

Protracted Dyking of High-Viscosity Magma into Near-Solidus Rock at Casiri, Peru

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1 Abstract

2 Dykes are a key mechanism for material transport within volcanic arcs and typically propagate over
3 days to months. We document an unusually slow dyking episode in the arc of southern Peru that
4 lasted for 1.5 years. The dyke opened up to 0.5 m from 2-8 km depth and was coupled with inflation
5 of an underlying arc-parallel magma reservoir. To reconcile the slow opening of such a thin dyke with
6 physical models of magma flow and heat transfer, magma with viscosity $\sim 10^6$ Pa s was likely injected
7 into wall rock within $\lesssim 100$ °C of the magma's solidus. Pre-heating of the shallow crust by previous
8 plutonic intrusions may have enabled the viscous magma to have propagated as a dyke to within 2 km
9 of the surface. Our findings suggest that significant pre-heating will be preferential for the transport
10 of evolved, silicic magmas in dykes to depths where magmatic-hydrothermal mineral deposits form.

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12 *This paper is a pre-print submitted to EarthArXiv, therefore has not finished peer review and is cur-*
13 *rently being considered for publication in Geophysical Research Letters.*

14 Plain Language Summary

15 Dykes are thin fractures that transport magmas through the Earth’s brittle crust. Most dykes observed
16 in nature accommodate rapid magma flow at rates of a few kilometres per day, and are inferred to be
17 filled with low-viscosity basaltic magmas. If the flow were slower, the magma would solidify. Here, we
18 use measurements of ground deformation in the volcanic arc of Peru to identify a shallow dyke that
19 propagated at rates of a few kilometres per year — far slower than observed previously. To explain
20 our observations, we propose that the dyke was filled with a viscous magma that propagated into
21 rocks that had been heated extensively by past intrusions of magma. Therefore, pre-heating of the
22 crust may be an important condition for viscous, silica-rich magmas to move through dykes into the
23 upper crust where they can contribute to forming important metal deposits.

24 Key Words

- 25 • We document a 1.5-year-long dyking episode in the Peruvian volcanic arc
- 26 • The slow dyke propagation indicates viscous silicic magma intruding into near-solidus rocks
- 27 • Extensive pre-heating of the upper crust enables silicic magmas to propagate as dykes in arcs

28 1 Introduction

29 Dykes transport magma through the upper crust of volcanic arcs and control the location and timing
30 of eruptions and seismicity over timescales of days to months and metal enrichment over geological
31 timescales. A key limitation with dyking as a mechanism of magma transport into the upper crust is
32 that thin (i.e. <1 m) dykes will cool and solidify within days without large magma fluxes that can
33 advect heat from a reservoir and keep temperatures above the magma solidus [Bruce and Huppert,
34 1989; Lister and Kerr, 1991; Rubin, 1993]. This simple physical limitation has been at the heart of
35 the debate about whether viscous, silicic magmas are transported into the upper crust primarily via
36 dykes or plutons [e.g. Petford et al., 2000; Cruden and Weinberg, 2018].

37 Modern geodetic observations have revealed that most dyking episodes do comprise short-lived, high
38 magma-flux events with propagation and inflation lasting for days to a few months, often with con-
39 temporaneous magma withdrawal from a nearby reservoir that acts as source of overpressure driving
40 magma flow [Segall et al., 2001; Wright et al., 2006; Irwan et al., 2006; Albino et al., 2019; Temtime

et al., 2020; Sigmundsson et al., 2024; Parks et al., 2025; Isken et al., 2025]. However, such rapid magma flow generally requires low viscosity magmas, as flow velocities through a tabular dyke of width w is proportional to $w^2 \Delta P / \eta l$ where $\Delta P / l$ is the magma overpressure gradient along the dyke and η the magma viscosity [Rubin, 1995b]. Silicic magmas common to arc settings that are orders of magnitude more viscous than basalts should therefore require relatively wide dykes and/or high magma overpressures in order to rapidly propagate into cool upper-crustal rocks before freezing [Petford et al., 1994; Rubin, 1995a; Annen et al., 2006]. Alternatively, they may propagate more slowly but need some other mechanism that prevents them from cooling and solidifying.

In this study, we analyse an unusually long-lived dyke intrusion in the volcanic arc of southern Peru between Casiri and Purupuruni (Figure 1a). The intrusion forms part of 20 years of increased seismicity and volcanic activity spread across the Peruvian arc following the 2001 M_w 8.4 Arequipa and 2014 M_w 8.1 Iquique earthquakes (Figure 1b). In Section 2, we present satellite radar measurements of the ground deformation associated with three clusters of seismicity in 2020-2021 near Casiri (Figure 1c). We show that the ground deformation reflects the intrusion of a thin dyke that propagated slowly over ~ 1.5 years, coupled with inflation of an underlying reservoir and shallow faulting (Section 3). In Section 4, we consider the conditions under which such a thin dyke could have propagated and inflated so slowly within the upper crust and explore the implications of these results for the transport of silicic magmas and their volatile and metal cargos in volcanic arcs.

2 Observations of Unrest at Casiri-Purupuruni

The first signs of unrest at Casiri-Purupuruni came from three distinct clusters of shallow (< 30 km hypocentral depth) earthquakes in 2020-2021 near Casiri detected by global the seismic network (S1-S3 in Figure 1c; Supplementary Figure 1). S1 included two M_w 5.0 events on the 25-26th July 2020 that have moment tensors in the Global Centroid Moment Tensor (gCMT) catalogue [Ekström et al., 2012]. After a delay of 8 months, cluster S2 began in March 2021 and lasted 3 months until the end of May 2021 and included a M_w 5.4 earthquake with a northwest-striking oblique normal-faulting mechanism. S2 is swarm-like, as it has no distinct mainshock [Hill, 1977]. The seismic activity ended with three small m_b 3.9-4.6 events in December 2021-January 2022 during S3. Although there is likely ~ 10 km uncertainty in the hypocentral locations of these earthquakes [Weston et al., 2011] making interpretation of their distribution tentative, the events were located along a NW-SE trend between Purupuruni and Casiri (Supplementary Figure 1).

We used Interferometric Synthetic Aperture Radar (InSAR) to measure the ground deformation between 2016 and 2023 covering the period prior to, during, and after the clusters S1-S3. We created 6, 12 and 18-day interferograms using Sentinel-1 SAR data collected along ascending track 149A and descending track 054D with the LiCSAR system [Lazecky et al., 2020]. Individual interferograms were multi-looked to generate ~ 30 m square pixels, filtered using the adaptive power-spectral filter of Goldstein et al. [1988] and unwrapped using the SNAPHU algorithm [Zebker and Lu, 1998]. To correct for the tropospheric contribution to the interferometric phase we applied GACOS corrections to each interferogram [Yu et al., 2018], though these corrections make little difference to the final time-series. Coherence in the region is excellent, therefore we used the short temporal-baseline interferograms to construct a time-series of line-of-sight (LOS) displacement using the New Small Baselines Subset algorithm [Berardino et al., 2002] implemented in LiCSBAS [Morishita et al., 2020]. The time-series were filtered using a 2-D spatial Gaussian-kernel filter of width 2 km and 1-D temporal Gaussian-kernel filter of 1 month and all pixels with a median coherence through time < 0.3 were removed. The full time-series for both tracks are shown in Supplementary Figures 2-3.

Between 2016 and 2020 we found there was no resolvable deformation in either satellite track, which is consistent with previous InSAR surveys of the region covering 1992-2002 [Pritchard and Simons, 2002, 2004]. However, around February 2020 a subtle asymmetrical displacement pattern emerged between Purupuruni and Achocollo equivalent to ~ 10 -15 mm of line-of-sight (LOS) motion towards the satellite visible in both tracks (Figure 2a,d and 2g,h). The onset of this signal appears to pre-date the earthquakes in S1 in July 2020, but this conclusion is sensitive to the temporal smoothing used in the time-series analysis and is not robust. Between July 2020 and August 2020, earthquakes in S1 led to short-wavelength (~ 3 km-wide) LOS displacements of up to +100 mm with a sharp offset across late Quaternary fault scarps mapped from a 1×1 m Digital Elevation Model derived from Pleiades bi-stereo imagery (Figure 2b,e) [Benavente et al., 2010; Vargas et al., 2020]. These fault scarps form a ~ 5 km-wide graben structure extending between Casiri and Purupuruni [Benavente et al., 2010]. Subsequently, from August 2020 to January 2022, a much longer-wavelength (~ 40 km-wide) elliptical LOS displacement pattern developed (Figure 2c,f), with the long-axis of the elliptical signal trending sub-parallel to the volcanic arc and the trend of the Quaternary fault scarps. In the area between the fault scarps, LOS displacements away from the satellite in both tracks indicate localised subsidence due to the M_w 4-5 normal-faulting earthquakes in sequences S1-S2. The LOS displacements stop after January 2022, around the same time as the earthquakes in S3, marking the end of the period of unrest.

The amplitude and wavelength of the LOS displacements at Casiri-Purupuruni between 2020 and 2022

far exceed those expected from the M_w 4-5 earthquakes in S1-S3 (compare Figure 2b,c and Figure 2e,f). The most likely alternative sources of deformation are magmatic inflation or hydrothermal activity that was contemporaneous with the clusters of seismicity. We test these hypotheses and determine the geometry and location of the deformation sources below.

3 Modelling of the Magmatic Deformation

3.1 Magma Architecture

The asymmetry and wavelength of the LOS displacement patterns in Figure 2c,f are indicative of at least one long, thin, inflating magma source in the top ~ 20 km of the crust. The source(s) must generate significant horizontal motions, consistent with dyking [e.g. Wright et al., 2006; Segall, 2010]. However, south of Casiri there are positive LOS motions in both ascending and descending tracks indicative of uplift where a dyke would generate relative subsidence [Rubin and Pollard, 1988] implying multiple magma sources, or hydrothermal poroelastic deformation, may have been active.

To constrain the location and shape of the magma source(s), first we conducted bounded, non-linear Bayesian inversions [Bagnardi and Hooper, 2018] making the assumption that the deformation can be modelled as a combination of inflating ellipsoids [Yang et al., 1988] and/or tensile dislocations [Okada, 1985] in an elastic half-space. Within the inversions, we set broad, uniform priors to explore a range of potential magma source geometries and locations (see Supplementary Table 1 for details). To assess the potential contribution from poro-elastic deformation in the hydrothermal system, we compare the data-model residuals with the distribution of argillic alteration visible at the surface in optical satellite imagery, the distribution of hot springs from Ramírez et al. [2023] and zones of high electrical resistivity mapped in Velarde Benavente et al. [2021].

In the first set of inversions, we masked out the region around the Quaternary fault scarps where we presume displacements are primarily related to fault slip, and attempted to reproduce the longer-wavelength LOS displacement pattern between August 2020 and January 2022 (Figure 2c,f). Inversions including only an inflating ellipsoidal source (Supplementary Figure 4) or only a dyke with constant opening (Supplementary Figure 5) result in systematic misfits with the data suggesting that they cannot account for the observed deformation. Including spatially variable opening along the dyke's length using the approach of Wright et al. [2006] does slightly improve the data fit, but still cannot account for the uplift around Casiri needed to explain the positive LOS motion in both the ascending

and descending tracks (Supplementary Figure 6). However, models with both an inflating ellipsoid and an opening dyke can better account for the observed displacements, particularly the positive LOS motions in both tracks indicative of uplift near Casiri and over the dyke's top edge, and lead to a 50% reduction in the root-mean-square (RMS) misfits compared to the ellipsoid-only or dyke-only models (Supplementary Figure 7). Taking the range of model parameters as the 5th-95th percentiles in the posterior probability density functions (PDFs), the ellipsoidal magma reservoir in these best-fit models is 6-8 km deep and 14-16 km long with a dimensionless excess pressure increase $\Delta P_e/\mu$ of $0.2\text{-}6\times 10^{-4}$ (where μ is the shear modulus). The ellipsoidal reservoir is overlain by a 30-34 km-long dyke that opens by 0.15-0.25 m and which reaches from 8 km deep up to 2 km below the surface, equivalent to a dyke volume of $0.04\text{-}0.07\text{ km}^3$ (posterior PDFs shown in Supplementary Figure 8).

When unmasking the LOS displacements caused by faulting, further improvements to the data fit can be made by including inward-dipping normal faults aligned with the Quaternary fault scarps to the north of Casiri and by allowing distributed opening along the dyke (Figure 3). Modelling the displacement pattern for the largest earthquake during S1 (Figure 2b,e) indicates the normal faults have dips of $55\text{-}60^\circ$, a rake of $65\text{-}70^\circ$ and slip at 0-4 km depth (Supplementary Figure 9; see also Vargas et al. [2020]), implying NE-SW extension across the graben as expected from the earthquake focal mechanisms in the gCMT catalogue [Ekström et al., 2012]. In this, our preferred model, the cumulative fault slip reaches up to ~ 30 cm and releases the equivalent of a M_w 5.2-5.4 earthquake. The dyke volume is $0.025\text{-}0.030\text{ km}^3$, with the range derived by varying the Laplacian smoothing used to regularise the inversion. The dyke volume is smaller than from the non-linear inversions described above because fault slip trades-off with opening on the top ~ 4 km of the dyke. Maximum dyke opening is 0.5 m and $\Delta P_e/\mu$ in the ellipsoidal source is 5×10^{-4} .

Only small (<0.5 cm) residual signals are present, mostly around the surface projection of the faults where the fault discretisation is too coarse to account for subtle near-surface variations in fault slip. Small residuals are also present around Casiri and have a spatial pattern similar to the distribution of argillic alteration at the surface, suggesting they may reflect shallow poroelastic deformation in the hydrothermal system (Supplementary Figure 10). The distribution of cumulative dyke opening and fault slip over the duration of the unrest is shown in Figure 3g. A summary of the different model combinations, their best-fit parameters and model-data fits is provided in Supplementary Table 2.

3.2 Temporal Evolution of Dyking and Reservoir Inflation

Although the amplitude of the displacement pattern observed in the InSAR time-series does change over time, the general pattern does not (Supplementary Figures 2-3), suggesting that the same magma reservoirs have been active over the period of unrest. Using a fixed geometry for the dyke and ellipsoid based on the model in Figure 3, we conducted inversions for the cumulative dyke volume V_d and dimensionless excess pressure $\Delta P_e/\mu$ within the ellipsoidal source over nine different epochs. We find that V_d and $\Delta P_e/\mu$ increased proportionally during the unrest until displacements rates become negligible in January 2022 (Supplementary Figure 11), and there is no robust evidence for propagation of dyke opening along-strike. The inversions do not have sufficient resolution to identify any vertical propagation of dyke opening. For each epoch, there is only a small increase in the RMS misfit between these models and the data and no visible deterioration in the model-data fit, supporting the view that the sources of magma were not changing significantly over time.

4 Physical Implications of Protracted Dyking

Our modelling of the surface deformation and mapping of Quaternary fault scarps all support the interpretation that a dyke (or sub-parallel dykes) injected into the upper 8 km of the crust between Casiri and Purupuruni from June 2020 to January 2022 triggering normal-faulting on a shallow, ~ 5 km-wide graben. Unusually, the dyke at Casiri appears to have been coupled to inflation of an underlying arc-parallel magma reservoir, and that the dyke and magma reservoir were physically linked.

4.1 Dyke-Induced Faulting

The pattern and timing of the shallow normal-faulting earthquakes overlying the dyke are consistent with being triggered by dyke-induced stresses, similar to the patterns observed in other extensional settings in Iceland, Hawaii and the East African Rift [e.g. Rubin and Pollard, 1988; Wright et al., 2006]. Our modelling indicates fault slip was primarily towards the north of the peak dyke opening and between 0-4 km depth, consistent with the prediction that faults ahead of the dyke tip will be moved towards failure and can extend to greater depths than the dyke's upper tip [Rubin and Pollard, 1988]. Notably, geomorphological evidence for normal faulting does not extend east or west of the graben above the dyke's surface trace, suggesting these faults exclusively accommodate dyke-induced strain rather than any component of tectonic extension within the arc. The estimated heave across

the normal faults from our model is 0.1-0.3 m, which is similar to the average dyke opening (0.2-0.3 m) implying that there is a horizontal strain balance between the dyke and shallow faults.

4.2 Dynamics of Coupled Dyke Intrusion and Inflation

A first-order analysis of pressure equilibrium between the ellipsoidal magma reservoir and dyke supports the proposition that they were hydraulically connected and that dyke inflation was driven primarily by the reservoir pressure and not magma buoyancy. The pattern of opening above the magma reservoir shows that the dyke's height h is much smaller than its length l . Under these simplifications, dyke opening w under overpressure ΔP_m is given by $w = \Delta P_m h(1-\nu)/\mu$, where μ is the shear modulus ν the Poisson's ratio [Lister and Kerr, 1991]. Assuming overpressure in the dyke was in equilibrium with the pressure increase in the ellipsoidal reservoir, then $\Delta P_e/\mu = \Delta P_m/\mu$, yielding $w = 1$ m for a dyke of height $h = 6$ km and assuming $\nu = 0.25$. This overestimates our observed dyke opening by at least a factor two, demonstrating that the inferred pressure change in the magma reservoir (~ 3 -10 MPa for $\mu = 5$ -20 GPa) was large enough to account for the dyke's opening and height. As the reservoir continued to inflate whilst magma flowed into the dyke, then some geodetically undetected source at depth within the transcrustal arc magmatic system [Cashman et al., 2017] may have been feeding the reservoir at 8 km to maintain a positive pressure change.

4.3 Thermal Model of Protracted Dyking

Progressive inflation of the magma reservoir provides a source of pressure to drive protracted dyke opening, but it cannot account for its unusually long duration given the dyke's narrow width. Dykes can propagate as long as the heat advection during magma flow and latent heat released by crystallisation outpaces heat loss via conduction to the wall rock [Rubin, 1993]. One possible explanation of the long-lived dyking at Casiri is that the observations reflect a swarm of rapidly-propagating dykes in the same location that were injected too quickly to be observed by the time intervals between SAR acquisitions. However, the possibility of multiple rapidly-propagating dykes seems unlikely, given the smooth temporal evolution of the LOS displacement in both the temporally filtered and unfiltered time-series (Figure 2). In addition, there is no evidence for pulses of dyke-related deformation within individual 12-day or month-long interferograms. Therefore, we interpret the observations to represent a single, prolonged episode of dyke inflation, and we examine the conditions necessary to account for the 1.5-year-long dyke propagation below.

The rate of magma flow in a dyke is dependent on its width, therefore we can define a critical width w_c below which a dyke would not propagate as slow magma flow would cause solidification. Bruce and Huppert [1989] and Petford et al. [1994] derived a relationship for the critical width by comparing the balance between conductive cooling and advective meltback at the dyke tip:

$$w_c = 1.5 \left(\frac{S_m}{S_\infty^2} \right)^{3/4} \left(\frac{\eta \kappa h}{\Delta P/h} \right)^{1/4}$$

$$S_m = \frac{L}{c(T_m - T_s)}$$

$$S_\infty = \frac{L}{c(T_s - T_w)} \quad (1)$$

where T_m , T_s and T_w are the melt, solidus and wall-rock temperatures, c is the specific heat capacity of the magma, L is the latent heat of crystallisation, κ is the wall-rock diffusivity, and h is the magma column height. Here we have replaced $\Delta \rho g$ in Equation 2 of Petford et al. [1994] with a linear approximation of the overpressure gradient along the dyke $\Delta P/h$ to account for source-pressure-driven flow at Casiri. Equation 1 shows that the critical width w_c is most strongly dependent on the excess melt temperature above the solidus $T_m - T_s$ and the difference between the solidus and wall-rock temperature $T_s - T_w$. Magma viscosity will also be an important control on w_c , as it can vary over ~ 14 orders of magnitude [Giordano et al., 2008]. For the critical dyke width in Equation 1, the average magma flow velocity $\bar{v}$ within a tabular dyke would be [Lister and Kerr, 1991]:

$$\bar{v} = \frac{w_c^2}{12\eta} \frac{\Delta P}{h}. \quad (2)$$

Our observations suggest that the Casiri dyke opened with no detectable change in length. Therefore, we assume opening was caused by vertical magma flow from the underlying reservoir, such that the rate of volume change can be related to the rate of magma flow assuming mass conservation: $dV_d/dt = w\bar{v}l$. From our measurements of V_d through time in Section 3, $\bar{v} \sim 2\text{-}4$ km/yr. A similar magma flow velocity comes from assuming that the dyke propagates its 4-6 km vertical height over the 1.5-year-long unrest.

For a magma flow velocity of 2-4 km/yr, and assuming an overpressure gradient of 3-10 MPa over 4-6 km and a 0.3-0.5 m wide dyke, the effective viscosity of the magma would have been $\sim 10^6$ Pa s. These viscosities are consistent with more silicic magmas [Giordano et al., 2008] with compositions similar to the Holocene-age dacitic and andesitic lava domes at Purupuruni [Bromley et al., 2019; Mamani et al., 2010]. Although high-viscosity dyking episodes have been observed before in rifts [e.g. Temtime et al., 2020], they are surprisingly rare in arc settings. The majority of previous estimates of the magma

flow velocity from well-studied dykes using the time-progression of seismicity or the propagation of opening are much faster — on the order of km/day and with inferred magma viscosities of $\sim 10^0$ - 10^3 Pa s which is consistent with mafic magmas [e.g. Segall et al., 2001; Sigmundsson et al., 2015].

Figure 4a shows calculations for w_c and $\bar{v}$ for representative values of the intrinsic thermal properties and a $T_m - T_s = 300$ °C. We find that, in order for a dyke with width <0.5 m and height 6 km to accommodate magma flow velocities of 2-4 km/yr, the temperature difference between the magma solidus and wall rocks $T_s - T_w$ was likely $\lesssim 100$ °C (Figure 4a). Varying the assumed intrinsic thermal properties c , L , κ within reasonable values does not significantly change this conclusion by more than a few tens of degrees. Assuming a melt solidus temperature $T_s = 700$ °C for silicic magma then the wall rocks will be $T_w \gtrsim 600$ °C, which is 350-500 °C hotter than typical steady-state geotherms over the depth range in which the dyke was injected. Equivalent curves for magma flow velocities of 1-10 km/day, which are more representative of the well-studied mafic dykes in Iceland, indicates a much larger range of allowable $T_s - T_w$ more consistent with the injection of metre-wide basaltic dykes with $T_s \sim 1000$ °C into the top 2-5 km of the crust of the Icelandic rift zones ($T_w \sim 200$ -500 °C).

5 Discussion and Conclusions

Our thermal modelling indicates that the upper ~ 8 km of crust between Casiri and Purupuruni is anomalously hot compared to a steady-state crustal geotherm dominated by vertical conduction. This is perhaps unsurprising, given that there are active hydrothermal systems [Ramírez et al., 2023] and Holocene lava flows and lava domes [Bromley et al., 2019] at both localities indicating shallow, hot magmatic systems. However, it raises the question: what mechanisms enable the shallow crust to be so hot? Two simple explanations are that the dyke injected into wall rocks that have been heated by previous dyke intrusions, or that it was injected into now solidified, but young, plutonic intrusion(s) that are yet to thermally equilibrate with its surroundings.

The fault-related geomorphology at Casiri-Purupuruni suggests there has been repeated dyke injections into the shallow crust, and can place loose constraints on the magma flux rates and therefore rates of heat advection through dyking. We found that the 0.2-0.5 m of opening in the 2020-2022 dyke led to 0.3 m of slip on faults that have late Quaternary fault scarps with 10-40 m of cumulative offset. Assuming that the order-of-magnitude relationship between the dyke opening and fault slip holds over repeated intrusions [e.g. Magee and Jackson, 2021], and that the scarps most likely formed since the last major glacial period in the western Andes ~ 10 -15 ka [Clapperton, 1983], yields an

order-of-magnitude constraint on the one-dimensional magma flux rate Q_d of $\sim 3 \times 10^{-3}$ m/yr. These magma flux rates can be related to temperature change in the wall-rock through a one-dimensional balance between the heat input from dyking and heat loss to wall rock via horizontal conduction [e.g. Michaut and Jaupart, 2006]:

$$\Delta T \sim \frac{T_m - T_w}{Q_d} \left(\frac{\kappa}{\tau} \right)^{1/2}, \quad (3)$$

where ΔT is the temperature change of the thermal boundary layer of thickness $\sqrt{\kappa\tau}$ following injection of magma with temperature T_m into wall rock with temperature T_w over total duration τ . Taking representative parameters of $\kappa = 10^{-5}$ - 10^{-6} m²/s and $\tau = 10$ -15 kyrs and a $Q_d \sim 10^{-3}$ m/yr would lead to negligible heating of the wall rock ($\Delta T \ll 100$ °C). This simple analysis is consistent with full two-dimensional solutions to the heat diffusion equation in Daniels et al. [2014], which show that for similar magma flux rates the top ~ 8 km of the crust would require 1-10 million years of dyking to heat the wall rocks by the required 350-600 °C. There is no geological or geomorphological evidence that such prolonged and volumetrically extensive shallow dyking has occurred at Casiri-Purupuruni.

The alternative explanation is that the dyke intruded into a pre-existing, yet solidified, plutonic intrusion that forms the top of a trans-crustal magma system [e.g. Cashman et al., 2017]. Wider plutonic intrusions carry more heat than dykes, cool slower and can therefore retain near-solidus temperatures for longer. Shallow, arc-parallel plutons are also common in the geological record of exhumed arcs. For example, within the well-explored Paleogene Cu-porphyry systems common throughout southern Peru (Figure 1a), circular stocks and plutons crystallised at paleodepths of ~ 2 -5 km and are often cross-cut by felsic dyke swarms [Tosdal and Richards, 2001; Sillitoe, 2010; Loucks, 2021; Farrar et al., 2023]. In addition, the wall rocks in porphyry systems are extensively brecciated and contain hydrothermal mineralisation, indicating that pressurised fluids were being released by the magma driving rock fracture [Sillitoe, 2010]. These features of porphyry systems closely mimic the magma architecture and patterns of seismicity during the recent unrest at Casiri (Figure 4b), with the earthquake swarm S2 potentially being a modern analogue for the hydrothermal brecciation process [e.g. Hill, 1977].

Our analyses lead to implications regarding the conditions that enable the transport viscous, silicic magmas to upper-crustal levels where magmatic-hydrothermal mineral deposits form (e.g. Cu-porphyries). We have shown that viscous magmas can be transported via dykes to depths of ~ 2 -8 km within arcs, but this needs significant pre-heating of the upper crust to prevent the dykes from solidifying. Outside of areas where there is significant tectonic extension within the arc, this pre-heating is most likely to be achieved through earlier plutonic intrusions and long-lived intrusive magmatism.

302 6 Data Availability Statement

303 All the data and software needed to reproduce the results of this study are freely available. The full
304 InSAR time-series for tracks 054D and 149A, as well as the software used to reduce the data, perform
305 the kinematic modelling and the thermal modelling, are available from: [10.5281/zenodo.23696091](https://doi.org/10.5281/zenodo.23696091).
306 The latest version of the non-linear inversion software GBIS is available from: [https://comet.nerc.](https://comet.nerc.ac.uk/geodetic-bayesian-inversion-software-gbis/)
307 [ac.uk/geodetic-bayesian-inversion-software-gbis/](https://comet.nerc.ac.uk/geodetic-bayesian-inversion-software-gbis/).

308 7 Acknowledgements

309 SW, TC, SE, ML and CNL were supported by COMET the NERC Centre for the Observation and
310 Modelling of Earthquakes, Volcanoes and Tectonics, a partnership between UK Universities and the
311 British Geological Survey. PYH was supported by a COMET Summer Internship grant. TC was
312 supported by the Royal Society under URF\R1\180088, URF\R\231019 and RF\ERE\210041. CB
313 was supported by the INGEMMET Neotectonics programme. SW thanks Lin Way for providing the
314 code for performing linear inversions including a dyke and ellipsoidal source.

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Figures

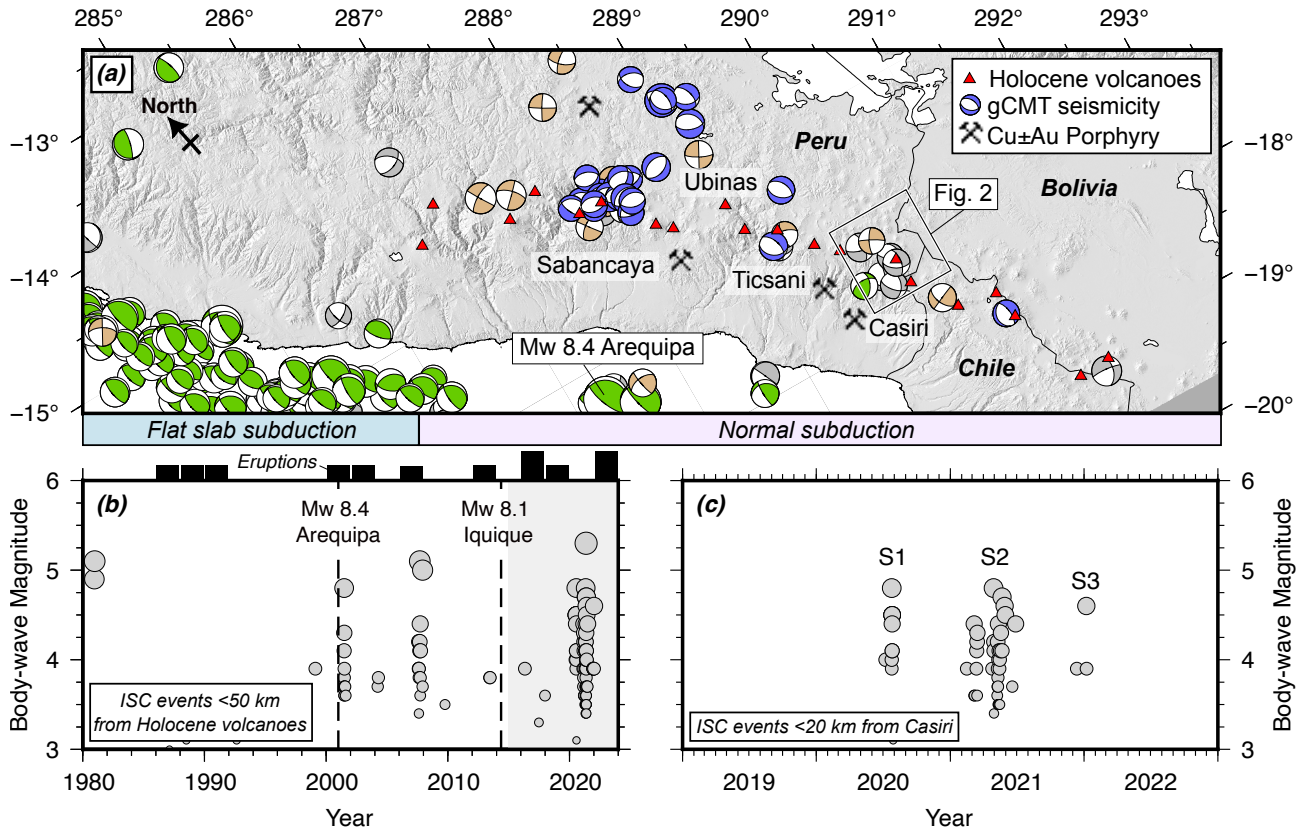


Figure 1: Overview of shallow seismicity and volcanism in southern Peru. (a) Moderate-magnitude ($M_w > 5$) earthquakes in the shallow crust reported by the gCMT catalogue [Ekström et al., 2012] are concentrated along the Quaternary volcanic arc. Locations of four of the largest Cu±Au porphyry mines in Peru (Toquepala, Cerro Verde, Quellevco, Antapaccay) are shown. (b) Temporal evolution of shallow-crustal (<30 km deep) seismicity within 50 km of the arc volcanoes in southern Peru from the ISC reviewed catalogue [Storchak et al., 2020]. The grey box marks the time period covered by the Sentinel-1 InSAR time-series in Figure 2. The black histogram shows the timing of the onset of eruptive activity at Sabancaya and Ubinas from the Global Volcanism Program [2025]. (c) Temporal evolution of seismicity around Casiri-Purupuruni volcanoes illustrating three distinct clusters (S1-S3) between June 2020 and January 2022.

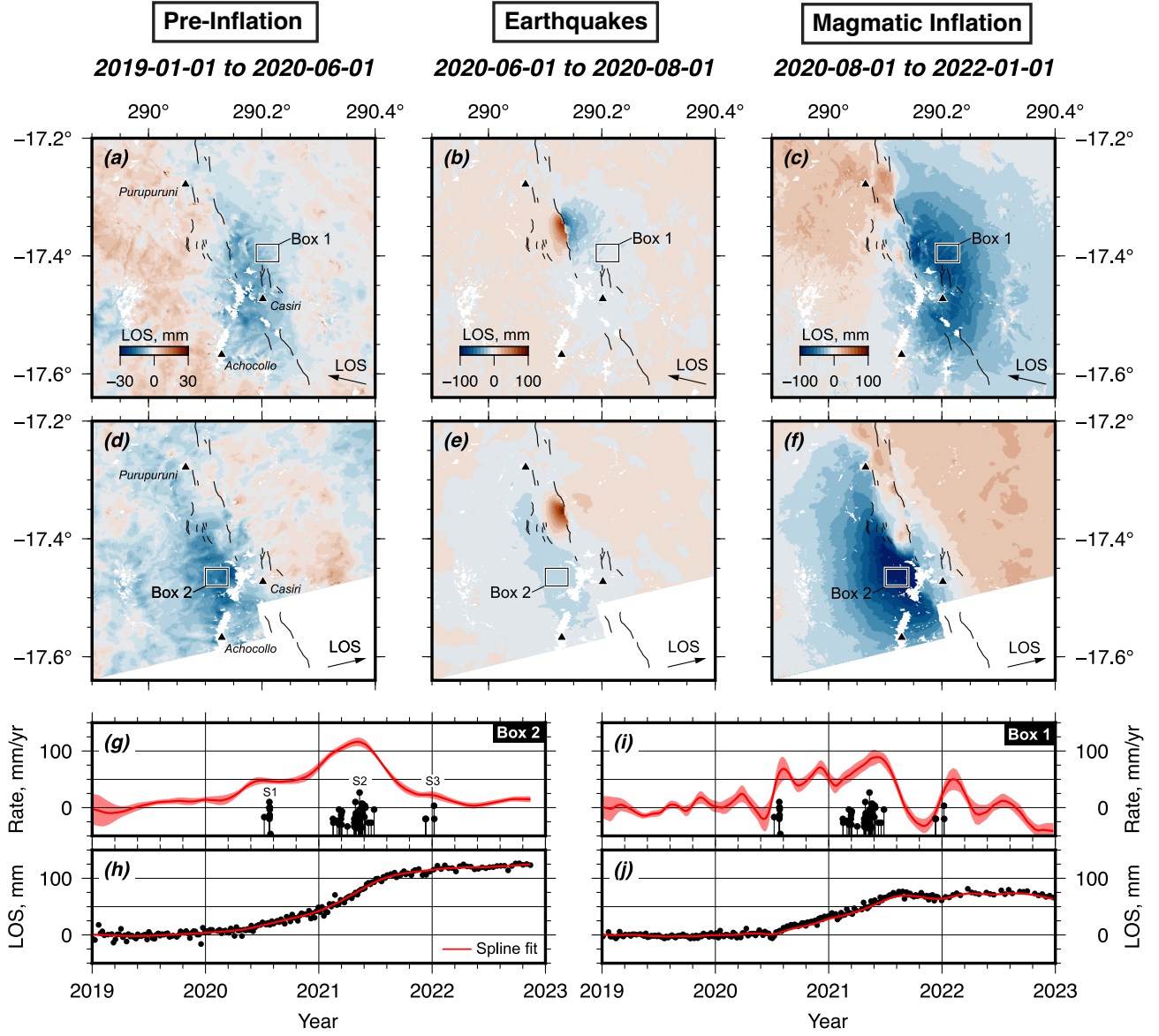


Figure 2: InSAR line-of-sight (LOS) displacement time-series at Casiri-Purupuruni. (a-c) Incremental LOS displacements measured for descending track 054D covering the period prior to the onset of the earthquake clusters (a), during cluster S1 (b), and between mid-2020 and 2022 during clusters S2 and S3 (c). (d-f) Equivalent plot for the ascending track 149A. Thin black lines in (a-f) delineate fault scarps mapped from optical satellite imagery and a 1 m Digital Elevation Model. (g-h) LOS displacement and displacement rate for pixels within Box 2 on the ascending track. The black stem diagram in (g) shows the timing of earthquake clusters S1-S3. Displacement rates are calculated from the gradient of a spline fit to the displacement time-series, with 2σ uncertainties determined using bootstrapping. (i-j) shows the equivalent of (g-h) but for Box 1 on the descending track.

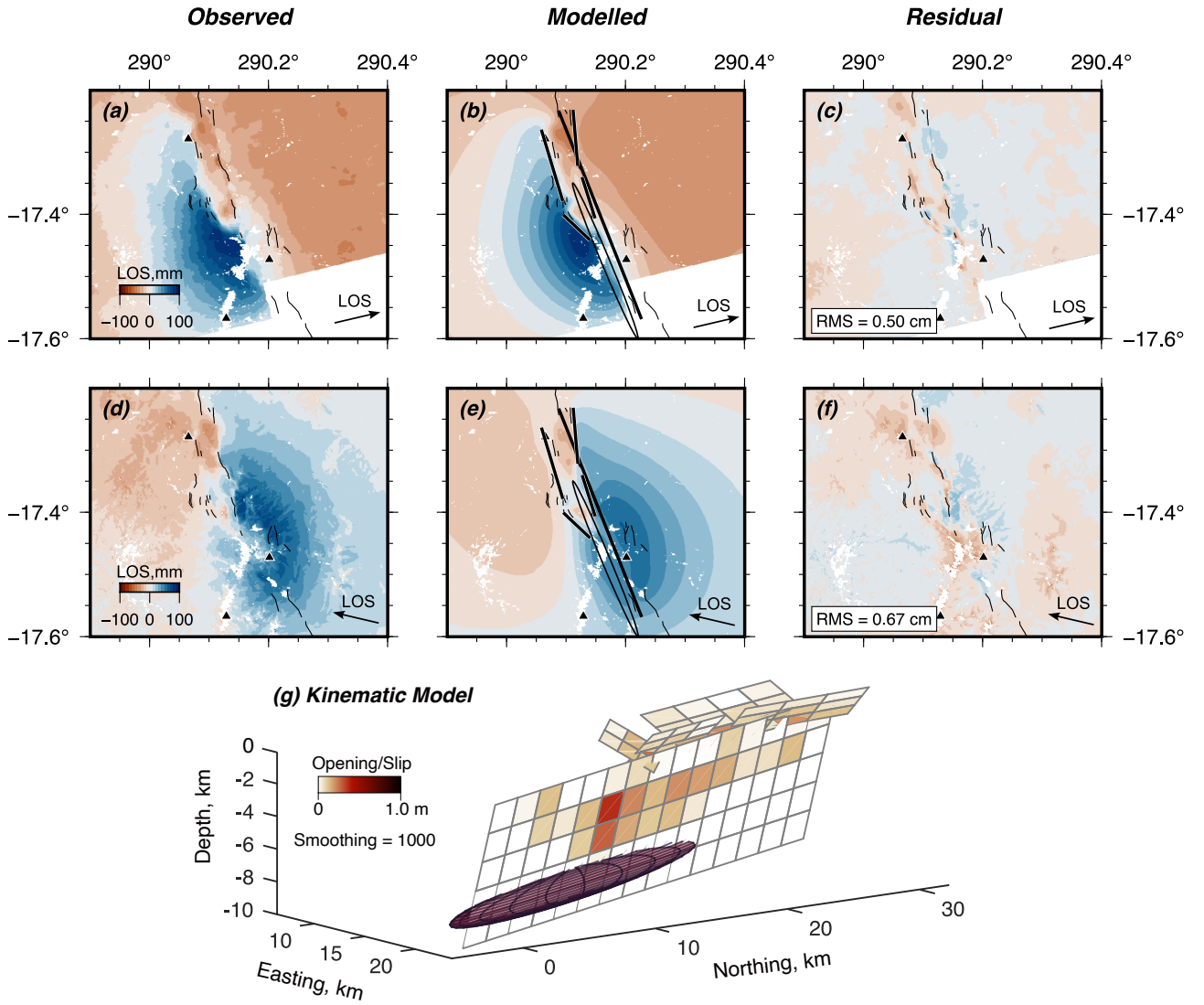


Figure 3: Kinematic dyke, fault and ellipsoid model for the main period of deformation between 14th August 2020 to 1st January 2022. (a-c) Observed, modelled and residual line-of-sight (LOS) displacements in the ascending track. (d-f) Observed, modelled and residual LOS displacements in the descending track. Thick back lines in (b) and (e) denote the surface projection of faults and dyke used in the modelling and the ellipsoid denotes the map projection of the Yang-type ellipsoidal source. (g) Three-dimensional perspective view of the dyke and fault geometry, including the distribution of dyke opening and fault slip. Colours correspond to opening on the near-vertical dyke and total slip on the shallow, dipping faults.

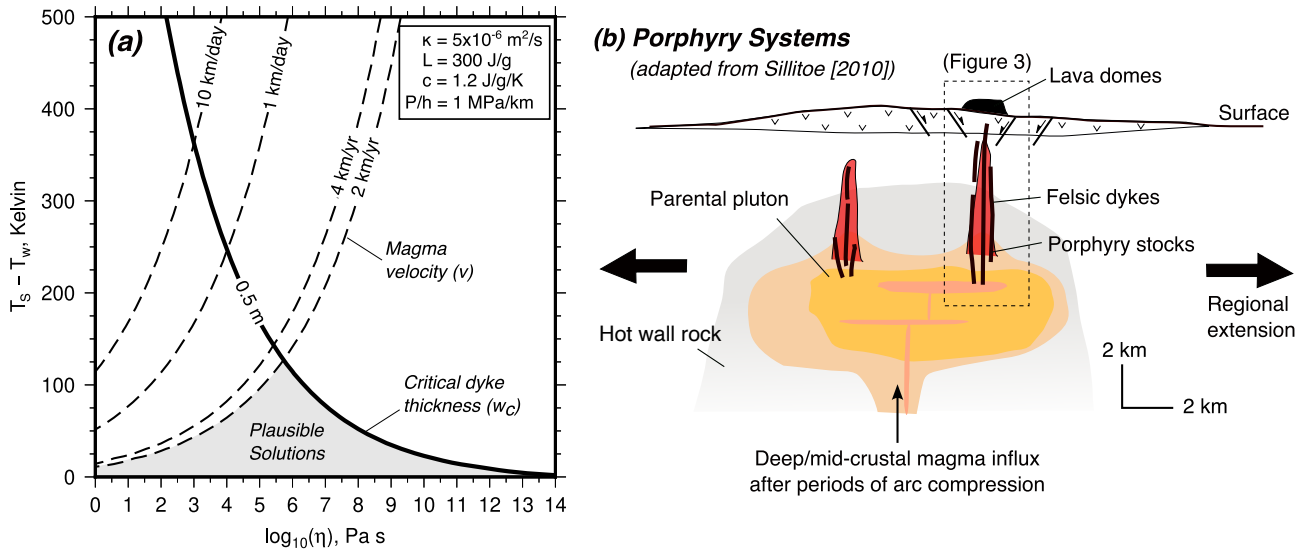


Figure 4: Thermal modelling of protracted dyking at Casiri. (a) Calculations for the critical dyke width w_c needed for dyke propagation and the resulting magma flow velocity $\bar{v}$ as a function of magma viscosity η and the temperature difference between the magma solidus and the wall rock $T_s - T_w$, following Petford et al. [1994]. For our observed w of 0.5 m, and inferred magma flow velocities of 2-4 km/yr, the magma solidus was likely $\lesssim 100^\circ\text{C}$ than the wall rock temperature. Curves for higher magma flow velocities of 1-10 km/yr that have been measured for mafic dykes are also shown. (b) Sketch of the spatial association between dykes, stocks and the underlying parental magma body in exhumed Cu-porphyry systems, adapted from Sillitoe [2010]. The magma architecture and tectonic setting in this generic model of porphyry systems is similar to the dyke and magma reservoir structure observed at Casiri.