

Finite element modelling of crustal anisotropy in volcanic systems using transversely isotropic media

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ABSTRACT

Numerical models are essential tools in volcanology for understanding stress redistribution, ground deformation, and eruption precursors associated with magma migration within the crust. Although many models assume an isotropic crust, natural crustal rocks are commonly mechanically anisotropic due to layering, foliation, or aligned fracture networks. This study uses three-dimensional finite element modelling to examine how elastic anisotropy influences stress distribution and surface deformation induced by pressurized magma chambers. The crust is represented as a transversely isotropic medium, capturing directional stiffness without explicitly modelling mechanical layering. Spherical, horizontally elongated, and vertically elongated chambers are simulated under different anisotropy orientations and overpressure conditions. Results show that anisotropy generates asymmetric stress and deformation patterns, including lateral shifts of tensile stress, displacement maxima, and stress rotations of up to 90°. The locations of maximum stress and deformation are largely independent of overpressure and are mainly controlled by chamber geometry and crustal fabric.

RESUMEN

Los modelos numéricos son herramientas esenciales en volcanología para analizar la redistribución de esfuerzos, la deformación del terreno y los precursores eruptivos asociados a la migración de magma. Aunque muchos modelos asumen una corteza isotrópica, las rocas suelen ser mecánicamente anisotrópicas debido a su estratificación, foliación o familias de fracturas. En este estudio se emplea modelación tridimensional mediante elementos finitos para evaluar cómo la anisotropía elástica controla la distribución de esfuerzos y la deformación superficial inducidas por cámaras magmáticas presurizadas. La corteza se representa como un medio transversalmente isotrópico, que captura rigideces direccionales sin modelar explícitamente la estratificación. Se simulan cámaras esféricas, alargadas horizontal y verticalmente, bajo distintas orientaciones de anisotropía y condiciones de sobrepresión. Los resultados muestran patrones asimétricos de esfuerzo y deformación, incluyendo desplazamientos laterales de la deformación máxima y rotaciones de esfuerzo de hasta 90°, controlados principalmente por la geometría de la cámara y la fábrica de la roca.

KEYWORDS: Finite Element Method; Rock anisotropy; Transversely Isotropic; Volcano deformation.

1 INTRODUCTION

Volcanic unrest and failure are strongly controlled by the redistribution of stresses around magma reservoirs [Hutton 1988; Rubin 1995; Gudmundsson 2011]. Pressure changes within shallow magmatic bodies, whether through magma accumulation or withdrawal, generate heterogeneous stress fields that influence dyke propagation, fault activation, and ground deformation [e.g. Pinel and Jaupart 2003; Gray and Monaghan 2004; Currenti and Williams 2014; Pascal et al. 2014; Villarroel et al. 2024]. These interactions are affected by the mechanical structure of the crust, including variations in stiffness, stratigraphy, and anisotropy [Trasatti et al. 2003; Manconi et al. 2007; Forbes Inskip et al. 2020; Gudmundsson et al. 2022; Drymoni et al. 2023]. Representing the mechanical properties of the crust surrounding magmatic systems is therefore essential for understanding geophysical signals during volcanic unrest and

improving volcanic hazard forecasts [Sparks 2003; Segall 2013; Browning et al. 2015; Heap et al. 2020].

Previous Finite Element Method (FEM) studies have shown that host rock layering, geometry, and mechanical contrasts significantly influence the location, magnitude, and orientation of stresses around pressurized magma chambers [Gudmundsson 2006; Masterlark 2007; Long and Grosfils 2009; Clunes et al. 2024]. Gudmundsson [2006] showed that dipping layers with contrasting mechanical properties can cause lateral shifts in tensile stress maxima and stress rotations, especially under high layer stiffness contrast conditions. Clunes et al. [2024] analysed this effect through a systematic comparison and found that dipping mechanically stratified crustal rocks, with stiffness contrasts up to two orders of magnitude, can induce lateral shifts of more than 2 km in the location of maximum tensile stress (σ_3) at the surface, and stress rotations of up to 30° in the orientation of principal stresses above the magma chamber. These stress rotations were especially pronounced when the dip of the elastic layers ranged between

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30° and 60°. While these models effectively capture the influence of stratigraphy, they require detailed geometric assumptions and computationally expensive mesh refinement when multiple layers and dips are modelled, an approach that may not always be practical or geologically justified. Moreover, dipping mechanically heterogeneous rock layers have also been shown to significantly affect ground deformation patterns during dyke emplacement, with displacements up to 40 times larger and offsets exceeding 1 km depending on layer dip and stiffness contrast, compared to the expected for a non-layered crust [Clunes et al. 2023]. These findings underscore the need to assess magma-induced ground deformation signals in models that incorporate more complex crustal properties. Forbes Inskip et al. [2020] further demonstrated using FEM that fracture arrest in layered sequences is governed by contrast in Young's modulus between the layer hosting the fracture tip and the layer ahead, as well as by the proximity of the fracture tip to the interface. When the layer ahead is stiffer, tensile stress at the fracture tip decreases as it approaches the interface, promoting arrest before the interface is reached. These findings are directly relevant to the development of stress rotation and fracture arrest processes, where elastic anisotropy plays an analogous role in controlling dyke propagation paths.

To address these challenges, we explore an alternative numerical representation of a layered crust using an equivalent continuous, homogeneous, and transversely isotropic (TI) medium [Salamon 1968; Barla 1974]. This approach enables directional stiffness to be defined without the need for explicit layering, offering a simplified yet physically meaningful way to capture anisotropic crustal mechanical behaviour. Moreover, most crustal rocks exhibit some degree of elastic anisotropy [Barton and Quadros 2015]. Although elastic anisotropy is commonly associated with sedimentary and metamorphic rocks [e.g. Baud et al. 2005], even volcanic rocks can display significant elastic anisotropy due to the preferred orientation of pores and microcracks [e.g. Douglass and Voight 1969; Dai and Xia 2010]. Recent studies have shown that Young's modulus in volcanic rocks can vary up to 50% depending on the measurement direction relative to the dominant pore axis [Vairé et al. 2024], with pore geometry and orientation exerting a first-order control in both stiffness and strength [Bubeck et al. 2017; Griffiths et al. 2017; Merchán et al. 2026].

It is worth noting that elastic anisotropy in crustal rocks operates across a range of scales. At the sample scale, anisotropy arises from microstructural features such as aligned pores, microcracks, and mineral laminations, which impart directional dependence to elastic properties. At the rock mass scale, anisotropy emerges from larger-scale structures such as sedimentary bedding, metamorphic foliation, and networks of aligned fractures [e.g. Barton and Quadros 2015]. In volcanic settings specifically, lava flows can also develop anisotropy through preferred alignment of vesicles and crystals during emplacement, differential cooling, and the inclined geometry of successive flows on stratovolcano flanks or lava domes [e.g. Grosse et al. 2014; Heap et al. 2014; Merchán et al. 2026]. The transversely isotropic medium adopted in this study oper-

ates at the rock mass scale, capturing the aggregate directional stiffness of a crustal volume without resolving individual microstructural features. This formulation is therefore applicable to a broader range of geological contexts in which anisotropy is expressed at scales relevant to magma chamber-induced stress and deformation.

TI materials represent the simplest form of elastic anisotropy and are characterized by five independent elastic constants: two Young's moduli, two Poisson's ratios, and one shear modulus. They have been used in geomechanical modelling to represent rock fabrics such as foliation, sedimentary bedding, or aligned microcracks [e.g. Schoenberg 1994; Zhao et al. 2018; Li et al. 2020; Liu et al. 2024; Cortez et al. 2026], and have also been applied in models of crustal deformation and fluid flow [e.g. Nishizawa 1982; Chen and Nur 1992; Cheng 1993; Mukerji and Mavko 1994; Ciz and Shapiro 2009; Moukhtari et al. 2020]. Field examples illustrating the interaction between magmatic intrusions and rock fabrics, particularly foliation, are shown in Figure 1, where dykes exhibit highly variable and non-planar geometries, which appear to be strongly influenced by the anisotropic mechanical behaviour imparted by the host rock foliation. These complex geometries include deflection and branching patterns, suggesting that intrusion propagation was mechanically guided by fabric anisotropy at multiple scales.

Experimental studies have further illustrated the role of crustal anisotropy in volcanic systems through analogue models using layered gelatine, embedded fibres, or welded zones, showing that the orientation and strength of internal structures affect fracture propagation, dyke paths, and failure patterns [e.g. Rivalta et al. 2005; Kavanagh et al. 2006; Galland et al. 2009; Kavanagh et al. 2015; Montanari et al. 2020]. Recently, Giro et al. [2025] introduced a TI gelatine-based analogue material, demonstrating that mechanical properties such as stiffness, cohesion, and tensile strength, vary systematically with the angle between loading direction and internal fabric, influencing fracture nucleation and geometry. These findings provide an experimental foundation for using continuous anisotropic media in numerical simulations of volcano deformation.

In this study, we revisit a well-characterized mechanical scenario: pressurized magma chambers with different geometries embedded in a mechanically stratified crust with dipping layers. We compare stress distributions obtained from layered elastic models and continuous TI media. Using FEM simulations, we assess whether TI models can reproduce key stress localization patterns observed in layered models, including stress peaks, principal stress orientations, and ground deformation. Our aim is to evaluate the potential of continuous TI models as computationally efficient and geologically plausible alternatives for representing crustal anisotropy in active volcanic regions. Understanding how crustal anisotropies influence stress localization and magma-induced ground deformation is essential for improving forecasts of volcanic unrest and intrusion-related seismicity. This is particularly relevant in stratovolcanoes and calderas, where anisotropic crustal structures are common due to inclined and mechanically heterogeneous layers, as well as contrasts in lithology, folding, and

faulting [e.g. Gudmundsson 2006; Grosse et al. 2014; Clunes et al. 2021; 2024].

2 METHODOLOGY

In order to evaluate the influence of elastic anisotropy on stress and displacement fields, a parametric analysis using a three-dimensional model was developed in the finite element method in COMSOL Multiphysics 6.2. A three-dimensional approach was adopted because oblique anisotropy orientations ($0^\circ < \beta < 90^\circ$) break radial symmetry, making axisymmetric formulations inadequate. The three-dimensional formulation allows simultaneous evaluation of the XZ and YZ planes, whose contrasting responses to crustal anisotropy are key outcomes of this study. The stress tensor was computed based on the following equations:

$$-\nabla \cdot \boldsymbol{\sigma} = f \quad (1)$$

$$\boldsymbol{\epsilon} = \mathbf{S} : \boldsymbol{\sigma} \quad (2)$$

$$\boldsymbol{\epsilon} = \frac{1}{2} (\nabla \mathbf{u} + (\nabla \mathbf{u})^T) \quad (3)$$

where $\boldsymbol{\sigma}$ is the stress tensor, $f = (0, 0, -\rho g)$ is the body force per unit volume, $\boldsymbol{\epsilon}$ is the symmetric strain tensor, $\mathbf{S}$ is the fourth-order compliance tensor, $\mathbf{u}$ is the displacement vector field and $(\cdot)^T$ represents a transpose. The geomechanics convention for displacements, strain, and stress have been used in accordance with Brady and Brown [2006], where compressive normal stresses and contractile normal strains are taken as positive.

The crust was modelled using a TI media, which compliance tensor in three dimensions using Voigt notation is [Barla 1974]:

$$\begin{pmatrix} \epsilon_{xx} \\ \epsilon_{yy} \\ \epsilon_{zz} \\ \epsilon_{xy} \\ \epsilon_{yz} \\ \epsilon_{xz} \end{pmatrix} = \begin{pmatrix} s_{11} & s_{12} & s_{13} & 0 & s_{15} & 0 \\ s_{21} & s_{22} & s_{23} & 0 & s_{25} & 0 \\ s_{31} & s_{32} & s_{33} & 0 & s_{35} & 0 \\ 0 & 0 & 0 & s_{44} & 0 & s_{46} \\ s_{15} & s_{25} & s_{35} & 0 & s_{55} & 0 \\ 0 & 0 & 0 & s_{46} & 0 & s_{66} \end{pmatrix} \begin{pmatrix} \sigma_{xx} \\ \sigma_{yy} \\ \sigma_{zz} \\ \sigma_{xy} \\ \sigma_{yz} \\ \sigma_{xz} \end{pmatrix} \quad (4)$$

and each component of the compliance tensor was calculated using:

$$s_{11} = \frac{1}{E_1} \quad (5)$$

$$s_{12} = -\left(\frac{\nu_1}{E_1} \cos^2 \beta + \frac{\nu_2}{E_2} \sin^2 \beta\right) \quad (6)$$

$$s_{13} = -\left(\frac{\nu_1}{E_1} \sin^2 \beta + \frac{\nu_2}{E_2} \cos^2 \beta\right) \quad (7)$$

$$s_{15} = 2\left(\frac{\nu_2}{E_2} - \frac{\nu_1}{E_1}\right) \sin \beta \cos \beta \quad (8)$$

$$s_{22} = \frac{1}{E_1} \cos^4 \beta + \left(\frac{1}{G_2} - \frac{2\nu_2}{E_2}\right) \sin^2 \beta \cos^2 \beta + \frac{1}{E_2} \sin^4 \beta \quad (9)$$

$$s_{23} = \left(\frac{1}{E_1} + \frac{1}{E_2} + \frac{2\nu_2}{E_2} - \frac{1}{G_2}\right) \sin^2 \beta \cos^2 \beta - \frac{\nu_2}{E_2} \quad (10)$$

$$s_{25} = 2\left(\frac{\cos^2 \beta}{E_1} - \frac{\sin^2 \beta}{E_2}\right) - \left(\frac{1}{G_2} - \frac{2\nu_2}{E_2}\right) (\cos^2 \beta - \sin^2 \beta) \sin \beta \cos \beta \quad (11)$$

$$s_{33} = \frac{1}{E_1} \sin^4 \beta + \left(\frac{1}{G_2} - \frac{2\nu_2}{E_2}\right) \cos^2 \beta \sin^2 \beta + \frac{1}{E_2} \cos^4 \beta \quad (12)$$

$$s_{35} = \left(2\left(\frac{\sin^2 \beta}{E_1} - \frac{\cos^2 \beta}{E_2}\right) + \left(\frac{1}{G_2} - \frac{2\nu_2}{E_2}\right) (\cos^2 \beta - \sin^2 \beta)\right) \sin \beta \cos \beta \quad (13)$$

$$s_{44} = \frac{2(1 + \nu_1)}{E_1} \cos^2 \beta + \frac{1}{G_2} \sin^2 \beta \quad (14)$$

$$s_{46} = \left(\frac{2(1 + \nu_1)}{E_1} - \frac{1}{G_2}\right) \sin \beta \cos \beta \quad (15)$$

$$s_{55} = 4\left(\frac{1}{E_1} + \frac{1}{E_2} + \frac{2\nu_2}{E_2} - \frac{1}{G_2}\right) \sin^2 \beta \cos^2 \beta + \frac{1}{G_2} \quad (16)$$

$$s_{66} = \frac{2(1 + \nu_1)}{E_1} \sin^2 \beta + \frac{1}{G_2} \cos^2 \beta \quad (17)$$

where β is the dip angle measured counterclockwise from the horizontal plane of the main structural feature contributing to the anisotropy of the crust, such as bedding, foliation or fault sets. These dip angles span a range of geologically plausible configurations: $\beta = 0^\circ$ represents horizontal stratification, typical of caldera fill or flat-lying sedimentary sequences; intermediate values ($\beta = 30^\circ$ – 60°) represent inclined layers characteristic of stratovolcano flanks and lava domes, dipping metamorphic foliation, or tilted sedimentary units; and $\beta = 90^\circ$ represents subvertical structures such as steeply dipping metamorphic foliation, dykes or fault sets. E_1 and ν_1 are the Young's moduli and Poisson's ratio of the isotropic plane, and E_2 , ν_2 and G_2 are the Young's moduli, Poisson's ratio and shear modulus of the TI plane, respectively (see Figure 2).

The model domain, representing a portion of the crust, consists of a parallelepiped measuring 5 km $\times$ 5 km at its base and extending to a height of 3 km. Within this domain, a magmatic chamber was embedded at a centroidal depth of 1 km. The chamber was modelled using three geometric configurations: a spherical body, a horizontally elongated ellipsoidal body (as shown in Figure 2A), and a vertically elongated ellipsoidal body. The chamber's dimensions, denoted as L_x , L_y , and L_z , were varied in the parametric analysis within a range of 100 m to 500 m. Boundary conditions were defined as follows: normal displacements were constrained along the lateral boundaries, all degrees of freedom were fixed at the bottom, and the upper surface was left free to deform. Additionally, a uniform internal overpressure (p_0) of 5 MPa and 10 MPa was applied to the walls of the magma chamber.

This formulation is particularly useful for representing stratified or layered rock masses, where alternating layers with contrasting elastic properties can be approximated to an equivalent homogeneous rock mass as a transversely

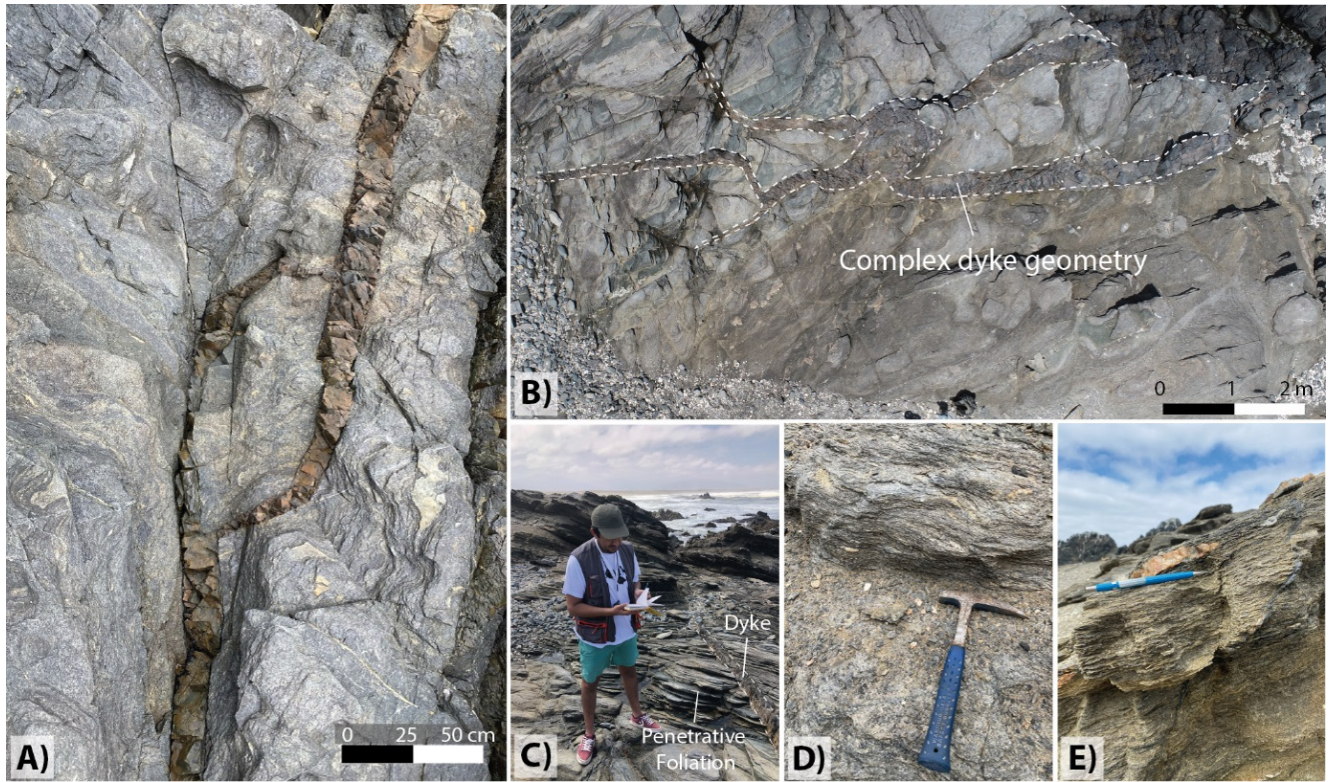


Figure 1: Field photographs showing examples of magma intrusions (dykes) into foliated metamorphic host rocks at Punta de Choros, Chile. [A] Steeply dipping dyke intruding folded and foliated metasediments. [B] Complex dyke geometry influenced by structural anisotropy. [C] Dyke cutting across penetrative foliation. [D]–[E] Detail of foliation planes in the host rock.

anisotropic medium using the relationships proposed by [Salomon \[1968\]](#). If all layers are isotropic with the same Poisson's ratio (ν), the five independent constants of the TI medium are:

$$\nu_1 = \nu_2 = \nu \quad (18)$$

$$E_1 = \sum \phi_i E_i \quad (19)$$

$$E_2 = \frac{(1 - \nu)E_1}{(1 + \nu)(1 - 2\nu)E_1 \sum \frac{\phi_i}{E_i} + 2\nu^2} \quad (20)$$

$$G_2 = \frac{1}{2(1 + \nu) \sum \frac{\phi_i}{E_i}} \quad (21)$$

where $\phi_i = t_i/L$, t_i is the thickness of the i -th layer, and L is the edge dimension of a cube, implying that $\sum \phi_i = 1$.

To account for the influence of elastic anisotropy, a single degree of anisotropy was considered, defined as the ratio $R = E_1/E_2$, where E_1 and E_2 are the Young's moduli parallel and perpendicular to the plane of isotropy, respectively (see [Figure 2B](#) and [2C](#)). In this study, a constant anisotropy ratio of $R = 3$ was adopted, with $E_1 = 75$ GPa, $E_2 = 25$ GPa, $\nu_1 = \nu_2 = 0.3$, and $G_2 = 8$ GPa. This ratio is consistent with values reported for foliated metamorphic rocks [[Homand et al. 1993](#)] and falls within the range of anisotropy commonly documented in crustal rock masses [[Barton and Quadros 2015](#)]. Furthermore, to explore the effect of anisotropy orientation, the dip angle of the plane of isotropy β was varied from 0° to 90° in 10° increments. The isotropic reference model uses $E =$

75 GPa, $\nu = 0.3$, and $G = 28.8$ GPa, consistent with the Young's modulus of the plane of isotropy (E_1) of the TI medium.

3 RESULTS

In an isotropic and homogeneous crust, the stress field surrounding a spherical or ellipsoidal magma chamber exhibits radial symmetry. Zones of maximum tensile stress (defined here as regions where $\sigma_3 < 0$, according to the sign convention adopted in this study) develop preferentially along the lateral margin of the chamber, where tensile fractures may initiate once the tensile strength of the surrounding rock is exceeded. In such cases, the iso-surfaces of tensile stress display a toroidal geometry. At the ground surface, tensile stresses are centred directly above the magma chamber and decay radially, approaching negligible values at distances beyond the chamber's influence. By contrast, the introduction of elastic anisotropy results in a markedly more complex stress field. At depth, the iso-surfaces of tensile stress become asymmetric and geometrically distorted, deviating substantially from the toroidal symmetry characteristic of the isotropic case. In shallower regions, these iso-surfaces exhibit more intricate topologies, in some instances resembling closed non-orientable surfaces such as a Klein bottle [e.g. [Franzoni 2012](#)]. At the surface, the distribution of tensile stress becomes elongated along the direction of anisotropy inclination, and the maximum tensile stress is displaced laterally from the vertical axis of the chamber. The orientation of the maximum and minimum principal stresses (σ_1 and σ_3) shows only minor deviations be-

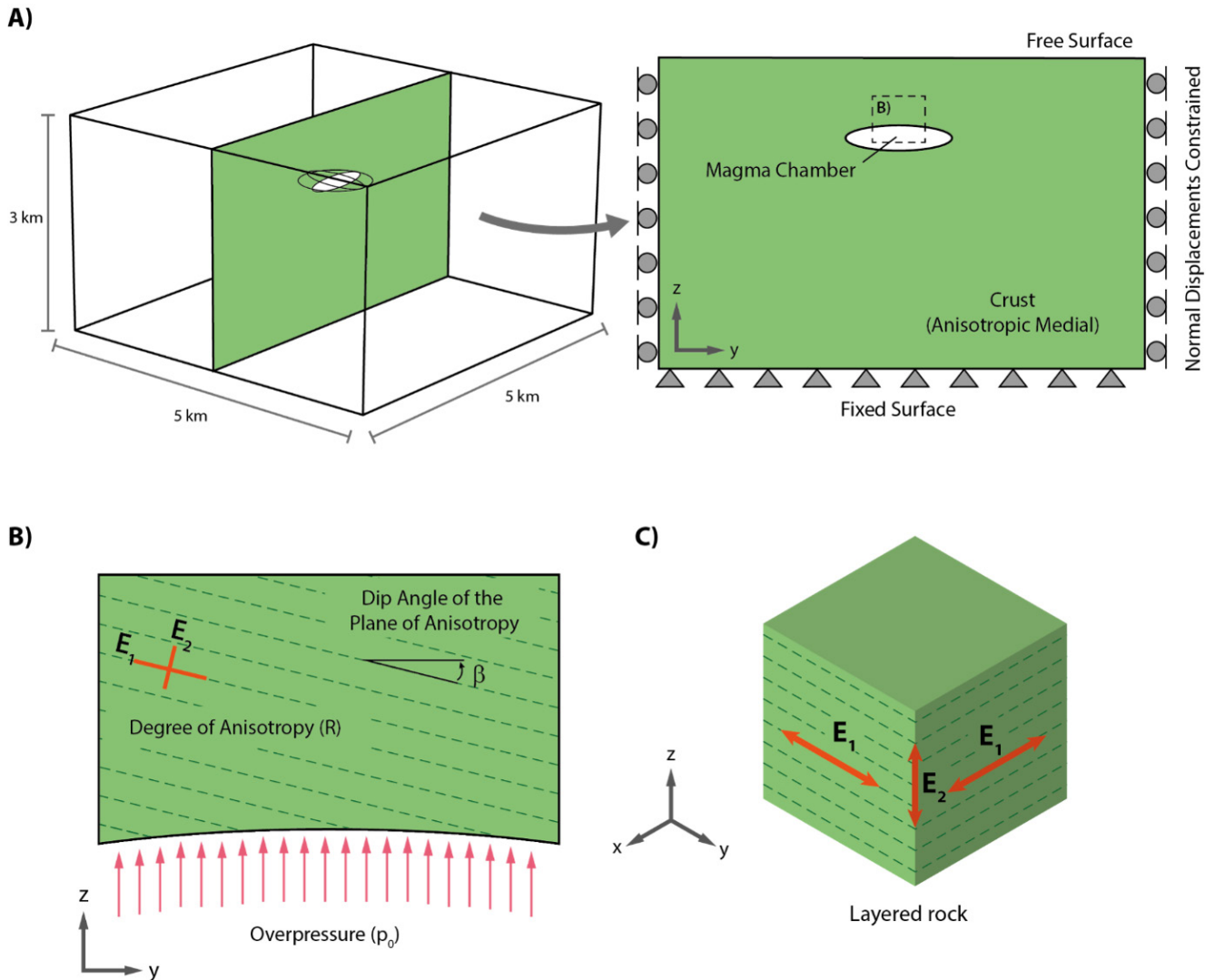


Figure 2: Numerical modelling setup for simulating crustal anisotropy using a transversely isotropic (TI) medium.

tween isotropic and anisotropic models at the ground surface. However, at depth, certain regions exhibit a complete rotation of the principal stress directions under anisotropic conditions. Figure 3 illustrates the distribution of tensile stress in both scenarios, at depth and at the surface, and includes the surface orientation of σ_1 and σ_3 for comparison.

For more details, Figure 4 illustrates the spatial distribution and orientation of the tensile stress (minimum principal stress, σ_3) in cross-sectional views of the YZ and XZ planes (left and right columns, respectively) around a horizontally elongated magma chamber located at a depth of 1 km. Figure 4A and 4B show the isotropic reference case, where the stress field is symmetric in both planes, with stress concentration lobes distributed evenly around the chamber's lateral margins. Figure 4C and 4D correspond to the anisotropic case with a dip angle of $\beta = 0^\circ$; in this configuration, the plane of isotropy is horizontal, producing stress patterns that remain relatively symmetric but begin to exhibit elongation along the anisotropy plane. Figure 4E and 4F represent the case with $\beta = 30^\circ$, where stress localization becomes more asymmetric. In Figure 4E, σ_3 orientation is deflected down-

ward, following the anisotropy orientation, resulting in an inclined stress distribution. Figure 4F retains a more symmetric pattern, although slight anisotropy effects are visible. In Figure 4G and 4H ($\beta = 60^\circ$), the anisotropy becomes more vertical, enhancing vertical stress redirection in the YZ plane (Figure 4G), while again, the XZ plane (Figure 4H) shows minor deviations. Finally, Figure 4I and 4J ($\beta = 90^\circ$) correspond to vertical anisotropy, where the stress field in the YZ plane (Figure 4I) shows highly vertical alignment of σ_3 orientation, and the stress concentration zones along the lateral flanks of the chamber are significantly reduced. The XZ plane (Figure 4J), in contrast, continues to show a nearly symmetric distribution, with more pronounced stress concentration zones than those observed in the YZ plane.

Throughout this study, we focus on the vertical component of surface displacement (u_z), as it represents the most directly observable signal in geodetic monitoring of volcanic unrest, being measurable by InSAR and GNSS, and is the component most sensitive to magma chamber pressurization. As with the stress field, when the dip angle of the anisotropy plane (β) differs from 0° and 90° , the vertical displacement field becomes

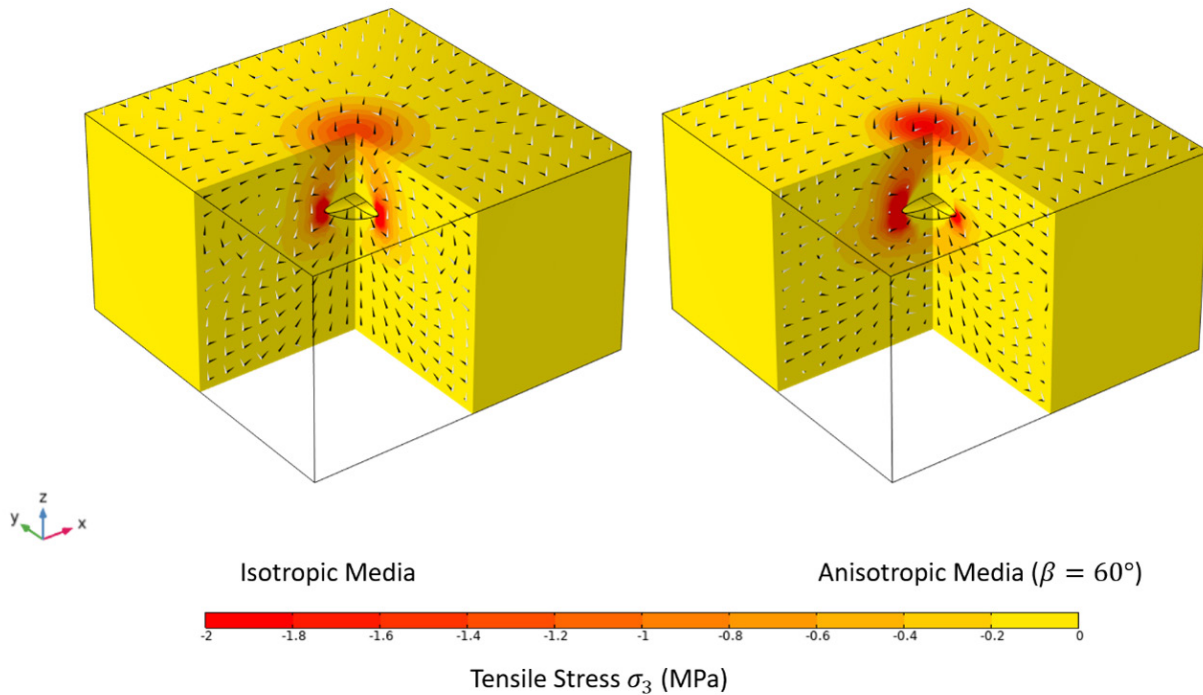


Figure 3: Iso-surfaces of the tensile stress (minimum principal stress, σ_3) around a horizontally elongated magmatic chamber embedded at 1 km depth and 5 MPa of overpressure. Black arrows denote the orientation of the maximum principal stress (σ_1), whereas white arrows indicate the orientation of the minimum principal stress (σ_3).

asymmetric in the YZ plane, while it remains symmetric in the XZ plane. Figure 5 presents u_z in both cross-sections for the isotropic model and for anisotropic models with dip angles of $\beta = 0^\circ$, 30° , 60° and 90° . The model corresponds to the case of a horizontally elongated magmatic chamber emplaced at 1 km depth, subject to a uniform overpressure of 5 MPa. In the isotropic case (Figure 5A and 5B), the displacement field is radially symmetric around the chamber, with a maximum surface uplift of 14 mm. For horizontal anisotropy ($\beta = 0^\circ$, Figure 5C and 5D), the reduced vertical stiffness leads to stronger vertical deformation, resulting in a surface uplift of up to 50 mm, with a broad and vertically elongated contour. This represents the highest displacement among all cases. For $\beta = 30^\circ$ (Figure 5E and 5F), which corresponds to a shallowly dipping anisotropy, the vertical displacement in the YZ plane becomes clearly asymmetric. The uplift contours are skewed toward the up-dip direction, with a peak value of 34 mm located approximately 250 m to the left of the chamber's central axis. In the XZ plane, the displacement distribution remains nearly symmetric. In contrast, the model with a steeply dipping anisotropy plane ($\beta = 60^\circ$, Figure 5G and 5H) the vertical displacement shifts in the opposite direction, toward the down-dip side, with a lower peak value of 19 mm located roughly 185 m to the right of the chamber axis. The deformation contours are narrower and more localized. Finally, in the case of vertical anisotropy ($\beta = 90^\circ$, Figure 5I and 5J), the highest stiffness is aligned with the vertical direction, which significantly suppresses vertical displacements compared to the horizontally anisotropic case. As a result, the uplift field is

once again symmetric, with a maximum surface displacement of only 23 mm.

The geometry of the magma chamber plays a key role in controlling both the distribution of tensile stresses and the pattern of vertical surface displacements. Figure 6 presents a comparison of these variables for anisotropic crustal models featuring different chamber shapes: horizontally elongated (Figure 6A and 6B), spherical (Figures 6C and D), and vertically elongated (Figures 6E and F) magma chambers.

In the case of a horizontally elongated ellipsoidal magma chamber (Figure 6A), the magnitude of tensile stress is lower in the anisotropic models compared to the isotropic case. The minimum surface tensile stress occurs when the anisotropy planes are vertical ($\beta = 90^\circ$), reaching approximately -2.6 MPa above the chamber's symmetry axis. A similar result is obtained when the anisotropy planes are horizontal ($\beta = 0^\circ$), with a minimum value of about -2.5 MPa, also located directly above the chamber axis. When the anisotropy planes are inclined, the minimum σ_3 shifts upward (up-dip) for low dips ($\beta = 30^\circ$) and downward (down-dip) for steeper dips ($\beta = 60^\circ$), reaching values near -1.7 MPa in both cases.

In contrast, when a spherical magma chamber is modelled (Figure 6C), the tensile stresses are substantially lower than those in the horizontally elongated ellipsoidal configuration. The lowest tensile stress occurs when the anisotropy planes are horizontal ($\beta = 0^\circ$), attaining -0.05 MPa at the surface directly above the chamber. As the dip of the anisotropy planes increases ($\beta = 30^\circ$ and 60°), the minimum tensile stress reaches values of approximately -0.03 and -0.055 MPa, respectively. Additionally, the location of the minimum tensile stress peak

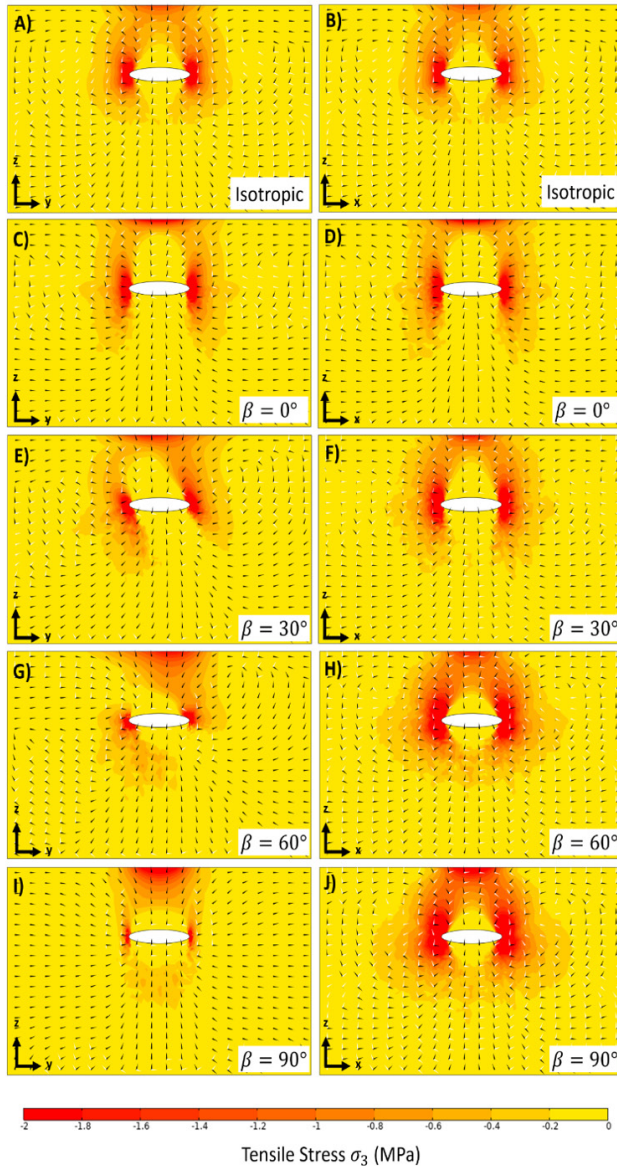


Figure 4: Distribution and orientation of the tensile stress (minimum principal stress, σ_3) around a horizontally elongated magmatic chamber embedded at 1 km depth. Black arrows denote the orientation of the maximum principal stress (σ_1), whereas white arrows indicate the orientation of the minimum principal stress (σ_3). The left and right columns show cross-sections in the YZ and XZ planes, respectively. Subplots [A] and [B] corresponds to the isotropic case, while the subsequent subplots [C] to [J] represent anisotropic media with varying dip angles (β) of the plane of anisotropy: 0° , 30° , 60° , and 90° . Elastic anisotropy is characterized by a constant stiffness ratio of $R = E_1/E_2 = 3$ and magma chamber geometry defined by $L_x = L_y = 500$ m and $L_z = 100$ m.

shifts up-dip by up to 500 m relative to the vertical axis of the magma chamber. When the anisotropy planes are vertical ($\beta = 90^\circ$), the minimum tensile stress again lies directly above the chamber, with a value of approximately -0.02 MPa. For the vertically elongated chamber configuration (Figure 6E), two distinct behaviours are observed. When the anisotropy planes

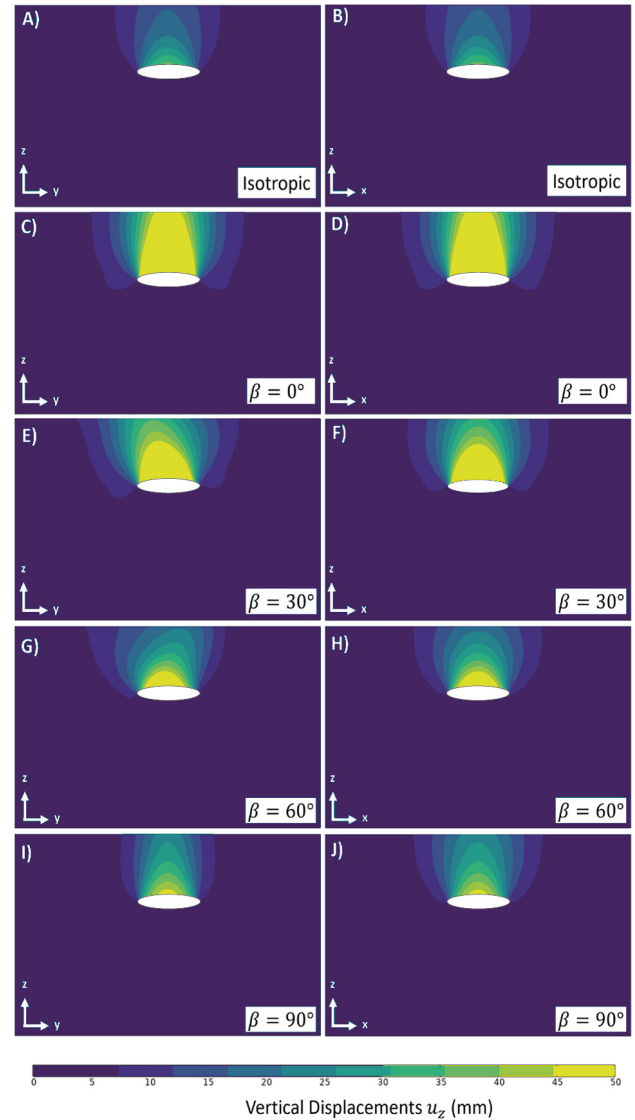


Figure 5: Vertical displacement field (u_z) around a horizontally elongated magmatic chamber embedded at 1 km depth. The left and right columns show cross-sections in the YZ and XZ planes, respectively. Subplots [A] and [B] corresponds to the isotropic case, while the subsequent subplots [C] to [J] represent anisotropic media with varying dip angles (β) of the plane of anisotropy: 0° , 30° , 60° , and 90° . Elastic anisotropy is characterized by a constant stiffness ratio of $R = E_1/E_2 = 3$ and magma chamber geometry defined by $L_x = L_y = 500$ m and $L_z = 100$ m.

are horizontal ($\beta = 0^\circ$), the minimum surface tensile stress is -0.18 MPa, located directly above the chamber. Its magnitude increases slightly and shifts up-dip as the dip of the anisotropy planes increases. However, under vertical anisotropy ($\beta = 90^\circ$) and in the isotropic model, the surface stress distribution exhibits two symmetric minima relative to the chamber's vertical axis.

In the case of surface vertical displacements, when the magma chamber is horizontally elongated (Figure 6B), the maximum uplift reaches 50 mm for models with horizontal

anisotropy planes ($\beta = 0^\circ$), located directly above the chamber. When the anisotropy planes are inclined at 30° , the maximum displacement decreases to 35 mm and shifts in the up-dip direction. For steeper inclinations ($\beta = 60^\circ$), the maximum uplift further decreases to 20 mm, displaced in the down-dip direction. In the isotropic model, the maximum displacement occurs directly above the chamber and only reaches 15 mm.

For the spherical chamber configuration (Figure 6D), surface vertical displacements are much smaller than those observed in the horizontally elongated case. In anisotropic models, the maximum uplift is approximately 0.7 mm and occurs when the anisotropy planes are horizontal ($\beta = 0^\circ$). As the dip angle increases, the magnitude of vertical deformation decreases and shifts progressively up-dip. Notably, when the anisotropy planes are vertical ($\beta = 90^\circ$), the resulting displacement is smaller than that observed in the isotropic model.

When the magma chamber is vertically elongated (Figure 6F), surface vertical displacement is also significantly reduced compared to the horizontally elongated configuration. Both the models with horizontal anisotropy ($\beta = 0^\circ$) and those with 30° inclination exhibit a maximum uplift of about 2.3 mm, however, in the latter case, the maximum is displaced up-dip from the chamber's symmetry axis, accompanied by minor subsidence down-dip. For steeper anisotropy ($\beta = 60^\circ$), the maximum vertical displacement reaches 1.3 mm in the up-dip direction, while subsidence down-dip attains approximately -0.4 mm. Under vertical anisotropy ($\beta = 90^\circ$), subsidence develops symmetrically above the chamber, reaching -0.5 mm at the surface.

Figure 7 summarizes the shift in the location of maximum vertical displacement (u_z) and maximum tensile stress (σ_3) as a function of the dip angle of the anisotropy planes (β) and chamber aspect ratio (L_y/L_z) for two different overpressure conditions (5 MPa and 10 MPa). The shift is defined as the horizontal distance between the position of the peak value and the vertical symmetry axis of the magma chamber, reported in metres. Negative values indicate up-dip migration of the peak; positive values indicate down-dip displacement. Since the models assume linear elasticity, doubling the applied overpressure scales the magnitudes of u_z and σ_3 proportionally but does not alter the position of their peaks. In this figure, aspect ratios $L_y/L_z > 1$ correspond to horizontally elongated chambers, $L_y/L_z = 1$ represents a spherical chamber, and $L_y/L_z < 1$ corresponds to vertically elongated chambers.

In Figure 7A and 7B, horizontally elongated chambers emplaced in anisotropic media with low dip angles ($\beta < 50^\circ$) exhibit an up-dip shift of the vertical displacement peak, which increases both with dip angle and with decreasing chamber aspect ratio. For steeper anisotropy orientations ($\beta > 50^\circ$), the displacement peak shifts down-dip, and its magnitude is roughly half that observed at lower dip angles, decreasing progressively as β increases. Spherical chambers display up-dip shifts up to $\beta = 70^\circ$, with magnitudes greater than those of horizontally elongated chambers, while at $\beta = 80^\circ$ the shift reverses down-dip. Vertically elongated chambers, in contrast, show only up-dip shifts, which increase rapidly with dip angle and remain approximately constant for $\beta > 60^\circ$.

The shift of tensile stress (Figure 6C and 6D) follows a similar trend to that of vertical displacement for spherical and vertically elongated chambers, though it occurs approximately 200 m closer to the vertical symmetry axis when the shift is up-dip, and coincides with the displacement peak when directed down-dip. In horizontally elongated chambers, up-dip shifts occur for low dip angles ($\beta \leq 40^\circ$) in models with aspect ratios of 4 and 5, with the tensile stress peak located 50–100 m nearer to the chamber's vertical axis than the displacement peak. Comparisons between the 5 MPa and 10 MPa overpressure cases indicate that the localization of both σ_3 and u_z peaks is independent of overpressure, being primarily controlled by chamber geometry and the anisotropic fabric of the surrounding rock.

4 DISCUSSION

4.1 Crustal anisotropy as a proxy for layering: implications for stress trajectories and magma propagation

Our results suggest that TI media can replicate many of the key stress and displacement patterns observed in models with explicit mechanical layering [e.g. Gudmundsson 2006; Clunes et al. 2024], including lateral shifts in tensile stress maxima, stress rotations, and asymmetric surface deformation patterns. While layered models offer finer control over heterogeneities at layer interfaces, they require high-resolution meshing and detailed knowledge of geometries that are often unavailable in volcanic settings. By contrast, TI models capture directional stiffness effects with fewer computational constraints, providing a geologically plausible and numerically efficient alternative for simulating crustal anisotropy in active volcanic regions.

A key outcome of our simulations is the abrupt rotation of the principal stress directions were observed in certain regions around the magma chamber, driven by elastic anisotropy of the host rock. If dyke intrusion follows mode I fractures [e.g. Rubin 1993], where fractures propagate parallel to the maximum principal stress (σ_1) and perpendicular to the minimum principal stress (σ_3), the orientation of stress has direct implications for the final geometry of the intrusion. When evaluating the propagation of planar fractures in anisotropic media, three principal fracture orientations can be defined [Chong et al. 1987; Forbes Inskip et al. 2018]: short-transverse, arrester, and divider (Figure 8). In the short-transverse orientation, both the fracture plane and the propagation direction are parallel to the structural anisotropy (i.e. layering, bedding, foliation). In contrast, the arrester orientation is characterized by both the fracture plane and the propagation direction being perpendicular to the anisotropy. In the divider orientation, the fracture plane is perpendicular to the anisotropy, but propagation occurs parallel to it.

To classify the fracture orientation mode, we evaluated the angular relationship between the fracture propagation direction ($\mathbf{v}$, assumed to align with σ_1), the orientation normal to fracture plane ($\mathbf{u}$, aligned with σ_3), and an orientation normal to the anisotropy plane ($\mathbf{n}$), which defines the axis of elastic anisotropy. A dyke was classified as divider when both $\mathbf{v}$ and $\mathbf{u}$ were approximately normal to the anisotropic plane

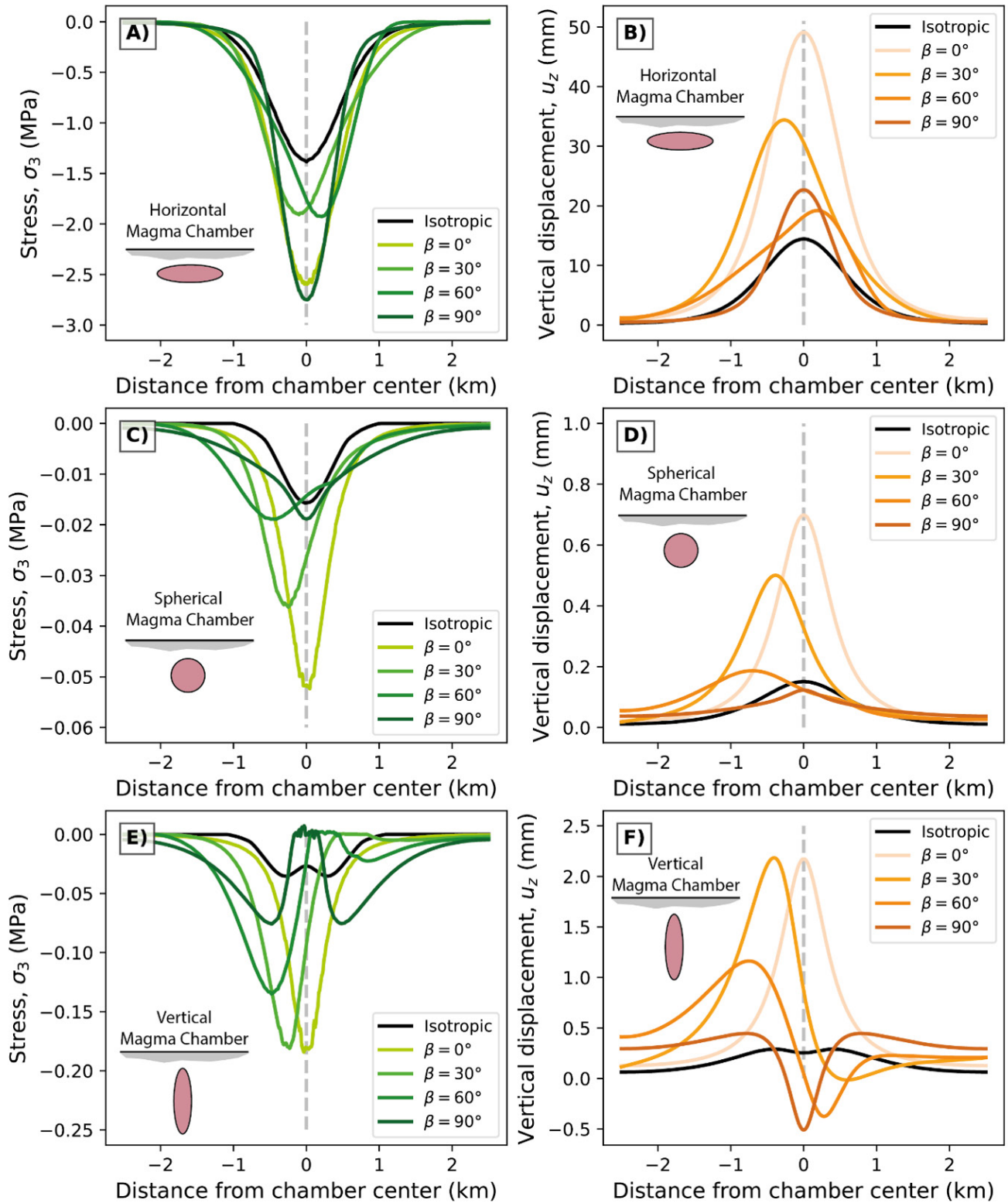


Figure 6: Comparison of tensile stress (σ_3) and surface vertical displacement (u_z) for different magma chamber geometries and anisotropy orientations. Subplots (A–B) correspond to a horizontally elongated ellipsoidal chamber, (C–D) to a spherical chamber, and (E–F) to a vertically elongated ellipsoidal chamber. The left column shows the distribution of tensile stress, while the right column presents the corresponding vertical surface displacement profiles. Elastic anisotropy is characterized by a constant stiffness ratio of $R = E_1/E_2 = 3$ and magma overpressure of 5 MPa. Magma chamber geometry is defined by $L_x = L_y = 500$ m and $L_z = 100$ m in the case of horizontally elongated ellipsoidal chamber, $L_x = L_y = L_z = 100$ m in the spherical model, and $L_x = L_y = 100$ m and $L_z = 500$ m in the case of the vertically elongated ellipsoidal chamber.

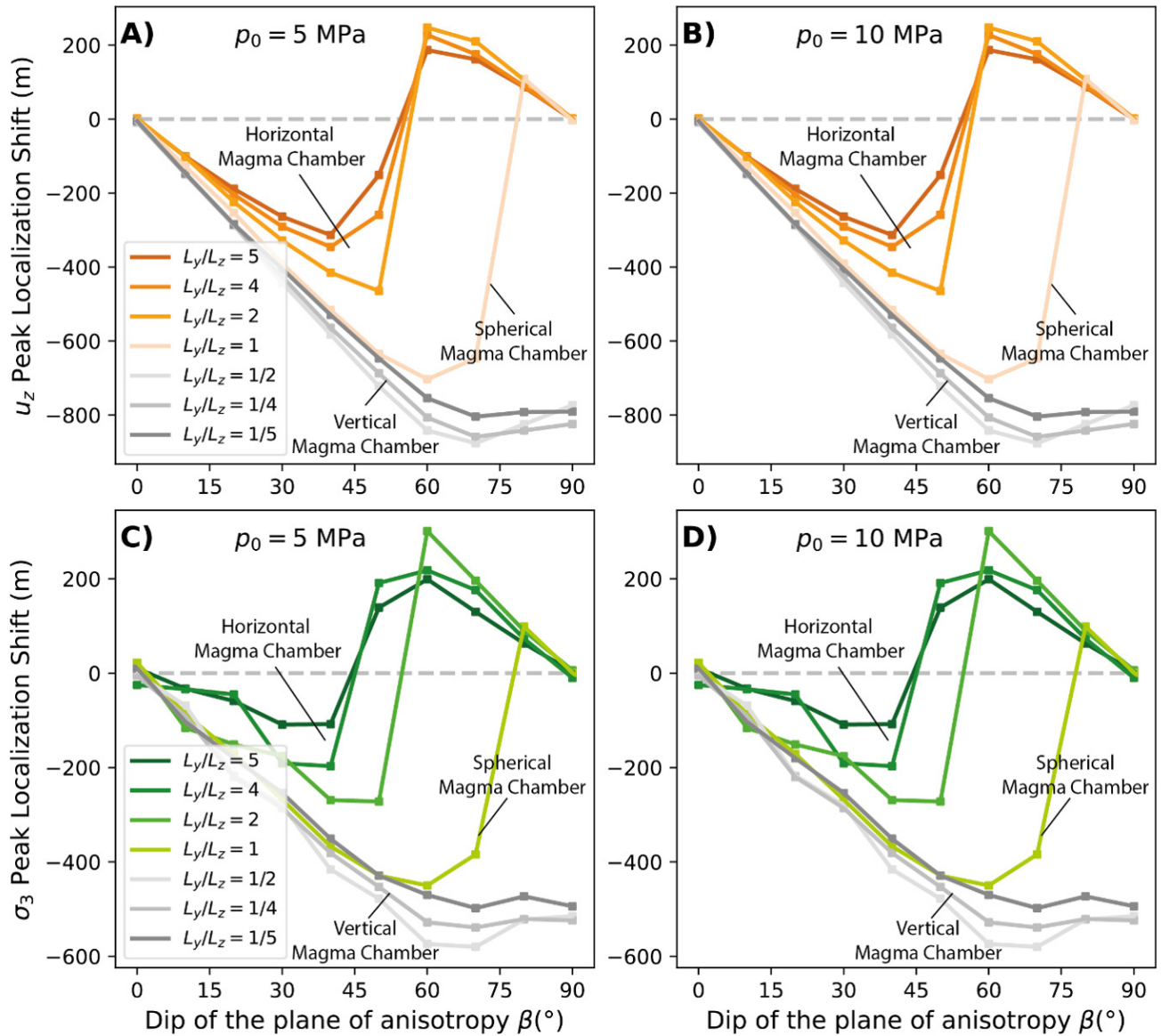


Figure 7: Shift in the localization of maximum vertical displacement (u_z) and tensile stress (σ_3) as a function of the dip of the anisotropy plane (β) and the aspect ratio of the magma chamber (L_y/L_z) for two magmatic overpressure conditions ($p_0 = 5$ MPa and $p_0 = 10$ MPa). Subplots [A] and [B] show the displacement shift, while [C] and [D] illustrate the tensile stress shift. Negative values indicate up-dip migration of the peak, whereas positive values denote down-dip displacement. Seven curves are shown, corresponding to aspect ratios $L_y/L_z = 5, 4$, and 2 (horizontally elongated chambers, orange tones), $L_y/L_z = 1$ (spherical chamber, intermediate tone), and $L_y/L_z = 1/2, 1/4$, and $1/5$ (vertically elongated chambers, grey tones). Category labels within each subplot identify the corresponding groups of curves.

(i.e. $\mathbf{u} \cdot \mathbf{v} \approx 0$ and $\mathbf{v} \cdot \mathbf{n} \approx 0$), as arrester when the fracture plane was normal and propagation occurred parallel to the anisotropic plane ($\mathbf{u} \cdot \mathbf{n} \approx 0$ and $\mathbf{v} \cdot \mathbf{n} \approx 1$), and as short-transverse when both the fracture plane and the propagation direction were parallel to the anisotropic plane ($\mathbf{u} \cdot \mathbf{n} \approx 1$ and $\mathbf{v} \cdot \mathbf{n} \approx 0$). These dot product conditions were calculated from the numerical output of principal stress orientations, and a tolerance of ± 0.1 was used to account for angular deviations due to local stress perturbations.

According to the numerical results, dyke propagation from the pressurized magma chamber is strongly influenced by the

dip of the anisotropic plane. Figure 9 presents the results for models with anisotropy dipping by 30° and 60° , respectively. The figure also highlights the regions of tensile stress concentrations, which are interpreted as the most favourable zones for fracture nucleation and the subsequent propagation of magma-driven fractures (dykes, inclined sheets or sills). Additionally, the trajectories of the maximum principal stress (σ_1) are shown, indicating the potential paths that dykes may follow during their ascent, as they are inferred from the static stress field and do not represent a dynamic simulation of fracture propagation. In both models, dyke propagation is pre-

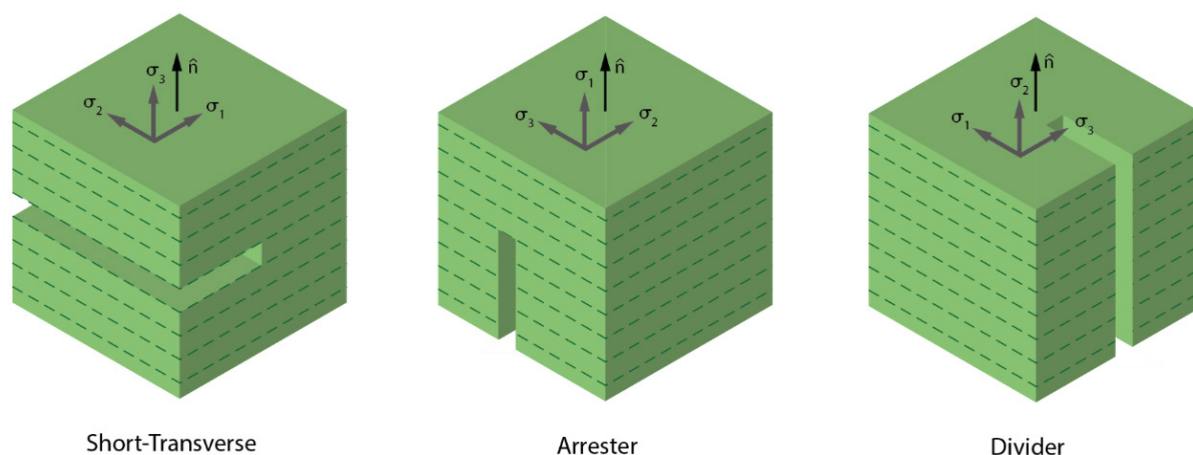


Figure 8: Schematic representation of the three principal fracture orientations with respect to planes of anisotropy: Short-transverse, Arrester, and Divider [Modified from Chong et al. 1987; Forbes Inskip et al. 2018].

dominantly governed by the arrester fracture mode (in blue). In the 30° model, this predominance occurs along the down-dip direction, particularly in the roof of the magma chamber, while in the 60° model it is more pronounced in lateral zones, adjacent to the chamber. In contrast, dykes propagate toward the surface as short-transverse or divider fractures primarily occur along the up-dip and down-dip direction of the chamber's central axis. Notably, short-transverse dykes may undergo abrupt rotation into the divider mode as they approach the free surface. In all cases, stress rotation may produce dyke arrest.

In Figure 9A, Dyke 1 initiates with a downward trajectory from the magma chamber, oriented obliquely with respect to the anisotropy plane. As it propagates, it progressively rotates, eventually aligning parallel to the anisotropy and adopting a short-transverse fracture mode. Near the surface, it transitions into a divider orientation. Dyke 2, which also originates with a downward trajectory, reorients upward during propagation. It initially crosses the anisotropy plane obliquely, then propagates perpendicularly to it, consistent with an arrester mode, before shifting to a short-transverse configuration near the surface. In Figure 9B, among the three simulated intrusions, Dyke 3 propagates laterally as a sill, initially at an oblique angle and subsequently perpendicular to the anisotropy plane, maintaining an arrester fracture mode throughout. In contrast, Dykes 4 and 5 initially rotate to align perpendicularly to the anisotropy but subsequently undergo abrupt reorientations: Dyke 4 transitions into a divider fracture, while Dyke 5 evolves into a short-transverse fracture. It is important to note that, in natural settings, rocks are not only anisotropic but also heterogeneous. Spatial variations in elastic properties, even over short distances, can further contribute to local stress field rotations, potentially influencing fracture trajectories and dyke propagation paths.

4.2 Implications for the interpretation of precursor signals

The observed stress deflections and surface displacement asymmetries have direct implications for interpreting geo-

physical monitoring data during volcanic unrest. Lateral offsets in zones of maximum tensile stress may account for asymmetric uplift patterns commonly detected by InSAR or GNSS, without requiring eccentric or irregularly shaped magma chamber geometries. In our models, such asymmetries arise most clearly in cases where the anisotropy plane is oblique to the vertical axis ($\beta = 30^\circ$ or 60°). These configurations break radial symmetry and cause both the peak tensile stress and the maximum vertical displacement to shift laterally from the reservoir axis, producing tilted or asymmetric uplift profiles. In contrast, horizontal ($\beta = 0^\circ$) and vertical ($\beta = 90^\circ$) anisotropy preserve axisymmetry and result in symmetric stress and displacement distributions. The oblique anisotropy cases therefore provide a plausible mechanical explanation for observed asymmetric deformation signals, even in volcanic systems with otherwise simple geometry. For example, horizontally anisotropic crust produces enhanced and laterally extended vertical displacement due to lower vertical stiffness, potentially explaining anomalously large surface deformation for relatively modest overpressures. Conversely, vertical anisotropy leads to reduced vertical displacements, possibly concealing inflation signals and contributing to underestimation of magmatic activity in geodetic inversions.

Elastic anisotropy also causes rotations in the orientation of principal stresses at depth, which may influence where and how fractures initiate. These stress rotations can guide dyke propagation along curved or segmented paths, resulting in seismicity patterns that deviate from the expected radial or vertical trajectories around a magma reservoir. Such effects may contribute to swarms of volcano-tectonic earthquakes that migrate laterally, bifurcate, or appear offset from the inflation source. In anisotropic crust, dyke segments may propagate along directions that are mechanically favourable but geometrically oblique, complicating the interpretation of seismic source mechanisms and their relationship to surface deformation.

Field-based and numerical studies at Piton de la Fournaise (Réunion Island) and Tungurahua (Ecuador) volcanoes

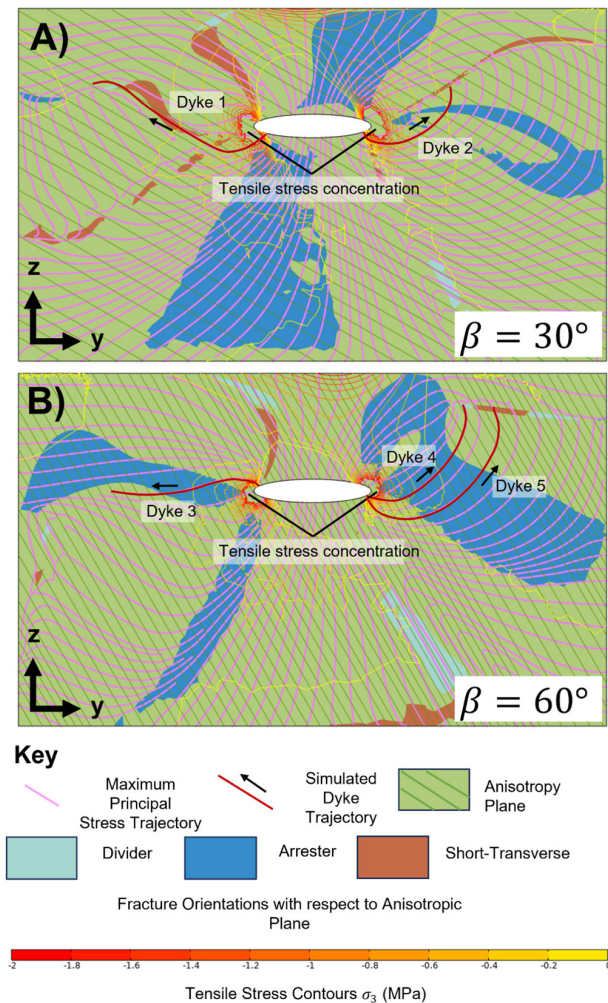


Figure 9: Simulated trajectories of dykes from the magma chamber in an anisotropic medium. Zones where fracture propagation is classified as Short-transverse, Arrester or Divider are indicated. Note that the depicted dyke trajectories are inferred from the static stress field and do not represent a dynamic simulation of fracture propagation.

have highlighted similar challenges in interpreting asymmetric ground deformation. [Letourneur et al. \[2008\]](#) showed that neglecting crustal heterogeneity leads to underestimation of dyke dips and misrepresentation of displacement asymmetry. More recently, [Hickey et al. \[2020\]](#) demonstrated that lateral variations in Young's modulus, related to altered or fractured rock within the edifice, explain pre-eruptive deformation asymmetries at Tungurahua without invoking complex source geometries. Additional numerical modelling by [Russo and Giberti \[2004\]](#) further supports these findings, showing that asymmetric deformation can result from the presence of a lateral rigid body beneath Vesuvius (Italy), emphasizing how elastic heterogeneity alone can distort uplift profiles even when the magmatic source is symmetric. Similarly, at Laguna del Maule (Chile-Argentina border), [Le Mével et al. \[2016\]](#) documented uplift centred southwest of the caldera, which they attributed to the position of an inflating source. While not explicitly linked to crustal structure, such laterally displaced deforma-

tion patterns may also be influenced by variations in host rock stiffness, an effect our TI models replicate even with simple source geometries. These results reinforce the idea that directional stiffness contrasts, whether due to internal heterogeneity or anisotropy, play a first-order role in shaping precursor signals and should be integrated into deformation modelling frameworks.

Remote sensing studies have documented asymmetric deformation structures at several volcanoes. [Kerle and van Wyk de Vries \[2001\]](#) used high-resolution DEMs and radar imagery to map gravitational scarps, grabens, and thrust-fronted slump structures on the southern flank of Casita volcano (Nicaragua), an edifice with inherently anisotropic mechanical stratigraphy due to alternating lava and tephra layers cut by hydrothermal alteration zones, attributing their asymmetric distribution solely to gravitational spreading. [Biggs et al. \[2009\]](#) detected surface grabens and asymmetric uplift patterns above a dyke intrusion in Tanzania using InSAR, noting the influence of pre-existing mechanical heterogeneities on fracture geometry but without invoking directional stiffness contrasts. At Mt. Etna (Italy), [Lundgren et al. \[2003\]](#) documented an asymmetric deformation field that required multiple source components to fit InSAR data, explicitly acknowledging the limitations of assuming a homogeneous isotropic half-space in a structurally complex stratovolcano. Similarly, [Bathke et al. \[2011\]](#) found that the inferred deformation source at Llaima volcano (Chile) was laterally offset from the summit crater, a pattern consistent with the displacement shifts predicted by our oblique anisotropy models yet modelled within an isotropic framework due to insufficient constraints on subsurface mechanical properties. Finally, [Richter and Froger \[2020\]](#) report that high-resolution InSAR interferograms at Piton de la Fournaise (Réunion Island) revealed previously unmapped structural lineaments confirmed only by subsequent fieldwork, illustrating how the mechanical fabric of a layered volcanic edifice can control fracture geometry in ways not yet integrated into deformation models. Laboratory analogue experiments by [Abdelmalak et al. \[2012\]](#) further demonstrated that asymmetric surface deformation patterns above dyke intrusions are sensitive to mechanical discontinuities in the host medium and explicitly identified the role of mechanical layering on fracture paths as a key open question. In none of these cases was mechanical anisotropy invoked as a controlling factor for the observed structural geometry. This is noteworthy given that anisotropy may reside not only in the volcanic edifice itself, but also in the underlying basement rocks. Our TI models suggest this represents a significant interpretive gap: the stress rotations and displacement asymmetries we document offer a mechanically consistent and computationally efficient explanation for such structures and warrant systematic testing against existing remote sensing datasets.

Accounting for these effects is essential for improving unrest interpretation and hazard assessment. In deformation source modelling, incorporating anisotropic stiffness can reduce misfits and misallocations by better capturing the shape and asymmetry of observed displacement fields. In seismic monitoring, recognizing anisotropy-driven stress rotations may help anticipate the direction of dyke migration or the emergence of

off-axis fracture zones. This highlights the need to consider crustal anisotropy as a potential source of deformation asymmetry before attributing such patterns to complex source geometries. This is especially relevant in stratovolcanoes and calderas where host rocks may exhibit strong structural fabrics, such as layering, foliation, bedding, or aligned fracture sets, that influence mechanical behaviour. Ultimately, integrating crustal anisotropy into stress and deformation models offers a path toward more accurate forecasting of failure locations, fracture trajectories, and eruptive potential.

4.3 Limitations and future perspectives

While TI models offer a computationally efficient framework for exploring mechanical consequences of crustal anisotropy, they also involve important simplifications. The assumption of homogeneous anisotropy orientation and magnitude throughout the crust is one of them, which contrasts with the heterogeneous and structurally complex nature of real volcanic systems. Variations in anisotropy direction, strength, and scale, due to faults, alteration zones, or inherited stratigraphy, may produce additional stress perturbations and fracture path complexities not captured by the models presented here. Moreover, our simulations do not include material failure or nonlinear fracture propagation, and they assume linear elasticity throughout. While this allows for isolating the first-order effects of stiffness anisotropy on stress distributions and deformation patterns, it limits the predictive capacity for processes such as dyke arrest, branching, or interaction with heterogeneous zones. Similarly, we do not account for time-dependent effects (e.g. viscoelasticity or fluid pressure evolution) that may modulate both stress accumulation and magma propagation paths during unrest. Another limitation is the idealized chamber geometry and pressurization. The models use simple geometry sources with uniform overpressure, whereas natural reservoirs may display complex shapes, evolving pressurization histories, and feedback with host rock deformation. These factors could interact nonlinearly with crustal anisotropy, reinforcing or dampening the effects we document. The influence of the degree of anisotropy on stress and deformation patterns was not explored here, as a single value of $R = 3$ was used throughout. Sensitivity analyses varying R across a wider range of geologically plausible values represent an important avenue for future work. The assumption of equal Poisson's ratios ($\nu_1 = \nu_2 = 0.3$) simplifies the compliance tensor by eliminating differential lateral coupling between the plane of isotropy and the perpendicular direction. Models with independent Poisson's ratios represent an important avenue for future work to assess whether this simplification influences the magnitude or spatial extent of the subsidence observed in the vertically elongated chamber configurations.

Despite these limitations, our findings highlight the value of TI models as a conceptual and practical tool for probing how directional stiffness may bias stress fields and intrusion trajectories. A particularly tractable avenue in this regard is the direct comparison of TI model predictions against published geodetic and remote sensing datasets from well-documented cases of asymmetric volcanic deformation, including Laguna

del Maule [Le Mével et al. 2016], Llaima [Bathke et al. 2011], Mt. Etna [Lundgren et al. 2003], and Piton de la Fournaise [Richter and Froger 2020], among others. In these volcanic systems, re-inversion of existing data using TI host rock properties could test whether crustal anisotropy reduces source model non-uniqueness and improves fit to observed displacement fields and deformation structures. Such comparisons require combining the TI framework developed here with site-specific source geometries, realistic topography, and field constraints on anisotropy orientation, and represent a natural extension of this work.

5 CONCLUSIONS

This study examined the mechanical implications of transversely isotropic crustal properties for stress and deformation around pressurized magma chambers. Using a suite of finite element models with varying anisotropy orientations and chamber geometries, we explored how directional stiffness alters the distribution and orientation of tensile stresses, the shape of uplift signals, and the potential magma propagation paths. The key findings are summarized as follows:

- Elastic anisotropy generates a complex and asymmetric stress field around the magmatic chamber. At depth, tensile stresses show irregular distributions and the orientation of principal stresses may rotate locally. Toward the surface, the tensile stress field is elongated and shifted laterally in the direction of anisotropy.
- Magmatic overpressure modifies the magnitude of tensile stresses and vertical displacements at the surface but does not affect the location of their peaks, which depend solely on the geometry of the magma chamber and the anisotropic properties of the surrounding rock. The peaks of maximum vertical displacement and tensile stress do not coincide at the surface, highlighting the influence of elastic anisotropy on the distribution of stress and deformation.
- Surface subsidence is observed only in anisotropic media when the magma chamber has a vertically elongated ellipsoidal shape, whereas all other chamber geometries produce surface uplift.
- The model shows stress rotations of up to 90° , consistent with the orientation of dykes observed in highly anisotropic rocks.
- Transversely isotropic models reproduce key stress and deformation patterns characteristic of mechanically layered media, including lateral stress shifts, stress rotations, and asymmetric surface deformation. They offer a geologically realistic and computationally efficient alternative for representing crustal anisotropy when detailed structural information is unavailable.

AUTHOR CONTRIBUTIONS

MC and JC led the conceptualization of the study and contributed to the development of methodology, software implementation, model validation, formal analysis, investigation,

and preparation of the original draft. JC configured meshing parameters and generated figures from COMSOL outputs. VR supported the study through contributions to methodology and software development. JB contributed through critical review and editing of the manuscript and provided overall supervision of the research.

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DATA AVAILABILITY

The intermediate data files and exported data products from COMSOL simulations used to generate the figures within this study are no longer accessible to the authors, but all details necessary to reproduce them are provided in the Methodology, which fully specifies the governing equations, elastic parameters, model domain ($5 \times 5 \times 3$ km), chamber geometries, and boundary conditions.

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