



This is a non-peer-reviewed preprint submitted to EarthArXiv.

Krasheninnikov eruption facilitated by M8.8 Kamchatka earthquake and hidden critical state

James Hickey¹, Rami Alshembari¹, Lorenzo Mantiloni¹, Gilles Seropian¹, Yasser Maghsoudi¹, Gino Gonzalez^{2,3,4}, Brendan McCormick Kilbride⁵, Ben Esse⁵, and Mike Burton⁵

¹ Department of Earth and Environmental Sciences, University of Exeter, UK

² Istituto Nazionale di Geofisica e Vulcanologia, Osservatorio Vesuviano, Napoli, Italy

³ Volcanes sin Fronteras, Costa Rica

⁴ Laboratorio Nacional de Nanotecnología (LANOTEC), Centro Nacional de Alta Tecnología (CeNAT-CONARE), Costa Rica

⁵ COMET, Department of Earth and Environmental Sciences, University of Manchester, UK

Corresponding author: James Hickey, j.hickey@exeter.ac.uk

Abstract

Large earthquakes can trigger volcanic eruptions, but the required conditions and mechanisms remain uncertain. Here, we investigate whether the August 2025 eruption of Krashennnikov volcano, Russia, was facilitated by the July 29th, 2025, M_w 8.8 Kamchatka megathrust earthquake. InSAR observations from 2016 to 2025 show no pre-earthquake inflation and satellite SO_2 data indicate no detectable pre-eruptive degassing, suggesting apparent pre-earthquake, and pre-eruption, volcanic quiescence. Following the earthquake, InSAR data reveal a dyke intrusion beginning August 1st, shortly before eruption onset on August 2nd. Deformation modelling indicates a $\sim 3.2 \times 10^7$ m³ dyke fed by a ~ 6 -km-deep reservoir containing compressible, gas-rich magma. Stress modelling shows that static seismic stress and strain changes were small, mechanically mixed, and unlikely to have directly caused the dyke intrusion and eruption. Dynamic stresses were larger, reaching $\sim 1.0 \pm 0.6$ MPa, but still probably insufficient to trigger direct magma reservoir failure. Instead, the prolonged earthquake shaking likely promoted time-dependent volatile exsolution, bubble growth, and eventual reservoir destabilisation in a system already mechanically susceptible to failure. We show that apparently quiescent volcanoes may conceal hidden critical states prone to seismic perturbation, particularly in extensional arc settings with volatile-rich magmas exposed to repeated megathrust earthquakes.

Large-magnitude earthquakes can perturb magmatic and hydrothermal systems and trigger volcanic eruptions^{1–6}. Earthquake-induced stress field perturbations underpin most inferred triggering mechanisms^{7,8}, with permanent static stress and strain changes acting at near-field (10's of kilometres) distances and transient dynamic stresses more dominant in the far-field (100's of kilometres)^{1,5}. However, global analyses and individual studies suggest that eruption triggering is rare and conditional, not deterministic^{1,9–11}, with seismic perturbations acting as accelerants at volcanoes already primed for eruption and in a so-called “critical state”^{6,12}. Magnitudes of inferred seismic stress changes are typically less than 1 MPa, reaching up to 3 MPa in only the most extreme cases⁵. Typical estimated magma reservoir failure pressures of ~ 1 -10 MPa^{13–15} suggest seismic stress changes are insufficient to directly trigger an eruption without a magmatic system already being near failure, i.e., the “critical state”. Although widely applied, the concept of a “critical state” remains loosely defined and varies across studies^{1,2,5,6,16,17}. The influence of seismic stress changes at volcanoes is related to magma viscosity, local stress fields, magma plumbing system geometry, edifice geometry, the presence or absence of a hydrothermal system, and whether the volcano is open to degassing¹, though these aspects can be hard to assess. A range of eruption-triggering mechanisms have been proposed^{1,6}, including bubble nucleation and exsolution⁵, sloshing¹⁸, magma mobilization¹⁹, edifice resonance²⁰, permeability changes²¹, and hydrothermal perturbations⁵. In subduction zones, seismically induced eruption triggering may be favoured by the recurrence of large magnitude earthquakes²² and the presence of volatile-rich, hydrous magmas²³ stored in crystal-mush reservoirs²⁴ that are sensitive to pressure oscillations promoting volatile exsolution, bubble growth, and magma

mobilisation⁵. Here, we investigate the possible triggering of the 2025 Krasheninnikov volcano eruption by the M_w 8.8 Kamchatka earthquake.

The Kamchatka Peninsula, Russia, lies above the Kurile–Kamchatka subduction zone, where the Pacific Plate subducts beneath the Okhotsk (North American) Plate (Figure 1)²⁵ generating large megathrust earthquakes and substantial arc volcanism²⁶. On July 29th at 23:24 UTC (July 30th 11:24 local time), 2025, a M_w 8.8 reverse-fault earthquake occurred ~100 km south-east of Petropavlovsk-Kamchatsky at a depth of ~35 km^{27–29}. Krasheninnikov volcano is located ~230 km north of the earthquake epicentre and composed of two overlapping Holocene stratovolcanoes on a NE-SW trending fissure^{30,31} (Figure 1), within a caldera dated at ~30 ka³². Eruptive products of the past 12,000 years range from basalt to dacite³⁰. Before 2025, the most recent eruption was thought to have occurred in 1550 within the northern crater of the edifice^{30,31}. The 2025 eruption began on August 2nd at 16:50 UTC (August 3rd 04:50 local time)³¹ and is continuing as of August 2026. It has been suggested that the 2025 Krasheninnikov eruption was influenced by the M_w 8.8 earthquake²⁸, but a detailed investigation is lacking. We test this hypothesis by investigating: activity at Krasheninnikov since 2016; stress transfer during the earthquake; and any evidence for a “critical state” that could have primed the volcano for triggering by the earthquake.

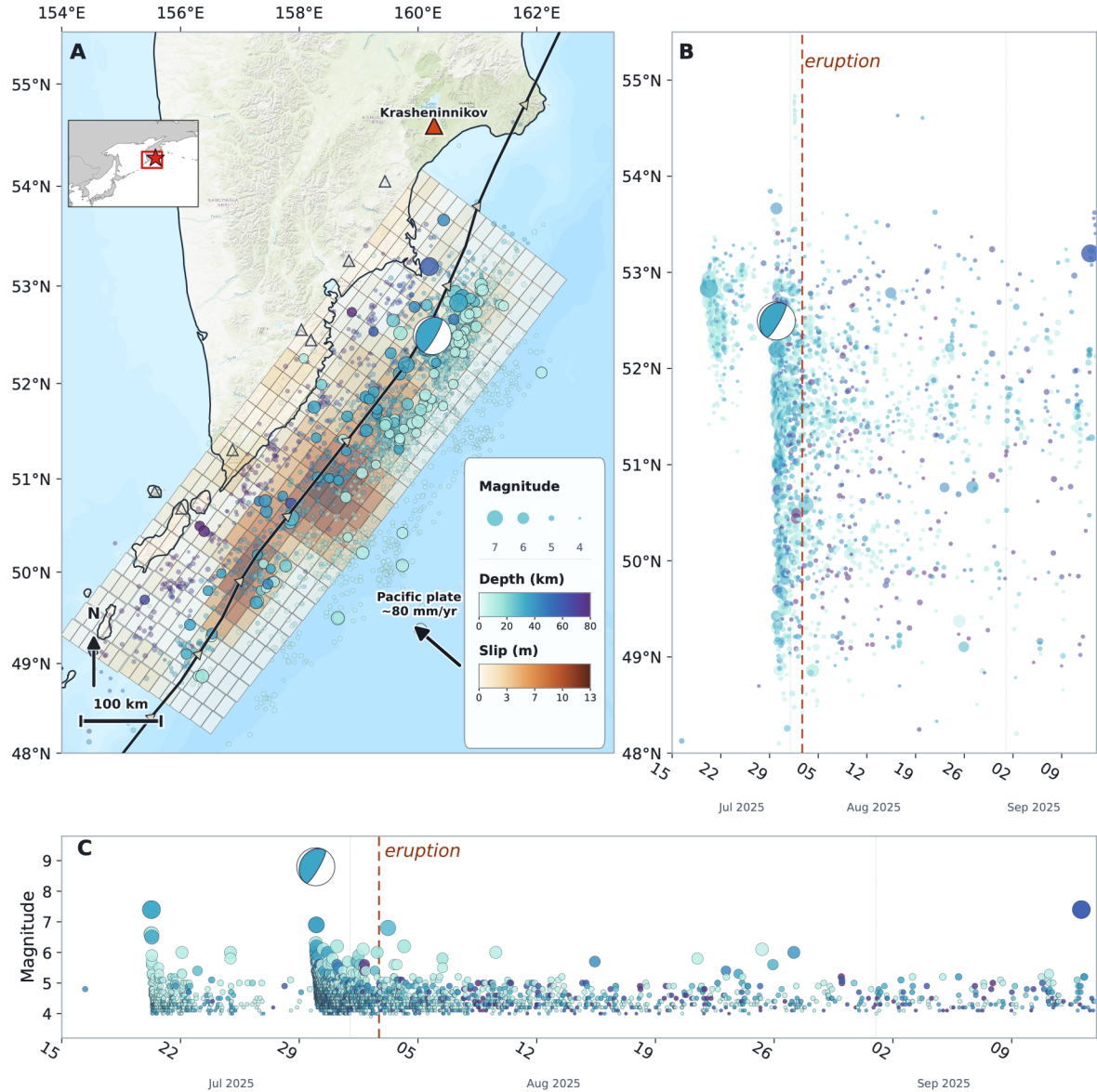


Figure 1: Kamchatka earthquake and Krasheninnikov volcano geographic context and earthquake sequence. (a) Map of southern Kamchatka showing location of Krasheninnikov (red triangle) and a fault slip model solution for the Mw 8.8 earthquake²⁸. Earthquake epicentres are shown, coloured by depth and sized by magnitude. A focal solutions is shown for the Mw 8.8 earthquake²⁸. Inset shows wider regional context. (b) Temporal and geographic pattern of earthquake sequence. (c) Earthquake magnitude shown against time. Seismicity data are from the USGS²⁷.

Apparent Pre-Earthquake Volcanic Quiescence

Interferometric Synthetic Aperture Radar (InSAR) analysis of ground deformation from June 2016 to July 2025 shows that prior to the Mw 8.8 earthquake Krasheninnikov was not displaying any signs of inflation (Figure 2). Surface inflation at volcanoes commonly reflects magma migration or accumulation, which could have suggested the volcano was in a state of pre-eruptive unrest^{33,34}. In

contrast, observations show a slow subsidence of ~ 4 mm/yr over the 9 years preceding the earthquake (Supplementary Figure 1). Ground subsidence at volcanoes can result from magma reservoirs cooling and contracting³⁵, but the lack of accompanying horizontal ground motion towards the centre of vertical subsidence makes this mechanism unlikely. Instead, horizontal ground motion away from the central subsidence, and the locations of peak ground movement on the edifice flanks, suggests the volcano was experiencing continuous, shallow-seated, small-scale flank sliding, perhaps due to permafrost melting³⁶ rather than a deeper-seated magmatic process.

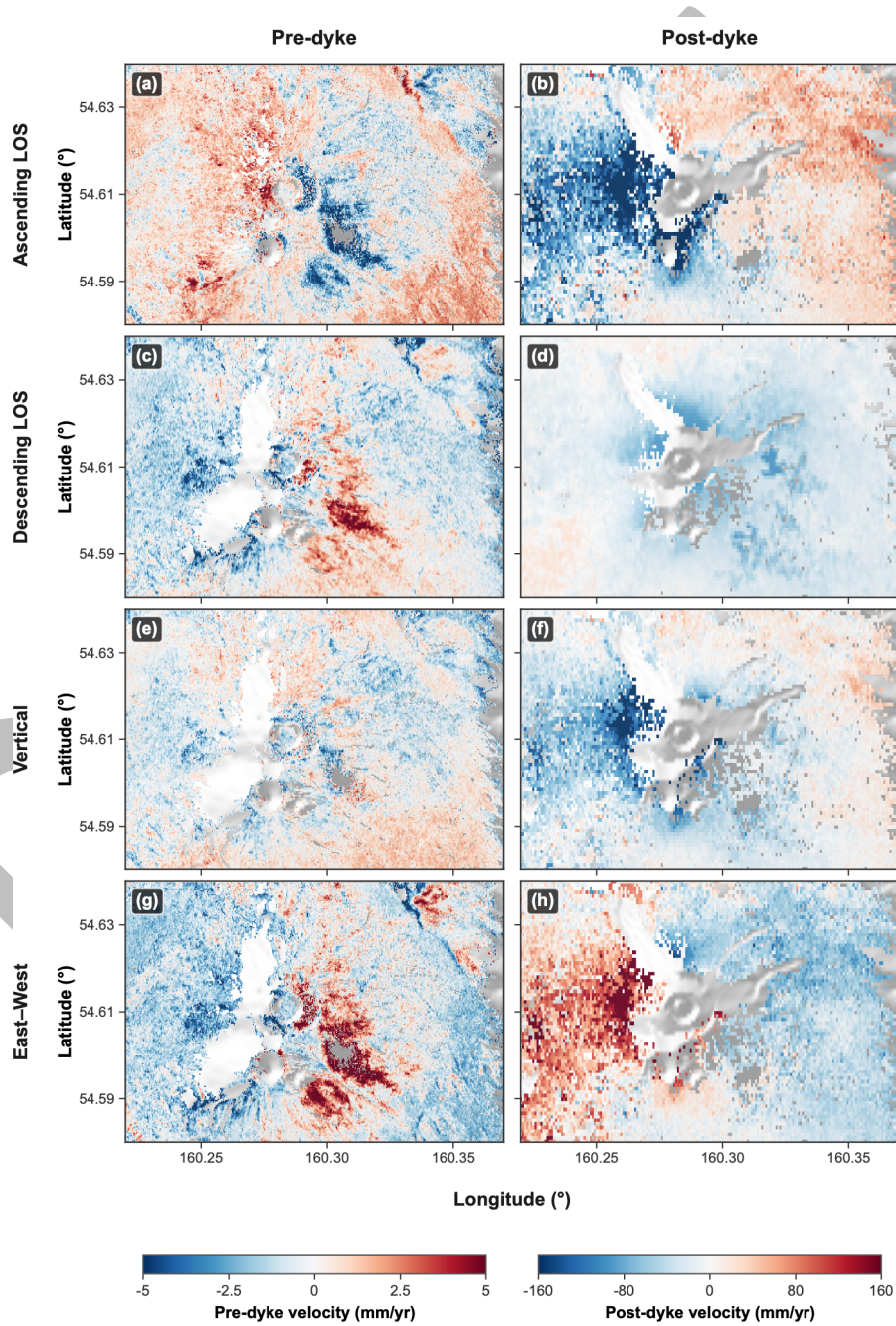


Figure 2: Pre- and post-dyke InSAR observations. Left column shows the pre-dyke surface displacement velocities between February 2016 and July 2025, for: (a) ascending satellite line-of-sight (LOS); (c) descending satellite LOS; (e) near-vertical decomposition; and (g) near-east-west decomposition. Right column shows the

post-dyke surface displacement velocities between August 2025 and March 2026, for: (b) ascending satellite LOS; (d) descending satellite LOS; (f) near-vertical decomposition; and (h) near-east-west decomposition. Note the difference in magnitude between the two time periods.

Satellite-derived observations from the TROPOspheric Monitoring Instrument (TROPOMI) give no indication of gas emissions at Krasheninnikov prior to the M_w 8.8 earthquake (Figure 3 and Supplementary Figure 2). Sulphur dioxide (SO_2) emissions, indicative of shallow-stored magma and an open conduit system^{37,38}, are evident in TROPOMI data from multiple volcanoes on the peninsula (Figure 3), but no comparable signal is present in the atmosphere over Krasheninnikov. From the data available, and the lack of precursory activity reported by the locally responsible monitoring agency KVERT (Kamchatka Volcanic Eruption Response Team), there are no indications that the magmatic system at Krasheninnikov was experiencing measurable unrest prior to the July 2025 M_w 8.8 earthquake.

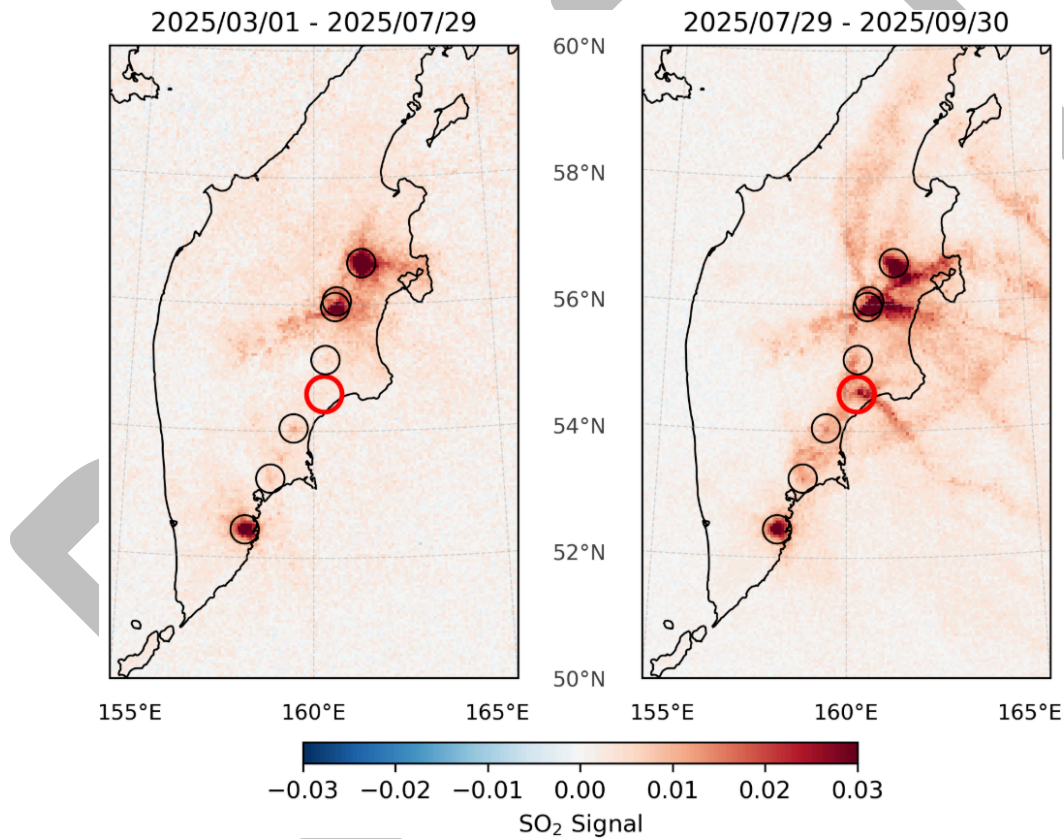


Figure 3: Mean atmospheric SO_2 concentrations over Kamchatka, based on regridded Level 2 SO_2 total column product. The assumed plume centre of mass altitude is 7 km. Left hand panel shows the pre-earthquake interval, 1st March to 29th July; right hand panel shows the post-earthquake interval, 30th July to 30th September. Circles indicate the location of selected volcanoes, from north to south Shiveluch, Klyuchevskoy, Bezymianny, Kizimen, Krasheninnikov (in red), Karymsky, Avachinsky, and Mutnovsky.

Dyke Intrusion, Eruption Initiation, and Evolving Surface Deformation

After the July 29th M_w 8.8 earthquake we detect an abrupt shift in the observed surface deformation. Interferograms derived between July 30th and August 5th (ascending satellite track) and July 31st and August 6th (descending satellite track), 2025, show characteristic “butterfly” ground deformation patterns, indicative of a dyke intrusion (Figure 4). Interferograms derived from data acquired before July 30th do not show this pattern (Supplementary Figure 3), constraining dyke onset to no earlier than July 31st, and confirming the dyke was emplaced after the M_w 8.8 earthquake. Despite the lack of local ground-based monitoring, regional seismic networks detected twelve $M_w \sim 4$ earthquakes around Krashennnikov on August 1st between 12:37 and 17:40 UTC³⁹. Their depths are unconstrained but similar magnitude seismicity commonly accompanies dyke intrusions at other volcanoes^{40,41}. Therefore, we interpret these earthquakes as associated with dyke intrusion, and suggest the intrusion began on August 1st, 2025.

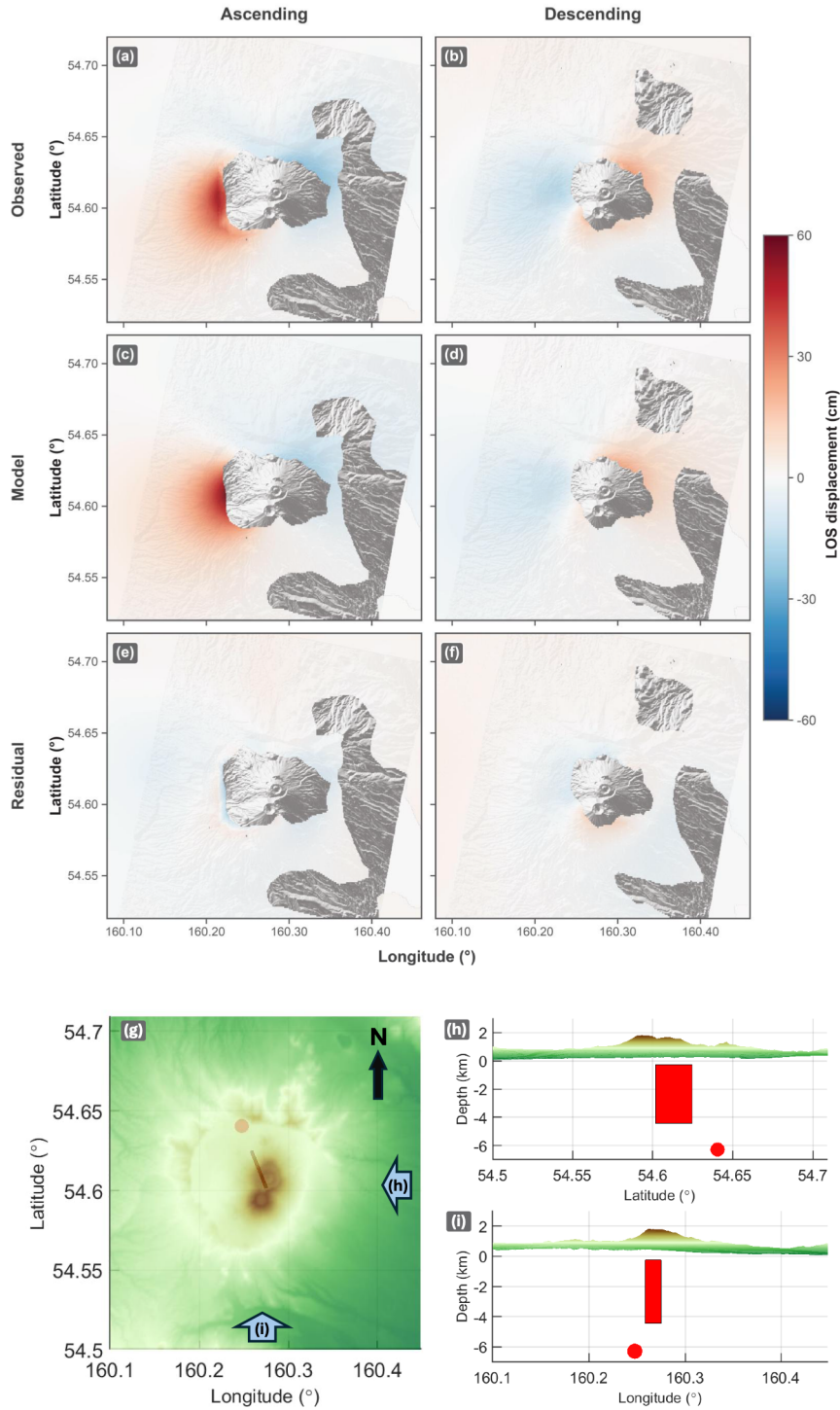


Figure 4: Dyke intrusion InSAR observations and modelling. Measured displacements for (a) ascending (July 30 – August 5, 2025) and (b) descending (July 31 – August 6, 2025) satellite LOS. Inverse geodetic modelling results of the measured displacements assuming a point source reservoir and dyke intrusion geometry for (c) ascending and (d) descending satellite LOS. Model residuals for (e) ascending and (f) descending satellite LOS. The model geometry and geographic context are shown in: (g) plan view; (h) perspective view from the east; and (i) perspective view from the south.

Inverse modelling of the surface deformation between July 30th and August 6th provides first-order estimates (and uncertainty) of the dyke geometry and magmatic feeding system (Figure 4,

Supplementary Figure 4 and Supplementary Table 1). The dyke is inferred to be striking SSE-NNW with a nearly vertical dip and an intruded volume of $\sim 3.2 \times 10^7 \text{ m}^3$. The dyke originates from a reservoir at $\sim 6.3 \text{ km}$ depth which undergoes a volume loss of $\sim 1.3 \times 10^7 \text{ m}^3$. The difference between the volume lost by the reservoir and that gained by the dyke can be explained by the compressibility of the magma^{42,43}. The ratio of the intruded volume to the volume lost by the reservoir is ~ 2.46 . Keeping the assumption of a spherical reservoir and accounting for a host-rock shear modulus of 1–10 GPa constrains the magma compressibility⁴² to be between 1.6×10^{-9} and $1.6 \times 10^{-10} \text{ Pa}^{-1}$, the upper range of which suggests a gas-rich compressible magma⁴³ consistent with the subduction zone setting.

After the dyke intrusion, the eruption began at 16:50 UTC on August 2nd, with explosions from the northern crater initially sending ash to $\sim 3\text{--}4 \text{ km}$ asl and later rising to $8\text{--}10 \text{ km}$ asl^{31,44}. Simultaneously, a fissure opened on the northwest flank producing a lava flow^{31,44}. Minor explosive activity with ash emission continued through to mid-November 2025³¹, while effusive activity migrated from the northwest fissure to the northern cone and eastern flank³¹, where lava continues to flow at the time of writing (August 2026). Both explosive and effusive eruptive products have been determined to be dacitic in composition³¹.

The eruption was accompanied by substantial gas emissions. TROPOMI detected SO_2 plumes emanating from Krasheninnikov from August 3rd onwards, typically extending for hundreds of kilometres downwind to the southeast, in contrast to the absence of gas prior to the eruption (Figure 3). Elevated post-earthquake SO_2 emissions are also evident from several other volcanoes on the peninsula (Figure 3). We calculate an initial (by 01:00 UTC on August 3rd) SO_2 mass release from Krasheninnikov of $\sim 14 \text{ kt}$ ($\pm \sim 5.6 \text{ kt}$), assuming a plume height of $\sim 7 \text{ km}$. Subsequent mass estimations are not possible due to the merging of plumes from multiple sources downwind.

Surface deformation during the eruption shows broader scale and higher rate subsidence compared to the pre-earthquake period, and is accompanied by horizontal motion directed in towards the central subsidence; a reversal compared to the pre-earthquake period (Figure 2). Inverse modelling of co-eruptive surface deformation identifies a magma reservoir with a volume loss of $\sim 1.26 \times 10^7 \text{ m}^3$ for Aug 2025 – March 2026, presumably feeding the eruption (Supplementary Table 2). The similarity in horizontal and vertical location (within the uncertainties) suggests the same reservoir that fed the dyke intrusion continued to supply the initial stages of the eruption (Supplementary Figures 5 and 6).

M_w 8.8 Earthquake Stress Transfer and Critical State

The close timing of the earthquake and magmatic activity warrants investigation of their potential geophysical coupling. The complex slip distribution and rupture velocities of the earthquake produced significant energy in the low-frequency range ($\sim 0.02\text{--}0.5 \text{ Hz}$)⁴⁵. However, in the magma reservoir size inferred by our geodetic modelling, where the calculated volume changes are equivalent to reasonable pressure changes ($\sim 1\text{--}10 \text{ MPa}$)^{13–15} in a 1000–1500 m radius reservoir⁴⁶, even lower

frequencies (<0.03 Hz) would be required to promote resonance¹⁸. Sloshing is also unlikely¹⁸ given a probable crystal-rich mush beneath Krasheninnikov, consistent with the reported multi-centennial eruptive cadence and prolonged intervals of crystal differentiation producing dacitic magma compositions³⁰. The recorded earthquake frequencies could have induced resonance in the edifice of Krasheninnikov instead and aided migration of fluids, where its width of 5700-8000 m and shear-wave velocities of 2000-3000 m/s produce resonance frequencies of 0.18-0.37 Hz²⁰, coincident with the highest radiation powers of the low-frequency earthquake energy⁴⁵. However, we have neither evidence nor independent inference of pre-existing fluids (magmatic or hydrothermal) stored within the edifice prior to the earthquake, so this mechanism is unlikely.

Stress modelling suggests the volcano was subjected to a coseismic static Coulomb stress change of 0.01-0.05 MPa, an extensional normal stress change on the plane the dyke eventually intruded along of 0.002-0.006 MPa, and a compressional volumetric strain of $\sim 2 \times 10^{-7}$ (Figure 5). Therefore, the earthquake produced an overall contraction in rock bulk volume at the volcano, but containing a component of extension in the direction perpendicular to the eventual dyke plane. Our results support early findings that the volcano was subjected to a compressive “rock pressure” of up to 0.03 MPa²⁸. Our calculated static stress changes are small, comparable to ocean tidal stresses⁵, and 2-3 orders of magnitude weaker than the magma overpressures typically assumed for magma reservoir failure (~ 1 -10 MPa)¹³⁻¹⁵, and thus played a negligible role in promoting the eruption. An overall compressive strain field would serve to make a magma reservoir relatively more stable¹³, but is slightly offset by a component of extension that may have unclamped fractures or faults and facilitated intrusion of the dyke^{1,5}. Together, the impacts of static stress and strain changes on the magmatic system (and its surroundings), and their influence on the eruption, are mechanically ambiguous and likely limited.

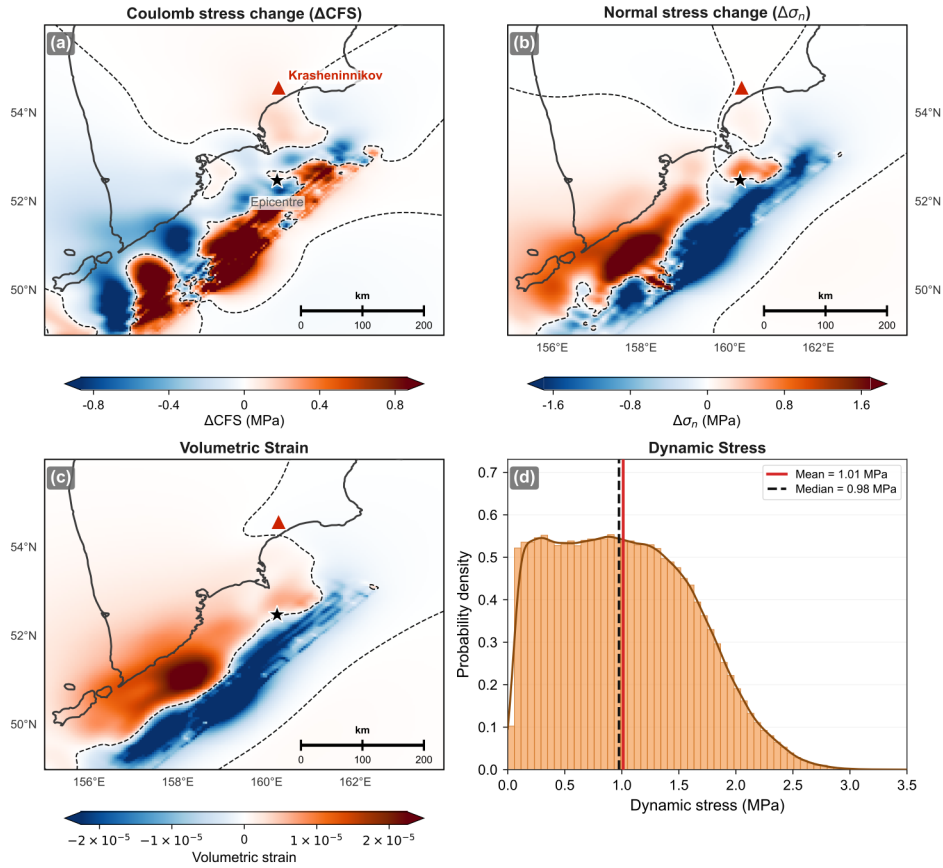


Figure 5: Earthquake-volcano stress transfer modelling results. (a) Coulomb stress change (ΔCFS) where the receiver fault is the geodetically-constrained eventual dyke plane (strike 158.2° , dip 81.2°). (b) Normal stress change on the eventual dyke plane ($\Delta\sigma_n$). (c) Volumetric strain. The red triangle denotes Krashennnikov volcano, while the black star indicates the earthquake epicentre. (d) Distribution of peak dynamic stress received at Krashennnikov volcano derived from estimated earthquake peak ground velocities.

The peak dynamic stress received at the volcano is estimated at $\sim 1.0 \pm 0.6$ MPa, with values greater than 2 MPa only possible with the upper potential ranges for peak ground velocity and shear modulus (Figure 5, Supplementary Figure 7). Given the ~ 190 – 280 s rupture duration^{28,29}, the volcano and magmatic system likely experienced relatively prolonged⁴⁷ dynamic stress changes. Our upper estimates for the peak dynamic stress imparted on the magmatic system by the earthquake overlap with the lower limits of magma reservoir failure overpressures (~ 1 – 10 MPa)^{13–15} and are in the range of dynamic stresses previously inferred to have triggered dyke intrusions^{2,48}. It is therefore theoretically feasible that the dynamic stress could have directly promoted magma reservoir failure. However, this scenario is unlikely for a number of reasons: (1) the overlap between the dynamic stress and reservoir failure values is narrow, and hence low probability; (2) the calculated dynamic stress values represent peak magnitudes, such that prolonged dynamic stressing could have been lower in magnitude and below the assumed thresholds for reservoir failure; and (3) the likely timing of the dyke intrusion, ~ 1.5 – 2.5 days after the earthquake, suggests a temporal lag between the dynamic stressing and dyke emplacement rather than direct failure immediately following, and caused by, the earthquake.

Given the earthquake-induced static and dynamic stresses alone are small and seem unlikely to have directly promoted reservoir failure, the magmatic system may already have been in a critical state. However, we have no evidence to support the presence of a critical state: there was no pre-earthquake volcano seismicity, ground deformation, or degassing, all of which can indicate a state of unrest that could potentially lead to an eruption^{49,50}. Therefore, this suggests either the magmatic system was not in a critical state, or such a state is currently hard to detect with remote sensing observations alone.

Volatile processes in magmatic systems have long been implicated in cases of earthquake-volcano interactions^{1,5}. Our inference of a compressible gas-rich magma from the volcano deformation modelling and substantial gas emissions accompanying the eruption³¹ (Figure 3) support a volatile-led process at Krasheninnikov, which could have been invisible to pre-earthquake remote sensing. In the presence of pre-existing bubbles in a volatile-rich magma, accumulated through centuries of crystallisation of anhydrous minerals, further volatile diffusion and bubble growth can be promoted by dynamic stresses^{1,5}. The prolonged duration of shaking from the earthquake, and high peak dynamic stresses, could therefore have facilitated substantial bubble growth and eventual destabilisation of the magma reservoir. Additionally, the abundance of nucleation sites present in a crystal-rich mush beneath Krasheninnikov should favour further bubble nucleation induced by dynamic pressure changes of a few MPa^{1,51,52}, coincident with our upper estimates. Together, it is plausible that time-dependent volatile processes contributed to the magma reservoir destabilisation, eventually leading to dyke intrusion and eruption. In this interpretation, pre-existing (and pre-earthquake) volatile-rich magma reservoir conditions maintained an apparently unobservable critical state.

An additional factor contributing towards a potentially hard-to-detect critical state relates to the magma-storage stress environment at Krasheninnikov. Magma-mush reservoirs⁵³, as well as melt-dominated chambers¹³, may be inherently unstable in extensional environments where the lithostatic stress could be insufficient to compensate for the internal load of magma, inducing a state of tension^{13,54}. Krasheninnikov sits within the extensional Kamchatka island arc⁵⁵, which may contribute to its magma reservoir being relatively mechanically unstable, and sensitive to additional perturbations that facilitate reservoir failure and eruption⁵⁶. From a regional perspective, the rare combination of a volcanic arc hosting volatile-rich magma susceptible to gas-led destabilisation, and exhibiting significant arc-internal extension directly above a repeatedly rupturing M_w 8–9 megathrust, may explain the relatively high likelihood of earthquake-facilitated eruptive activity in Kamchatka compared to other regions globally⁵⁷.

At Krasheninnikov in particular, remote sensing observations suggest the volcano was in a state of apparent quiescence prior to the July 29th, 2025, M_w 8.8 earthquake, which was closely followed by a dyke intrusion around August 1st, and initiation of eruption on August 2nd. The observed quiescence may have concealed a critical state related to the volcano-tectonic stress environment and volatile-

rich magma conditions. Ultimately, the earthquake likely facilitated the eruption not by directly forcing immediate reservoir failure, but by imposing prolonged dynamic stresses on a volatile-rich and mechanically susceptible magmatic system, prompting delayed volatile exsolution, bubble growth, reservoir destabilisation, and later dyke intrusion and eruption.

DRAFT

References

1. Seropian, G., Kennedy, B. M., Walter, T. R., Ichihara, M. & Jolly, A. D. A review framework of how earthquakes trigger volcanic eruptions. *Nat. Commun.* **12**, 1–13 (2021).
2. Hamling, I. J. & Kilgour, G. Goldilocks conditions required for earthquakes to trigger basaltic eruptions: Evidence from the 2015 Ambrym eruption. *Sci. Adv.* **6**, eaaz5261 (2020).
3. González, G. *et al.* Increment in the volcanic unrest and number of eruptions after the 2012 large earthquakes sequence in Central America. *Sci. Rep.* **11**, 22417 (2021).
4. Walter, T. R. How a tectonic earthquake may wake up volcanoes: Stress transfer during the 1996 earthquake–eruption sequence at the Karymsky Volcanic Group, Kamchatka. *Earth Planet. Sci. Lett.* **264**, 347–359 (2007).
5. Manga, M. & Brodsky, E. Seismic triggering of eruptions in the far field: Volcanoes and Geysers. *Annu. Rev. Earth Planet. Sci.* **34**, 263–291 (2006).
6. Hill, D. P., Pollitz, F. & Newhall, C. Earthquake–Volcano Interactions. *Phys. Today* **55**, 41–47 (2002).
7. Koyama, M. Mechanical Coupling Between Volcanic Unrests and Large Earthquakes. *J. Geogr. Chigaku Zasshi* **111**, 222–232 (2002).
8. Sulpizio, R. & Massaro, S. Influence of Stress Field Changes on Eruption Initiation and Dynamics: A Review. *Front. Earth Sci.* **5**, (2017).
9. Bebbington, M. S. & Marzocchi, W. Stochastic models for earthquake triggering of volcanic eruptions. *J. Geophys. Res. Solid Earth* **116**, (2011).
10. Sawi, T. M. & Manga, M. Revisiting short-term earthquake triggered volcanism. *Bull. Volcanol.* **80**, 57 (2018).

11. Jenkins, A., Rust, A. & Biggs, J. The relationship between large earthquakes and volcanic eruptions: A global statistical study. *Volcanica* **7**, 165–179 (2024).
12. Yamashina, K. & Nakamura, K. Correlations between tectonic earthquakes and volcanic activity of Izu-Oshima volcano, Japan. *J. Volcanol. Geotherm. Res.* **4**, 233–250 (1978).
13. Grosfils, E. B. Magma reservoir failure on the terrestrial planets: Assessing the importance of gravitational loading in simple elastic models. *J. Volcanol. Geotherm. Res.* **166**, 47–75 (2007).
14. Gudmundsson, A. How local stresses control magma-chamber ruptures, dyke injections, and eruptions in composite volcanoes. *Earth-Sci. Rev.* **79**, 1–31 (2006).
15. Currenti, G. & Williams, C. A. Numerical modeling of deformation and stress fields around a magma chamber: Constraints on failure conditions and rheology. *Phys. Earth Planet. Inter.* **226**, 14–27 (2014).
16. Walter, T. R. & Amelung, F. Volcanic eruptions following $M \geq 9$ megathrust earthquakes: Implications for the Sumatra-Andaman volcanoes. *Geology* **35**, 539–542 (2007).
17. Nishimura, T. Volcanic eruptions are triggered in static dilatational strain fields generated by large earthquakes. *Sci. Rep.* **11**, 17235 (2021).
18. Namiki, A., Rivalta, E., Woith, H. & Walter, T. R. Sloshing of a bubbly magma reservoir as a mechanism of triggered eruptions. *J. Volcanol. Geotherm. Res.* **320**, 156–171 (2016).
19. Namiki, A. *et al.* Crystal-Rich Magma Is Solid-Like and Liquefies When Deformed. *J. Geophys. Res. Solid Earth* **130**, e2024JB030483 (2025).

20. Namiki, A. *et al.* Volcanic activities triggered or inhibited by resonance of volcanic edifices to large earthquakes. *Geology* **47**, 67–70 (2019).
21. Lupi, M. & Miller, S. A. Short-lived tectonic switch mechanism for long-term pulses of volcanic activity after mega-thrust earthquakes. *Solid Earth* **5**, 13–24 (2014).
22. Wirth, E. A., Sahakian, V. J., Wallace, L. M. & Melnick, D. The occurrence and hazards of great subduction zone earthquakes. *Nat. Rev. Earth Environ.* **3**, 125–140 (2022).
23. Zellmer, G. F., Edmonds, M. & Straub, S. M. Volatiles in subduction zone magmatism. *Geol. Soc. Lond. Spec. Publ.* **410**, 1–17 (2015).
24. Cashman, K. V., Sparks, R. S. J. & Blundy, J. D. Vertically extensive and unstable magmatic systems: A unified view of igneous processes. *Science* **355**, (2017).
25. Argus, D. F., Gordon, R. G. & DeMets, C. Geologically current motion of 56 plates relative to the no-net-rotation reference frame. *Geochem. Geophys. Geosystems* **12**, (2011).
26. Koulakov, I., Bergal-Kuvikas, O., Voronova, M. & Jakovlev, A. Deep magma sources beneath Central Kamchatka inferred from teleseismic tomography. *Sci. Rep.* **15**, 16691 (2025).
27. U.S. Geological Survey National Earthquake Information Center. M 8.8 - 2025 Kamchatka Peninsula, Russia Earthquake.
<https://earthquake.usgs.gov/earthquakes/eventpage/us6000qw60/executive>.
28. Liu, C. *et al.* Simple unilateral rupture of the great Mw 8.8 2025 Kamchatka earthquake. *Science* **391**, 812–817 (2026).
29. Yin, H. Z. *et al.* Rapid Characterization of the 2025 Mw 8.8 Kamchatka, Russia Earthquake. *Seism. Rec.* **6**, 1–12 (2026).

30. Volynets, O. N., Ponomareva, V. V. & Tsyurupa, A. A. Petrological and Tephrochronological Studies of Krasheninnikov Volcano, Kamchatka. *Int. Geol. Rev.* **31**, 1107–1122 (1989).
31. Gorbach, N. V., Ozerov, A. Yu., Rogozin, A. N., Tolstykh, M. L. & Ovsyannikov, G. N. First Historical Eruption of Krasheninnikov Volcano (Eastern Kamchatka): Field Observations and Composition of Lavas Erupted in August–September 2025. *J. Volcanol. Seismol.* **20**, 1–14 (2026).
32. Ponomareva, V. V. et al. Krasheninnikov Caldera (Eastern Kamchatka): Age and Magnitude of Eruption. *Russ. J. Pac. Geol.* **19**, S42–S45 (2025).
33. Biggs, J. & Pritchard, M. E. Global volcano monitoring: What does it mean when volcanoes deform? *Elements* **13**, 17–22 (2017).
34. Hickey, J. et al. The Ups and Downs of Volcanic Unrest: Insights from Integrated Geodesy and Numerical Modelling. in *Advances in Volcanology* (Springer, 2017).
35. Parker, A. L., Biggs, J. & Lu, Z. Investigating long-term subsidence at Medicine Lake Volcano, CA, using multitemporal InSAR. *Geophys. J. Int.* **199**, 844–859 (2014).
36. Patton, A. I., Rathburn, S. L., Capps, D. M., McGrath, D. & Brown, R. A. Ongoing Landslide Deformation in Thawing Permafrost. *Geophys. Res. Lett.* **48**, e2021GL092959 (2021).
37. Edmonds, M., Liu, E. J. & Cashman, K. V. Open-vent volcanoes fuelled by depth-integrated magma degassing. *Bull. Volcanol.* **84**, 28 (2022).
38. Oppenheimer, C., Scaillet, B. & Martin, R. S. Sulfur Degassing From Volcanoes: Source Conditions, Surveillance, Plume Chemistry and Earth System Impacts. *Rev. Mineral. Geochem.* **73**, 363–421 (2011).

39. U.S. Geological Survey. Latest Earthquakes. *Earthquake Catalog*
<https://earthquake.usgs.gov/earthquakes/search>.
40. Lewi, E. *et al.* Scientific response to the 2024–2025 dyke intrusions in the Fentale-Dofen Region, Ethiopia: geophysical monitoring, surface manifestations, and hazard mapping. *Bull. Volcanol.* **87**, 64 (2025).
41. Kugaenko, Y., Titkov, N. & Saltykov, V. Constraints on unrest in the Tolbachik volcanic zone in Kamchatka prior the 2012–13 flank fissure eruption of Plosky Tolbachik volcano from local seismicity and GPS data. *J. Volcanol. Geotherm. Res.* **307**, 38–46 (2015).
42. Rivalta, E. & Segall, P. Magma compressibility and the missing source for some dike intrusions. *Geophys. Res. Lett.* **35**, 1–5 (2008).
43. McCormick Kilbride, B., Edmonds, M. & Biggs, J. Observing eruptions of gas-rich compressible magmas from space. *Nat. Commun.* **7**, 13744 (2016).
44. Global Volcanism Program. *Report on Krasheninnikov (Russia)*.
<https://volcano.si.edu/showreport.cfm?wvar=GVP.WVAR20250730-300190&utm>
(2025).
45. Tarumi, K. & Yoshizawa, K. Multi-frequency Teleseismic P-wave Back-projection of the 2025 Mw 8.8 Kamchatka Peninsula Earthquake.
<https://eartharxiv.org/repository/view/11283/> (2025).
46. Blake, S. Volcanism and the dynamics of open magma chambers. *Nature* **289**, 783–785 (1981).
47. Cui, Y., Li, S., Chen, L. & Aoki, Y. Large Megathrust Earthquakes Tend to Sustain an Increasingly Longer Duration Than Expected. *Geophys. Res. Lett.* **52**, e2024GL112985 (2025).

48. Schuler, J. *et al.* The Response of Taupō Volcano to the M7.8 Kaikōura Earthquake. *J. Geophys. Res. Solid Earth* **129**, (2024).
49. Sparks, R. S. J., Biggs, J. & Neuberg, J. W. Monitoring Volcanoes. *Science* **335**, 1310–1311 (2012).
50. Phillipson, G., Sobradelo, R. & Gottsmann, J. Global volcanic unrest in the 21st century: An analysis of the first decade. *J. Volcanol. Geotherm. Res.* **264**, 183–196 (2013).
51. Carey, R. J. *et al.* Externally triggered renewed bubble nucleation in basaltic magma: The 12 October 2008 eruption at Halema‘uma‘u Overlook vent, Kīlauea, Hawai‘i, USA. *J. Geophys. Res. Solid Earth* **117**, (2012).
52. Hurwitz, S. & Navon, O. Bubble nucleation in rhyolitic melts: Experiments at high pressure, temperature, and water content. *Earth Planet. Sci. Lett.* **122**, 267–280 (1994).
53. Mantiloni, L., Hickey, J. & Alshembari, R. Stress and Strain in Magma-Mush Reservoirs: Implications for Reservoir Failure and Magma Propagation. *J. Geophys. Res. Solid Earth* **131**, e2025JB032951 (2026).
54. Mantiloni, L., Hickey, J. & Alshembari, R. The role of gravity in the state of stress and pore pressure of poroelastic rocks. *Geophys. J. Int.* **242**, ggaf169 (2025).
55. Kozhurin, A. & Zelenin, E. An extending island arc: The case of Kamchatka. *Tectonophysics* **706–707**, 91–102 (2017).
56. Prejean, S. G. & Hill, D. P. The influence of tectonic environment on dynamic earthquake triggering: A review and case study on Alaskan volcanoes. *Tectonophysics* **745**, 293–304 (2018).

57. Eggert, S. & Walter, T. R. Volcanic activity before and after large tectonic earthquakes: Observations and statistical significance. *Tectonophysics* **471**, 14–26 (2009).
58. Lazecký, M. *et al.* LiCSAR: An Automatic InSAR Tool for Measuring and Monitoring Tectonic and Volcanic Activity. *Remote Sens.* **12**, 2430 (2020).
59. Maghsoudi, Y., Hooper, A. J., Wright, T. J., Lazecky, M. & Ansari, H. Characterizing and correcting phase biases in short-term, multilooked interferograms. *Remote Sens. Environ.* **275**, 113022 (2022).
60. Lazecký, M. *et al.* Strategies for improving and correcting unwrapped interferograms implemented in LiCSBAS. *Procedia Comput. Sci.* **239**, 2408–2412 (2024).
61. Lazecky, M., Fang, J., Hooper, A. & Wright, T. Improved Phase Unwrapping Algorithm Based on Standard Methods. in *IGARSS 2022 - 2022 IEEE International Geoscience and Remote Sensing Symposium* 743–746 (2022).
doi:10.1109/IGARSS46834.2022.9884337.
62. Morishita, Y. *et al.* LiCSBAS: An open-source insar time series analysis package integrated with the LiCSAR automated sentinel-1 InSAR processor. *Remote Sens.* **12**, 5–8 (2020).
63. Yu, C., Li, Z., Penna, N. T. & Crippa, P. Generic Atmospheric Correction Model for Interferometric Synthetic Aperture Radar Observations. *J. Geophys. Res. Solid Earth* **123**, 9202–9222 (2018).
64. Veefkind, J. P. *et al.* TROPOMI on the ESA Sentinel-5 Precursor: A GMES mission for global observations of the atmospheric composition for climate, air quality and ozone layer applications. *Remote Sens. Environ.* **120**, 70–83 (2012).

65. Theys, N. *et al.* A sulfur dioxide Covariance-Based Retrieval Algorithm (COBRA): application to TROPOMI reveals new emission sources. *Atmos. Chem. Phys.* **21**, 16727–16744 (2021).
66. Carn, S. A., Fioletov, V. E., McLinden, C. A., Li, C. & Krotkov, N. A. A decade of global volcanic SO₂ emissions measured from space. *Sci. Rep.* **7**, 44095 (2017).
67. Fioletov, V. E. *et al.* Version 2 of the global catalogue of large anthropogenic and volcanic SO₂ sources and emissions derived from satellite measurements. *Earth Syst. Sci. Data* **15**, 75–93 (2023).
68. Okada, Y. Surface deformation due to shear and tensile faults in a half-space. *Int. J. Rock Mech. Min. Sci. Geomech. Abstr.* **75**, 1135–1154 (1985).
69. Mogi, K. Relations between the eruptions of various volcanoes and the deformations of the ground surfaces around them. *Bull. Earthq. Res. Inst.* **36**, 99–134 (1958).
70. Trasatti, E. Volcanic and Seismic Source Modeling: An Open Tool for Geodetic Data Modeling. *Front. Earth Sci.* **10**, (2022).
71. Sambridge, M. Geophysical inversion with a neighbourhood algorithm—I. Searching a parameter space. *Geophys. J. Int.* **138**, 479–494 (1999).
72. Hickey, J., Pascal, K., Syers, R. & Alshembari, R. Topographical Effects on Volcano Deformation Signal Intensity: Implications for GPS Network Configuration. *Geophys. Res. Lett.* **51**, e2024GL108812 (2024).
73. Hickey, J., Gottsmann, J., Nakamichi, H. & Iguchi, M. Thermomechanical controls on magma supply and volcanic deformation: application to Aira caldera, Japan. *Sci. Rep.* **6**, 32691 (2016).

74. Head, M., Hickey, J., Gottsmann, J. & Fournier, N. Exploring the Impact of Thermally Controlled Crustal Viscosity on Volcanic Ground Deformation. *J. Geophys. Res. Solid Earth* **126**, 1–22 (2021).
75. Alshembari, R., Hickey, J., Williamson, B. J. & Cashman, K. Unveiling the Rheological Control of Magmatic Systems on Volcano Deformation: The Interplay of Poroviscoelastic Magma-Mush and Thermo-Viscoelastic Crust. *J. Geophys. Res. Solid Earth* **128**, e2023JB026625 (2023).
76. Okada, Y. Internal deformation due to shear and tensile faults in a half-space. *Bull. Seismol. Soc. Am.* **82**, 1018–1040 (1992).
77. Developers, T. P. Pyrocko: A Versatile Seismology Toolkit for Python. <https://doi.org/10.5880/GFZ.2.1.2017.001>.
78. Hill, D. P. et al. Seismicity Remotely Triggered by the Magnitude 7.3 Landers, California, Earthquake. *Science* **260**, 1617–1623 (1993).
79. Shapiro, N. M., Gorbato, A. V., Gordeev, E. & Dominguez, J. Average shear-wave velocity structure of the Kamchatka peninsula from the dispersion of surface waves. *Earth Planets Space* **52**, 573–577 (2000).
80. Corchete, V. 3D S-Wave Velocity Model of the Crust and Upper Mantle beneath the Sea of Okhotsk and the Kamchatka Peninsula. *Lithosphere* **2022**, 7323670 (2022).
81. Heap, M. J. et al. Towards more realistic values of elastic moduli for volcano modelling. *J. Volcanol. Geotherm. Res.* **390**, (2020).
82. Gudmundsson, A. *Rock Fractures in Geological Processes*. (Cambridge University Press, 2011). doi:<https://doi.org/10.1017/CBO9780511975684>.

Methods

InSAR Processing

We used C-band Sentinel-1A Single Look Complex (SLC) data acquired in Interferometric Wide (IW) swath mode with VV polarisation from ascending track 38 and descending track 60. The data were processed using the COMET-LiCSAR system⁵⁸ to generate interferograms over Krasheninnikov volcano. Standard COMET-LiCSAR products are generated at approximately 100 m pixel spacing using multilooking factors of 20 and 5 in the range and azimuth directions, respectively. Here, we reprocessed the data at higher spatial resolution using multilooking factors of 7 in range and 2 in azimuth, producing interferograms with an approximately 30 m pixel spacing. This higher-resolution processing was used to better resolve localised deformation on the volcanic edifice and to reduce the impact of InSAR phase bias in decorrelating terrain⁵⁹.

Topographic phase contributions were removed using the Copernicus 30 m DEM, and the Sentinel-1 scenes were cropped to the target area. Interferograms were generated using a small-baseline network, with a minimum temporal baseline of 12 days and up to five interferometric connections per acquisition. The ascending dataset included 282 acquisitions from 30 May 2016 to 27 March 2026, producing 2,026 interferograms, while the descending dataset included 271 acquisitions from 13 February 2016 to 28 March 2026, producing 1,924 interferograms. The interferogram networks were separated into pre-dyke and post-dyke subsets, with the pre-dyke time series extending to 31 July 2025 and the post-dyke time series beginning after 5 August 2025.

To improve phase continuity and reduce unwrapping errors in areas affected by vegetation-related decorrelation, we also applied an advanced re-unwrapping procedure^{60,61}. Line-of-sight displacement time series and mean velocities were estimated independently for each orbit using LiCSBAS⁶². However, the displacement time series estimate for the post-dyke period alone (Aug 2025 – March 2026) should be deemed preliminary due to the short observation period and relative lack of satellite retrievals in this period. GACOS corrections were applied to reduce tropospheric artefacts⁶³. Statistical quality checks, small-baseline inversion, uncertainty estimation using percentile bootstrapping, and pixel masking based on noise indices were then applied following the LiCSBAS workflow⁶². This processing strategy produced high-resolution deformation maps suitable for identifying localised pre- and post-dyke displacement patterns across Krasheninnikov volcano. Additionally, ascending and descending data were combined to exploit the two satellite lines of sight and decompose the surface deformation to vertical and east-west motion, assuming negligible north–south displacement due to the limited sensitivity of Sentinel-1 to motion in this direction.

We analysed satellite-derived observations of SO₂ in the atmosphere over Kamchatka to detect volcanic gas emissions from Krasheninnikov prior to, and following, the July 2025 earthquake. We utilise SO₂ imagery from the TROPospheric Monitoring Instrument (TROPOMI), carried aboard the European Space Agency Sentinel-5P satellite⁶⁴. TROPOMI is carried in a sun-synchronous polar orbit, with a swath width of 2600 km, and spatial resolution at nadir of 5.5 x 3.5 km, thus achieving near daily global coverage. In this study, we use the updated Covariance Based Retrieval Algorithm (COBRA) SO₂ product⁶⁵.

TROPOMI provides column concentrations of SO₂ in the atmosphere over a volcano and the observation represents a 'snapshot' at the time of overpass with no information on the SO₂ source, plume altitude, or time and altitude of gas emission. For long-term observation windows of weeks to months, as we adopt here, the persistent presence of elevated SO₂ concentrations in the atmosphere above a volcano can be inferred to represent regular or sustained emissions^{66,67}.

We generated long-term (multi-month) average maps of SO₂ VCD (vertical column density) over Kamchatka from level 2 TROPOMI SO₂ total column data, accessed via the WEKEO portal (<https://wekeo.copernicus.eu/>). TROPOMI SO₂ column concentrations are provided assuming different plume altitudes, and here we have used the 7 km product, which we judge to be best suited to eruptive plumes over Kamchatka, but which will underestimate the true VCD of SO₂ in any plumes at lower (<7 km) altitudes.

We defined an area of interest encompassing the Kamchatka peninsula (Figure 3) and analysed all TROPOMI swaths that overlapped with this region for two intervals, 1st March to 29 July (pre-earthquake) and 30 July to 30 September (post-earthquake). The boreal winter is subject to poor signal to noise for UV sensors owing to low light conditions, hence our omission of January to February and October to December. We regridded the TROPOMI data to a fixed grid of ~5 km pixel resolution by nearest-neighbour interpolation, selecting the orbit with lower viewing zenith angle when two orbits were overlapping. The maps shown in Figure 3 therefore represent the average SO₂ column density across the region during the year of 2025.

In addition to the regridded data shown in Figure 3, we present single swath SO₂ maps, again assuming a plume altitude of 7 km, for a sequence of eight days extending from July 29th (immediately following the earthquake) to August 5th (one day prior to escalation of eruptive activity at Klyuchevskoy). These maps (Supplementary Figure 2) illustrate changes in SO₂ emissions from the volcanoes in Kamchatka on a daily basis. Plumes emanating from a given volcano can be assumed to represent that volcano's emissions, though with increasing distance downwind we see plumes merging (e.g. Supplementary Figure 2c, 2e, 2g, 2h). Large co-eruptive plumes may persist within the region covered by a single TROPOMI swath for longer than 24 hours, in which case subsequent

overpasses may contain both fresh and relict plumes, complicating the computation of total scene mass, or subsequent derivations of SO₂ emission rate. SO₂ mass in a plume is calculated by integrating the plume VCD values across the area of the pixels in that plume, with uncertainties calculated as a combination of the reported random and systematic uncertainties in the retrieved columns. Note that this does not include errors rising from the assumed SO₂ profile.

Deformation Modelling

Inversion of syn-dyke emplacement InSAR data was performed assuming that ground deformation was caused by a coupled dyke and reservoir system, in a homogeneous, isotropic, elastic half-space. We modelled the dyke as a rectangular tensile fault⁶⁸, with length, L_0 , width, W_0 , opening, t_0 , dip angle, δ_0 , and strike angle, ϕ_0 , measured clockwise from north. The dyke's location is defined by the depth, D_0 , and the longitude and latitude, x_0 and y_0 , respectively, of the top, northernmost corner of the dislocation. The reservoir was represented by an isotropic deformation point-source⁶⁹ defined by its longitude, x_{M1} , latitude, y_{M1} , depth, D_{M1} , and volume change, ΔV_{M1} . Both sources are embedded in a homogeneous, elastic half-space.

To run the inversion we used the open-source, Python-based toolbox “Volcanic and Seismic Source Modelling” (VSM)⁷⁰. We employed a global optimisation sampling algorithm (the Neighbourhood Algorithm⁷¹) to constrain the parameters of the two sources aiming to minimise the misfit between InSAR data and modelled deformation, with equal weight assigned to the dyke and reservoir sources. The inversion returns a best-fit model for the observed deformation, as well as the sampled distributions of source parameters providing an estimation of their probability density functions (PDFs), from which we also calculate 1σ uncertainties on the best-fit source parameters.

Inversion of the syn-eruptive ascending and descending InSAR data assumed a single isotropic deformation point-source⁶⁹ in a homogeneous, elastic half-space defined by its longitude, x_{M2} , latitude, y_{M2} , depth, D_{M2} , and volume change, ΔV_{M2} . The inversion was computed with the same approach described for the syn-dyke emplacement InSAR data.

Our deformation modelling neglects the roles of topography, subsurface heterogeneity, and inelastic crustal rheologies, which can all impact modelled surface deformation predictions^{72–75}. However, our focus here is not on producing a set of models to accurately describe the magmatic system and complex deformation patterns, but on acquiring first-order estimates for volumes and locations of magma intrusion and withdrawal to help assess the role of the earthquake in promoting the eruption.

Static Stress and Strain Calculations

Co-seismic static stress and strain changes were computed from the Liu et al. (2026) distributed-slip model of the 2025 M_w 8.8 Kamchatka earthquake²⁸. The crust was treated as a homogeneous,

isotropic, elastic half-space with a shear modulus of 40 GPa and Poisson's ratio of 0.25. Static deformation was evaluated using the analytic closed-form solution of Okada (1992) for rectangular dislocations in an elastic half-space⁷⁶, implemented in the compiled pyrocko toolbox⁷⁷. The slip model was used to compute the displacement field and the full displacement-gradient tensor throughout the half-space, evaluated on a regular grid at a fixed depth of 6 km, chosen to coincide with the inferred magma-storage and dyke-feeding region beneath Krasheninnikov.

From the displacement-gradient tensor we formed the symmetric strain tensor (ε_{ij}):

$$\varepsilon_{ij} = \frac{1}{2} \left(\frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right)$$

where u is the displacement component, x is the coordinate, and i and j are coordinate directions. We then obtained the volumetric (dilatational) strain (Δ) as its trace:

$$\Delta = tr(\varepsilon) = \varepsilon_{11} + \varepsilon_{22} + \varepsilon_{33}$$

with positive values denoting dilation and negative values compression. The corresponding Cauchy stress tensor (σ_{ij}) was computed via Hooke's law for a linear-elastic isotropic medium:

$$\sigma_{ij} = \lambda \delta_{ij} tr(\varepsilon) + 2\mu \varepsilon_{ij}$$

where δ_{ij} is the Kronecker delta, λ is the First Lamé parameter, and μ is the shear modulus.

Stress changes were then resolved onto a receiver fault representing the Krasheninnikov dyke as derived from the earlier deformation modelling, with strike $\varphi = 158.2^\circ$, dip $\delta = 81.2^\circ$, rake $\lambda = 0^\circ$ (pure tensile opening), and effective friction coefficient $\mu' = 0.4$. In the North–East–Down frame the fault-plane unit normal ($\hat{n}$) and slip-direction ($\hat{s}$) vectors are:

$$\hat{n} = (-\sin \delta \sin \varphi, \sin \delta \cos \varphi, -\cos \delta), \quad \hat{s} = (\cos \varphi, \sin \varphi, 0).$$

The traction acting on the fault plane is $t = \sigma \cdot \hat{n}$ where σ is the stress change tensor, from which the normal stress change is calculated ($\Delta\sigma_n$):

$$\Delta\sigma_n = \hat{n} \cdot \sigma \cdot \hat{n}$$

and the shear stress change ($\Delta\tau$) resolved in the slip direction is:

$$\Delta\tau = \hat{s} \cdot \sigma \cdot \hat{n}.$$

The Coulomb failure stress change (ΔCFS) was then computed as:

$$\Delta CFS = \Delta\tau + \mu' \Delta\sigma_n.$$

We adopt the convention that a positive $\Delta\sigma_n$ corresponds to unclamping (reduced normal stress, favouring dyke opening), and a positive ΔCFS indicates that the receiver fault is brought closer to failure, with negative values denoting clamping and stabilisation. Stress quantities are reported in MPa and volumetric strain as a dimensionless ratio.

Dynamic Stress Calculations

Peak dynamic stress (σ_D) is calculated according to^{3,78}:

$$\sigma_D = \frac{PGV * G}{V_s}$$

where PGV is the peak ground velocity, G is the shear modulus, and V_s is the seismic shear wave velocity. Due to a lack of locally-derived observations of PGV , we extract the estimated PGV from the USGS ShakeMap solution²⁷, which has a value of 0.122 m/s with a standard deviation (σ) of 0.018 m/s. Shear wave velocities in the region are estimated between 2000 and 3000 m/s^{79,80}, and we assess values of the shear modulus between 1 GPa and 40 GPa to account for varying rock properties along the seismic path^{81,82}. A Monte Carlo approach was used to propagate parameter uncertainty into estimates of peak dynamic stress. PGV was treated as a normally distributed variable centred on the observed value, with variability defined by its standard deviation, and was truncated at $\pm 3\sigma$ to avoid unphysical values. In contrast, G and V_s were treated as uniform distributions bounded by their lower and upper limits. A total of 200,000 samples were generated, and we report the mean and standard deviation derived from the resulting distribution of dynamic stress values in the manuscript text.

Acknowledgements

JH, RA, LM, YM, and BMK are supported by the Natural Environment Research Council (NERC) (NE/X013944/1). BE is supported by NERC (NE/S004106/1) and the UK Centre for Observation and Modelling of Earthquakes, Volcanoes, and Tectonics (COMET). GS is supported by a Royal Society Newton International Fellowship.

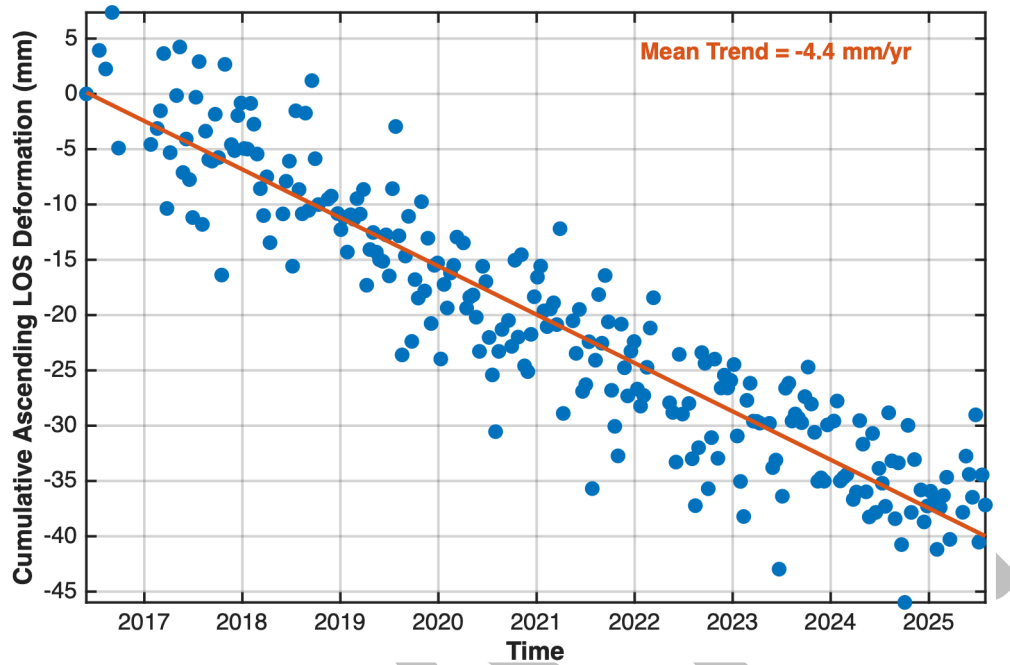
For the purpose of open access, the author has applied a 'Creative Commons Attribution (CC BY)' licence to any Author Accepted Manuscript version arising from this submission.

Source Data

Sentinel-1 SAR data used for the InSAR analysis are freely available through the Copernicus Data Space Ecosystem (CDSE, <https://dataspace.copernicus.eu>). The derived interferometric products were generated using the COMET-LiCSAR processing system. All TROPOMI data are freely available through the CDSE.

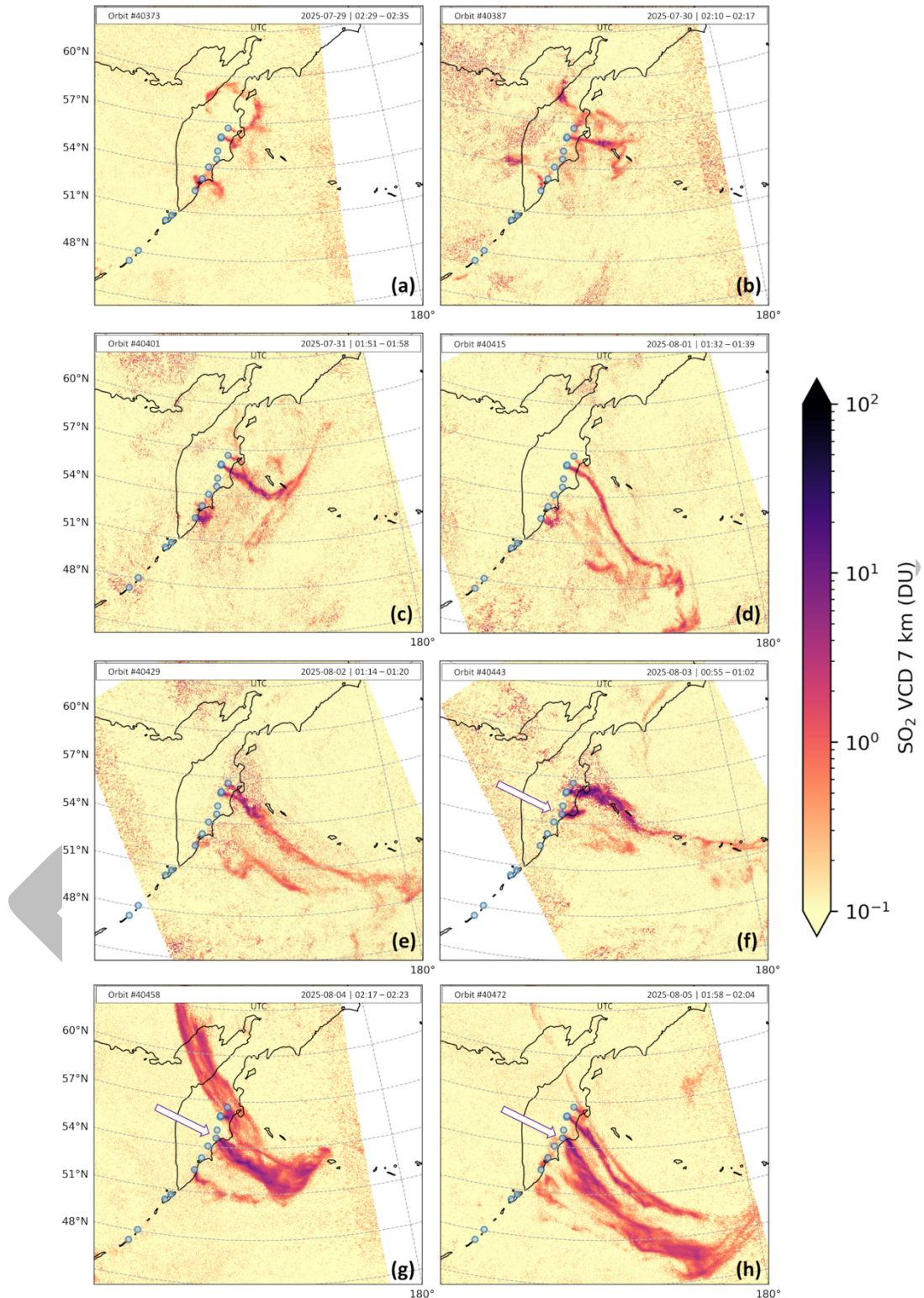
Supplementary Figures

Supplementary Figure 1



Supplementary Figure 1: Pre-dyke InSAR deformation time series from 13 February 2016 to 31 July 2025 using the ascending data set for a point on the eastern flank of the volcano. The red line indicates the mean trend of ~ 4.4 mm/yr of subsidence.

Supplementary Figure 2

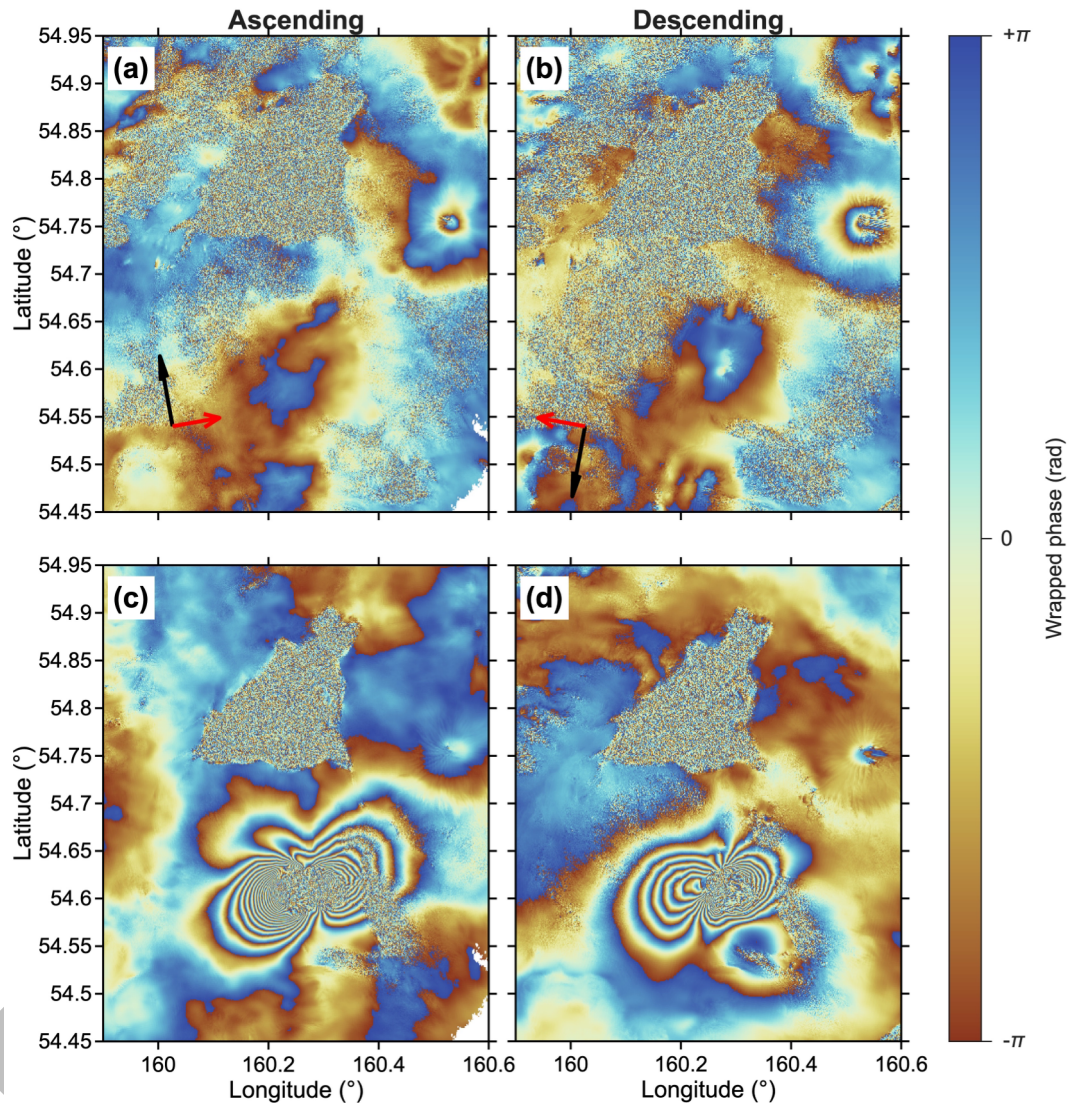


Supplementary Figure 2: Sequence of daily SO₂ plume maps over Kamchatka, generated from TROPOMI 7 km VCD (vertical column density) COBRA SO₂ product¹. The images run from August 29th (overpass ~8 hours after earthquake). The first SO₂ emissions from Krasheninnikov (indicated with white arrow) occur in panel (f), ~4 days post-earthquake and ~8 hours post-eruption of

Krasheninnikov. Strong SO₂ emissions can also be seen from Shiveluch, Klyuchevskoy, and Mutnovsky. The volcanoes indicated by circle icons are, north to south, Shiveluch, Klyuchevskoy, Bezymianny, Krasheninnikov, Kizimen, Karymsky, Avachinsky, Gorely, Mutnovsky, Ebeko, Chikurachiki, Sarychev Peak and Ketoi. DU = Dobson Unit.

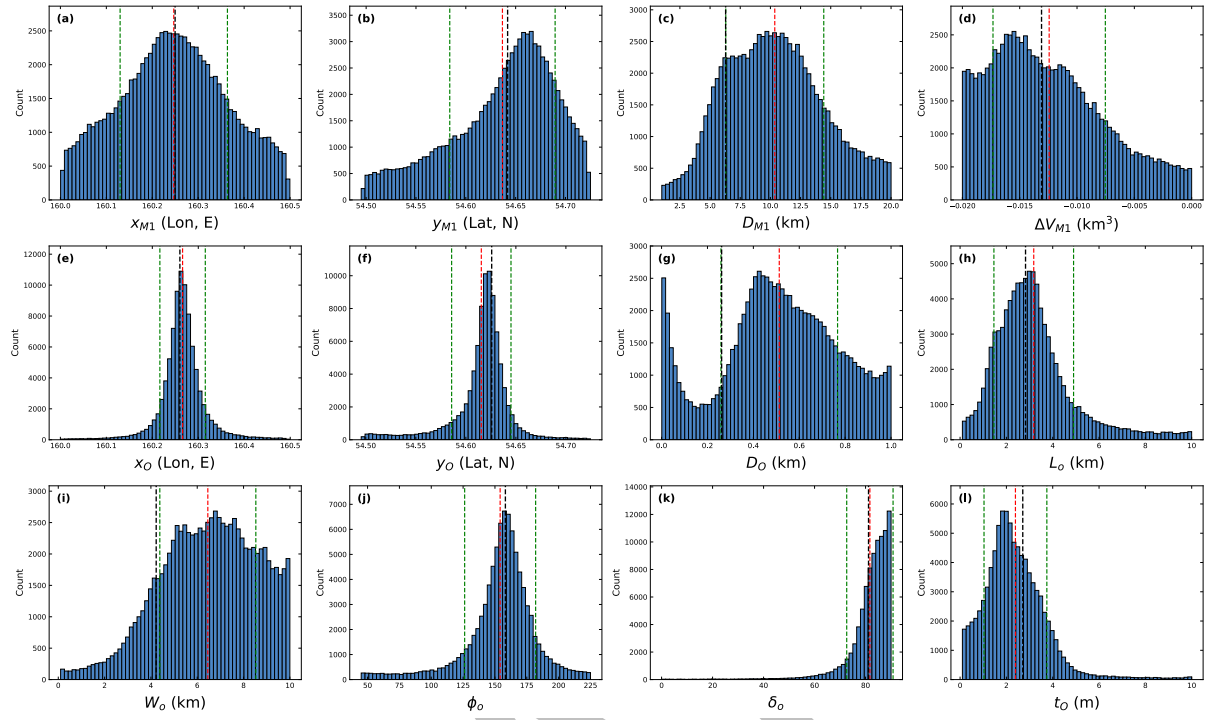
DRAFT

Supplementary Figure 3



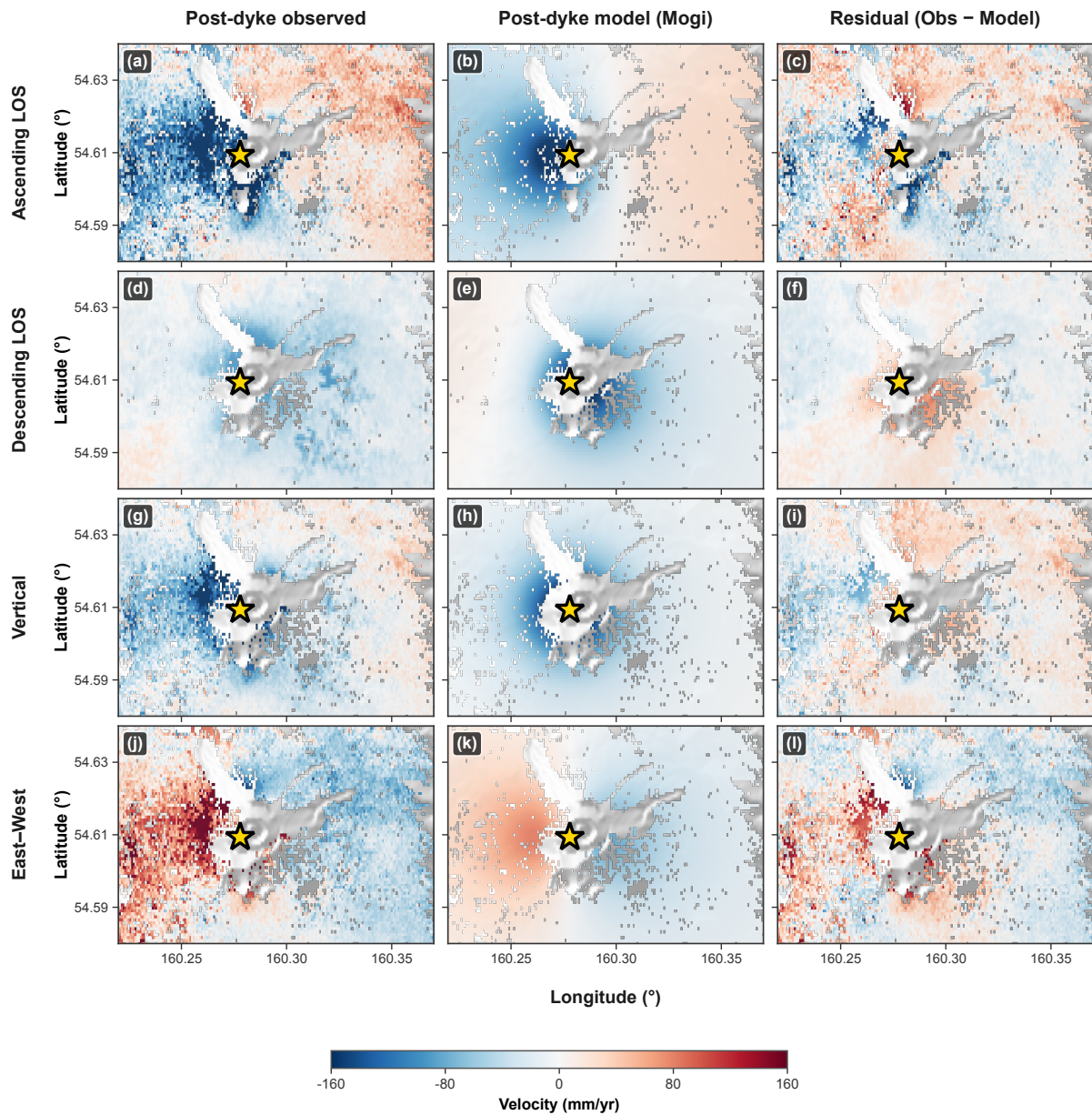
Supplementary Figure 3: Wrapped interferograms spanning the period prior to (top-row) and following (bottom-row) the July 29th, 2025, M_w 8.8 earthquake. (A) Ascending interferogram spanning July 6th, 2025, to July 30th, 2025. (B) Descending interferogram spanning July 7th, 2025, to July 31st, 2025. (C) Ascending interferogram spanning July 30th, 2025, to August 5th, 2025. (D) Descending interferogram spanning July 31st, 2025, to August 6th, 2025. In (A) and (B) the black arrow indicates satellite flight direction, and the red arrow the horizontal projection of the satellite line of sight. Mean incidence angles are 39.3° for the ascending data and 33.6° for the descending data. The surface manifestation of the dyke intrusion is only evident in (C) and (D) as the butterfly-shaped phase fringes.

Supplementary Figure 4



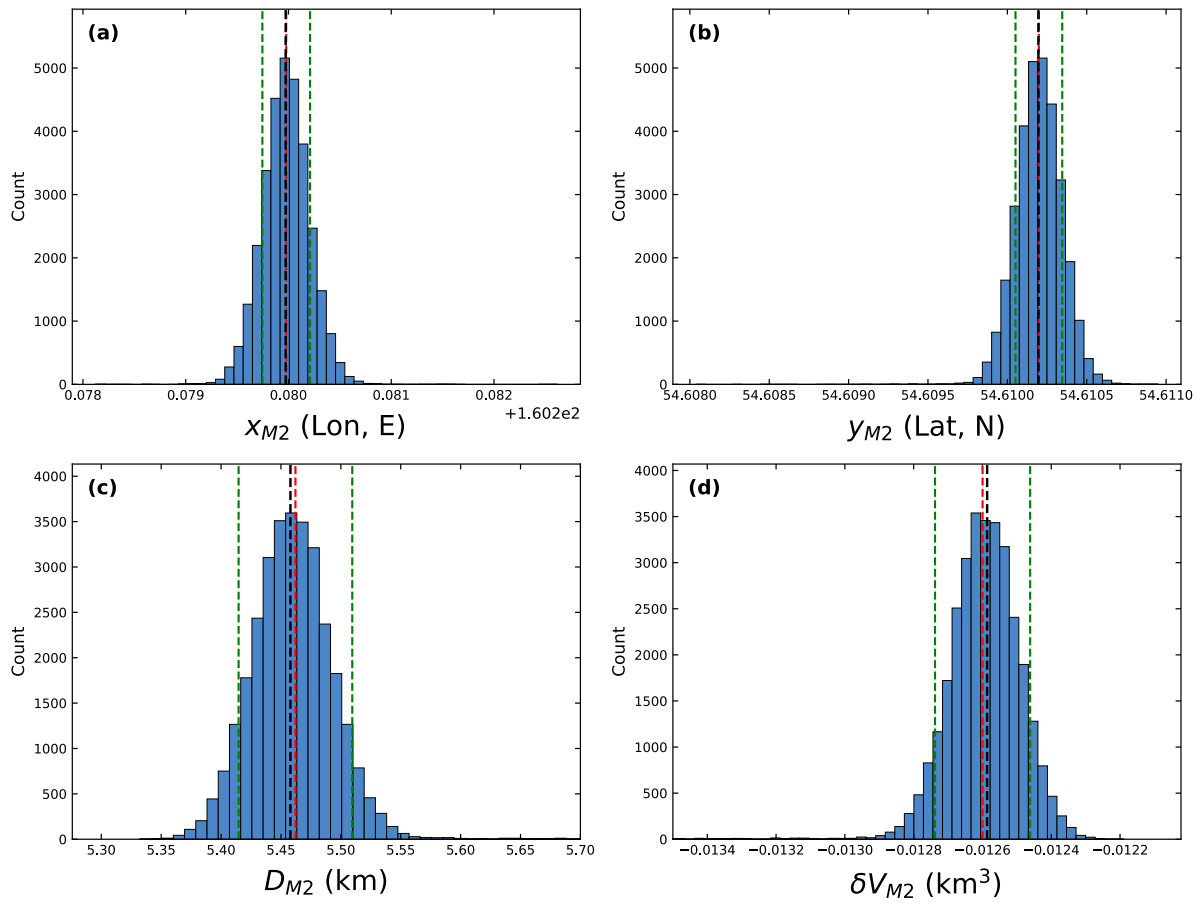
Supplementary Figure 4: Inverse geodetic modelling result probability density functions for the syn-dyke period (July 30th 2025 to August 6th 2025) assuming a point-source reservoir and a dyke intrusion model geometry. Dashed lines mark the mean (red) and best-fit (black) values, and the $\pm$ standard deviation intervals (green). Model parameters are as defined in supplementary table 1.

Supplementary Figure 5



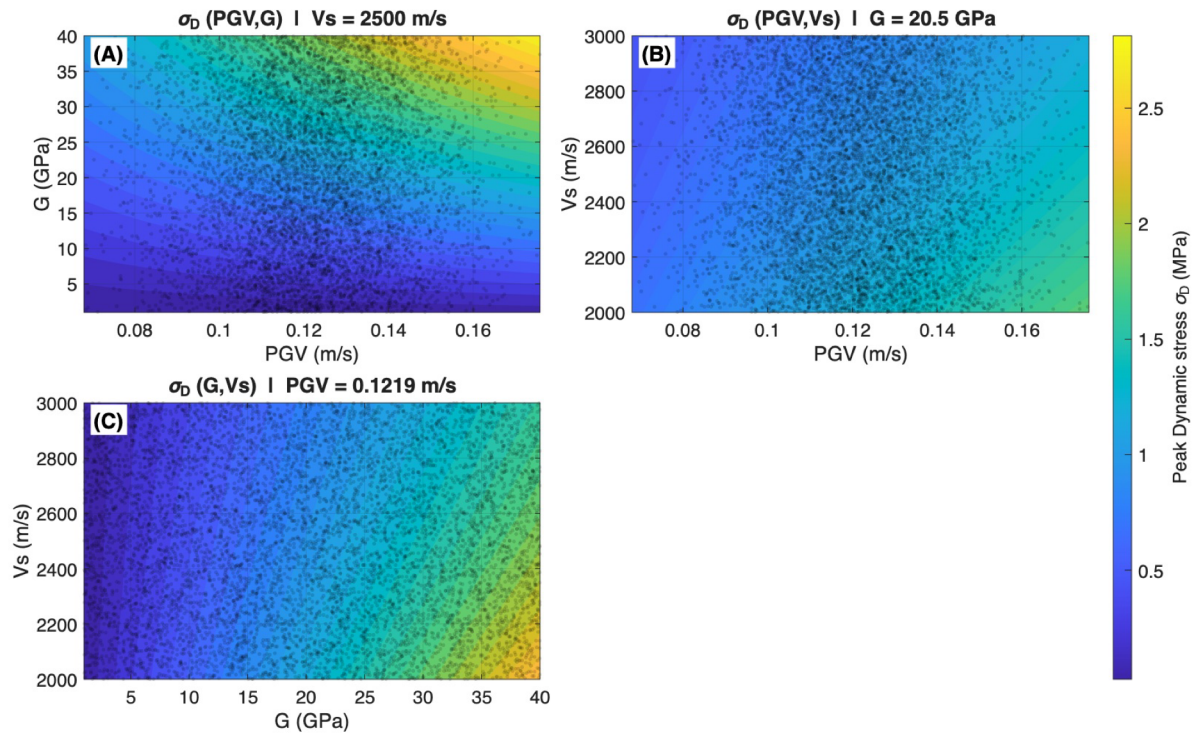
Supplementary Figure 5: Post-dyke intrusion InSAR observations and modelling. Left column shows the post-dyke surface displacement velocities between August 2025 and March 2026, for: (a) ascending satellite LOS; (d) descending satellite LOS; (g) near-vertical decomposition; and (j) near-east-west decomposition. Middle column shows inverse geodetic modelling results of the measured displacements assuming a point source reservoir, with the reservoir location marked with a star. Right column shows the model residuals.

Supplementary Figure 6



Supplementary Figure 6: Inverse geodetic modelling result probability density functions for the post-dyke period (August 2025 to March 2026) assuming a single, point-source reservoir. Dashed lines mark the mean (red) and best-fit (black) values, and the $\pm$ standard deviation intervals (green). Model parameters are as defined in supplementary table 2.

Supplementary Figure 7



Supplementary Figure 7: Peak dynamic stress (σ_D) calculation sensitivities from a Monte Carlo approach (see Methods). (A) Peak dynamic stress values for a seismic shear wave velocity (V_s) of 2500 m/s, and a range of peak ground velocities (PGV) and shear moduli (G). (B) Peak dynamic stress values for $G = 20.5$ GPa, and a range of PGV and V_s . (C) Peak dynamic stress values for $PGV = 0.1219$ m/s, and a range of G and V_s . Black circles represent individual solutions from the Monte Carlo approach.

Supplementary Tables

Supplementary Table 1: Geodetic modelling results for the dyke intrusion period

Parameter	Symbol	Best-fit value	Uncertainty	Unit
Okada dislocation (pure opening)				
Longitude (northernmost top corner)	x_o	160.26	± 0.05	
Latitude (northernmost top corner)	y_o	54.63	± 0.03	
Length	L_o	2827	± 1720	m
Width	W_o	4228	± 2073	m
Depth	D_o	262	± 254	m
Strike angle	φ_o	158.2	± 27.8	°
Dip angle	δ_o	81.2	± 9.1	°
Opening	t_o	2.7	± 1.4	m
Volume change	ΔV_o	3.2×10^7	$\pm 1.7 \times 10^7$	m ³
Mogi source				
Longitude (centre)	x_{M1}	160.25	± 0.12	
Latitude (centre)	y_{M1}	54.64	± 0.05	
Depth	D_{M1}	6290	± 4060	m
Volume change	ΔV_{M1}	-1.3×10^7	$\pm 0.5 \times 10^7$	m ³

Supplementary Table 2: Geodetic modelling results for the syn-eruption/post-dyke period

Parameter	Symbol	Best-fit value	Uncertainty	Unit
	Mogi source			
Longitude (centre)	x_{M2}	160.280	± 0.0021	
Latitude (centre)	y_{M2}	54.610	± 0.0012	
Depth	D_{M2}	5460	± 320	m
Volume Change	ΔV_{M2}	-1.26×10^7	$\pm 1.1 \times 10^6$	m ³