Water plays a key role in magma genesis, differentiation, ascent and, finally, eruption. Despite the recognized crucial function of water, there are still several issues that continue to blur our view about its role in magmatic systems. What are the timescales of H₂O accumulation in crystallizing magmas? What are the ascent rates of water-rich residual melts leading to explosive eruptions? Here, we track the timescale of water accumulation in a residual melt resulting from crystallization of a hydrous CO₂-bearing magmatic mass stored at mid- to deep-crustal levels in a subduction-related geodynamic setting. Our results indicate that, after a repose period ranging from few to several thousand years, water-rich melts with water concentrations larger than 6–9 wt.% can migrate towards the Earth surface in very short timescales, on the order of days or even hours, possibly triggering explosive eruptions with short warning times and devoid of long-term geophysical precursors.
The presence of volatiles in subduction zones plays a fundamental role in many processes involved in magma genesis and differentiation. Among them, water is the most abundant, and it has a significant effect on the rheological behavior of magmas, on the chemistry of growing crystals, and as a consequence, on the lifetime evolution of a magmatic system. The second volatile species in magmas is carbon dioxide, which also plays a significant role in subduction zones, modulating, for example, the solubility of water and controlling the initial stages of degassing of ascending magmas.

Also, at depths where the magmatic pressure is below the saturation threshold, volatile species can exsolve rapidly, potentially causing magma chamber rejuvenation and triggering explosive eruptions. The volatile of volatile exsolution can be strongly affected by the ascent rate of volatile-rich magmas, finally exerting a significant role in modulating the eruptive style.

Melt inclusions in minerals can provide useful information about volatile contents in magmatic systems. During their growth in magmas, minerals occasionally trap small parcels of melt (preserved as inclusions), providing a record of the volatile content of the system. Available data from arc volcanoes, point to the presence of higher H₂O contents than those recorded by melt inclusions. For example, Edmonds et al. hypothesized the existence of composite bodies of water-rich (6–9 wt.%) eruptible magma extending vertically (down to 16 km) within the upper crust beneath the volcanic island of Montserrat (Caribbean Sea). The presence of water-rich magmas (H₂O ≥ 8.0 wt.%) at mid-crustal levels is also supported by recent studies combining geophysical investigations, laboratory experiments, and petrological data to constrain deep electrical conductivity anomalies beneath the Uturuncu Volcano (Central Andes). Also, similar features have been observed beneath the Cascades volcanic arc and Taupo Volcanic Zone, suggesting that the presence of water-rich magmas could be a common feature of active continental arcs. As for carbon dioxide, Blundy et al. suggest, in agreement with Wallace, an average content of ~0.3 wt.% in arc parental magmas.

Despite the developments mentioned above, our knowledge about the fate of volatile-rich magmas in the Earth crust is still actively debated and deserves further investigation. Key questions include: What are the timescales for volatiles accumulation in crystallizing magmas at mid- to deep-crustal levels? What are the ascent velocities of volatile-rich residual melts (produced by the crystallization of volatile-bearing parental magmas at mid- to deep-crustal levels) toward the Earth surface, eventually culminating in explosive volcanic eruptions?

To answer the first question, we present new numerical simulations, targeted at reproducing the thermal and petrological evolution of magmas at mid- to deep-crustal levels. In particular, the thermochemical behavior of a subduction-related, volatile-bearing, magmatic body is investigated.

The model accounts for different initial chemical compositions of the parental magma and initial water contents ranging from 2 to 4.5 wt.% (please refer to the “Methods” for a complete description of the model development and parameterization) both in presence and absence of CO₂. It also accounts for different storage conditions (0.7–1.0 GPa corresponding to ~20–35 km in depth) of both closed and open magmatic systems.

In particular, we first investigated a closed magmatic system exchanging heat with the hosting rocks only by conduction. For comparison, an open system consisting of progressive refilling by new parental magma is also modeled. Finally, we studied the effect of the development of convective motions within the system, accounting for the non-Newtonian rheology of the magma during the crystallization and considering Rayleigh numbers (Ra) up to 10⁹. The numerical simulations started at the liquidus temperature (i.e., no crystals are present within the system) and continued until the attainment of a rheological behavior close to the jamming conditions, i.e., when the melt extraction from the magmatic mush becomes most probable. The fate of the volatile-rich residual melts segregated from the crystal-rich mushes is also investigated. In particular, we provide a comprehensive discussion of the factors promoting and contrasting the rising of volatile-rich melts in the crust. Finally, the timescales for the transfer of volatile-rich melts toward the Earth surface, eventually culminating in volcanic eruptions, are estimated. Results indicate that water-rich melts can ascend rapidly to the surface over timescales of hours to days with very short warning times.

**Results**

**H₂O and CO₂ solubilities in arch magmas.** In order to investigate the timescales of volatile accumulation in subduction zones at mid- to deep-crustal levels, the H₂O and CO₂ solubilities in mafic primitive arc magmas above the liquidus temperature (i.e., before the beginning of the crystallization process) must be considered. We focus on pressures between 0.7 and 1.0 GPa, corresponding to crustal depths ranging between ~20 and 35 km. Figure 1a displays the saturation conditions for one of the starting magmatic compositions considered here as a proxy for primitive mafic arc magmas. The same curves for the other parental magma compositions that we considered are reported as Supplementary Material (Supplementary Fig. 1a, b). Figure 1a shows that the maximum solubility of carbon dioxide in the system at 1.0 GPa is ~0.90 wt.%, corresponding to a water content of ~3.8 wt.%. H₂O and CO₂ solubilities progressively drop down as pressure decreases (Fig. 1a). In a magmatic system subjected to isobaric cooling (i.e., crystallizing at a constant depth), melt solubilities for water and carbon dioxide will progressively evolve in response to thermal and chemical changes. As an example, Fig. 1b reports the evolution of the H₂O and CO₂ saturation conditions for the residual melt of a magmatic system crystallizing at 0.7 GPa. Figure 1b highlights that, during the crystallization process, the maximum solubility of water progressively increases from ~10.2 to ~12.0 wt.%. On the contrary, carbon dioxide solubility gradually decreases from ~0.53 to ~0.36 wt.%.

**Thermodynamic modeling.** Thermodynamic modeling can be also used to track the evolution of H₂O and CO₂ during the crystallization of a parental magma at pressures ranging from 0.7 to 1.0 GPa (see the “Methods” section for further details). As an example, we report the cases for magmas characterized by an
The melt phase progressively accumulates H$_2$O and CO$_2$ up to the attainment of the saturation conditions due to the development of H$_2$O- and CO$_2$-free phases in the early evolution of the crystallization process. Hence, the melt starts exsolving a fluid phase, mostly composed of the least soluble volatile component (i.e., CO$_2$). As a consequence, after reaching the saturation conditions, CO$_2$ concentrations are buffered at the saturation levels ($\sim$0.33 wt.% in Fig. 1c). In contrast, the evolution of H$_2$O is slightly affected by the attainment of the saturation conditions. At 1.0 GPa (Fig. 1d), the overall behavior of the system during an isobaric cooling is similar to that observed at 0.7 GPa (Fig. 1c). In this case, due to the higher solubilities for both water and carbon dioxide, the maximum CO$_2$ concentration is $\sim$0.79 wt.% at a crystal mass fraction ($X_{\text{cryst}}$) approaching 0.5 (Fig. 1d). The behavior of a parental magma richer in H$_2$O ($X_{\text{H2O}}$ = 4.5 wt.%) is similar; the detailed description of the simulations is provided as Supplementary Material (Supplementary Fig. 2). In this case, the maximum final concentration of carbon dioxide in the melt phase is $\sim$0.8 and 0.3 wt.% for pressures of 1.0 (Supplementary Fig. 2a) and 0.7 GPa (Supplementary Fig. 2b), respectively. The impact of the attainment of volatile saturation conditions on the magmatic system can be tracked monitoring the relative increment of H$_2$O ($\Delta$H$_2$O; defined as the ratio between the H$_2$O content and its initial value) in the melt phase plotted against the melt fraction ($X_{\text{melt}}$; Fig. 2a). Figure 2a shows that, at 1.0 GPa, the melt phase of the studied systems (3.0 $< X_{\text{H2O}}$ $< 4.5$ wt.% and 0.0 $< X_{\text{CO2}}$ $< 0.6$ wt.%) accumulates water, behaving essentially as an under-saturated system (i.e., no H$_2$O release). On the contrary, at 0.7 GPa, CO$_2$-rich melts may release a larger amount of water, slightly deviating from the behavior of the under-saturated system (Fig. 2b). Also, the crystallization of hydrous phases (e.g., hornblende) may potentially affect the process of water accumulation in the melt phase during crystallization of water-rich arc magmas. For example, hornblende is a ubiquitous phase in volcanic and plutonic rocks produced in magmatic arc systems$^{34}$. Although, amphibole does not crystallize massively in the experiments considered here$^{30-33}$ (please refer to the “Methods” section for further details), its potential effect on the evolution of the system needs to be considered.

Fig. 1 Water and carbon dioxide solubilities in the melt phase. Solubility curves for water and carbon dioxide calculated for the mid-MgO system$^{30}$: a Solubilities of the parental melt at pressures of 0.7 and 1.0 GPa and liquidus temperatures; b Solubilities of the residual melt during the crystallization process at a pressure of 0.7 GPa. The SiO$_2$ content in the residual melt on dry basis is 50.3, 50.9, and 57.2 wt.% respectively; c Equilibrium thermodynamical models performed by rhyolite-MELTS$^{59}$ at 0.7 GPa displaying the evolution of H$_2$O and CO$_2$ in the residual melt during the crystallization process; d Equilibrium thermodynamical models performed by rhyolite-MELTS$^{59}$ at 1.0 GPa displaying the evolution of H$_2$O and CO$_2$ in the residual melt during the crystallization process. The parameter $X_{\text{cryst}}$ reported in c, d is the crystal mass fraction.
Mass balance modeling. The evolution of water content in the melt phase during amphibole crystallization can be tracked by mass balance modeling (Fig. 2c; please refer to the “Methods” section for further details). Figure 2c compares the evolution of $\Delta H_2O$ vs. the melt mass fraction ($X_{\text{melt}}$) during the crystallization of only dry phases (e.g., the sequence olivine, pyroxene, spinel, and plagioclase), the crystallization of a hornblende-bearing magma (e.g., the sequence olivine, pyroxene, spinel, plagioclase, and hornblende), and a hornblende-bearing over-saturated magma where volatile exsolution also contributes to the process of water subtraction to the melt phase. Data extracted from the used experimental petrology data set are also reported in Fig. 2c. The plot shows that crystallization of amphibole does not produce substantial changes in $\Delta H_2O$ of the magmatic system as indicated by the fact that crystallization of amphibole and volatile exsolution might decrease the $\Delta H_2O$ parameter by maximum $\sim$0.2 at the attainment of crystal mass fractions ($X_{\text{cryst}}$) of 0.5. This being so, in the following sections we will consider the crystallization of dry phases in an under-saturated system up to the attainment of $\Delta H_2O$ equal to 2.0 (i.e., a crystal mass fraction of 0.5).

Timescales for $H_2O$ accumulation in the residual melt. In order to define the timescales for $H_2O$ accumulation, we performed thermal numerical simulations (please refer to the “Methods” section for further details) monitoring the evolution of the $\Delta H_2O$ parameter (calculated as ratio between the water content in the residual melt and the initial $H_2O$ concentration in the system) against the melt mass fraction ($X_{\text{melt}}$). a Results of equilibrium thermodynamical simulations performed by rhyolite-MELTS showing the evolution of the $\Delta H_2O$ parameter in the residual melt of a system crystallizing at 1.0 GPa. b Results of equilibrium thermodynamical simulations performed by rhyolite-MELTS showing the evolution of the $\Delta H_2O$ parameter in the residual melt of a system crystallizing at 0.7 GPa. c Mass balance modeling of the $\Delta H_2O$ parameter displaying the evolution of $\Delta H_2O$ parameter during the crystallization of a magmatic system. Different curves are reported for the crystallization of an anhydrous crystal assemblage (purple curve), a hornblende-bearing crystal assemblage (yellow curve), and an over-saturated hornblende-bearing crystal assemblage (brown curve). The data extracted from the investigated petrological data set are also reported in the picture with error bars expressed as one standard deviation.

**Fig. 2** Water accumulation in the melt phase during crystallization. Evolution of the $\Delta H_2O$ parameter (calculated as ratio between the water content in the residual melt and the initial $H_2O$ concentration in the system) against the melt mass fraction ($X_{\text{melt}}$). a Results of equilibrium thermodynamical simulations performed by rhyolite-MELTS showing the evolution of the $\Delta H_2O$ parameter in the residual melt of a system crystallizing at 1.0 GPa. b Results of equilibrium thermodynamical simulations performed by rhyolite-MELTS showing the evolution of the $\Delta H_2O$ parameter in the residual melt of a system crystallizing at 0.7 GPa. c Mass balance modeling of the $\Delta H_2O$ parameter displaying the evolution of $\Delta H_2O$ parameter during the crystallization of a magmatic system. Different curves are reported for the crystallization of an anhydrous crystal assemblage (purple curve), a hornblende-bearing crystal assemblage (yellow curve), and an over-saturated hornblende-bearing crystal assemblage (brown curve). The data extracted from the investigated petrological data set are also reported in the picture with error bars expressed as one standard deviation.
The static pure conductive case (PCC) scenario is also reported as a reference. The dynamic cases are dramatically more efficient in accumulating H$_2$O in the residual melt (Fig. 3e) than the static configurations because of the faster cooling due to convection, resulting in higher crystallization rates for olivine, clinopyroxene, and plagioclase. Residual melts deriving from DC-1, DC-2, and DC-3 reach the H$_2$O concentration threshold at $\tau$ of ~0.069, 0.057, and 0.048, respectively, up to ~2.3 times faster than the static PCC scenario.

The relationships emerging from the numerical experiments equip us with a new tool to understand the timescales of water accumulation in subduction-related residual melts. In dimensional terms, the following relationships, linking the size of the magmatic reservoir ($L$) and the time ($t$) required to reach the threshold water content (i.e., $\Delta$H$_2$O larger than 2.0), hold:

\begin{align}
\text{PCC} &: \log_{10}[t] = 2\log_{10}[L] + 0.63, \\
\text{DC-1} &: \log_{10}[t] = 2\log_{10}[L] + 0.44, \\
\text{DC-2} &: \log_{10}[t] = 2\log_{10}[L] + 0.35, \\
\text{DC-3} &: \log_{10}[t] = 2\log_{10}[L] + 0.28,
\end{align}

where $L$ and $t$ are expressed in km and ky, respectively. Concerning a possible natural scenario, these equations indicate that a magma body with a lateral dimension ($L$) equal to 5.0 km will attain the water concentration threshold in 69, 56 and 48 ky.
Table 1 Initial thermophysical properties of the magmatic system

| Property                  | Value range   | Unit  |
|---------------------------|---------------|-------|
| Density, \( \rho \)        | 2540-2750     | kg m\(^{-3}\) |
| Dynamic viscosity, \( \eta \) | 1.2-8         | Pa s  |
| Thermal diffusivity, \( \alpha \) | \(8.0 \times 10^{-7}-8.7 \times 10^{-7}\) | m\(^2\) s\(^{-1}\) |
| Thermal conductivity, \( k \)     | 2.2           | W m\(^{-1}\) K\(^{-1}\) |
| Specific heat, \( c_p \)         | 1000          | J kg\(^{-1}\) K\(^{-1}\) |
| Latent heat of crystallization, \( L_m \) | \(3.5 \times 10^5\) | J kg\(^{-1}\) |
| Temperature difference, \( \Delta T \) | 105-190       | K     |
| Prandtl number, \( Pr \)        | 545-3636      | –     |
| Stephan number, \( Ste \)       | 0.3-0.54      | –     |

Ranges of thermophysical properties of the magmatic system at the initial state as reported in Supplementary Table 1.

Discussion

Numerical experiments indicate that the time evolution of \( \text{H}_2\text{O} \) concentration is strongly dependent on the dynamic conditions of the magmatic system, with convection playing a key role.

Figure 5 compares the variation of viscosity as a function of temperature for crystal-bearing magmas with a constant \( \text{H}_2\text{O} \) content of 3.0 wt.% (Fig. 5a) and 4.5 wt.% (Fig. 5b) at 0.7 and 1.0 GPa, respectively. Different aspect ratios ranging from 1 to 3 have been investigated for forming minerals, in agreement with aspect ratios expected during crystallization of olivine, pyroxene, and amphibole. Notably, the rheological behavior of the crystal-bearing magmas does not change significantly for these aspect ratios (Fig. 5). Also, the adopted rheological model agrees well with that reported by Laumonier et al. The viscosity of the hydrous residual melts (where \( \text{H}_2\text{O} \) is progressively accumulating with time) are also reported in Fig. 5. At the attainment of the jamming conditions (i.e., close to the maximum packing fraction), the viscosities of the residual melts are \(~40-50\) and \(7-10\) Pa s, respectively. These values, calculated following Giordano et al., are several orders of magnitude lower than those of the crystal-bearing magma (hydrous melt phase plus crystals). In agreement with Mader et al., the jamming condition occurs at a crystal volume fraction \( (\phi) \) of \(~0.5\), which represents the lower boundary of the crystallinity window \((0.5 < \phi < 0.7)\), where the extraction of melt from the crystal mush is statistically most probable. This scenario, the produced residual melts are characterized by very low viscosities \((\eta \text{ equal to } 7-10 \text{ and } 40-50 \text{ Pa s, respectively})\) and high water contents, which make them suitable for migrating to shallower crustal levels. This implies that the ascending melt can potentially refill shallow magmatic reservoirs or rise directly to the Earth surface causing an eruption.

Two main scenarios are portrayed in Fig. 6. The first scenario is a melt ascending from a depth of 25 km (Fig. 6a, c) and characterized by a viscosity of 50 Pa s. The second scenario considers a melt moving from 35 km depth (Fig. 6b, d) with a viscosity of 10 Pa s. These viscosity values are in agreement with the results reported in Fig. 5a. Figure 6a, b indicates that buoyancy-driven, water-rich, magmas produced at mid- to deep-crustal levels, can arrive at shallow crustal levels with velocities ranging from 0.15 to 8 m s\(^{-1}\). However, many studies suggest that the buoyancy is unlikely to be the sole process driving magma to the surface. Indeed, in the magma-driven formulation, the dynamic of the crack propagation is governed by both the buoyancy of the fluid and the reservoir’s excess pressure. Figure 6a, b indicates that magma-driven, water-rich melts can reach the Earth surface with velocities ranging from \(~1.8\) to \(\sim 20\) m s\(^{-1}\) for overpressures \( (\Delta P_{\text{excess}}) \) ranging from 2 to 3 MPa. These overpressures are in agreement with typical \( \Delta P_{\text{excess}} \) values estimated for natural magmatic systems. Enhanced ascent rates are also expected due to magma vesiculation at shallow levels (below \(~10\) km), where the presence of gas bubbles contribute to lower melt density and, therefore, enhance the buoyancy effect. We estimated the effect of magma vesiculation (i.e., compressible fluid...
behavior (in agreement with Taisne and Jaupart46 (Fig. 6c, d). Figure 6c, d highlights a significant increase of rising velocities at depths lower than ~10 km. In particular, ascending velocities are increased by a factor ~2 and 10 at depths corresponding to ~4 and 1 km, respectively. The obtained magma ascent velocities are in agreement with estimates (velocity in the range 0.1–25 m s\(^{-1}\)) provided for natural systems46–54. As an example, Lloyd et al.50 suggested rising velocities up to ~20 m s\(^{-1}\) for Volcán de Fuego eruption occurred on October 17, 1974, based on multi-volatile profiles obtained from olivine-hosted melt embayments.

Another process that can affect the ascent velocity, playing the opposite role of buoyancy, is decompression-driven melt crystallization41. The increase of the crystal volume fraction during ascent produces an increase of the viscosity of the mixture, which, in turn, generates a decrease in the ascent velocity. The petrological evolution of ascending water-rich melts was considered by Annen et al.41 who postulated a viscous death of water-rich ascending magmas at depths of ~10–15 km due to depressurization-induced crystallization. The model proposed by Annen et al.41 successfully describes the system at physical conditions close to the thermodynamical equilibrium. However, the estimated rising velocities and those reported in the literature for natural systems suggest that the ascending magmas could not be at equilibrium conditions52. In particular, La Spina et al.52 indicated that basaltic magmas ascending from ~10 km depth need approximately 2 h to reach the equilibrium crystal volume content. Therefore, when the average ascent velocity is higher than ~1 m s\(^{-1}\), the system does not have enough time to reach the equilibrium condition, resulting in a delayed crystallization and leading to a lower viscosity of the mixture. These estimates indicate that, if the ascent velocity is high enough, crystallization can be inhibited and, therefore, water-rich ascending magmas can readily reach the Earth surface.

A potential scenario for the evolution of a crustal arc section is illustrated in Fig. 7. An initial batch of magma containing ~3–4.5 wt.% of water, stored at mid- to deep-crustal levels (~20–35 km), can evolve to a water-rich (H\(_2\)O equal to ~6–9 wt.%) residual melt in timescales ranging from a few to several thousand years (see Eq. (6)). These results are in agreement with typical timescales obtained using crystal residence times based on 238U–230Th and 230Th–226Ra disequilibrium40. If the ascent velocities are too slow (e.g., below the critical velocity), magma storages at mid- to shallow-crustal levels feeding the incremental growth of batholiths might be favored41,55. On the contrary, if rising velocities exceed \(v_c\) and are large enough to maintain the system far from the thermodynamic equilibrium, water-rich magmas can reach the Earth surface in short timescales (of the order of days or even hours)42.

In the most hazardous scenario (Fig. 7), after a storage of several thousand years at mid- to deep-crustal levels, water-rich melts can migrate to the surface in a few hours, possibly triggering explosive eruptions with little warning time and devoid of long-term geophysical warnings that can be used for alerting purposes4,36,37.

Our results suggest a strategy for the risk mitigation of these sudden eruptions. Geophysical investigations should be targeted at detecting the presence of crystal-rich magmatic mushes at mid- to deep-crustal levels characterized by water-rich residual melts, as well as recognizing possible signs of water-rich magmatic systems deep within the Earth crust that might rapidly evolve toward explosive eruptions.

**Methods**

**Conceptual model.** The evolution of a magmatic system crystallizing at mid- to deep-crustal levels was modeled considering two main scenarios: a closed system where the magmatic body remains isolated from external magmatic inputs and an...
The initial conditions for the closed system scenario are schematically shown in Fig. 8a. It consists of a two-dimensional (2D) rectangular domain initially filled with a magma at the liquidus temperature27 (Fig. 8a). During its evolution, the magmatic system cools due to heat exchange with the host rock (Fig. 8c–f). The open-system behavior is simulated in agreement with the progressive incremental growth model proposed by Annen et al.41,55.

The initial configuration for the open-system scenario (Fig. 8b) consists of a magmatic sill initially emplaced at its liquidus temperature (Fig. 8b) and progressively grows by the periodical addition of new sills at the base of the system (Fig. 8g–j)41,55.

**Petroleum data set**. The petrological data set considered for the parametrization used in the present study consists of published experimental data dealing with crystallization of hydrous magmas at mid- to deep-crustal levels (~20–35 km in depth) corresponding to pressures between 0.7 and 1.0 GPa30–33. As the focus is on hydrous magmas, experiments were filtered to consider only those where the initial water content was between 2.0 and 4.5 wt.% (Supplementary Table 1). This allowed us to develop a model parametrization based on a comprehensive set of 26 experiments30–33 with starting materials having compositions of primitive hydrous mafic arc magmas crystallizing at mid- to deep-crustal levels (Supplementary Table 2). In detail, the starting materials are a primitive high-MgO (14.6 wt.% MgO) basalt from Soufrière (St. Vincent)31,32, a near-primary olivine–tholeiite mid-MgO (8.7 wt.% MgO) dike composition from the southern part of the Tertiary Adamello batholith (Northern Italy)32, and a relatively magnesian basalt (8.7 wt.% MgO) typical of magmatic magmas erupted in the Cascades, near Mount Rainier (Washington, USA)33. The investigated starting materials represent typical compositions for super- and near-liquidus magmas in subduction-related environments33. The experiments are described in detail in literature30–33 and they are characterized by two key features. The first is that dry phases dominate the crystallization process before attaining the jamming conditions (i.e., a crystal volume fraction, ϕc, close to 0.5). The second is that the first hydrous phase is always hornblende30–33, mostly appearing in experiments at crystal volume fractions larger than 0.5.

**H2O and CO2 solubility model**. The solubilities and the saturation conditions for H2O and CO2 in the melt phase were estimated using the formulation proposed by Ghiorso and Gualda58. It consists of a thermodynamic model for estimating the saturation conditions of H2O–CO2 mixed fluids in multi-component silicate liquids58. As reported by Ghiorso and Gualda58, the model is calibrated from published experimental data on water and carbon dioxide solubility, and mixed fluid saturation in silicate liquids. The model assumes that H2O dissolves to form a hydroxyl melt species, and that CO2 dissolves to compose both a molecular species...
and a carbonate ion, the latter complexed with calcium. The model is restricted to liquids of natural compositions over the pressure range 0–3 GPa making it suitable for the study of magmatic systems crystallizing at pressures ranging between 0.7 and 1.0 GPa. It is also suitable to study the H2O and CO2 solubilities during the rising of magmas from 1.0 GPa to the Earth surface.

Thermodynamical evolution of H2O and CO2 in the melt phase. The evolution of H2O and CO2 in the melt phase during the crystallization process have been modeled using rhyolite-MELTS 1.2. Rhyolite-MELTS is a software package that combines the potentials of the MELTS package for thermodynamic modeling with the H2O–CO2 solubility model proposed by Ghiorso and Gualda.

Specifically, to model the evolution of H2O and CO2 during the crystallization process at mid- to deep-crustal levels, we used the dry chemical compositions of the starting materials reported in Supplementary Table 2 and we added different amounts of water and carbon dioxide in the ranges of 2 < H2O < 4.5 wt.% and 0 < CO2 < 0.6 wt.%, respectively. Simulations were started at the liquidus composition and stopped at a crystal mass fraction (~0.5). The NNO buffer has been adopted to constrain the oxygen fugacity, and the numerical experiments have been performed at equilibrium conditions fractionating the solid phase.

Mass balance modeling. Nominal H2O concentrations in the residual melts were estimated by the mass balance method, an application of the mass conservation law.
that accounts for the material entering and leaving the system. In the specific case of a crystallizing magmatic system, the mass balance equation for H$_2$O can be expressed as follow:

$$C_{\text{H}_2\text{O}}^{\text{tot}} = \sum_{i=1}^{N} C_{\text{H}_2\text{O}}^{i} X_{i}, \quad (7)$$

where $C_{\text{H}_2\text{O}}^{\text{tot}}$ is the total H$_2$O concentration in the system, $N$ is the number of phases able to host a significant amount of H$_2$O, $C_{\text{H}_2\text{O}}^{i}$ is the H$_2$O concentration in the phase $i$ and $X_{i}$ is its mass fraction. In the studied system (i.e., a magmatic system crystallizing at mid- to deep-crustal levels in subduction-related environment), the only phases able to host a significant amount of H$_2$O are the residual melt, hornblende (hbl), and the fluid phase. Although biotite can potentially host a significant amount of water, it starts crystallizing only at later stages much beyond the attainment of a crystal mass fraction equal to 0.5. As our numerical experiments stop when $X_{\text{cryst}}$ reaches 0.5, modeling of this mineral phase is not considered. The mass balance Eq. (7) can be written as:

$$C_{\text{H}_2\text{O}}^{\text{tot}} = C_{\text{H}_2\text{O}}^{\text{melt}} + C_{\text{H}_2\text{O}}^{\text{hbl}} + C_{\text{H}_2\text{O}}^{\text{fluid}}, \quad (8)$$

that yields:

$$C_{\text{H}_2\text{O}}^{\text{melt}} = C_{\text{H}_2\text{O}}^{\text{tot}} - C_{\text{H}_2\text{O}}^{\text{hbl}} X_{\text{hbl}} - C_{\text{H}_2\text{O}}^{\text{fluid}} X_{\text{fluid}}. \quad (9)$$

Finally, if the total concentration of water does not change with time in the system ($C_{\text{H}_2\text{O}}^{\text{tot}} = C_{\text{H}_2\text{O}}^{\text{init}}$) the relative H$_2$O increment to the initial composition ($\Delta H_2O$) is:

$$\Delta H_2O = \frac{C_{\text{H}_2\text{O}}^{\text{init}}}{C_{\text{H}_2\text{O}}^{\text{tot}}} = \frac{C_{\text{H}_2\text{O}}^{\text{tot}} - C_{\text{H}_2\text{O}}^{\text{hbl}} X_{\text{hbl}} - C_{\text{H}_2\text{O}}^{\text{fluid}} X_{\text{fluid}}}{C_{\text{H}_2\text{O}}^{\text{init}} X_{\text{melt}}}. \quad (10)$$

The mass balance has been utilized in the present study to model the evolution of $\Delta H_2O$ during the crystallization process. In detail, in the case of a volatile under-saturated system crystallizing anhydrous phases only, Eq. (10) reduces to (purple line in Fig. 2):

$$\Delta H_2O = \frac{1}{X_{\text{melt}}}. \quad (11)$$

In order to test the maximum potential impact of hornblende crystallization on the water content of the magmatic system, a model has been developed considering the appearance of this mineral at $X_{\text{melt}}$ equal to 0.7 (Fig. 2). The modal proportion of hornblende was assumed to increase linearly from 0 to 30% when $X_{\text{melt}}$ passes to 0.4.

A further test was performed to evaluate, in addition to hornblende crystallization in the framework of the extreme case described above, the loss of H$_2$O in an over-saturated system at 0.7 GPa considering the results from thermodynamical models. In detail, the results of the numerical simulation reported in Fig. 2b were used to constrain the maximum mass fraction of H$_2$O potentially subtracted to the system by volatile loss. These values are then introduced in Eq. (10) and reported in Fig. 2c (brown line). The latter must be considered as a very extreme case, as it combines the maximum amount of water potentially subtracted by the development of the hornblende phase at high pressure and high water contents with those related to volatile exsolution at lower pressures (i.e., 0.7 GPa).

**Numerical model for the thermal evolution of the systems.** The numerical model adopted for the study of the thermal evolution of the system in the dynamic case is governed by the conservation equations (mass, momentum, and energy)}
reported below in their dimensionless form:

\[ \nabla \cdot \mathbf{v} = 0. \]  

\[ \frac{\partial \mathbf{v}}{\partial t} + (\mathbf{v} \cdot \nabla) \mathbf{v} = -\nabla p + \nabla \cdot \left( \rho \left( f^i (\nabla \mathbf{v}) + (\nabla \mathbf{v})^T \right) \right) + \text{Gr} \; T \; \mathbf{g}. \]  

where \( \mathbf{v} \) is the non-dimensional dynamic viscosity; the index \( r \) indicates that the considered thermophysical property is taken at the initial reference temperature \( T_r \). The system does not consider multi-phase flows, i.e., separate flows of melt and solid phases are not accounted for. The dimensionless temperature is defined as:

\[ T^* = \frac{T - T_w}{T_{\infty} - T_w \Delta T_r}, \]  

where \( T_r \) and \( T_w \) represent the far-field and the initial temperature, respectively. The volume fraction of the melt phase is given by \( \phi (1 - \varphi) \), where \( \varphi \) is the crystal volume fraction. The parameter \( f \) can be linked to the melt mass fraction \( (X_{\text{melt}}) \) by the following relation:

\[ X_{\text{melt}} = \frac{\rho_0 \mu_{\text{melt}} + \varphi \mu_{\text{cryst}}}{\rho_0 \mu_{\text{melt}} + \varphi \mu_{\text{cryst}}} \times \frac{f - \mu_{\text{melt}}}{\varphi \rho_0 + 1 - f}. \]  

The last term in the heat transport equation expresses the release of heat during crystallization of minerals in the magma, which is inversely proportional to the Stephan number and directly proportional to the variation of the liquid fraction \( f \) with time \( (\partial f/\partial t) \). The Stephan number is:

\[ \text{Ste} = \frac{c_p \Delta T_r}{L_m}. \]  

where \( L_m \) the latent heat of crystallization for the basaltic magmatic mass. The strength of the buoyancy forces can be characterized by the Grashof number or by the Rayleigh number:

\[ \text{Ra} = \text{Gr} \; \text{Pr} = \frac{\rho_0 |\mathbf{g}| \beta_A^* |T| L_m^3}{\alpha \eta}, \]  

which is fluid dependent. In Eq. (19), the expansion coefficient \( \beta \) takes into account both thermal expansion and crystallinity changes within the system and can be expressed as follows:

\[ \beta = \frac{\rho_{\text{cryst}} - \rho_{\text{melt}}}{\rho_r} \frac{c_p}{c_v} + \beta_i, \]  

where \( \beta_i \) is the thermal expansion coefficient.

The effective temperature difference, \( \Delta T \), driving convection for a magmatic system under cooling is linked to the temperature difference between the roof of the magma chamber and the mean temperature of the convecting magma. Thus, the Rayleigh number will decrease rapidly during the cooling of the magma chamber due to the decrease of the temperature difference, as well as to the evolution of the thermophysical properties of the magma (mainly viscosity). Equations (12), (13), and (14) have been solved by direct numerical simulation using the in-house Tamari CFD code. It is an unsteady finite volume parallel solver for anisothermal incompressible fluids flow based on the SIMPLE algorithm and on second-order accurate spatial and temporal schemes. The Herschel-Bulkley model was used to describe the magma rheological behavior:

\[ \eta = K (T, \varphi)^{\alpha - 1} + \frac{\tau_n (T, \varphi)}{f}, \]  

where \( \eta \) is the bulk dynamic viscosity, \( K \) is the consistency, a function of the crystal volume fraction \( \varphi \) and temperature \( T \), whereas the yield stress \( (\tau_n) \) and flow index \( (\alpha) \) are both functions of \( \varphi \). More details about this bulk viscosity model can be found in Mader et al. In the numerical model, this viscosity formulation was regularized using the Papanastasiou’s method. A source-based enthalpy method used to model the liquid–solid phase change (i.e., the crystallization) by ensuring that the relationship between temperature and the liquid fraction \( f \) showed in Supplementary Fig. 3 and discussed in the section entitled “Parameterization and validation of the thermal model” was satisfied. The computational mesh is composed of 54,000 cells. This mesh size was selected after a systematic study of the dependence of the results on mesh size. Code validations are reported and extensively discussed in several previous heat and fluid flow studies.

On the other hand, the static case was modeled by a conductive heat transfer model, where the non-dimensional equation writes:

\[ \frac{\partial T^*}{\partial t^*} = \frac{c_p}{\rho c_0} \left( T^* - \frac{T^*}{T_{\infty}} \right) \]  

In this equation, the non-dimensional time \( t^* = r t \). This equation was solved by a finite volume method in a structured grid coupled with an algorithm of phase change. Furthermore, the relation between \( T^* \) and \( f \) has been solved in a similar to the dynamic case. As in the dynamic case, we considered a two-dimensional computational domain and the same thermophysical properties.

However, to model an open magmatic system re-experiencing an incremental growth by addition of successive sills, the computational domain was periodically extended from below by the addition of new rectangular blocks of computational cells carrying initial magma properties and conditions (Fig. 8g–j). The growth is operated at different rates until the final size of the chamber is reached, which is equivalent to the size of the closed system (Fig. 8a).

Parametrization and validation of the thermal model. In the following we parametrize the thermal model. It allowed us to calibrate the simulation for different values of pressure, initial water content, and chemical composition of the parental magma that are relevant for the study of the fate of arc magmas. In detail, the first step to solve Eqs. (12), (13), and (14) consists in the definition of the dimensionless temperature.

For each experimental set (i.e., a group of experiments characterized by the same conditions of pressure, starting material and initial water content), the initial temperature \( T_r \) was fixed at the liquidus temperature (Supplementary Table 1). We assume that the crystal-bearing magma is surrounded by rocks characterized by a significantly lower viscosity \( \eta_0 \). We have parametrized the temperature \( T_r \) in accordance with Annen et al. The latent heat of fusion of the melt \( (L_m) \) is taken equal to \( 3.5 \times 10^5 \text{ J/kg} \). The temperature difference \( \Delta T_r \) ranges from 105 to 190 °C. Stephan and Prandtl numbers are in the ranges 0.30–0.54 and 350–3650, respectively. Numerical simulations have been performed for representative Stephan and Prandtl numbers of 0.43 and 2577, respectively. The latent heat of crystallization \( (L_m) \) is taken equal to \( 3.5 \times 10^5 \text{ J/kg} \) in accordance with Annen et al. and the viscosity of the melt phase is calculated using the method proposed by Giordano et al. The selected values strictly describe the physical conditions of the experiment reported by Nandekar et al. Moreover, they can be assumed as proxies for the whole range of investigated thermophysical conditions. Indeed, at high Pr numbers, there is a negligible dependence of the flow and heat transfer on Pr. This relatively high value of Prandtl number indicates that at the beginning of the cooling process, the diffusion of momentum by viscous action is predominant over the diffusion of heat inside the magmatic chamber. The value of the Stefan number just below one confirms the relative importance of the latent heat transfer in comparison with the sensible heat transfer. We performed a set of simulations to investigate the effect of the Stefan number variations in the range 0.20–0.54, obtaining timescales estimation within 19% of those obtained at Ste equal to 0.43, as indicated in Supplementary Fig. 4.
To provide a solution for the latent heat term in Eq. (14), we defined an empirical relation between the melt fraction ($f$) and the dimensionless temperature ($T'$). This relation is based on the petrological data set described in the section titled "Petrological data set" (Supplementary Fig. 3). Results from experimental data have been modeled using both a linear relation (reported in purple on Supplementary Fig. 3) and a polynomial curve (reported in yellow on Supplementary Fig. 3). Since the use of the two different models provide similar results (Supplementary Fig. 5), only the results obtained from the linear relationship between $f$ and $T'$ are reported. In practice, the slope of the line reported in Supplementary Fig. 3 provides an estimation of $df/dT'$ at each $T'$ value. Also, the empirical relationship linking the melt fraction ($f$) and the dimensionless temperature ($T'$) reported in Supplementary Fig. 3 allowed us to constrain the rheological behavior of the system for the simulation of the dynamic configuration. In these cases, the rheological behavior of the system is modeled applying the Herschel–Bulkley formulation28. A wide range of thermodynamical conditions are investigated. In detail, we defined three buoyancy-driven convective systems characterized by different Rayleigh numbers ($Ra$) of $10^7$ (dynamic case 1, DC-1), $10^8$ (dynamic case 2, DC-2), and $10^9$ (dynamic case 3, DC-3)87.

The parameter $\beta$ in the Eq. (19) is assumed as a constant and varied in accordance with $Ra$ variations. This choice is in agreement with the linear parametrization reported in Supplementary Fig. 3, and it assumes that relative density variations in the melt phase and the crystal assemblage do not significantly change the value of the parameter $\beta$ during the evolution of the system.

To deal with the complex behavior of the studied system, we consider a non-linear model coupling the flow induced by buoyancy forces, resulting from the thermal expansion of the melt, the heat transfer, the phase change (i.e., formation of crystals), and the non-Newtonian rheological behavior of the fluid. The latter is a function of shear, temperature, and crystal volume fraction $\varphi$, accounted for by a Herschel–Bulkley model27.

Figure 9 displays the temperature field of the system at four dimensionless times $t' = 10$, 25, 50, and 100 corresponding to $t = 3.68 \times 10^{-3}$, $9.7 \times 10^{-3}$, $1.94 \times 10^{-2}$, and $3.88 \times 10^{-2}$). Convective motions generate upside-down mushroom-like thermal plumes. These plumes originate at the roof of the system and progressively descend to the floor of the magma chamber. At low $Ra$ (i.e., $10^5$), the location of consecutive plumes is almost regular; however, at higher $Ra$, the development of new plumes becomes irregular. Figure 9 also shows that the frequency of plume formation decreases with time, leading to the generation of a region of thermal stratification on the floor of the magma chamber.

The observed variations in the temperature fields at different $Ra$ are due to different evolutions of the velocity field in the system. Supplementary Fig. 6 shows the velocity field for the same configurations and times shown in Fig. 9. As reported in Supplementary Fig. 6, the module of the velocity field varies by about an order of magnitude between $Ra = 10^5$ and $Ra = 10^9$ from the early stages of the simulation (e.g., $t' = 10$). The observed increase in the vigor of the convection at growing $Ra$ results in an increase of thermal mixing, leading to a faster decrease of the mean temperature of the system (Supplementary Fig. 6).

The final step to parameterize the evolution of $\Delta H_{\text{fH}_2\text{O}}$ over the non-dimensional time consists in the transformation of the volume fraction $\Delta f$ utilized in the simulations to a mass fraction $X_{\text{melt}}$ (Eq. (16)). The obtained $X_{\text{melt}}$ values are then introduced in Eqs. (10) and (11) to obtain $\Delta H_{\text{fH}_2\text{O}}$. The reported results are developed adopting $R_a$ value equal to 0.86, corresponding, for example, to a melt density ($\rho_{\text{melt}}$) equal to $2750$ kg m$^{-3}$ and an average density of the crystal assemblage ($\rho_{\text{crystal}}$) equal to $3200$ kg m$^{-3}$. The values for the densities of melt and crystals are in agreement to those characterizing the modeled petrological data set30–33. A reduction of the $R_a$ parameter of 0.07 is consistent with a decrease of the melt density ($\rho_{\text{melt}}$) of 250 kg m$^{-3}$. Moreover, a variation of the $R_a$ parameter of 0.07 will result in a change of the estimated timescales, calculated at $\Delta H_{\text{fH}_2\text{O}}$ equal to 2.0, no larger than 15%.

Melt migration velocities. Magma velocities are estimated using two different approaches, based respectively on the buoyancy-driven32,34 and magma-driven32,36 ascending dike theory. With the former approach, the dike is assumed to be isolated and its propagation is due to buoyancy forces only. Indeed, using the elastostatic theory, Weertman36 demonstrates that buoyancy forces are sufficient to allow vertical fractures to propagate upward, as soon as they are longer than a critical length41. In the latter, propagation is driven not only by buoyancy forces, but also by the excess pressure of a reservoir connected to the dike. Therefore, due to the force applied by the excess pressure, even dikes smaller than the critical length can propagate toward the surface. The computation of magma ascent velocities are disjointed from the thermal numerical simulations, allowing us a more comprehensive description of the fate of water-rich melts in subduction zones.

Following Dahm (2000)31, the propagation velocity ($v$) of a buoyancy-driven ascending dike is characterized by a height $2a$ and an average width $2h$ can be estimated using the following equation:

$$v_{\text{Dahm}} = \frac{\Delta P_{\text{visc}}}{3\eta}\frac{1}{f' - (7/3) + (2/3)C} \frac{1}{(1 - \gamma_0)}$$  (23)

where $\eta$ is the dynamic viscosity; $\Delta P_{\text{visc}}$ and $h(z)$ are defined as follows:

$$\Delta P_{\text{visc}} = 2\pi\Delta\gamma\left(1 - f'\frac{K_i}{\sqrt{\pi\Delta\gamma}}\right)^{3/2}$$  (24)

$$h(z) = h(z) + \frac{(1 - \nu)K_i}{2G} \left[\frac{(1 + 2\nu)/\nu}{(1 - (2\nu)/\nu)}\right]$$  (25)

In Eqs. (24) and (25), $\Delta\gamma$ is the density difference between crustal rocks and the rising melt, $\gamma$ is the modulus of the gravitational acceleration, and $K_i$ is the fracture toughness. Furthermore, $G$ is the shear modulus, $\nu$ is the Poisson’s ratio, and $D$ is an experimentally estimated dimensionless thickness, which it is assumed to be in the range 0.01–0.1 (in agreement with Dahm31). The upper and the lower limits of the blue and red areas reported in Fig. 6a, b correspond to a $D$ value equal to 0.05 and 0.1, respectively. The $f'$ constant is a geometrical factor equal to 1 and $\nu/t$ in two and three dimensions, respectively. Finally, the half width of the dike $h$ at the depth $z$ is defined as follows32,34:

$$h(z) = \left[\frac{(1 - \nu)K_i}{2G} \left[\frac{(1 + 2\nu)/\nu}{(1 - (2\nu)/\nu)}\right] \right]^{1/2}$$  (26)

Crust rheological parameters are reported in Supplementary Table 370.

The average flow velocity estimated by Rubin63, also obtained adopting the magma-driven approach, has a similar formulation compared to that reported in Eq. (23):

$$v_{\text{Rubin}} = \frac{\Delta P_{\text{excess}}}{3\eta} \frac{2h}{2a}$$  (27)

where $\Delta P_{\text{excess}}$ is the uniform excess pressure and the average half thickness $h$ is calculated as:

$$h = \frac{\Delta P_{\text{excess}}}{\eta(1 - \nu)}$$  (28)

The estimated ascent velocities reported previously are computed assuming an incompressible magma. However, the presence of volatiles in the magmatic mixture increase the ascent velocity, since, due to volatile exsolution, the density of the mixture decreases with pressure, increasing, as a result, the buoyancy forces. Tsai and Jaupart30 have estimated the increase of the ascent velocity due to the presence of exsolved volatile phases using the following relation:

$$v = v_{c} \rho_{\text{melt}} \left(\frac{\rho_{s} - \rho_{\text{melt}}}{\rho_{s} - \rho_{\text{gas}}}\right)^{1/3}$$  (29)

In the previous equation, $\rho_{\text{melt}}$ is the bubble-free melt density, $\rho_{s}$ is the density of the surrounding rocks, while $\rho$ is the density of the mixture calculated as:

$$\rho = (1 - \alpha_{\text{gas}})\rho_{\text{melt}} + \alpha_{\text{gas}}\rho_{\text{gas}}$$  (30)

where $\alpha_{\text{gas}}$ and $\rho_{\text{gas}}$ are, respectively, the volume fraction and the density of the exsolved gas mixture. In our computations, the density of the bubble-free melt and of the surrounding rocks are assumed constant, while $H_2O$ and CO$_2$ densities vary accordingly to the ideal gas law. The exsolved volatile contents and the relative proportion of exsolved H$_2$O and CO$_2$ are calculated at different pressures using the mixture solubility law proposed by Ghiorso and Gualda38.

Finally, the critical velocity $v_c$, defined as the minimum velocity at which the magma is able to arrive to the Earth surface, is defined as62:

$$v_c = \frac{H}{t_s}$$  (31)

where $H$ (25 and 35 km in our case) is the initial depth of the rising magma and $t_s$ is the solidification time defined as:

$$t_s = \frac{H^2}{4a\sigma^2}$$  (32)

where $\alpha_s$ is the thermal diffusivity of the host rocks and $\sigma$ is given by the transcendental equation and fixed to a value of 0.5462.

Data availability. The raw results of numerical simulations are available at the following repositories: https://goo.gl/GrNPWbo and https://goo.gl/9C9rBZt. Other the supporting information of this study are available as Supplementary Material and from the corresponding authors upon request.

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Author contributions
M.P. conceived and supervised the study. M.P., L.S., and D.P. investigated the thermo-dynamical evolution of the studied systems and developed the H2O–CO2 solubility computations. M.P., K.E.O., and Y.L.G. constrained the initial and boundary conditions for the thermal model. K.E.O. and Y.L.G. performed the numerical simulations for the thermal modeling. M.P., L.S., and G.L.S. managed the estimation of magma ascent rates. All the authors contributed to the data interpretation and participated in the final version of the article.

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