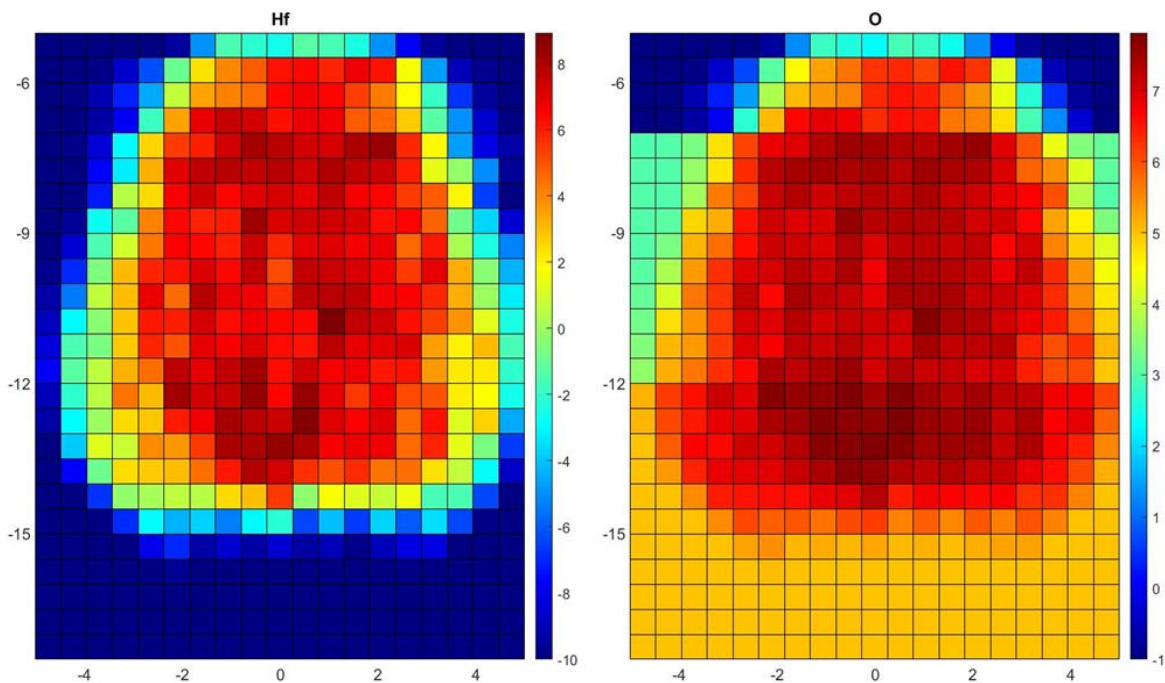
761 Figure 12. Distribution of Hf and O isotopes averaged in 500x500 m grid. Magma Hf +10, O +8

762 **4 Discussion**

763 Formation of crustal magmatic systems involves magma injection from depth, heat and mass

764 transfer, melt and gas separation and episodic volcanic eruption (e.g., McBirney, 2006). It is

765 obvious that no single model can capture complexity of these diverse processes, that requires

766 solution of conjugate fluid-solid-heat problem coupled with and chemical problem (phased

767   diagram, rock compositions). Previously widely applied 1D or 2D cylindrical magma chambers

768   models with underplating (Annen and Sparks, 2002; Dufek and Berganz, 2005; Annen et al.

769   2006, Caricchi et al. 2014; 2016) lead to the development of quantitative intuition of our

770   understanding of the relationship of the magma flux rates and melting efficiencies at different

771   geotherms and depths. In all current models, a simple kinematics is prescribed, as the whole

772   crustal column with previously intruded sills is moving vertically down upon emplacement of

773   each additional sill intrusion, or horizontally sideways upon intrusion of a dike.

774   Previously proposed and widely utilized magma body formation by relying on axisymmetric

775   geometries of intrusion with prescribed position of magma injections and rock associated

776   displacements taught us important first-order trends in magma body formation and evolution.

777   The model and software presented in this paper a new model of 2D magma chamber formation

778   by dike and sill injection and this that allows us and potential users to investigate spatio-temporal

779   details of magma body growth and evolution, including zircon histories. Furthermore, we

780   simulated elastic response to dike and sill injection, realized via spreading of markers which

781   creates accommodation for new intrusions by moving them apart. Incorporation of realistic

782   displacements in the 2D model makes a dramatic difference in the evolution of the temperature

783   field, and melt production, and melt distribution (Fig. 3). We believe that our model provide a

784   realistic and high-resolution description of thermal, compositional, and melt % distribution

785   without relying on prescribed whole-system behaviors.

786   After formation of large and interconnected melt regions, thermal and compositional convection

787   can significantly intensify heat transfer, but this process is not currently considered in this model,

788   or previously published thermal-elastic models. In nature, unmolten rock screens surrounded by

789   magma would likely sink leading to segregation of melt, and magma in large melt fraction areas

790   in between cohesive solid framework would flow. Instead, our model assumes possibility of dike

791   penetration through areas that are >60% molten. Similar behavior is accepted in previously

792   proposed models of Annen et al. (2006), Annen (2009) and Caricchi et al. (2016), especially

793   those including intraplating and overplating. In a certain way, the lack of internal segregation in

794   these models is compensated by the possibility of internal diking, balancing the overall heat and

795   mass balance. While this may appear as locally physically unrealistic, delivery of magma

796   through a melt layer may instead proceed via magma addition, magma mixing and convective

797   advection of heat and mass. The latter have similar overall effect on the thermal evolution of the

798   system as a whole, especially in its central fully molten part. Segregation and magma mixing

799   processes will homogenize melt fractions and temperatures to a greater degree than is shown,

800   while global % melt and temperature evolution will remain similar.

801   The bulleted points below outline broad implications of our new model to <u>magma body</u>

802   <u>formation</u> in nature.

803   1) As expected, magma flux controls the efficiency of magma production via melt

804   accumulation and country rock melting. Equally important is melt focusing, which

805   depends on the width of the injection region. Magma injection with the low discharge

806   rate, or in a wide region, can produce no eruptible melt, while significant volume of melt

807   forms for focused dike emplacement. At higher discharge rates the melt production

808   efficiency increases but the influence of the domain size remains significant.

809   2) The model predicts temperature-time histories of melts formed by crystallization and

810   partial melting of host rocks. We observe highly non-monotonic temperature distributions

811   in space and time, especially in the beginning of magma chamber formation that

812   corresponds to individual dike emplacements into the cold crust. Formation of the

813   significant volume of molten magma later in the course buffers temperature oscillations,

814   leading to "McBirney-type" liquid core magma body.

815   3) Depending on the magma flux, melt focusing generates fully molten cores filled with

816   liquidus magma and surrounded by thick "mushified" rinds and partial melt as predicted

817   for "classic" drawings of textbook magma chambers (McBirney, 2006; Marsh, 1981).

818   Defocused systems, and systems with low flux, generate much less melt and such

819   systems may cool as plutons (e.g., Glazner et al. 2004), and subsequently exhibit a

820   variety of observed intraplutonic processes of mass-redistribution (Bartley et al. 2020)

821   initial stages of which can be understood via our model of randomized dike intrusion.

822   4) Ordinary magma flux rates comparable to those observed in island arcs are capable of

823   rapid generation of large, 10^2 km^3 volumes of silicic melt over ~4×10^3 years, especially if

824   this magma injection is focused in a relatively narrow zone, and mostly due to

825   accumulation of such melts.

826   5) The magma system that is getting formed reminds network of dikes, sills and magma

827   bodies interconnected over large vertical domains in the crust. Such systems are able to

828   provide high silicic magmas discharge rates in the course of supervolcanic eruptions
829   (Sparks et al. 2017) or voluminous Yellowstone lava flows (Loewen et al. 2017).

830 6) Mapping of zones of melt with >50% melt in the model demonstrates that these are
831   distributed in a tortuous pattern with a great deal of both vertical and lateral
832   interconnectivity. Such configuration would enable easy flow of magma averaging out
833   regional heterogeneities on magma flow timescales.

834 7) Melting by basalts or andesites vs melting by hot rhyolites (including addition of
835   rhyolites from below) has surprising similarities in the total volume of generated melt
836   (Fig. 7).

837 Our new model is tested to its ability to generate a) crustal melt and b) total melt, and describes
838 melt volume-time histories of such melts during the growth of the intrusion, and its subsequent
839 cooling. Points below outline specific implications for our model with respect to crustal melting.

840 1) An important and previously unexplored result is in our 2D model is that simulated rock
841   spreading and heat propagation leads to less efficient crustal melting. During the initial
842   stages of intrusion (crustal preconditioning) magma is capable of melting only up to 15-
843   20% of the ambient crust. Subsequent intrusion will have to melt this already diluted
844   crust-dike (e.g. "crust-mantle") mixtures. Furthermore, after the establishing the partial
845   rind zone with 30% crustal melt, subsequent magmatism is dominated chemically and
846   isotopically by the intruded dikes. This result is in line with observations that even in the
847   hot-spot related examples of magmatism, such as Iceland or the Snake River Plain,
848   where high temperature mantle-derived Yellowstone-plume magmas intrude the
849   Archean crust, they assimilate at most 10-20% of such crust based on Nd-crustal index
850   (Nash et al. 2006), and based on earlier reported results of crust-wide thermomechanical
851   modeling (e.g. Colón et al. 2019). Therefore, the high-resolution model in this paper
852   provides a more realistic estimates of maximum possible extent of crustal melting by
853   dike and sills intrusion.

854 2) We thus revise downward the maximum possible "melting efficiency" from previously
855   published models, even though our model includes hot liquidus magmas. The undoubted
856   presence of high-degree or pure crustal melts in the geologic record, commonly with a
857   highly peraluminous nature (Bucholz et al. 2018) negates their generation by dike

36

858   intrusion and must be related to radioactive heating (migmatization and assembly) at

859   deep crustal zones.

860   3) The model also predicts lateral and vertical heterogeneities, due to naturally-generated

861   crustal melts vs newly injected melt, for both the rhyolitic intrusions and andesitic

862   intrusion cases. It identifies areas with >50% melt, potentially eruptible magma, which

863   contain variable proportions of crustal melt vs original melt. In nature this would

864   correspond to a) varying isotopic composition b) variable proportion of inherited zircons

865   c) in the case of andesitic partial melt-silicic remelt, various bulk melt compositions.

866

867 Integration of computed thermal and melting histories with zircon crystallization software

868 allows us to predict zircon dissolution/crystallization conditions in 25,000 markers across the

869 model (Fig. 11) and the dissolution/overgrowth destinies of inherited zircons. Such modeling

870 helps in understanding zircon record in magmatic rocks that range from extremely diverse to

871 extremely homogeneous with respect to age. Examples of extremely diverse zircon record

872 include small magmatic systems such as Mt St Helens (Claiborne et al. 2010a); and large size

873 systems such as Peach Spring (Claiborne et al. 2010b) and the Fish Canyon Tuffs (Wotzlaw

874 et al. 2013). Examples of very homogeneous with respect to age, but very diverse with

875 respect to crust/mantle proportion (diverse O and Hf isotopes) systems include most of

876 Yellowstone hotspot track magmas (Bindeman and Simakin, 2014; Wotzlaw et al. 2014,

877 2015). The bulleted points below outline specific results of our modeling.

878   1) Complex zircon growth patterns occur in the beginning of the magma chamber

879   formation with several episodes of complete or partial dissolution and regrowth.

880   Early injected dikes will cool quickly and stay in the periphery of the forming

881   magma chamber resulting in very restricted ages of zircons corresponding to active

882   diking period.

883   2) At early stages of magma body formation, and later around its periphery, both host

884   rock and magmatic zircons will show a variety of ages and age distribution between

885   cores and rims. Most of host-rock zircons are shifted towards the periphery of the

886   growing magma body, which experience only partial melting followed by further

887   regrowth.

888   3) Zircons in the central part of the intrusion are mostly fully dissolved and
889   reprecipitated, leaving no inherited core. Only small portion of the intrusion keeps
890   some part of an old core.

891   4) Zircons grow and age together with magma cooling and crystallization and thus rim
892   age corresponding to the cooling interval of the intrusion, unless it is interrupted by
893   the eruption of magma. Thus, crystals in the center of the intrusion will have a long
894   growth history spanning a few 100 ky age difference between the core and the rim.
895   This concentric age zoning must be taken into account during interpretation of large
896   volume ignimbrites or plutonic rocks.

897   5) Melting by hotter magmas (basalts and andesites) should lead to more aggressive
898   reset of zircons especially in the central parts of the intrusion due to higher T and M
899   factors; however lateral intrusion heterogeneity and zircon diversity in age and O
900   and H isotopes will persist.

901   6) When % of crustal melt vs dike melt is translated into crustal-mantle O and Hf
902   isotopic end-member values (e.g. Fig. 12) in the whole system, and in areas with
903   >50% melt across the system, we observe natural development of diversity of O and
904   Hf isotopes in different areas within the system.

905   7) Depending on the magma flux and melt focusing two end-member scenarios are
906   possible for zircon record.

907   8) At low magma fluxes and melt defocusing, zirconal record reminds "Fish Canyon
908   type" (Wotzlaw et al. 2013) when large volume of crystal-rich rhyodacitic magmas
909   with 40-45% crystals inherited from variable crust-mantle mixtures, favoring the
910   crust. Many zircons got inherited tailing back to many hundreds of thousands of
911   years signifying long mush residence in the crust (Bachmann et al. 2002). Each hand
912   specimen contains crystals delivered from isotopically diverse (e.g. O and Hf)
913   country rocks including Archean basement. Observed high level of inheritance
914   suggests that incorporated magma volumes were not fully molten. Each hand
915   specimen represents a heterogeneous mixture of these; while laterally, bulk chemical
916   composition is "monotonous intermediate" homogenous mix, suggesting effective
917   remobilization of >50% melt fractions with suspended convection of crystal-melt
918   mixtures.

919   i)   At high magma flux and melt focusing, zirconal record becomes "Yellowstone"

920   type (Bindeman and Simakin, 2014; Wotzlaw et al. 2014; 2015) when large

921   volumes of near liquidus, crystal poor magmas are erupted. These exhibit zircons

922   of identical age, within time-resolution of $\pm500\text{-}1000$ years using the best

923   geochronological methods available CA-ID-TIMS. however, they exhibit extreme

924   diversity of crust/mantle proportions (via O and Hf isotopes in zircons) suggesting

925   that magmas were generated simultaneously in different parts of the fully molten

926   system but with different proportions of crust/mantle there, and erupted without

927   delay. These observations suggest that pockets of melt with highly diverse crust-

928   mantle proportions got rapidly heated above zircon saturation, zeroing the ages,

929   followed by zircon crystallization in the same isolated pockets. This was followed

930   by a rapid assembly of such pockets and eruption. Time between crystallization of

931   zircons is $<\sim5000$ years.

932

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2 *[Journal of geophysical research. Solid Earth]*

3 Supporting Information for

4 **Magma chamber formation by dike accretion and crustal melting: 2D thermal model**
5 **with emphasis on zircon record**

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11 **Contents of this file**
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13

14 **Introduction**

15 Supporting information provides phase diagrams for andesitic and rhyolitic magmas that were
16 used in the simulation and additional figures for andesitic magma injection into the granitic
17 crust. All discussion and references to figures are provided in the main text.
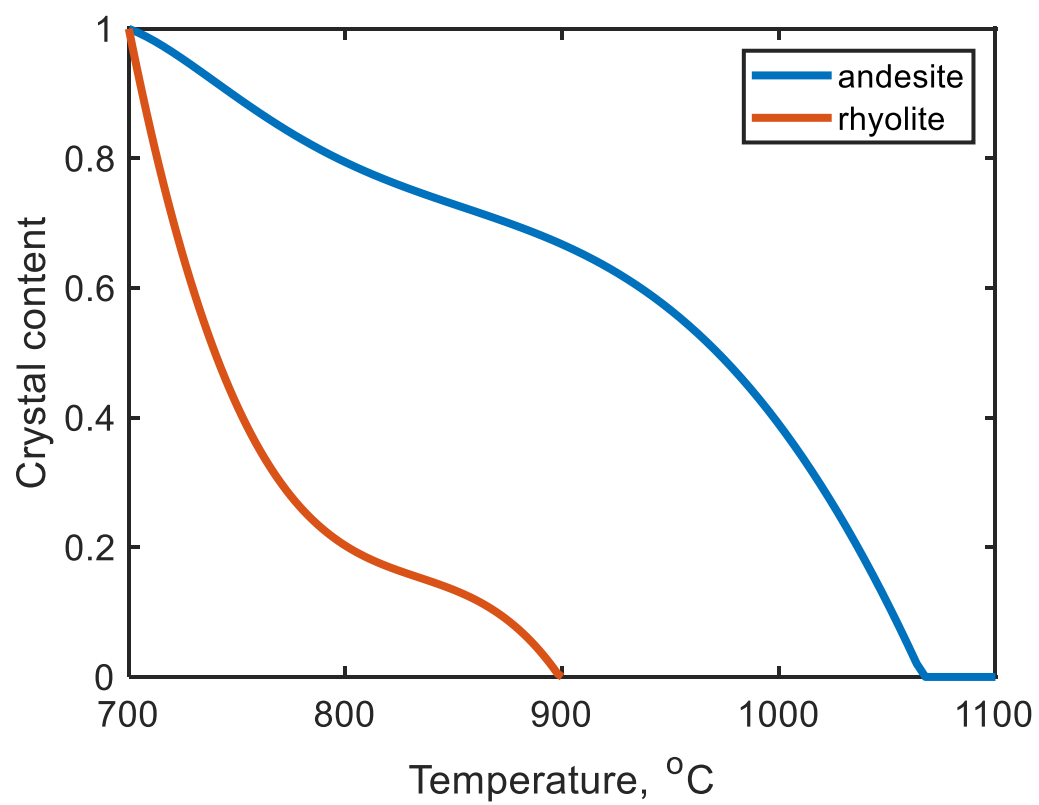
**Figure S1.** Phase diagrams used in the simulations.

Melt fraction   **Temperature, ⁰C**

a – Q= 0.5 m³/s, W=4 km
b – Q= 1 m³/s,   W=2 km
c – Q= 1 m³/s,   W=4 km
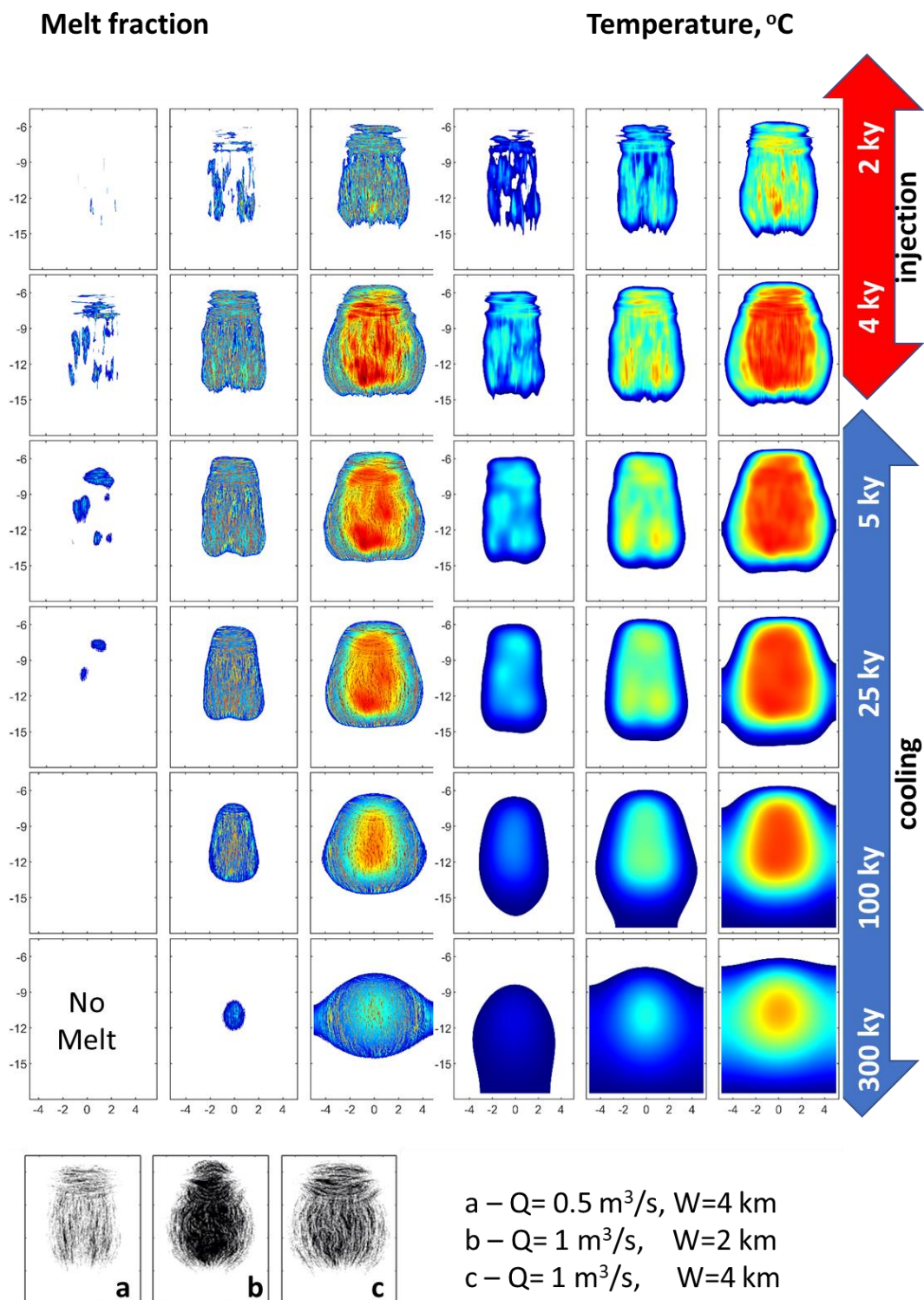
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22   **Figure S2.** Met fraction and temperature distribution for andesitic magma injection in granitic
23   crust.

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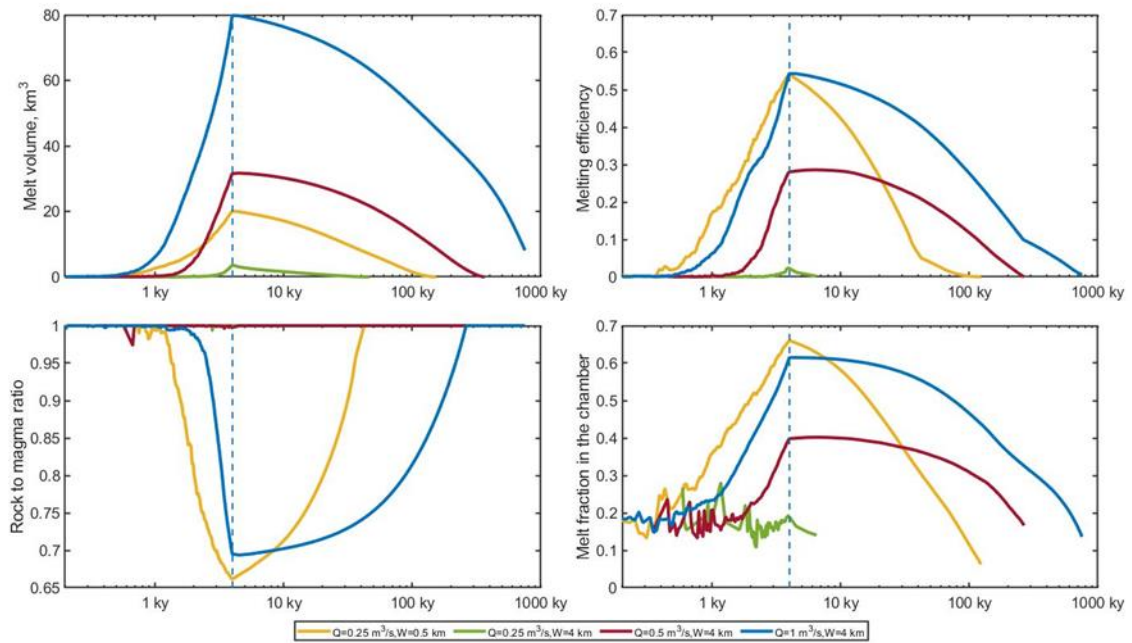
Figure S3. Melt volume (A), melting efficiency defined as the volume of eruptible melt divided
by the amount of injected magma (B), rock to magma ratio in eruptible magma (C) and
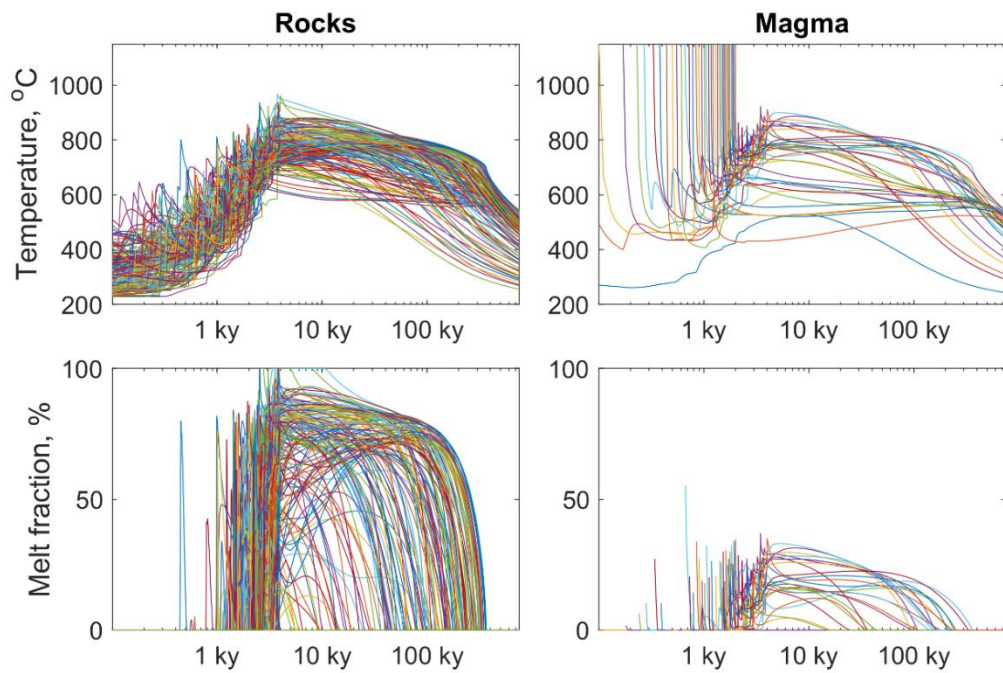averaged melt fraction in the magma chamber.

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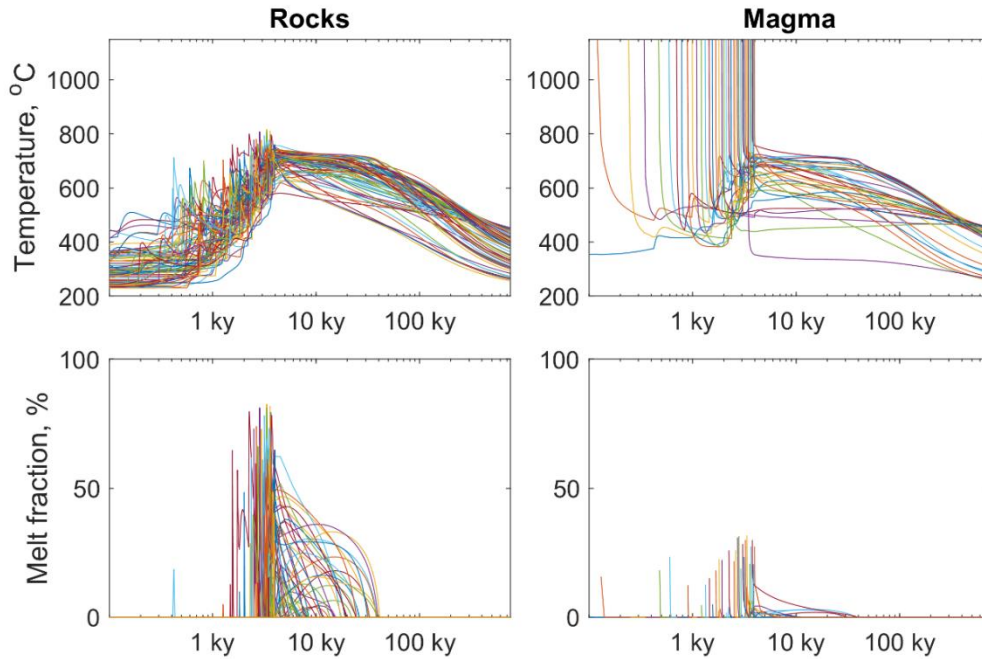
Figure S4a. Temperature histories of rock and andesitic magma particles for Q=0.25 m³/s,
W=4 km.

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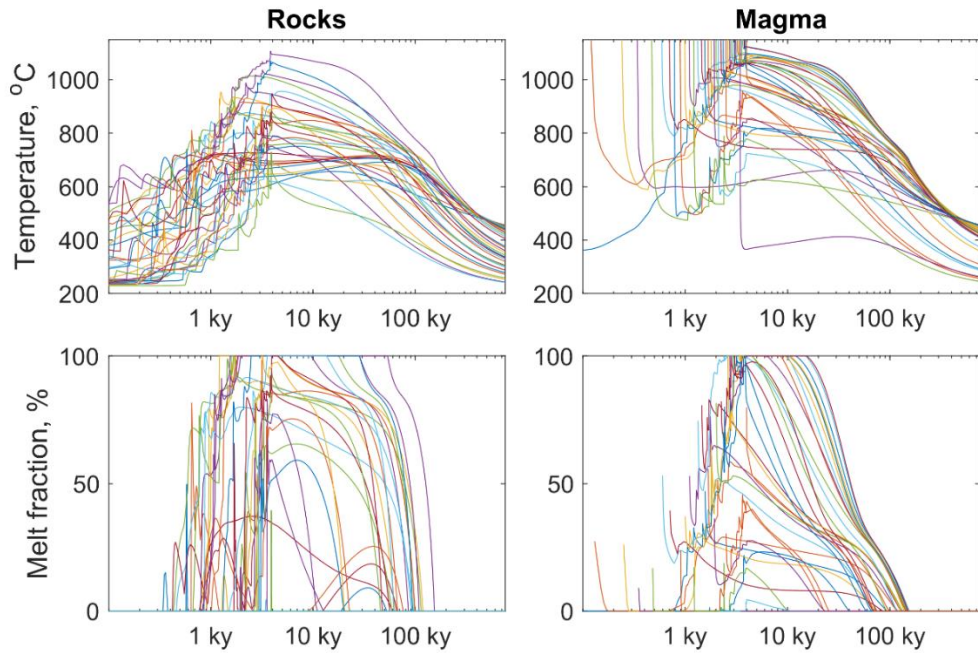
Figure S4b. Temperature histories of rock and magma andesitic particles for Q=0.25 m³/s, W=0.5 km.
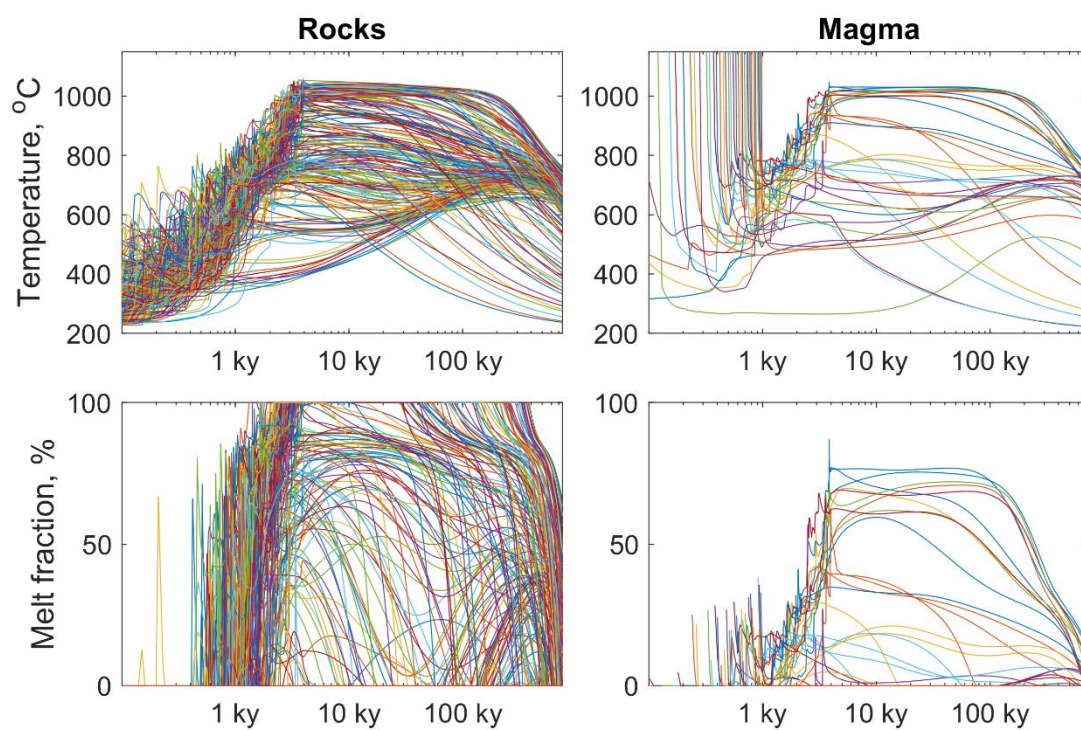

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Figure S4c. Temperature histories of rock and andesitic magma particles for Q=0.5 m³/s, W=4 km.

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40 **Figure S4d.** Temperature histories of rock and andesitic magma particles for Q=1 m³/s, W=4
41 km.

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