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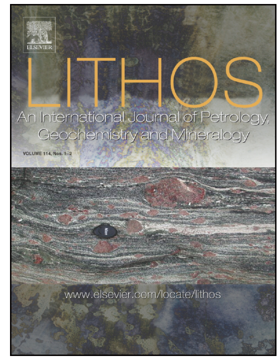

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Thermal Model of Successive Dike Injections and Implications for the Development of Intraplate Volcanoes

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ABSTRACT

Temperatures in the root zones of volcanoes play a critical role in the development and persistence of shallow-level magmatic reservoirs in the crust. Here, we present a 1D thermal model allowing evaluation of the thermal impact of magma travelling in conduits to the surface on the root zone of a volcano. This thermal model has been developed to better understand the formation of a vertical intrusion located in the root zone of a dismembered Miocene volcano on Fuerteventura, Canary Archipelago. This intrusion, named PX1, constitutes an almost pure amalgamation of dikes of either clinopyroxenitic or gabbroic composition. Both types of dikes display cumulate textures and are interpreted as resulting from the protracted crystallization of a mafic magma. The formation of clinopyroxenitic, in contrast to gabbroic dikes, requires that the residual melt was extracted at high temperature (>1050°) to avoid plagioclase crystallization.

Simulations of multiple dike injections show that the temperature in the root zone increases significantly with the addition of dikes, but the maximum temperature reached in the system depends on the duration of magma flow in the conduits and the time interval between dike injections (i.e., repose period). Active flow is the critical parameter that distinguishes instantaneous dike injection from a magmatic conduit. Without significant magma flow (>1 month), high-temperature conditions (>1000°C) cannot be maintained in the pluton unless dikes are very thick and the repose period is extremely small. On the other hand, magma flow times of one to several months, combined with short time intervals between dike injections (<25 years), which are conditions comparable to those recorded for historical eruptions of oceanic island volcanoes, allow the production and preservation of temperatures above the plagioclase liquidus for significant durations, as required to generate clinopyroxenitic dikes such as those observed in the PX1 pluton. Persistent high temperature in the vicinity of magma conduits limits the differentiation of melts in transit to the surface, providing a potential explanation for why lavas of mafic to intermediate composition predominate in intraplate volcanoes such as Fuerteventura or Fogo Island (Cape Verde

Archipelago). In extreme cases, when temperatures over 1000-1050°C in the central part of the feeding zone are maintained for years, the remaining magma in the conduit does not solidify but is preserved in a mushy state. New pulses of magma would not be able to cross this zone but would rather amalgamate in the incipient magma reservoir. The present model differs from previous models of sill intrusion in that magmas do not need to pond at depth to create a reservoir but merely supply heat while travelling to the surface. Depending on the time interval between dike injections and the duration of magma flow through the crust, magma rising in vertical conduits could directly feed the volcanic edifice or could lead to the formation of magma reservoirs. This process may explain why some volcanoes erupt mafic or differentiated magmas during distinct periods of activity.

1. Introduction

Magmatic differentiation is the major process to account for the generation of various rock types at mid-ocean ridge, arc, or intraplate settings. However, the depth at which fractional crystallization takes place is still a matter of discussion. Batches of magma are extracted from the source and rise. These magmas are then expected to be stored in magmatic reservoirs, which could be localized at different depths from the upper mantle to the upper crust. Important questions are how a magma chamber develops and what the thermal structure of the edifice is to maintain a magma chamber active for sufficient time to allow the magma to differentiate (e.g., Annen, 2009; Karakas et al., 2017; White et al., 2006). Different models for the locations and shapes of magma chambers have been proposed in the context of intraplate oceanic islands. Studies on Hawaiian volcanoes have suggested the presence of large magma chambers, but the depth at which these magma chambers developed varies depending on the studies. Clague and Dixon (2000) and Gudmundsson (2012) argued for the development of a deep magma chamber at the crust-mantle boundary, while a shallow magma reservoir ca. 3 km below the summit of the volcano (Decker, 1987) is expected to form at the late preshield stage. This shallow magma chamber is assumed to cool and solidify progressively, while the magma supply decreases during the postshield growth stage (Clague and Dixon, 2000). In contrast, studies on other basaltic intraplate volcanoes have shown that this representation of the plumbing system as large magmatic chambers may not be extrapolated to all ocean islands, in particular for volcanoes such as those in the Canary, Cape Verde or Madeira Islands characterized by low magma fluxes. There is no evidence for long-lived shallow magma reservoirs for most of these islands (Galipp et al., 2006; Hildner et al., 2012; Klügel et al., 2000). One hypothesis to explain the lack of a persistent shallow magma reservoir is the development of giant landslides that would have affected the tectonic stress fields of these volcanoes and thus prevented the preservation of shallow magma reservoirs (Longpre et al., 2008). However, most authors instead argue that this lack is the consequence of a low magma supply rate, which would be insufficient to prevent the cooling of these reservoirs (Clague and Dixon; 2000, Gudmundsson, 2012; Klügel et al., 2015). Thermobarometry studies on lavas from the western Canary Islands (Galipp et al. 2006; Hansteen et al., 1998; Klügel et al., 2005; Marti et al.,2013) and Fogo Island (Cape Verde Archipelago; Hildner et al.,2012) indicate that magmatic differentiation starts within the upper mantle at 15-30 km depth. These deep levels are interpreted as places of prolonged magma storage where fractional crystallization could take place. Multistage magma ascent models have thus been developed for Canary or Cape Verde Archipelago volcanoes (Galipp et al., 2006; Hansteen et al., 1998; Hildner et al., 2012; Hildner et al., 2011; Klügel, 1998; Klügel et al., 2005; Klügel et al., 1997; Klügel et al., 2000; Longpre et al., 2014; Stroncik et al., 2009). These models do not involve long-term magma stagnation at shallow depths but rather a complex system of magma reservoirs or pockets at upper mantle and crustal levels, where batches of melt can evolve separately (Gudmundsson, 2012). The magma is collected in these reservoirs and stagnates before rapid vertical ascent toward another reservoir or directly to the surface through hydraulic fractures. In these models, no fractional crystallization is expected during vertical transport of magma; magma is rather considered to ascend quickly to preserve high temperatures, thus preventing dike solidification (Bruce and Huppert, 1990; Menand et al., 2015). In contrast, the study of plutonic bodies in Fuerteventura (Canary Islands) volcanoes supports mineral fractionation during magma ascent in vertical conduits (Tornare et al., 2016), which is an important process to account for the variability in lava composition and for the texture observed at the surface.

Fuerteventura Island provides a unique opportunity to study magmatic differentiation and the connection between plutonic and volcanic activities, as a giant landslide has cut the volcano to a depth of ~3 km (Stillman, 1999). Tornare and coworkers (2016) have established a genetic relationship between a shallow-level, vertically layered clinopyroxenite-gabbro pluton, named PX1, and its contemporaneous basanite to basaltic trachyandesite lavas. PX1 is located in the root zone of a volcano and consists of alternating vertical lithological sequences varying from highly cumulate clinopyroxenitic dikes to less cumulate compositions dominated by gabbroic mineral assemblages. The similar mineral composition and zoning observed in clinopyroxene (cpx) from dikes and concomitant emitted lavas supports the hypothesis that dikes composing the PX1 pluton represent distinct magmatic conduits packed together to produce a ~3.5 x 5.5 km pluton. The formation of variable lithologies as observed in PX1 implies that crystal segregation is efficient during ascent of the magma to the surface. The different plutonic lithologies varying from olivine (ol)-rich wehrlite to gabbro may suggest that they formed during a progressive alkaline differentiation trend, e.g., ol-rich wehrlites could represent high-temperature ol-cpx cumulates, while gabbros may form at lower temperature when plagioclase reaches its liquidus. However, the similar ranges of clinopyroxene and olivine compositions, including Mg#, observed in all plutonic rocks suggest that these rocks represent cumulates formed under similar conditions but with variable interstitial melt extraction. The range in compositions of cumulate rocks from ol-rich wehrlites to gabbros reflects the efficiency of interstitial melt extraction before plagioclase crystallization (Tornare et al., 2016). Tornare et al. (2016) hypothesized that mineral growth and segregation in these vertical conduits play important role s in explaining the evolution from mafic to intermediate lava compositions and suggested that the thermal structure of the volcanic edifice is crucial to allow for interstitial residual melt in dikes to be extracted to form cumulate rocks. In this paper, we investigate the evolution of the thermal structure of the shallow part of the volcanoes associated with variations in magma supply and transport to the surface. We show that fluctuations in the repose time between eruptions and the duration of these eruptions play an important role in the development of the internal thermal structure of the volcanoes, which, in turn, could impact magma differentiation and the development of magma reservoirs in intraplate volcanoes, as previously suggested by various authors for other geological settings (e.g., Annen, 2009; Karakas et al., 2017; White et al., 2006).

Modeling of the thermal evolution of incrementally grown plutons by multiple dike injection has not yet been investigated in uppermost crustal environments. Existing numerical models have focused on the postemplacement cooling of magma bodies, sills or dikes and their effects on the country rock. Many models consider that magma is emplaced instantaneously in one batch, although some studies have focused on the thermal effect of successive magma injections, especially sills (Annen, 2011; Annen et al., 2006; Annen and Sparks, 2002; Michaut and Jaupart, 2006; Michaut and Jaupart, 2011; Philpotts and Asher, 1993; Caricchi et al. 2014). Karakas et al. (2017) investigated the thermal effect of multiple dike and sill injections at 5-15 km depth to generate large magma chambers. Their paper demonstrates the importance of the initial upper crustal thermal structure to produce such magma reservoirs. Magma transport in vertical conduits has been modeled in a single dike by investigating the thermal impact of continuous magma flow and magma advection along the dike over a given period of time (Bruce and Huppert, 1990; Delaney and Pollard, 1982; Petcovic and Dufek, 2005). Floess and Baumgartner (2015) investigated the thermal effect of successive dike emplacement on the contact metamorphic aureole of a pluton, considering several variables such as magma flow rates, periods of magma inactivity, dike thickness and magma temperature. However, these authors have not considered the impact of these parameters on plutonic rock compositions and the long-term temperature evolution within the pluton.

Here, we explore a thermal model of pluton growth by successive dike injections within the upper crust. We use variables such as magma flow duration, time between magma injection and dike thickness. Numerical predictions are compared with field observations to test whether our petrological model can account for the development of a migmatitic contact aureole and a clinopyroxenite-gabbro alternation as documented around and within the PX1 pluton, respectively. We evaluate the importance of multiple dike injection on the thermal boundary conditions associated with magma differentiation processes.

2. Geological context

This numerical study is motivated by geological observations made on Fuerteventura Island. In this section, we summarize information that we use to test our thermal model and its implication for magma differentiation in intraplate volcanoes. Fuerteventura is a volcanic ocean island in the Canary Archipelago and consists of three eroded Miocene volcanoes (Fig. 1). The Central volcano reveals a complex plutonic root zone made of several imbricated alkaline plutons. One of these plutons, named PX1, is a complex, vertically layered association of cumulate rocks linked with remnants of contemporaneous lava flows. PX1 rocks are interpreted as resulting from the accumulation and coarsening of phenocrysts in juxtaposed volcanic conduits during the movement of crystal-bearing magmas to the surface (Fig. 1d) (Tornare et al., 2016).

The PX1 pluton is interpreted as a dynamic volcanic feeder zone located in a tectonically active transtensional shear zone. Ar/Ar and U/Pb isotope dating on kaersutite, zircon and baddeleyite yields ages of 22.10 ± 0.07 and 21.8 ± 0.5 Ma, respectively (Allibon et al., 2011a). PX1 is approximately 3.5 by 5.5 km in size (~>10 km^2) and the emplacement pressure was empirically estimated ca. 0.1 – 0.2 GPa (Allibon et al., 2011b; Holloway and Bussy, 2008; Holloway et al., 2008), in line with the expected lithostatic pressure at the base of the ca. 3000 m-high (Stillman 1999) overlying volcano. Subvertical layering is observed at all scales and throughout the whole pluton. At the pluton scale, the layering is expressed by main lithological sequence alternation (Fig. 1c). In the field, layering is defined by subvertical dikes or bodies, which have been emplaced successively against each other and/or intruded among each other (Tornare et al., 2016). The thickness of these dikes or bodies varies from several centimeters to a few meters. The layering is oriented NNE-SSW, except in the northwestern part of the pluton, where it changes to WNW-ESE (Allibon et al., 2011b). These orientations reflect the extensional rift-like regime that affected the island during the Lower Miocene (Fernández et al., 1997).

The PX1 intrusion consists of lithologies ranging from olivine-rich wehrlites to clinopyroxenites to gabbros. Figure 1c reports predominant lithologies subdivided into sequences according to plagioclase modal proportions. Some areas are dominated by plagioclase-rich clinopyroxenites and gabbros (referred to as gabbroic sequences in blue in Fig. 1c), while other areas consist mainly of clinopyroxenite (± olivine and minor plagioclase) to wehrlite lithologies (in orange in Fig. 1c). The study of mineral composition, mostly cpx, suggests that the sequence from ol-rich wehrlite to gabbro does not represent progressive cumulates linked to fractional crystallization but similar lithologies with variable residual melt extraction (Tornare et al., 2016). The ol-wehrlites represent cumulates from which most of the residual liquid has been extracted, while gabbros seem to have retained their residual liquid until complete crystallization. Tornare et al. (2016) hypothesized that this variable melt extraction efficiency could be linked to a difference in thermal conditions: wehrlite formed at temperatures sufficiently hot to prevent plagioclase crystallization (>1050°C) and to promote residual melt extraction. The physical processes associated with residual melt extraction from the dikes are mostly unconstrained. As the regional tectonic setting during PX1 growth was transtensive to extensive (Fernández et al., 1997), it is unlikely that pressure gradients were induced in this way. Allibon et al. (2011b) proposed that forceful injections in previously emplaced crystal-rich channels could generate the required horizontal pressure gradient through mush compaction, producing residual melt extraction. Extraction could also be simply due to the effect of lithostatic pressure.

Various observations suggest that PX1 was formed at high temperatures. Layering is never sharp, and dike edges are not always clearly distinguishable. Dikes are quite variable in length and thickness. They are tortuous, sometimes merging and sometimes subdividing into several dikelets. Other observations indicate that rising melt interacted with surrounding dikes at the time of new dike emplacement. Enclave swarms of cumulate rocks are found locally (Fig. 2a), probably originating from dismembering nearby cumulate dikes. In clinopyroxenite zones, these enclaves consist of rounded wehrlites and clinopyroxenites embedded in a heterogeneous olivine-clinopyroxene matrix. Enclave rims sometimes show crystals tearing off, attesting to supersolidus conditions during remobilization (Fig. 2b). Enclaves hosted in some gabbroic rocks show brittle textures tending toward magmatic breccia (Fig. 2c). This texture suggests that some gabbros were emplaced while the host rock was cooler than when clinopyroxenites formed, i.e., at temperatures closer to the brittle/ductile transition.

Pluton emplacement had a strong thermal impact on the country rock, which developed a contact metamorphic aureole affecting preexisting clinopyroxenitic, gabbroic and carbonatitic lithologies (Migmatite zone in Fig. 1c, Hobson et al., 1998). This metamorphism resulted in deformation and recrystallization of the host rock at temperature estimated as ca. 700-890°C (Muñoz and Sagredo, 1989) to 900-1000°C (Koepke et al., 2003). These high temperatures induced partial melting and melt segregation in hydrated lithologies surrounding the western and northern borders of the pluton over a width of 200 m (Holloway and Bussy, 2008; Holloway et al., 2008). Contact metamorphism extends over a width of more than 1 km with evidence of a greenschist-facies overprint (Stillman, 1987).

3. Model setup

We performed 1D-thermal simulations of incremental pluton growth by random magma channel emplacement. Figure 3 shows a schematic view of the modeling technique we used. This model simulates pluton growth based on the random location of channel emplacement in a defined area (Fig. 3a and b). The pluton growth is simulated to a size of 600 m corresponding to the minimum width observed in the PX1 pluton (Fig. 1c). To model the thermal effect of magma flowing in vertical channels, we maintain the temperature of the dike at 1150°C for a given amount of time (Fig. 3b). Then, the magma supply ceases, and the channel starts to cool by thermal diffusion during an period of inactivity (Fig. 3c). After this time lapse, a new channel forms adjacent to the first channel (Fig. 3d). We ran several simulations with this model, testing different parameters such as the dike thickness, the magma flow duration (flow time) and the time interval between dike emplacement (repose period; t_RP), which is inversely proportional to the dike injection frequency, to evaluate and compare the thermal effect of each parameter separately.

Thermal calculations were performed with a one-dimensional finite difference scheme in a MATLAB code. A regular grid with a numerical resolution of 0.5 m was employed for an initial domain size of 10000 m. Time steps were 1 day.

The heat equation used in the MATLAB code is a simple adaptation of the diffusion equation: $[\frac{\partial T}{\partial t}] = k\,[\frac{\partial^2 T}{\partial x^2}]$ with a constant thermal diffusivity of 10^{-6} m^2s^{-1}.

We assume an initial temperature of 200°C for the system. This initial temperature is clearly higher than expected from a standard geothermal gradient for a depth of 3000 to 4000 meters (100-120°C). However, it is reasonable to consider such a high initial temperature in a volcanic root zone, where earlier magmatic activity is documented. On the other hand, initial temperatures were likely below greenschist-facies conditions (~300 to 400°C), since such conditions were reached 1 km away from the pluton in the contact metamorphism aureole induced by pluton emplacement. Consequently, an initial temperature of 200°C was considered reasonable in the model.

We applied Dirichlet boundary conditions fixed at 200°C, corresponding to the initial temperature conditions. We established a 5000 m-thick host rock zone on both sides of the intrusion. The left contact between host-rock and intrusion was fixed, while the second contact was mobile and moved progressively toward the right along with new dike injection. The fixed contact corresponded to the location of the first channel (Fig. 3a and b). The size of the host rock zone was defined after a few tests because a fixed Dirichlet boundary condition of 200°C could potentially affect the thermal profiles if the limit is too close to the intrusion. Tests performed using a limit fixed at 10000 m on both sides of the intrusion show no difference from tests with a limit fixed at 5000 m, even in the longest duration diffusion experiments (see Fig. S1). The simulation of channel volume emplacement is due to the shifting of the preexisting thermal profile to the right over a distance equal to the new dike thickness (Fig. 3d). The potential position of the new channel is defined randomly in the area, which corresponds to the already emplaced dikes extended by the width of one dike.

Plagioclase phenocrysts are mostly absent in Fuerteventura lavas contemporaneous with PX1 formation, while cpx phenocrysts are always present (Tornare et al., 2016). A constant magma temperature of 1150°C was selected in order to allow clinopyroxene crystallization while keeping the temperature above the plagioclase liquidus temperature during the phase of magma transit in the channel (Fig. 3b). To predict liquidus temperatures for cpx and plagioclase, we used the rhyolite-MELTS software (Gualda et al., 2012) assuming an H_2O-undersaturated alkaline basanite as the starting material, differentiating at pressures between 0.1 and 0.2 GPa. Rhyolite-MELTS calculations predict an increase in the cpx liquidus from 1140 to 1185°C with increasing differentiation pressure from 0.1 to 0.2 GPa. Complex zoning in cpx from PX1 plutonic lithologies suggests, however, that cpx starts to crystallize during magma ascent (Tornare et al., 2016); we thus used a temperature of 1150°C for the rising magma, which is slightly higher than the minimum temperature predicted by rhyolite-MELTS at 0.1 GPa. In Figs. 4, 5 and 6, we report the 1050°C isotherm corresponding to our estimate for the plagioclase liquidus temperature. This temperature corresponds to the MELTS-predicted plagioclase liquidus temperature for a magma differentiating at 0.1 GPa with 1 wt% H_2O. A lower solidus plagioclase temperature is predicted with increasing water content, but as it is difficult to predict the water content in the magma, we selected the highest predicted value for comparison with the thermal prediction of pluton evolution associated with multiple dike injection.

This model does not take into account the latent heat of crystallization. Unlike models assuming "instantaneous" emplacement of dikes or sills within the crust (e.g., Annen and Sparks, 2002), our model considers that magma continuously rises during the eruption time, therefore making it difficult to estimate the amount of heat released by mineral crystallization during the process. Nevertheless, we assume that the latent heat of crystallization is minor relative to the magma heat input and would not noticeably influence the final results, as this potential heat addition would only increase the effect of magma flux to the surrounding edifice.

The channel width was set to 1, 2 or 10 meters. Dikes thinner than 1 m would require higher numerical resolution and calculation time, which were unreasonable for the needs of the model. At the high end, dike widths greater than 10 m are scarcely documented in the field.

Repose periods between dike injections (t_RP) of 5, 15, and 25 years were tested for all dike widths. Two additional t_RP values (50 and 100 years) were tested for the 1 m-thick dike emplacement simulations. The flow time alternatives were 1, 3 and 6 months. These durations are plausible and in agreement with natural observations on fissure eruptions as recorded in Iceland (Thordarson and Larsen, 2007), Cape Verde (Amelung and Day, 2002), and the Canary Islands during historical eruptions (Carracedo et al., 2015; Gonzalez et al., 2013). We also performed simulations without flow time (flow time: 0 month) to compare the effect of continuous magma supply versus a one-time magma injection. This situation was tested only for 1 m-thick and 10 m-thick dike emplacement models.

4. Results

The complete data set for all simulations is provided in supplementary Fig. S2, and Figs. 5 to 9 report only the most interesting simulations for the different parameters. All temperature data and profiles or diagrams are related to the thermal situation at the end of a magmatic cycle. These data display the results of heat diffusion after a flow period plus a repose period for each cycle (active flow and inactive flow periods for each dike emplacement). The results are presented in three different ways and synthesized in Fig. 4.

The simulation illustrated in Fig. 4 shows that a hot zone is progressively developed where the pluton is growing. In this area, the temperature profile tends to follow a Gaussian distribution with a maximum located in the pluton (Fig. 4a). Temperatures at the center of the pluton largely exceed the solidus (900°C) and decrease toward the contacts. The peak of temperature is not located exactly in the center of the pluton but where the last injection or several recent injections occurred and would rather correspond to their remnant heat (Fig. 4a). This effect is also shown in Fig. 4b, where high-temperature spots appear through time and are not located exactly at the center of the pluton. The temperature decreases strongly outside the pluton as a function of thermal diffusion. Thermal profiles on both sides of the pluton are not exactly similar, as the volume increase associated with dike injection is compensated only on the right side of the pluton. Contact metamorphic conditions exceed the solidus temperature (900°C) over ca. 50 m and then decrease progressively over ca. 1500 m. Figure 4b shows that a central hot area develops through time where the pluton grows. After 1100 years and emplacement of 75 dikes, some areas within the pluton exceed the solidus temperature (900°C). This area corresponds to a pluton width of 150 m (Fig. 4c). The temperature increase is strong and fast in the early stages of pluton growth and then slows down at higher temperatures. Figure 4c also shows that a maximum temperature over 1050°C corresponding to the plagioclase liquidus temperature is exceeded at the end of pluton growth.

4.1. Effect of the duration of magma flow (flow time)

Figure 5 reports the time evolution of maximum temperatures (T max) recorded in the pluton as a function of dike thickness, repose period and flow times (see Fig. S2 for the complete data set). The occurrence or lack of flow time, i.e., the duration of an eruption, significantly influences the thermal evolution of the system. For instantaneous dike injection, the system is primarily sensitive to the dike size. In the case of 1 m-thick instantaneous dike emplacement, temperatures are systematically well below all other simulations that integrate flow times (Fig. 5a). In the case of 10 m-thick dike emplacement, temperatures remain close to those calculated in simulations that integrate flow times. For instantaneous magma injection, only the 10 m-thick dike simulations with short t_RP (5 years) are able to heat the system over 1050°C, but this effect implies a short and probably unrealistic time scale for pluton formation (300 years to produce a 600 m-thick pluton, Fig. 5d). The flow time has a major influence on the evolution of T max recorded by the pluton: the longer the flow time is, the shorter the time required to sustain high temperatures in the pluton.

Variations in flow duration do not significantly impact the width of the metamorphic contact aureole but do affect the intensity of metamorphism (Fig. 6). Profiles in Fig. 6a and b show that for different flow times but similar repose periods (same color), the outer boundaries of the zone influenced by contact metamorphism are very close to each other. The effect of the flow time is more significant in the pluton itself than in the contact aureole. For a dike injection every 25 years, the increase in flow time from 0 to 6 months increases the temperature within the pluton from ca. 650°C to more than 1000°C before the arrival of the new dike. This difference is significant and could modify the potential interaction with the magma previously injected into the system. The effect is less important away from the intrusion aureole, where thermal profiles are only slightly lower in simulations using short flow times than in those using longer flow times. Temperatures also decrease dramatically for both simulation types when using long t_RPs (50 and 100 years Fig. 6c) and for 1 m-thick dike simulations without flow times (Fig. 6a).

4.2. Effect of dike thickness

Figure 5 shows that when simulations integrate magma flow (flow time >0 month), the dike width does not significantly influence the maximum temperature reached in the pluton (see also Fig. S2). The maximum temperatures obtained as the pluton progressively grows are very similar, whether dikes are 1 m thick or 10 m thick (Fig. 5). This result is especially true for short flow times (1 month) and short t_RPs (5 years). Dissimilarities between temperature profiles appear during the first stages of pluton growth for longer flow times and t_RP simulations, but this effect is minor. In contrast, dike thickness has a great impact on the thermal evolution of the system in simulations without flow time; 10 m-thick-dike plutons reach systematically higher temperatures than 1 m-thick-dike plutons (Fig. 5) according to the amount of hot material added to the system. This effect would even be reinforced if the latent heat of crystallization were taken into account. Dike thickness has also a clear effect on the timing of pluton growth (Fig. 7). The thicker the dikes, the faster a pluton reaches a given size at equal t_RP and the faster high temperatures are reached within the pluton.

Considering the overall thermal evolution of the pluton through time (Fig. 8), we observe that the growing pluton shows a more homogeneous and gradual thermal evolution in systems formed by 1 m-thick dikes than in those formed by 10 m-thick dikes, and the central hot area is generally wider. The thermal evolution of a pluton growing by emplacing 10 m-thick dikes shows a patchy structure in the central hot area. These hot spots reflect the locations where large volumes of magma are emplaced and progressively release their heat through thermal relaxation.

4.3. Effect of repose period (t_RP)

Figure 5 shows that the time interval between dike injections, which corresponds to the repose period (t_RP), clearly influences the maximum temperature reached in the pluton (see also Fig. S1). Short t_RPs systematically induce hotter evolution of growing plutons. Maximum temperatures exceed 1050°C in all simulations using short t_RPs of 5 years, independent of the other parameters. Simulations using longer t_RPs (15 and 25 years) are also able to reach this temperature but only when they are combined with long flow times. Figure 5 also indicates that simulations using short t_RPs produce a sharper and quicker temperature increase during the early stages of pluton growth, and these systems reach hotter thermal maturation. For instance, if a pluton is built with 1 m-thick dikes active for 6 months each (flow time), a temperature of 1050°C is reached after 400 years and 80 dikes, assuming a t_RP of 5 years. This temperature is reached only after 3450 years and emplacement of 230 dikes if the t_RP is 15 years and after 8900 years and 356 dikes if the RP is 25 years.

The contact aureole width also varies as a function of t_RP in association with the growth rate and heating rate of the pluton (Fig. 6). Contact aureoles are significantly wider in simulations using longer t_RPs, which are related to longer pluton growth and heat diffusion times (Figs. 6 and 8).

5. Discussion

5.1 Thermal simulation applied to the emplacement of the PX1 pluton

The thermal model presented here has been developed to account for geological characteristics observed in plutonic rocks in Fuerteventura. Gabbroic dikes are interpreted as crystallized melts, while clinopyroxenitic dikes attest to mineral (cpx, ol, ± Fe-Ti oxides) segregation and accumulation in vertical conduits. The absence of plagioclase in these cumulate rocks implies that residual liquid was extracted before the temperature reached the plagioclase liquidus. This result suggests that the system stayed above this critical temperature long enough to allow such melt extraction. Additionally, the contact aureole surrounding PX1 requires high heat transfer to the enclosing rocks, as mafic rocks are partially remelted (Muñoz and Sagredo, 1989; Hobson et al., 1998; Koepke et al., 2003). Finally, the construction of a large volcanic edifice (3000 m high; Stillman, 1999) simultaneous with the emplacement of PX1 suggests that large volumes of basaltic magma flowed through the pluton to the surface during pluton formation.

The simulation of the thermal evolution of a 600 m-wide zone built by multiple dike injections (Figs. 5-8) demonstrates that such a process could produce a significant increase in temperature within the pluton and the surrounding rocks. Magma flow in the system is a critical parameter because simulations performed without flow time were not able to heat the pluton to more than 900°C, except for the case of large dike injection (10 m) and repose time lower than 15 years (Figs. 5 and S1). Adding the latent heat of crystallization would increase the thermal effect of instantaneous dike injection, particularly for large dikes (Jaeger, 1964; Furlong et al., 1991; Annen and Sparks, 2002). However, as mentioned in the previous paragraph, the formation of PX1 by multiple instantaneous thick dike injections seems unlikely. In contrast, all simulations including magma flow predict an internal temperature of the pluton greater than 900°C after the incremental zone has reached 600 m in width.

The dike injection rate has been identified as an important factor controlling the thermal evolution of a magmatic system (Annen, 2011; Annen and Sparks, 2002; Michaut and Jaupart, 2006; Michaut and Jaupart, 2011), especially when magma bodies are emplaced closely or against each other. Continuous magma supply (flow time ≥1 month) in channels also has a strong influence on the thermal state of the system (Figs. 5 to 8). A protracted heat input in a channel allows the injection of more heat into the overall system regardless of the channel volume (dike width). Indeed, while the dike remains at the maximum magma temperature during flow time, heat is continuously transferred to the surrounding host rocks by conduction through dike walls, heating the host rocks progressively. Therefore, the heat supply during flow times is added to the primary amount of heat released by the dike volume.

The results of our thermal simulations support the model proposed by Tornare et al. (2016) for the formation of the PX1 pluton by multiple injections through magmatic conduits; the results also allow making some predictions about the magma supply associated with the development of the pluton. Simulations show that temperatures greater 1050°C can indeed be reached and preserved in the pluton when dike injection is frequent (every 5 or 15 years, Figs. 5 and S1) and is combined with long flow times (i.e., eruption times). If the flow time is limited to 1 month, only very frequent eruptions can preserve high-temperature conditions in the subsurface pluton. In contrast, only a long flow time (≥6 months) can produce these conditions if eruptions occur only every 25 years (Fig. 5). Simulations performed with longer repose periods of 50 or 100 years (Fig. S1) suggest that the pluton cools below 1000°C and 900°C, respectively, before the subsequent dike is injected, even if the flow time is as long as 6 months.

The time interval between eruptions and their duration involved in the simulations seem to agree with time constraints for historical and prehistorical volcanic eruptions observed in mafic oceanic islands (White et al., 2006). As an example, historical eruptions in the Canary Islands have lasted between 1 and 3 months on average, some much longer such as in Tenerife (>13 months in 1704-1705) or in Lanzarote (>68 months in 1730-1736) (Carracedo et al., 2015; Sobradelo et al., 2011). However, intervals between eruptions on these specific islands could be much longer, ranging from 20-60 years to ca. 100 years to 200 years (Carracedo et al., 2015; Sobradelo et al., 2011). In contrast, historical fissure eruptions recorded at Fogo, Cape Verde, show a much shorter repose period (average of eruption intervals: 14 years, with one longer interval between the 1857 and 1951 eruptions; Amelung and Day, 2002). Etna shows an even shorter interval with one or more eruptions every year since 2000. Therefore, we can assume that the durations of 3 to 6 months and repose periods of 5 to 25 years used in our model are realistic.

The presence of ol-wehrlite and clinopyroxenite in the PX1 pluton suggests that residual liquid could have been extracted before plagioclase started to crystallize. Whatever the cause of the pressure gradient responsible for squeezing the residual melt out of the dike, it is necessary that the dike remains partially molten for some time, i.e., at temperatures higher than 1050°C to avoid plagioclase crystallization. The maximum temperatures calculated during simulations of pluton growth (Figs. 5 to 8) show that it is reasonable to assume temperatures in the PX1 pluton exceeded 1050°C long enough to allow extraction of residual melt, as required to generate the cumulate rocks. Moreover, the spatial coexistence of gabbroic and pyroxenitic lithologies could be explained by small variations in the time interval between dike injections and in the duration of eruptions.

5.2. Contact metamorphism and generation of the migmatite aureole

Periods of protracted heat supply affect not only the thermal evolution of a growing pluton but also its country rock. According to our calculations (Fig. 6), the emplacement of a 600 m-wide pluton develops a contact aureole whose characteristics strongly depend on the time interval between magma injections, the magma flow times and the total number of dikes. Flow times of 3 to 6 months and 600 1 m-thick dikes (Fig. 6a, c) would induce a temperature above 800°C in the country rock over a distance of more than 200 m from the contact. In contrast, 60 10 m-thick dikes would not produce such a wide and intense metamorphic aureole (Fig. 6d). Indeed, the intensity of the contact metamorphism, especially in the immediate vicinity of the pluton, is largely controlled by the total amount of heat injected into the system and by the capacity to accumulate this heat through high injection rates, as shown by Floess and Baumgartner (2015). The thickness of the contact aureole and the intensity of the metamorphism far from the contact are also related to this amount of heat injected into the system (Fig. 6). By contributing significantly to the heat input into the system, protracted magma flow thus has a very important impact on contact metamorphism development. Field observations demonstrate that the emplacement of PX1 induced strong contact metamorphism with partial melting of gabbroic lithologies over a distance of ca. 200 m around the contact (migmatite zone in Fig. 1; Hobson et al., 1998; Holloway and Bussy, 2008; Holloway et al., 2008). Estimates of recrystallization temperatures are approximately 700-1000°C (Holloway and Bussy, 2008; Holloway et al., 2008; Koepke et al., 2003; Muñoz and Sagredo, 1989). Figure 6 shows that this temperature range is almost never reached in simulations without flow times, except in the case where the 600 m-wide pluton is formed by the emplacement of a 1 m-thick dike every 5 years (Fig. 6a). By adding flow times, it is possible to develop a contact aureole in the 700 to 1000°C temperature range even using a relatively long time interval between dike injections (Fig. 6c). The emplacement of thick dikes seems unable to allow the development of migmatites over the expected distance from the contact (200 m, Fig. 6d), but we do not exclude that higher temperature could be reached if the latent heat of crystallization is taken into account (Annen and Sparks, 2002) and thermal diffusion after the end of pluton emplacement is considerate. In contrast, 1 m-thick dike simulations tend to produce contact aureoles twice as large (Fig. 6a). In addition, with thin dike injections, temperatures exceed greenschist-facies conditions 1 km away from the contact. Our calculations suggest that the right thermal conditions across the entire contact aureole are produced by the injection of 2 m-thick dikes. However, the model is representative of the emplacement of a 600 m-thick pluton, while as illustrated in Fig. 1c, some parts of PX1 consist of a succession of main lithological sequences totaling 3000-3500 m in size, which might significantly modify the thermal effect on the country rock. We suggest that the pluton potentially did not form in a single pulse of successive dike emplacement but was more likely formed incrementally by successive main sequences. It is possible that these sequences also reflect periods of eruptive activity that varied over time. The intensity of contact metamorphism is mainly controlled by the growth of the sequence close to the host rock. Sequences emplaced in the central part of the pluton have only a moderate effect on the aureole. On the other hand, if successive sequences are emplaced incrementally against the host rock, the thermal effect on the latter would be amplified (Floess and Baumgartner, 2015), unless the intervals between emplacements are long enough to significantly cool the host rock. Since no chronology of sequence emplacement could be established, it is not possible to determine which portion of the intrusion is responsible for the development of the contact metamorphic aureole. Nevertheless, the conditions estimated for the formation of cumulate clinopyroxenitic rocks in the pluton, i.e., the injection of 1-2 m-thick dikes at frequencies of 5 to 25 years and durations between 3 and 6 months seems to be in agreement with the development of a 200 m-thick migmatitic zone directly in contact with the pluton and of greenschist-facies conditions 1 km away from the contact. This result represents strong support for our model of pluton formation.

5.3. Juxtaposition of clinopyroxenitic and gabbroic dikes

Field observations suggest that the PX1 pluton was formed by amalgamation of a very large number of dikes. The map presented in Fig. 1b indicates that some areas are dominated by gabbroic lithologies, while other areas are more clinopyroxenitic. However, clinopyroxenitic and gabbroic dikes frequently coexist at the local scale. This section focuses on the feasibility of achieving the specific thermal conditions required for the production of large volumes of cumulate rocks of clinopyroxenitic composition.

As discussed above, rocks of clinopyroxenitic composition lacking plagioclase require interstitial melt extraction while the temperature is still above the plagioclase liquidus, i.e., >1050°C. These thermal conditions are met in simulations with frequent injections and flow times >0 month (Figs. 5, 7 and 8) but only after the emplacement of a series of early dikes, which warm up the system (Figs. 5, 7 and 8). These early dikes are expected to quickly cool below the plagioclase liquidus temperature and to crystallize as gabbros, resulting in noncumulate compositions. Thus, the proportion of gabbroic dikes observed in PX1 should at least be as high as the proportion of early dikes required by the thermal model. Once the growing pluton has reached a thermal threshold above the plagioclase liquidus temperature, the system can potentially produce clinopyroxenitic dikes assuming that the residual liquid could be extracted. Calculations show that temperatures in the pluton exceed the solidus temperature of plagioclase after the emplacement of a series of early dikes corresponding to a pluton width of >50 m with injection intervals of 5 years and flow times of 6 months (Fig. 5a). This width increases to 100-250 m if a 25 year time interval is used (Fig. 5c). Considering the final size of the simulated pluton (600 m), injection intervals longer than 5 years would imply that a large proportion of noncumulate gabbroic dikes are emplaced before the first cumulate rock might form, which is not what is observed in PX1, where some zones are dominated by pyroxenitic compositions (Fig. 1c). In other words, pyroxenitic lithologies could potentially be generated by an additional process, i.e., the remelting of earlier, plagioclase-bearing gabbroic dikes, if we consider the very high ambient temperatures generated by large magma fluxes. This process might be considered an extension of dike-wall rock partial melting processes (Petcovic and Dufek, 2005) or progressive channel widening in individual dikes (Bruce and Huppert, 1990) due to repeated magma injection through the same conduit. Field observations do indeed attest to ductile deformation during dike emplacement in some gabbroic areas (Fig. 2e). This observation illustrates that rising magmas did actually interact with the neighboring rocks and suggests that gabbros were emplaced under cooler conditions than clinopyroxenites.

Figure 9 presents simulations using random flow times (1 to 6 months) with variable repose periods (1 to 25 years) to mimic natural fluctuations recorded in intraplate volcanoes (e.g., White et al., 2006). The initial step of pluton growth is associated with a rapid increase in temperature to 900-1000°C in the central zone of the pluton (Fig. 9, time from 0 to ~3000 years). This zone grows progressively by new dike injections and maintains its high temperature until the end of pluton growth (Fig. 9). After ca. 3500 years, the randomly calculated repose period decreases slightly over a few hundred years (dotted line in Fig. 9). This time lapse coincides with the development of a hot zone where the temperature exceeds 1050° during ca. 500 years. The temperature in this zone then decreases to ca. 1000°C before the zone is reheated by new dike injections. As we use a random location for dike injection in simulations, the hot zones most likely include previously solidified dikes. Such a process could partially remelt dikes of gabbroic composition, turning them into plagioclase-free residual cumulate rocks with resorption mineral textures, provided the resulting interstitial melts are extracted. These processes of remelting or extraction of residual liquid from early dikes might well contribute to the development of large volumes of residual rocks of clinopyroxenitic composition. During the last period of pluton growth, the simulation suggests that a large, more than 300 to 400 m-thick, hot zone develops with temperature that could even exceed 1080-1100°C in some place (Fig. 9). Under such conditions, it is not clear whether the central part keeps its internal structure, as the proportion of liquid is expected to be quite high. As we hypothesize below, this zone could correspond to a magmatic reservoir. Simulations imply a high magma supply rate to account for eruptions every 50 years or less. We might expect that, in natural systems, the magma supply rate could fluctuate over the long run (500-2000 years), in which case, hot zones in the pluton would accordingly be ephemeral. In summary, the large variety of lithologies in and the high level of structural complexity of PX1 can be accounted for by various combinations of time intervals between magma injection, magma flow duration, and efficiency of residual melt extraction. The complex zoning observed in clinopyroxene of both gabbroic and pyroxenitic dikes and associated lavas (normal and reverse zoning, resorption textures; Tornare et al., 2016) supports the interpretation that rising magma interacts with previously crystallized dikes.

5.4. Magma differentiation and potential development of shallow-level magma reservoirs

Alkaline oceanic volcanoes differ from place to place, with some islands such as Fuerteventura (Carracedo et al. 2001, Ancochea et al. 1996) or Fogo in the Cape Verde Archipelago (Eisele et al. 2015, Hildner 2011, Hildner et al. 2012) dominated by mafic to intermediate compositions, while other islands such as Tenerife (Araña and Brändle, 1969, Sliwinski et al., 2015, Wolff et al., 2000) have more differentiated compositions (Fig. 10). Figure 10 also shows that most differentiation trends are characterized by a compositional scarcity, or "gap", between moderately differentiated compositions and highly differentiated compositions, and the proportion of lava with compositions ranging from basaltic to basanitic to trachy-andesitic and phonotephritic compositions is significantly higher than that of highly differentiated lavas. This difference would even be larger if the respective volumes of magmas were considered instead of the number of analyses, as minor lithologies are expected to be oversampled with respect to dominant lithologies. Discontinuity in alkaline lava compositions is well known from the pioneering work of Daly (1925). As summarized by Dufek and Bachmann (2010), different hypotheses have been proposed to explain the gap between mafic and differentiated rocks observed in various settings. A first explanation is to consider that mafic and differentiated rocks have different origins, with mafic melts being extracted from the mantle and differentiated melts from the partial melting of the crust (e.g., Chayes, 1963); a second hypothesis supports the primary role of fractional crystallization (e.g., Marsh, 1981), while a third hypothesis considers the dominant role of melt-crystal segregation mechanisms (Dufek and Bachmann, 2010). The present model of multiple dike injection and the resulting thermal evolution provides a new perspective to account for the discontinuity in lava compositions for the specific case of alkaline oceanic islands.

The range of lava compositions observed in oceanic islands is usually interpreted as resulting from differentiation in deep-seated magma chambers. The evolved melts would subsequently quickly rise through vertical conduits and erupt at the surface, without significant differentiation and cooling in the conduits and/or in shallow-level reservoirs (Klügel et al., 2000; Klügel et al., 2015; Stroncik et al., 2009). However, Tornare et al. (2016) have shown that in Fuerteventura, mineral crystallization and segregation processes could indeed occur during magma transport in vertical conduits and account for the unusual proportion of mafic minerals in rocks from the root zone of the volcano. The thermal model we present here provides an additional potential mechanism to account for the large proportion of mafic to intermediate lavas in Fuerteventura. Figures 8 and 9 illustrate that the root zone of the volcano could reach temperatures over 1000°C through multiple dike injections, assuming a repose period similar to the ones observed in active volcanoes. The generation and preservation of such high temperatures throughout the conduits would limit magma crystallization and thus impact the composition of erupted lavas. MELTS calculations indicate that differentiation from a liquid similar to the Fuerteventura parental melt to 1050-1000°C produces melts with Mg# approximately 37 to 35 with SiO_2 content of 54 to 58 wt.%. This predicted compositional range is in agreement with the compositions of the most differentiated lavas observed in Fuerteventura, except for a few differentiated lavas with Mg# <10, which represent a negligible volume (<<1%). In other words, after an initial time required to heat up the plumbing system of the volcano, temperatures would be so high that they would not allow any more magmas to crystallize and fractionate the mineral phases necessary to generate highly differentiated lavas (low Mg# cpx, amphibole, Ti-oxides and plagioclase). Thus, the thermal structure of the root zone of oceanic island volcanoes could be the controlling factor for the predominance of mafic magmas over differentiated ones.

The next question is how to explain why some islands are dominated by mafic to intermediate composition while others, such Tenerife island (Araña and Brändle, 1969; Sliwinski et al., 2015; Wolff et al., 2000), erupt differentiated lavas. White et al. (2006) have shown, based on a compilation of repose periods and duration of eruptions from various volcanoes, that a positive correlation exists between the repose time and the SiO_2 content of the erupted magmas. If the repose time is shorter than ca. 100 years for basaltic magmas, it increases to more than a thousand years for more silica-rich magmas. White and coworkers postulated that this difference is related to the time required by fractional crystallization and assimilation processes to produce differentiated liquids. These researchers argue that volcanic systems must be open to magma recharge, except for the most frequently erupting basaltic volcanoes. This difference in the repose period between eruptions of mafic and differentiated lava suggests a change in the magma storage and eruption dynamics. Figure 11 illustrates how multiple dike injections could initiate the formation of a magmatic reservoir within the oceanic crust. This schematic view is presented in the context of Fuerteventura Island (Fig. 11a), as the PX1 pluton provides petrological observations to test predictions based on thermal modeling.

5.5. A model for the development of shallow level magma reservoirs

Stage 1. Initial stage of pluton formation

Stage 1 in Fig. 11b corresponds to the initial stage of pluton formation when the thermal conditions are still relatively cold. According to calculations reported in Fig. 9, these conditions are expected between 500 and 3000 years of pluton growth when new dikes are injected within crystallized dikes. Interaction between the solidified dikes and the newly injected magma could happen but would be relatively minor. This stage is illustrated in Fuerteventura by the juxtaposition of dikes of distinct compositions with relatively straight contacts (Fig. 2d). These thermal conditions are also expected if the repose period increases and allows previous dikes to cool and crystallize.

What happens at the very beginning (0 to 500 years in Fig. 9) is not quite clear, as we held the magma temperature constant in our 1D simulations and did not take into account the possible cooling and crystallization of the dike during magma flow. If heat advection by magma is slower than conduction, this situation would lead to magma solidification and the cessation of magma flow and dike propagation (Menand et al., 2015). When magma starts flowing in a new channel, solidification might occur at the dike walls, where the thermal contrast between the cool host rock and the magma is maximum (Bruce and Huppert, 1990; Delaney and Pollard, 1982). This effect might reduce or even clog magma transport in the channel. This phenomenon is expected to occur especially in thin channels (<2 m) if the magma flux is low (Menand et al., 2015) and/or if the thermal contrast with the wall rock is high, as solidification would proceed quickly inwards (Bruce and Huppert, 1990). In a volcanic system, channel obstruction is expected to occur at a rather shallow level, where the thermal contrast with the host rock is highest. As soon as a channel becomes obstructed at some place, the magma stops moving upward. However, as subsequent dikes are emplaced, even without magma transit, the growing pluton is progressively heated and is expected to quickly become hot enough to prevent rapid solidification in any small dikes. Indeed, as soon as the host rock temperature reaches ~400°C, complete solidification is prevented in dikes as thin as 1 m, and the magma is expected to keep flowing until the magma supply ceases (Bruce and Huppert, 1990).

Simulations reported in Fig. 5 show that by incremental emplacement of 1 m-thick dikes, without magma transit, a global pluton temperature of ~400°C is reached after only 20 dikes (~100 years) when the time lapse between dike injections is 5 years and after 72 dikes when the time lapse is 25 years (~1800 years). This discussion illustrates the importance of thermal maturation of the system to develop volcanic activity.

Stage 2. Pluton growth

According to Bruce and Huppert (1990), with a dike thickness greater than ca. 2 m, as well as in the case of low thermal contrast between the host rock and the magma channel, the continuous heat supply of a magma flow is thought to exceed the conductive heat loss to the host rock. Solidification and constriction might still occur in the channel but only during the early stage of magma flow and not sufficiently to completely obstruct the channel. Then, if heat is continuously supplied to the system, remelting of adjacent solidified dikes can be expected, leading to a potential widening of the active channel until the magma flow ceases (Bruce and Huppert, 1990; Petcovic and Dufek, 2005). Stage 2 (Fig. 11c) represents such hot conditions when new magma passes between preexisting dikes, which are mostly consolidated but also still ductile. In this case, significant interactions between former dikes and flowing magma occur. This process seems to be the dominant situation in the PX1 pluton. Most contacts between dikes are tortuous with significant variations in dike thickness. Enclaves of various sizes are frequently seen included in new magmas (Fig. 2c). The contact between dikes is frequently transitional, and it is not rare to see remelting of previous gabbroic dikes in contact with new magmas (Fig. 2b). We hypothesize that such a process is also important to add crystals with complex chemical zoning to the erupted lavas (see Tornare et al., 2016). Such crystals would be very difficult to distinguish from “true” phenocrysts, as the chemistry of these crystals would be inherited from similar magmas but injected earlier into the system. The high temperature would prevent differentiation, but some crystal segregation in the vertical conduits is expected, particularly if the vertical velocity of the magma is low or decreases during the last stages of the eruption.

Stage 3. Development of magmatic reservoirs

Figure 9 illustrates that a hot zone could develop in the central part of the pluton if frequent magma injections are associated with long-lasting magma flow for thousands of years. Stage 3 (Fig. 11d) illustrates these conditions when the ambient temperature is sufficiently high to remelt the previously solidified gabbroic dikes and to prevent cooling of the melt below the plagioclase liquidus temperature. As long as these conditions are maintained, large volumes of the pluton are thus partially molten (several tens of meters in width in our simulations, Fig. 9). This effect could even be enhanced if the location of new dike injection is not random within the forming pluton, as presented in the model, but follows preferential paths of crustal weaknesses. As weak zones are likely associated with previous dike injection, we hypothesize that new magma pulses may follow the same path as previous dikes, producing even higher thermal effects than those observed in the simulations presented in Fig. 9. Although simulations allow only a sectional view through the stacks of channels, these partially molten mushy volumes would extend vertically along the active channels. We expect that these volumes may lose their internal cohesion and turn into large unconsolidated olivine-clinopyroxene crystal mush zones. This phenomenon would lead to the creation of ephemeral magma reservoirs (Fig. 11d). As a result, subsequent magma injections would not reach the surface but would be injected into the crystal mush, thus remobilizing crystal clots and mushy lumps.

PX1 reveals several large areas characterized by unclear or even absent vertical layering, which host numerous rounded enclave swarms of mostly olivine-rich clinopyroxenitic lithologies (Fig. 2a). The matrix surrounding these enclaves is heterogeneous, and the contacts between these enclaves and the matrix most often show progressive modal transition, suggesting progressive crystal separation from the enclaves (Fig. 2b). This observation suggests remobilization of olivine-rich unconsolidated material under supersolidus conditions. These areas revealed by the PX1 intrusion may represent the potential "ephemeral-reservoirs" as suggested by the model.

An interesting implication of the presence of a hot zone in the central part of the pluton is that new magma pulses are unlikely to cross this crystal mush zone. New magma pulses would instead recharge the reservoir and keep the internal temperature high (Fig. 11d). However, in the same way, the absence of magma flux in the upper part of the crust (see Fig. 11d) would allow the ambient temperature in the upper part of the edifice to decrease. This potential temperature contrast between the upper and lower parts of the crust might possibly generate differentiation processes. The eruption style is likely to change from a situation controlled by magma reservoirs fed by mantle melts at the base of the crust to crystal-melt extraction processes in the upper crust, inducing production and eruption of differentiated lavas (e.g., Bachmann and Huber, 2016). This change in the eruption style may explain the differences in the repose period between silica-rich and silica-poor lavas, as pointed out by White et al. (2006).

6. Conclusions

Thermal simulations of multiple dike injections into the shallow crust show that the time interval between dike injection and the duration of magma flow are two major factors controlling the thermal evolution of the root zones of intraplate volcanoes. Protracted magma flow in dikes increases the temperature within the feeding zone well beyond what would be achieved by instantaneous magma injections, especially when dikes are 1-2 m thick. This process can account for the development of the PX1 subvolcanic pluton in Fuerteventura, which requires a high and sustained thermal regime to account for a high proportion of cumulate clinopyroxenitic dikes, which formed above the plagioclase liquidus temperature of 1050°C, and for the development of a very high-grade migmatite contact aureole around the pluton.

Preservation of high-temperature regimes in volcanic conduits and, more generally, in the whole feeding zone of a volcano might have several consequences for lava compositions. As fractional crystallization is mostly driven by magma cooling, magmatic differentiation would be limited if temperatures around magma conduits are high. In other words, the temperature in the root zone of a volcano controls the degree of differentiation of the melts. Although this process cannot account for the "Daly" gap observed in alkaline lavas, the process does provide a potential explanation for the predominance of mafic to intermediate lava compositions over differentiated compositions in intraplate volcanoes such as Fuerteventura or Fogo Islands.

The thermal regime in the root zone of a volcano might also impact the potential development of shallow-level magmatic reservoirs. Our model shows that magmas do not necessarily need to stop and stagnate within the upper crust to initiate a magmatic reservoir. If dikes are injected at small time intervals, this process could heat the central part of the feeding zone of a volcano to temperatures greater than 1000-1050°C for years. These conditions would keep magmas in a mushy state in the vertical conduits. Only during a subsequent stage would new magma injections pond at the base of this mushy zone and feed the incipient magmatic reservoir. The latter could survive for thousands of years if the repose period between eruptions is maintained. Mineral fractionation could then occur within such a reservoir, leading to the potential eruption of differentiated lavas. In other words, the multiple dike injection process could generate ephemeral magma reservoirs whose lifetimes would be controlled by the duration of periods associated with frequent dike injection and could explain why volcanoes erupt mafic and differentiated lavas during distinct periods of activity.

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Fig. 1: (a) Geographic setting of the Canary Islands. (b) Geological map of Fuerteventura showing Miocene edifices and effusive centers (northern, central and southern volcanic complexes) with locations of Basal Complex outcrops, Miocene intrusions and Miocene Basaltic Series; this map was compiled from data in Casillas et al. (2008), Ancochea et al. (1996), Stillman (1999), Muñoz et al. (2003) and Le Bas et al. (1986). (c) Schematic map of the PX1 intrusion modified after Hobson et al. (1998) and Tornare et al. (2016). (d) Location of the PX1 intrusion in the plumbing system of the Miocene Central volcano of Fuerteventura. The volcano cross section was interpreted following Gutierrez et al. (2006), Dañobeitia and Canales (2000), Stillman (1999), and Steiner et al. (1998).

Fig. 2: (a) A PX1 pluton area dominated by clinopyroxenitic lithologies. This zone is characterized by a chaotic structure with numerous enclaves of wehrlitic lithologies enclosed in clinopyroxenite. The larger enclaves, outlined by white dashed lines, may reach several meters in size. (b) Closer look at the contact between the olivine-rich wehrlitic enclave shown in panel (a) and the enclosing clinopyroxenitic matrix. This contact is not sharp, and the olivine proportion decreases progressively toward the clinopyroxenitic matrix, suggesting possible crystal separation from the enclave contact. (c) Light-colored gabbro dikes carrying angular clinopyroxenite enclaves (outlined by black lines) intruding wehrlitic host rock. (d) Relatively straight contacts observed between banded gabbro and intruded ol-rich cumulate lithologies (wherlite). The wherlites are affected by subsolidus deformation, including olivine fracturing and clinopyroxene lineation.

Fig. 3: Model setup for the emplacement of a pluton by random location of channel-like intrusions. (a) Location of the first channel in all simulations (5000 m from the left limit of the box). (b) The simulated magma flow inside the channel is maintained for a certain amount of time (flow time) during which the channel remains at constant magma temperature (1150°C). Heat diffusion starts at this time. (c) The magma supply ceases, the channel cools, and heat diffuses during a repose period (t_RP). The thermal profile corresponds to the situation at the end of the repose period. A

subsequent channel is then emplaced randomly within the pluton area, leading to an additional dike width. (d) Thermal structure of the pluton during the active magma flow in the new dike. The change in pluton volume is solved by shifting the preexisting thermal profile to the right over a distance equal to the new dike width. (e) Thermal profile at the end of the repose period linked to the second dike emplacement. The arrow points to the location of the maximum temperature preserved in the system after two dike emplacements.

Fig. 4: Simulation of multiple 2 m-thick dike injections with 3 month flow times and 15 years between dike injections (repose period, t_RP). (a) Snapshot of final thermal profiles at the end of simulations. The curve shows the temperature as a function of distance, with the location of the pluton (gray field) and the emplacement of the first dike (red line). (b) Thermal evolution of the system throughout the entire duration of a simulation, from the first dike emplacement (time = 0 years) to the end of the 600 m-thick pluton formation (time = 4500 years). The temperature indicated by the color bar is given as a function of the distance relative to the model domain (x-axis) and the time (y-axis). (c) Evolution of the maximum temperature reached in the system as a function of time, pluton thickness or magmatic cycles. The red horizontal line indicates the choice of temperature for magma rising in the dikes (1150°C), and the black horizontal line indicates the estimated plagioclase liquidus temperature (1050°). The black arrows highlight the connection between graphs shown in panels (b) and (c). Note: all data correspond to the thermal situation at the end of the magmatic cycle, composed of an active magma flow period plus a repose period.

Fig. 5: Evolution of T max as a function of time, pluton thickness or magmatic cycles for various simulations of multiple dike injections. The different diagrams report T max evolution depending on dike thickness (1 – 10 m), repose period (5, 15, or 25 years), and flow time (0, 1, 3, or 6 months). Simulations using 2 m-thick dikes or longer repose periods are presented in Fig. S2.

Fig. 6: Thermal profiles showing the final situation of 600 m-thick pluton emplacement simulations. (a) Simulations of pluton construction using 1 m-thick dikes, flow times between 0 and 6 months, and 5, 15 or 25 years repose periods. (b) Simulations of pluton construction using 2 m-thick dikes and parameters for simulations similar to those reported in (a). (c) Simulations of pluton construction using 1 m-thick dikes but longer repose periods (50 and 100 years). (d) Simulations of pluton construction using 10 m-thick dikes. The shaded rectangles represent the natural contact metamorphic conditions applied to the simulations. The yellowish rectangle indicates the conditions of the 200 m-wide migmatite aureole formation described by Hobson et al. (1998); the orange rectangle corresponds to the greenschist facies recognized by Stillman (1987) 1 km away from the PX1 border. The upper red arrow indicates the magma temperature at 1150°C, and the upper black arrow indicates the plagioclase liquidus temperature (1050°C). The gray area indicates the final pluton thickness of 600 m. The vertical red line indicates the 1000 m distance, which corresponds in the simulation to the first dike emplacement area and the fixed contact between the host rock and the pluton.

Fig. 7: Effect of dike thickness on maximum temperature (T max) observed in the pluton as a function of time (or magmatic cycles). The different diagrams represent simulation with distinct repose periods and flow times.

Fig. 8: Thermal evolution of the pluton throughout the entire duration of a simulation, from the first dike emplacement (time = 0 years) to the end of the 600 m-thick pluton formation (time varies from 300 to 15000 years depending the dike thickness and repose period). Simulations reported in diagrams (a) to (f) refer to a dike injection rate of 5 years (i.e., repose period), while diagrams (g) to (l) refer to a repose period of 25 years. The different diagrams illustrate the effects of dike thickness and flow time on the thermal evolution of the pluton and surrounding rocks.

Fig. 9: (a) Thermal evolution of the pluton throughout the entire duration of a simulation, from the first dike emplacement to the end of the 600 m-thick pluton formation. The flow time and repose period vary randomly between 1 and 6 months and between 1 and 25 years, respectively. The black line marks the 1050°C isotherm. (b) Variation in the repose period as a function of time during the simulation. Blue and red lines indicate the effective repose period variation and the average of 10 time intervals between dike injections, respectively. (c) Color code used in panel (a).

Fig. 10: (a) Total alkali–silica (TAS) diagram and (b) alkali vs Mg# for lavas representative of oceanic island volcanism. All compositions have been selected from the Georoc database (http://georoc.mpch-mainz.gwdg.de/georoc/) and have totals between 98.5 and 101 wt.%. (c) Histogram showing the distribution of lava as a function of Mg# in the different oceanic islands. These diagrams illustrate the abundance of mafic to moderately differentiated lava with respect to that of differentiated lavas in oceanic islands.

Fig. 11: Schematic model for the evolution of a pluton by multiple dike injection. (a) Geological setting based on the case study of Fuerteventura Island (see Fig. 1). (b) Stage 1. Magma flow within solidified gabbroic pluton showing limited interaction between preexisting dikes and the rising magma. This situation corresponds to the initial steps of pluton formation. (c) Stage 2. The temperature within the pluton is high, and preexisting dikes are still close to their solidus temperature. Significant interactions are expected between preexisting dikes and the new rising magma. Enclaves and crystals from the preexisting dikes are entrained in the new magma. Partial melting of plagioclase from preexisting gabbroic dikes is possible, helping to produce large areas of clinopyroxenitic lithologies. (d) Stage 3. Conditions created after a significant period of frequent dike injection. In such a case, the temperature within the pluton is sufficiently high to prevent complete crystallization of the melt. A portion of the pluton stays in a mushy state where crystals do not create

a rigid framework. New magma pulses are unlikely to cross this zone to produce eruption at the surface. In contrast, such magmas intrude the mushy zone and mix with it. The absence of magma flux within the upper part of the crust produces a thermal gradient between the bottom of the crust heated by magma fluxes and the upper crust, which cools by conduction. We hypothesize that this situation initiates magmatic reservoirs and provides suitable conditions to produce differentiated magmas.

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HIGHLIGHTS

- We present a thermal model of successive dike injections in the root zone of a volcano.
- The temperature in the root zone depends on the duration of magma flow and the repose time between dike injections.
- High temperatures in the root zone, >1000°C, could be reached if repose and flow times recorded by historical eruptions are applied.
- Such high temperatures limit differentiation and may allow magma chambers to develop.

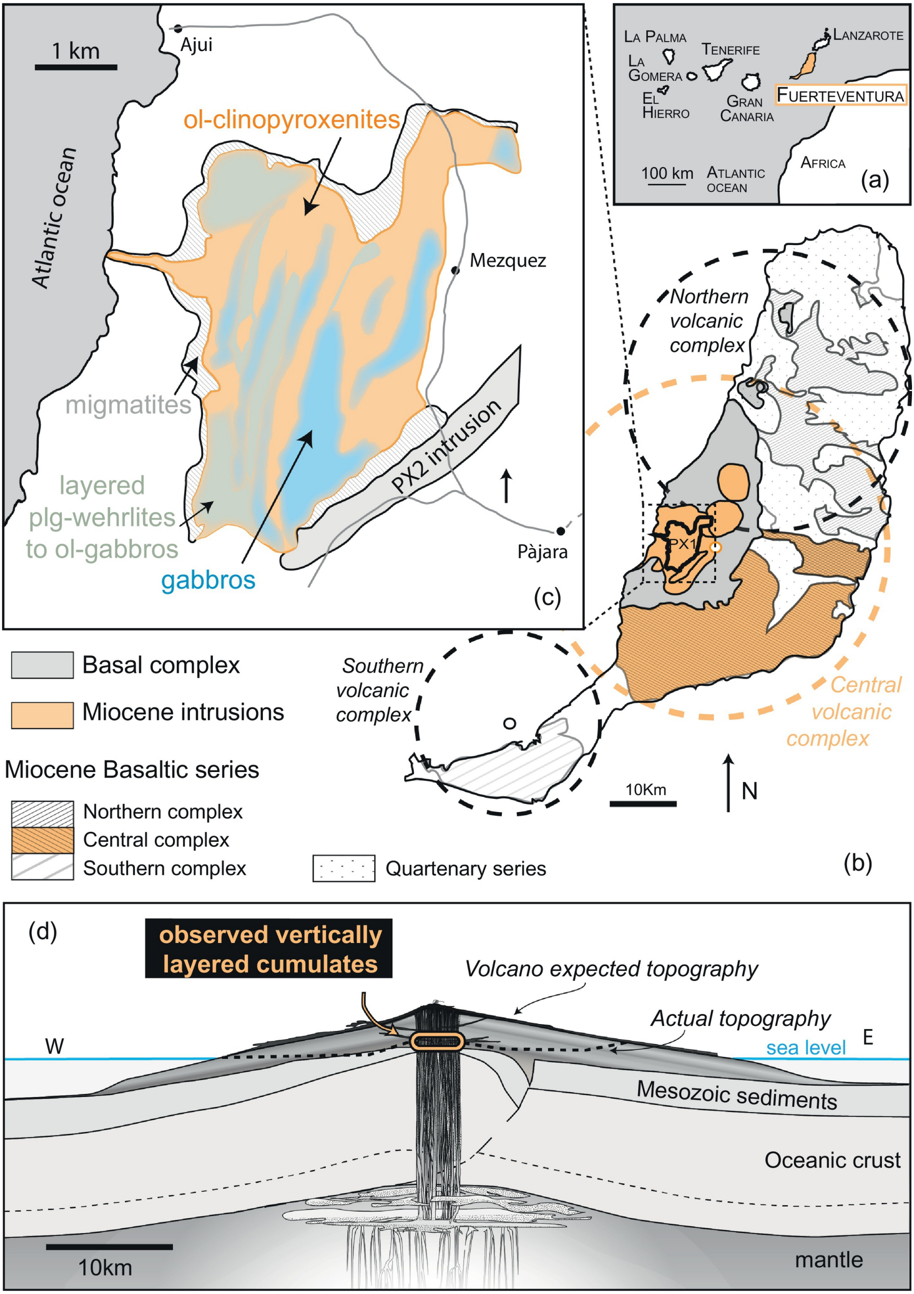

1 km
Ajui
Atlantic ocean
ol-clinopyroxenites
Mezquez
migmatites
layered
plg-wehrlites
to ol-gabbros
gabbros
PX2 intrusion
Pàjara
(c)
La Palma
La Gomera
El Hierro
Tenerife
Gran Canaria
Lanzarote
Fuerteventura
100 km
Atlantic ocean
Africa
(a)
Northern volcanic complex
PX1
Southern volcanic complex
Central volcanic complex
10Km
N
(b)
Basal complex
Miocene intrusions
Miocene Basaltic series
Northern complex
Central complex
Southern complex
Quartenary series
(d)
observed vertically layered cumulates
Volcano expected topography
Actual topography
W
E
sea level
Mesozoic sediments
Oceanic crust
mantle
10km

Figure 1

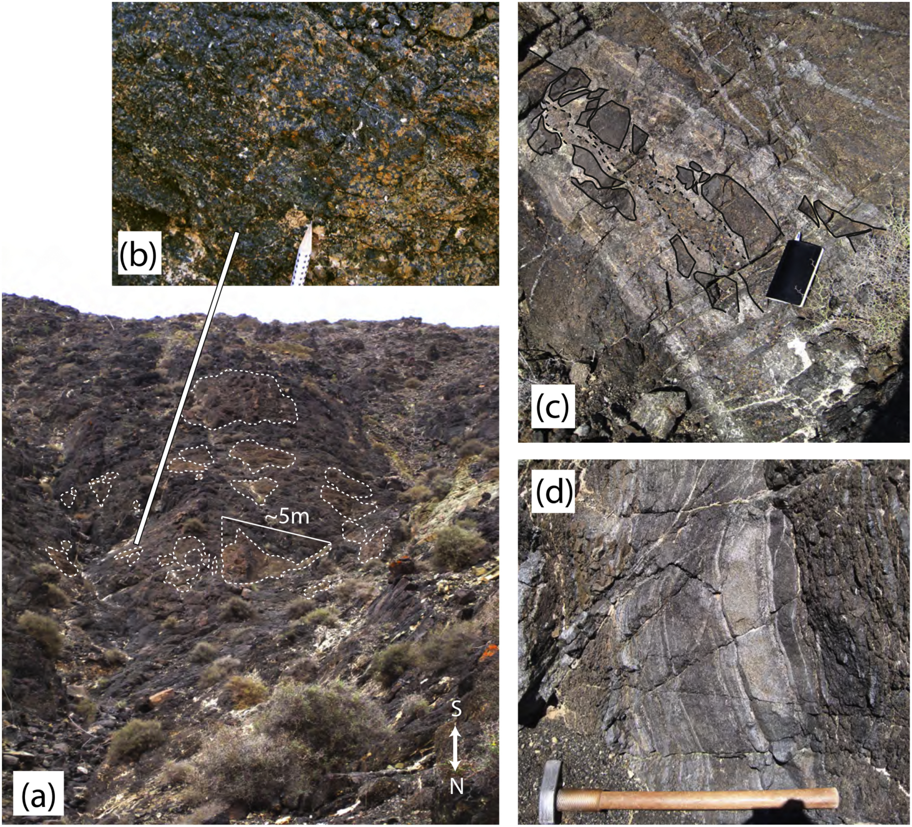


Figure 2

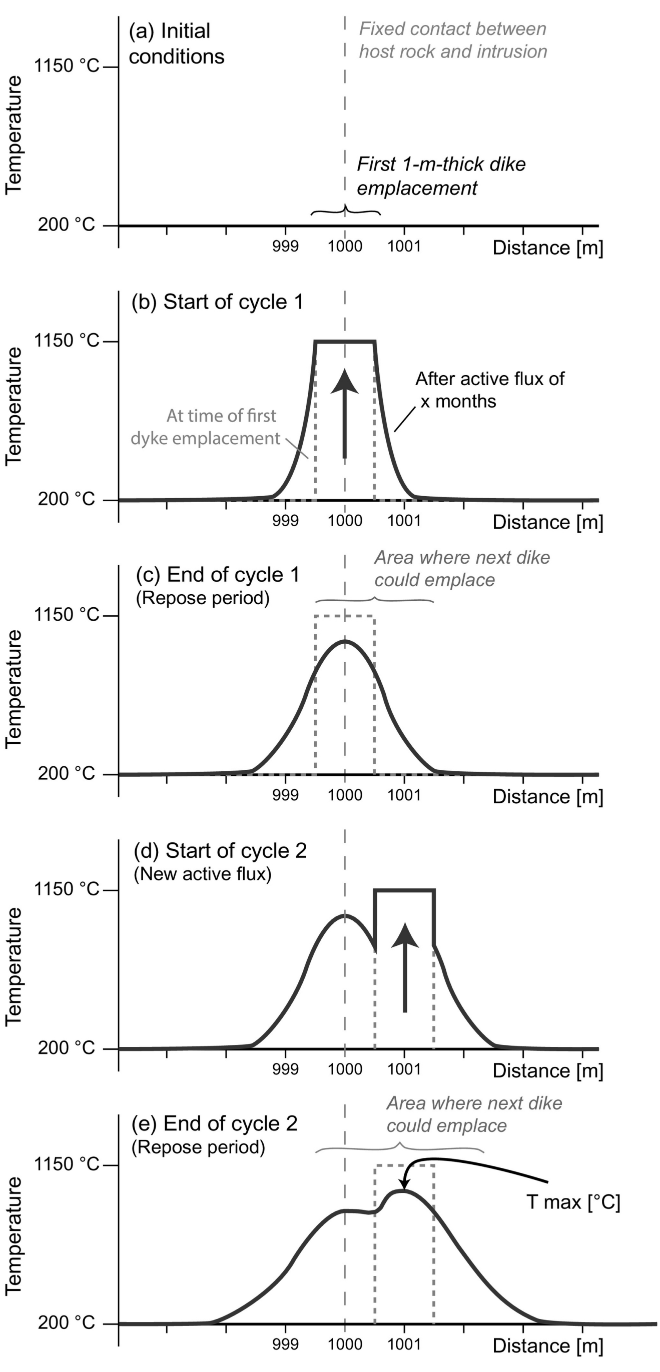
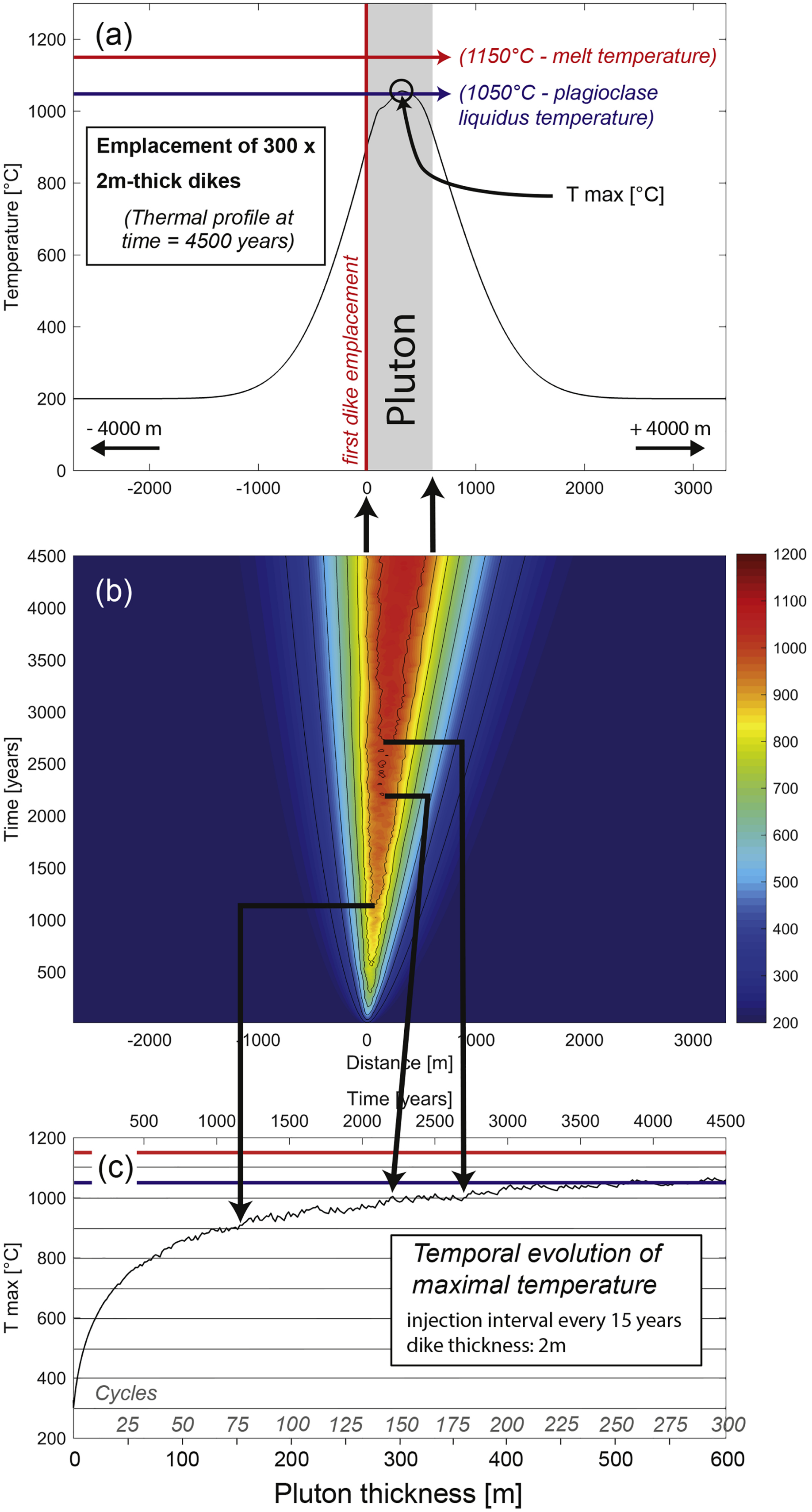

(a) Initial conditions
Fixed contact between host rock and intrusion
First 1-m-thick dike emplacement
1150 °C
200 °C
Temperature
999
1000
1001
Distance [m]
(b) Start of cycle 1
After active flux of x months
At time of first dyke emplacement
1150 °C
200 °C
Temperature
999
1000
1001
Distance [m]
(c) End of cycle 1
(Repose period)
Area where next dike could emplace
1150 °C
200 °C
Temperature
999
1000
1001
Distance [m]
(d) Start of cycle 2
(New active flux)
1150 °C
200 °C
Temperature
999
1000
1001
Distance [m]
(e) End of cycle 2
(Repose period)
Area where next dike could emplace
T max [°C]
1150 °C
200 °C
Temperature
999
1000
1001
Distance [m]

Figure 3



(a)
(1150°C - melt temperature)
(1050°C - plagioclase liquidus temperature)
Emplacement of 300 x 2m-thick dikes
(Thermal profile at time = 4500 years)
T max [°C]
Temperature [°C]
first dike emplacement
Pluton
- 4000 m
+ 4000 m
(b)
Time [years]
Distance [m]
(c)
Time [years]
T max [°C]
Temporal evolution of maximal temperature
injection interval every 15 years
dike thickness: 2m
Cycles
Pluton thickness [m]

Figure 4

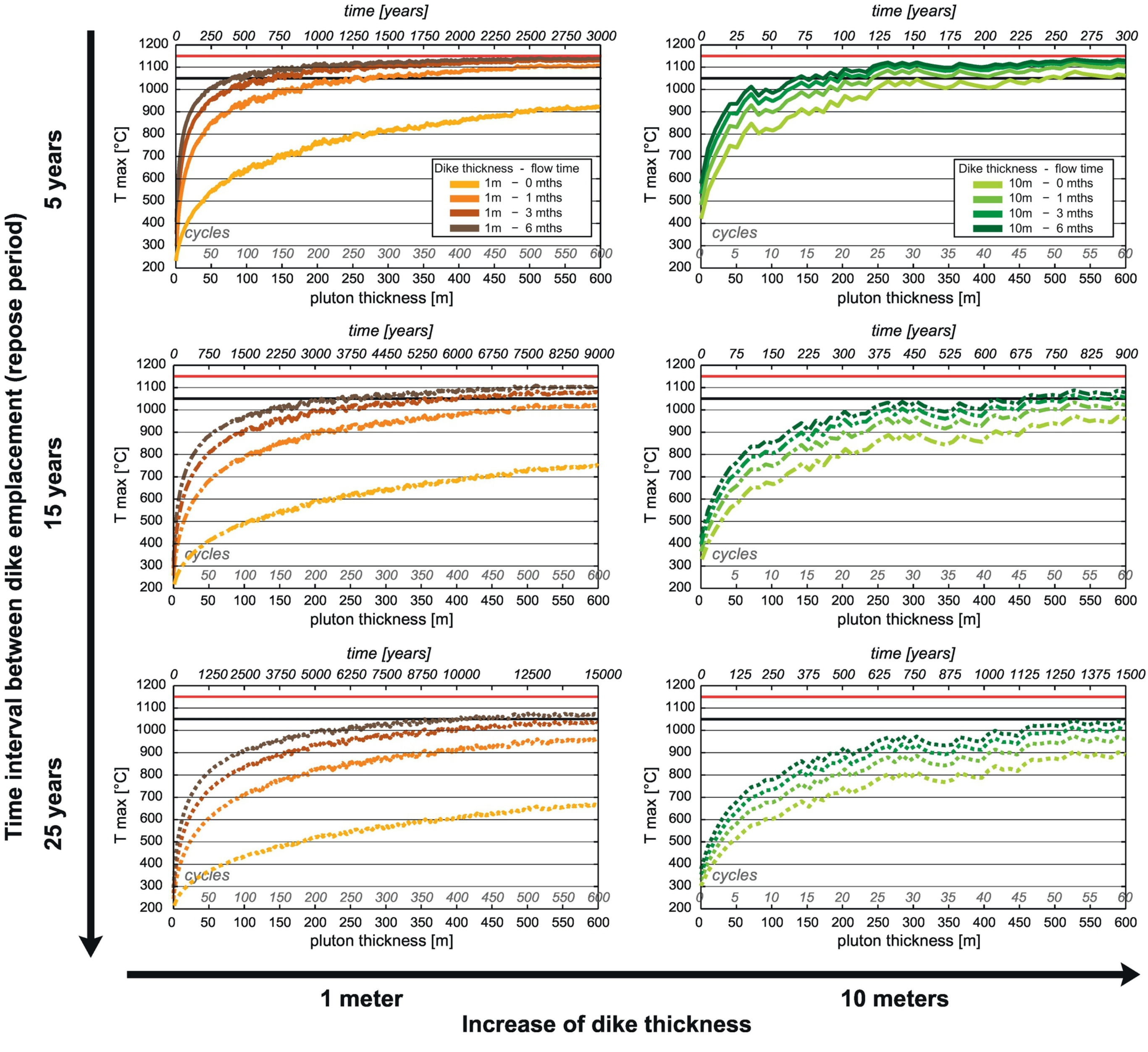
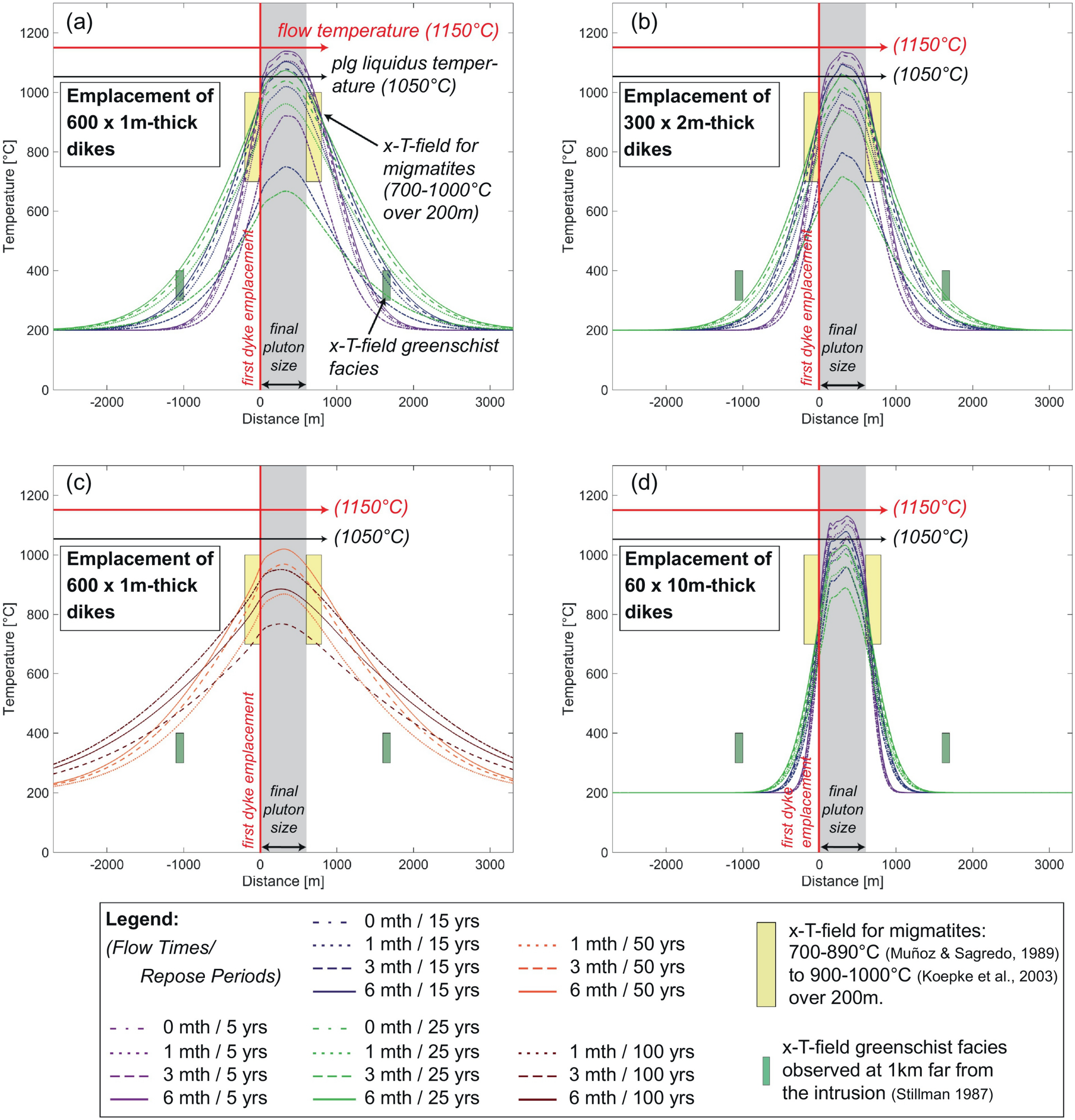

Time interval between dike emplacement (repose period)
5 years
15 years
25 years
time [years]
T max [°C]
pluton thickness [m]
cycles
Dike thickness - flow time
1m – 0 mths
1m – 1 mths
1m – 3 mths
1m – 6 mths
10m – 0 mths
10m – 1 mths
10m – 3 mths
10m – 6 mths
1 meter
10 meters
Increase of dike thickness

Figure 5



(a)
flow temperature (1150°C)
plg liquidus temper-
ature (1050°C)
Emplacement of
600 x 1m-thick
dikes
x-T-field for
migmatites
(700-1000°C
over 200m)
x-T-field greenschist
facies
first dyke emplacement
final
pluton
size
Temperature [°C]
Distance [m]
(b)
(1150°C)
(1050°C)
Emplacement of
300 x 2m-thick
dikes
(c)
Emplacement of
600 x 1m-thick
dikes
(d)
Emplacement of
60 x 10m-thick
dikes
Legend:
(Flow Times/
Repose Periods)
0 mth / 5 yrs
1 mth / 5 yrs
3 mth / 5 yrs
6 mth / 5 yrs
0 mth / 15 yrs
1 mth / 15 yrs
3 mth / 15 yrs
6 mth / 15 yrs
0 mth / 25 yrs
1 mth / 25 yrs
3 mth / 25 yrs
6 mth / 25 yrs
1 mth / 50 yrs
3 mth / 50 yrs
6 mth / 50 yrs
1 mth / 100 yrs
3 mth / 100 yrs
6 mth / 100 yrs
x-T-field for migmatites:
700-890°C (Muñoz & Sagredo, 1989)
to 900-1000°C (Koepke et al., 2003)
over 200m.
x-T-field greenschist facies
observed at 1km far from
the intrusion (Stillman 1987)

Figure 6

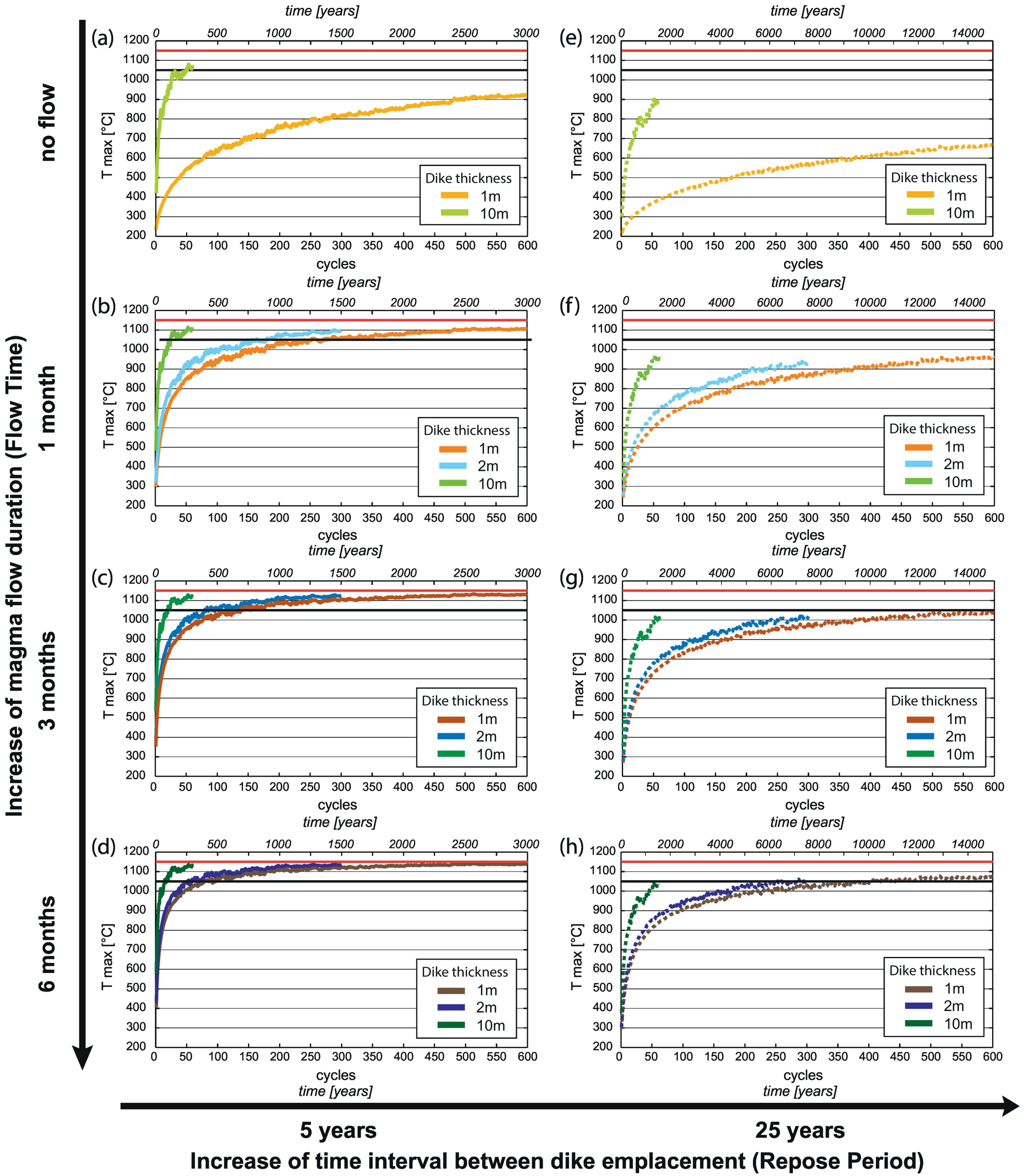
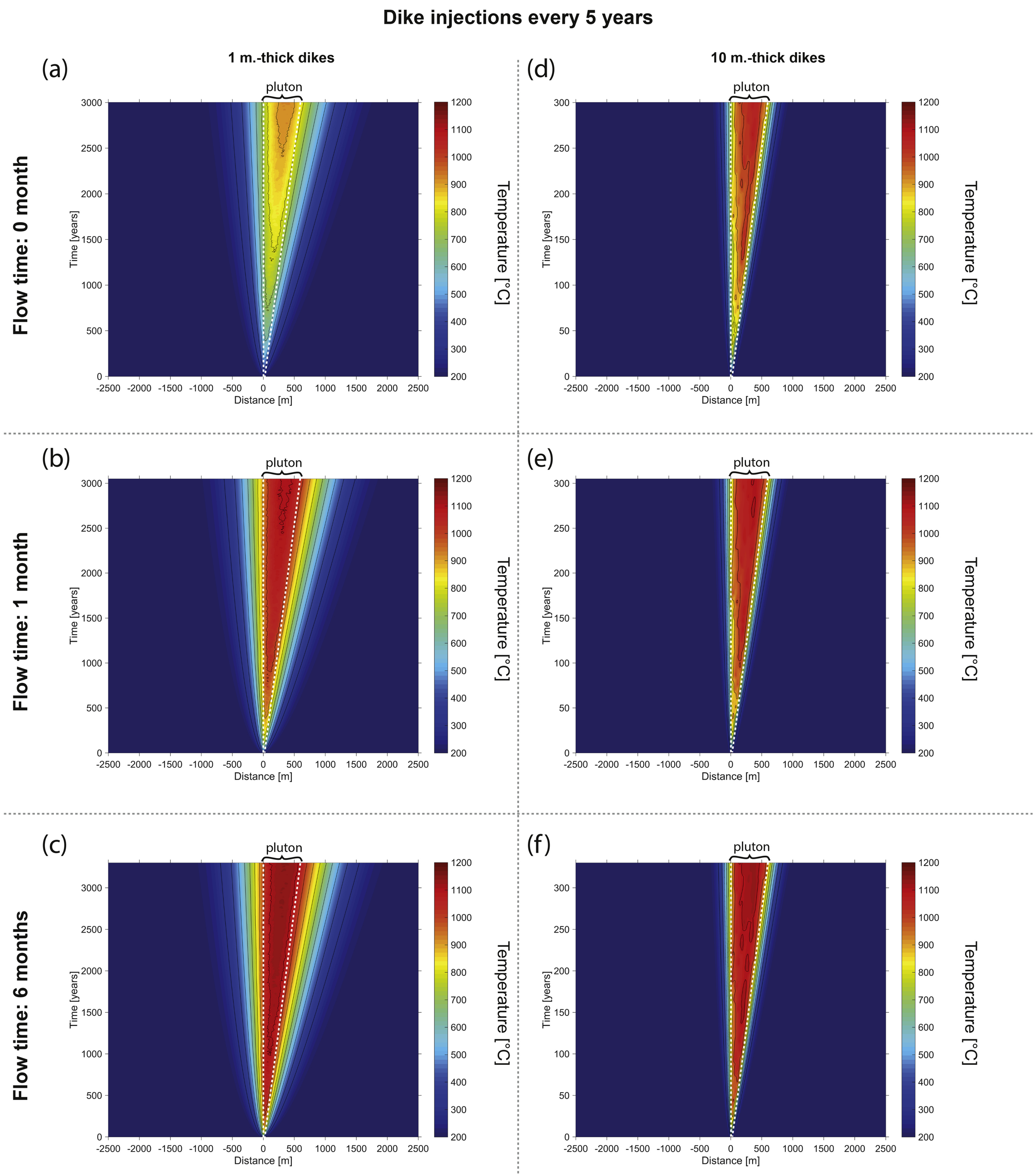
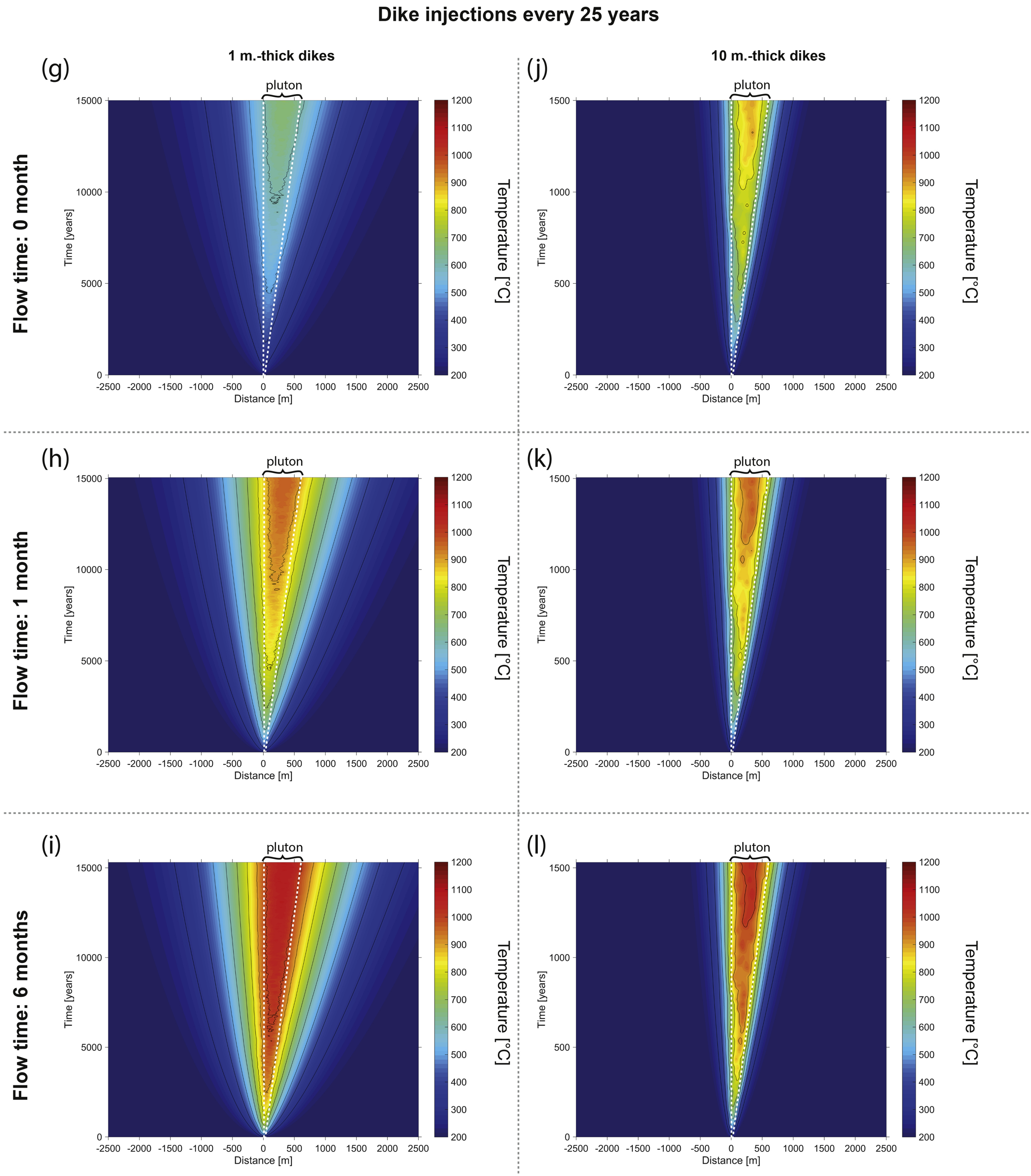

Increase of magma flow duration (Flow Time)
no flow
1 month
3 months
6 months
(a)
(b)
(c)
(d)
(e)
(f)
(g)
(h)
time [years]
cycles
T max [°C]
Dike thickness
1m
2m
10m
5 years
25 years
Increase of time interval between dike emplacement (Repose Period)

Figure 7



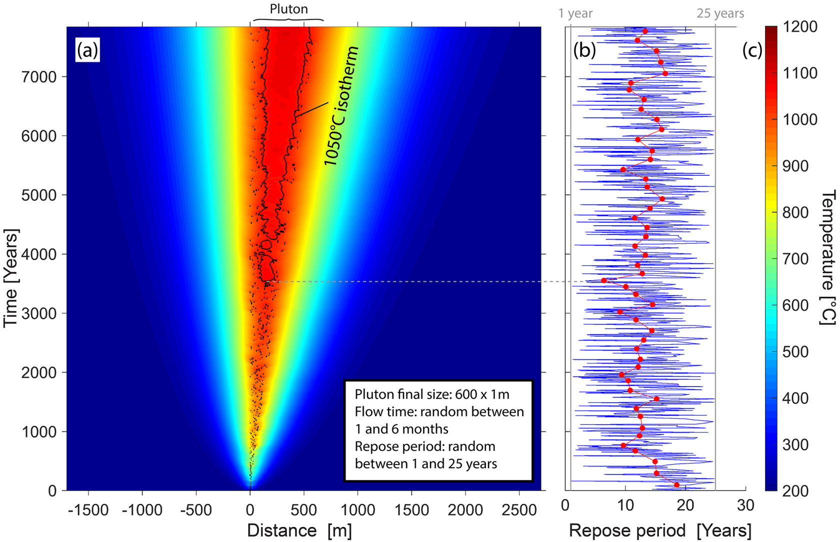
Dike injections every 5 years
1 m.-thick dikes
10 m.-thick dikes
(a)
(b)
(c)
(d)
(e)
(f)
Flow time: 0 month
Flow time: 1 month
Flow time: 6 months
pluton
Time [years]
Distance [m]
Temperature [°C]

Figure 8A



Dike injections every 25 years
1 m.-thick dikes
10 m.-thick dikes
(g)
(h)
(i)
(j)
(k)
(l)
Flow time: 0 month
Flow time: 1 month
Flow time: 6 months
pluton
Time [years]
Distance [m]
Temperature [°C]

Figure 8B



Pluton
(a)
(b)
(c)
1 year
25 years
1050°C isotherm
Time [Years]
0
1000
2000
3000
4000
5000
6000
7000
-1500
-1000
-500
0
500
1000
1500
2000
2500
Distance [m]
0
10
20
30
Repose period [Years]
Temperature [°C]
200
300
400
500
600
700
800
900
1000
1100
1200
Pluton final size: 600 x 1m
Flow time: random between 1 and 6 months
Repose period: random between 1 and 25 years

Figure 9

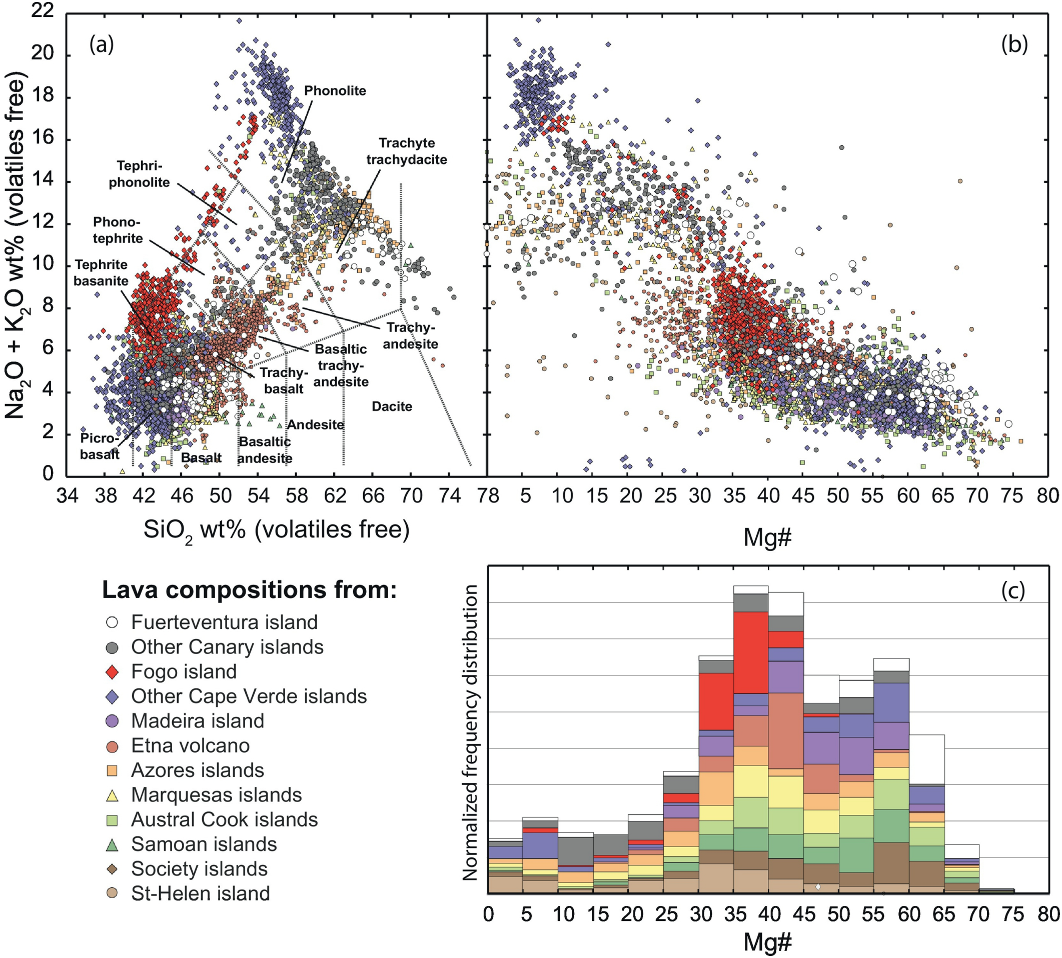

(a)
(b)
(c)
Phonolite
Trachyte trachydacite
Tephri-phonolite
Phono-tephrite
Tephrite basanite
Trachy-andesite
Basaltic trachy-andesite
Trachy-basalt
Andesite
Dacite
Picro-basalt
Basalt
Basaltic andesite
$Na_2O + K_2O$ wt% (volatiles free)
SiO_2 wt% (volatiles free)
Mg#
Normalized frequency distribution
Lava compositions from:
Fuerteventura island
Other Canary islands
Fogo island
Other Cape Verde islands
Madeira island
Etna volcano
Azores islands
Marquesas islands
Austral Cook islands
Samoan islands
Society islands
St-Helen island

Figure 10

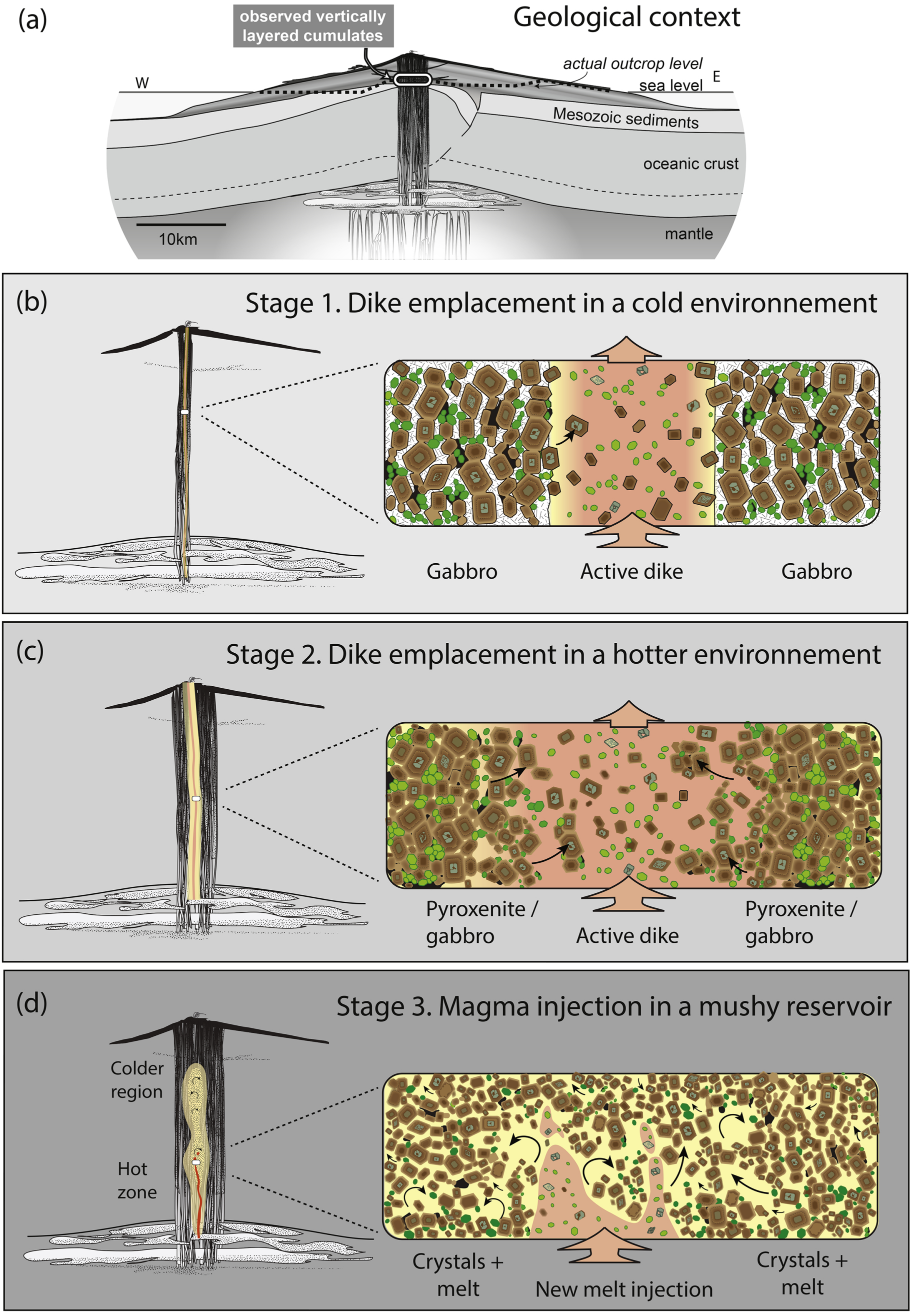

(a)
observed vertically layered cumulates
Geological context
actual outcrop level
W
sea level
E
Mesozoic sediments
oceanic crust
10km
mantle
(b)
Stage 1. Dike emplacement in a cold environnement
Gabbro
Active dike
Gabbro
(c)
Stage 2. Dike emplacement in a hotter environnement
Pyroxenite / gabbro
Active dike
Pyroxenite / gabbro
(d)
Stage 3. Magma injection in a mushy reservoir
Colder region
Hot zone
Crystals + melt
New melt injection
Crystals + melt

Figure 11