Melt Evolution and Residence in Extending Crust: Thermal Modeling of the Crust and Crustal Magmas

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Abstract

Tectonic extension and magmatism often act in concert to modify the thermal, mechanical, and chemical structure of the crust. Quantifying the effects of extension and magma flux on melting relationships in the crust is fundamental to determining the rate of crustal melting versus fractionation, magma residence time, and the growth of continental crust in rift environments. In order to understand the coupled control of tectonic extension and magma emplacement on crustal thermal evolution, we develop a numerical model that accounts for extension and thermal-petrographic processes in diverse extensional settings. We show that magma flux exerts the primary control on melt generation and tectonic extension amplifies the volume of melt residing in the crustal column. Diking into an extending crust produces hybrid magmas composed of 1) residual melt remaining after partial crystallization of basalt (mantle-derived melt) and 2) melt from partial melting of the crust (crustal melt). In an extending crust, mantle-derived melts are more prevalent than crustal melts across a range of magma fluxes, tectonic extension rates, and magmatic water contents. In most of the conditions, crustal

temperatures do not reach their solidus temperatures to initiate partial melting of these igneous lithologies. Energy balance calculations show that the total enthalpy transported by dikes is primarily used for increasing the sensible heat of the cold surrounding crust with little energy contributing to latent heat of melting the crust (maximum crustal melting efficiency is 6%). In the lower crust, an extensive mush region develops for most of the conditions. Upper crustal crystalline mush is produced by continuous emplacement of magma with geologically reasonable flux and extension rates on timescales of 10^6 years. Addition of tectonic effects and non-linear melt fraction relationships demonstrates that the magma flux required to sustain partially molten regions in the upper crust is within the range of estimates of magmatic flux in many rifting regions ($\sim 10^{-4}$ to 10^{-3} km³/yr) and at least an order of magnitude lower than previous modeling estimates. Our results demonstrate the importance of tectonics in augmenting melt production, composition, and crustal evolution in active magmatic systems.

Keywords: *thermal model; tectonic extension; crustal melting; fractionation; crystal mush; magma longevity*

1. Introduction

The combination of extensional tectonics and magmatic intrusion has long been recognized to alter the thermal and structural properties of the crust in rifting environments (e.g., Brown and Solar, 1998; Callot et al., 2001; Ebinger and Casey, 2001; McKenzie and Bickle, 1988; Mohr, 1982; Sengör and Burke, 1978). These two processes are strongly coupled because: 1) both crustal thinning and magmatic intrusions perturb the geothermal gradient, 2) tectonic extension favors vertical propagation of magma

following the path perpendicular to the least principal stress direction (e.g., Anderson, 1951), and 3) extensional processes in the crust can control the growth and emplacement of magmatic bodies (e.g., Corti et al., 2003; Gudmundsson, 2006; Jellinek and DePaolo, 2003). However, the interplay of long-term crustal extension and magma intrusions and their exact control on crustal growth, compositional evolution, and thermal evolution remains poorly constrained (e.g., Dufek and Bergantz, 2005; Fitton et al., 1991; Karlstrom et al., 2010; Thompson and Connolly, 1995).

A key question is how the mass and energy are balanced in the crust in response to magma emplacement and extensional tectonics. In many rift environments, the origin, longevity, and location of melt in the crust remain controversial (e.g., Deering et al., 2011; Schmitt and Vazquez, 2006). Numerical and geochemical models have suggested either partial melting of the crust (e.g., Bindeman and Simakin, 2014; Conrad et al., 1988; Fyfe, 1973) or fractional crystallization of the parental magma (e.g., Baker et al., 1977; Barberi et al., 1975; Jagoutz, 2010; Mahood and Baker, 1986; O'Hara, 1977) as the main process that explains crustal layering and petrologic diversity observed in volcanic settings. It is likely that both processes operate to different degrees (e.g., DePaolo, 1981b), but in many systems quantitative understanding can remain elusive.

Understanding the physical parameters that determine crustal melting versus residence of mantle-derived magmas in the crust has a bearing on the mass balance of rifts and the generation of new crustal material (e.g., Condie, 1982; Rudnick and Fountain, 1995; White et al., 2006).

A closely related issue is that of melt residence time in crustal domains. Observational evidence from exposed plutons and erupted volcanics constrains our

current understanding of crustal thermal evolution. Long residence times at low melt fractions have been often argued for crustal magmas (e.g., Bachmann et al., 2007; Hildreth, 2004; Reid et al., 1997). In opposition to this, others have argued that, if the magma bodies do not erupt on relatively short timescales, then they must completely solidify to form plutons (e.g., Glazner et al., 2004; Tappa et al., 2011). The key to understanding the generation and thermal and compositional evolution of crustal magmas is 1) determining the thermal budget of these systems, and 2) assessing the relative amount of crustal melt versus fractionated melt derived from the mantle.

Much of the energy driving these crustal processes derives from mantle melting. While there is a first-order understanding of the relationship between mantle upwelling and melt generation in the mantle, the amount of melt that enters the crust depends on complex non-linear dynamic processes during mantle melt generation, segregation, and transport (e.g., Liang and Parmentier, 2010). These processes are influenced by mantle heterogeneities and several other parameters such as rheology, rift duration, rifting velocity, mantle potential temperature, mechanical boundary layer thickness, decompression rate, presence of volatiles, strain rates, porosity, compaction length, separation velocity, rupture, and dike opening (e.g., Bialas et al., 2010; Brown and Solar, 1998; Buck, 2004; Foucher et al., 1982; McKenzie, 1985; McKenzie and Bickle, 1988; Pedersen and Ro, 1992; Ruppel, 1995; White et al., 1987; Ziegler and Cloetingh, 2004). The complex processes in the mantle lead to significant differences in the timing and volume of magmatism in different rifts and segments, ranging from minor magmatic contribution to intense volcanic activity (e.g., Ruppel, 1995). An open question is how

93 this variability in tectonic rates and mass and energy flux affect the thermal and
94 compositional evolution of crustal magma.

95 While in many settings both tectonics and magmatic intrusions clearly influence the
96 thermal state of the crust, these processes are often treated in isolation. Geophysical and
97 geodynamic models typically focus on deformation and thermal changes associated with
98 intrusion events and extensional tectonics (e.g., Baer et al., 2008; Corti et al., 2003).

99 Petrologic and thermodynamic studies often focus on the detailed thermal history of
100 magmatic regions by constructing models that consider magmatic intrusion,
101 differentiation, and crustal melting using either a single intrusion (e.g., Bergantz, 1989;
102 Huppert and Sparks, 1988) or multiple intrusion events (e.g., Annen et al., 2006; Annen
103 and Sparks, 2002; Dufek and Bergantz, 2005; Pedersen et al., 1998; Petford and
104 Gallagher, 2001; Wells, 1980). However, the thermal state of the crust and crustal
105 magmas have not been examined by models that include the interplay of varying tectonic
106 extension rates and detailed thermodynamic and petrologic processes.

107 To investigate the long-term thermal evolution of the crust that is subjected to
108 tectonic extension and dike injections derived from the mantle, we present a thermo-
109 mechanical numerical model that solves for coupled kinematic and thermodynamic
110 processes. This model is focused on crust-magma interaction, and we do not replicate
111 detailed mantle processes including generation, segregation, and transport of the melt but
112 rather we focus on the heat transfer processes in the crustal domain as a result of variable
113 extension rates and magma input. We systematically examine a parameter space for
114 magma flux and tectonic extension to quantify the length- and time-scales of magma-
115 crust interaction. We also use two different magmatic water contents to examine crustal

rifts both above relatively wet and dry mantle. The model accounts for crustal displacement due to intrusions and for extensional tectonics through a specified kinematic condition. In this way, our model differs from prior thermal-petrologic models. We examine the crustal growth and thermal evolution of crustal magmatic systems and give estimates for the origin, location, fraction, temporal evolution, and amount of melt that can be generated in the crustal section. Our model is applicable to rift environments with a pre-existing continental crust.

2. Thermo-mechanical model

A two-dimensional thermal model has been developed to simulate tectonic extension of the crustal column and repetitive magma emplacement in the crust to evaluate the crustal heat transfer and thermal evolution of magmas. This model was modified from the work of Dufek and Bergantz (2005) to account for extensional tectonics. In the model, heat transfer is dictated by two processes: 1) basaltic injections that are emplaced in the crust and transport mass and enthalpy from the mantle and 2) tectonic extension of the crust that accounts for crustal thinning and advects heat away from the axis of extension by displacing the previously heated material. Embedded in the model are parameterized thermodynamic relationships determined with the rhyolite-MELTS thermodynamic software (Gualda et al., 2012) to monitor the geochemical evolution of the system and to accurately partition the latent and sensible heat.

Basaltic injections are modeled as multiple dike and sill intrusions emplaced at different levels in the crust. We assume that each intrusion is emplaced in a stochastic manner; that is, the lengths and timing of intrusions into the crust follow a uniform

distribution and thus conform to a specified average flux of magma (see supplemental material). Dikes can extend at all levels in the crust and have widths of 50 m, which is comparable with several field observations (e.g., Williams et al., 1995). The sills are relatively smaller in size (100 m thick and 500 m wide) and are emplaced at the tip of dikes.

We treat the intrusion process as instantaneous as dike emplacement is relatively rapid (e.g., Petford et al., 1993; Rubin, 1992) compared to the much longer crustal thermal evolution considered in this study. The dikes and sills intrude the crust by displacing (effectively advecting) crustal material to accommodate their volume within the crust and cool conductively between successive intrusions. Conduction of heat between the intrusions and the surrounding crust can result in melting of the crustal material and cooling and crystallization of the intrusions. We refer to partially molten crust as the ‘crustal melt’ and the residual melt after partial crystallization of dikes as the ‘residual mantle-derived melt’. At each time step, we investigate the volume and distribution of residual and crustal melts separately.

Tectonic extension in the model is specified as a kinematic condition and satisfies conservation of mass. Feedback between thermal evolution, rheology, and strain rate is not considered in the present investigation nor do we consider the feedback between the crustal stress field and diking (e.g., Callot et al., 2002; England, 1983; Lynch and Morgan, 1990). However, by treating the kinematic condition independently we can examine a wide range of extension conditions and examine the sensitivity of the thermal and melting conditions relative to intrusion scenarios that lack deformation. The kinematic condition advects heat in the crust away from the axis of extension

continuously and is motivated by geodetic estimates from several rifting environments (e.g., Papazachos et al., 1992), discussed in section 2.2. The extension of the crust in this two-dimensional domain is accommodated by thinning of the crust in a near-axis deformation zone (having a half-width of 10 km, Fig. 1), and generates mantle upwelling to replace the displaced crustal material. Due to upwelling, crustal thickness in the model changes continuously over time, which increases the crustal temperatures as the crust thins.

The simplified model geometry is shown in Fig. 1. Although we included the uppermost mantle to treat evolving crustal thickness, our work focuses on how the crustal thermal structure interacts with the basaltic injections and tectonics; therefore our thermal calculations focus on the upper 30 kilometers. We use this structure as a base case for model comparison and to be generally applicable to a number of natural examples, where the continental crustal thickness typically varies between 10 to 35 kilometers in extensional settings (e.g., Platt, 2007; Zhu et al., 2006).

We assume tonalitic composition for the initial uppermost 10 km of the model as a proxy for a more silicic upper crust (e.g., Wolf and Wyllie, 1994). The remaining 20 km of the crust in the model is treated as amphibolite (e.g., Helz, 1982; Wolf and Wyllie, 1994) as a proxy for a relatively fertile end-member, similar to observations (e.g., Rudnick and Fountain, 1995) and experimental work (e.g., Beard and Lofgren, 1991; Rapp et al., 1991; Sen and Dunn, 1994; Wolf and Wyllie, 1994).

2.1. Time-dependent thermal modeling and parameterization

The initial steady state thermal profile of the crust is determined using radiogenic

heat production estimates ($\sim 8.0 \times 10^{-10}$ W/kg) and initial surface heat flux of 70 mW/m² (see supplemental material). The tectonic extension and injection of magma at its liquidus perturb the steady-state thermal profile of the crust. We use a two-dimensional, transient thermal model that accounts for conduction and advection by solving the conservation of thermal energy equation

$$r_{mix} c_p \frac{\partial T}{\partial t} + r_{mix} c_p v_i \frac{\partial T}{\partial x_i} + r_{mix} L \frac{\partial f}{\partial t} = k \frac{\partial^2 T}{\partial x_i^2}, \quad i=1, 2 \quad (1)$$

where, the velocity (v) of solid material is given by the kinematic deformation condition. The symbols are explained in Table 1. We discretize equation 1 using the finite volume approach described in Patankar (1980). The cooling of the basalt is treated as conductive in these large-scale simulations (low Biot number, Carrigan, 1988). We neglect internal convective processes and radiogenic heat production in the transient calculations (see supplemental material).

We use a non-linear relationship between the melt fraction and temperature (Fig. 2) using parameterized solutions of rhyolite-MELTS software and experiments. For intruding dikes, we use the basaltic andesite composition from Mt. Shasta primitive lava (sample #85-44, Müntener et al., 2001). For amphibolite, we use experimental results from a number of studies (e.g., Rushmer, 1991; Sisson et al., 2005; Wolf and Wyllie, 1994). We use experimental results by Piwinski and Wyllie (1968) for tonalite.

One of the important components of our model is the application of predictor-corrector algorithm following Voller and Swaminathan (1991) to calculate the numerical convergence for non-linear melt fraction versus temperature relationships gathered from rhyolite-MELTS solutions. This algorithm successfully partitions the total enthalpy change in the system into latent heat and sensible heat during phase changes at each

timestep (Petcovic and Dufek, 2005; Voller and Swaminathan, 1991).

Melting of the crust and the state of the intrusions are limited by the thermodynamic properties of the materials and the volume and heat transported by the injections. Following the study of Dufek and Bergantz (2005), we calculate the efficiency of crustal melting as the ratio between the volume of crustal melt based on numerical calculations to the volume of crustal melt based strictly on enthalpy balance arguments (assuming no heat is lost heating the crustal parcels that do not melt). This approach is useful for model comparison where assumptions of lithology and melting properties vary.

This is formulated by

$$E\% = 100 \cdot V_{crust}^{mod.} / V_{crust}^{eff.}, \quad (2)$$

$$V_{crust}^{eff.} = r_m V_m \left(C_p^m \Delta T_m + L_m \right) / r_c \left(C_p^c \Delta T_c + L_c f \right), \quad (3)$$

where, $V_{crust}^{mod.}$ is the volume of crustal melt that is generated when the enthalpy released from basaltic dikes is diffused to the crustal section, and $V_{crust}^{eff.}$ is the volume of crustal melt when the enthalpy released from basaltic material is used only on the crustal material that melts. Subscripts m and c refer to mantle-derived and crustal materials, respectively.

2.2. Parameter space

We varied extensional rates, magma input, and water content of the intruded magma with the same initial crustal conditions and examine the resulting thermal profile and melting relations in the crusts of different rift environments. Although different initial assumptions will result in variable melting characteristics, constant initial

conditions allow for comparison of model sensitivity to melt flux and extension rates. We varied the extension rates between 2, 5, 7, 10, and 15 mm/yr. The extension values of 2 mm/yr and 15 mm/yr are comparable to β values of ~ 1.5 and ~ 4 in McKenzie and Bickle (1988), respectively. This range is also comparable to estimated rates in different extensional settings (~ 1 to ~ 13 mm/yr, e.g., Bilham et al., 1999; Brothers et al., 2009; Dewey and Lamb, 1992; Morgan et al., 2008; Papazachos et al., 1992; Stamps et al., 2008). The simulations are conducted for up to 2 million years to explore the response of the crust to tectonic extension and dike intrusions over long time periods.

We varied the mean period of time between successive intrusion events from 15 to 100 kyr for each separate set of simulation. We also simulated a set of simulation with no magma flux (see supplemental material). Our choice of flux values corresponds to a range of 0 to 10^{-2} m³/m² per year and brackets the magma flux estimates suggested in seismic, gravity, and thermal studies (10^{-4} to 10^{-3} m³/m² per year, e.g., Courtillot and Renne, 2003; Crisp, 1984; Dimalanta et al., 2002; Grunder, 1995). These observations probably underestimate the rate of magmatic input (e.g., Jagoutz and Schmidt, 2013; Jicha et al., 2006) because thermal estimates usually assume efficient enthalpy transfer (e.g., Dufek and Bergantz, 2005) and geophysical calculations neglect some mechanisms, such as sediment subduction, tectonic erosion, and delamination/foundering (e.g., Ducea, 2002; Jagoutz and Schmidt, 2013; Kay and Kay, 1991; Plank and Langmuir, 1998). The flux values of 0.001 and 0.01 m³/m²/yr correspond to melt thicknesses of 1 km and 10 km per million years, which is also comparable to melt generation thickness estimates in McKenzie and Bickle (1988), within β values of ~ 1.5 to 4.

The effect of initial magmatic water content on the crystallization path of the

intruded melt is depicted in Fig. 2. In the thermodynamic model, amphibole first appears around 0.40 melt fraction compared to 0.47 melt fraction in the melting experiments of Müntener et al. (2001) at 12 kbar with ~5 wt% H₂O. Due to limited amphibole solid-solution data, the amphibole thermodynamic model only includes pargasite and potentially underpredicts amphibole stability in the hydrous end-member. However, the difference appears to be in an acceptable range (for calculating melt volumes) between rhyolite-MELTS calculations and Müntener et al. (2001) experiments. In our model, we use 0.5 wt.% H₂O (referred to as dry injections) and 5.6 wt.% H₂O (referred to as wet injections, comparable to sample #85-44 B686, Müntener et al., 2001) for the dikes to bracket the range observed in different rift settings (0.5 wt.% to ~6 wt.% H₂O, e.g., Dunbar et al., 1989; Moore and Carmichael, 1998; Roggensack, 2001; Stolper and Newman, 1994; Walker et al., 2003).

3. Results

We consider both the long-term crustal thermal evolution that is a result of cumulative intrusions and extension and short-term interactions associated with individual intrusions (Fig. 3). We quantify the volume of crustal and mantle-derived melts, melt fraction, longevity of connected melt pockets, and the efficiency of crustal melting as a function of magma flux, extension rate, and magmatic water content.

3.1. Melt thickness in the crust over time

To investigate the effect of magma flux and tectonics on crustal magmas, we examine: 1) a fixed tectonic extension rate with varied magma flux and water content, 2)

a fixed magma flux with varied extension and water content, and 3) varied tectonic extension, magma flux, and water contents (Figs. 4-6). Similar to earlier thermal models on melt emplacement in the crust (e.g., Annen et al., 2006; Dufek and Bergantz, 2005), we observe that the amount of melt in the crust is strongly controlled by the magma flux. For a given extension rate, the amount of crustal and residual mantle-derived melt increases with time and increasing flux because progressive accumulation of mantle-derived magma transports mass and enthalpy to the crust (Fig. 4). Mantle-derived melt progressively accumulates in the crust over time when the injection frequency is faster than the time required for conducting the heat away from the intrusion (e.g., Dufek and Bergantz, 2005; Petford and Gallagher, 2001). Due to enthalpy transfer by the intruded melt, the temperatures of crustal lithologies can increase above their solidi and partial melting processes initiate. Eventually, partially molten regions in the crust become a mixture of mantle-derived and crustal melts compositionally. In all conditions examined, the mantle-derived melt is significantly greater than the crustal melt although there is considerable variability (Fig. 4). The maximum amount of total melt is generated in 2 million years when the flux is $10^{-2} \text{ m}^3/\text{m}^2/\text{yr}$. In these conditions, the crustal melt to residual mantle-derived melt ratio is about 1:25 in environments with wet injections and about 1:3 in environments with dry injections.

We examine the effect of water content in all simulations due to its role in modifying the thermal and compositional evolution of magma (e.g., Green, 1972). As depicted in Fig. 4, in wet injection scenarios, the thickness of residual mantle-derived melt is higher (twice in maximum conditions) and crustal melt is lower (1/4th in maximum conditions) than dry injection scenarios. Such behavior reflects the

crystallization behavior of the injected basalt as a function of water content: 1) the liquidus temperature of dry magma is higher than wet magma and thus can induce greater rates of crustal melting and 2) the minerals in the wet magma (e.g., amphibole) are stable in their melt phases at lower temperatures, therefore the melt fraction of wet magma is higher than the dry magma at low temperatures (Fig. 2).

Extension of the crustal column results in mantle upwelling and heat advection by replacing the crustal material, which promotes thicker melt columns (Fig. 5). Extension transports the enthalpy laterally by advecting the heated parcels of the crust, and can increase the amount of melt generated in the crust by up to 5 times when the magma flux is intermediate. Here, we note that the mantle upwelling changes the boundary conditions but not the flux of melt as it is treated independently. In maximum extension conditions (0.015 m/yr extension at 2 million years), the ratio of crustal melt to residual mantle-derived melt thickness is about 1:50 in wet injection scenarios. This ratio is about 1:5 in the dry injection scenarios. The decrease in the amount of crustal melt after ~1.2 million years is attributed to the concentration of the extension and diking at the center of the domain. In the cases of high extension and high flux, the center of the domain is saturated with mantle-derived melt because the crustal material is pushed away by the extensional forces over time. Therefore, further melting of the crust is limited after ~1.2 million years. Similar to previous observations (Fig. 4), the residual mantle-derived and crustal melt amounts are strongly dependent on the water content of the injections. The ratio of crustal melt thicknesses in wet injection to dry injection scenario is 1:4. This ratio is about 2:1 for residual mantle-derived melt thickness.

When both magma flux and tectonic extension are maximized, large accumulations

of melt develop in the crust over time (Fig. 6). In the highest flux and extension conditions, the maximum melt thickness with wet intrusions is almost two times larger than the dry conditions. The influence of tectonic extension on melt generation is more prominent when the flux increases ($>0.004 \text{ m}^3/\text{m}^2/\text{yr}$ in wet and $>0.007 \text{ m}^3/\text{m}^2/\text{yr}$ in dry cases). In addition to the mantle upwelling, high magma flux at the center of the domain heats the crust significantly and extension advects the enthalpy in the crustal column. In comparison with the static models, addition of tectonic extension calculations implies that lower basalt fluxes can generate the same amount of melt volumes in the crust.

3.2. Melt residence in the crust

3.2.1. Mush regions in the lower and upper crust

An extensive area of crystal-rich mush is developed in the lower crust with high and intermediate flux at intermediate tectonic extension (Fig. 7). In wet injection scenarios, the melt fraction is higher in general and covers a more extensive area compared to dry injection scenarios. The melt fraction increases towards the center of the mush body along the axis of extension and is maximum at the base of the amphibolitic lower crust. This is also true for the tonalitic upper crust when the conditions permit upper crustal mush generation. In both cases, the partially molten region that straddles the center of the domain buffers cooling through the release of latent heat, and therefore, preserves the heat at the center.

The upper crust starts to develop a persistent melt domain in wet and high flux conditions after approximately one million years (Fig. 7a, c), although periodically small bodies form and solidify in the upper crust with shorter incubation periods. Additionally,

in dry injection conditions with intermediate flux, an upper crustal crystal-rich mush develops after two million years (Fig. 7d) due to high liquidus temperatures of the dry injections. This suggests that it is possible to create sustained upper crustal melt regions with geologically reasonable basalt fluxes ($>0.006 \text{ m}^3/\text{m}^2/\text{yr}$) on million year timescales. The upper crustal mush is larger and has higher melt fraction in wet injection scenarios compared to dry scenarios.

3.2.2. Melt storage in the crust

Although reasonable flux and extension rates can create extensive areas of melt in the crust, these melt regions often include very low melt fraction and/or can reside in multiple levels in the crust. In order to estimate the evolution, longevity, and continuous size of the melt bodies in the crust, we employ a metric that assumes that neighboring melted sections with melt fractions greater than 0.25 can combine to form individual melt storage systems. Our choice here represents melt exchange for moderate melt fraction in a crystalline (50-75% crystallinity) mush region (e.g., Bachmann and Bergantz, 2004; Mangan and Marsh, 1992; Marsh, 1988).

The water content of the injections exerts an important control on the formation of crustal magma bodies (Fig. 8). A large partially molten region dominates in the crust in wet injection conditions (Fig. 8a, b). This melt is located in the lower crust (Fig. 7a, b) and its growth rate accelerates after 0.5 million years. The temperature of the lower crustal section increases in response to the enthalpy provided by the dike intrusions and the area of melt extend over 250 km^2 (high flux) and 100 km^2 (intermediate flux) in 2 million years (Figs. 8a, b). These estimates are consistent with intrusive volume

estimations in the crust of Nicaragua (90 to 180 km³/km/Ma, Morgan et al., 2008) and Kenya rift zone (250 to 500 km², Morley, 1994).

In time, the surrounding small, isolated melt bodies either solidify or join the large partially molten zone. In the simulations with dry intrusions and the highest flux (Fig. 8c), there are a number of individual melt pockets; however, all of them are relatively small in size (<20 km²). The melt pockets are small (<1 km²) with dry intrusions and intermediate flux (Fig. 8d).

3.3. Efficiency of crustal melting

The crustal melting efficiency calculations show the relationship between the total heat transferred from the mantle by dikes and resulting melting in the crust. In all cases, the efficiency is low and means that almost all of the enthalpy from the intrusions is consumed to increase the sensible heat of the crust (Fig. 9). The maximum efficiency of crustal melting in an environment with wet injections is only ~1% with high flux and ~0.3% with intermediate flux (Fig. 9a, b). In dry injection cases, the maximum efficiency of crustal melting is relatively higher, being ~6% and ~4% when the flux is high and intermediate, respectively (Fig. 9c, d).

Peak crustal melting is observed over a limited time window and is constrained by tectonic extension. This is a result of how we define crustal and mantle-derived materials: we refer to newly crystallized intrusions as mantle-derived material, and the older crust present at the initiation of the simulations as the crustal material. In high flux and extension conditions, the enthalpy from the repeated intrusions that intrude from the center can only melt the previously intruded and solidified mantle-derived material.

Therefore, over time, the domain is saturated with mantle-derived material and diluted with the crustal material.

4. Discussion

4.1. How do extensional processes affect melt location and residence time in the crust?

The advection of crustal material (in response to thinning) enhances melt residence by increasing the thermal gradients compared to previous melting calculations into a static crust (e.g., Annen et al., 2006; Annen and Sparks, 2002; Dufek and Bergantz, 2005; Petford and Gallagher, 2001). Our work shows that the role of extensional processes in augmenting melting is most prominent when also coupled to high magma fluxes ($\sim 0.01 \text{ m}^3/\text{m}^2/\text{yr}$, Fig. 6). Under these conditions, tectonic extension amplifies the melt residence in the crust significantly (~ 3 times thicker melt columns), owing to mantle upwelling and heat advection. The effect of extension is even higher in low fluxes ($\sim 0.001 \text{ m}^3/\text{m}^2/\text{yr}$), producing >10 times thicker melt columns when the extension increases from 0.002 m/yr to 0.015 m/yr . However, these extension rates coupled with low melt fluxes cannot create extensive partially molten regions (0.7 km in wet and 0.25 km in dry scenarios) in one million years compared to thick melt columns ($>5 \text{ km}$) that can be generated by intermediate to high fluxes.

Therefore, while extensional processes have a significant effect on the thermal evolution of the crust, extension alone is not sufficient to create major crustal magmatic systems. Below a flux value ($\sim 0.001 \text{ m}^3/\text{m}^2/\text{yr}$ in wet and $\sim 0.004 \text{ m}^3/\text{m}^2/\text{yr}$ in dry cases), $>97\%$ of the emplaced melt solidifies between successive intrusions when extension is low (0.002 m/yr). This is due to the competing effects of: 1) the cooling timescale of the

intruded basalt and 2) the period of intrusions that carry enthalpy from the mantle (e.g., Petford and Gallagher, 2001). If the magma flux is high, the heat supply to the magmatic system by dikes will be faster than the heat lost by diffusion. Additionally, the temperature difference between crustal solidus and ambient crustal temperatures decreases with depth, which promotes increasing melting rates in the lower crust (e.g., Dufek and Bergantz, 2005).

It is thermally viable to create extensive partially molten regions in the crust (>10 km thick in two million years, Fig. 7) with geologically reasonable flux and extension rates ($>0.004 \text{ m}^3/\text{m}^2/\text{yr}$ and $>0.005 \text{ m/yr}$ for most cases). In the lower crust, these melts reside at low melt fraction (<0.2 in most of the cases) and persist in part because they are buffered by latent heat at high crystallinity (e.g., Dufek and Bergantz, 2005; Huber et al., 2009; Marsh, 1981). The assumptions of localized tectonic extension and repeated diking events likely contribute to heat preservation and construction of mush regions. There are relatively limited geophysical constraints on the expanse and residence time for lower crustal melt that focus on: 1) petrologic observations on exhumed lower crustal terrains that include for example Ivrea, North Italy (e.g., Sinigoi et al., 1994; Voshage et al., 1990; Wedepohl, 1995); Kohistan, Pakistan (e.g., Burg, 2011; Dhuime et al., 2009; Jagoutz et al., 2007; Jagoutz and Schmidt, 2012); Fraser and Musgrave Ranges, Australia (e.g., Fletcher et al., 1991; White et al., 1999); Sierra Nevada and Peninsular Ranges, California (e.g., DePaolo, 1981a; Pickett and Saleeby, 1993); Fiordland, New Zealand (e.g., Klepeis et al., 2003) and 2) seismic studies exemplified by Main Ethiopian Rift (e.g., Keranen et al., 2004) and Western United States (e.g., McCarthy and Thompson,

1988). However, while useful, these observations provide a limited picture of the temporal history of lower crustal magmas.

While relatively little is known about lower crustal mush regions, observations of eruptive products (e.g., Bachmann and Bergantz, 2008; Hildreth, 1979; Hildreth, 2004) provide valuable information about processes in the mid-upper crustal mush reservoirs. Although our study does not model magma segregation and extraction from crystalline residue, our results suggest thermal conditions that can allow for development of upper crustal mush systems within reasonable magma flux ($>0.006 \text{ m}^3/\text{m}^2/\text{yr}$), extension rate ($>0.007 \text{ m/yr}$), and timescales (10^5 - 10^6 years). High flux ($0.01 \text{ m}^3/\text{m}^2/\text{yr}$) and intermediate extension rate (0.007 m/yr) can create large upper crustal mush regions, therefore, increased extension rates will allow for more extensive upper crustal bodies. Additionally, dry injections create an upper crustal mush with intermediate flux due to their high liquidus temperatures. From the initial conditions specified it takes greater than one million years for mid-upper crust to construct mush zones, although certainly crust starting with a higher initial geotherm would necessarily develop mush regions more quickly. The long residence time of many of these systems at low melt fraction is consistent with preferential melt extraction and segregation at these low melt fractions. In the eruptive record, if melt is preferentially extracted at low fraction and some primary melt also erupts it should produce the compositional gaps that are seen in many volcanic series (Dufek and Bachmann, 2010). In this particular case, the compositional gap would be between the intruding basalt and melt in equilibrium with 60-90 % crystals by volume. Based on the experimental studies of similar compositions and conditions, this roughly corresponds to andesitic compositions (e.g., Müntener et al., 2001; Sisson et al., 2005).

The thermal residence time of mid-upper crustal magma bodies has been the topic of much recent work (e.g., Annen, 2009; Gelman et al., 2013) as these systems are important for the assembly of large, caldera forming eruptions. Although some geochemical studies suggest $>10^5$ - 10^6 years for longevity of upper crustal melt reservoirs (e.g., Christensen and DePaolo, 1993; Costa, 2008) with potentially short timescales of rejuvenation prior to eruption (e.g., Barboni and Schoene, 2014; Burgisser and Bergantz, 2011; Cooper and Kent, 2014), previous conductive thermal models indicate that systems of the size sufficient to produce caldera forming eruptions persist at those timescales only in high magma flux conditions (10^{-2} - 10^{-1} km³/yr, Annen, 2009). In a recent work, Gelman et al. (2013) suggested that these magma bodies could form and reside in the upper crust with lower fluxes (<5 - 8×10^{-3} km³/yr) for longer timescales (1.2 million years with 10^{-2} km³/yr). These magma emplacement rates are consistent or somewhat higher than estimates from individual large eruptions (10^{-3} to 10^{-2} km³/yr in Altiplano, de Silva and Gosnold, 2007; 10^{-2} km³/yr in Taupo Volcanic Zone, Wilson et al., 2006).

Yet, calculations of average magma emplacement rates inferred from long-term volcanic input rates are on the order of 10^{-3} - 10^{-4} km³/yr (Crisp, 1984) and previous studies suggested short residence times of magmas in these regimes (e.g., Tappa et al., 2011). In our model, assuming an injection area of 10 km x 10 km as representative of the areal footprint of large caldera forming eruptions, we estimate 10^{-3} km³/yr and 4 - 6×10^{-4} km³/yr for the highest and intermediate flux values, respectively. For these conditions, we do observe long-lived upper crustal magma reservoirs ($>10^5$ years), however, for the majority of their lifetimes they are at low melt fraction and might not be eruptible (Marsh, 1981).

Observations in various intraplate and arc settings show that crustal magmatic systems evolve through the contribution of two end-member processes: 1) residual mantle-derived melt from crystallization of intruded magma and 2) crustal melt generated by partial melting of the crust (e.g., Deering et al., 2008; DePaolo, 1981b; DePaolo et al., 1992; Hildreth et al., 1991; Lipman et al., 1978). Our results suggest a combination of these two processes, although hydrous intrusions interacting throughout the crustal column tend to heavily favor residence of mantle-derived melts over crustal melting. This is consistent with recent petrologic studies, particularly those focused on the intrusion of hydrous magmas into the crust (e.g., Bachmann and Bergantz, 2008; Deering et al., 2008; Jagoutz and Schmidt, 2013; Müntener et al., 2001). Trace element and isotopic evidence, where there is sufficient leverage to discern mantle and crustal melts, also indicate a dominant role for fractionation in many regions of relatively thin crust, low basalt flux, and hydrous magmas (e.g., Searchlight Pluton, Gelman et al., 2014; Kohistan, Jagoutz, 2010; Marianas, Peninsular Ranges, Andes, and Aegean Arc, Lee and Bachmann, 2014). However, assimilation of crustal melts occurs in almost all our simulations. While fractionated melts dominate the melt inventory, crustal melt assimilation plays a role in producing compositional diversity (e.g., DePaolo, 1981b), especially in cases where the intruding magma has a high liquidus relative to the solidus of the intruding lithologies (such as our dry intrusion simulations). We note that we only examine the interaction of melt with igneous crust and not with sediments filling the basin developed by extension, which can provide an additional source of material for melting (e.g., McKenzie, 1978).

The maximum efficiency of crustal melting in our simulations is lower than, but comparable to those calculated via convection and over-accretion models (e.g., Huppert

and Sparks, 1988; Younker and Vogel, 1976). Crustal melting is more favored in regions where dry melts ascend from the mantle particularly when intruding hydrous lithologies (e.g., Bindeman and Simakin, 2014). It is also significantly controlled by tectonic extension (Fig. 9).

4.2. Caveats and model limitations

Geophysical observations of extension rates are often limited to surface estimates and it is not trivial to calculate the extension at depth because tectonic extension may vary nonlinearly in the crustal column (e.g., Buck, 2004; Villamor and Berryman, 2001). Nevertheless, our model gives estimates of the melt generated in the crust using the surface measurements as a constraint. Similarly, estimating magma flux (and total mass budgets) for a particular region remains difficult and may vary in time. In our calculations, we focused on general proxy compositions for crustal layering to examine the overall sensitivity of the system to intrusions and extension. Due to these simplifications, caution should be used in making detailed comparisons with specific settings where the lithology may vary.

Additionally, our model does not include rheological feedback in determining dike propagation and emplacement in the crust. Structural discontinuities, heterogeneities, brittle-ductile transition, and magmatic stoping likely influence the melt emplacement depth (e.g., Clemens and Mawer, 1992; Gudmundsson, 2006; John, 1988; Marsh, 1982; Parsons et al., 1992; Petford et al., 1994).

4.3. Implications to dry and wet environments

Dry and wet injection scenarios in our model represent two different extensional settings. While highly variable, melt inclusion studies show that back-arc and rifted arcs typically have wet basalts (e.g., Hochstaedter et al., 1990). On the other hand, continental rift magmas, which often have MORB-like melting trends, are typically known as having drier magmas (e.g., Bailey, 1978; Michael, 1995).

We find that wet environments have less crustal contribution compared to dry environments because hydrous dike intrusions are less capable of melting the crust. On the other hand, wet environments have more residual mantle-derived melt compared to dry environments (Fig. 4) because the melt fraction of the dry melt decreases very quickly and stays at lower crystallinity within the same range of temperatures (Fig. 2). This suggests higher crustal melting efficiency in continental rifts compared to rifted arcs and back-arcs. However, we suggest that crustal melting of igneous crust in both wet (arcs) and dry (continental rift) conditions is very low. This is in agreement with petrologic studies in different rift settings, exemplified by the Taupo Volcanic Zone (e.g., Deering et al., 2008), Kos Plateau Tuff (e.g., Bachmann et al., 2007), and the Main Ethiopian Rift (e.g., Boccaletti et al., 1999).

Our model indicates two different compositional evolution scenarios in dry and wet magma injection conditions. While in wet environments only one large melt accumulation region evolves through time; a number of melt bodies formed isolated lenses that stagnate at multiple levels in the crust and evolve through time in dry environments. We suggest that these melt bodies will likely evolve compositionally differently because they stall and differentiate at multiple levels in the crust. Therefore, these bodies that come from the same source at approximately the same time are expected

to be compositionally distinct, suggesting greater compositional diversity in magmas of environments with dry injections. A similar mechanism was suggested by Bloomer et al. (1989), where they observed variable geochemistry of Kenya rift volcanic rocks that are spatially distributed over a small area. Bloomer et al. (1989) argued that this variety might show the presence of individual melt pockets at depth or variations in the degrees of partial melting or fractionation.

5. Conclusions

We developed a thermal model that estimates the time and lengthscales of the melt evolution in the crust in different rift settings given their basalt flux, extension rates, and magmatic water contents. Our simulations indicate that geologically reasonable fluxes ($>0.004 \text{ m}^3/\text{m}^2/\text{yr}$) of hydrous magmas (5.6 wt.% H_2O) with intermediate-high extension conditions ($>0.005 \text{ m/yr}$) promote extensive mush regions with long residence times ($>10^5$ years). The evolution of melt in the crust is primarily controlled by the magma flux. The control of tectonic extension is prominent in high flux conditions and increases the amount of melt significantly. The total melt in the crust is a mixture of crustal and mantle-derived material with mantle-derived melt dominating the overall volume. Dry injections result in higher crustal melting efficiency compared to wet basaltic injections, however, even with the optimal conditions, the maximum crustal melting efficiency is very low ($\sim 6\%$).

In most of the conditions tested, the melt stays at low melt fraction at both lower and upper crustal depths for the majority of its history. We find that under optimal conditions magma can stay molten in the mid to lower crust on million-year timescales.

Under the conditions tested in the model, magma flux exerts a dominant control on the volume and longevity of melt, with extensional tectonics and magmatic water content also imposing strong controls on melt generation and longevity in extending crust.

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Table 1. Model parameters.

Fig. 1. Conceptual model for thermo-mechanical simulations. The computational domain is an idealized proxy for the crust and uppermost mantle and has three layers: 0-10 km is tonalite, 10-30 km is amphibolite and 30-60 km is peridotite. Diking is localized over the 10 kilometers on the crust-mantle boundary (injection width). Normalizing the two-dimensional area of melt in the domain by the total intrusion width is defined as the normalized melt thickness (NMT). The ratio 'R' represents the ratio of NMT to the initial crustal thickness (ICT). Sills are formed at the tip of the dikes. Emplacement of dikes and sills are accommodated by displacement of the crustal material to conserve mass. Crustal extension is given by a uniform displacement through the crustal column by a defined extension rate. Extension is accommodated by thinning of the crust near-axis (constrained to 10 km half-width) by conservation of mass and induces mantle upwelling. Extension of the crust and mantle upwelling promotes advection of heat in the crust. An example deformation of a regular grid due to extension is shown in the inset. Drawing is not to scale.

Fig. 2. (a) Equilibrium crystallization of basaltic andesite based on Mt. Shasta primitive lava composition (sample #85-44, Muntener et al., 2001) at 9 kbar. Melt fraction as a function of temperature and water content (5.6 wt.% and 0.5 wt.%) is calculated with rhyolite-MELTS thermodynamic software (Gualda et al., 2012) and (b) Melt fraction as a function of temperature for the crustal lithologies. Parameterized melt fractions of tonalite is based on melting experiments of Piwiński and Wyllie (1968). Amphibolite parameterization is based on the experiments of Wolf and Wyllie (1994), Rushmer (1991), and Sisson et al. (2005).

Fig. 3. An example simulation that represents diking, tectonic extension and resulting crustal thermal profile at one point in time. (a) Stochastic dike intrusions from mantle into the crust and (b) temperature perturbation due to intrusion events. The simulation represents an environment with $0.006 \text{ m}^3/\text{m}^2/\text{yr}$ basalt flux and $0.002 \text{ m}/\text{yr}$ tectonic extension. Water content of the intruded magma is 5.6 wt.%.

Fig. 4. Evolution of crustal and residual mantle-derived melts as a function of time when the extension is fixed (intermediate, $0.007 \text{ m}/\text{yr}$). R is the ratio of normalized melt thickness (NMT, Fig. 1) to the initial crustal thickness (ICT, Fig. 1). A compilation of 16 sets of simulations with basalt fluxes ranging from 0 to $0.010 \text{ m}^3/\text{m}^2/\text{yr}$ are shown. In (a-b) wet basaltic injections and in (c-d) dry basaltic injections are simulated. Note the order of magnitude difference between crustal (a, c) and residual mantle-derived melts (b, d) in the color bars.

Fig. 5. Evolution of crustal and residual mantle-derived melts as a function of time when the basalt flux is fixed (intermediate, $0.006 \text{ m}^3/\text{m}^2/\text{yr}$). R is the ratio of normalized melt thickness (NMT, Fig. 1) to the initial crustal thickness (ICT, Fig. 1). A compilation of 5 sets of simulations with tectonic extension ranging from 0.002 to $0.015 \text{ m}/\text{yr}$ is shown. In (a-b) wet basaltic injections and in (c-d) dry basaltic injections are simulated. Note the

order of magnitude difference between the crustal (a, c) and residual mantle-derived melts (b, d) in the color bars.

Fig. 6. Normalized thickness of total melt (total of crustal and residual mantle-derived melts) in the crust as a function of basalt flux and tectonic rate. All simulated environments are shown with varying basalt flux (16 flux values, circles) and tectonic extension (5 extension rates, different colored lines). Time is 1 million years after the initiation of tectonic extension and diiking events. (a) wet (5.6 wt.% H₂O) and (b) dry (0.5 wt.% H₂O) basaltic injections are simulated.

Fig. 7. Distribution of melt fraction (f) in the two-dimensional crustal domain after 1 and 2 million years of tectonic extension and diiking events. Both wet (5.6 wt.% H₂O) and dry (0.5 wt.% H₂O) basaltic injection scenarios are shown. In (a, c) basalt flux is 0.010 m³/m²/yr; in (b, d) basalt flux is 0.006 m³/m²/yr. Tectonic extension is intermediate (0.007 m/yr) in all simulations.

Fig. 8. Longevity of interconnected melt bodies (f>0.25) and their area in the two-dimensional crustal domain. Both wet (5.6 wt.% H₂O) and dry (0.5 wt.% H₂O) basaltic injection scenarios are shown. In (a, c) basalt flux is 0.010 m³/m²/yr; in (b, d) basalt flux is 0.006 m³/m²/yr. Tectonic extension is intermediate (0.007 m/yr) in all simulations. Different colors represent individual melt bodies. Note the difference in y-scale in (a-d).

Fig. 9. Crustal melting efficiency as a function of time and tectonic extension. Both wet (5.6 wt.% H₂O) and dry (0.5 wt.% H₂O) basaltic injection scenarios are shown. In (a, c) basalt flux is 0.010 m³/m²/yr; in (b, d) basalt flux is 0.006 m³/m²/yr. Note the order of magnitude difference between wet (a, b) and dry injection scenarios (c, d) in the color bars.

Figure1

FIGURES

Figure2

- Piwiinski and Wyllie (1968)
- △ Wolf and Wyllie (1994)
- △ Rushmer (1991)
- Sisson et al (2005)

Figure3

Figure4

a

b

c

d

Figure5

a

b

c

d

Figure6

a

Wet basaltic injections (5.6 wt.% H₂O)

b

Dry basaltic injections (0.5 wt.% H₂O)

Figure7

Figure8

a

High Flux
Wet basaltic injections

b

Intermediate Flux
Wet basaltic injections

c

High Flux
Dry basaltic injections

d

Intermediate Flux
Dry basaltic injections

Figure9

a

b

c

d

Table 1
Model parameters.

Symbol	Parameter		Value	Unit
k	thermal conductivity	basalt	2.2	W m ⁻¹ K ⁻¹
		crust	3.6	
T	temperature			K
ρ	density	basaltic melt	2750	kg m ⁻³
		solidified basalt	3065	
		crustal melt	2700	
		solid crust	3050	
H ₀	radiogenic heat		8.0 x 10 ⁻¹⁰	W kg ⁻¹
l	radiogenic lengthscale		16000	m
c _p	specific heat			J kg ⁻¹ K ⁻¹
t	time			s
v _i	velocity			m s ⁻¹
L	latent heat		1.0 x 10 ⁵	J kg ⁻¹
f	melt fraction			
E	efficiency			%
V	volume			m ³
E _T	energy budget			J