Igneous sills as a record of horizontal shortening: The San Rafael Sub-Volcanic Field, Utah

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Abstract

Igneous sills can facilitate significant lateral magma transport in the crust, therefore it is important to constrain controls on their formation and propagation. Close spatial association between sills and dikes in layered (sedimentary) host rocks has led to a number of sill emplacement mechanisms that involve stress rotation related to layering; from horizontal extension and dike emplacement, to horizontal compression and sill emplacement. Here we use field observations in the San Rafael subvolcanic field (Utah, USA), on the Colorado Plateau, supported by mechanical modelling, to show that layering is not the dominant control in all cases of sill formation. We found no compelling evidence of large sills fed by dikes; all observed cases show that either dikes cut sills, or vice versa. Local sill contacts activate and follow host layer interfaces, but regionally, sills cut the stratigraphy at a low angle. The sills cut and are cut by reverse faults (1-3 m

displacement) and related fractures that accommodate horizontal shortening. Minor sill networks resemble extension vein meshes, and indicate that horizontal and inclined geometries were formed during coaxial horizontal shortening and vertical thickening. Although sills elsewhere may be related to mechanical layering during tectonic quiescence, our mechanical models show that the observed SRSVF geometries are favoured in the upper crust during mild horizontal shortening. We propose that sill geometry provides an indication of regional stress states during emplacement, and are not all sill geometry is a response to bedding. Constraining sill geometry may therefore present a useful tool in plate tectonic studies.

1. Introduction

Igneous sill complexes represent a significant volumetric contribution to upper crustal magma systems (e.g., Planke et al., 2005; Muirhead et al., 2011), and they can play an important role in basin development, petroleum system maturation, and greenhouse gas generation (e.g., Svensen et al., 2004). Although vertical igneous dikes are typically assumed as being the dominant subvolcanic supply route for effusive volcanism (e.g., Ebinger et al., 2008), recent studies have shown that sills can also act as an important regional transport network (e.g., Galland et al., 2007; Airoldi et al., 2011; Muirhead et al., 2011; Airoldi et al., 2016; Magee et al., 2016). Dikes are commonly inferred to represent magma-filled extension mode (mode I) cracks that accommodate crustal extension, with the dike plane forming parallel to the plane of minimum normal stress: the plane containing σ_1-σ_2 (in this paper stresses are reckoned positive when compressive, with $\sigma_1 \geq \sigma_2 \geq \sigma_3$). In contrast, sills require a σ_1-σ_2 plane that is approaching horizontal, with σ_3 (near-)vertical. Dikes and sills are commonly found in close spatial association, particularly in sedimentary basins, yet transitions are rarely observed, especially in terms of dikes feeding kilometer-scale sills (i.e., sills that are laterally continuous at the km-scale; see Valentine & Krogh (2006) and Eide et al. (2016) for possible examples of this).

The assumption that vertical dikes feed sills has important implications for emplacement mechanisms in that it requires the σ_1-σ_2 plane to rotate from vertical (dikes) to horizontal (sills). There are a number of models to explain this, including a level of neutral buoyancy (Francis, 1982), and various controls imposed by host mechanical layering (Gudmundsson, 2011; Schofield et al., 2012; Magee et al., 2016). Analogue injection models have not been able to reproduce a dike-fed sill solely as a function of the level of neutral buoyancy: in all cases, the dike ceases ascent, and begins lateral propagation in the vertical plane rather than flattening into a sill (e.g., Lister and Kerr, 1991). Most analogue models achieve a transition to sills using imposed mechanical layering (Kavanagh et al., 2006) in a hydrostatic stress state (i.e. σ_1=σ_2=σ_3), implying that sills are a consequence of intrusion into a sub-horizontally bedded or layered host rock stratigraphy (Galland et al., 2012). Layering is therefore considered the dominant cause of sill emplacement, with sills fed by dikes. It is commonly overlooked, however, that many dike and sill systems are emplaced within regions subjected to a regional tectonic stress, which can contribute to host rock failure, leading to specific geometries relative to the stress state. Models that account for tectonic stress show that it is possible to cause dike-to-sill transitions as a result of an applied horizontal shortening (e.g., Menand et al., 2010; Maccaferri et al., 2011), but these tectonic-origin models have not gained traction; in each case, layering is considered the dominant control, despite the homogenous host setup in both models.

Here we present remote sensing and field characterization of sills from the San Rafael Sub-Volcanic Field (SRSVF, Utah: **Fig. 1**). Contrary to previous interpretations (e.g., Richardson et al., 2015), we find no observable field evidence that the exposed dikes fed sills. Instead, cross-cutting relationships with dikes and tectonic faults suggest regional horizontal shortening during sill (and dike) emplacement. We use a mechanical, poroelastic model to show how tectonic compression, and related distributed low angle structures (i.e., thrust faults and horizontal extension fractures) could promote and facilitate sill formation. Our model for sill emplacement does not require horizontal mechanical layering for sill formation; the main control on sill geometry is tectonic stress, which could operate in tandem with local stress perturbations.

2. Background: observations and models for sill emplacement

2.1 Natural Sills

Many natural sills are described as exploiting stratigraphic contacts. This is demonstrably true at the local scale (i.e. meter- to hundreds-of-meter-scale), particularly in terms of identified sill segments, lobes, or fingers, inferred to represent the early stages of sill propagation before formation of a through-going sheet intrusion (Thomson and Hutton, 2004; Schofield et al., 2012). However, at the regional scale (i.e. kilometer-scale) many sills are shown to gently climb through stratigraphy (e.g., the Great Whin Sill, northern Britain: Francis, 1982; see also Walker, 1993). Such transgressive sills are commonly compared to 3-D seismic interpretations of sills, and field examples of exhumed *saucer-shaped* sills which exhibit a flat inner region, and transgress or cut up through stratigraphy as a series of ramp and flat segments (e.g., the Golden Valley Sill, Karoo Basin, South Africa: Malthe-Sørenssen et al., 2004; Polteau et al., 2008; Schofield et al., 2010). It is important to note however that transgression can result from a number of mechanisms that occur ahead of the propagating sill tip, including: (1) intrusion of magma into pre-existing faults or fractures; (2) intrusion of magma into new fractures or faults induced by the propagating sill (Magee et al., 2015); or (3) intrusion of tectonic faults and/or fractures formed coeval with magmatism. Exposed transgressive segments therefore are not unique to saucer-shaped sills.

2.2. Analogue and Numerical Models of Sill Emplacement

Galland et al. (2015) present a detailed review of analogue modelling of intrusion emplacement hence only a brief synopsis is provided here. Most modelling of horizontal or transgressive sill emplacement involves injecting fluid vertically into the experimental apparatus, either directly from a tube in the apparatus base plate, or via injection into an imposed vertical crack. Most experiments aim to impede vertical dike emplacement to form a sill, either with an experimental set up using two layers of contrasting material stiffness (e.g., Kavanagh et al., 2006), or by using a porous mesh at a particular level, which reduces cohesion within the material (e.g., Galland, 2012). The experiments in both cases use a static apparatus, in that the rigid walls of the box do not exert a tectonic stress on the experiment; the system should involve low deviatoric stress (i.e. approaching a hydrostatic system). It has been inferred that these types of experiment replicate the natural system, in which intrusions are commonly observed within bedded sequences. However, some horizontal intrusions cut vertically-oriented host layering or foliation (e.g., the Traigh Bhan na Sgurra Sill, Isle of Mull, Scotland: Preston, 2006; Holness and Humphreys, 2003), challenging the requirement for sub-horizontal mechanical layering (e.g. bedding).

Models have produced sills from vertical injection into homogenous media (e.g., Wyrick et al., 2014), depending on the apparatus configuration: in cases where the apparatus lateral boundaries were unconfined, dike geometries were most common. Confining the sides of the model (i.e. placing the host rock analogue sand within a box) led to a fluid-pressure-controlled differential stress state, such that the volume introduced by intrusion led to the generation of a horizontal σ_1, causing σ_3 to switch to the axis that is unconfined (i.e. vertical). This mechanism was originally proposed by Anderson (1951), in which forceful injection of magma into the crust, as dikes, would lead to compression in the surrounding host rock, and eventually lead to stress rotation and sill emplacement.

Excess pore fluid pressure (i.e. suprahydrostatic pressure) has been used to explain intrusion at depths greater than ~2 km, as an alternative to rigidity contrasts, or neutral buoyancy (e.g., Gressier et al., 2010). In such cases the pore fluid factor, λ_v (where $\lambda_v = P_f/\sigma_v$ - the pore fluid pressure divided by the vertical stress) is inferred to equal or exceed lithostatic values ($\lambda_v = 1$ or $\lambda_v > 1$ respectively), at very low differential stress. Gressier et al. (2010) applied this model to the Neuquen Basin, Argentina, which represents a Mesozoic rift basin that has been inverted during Aptian-to-present Andean compression (Cobbold & Rossello, 2003); the models did not involve horizontal compression as a function of an applied contraction, and the intrusive sheet orientation followed contrasting rheological properties of host layering, without which extension fractures and sheet intrusions would have no preferred orientation.

Some analogue models use differential stress to simulate tectonic stress conditions. Galland et al. (2007) modelled intrusion in a developing fold and thrust system, simulating a convergent plate margin, in which intrusions formed as inclined sheets along, and parallel to, developing thrust faults. It is important to recognise that although this model involved horizontal shortening, the formation of thrust faults will have relieved stress within the host medium, and the intrusion experiment was probably conducted during low deviatoric stress. Menand et al. (2010) applied horizontal compression, inducing a vertical minimum compressive stress (σ_3), in which a dike to sill transition was achieved. They concluded that because models did not achieve instantaneous rotation from a dike to a sill, the results did not scale to their observations of natural systems and as such, host rock mechanical layering would be required for sill emplacement. The model results are supported by numerical simulations by Maccaferri et al., (2011), who showed that dike to sill transition during shortening would occur vertically over a few kilometers. With these exceptions, sills and dikes in the majority of cases, are inferred to represent periods of low deviatoric stress. This inferred stress state fits with the assumption that dikes and sills will form in the σ_1-σ_2 plane, and has led to most sheet intrusions being treated as extension fractures, and therefore mechanically equivalent to joints.

3. The San Rafael Sub-Volcanic Field, Utah, USA

The SRSVF is located on the western margin of the Colorado Plateau; about 120 km from the Basin and Range Province (**Fig. 1**). The area is host to several Laramide-age folds with the SRSVF lying between the Waterpocket monocline and the San Rafael Swell (**Figs 1,2**). The Colorado Plateau has seen little reorganization since around ~8 Ma (Burchfiel et al., 1992; Faulds et al., 2008), and the SRSVF is generally considered to have been tectonically inactive since Laramide shortening. The SRSVF comprises about 200 dikes, sills, and volcanic breccia bodies, which were emplaced into lithified Jurassic sediments between 3.7 to 4.6 Ma, contemporaneously with mafic volcanism along the nearby margin of the western Colorado Plateau (**Fig. 1**). The SRSVF crops out over an area of about 1200 km^2, and occupies an observable elevation range of ~500 m, emplaced within the upper 1 km of the crust (Gartner, 1986). The intruded country rocks at outcrop are Middle Jurassic strata of the San Rafael Group, consisting of the Carmel Formation (limestones, sandstones, siltstones and mudstones), Entrada Sandstone, and Curtis (sandstone, siltstones and conglomerates) and Summerville (siltstones, mudstone and fine-sandstones) Formations, which were deposited in shallow/near-shore marine, paralic, and aeolian environments (Gilluly, 1927; Delaney and Gartner, 1995). The San Rafael Group represents deposition into a basin >100 million km^2, with remnants covering most of the Colorado Plateau. The sills are in places composite (basalt-syenite in composition; Gilluly, 1927), with the mafic rock similar in composition to dykes in the SRSVF (Gilluly, 1927; Williams, 1983).

Intrusions in the SRSVF were mapped via remote sensing analysis of high-resolution aerial imagery (~60 cm pixel resolution; National Agricultural Imagery Program for Emery, Sevier, and Wayne

Counties), and 1 m and 10 m resolution topographic data sets. Remote sensing was supported by existing geological maps (Doelling, 2004) and by field characterisation during this study. Dikes were identified by colour contrast in aerial imagery and manually picked remotely in ArcGIS™ for spatial distributions and dike segment strikes (**Figs 2 and 3**). Delaney and Gartner (1997) provide a very detailed analysis of dike geometry; only a short account is provided here. Sill top surfaces were picked where possible, using the top contact between the sill and host rock evident in aerial images; lines representing those top contacts were draped onto the digital elevation models to provide constraints on regional-scale sill geometry (**Fig. 4**). Idealized surfaces were projected through line segments for sill top contacts, from which plane attitudes were derived (**Fig. 4**).

3.1 Dikes

3.1.1 Observations

Dikes in the SRSVF comprise about 2200 observed segments within the Jurassic strata (**Figures 1-3**). The segments are stepped in plan and section view but no clear systematic en echelon left or right stepping is observed. Dike segments show a range of strikes, dominantly between NW-SE and NNE-SSW, with the mean and modal strike of segments being NNW-SSE (**Fig. 2B**). Dikes commonly intersect at a low angle (**Figs 2C,D** and **3B,D,E**), with the acute bisector oriented NNW-SSE to N-S (e.g., **Fig. 2C,D**). Dikes generally dip steeply (>80°) to the east or west, and no preferential dip direction was noted (see also Delaney and Gartner, 1997). Many dike margins preserve breccia of the country rock, which appear to be sourced from the adjacent wall rock, rather than from other parts of the stratigraphy.

Dike segments show a range of tip geometries, from tapered to blunt (**Fig. 3C**). In some cases these steps appear to show a close spatial relationship with the sedimentary layering (e.g., **Fig. 3C**), although this is not always the case (**Fig. 3B**). In other cases, dikes show minor deflection across host layers (**Fig. 3A**). We find no clear instances of dikes transitioning into sills, either at the local (m-scale), or regional (km-scale) scale. Where dikes and sills are observed together, dikes cut sills (e.g., **Figs 5 and 6C,D**) or sills cut dikes (e.g., **Fig 7A,D**).

3.1.2 Summary and Interpretations

Dikes observed at outcrop in the SRSVF are interpreted to represent the segmented peripheral extremity of connected dikes at depth. The vertical and horizontal stepping shows no preferential orientation and we infer that the stepping represents intrusion of fracture segments that formed ahead of the main dike tip, similar to the propagation and linkage of segmented faults and veins in layered materials (e.g., Crider and Peacock, 2004).

The range in dike strikes can be interpreted in three ways, which are not mutually exclusive: (1) reactivation of existing joints; (2) rotation of the principal stress axes; and/or (3) intrusion during tectonic extension. Models 1 and 2 can be rationalised best if considering the intrusions as opening mode fractures. Model 3 does not require that the dikes be opening mode, allowing the through-going dike to accommodate at least a minor component of shear. We infer that the acute angle observed between intersecting dikes could be achieved if the maximum compressive stress (σ_1) and minimum compressive stress (σ_3) are both horizontal; from **Fig. 2C and D**, σ_1 would be oriented N-S, and σ_3 oriented E-W. This extension direction is consistent with the findings of Delaney and Gartner (1997), who associated dike strikes in the SRSVF with the probable reactivation of host rock joint systems in the underlying (Triassic) Glen Canyon Group, and inferring emplacement during low horizontal deviatoric stress (i.e. invoking models 1 and 2 above). Dike emplacement in the SRSVF appears to have been via newly-formed fractures of intact rock, and the acute angle of intersection between dikes suggests elevated differential stress (i.e. greater than four times the tensile strength of the rock; Ramsay and Chester, 2004). Although it is possible that dikes inherited their strikes from the underlying joint systems, it is unclear why low deviatoric stress would not favour intrusion of joint sets at the level of exposure, particularly when low deviatoric stress is considered important in facilitating principal stress rotation to form the sills (Richardson et al., 2015) by way of exploitation of bedding planes.

We infer that variations in rock properties through the host stratigraphy have caused local deflection during dike propagation, but this does not appear to have been sufficient to cause deflection from dikes to sills. It is possible that dikes cutting sills observed at outcrop represent the feeders for sills above that have since been eroded. Alternatively dikes may have fed sills for an initial period, followed by a return to magma flow through vertical conduits; the present study cannot support or preclude either possibility based on field observations alone.

3.2. Sills

3.2.1 Observations

The sills are observed dominantly within the Entrada (sandstones and siltstones), but notable exceptions within the Carmel (siltstones and mudstones) and Summerville (siltstones, mudstone, and fine-sandstones) Formations do occur (**Fig. 1C**). Sills that cut formation boundaries are also observed, such as the Cedar Mountain sills (Entrada and Summerville Formations) and the Last Chance and Little Black Mountain sills (Carmel and Entrada Formations). Sills vary in thickness from <10 cm to about 30 m, and display vertical transgressions as steps along outcrop (**Fig. 7E**), as well as sub-horizontal and inclined sections (**Fig. 7A**). For the purposes of simplifying description, we will refer to sills that are <1 m thick as *thin* sills; those that are >1 m thick are termed *thick* sills.

Thin sills form complex networks of horizontal to inclined (~1° to 25°) sheets that are laterally continuous at the tens-of-metre to hundred-metre scale (**Figs. 7B and 8**). Individual sheets show abrupt steps (**Fig. 8A**) as well as flat and ramp geometries (**Fig. 8A**). Numerous localities show segmented sheets separated by relay zones (*cf.* the bridge structures of Hutton, 2009; **Figs. 7A and 8B,C**). Where dipping sills intersect, chilled contacts are observed, showing intrusion of the younger sill followed solidification of the earlier sill. Where thin sills cut vertical fractures, the fractures are not intruded (e.g., **Figs 7D and 8A**).

Thin sill networks are cut by thick sills (**Figs 7A and 9**). As with instances of thin sills cutting other thin sills (**Fig. 9C**), chilled contacts in the thick sill and/or breccia of the thin sills indicate multiple stages of intrusion (**Fig. 9B**). Thick sills show the same dip range as thin sills (i.e. 0° to 25°: **Figs 4** and **10**). Thick sills show large (>10 m) abrupt vertical steps (**Fig. 7E**) as well as overlaps in which sill tips are more tapered (**Fig. 6E, 7A**). No consistent stepping direction is noted, and no shear sense is inferred. Many thick sills show internal contacts as chilled margins, suggesting that some may represent multiple sills (**Figs 7 and 9**).

Although there are a number of locations where the sills are parallel to bedding (e.g., **Figs 6A,B,C, 7A, and 10A**) sill systems gently climb through the stratigraphy at a low angle: in the south of the SRSVF, sills dip generally northeast or southwest (**Figs 4B, 7, and 10**), and in the northern part of the area, sills dip generally northwest or southeast (**Fig. 4A and 6**), forming an acute angle about the horizontal plane (e.g., **Fig. 10A**). In some areas, and within the Entrada Formation, sandstones around the sills host deformation bands and thrust faults (low angle reverse faults), which in the south of the SRSVF, dip northeast or southwest (**Fig. 7, 10, and 11**). Inclined segments of sills are sub-parallel to reverse faults (**Figs 10E and 11D**), with the long-axis of sill steps oriented sub-parallel to the σ_2 axis derived for the reverse faults (**Fig. 10C inset**). Key localities show that sills have intruded thrust-parallel fractures (**Fig. 11A,B,D,E**), but breccia of sills within fault rock, gypsum veins that display dip-slip reverse motion, and low angle fractures within the sills, show that they are also cut by the faults (**Fig. 11B,F**).

3.1.1 Summary and Interpretations

Sills cross cut the bedding at a low angle. Shallowly-dipping mechanical discontinuities, such as bedding interfaces, faults, fractures, veins, and deformation bands have been intruded, whereas subvertical and vertical structures (fractures, faults, and joints; e.g. **Fig. 7A,B,D**) have not. In several key localities, linked sills are aligned parallel to reverse faults, and are cut by reverse faults (**Fig. 11**), suggesting that they were emplaced during horizontal shortening (**Fig. 10C inset**). The apparent bimodal to quadrimodal sill dip distribution is consistent with horizontal shortening either as a plane strain or during radial horizontal shortening respectively. The inclined sill segments do not occur towards the periphery of large sills, but rather occur at all scales: the SRSVF sills are not saucer shaped. Steps in the sills are sub-vertical, and show direct upward offset of the stratigraphy: the sills in almost all instances are accommodated by a relative vertical uplift of the sill roof (**Figs 7E, 10E, 11D**). Relative uplift of the roof is therefore accommodated by shear offset at inclined sill segments, rather than pure opening mode. Vertical fractures were not intruded, suggesting stress in the horizontal plane exceeds the magmatic pressure on the fracture plane. As shown in **Figure 10E**, some sills show a possible shear offset across the margins, which cannot be associated with an original thrust offset. In cases where the sills are horizontal, vertical opening represents an opening mode displacement (i.e. mode I extension), but for inclined sheets, vertical opening requires a component of shear offset (i.e. mixed mode opening). The opening direction is important as it suggests that the σ_3 axis was vertical in all cases; if all sills – inclined and horizontal – were purely opening mode, the σ_3 axis would have been inferred to rotate. Sills consistently dip in opposite directions – northeast and southwest, or northwest and southeast – respectively in the southern and northern parts of the SRSVF (**Fig. 4**). The σ_3 axis for these areas is inferred to be coaxial, and the mutual cross-cutting relationship between minor sills with opposing dips, suggests that they represent conjugate structures, with the σ_3 axis lying in the obtuse angle (**Fig. 12**). Because vertical structures are not intruded during sill emplacement (note that it is been established in section 3.1 that dikes and sills are cross-cutting) we can infer that the σ_1 and σ_2 axes are horizontal *and* greater than the vertical σ_3; i.e. that σ_2 is not equal to σ_3 and that σ_2 probably exceeded the magma pressure.

Analogue models have shown that dike and sill intrusion can cause deformation of the host rock, and in particular, that inclined sills may cause reverse-sense offsets within the host medium (e.g., Wyrick et al., 2014). These models involve dike and sill emplacement in which the magma pressure and magma volume drives differential stress and failure. It is possible that the reverse faults observed in the SRSVF emanate from intrusions that are not observed, however, it is noted that in cases with exceptional exposure around thick sills (e.g., the Last Chance and Cedar Mountain sills), no reverse faults of this kind are observed. In addition, where observed, the majority of dikes are later than the sills; Delaney and Gartner (1997) showed that in total, dikes in the SRSVF accommodated ~17 m of E-W extension, hence at the time of sill emplacement, volume change related to dikes is inferred to be minor.

4. A NEW MODEL FOR SILL EMPLACEMENT: INTRUSION DURING HORIZONTAL SHORTENING

4.1. Conceptual model

The SRSVF has been associated previously with dike emplacement accommodating ENE-WSW extension during a period of low deviatoric stress (Delaney and Gartner, 1997). Low deviatoric stress is important in their model, because it facilitates a range of dike strikes via activation of joints in the host rock, and during phases of elevated magma pressure (i.e. where the magma pressure exceeds the vertical principal stress) allows intrusion of sills along weak unit interfaces.

We have shown that sills intrude along bedding locally, but predominantly are at a low angle to it. The sills also show mutual cross cutting relationships with reverse faults, and do not intrude vertical fractures, suggesting that sills were emplaced at a time of horizontal shortening. Although sills have been shown previously in contractional settings (e.g. Galland et al., 2007; Tibaldi, 2008; Tibaldi, 2015), our descriptions represent an account of sills formed during tectonic contraction in a region that is generally considered tectonically inactive (Faulds et al., 2008), and adjacent to a major extensional province (the Basin and Range Province). Based on the close relationship between contractional faults and sills, we infer a propagation and inflation model for sill emplacement, similar to that presented by Walker (2016) for the Faroe Islands, on the NE Atlantic passive margin: (1) Regional compression, with a horizontal σ_1-σ_2 plane, and vertical σ_3 axis, resulted in the formation of distributed horizontal extension cracks (mode I) parallel to the σ_1-σ_2 plane, and localized thrust faults at a low angle to it (**Fig. 12a**); (2) magmatic activity resulted in local reactivation of preferentially-oriented pre-existing low-cohesion structures (i.e. those at a low angle to the σ_1-σ_2 plane), such as distributed microfractures, thrust faults, and lithological unit interfaces (**Fig. 12b**); and (3) inflation of individual segments, linked to create a through-going sill (**Fig. 12c-d)**. In this model, intrusions climb through stratigraphy at a low angle to the σ_1-σ_2 plane, but must also propagate laterally along the σ_2 axis. Magmatic inflation of segments is only possible where fractures become linked to the magmatic source (e.g., **Fig. 12b**). Sill propagation and magma flow may therefore be upward or downward, to link fractures in the vertical plane, and may also be horizontal, to link segments laterally. Our model differs from existing models for sill emplacement in two ways: (1) if present, layering serves as a local control only, and critically, is not *necessary* for horizontal intrusion; and (2) sills do not strictly need to form in the σ_1-σ_2 plane, but rather may form oblique to it overall; i.e. sills may not be magma-filled extension fractures, but rather magma-filled extensional-shear (or ‘hybrid’) faults. It is worth noting that under a horizontal compression imposed by tectonic stress, near-vertical faults and fractures may be opened if the magma pressure has sufficient effect to overcome the normal stress on the plane. As no steeply-dipping faults, fractures, or joints in the host rock appear to be intruded (note it is important to distinguish between dilation or intrusion, versus slip along a structure to facilitate linkage of a horizontal sheet), we infer that the maximum and minimum compressive stress within the horizontal plane (SH_{max} and SH_{min} respectively) was greater than the effect of an applied magma pressure.

4.1. Mechanical model

4.1.1. Mechanical model background and parameters

Intrusions are generally viewed as fluid-filled fractures, in which the simplified stress state generally considered for intrusion as an extension fracture, is that the magma pressure (P_m) inside the fracture must exceed the least compressive stress (σ_3) plus the rock tensile strength (T) (Jaeger and Cook, 1979):

$$P_m > \sigma_3 + T \qquad \text{Equation (1)}$$

Extension (mode I) fractures are an end member form of brittle failure that do not involve shear-offset of the fracture walls. Inclined intrusions in the SRSVF show vertical opening, and therefore accommodate a minor component of shear. Extensional and contractional shear failure (mode II) is commonly simplified to the Mohr-Coulomb criterion for failure (**Fig. 13A**) which, taking into account fluid pressure, can be written as:

$$\tau_f = S + (\mu\, \sigma_n') \qquad \text{Equation (2)}$$

where τ_f is the shear stress at failure, S is shear cohesion, μ is the coefficient of internal friction for intact rock, and σ_n' is the effective normal stress (i.e., the normal stress σ, minus the pore fluid pressure P_f: Terzaghi, 1943). In this model, pore fluid pressure and magma pressure (P_f and P_m respectively) would have essentially the same contribution to rock failure, although it should be noted that intrusion of hot magma will be via cracks, whereas pore fluid pressure gains can occur within the host rock primary pore space (Hubbert and Willis, 1957). Equations 1 and 2 assume isotropic poroelasticity, with the pore fluid, or magma, hosted within statistically spherical or equant pore space. Following Carroll (1979) and Chen & Nur (1992), Healy (2012) modelled the effect of changing the shape of the pore space, considered more generally as ellipsoidal cracks, to induce poroelastic anisotropy within the rock volume as observed around faults, within damage zones and fault cores. Healy (2012) showed that, depending on the crack density and the orientation of the cracks relative to the *in situ* stress, significant deviations from the isotropic response to changes in fluid pressure are predicted. Considering the close association of sills with faults in the SRSVF, and because the magma is not transmitted via the primary (intergranular) pore space, but rather via cracks, we apply this model here. For comparison, our models show the role of isotropic crack distribution (i.e. randomly oriented cracks) as well as for anisotropic crack distribution (i.e. parallel cracks). The anisotropic models involve cracks that are oriented in the horizontal plane. The crack density (ρ) is an important factor in the response of the rock to an applied stress (Healy, 2012), hence we show here the results for a crack density of 0.1 (10^5 one-centimeter-radius cracks per m^3) and 0.4 ($4x10^5$ one-centimeter-radius cracks per m^3).

For our illustrative mechanical models, we assume a depth of sill emplacement of ~1 km (see e.g. Richardson et al., 2015), giving a vertical lithostatic load ($\sigma_v = \sigma_3$) of 25 MPa. The tensile strength (T) of the sandstone host rock is overestimated at 10 MPa, and we apply a shear cohesion (S) of 2T = 20 MPa (**Fig. 13**). According to the classical theory of effective stress and brittle failure (e.g., Hubbert & Rubey, 1959; Sibson, 2003), if the differential stress ($\sigma_d = \sigma_1 - \sigma_3$) induced by tectonic loading is ≤ 4T (i.e., 0 MPa < σ_d ≤ 40 MPa), extensional failure of an isotropic rock will be achieved by applying a magmatic fluid pressure (P_m) that overcomes the vertical stress plus the tensile strength of intact rock: here P_m would need to be ≥35 MPa (**Fig. 13B**). In a truly horizontal compressional stress state, this will result in a horizontal extension (mode I) fracture (**Fig. 13**). At higher differential stress (i.e., where σ_d > 40 MPa), fracturing can be achieved only by shear (mode II) failure of the host rock (**Fig. 13A**). However, the Hubbert and Rubey (1959) model assumes a number of important parameters in terms of the response of the rock to changes in stress; in particular, the material compressibility (Poisson's ratio (ν□:the ratio of lateral strain to an applied axial strain) and the bulk modulus (the ratio of pressure increase to a decrease in volume). Additionally, following Nur & Byerlee (1971) the bulk modulus should be considered in terms of the total porous volume (*K*) and the bulk modulus of the solid components (K_s), as the Biot coefficient (α):

$$\alpha = 1 - K/K_s \qquad \text{Equation (3)}$$

Equations 1 and 2, and the classical model of Hubbert and Rubey (1959), assume that the Biot coefficient is 1, hence α is not shown in Equations 1 or 2. Where $\alpha = 1$, an applied fluid pressure of 35 MPa moves the Mohr circle by 35 MPa (i.e., **Fig. 13B**). Decreasing α towards 0 will decrease the effect such that, where $\alpha = 0$, the fluid pressure would have no effect. Poisson's ratio is also often overlooked in the approach to brittle failure. Poisson's ratio for isotropic rocks lies between $0 < \nu \leq 0.5$, ranging from the very compressible to the incompressible, respectively, with an assumed ν = 0.25 commonly applied. The classical model involves perfect incompressibility (i.e. ν = 0.5), and 35 MPa fluid pressure will move the Mohr circle by 35 MPa (**Fig. 13B**); decreasing ν will decrease the effect of an applied fluid pressure because more of the work done is for compression of the pore space itself, without changing the shape of the rock. For many rocks, even those without an obvious fabric, Poisson's ratio departs from the assumed value of $\nu = 0.25$. Well-cemented cohesive sandstones, many limestones, and crystalline granites display $\nu \leq 0.25$. Weaker, less well consolidated sedimentary rocks, coals, shales, and hydrothermally altered crystalline rocks often have $\nu >> 0.25$ (Gercek, 2007).

4.1.1. Mechanical model results

In our models, we apply 25 MPa fluid pressure (i.e. $\lambda_v = 1.0$), which could be considered as the pressure of a pore fluid in the host rock, or the magma pressure within a static (non-propagating) crack. The value for fluid pressure is specifically low, to illustrate the approach to failure only, and is substantially less than the required 35 MPa fluid pressure required for mode I failure in the Hubbert and Rubey (1959) model. The starting value for differential stress is equivalent to 4T for intact rock (i.e. $\sigma_d = 40$ MPa), which is probably lower than the differential stress implied by thrust faults that are interpreted to be coeval with the sills. For comparison, we also performed tests to simulate failure conditions as a function of elevated fluid pressure (**Table 1**).

Applying a fluid pressure increase of 25 MPa in isotropic rocks moves the Mohr circle towards the failure envelope, but maintains a constant differential stress (**Figs 13B and 14**): the effect of fluid pressure is equal in the σ_1 and σ_3 directions. For a compressible rock (e.g. $\nu = 0.11$), the fluid pressure is only sufficient to initiate a re-shear of existing cohesion-less surfaces if the crack density (ρ) is increased. In **Figure 14A** we can see that where $\rho = 0.4$, re-shear is possible on structures that are within an angular range of 11-42° to the σ_1 axis. Less compressible rocks ($\nu = 0.4$) have a significant response to 25 MPa fluid pressure, with re-shear of pre-existing cohesion-less structures possible within an angular range of 12-41° where $\rho = 0.1$, and 0-55° where $\rho = 0.4$ (**Fig. 14B**). The predicted increased dilatancy of the less compressible rock is important in controlling reactivation of existing surfaces, but as anticipated, neither of the *isotropic* poroelastic models predicts intact rock failure. Increasing fluid pressure to intact rock failure highlights the importance of crack density and Poisson's ratio, requiring fluid pressure to exceed the classical model prediction of 35 MPa in all cases. Rocks with low Poisson's ratio and low (isotropic) crack density require significantly greater fluid pressure to cause failure, than rocks with high Poisson's ratio and high crack density (**Table 1; Fig. 14**).

For rocks with patterns of parallel cracks (in this case, horizontal and aligned parallel to σ_1), shear stress and effective normal stress both change within the intact rock, with increases in fluid (magma) pressure (**Fig. 15A,B**): the fluid pressure has greater fracture surface area to act upon normal to σ_3 than there is to counteract σ_1 (and σ_2). This manifests itself in the poroelastic framework as a directionality in the values of α (now a 2nd rank tensor, **Table 1**), leading to an increase in differential stress. **Figure 15A** shows the effective stress change induced by a 25 MPa increase in fluid pressure for a host rock with $\nu = 0.11$. As fluid pressure increases, the differential stress increases, and the stress state is driven towards brittle shear failure (i.e. the shear failure envelope), with re-shear possible on cohesion-less structures within an angular range of 13-40° and 1-52° where $\rho = 0.1$ and $\rho = 0.4$ respectively. Failure of intact rock in this case is achieved at 75 MPa and 34 MPa for crack densities of 0.1 and 0.4 respectively, with the latter falling below the classical model prediction. Failure is predicted within the contractional shear portion of the failure envelope: i.e. a thrust fault, with planes forming at an ideal angle of 25° to the σ_1 axis. In **Figure 15B**, an alternative model is presented for a host rock with $\nu = 0.4$. In this case applying 25 MPa fluid pressure increases differential stress, but the effect is subdued compared to $\nu = 0.11$. Re-shear is possible within an angular range of 7-46° and 0-56° where $\rho = 0.1$ and $\rho = 0.4$ respectively. Intact rock failure is achieved with fluid pressure at 55 MPa and 32 MPa where $\rho = 0.1$ and $\rho = 0.4$ respectively. Notably, the latter falls below the predicted 35 MPa required in the classical model.

Failure occurs within the hybrid portion of the envelope; extensional shear is predicted (Ramsey & Chester, 2004) with planes forming at an ideal angle of 18° to the σ_1 axis (**Fig. 15B**).

4.1.1. Mechanical model summary

Our mechanical model results have two very significant implications for intrusions in general: (1) decreasing the host rock Poisson's ratio leads to a lesser response of the rock, when increasing fluid pressure; and (2) the failure plane, and therefore sills (and dikes), may not be parallel to the σ_1-σ_2 plane. The role of host rock Poisson's ratio in facilitating sill emplacement is important: a larger ratio of vertical ("lateral") strain to horizontal ("axial") strain for a horizontally applied load in a compressional thrust fault regime will promote more dilatancy for magma to occupy. Sills may not necessarily prefer weaker rocks (i.e., rocks of lower brittle strength), but they may preferentially intrude those with elevated values of Poisson's ratio (i.e., those that are more elastically compliant). In addition, even at low fracture densities, anisotropic poroelasticity will promote a deviation from mode I fracture. Our models involve low differential stress initially, but increases in fluid pressure will increase differential stress due to the directional variation in the Biot coefficient (**Table 1**), promoting shear failure. This effect may be particularly important at depth, where tectonically-driven differential stress may exceed our starting value of four times the tensile strength of the rock, and even in the near surface (~1 km) as indicated by the presence of minor thrust faults in the SRSVF.

These simple poroelastic models describe only the *approach to failure*. Brittle failure of rock, by shear or extensional fracturing, includes processes and deformation mechanisms that are not well modelled by poroelasticity, including cataclasis, pore collapse, and the coalescence of microcracks into through-going fractures. However, we maintain that the approach to failure, i.e. the effective stress path followed by the rock mass towards fracture formation or reactivation, is the critical part of the wider process of sill emplacement that we seek to address. We have only explored a few parameters (i.e. crack density, Poisson's ratio, and the Biot coefficient) that contribute to the response of a rock to an applied stress, rather than a full sensitivity analysis for elastic parameters.

5. Discussion

5.1. Is layering the primary control on sill formation in the SRSVF?

Sills in the SRSVF do activate layer interfaces (e.g., **Fig. 10A**), and our model like others before it, shows that stiff materials will fail at lower applied stress than soft materials (Eisenstadt and DePaor, 1987; Ferrill et al., 2016). Sills also appear to intrude pre-existing fractures and thrusts (**Fig. 11**). Material properties, and existing discontinuities can have a strong control on the positioning of intrusions, and their geometry (McCaffrey and Petford, 1997; Schofield et al., 2012). It is important to recognize that this does not mean the layer interface, or existing discontinuity is the cause of the sill in all cases. Dikes in the SRSVF are observed at the same stratigraphic level as the sills - within the same host units - and show cross-cutting relationships indicating dikes and sills are not connected. Perhaps the greatest physical and mechanical property contrast in the observed sequence is that between the sills and the sedimentary host rocks: dikes cut sill contacts without major deflection. Hence other factors must control the formation of the SRSVF sills. Notably, the orientation of dikes, and concurrent emplacement of sills, has been inferred to relate to low deviatoric stress, and activation of existing joint sets in the underlying strata (Delaney and Gartner, 1997); we have not observed any clear instances of sills feeding dikes; vertical fractures are not intruded via sills. Where dikes cross unit interfaces, and formation boundaries, they exhibit minor deflections in dip and/or strike, indicating interaction with the host mechanical stratigraphy (**Fig. 3**). There are a number of well-accepted models for sill emplacement that involve mechanical stratigraphy, including stress barrier configurations (Gudmundsson, 2006), elastic mismatch (Dundurs, 1969) or material toughness variation (Kavanagh et al., 2006), and Cook-Gordon delamination (Cook et al., 1964). In each case, a dike propagates through layering before one of the above mechanisms causes the σ_3 axis to rotate from horizontal to vertical. Each mechanism is strongly dependent on the host rock mechanical variation, including the material toughness and fracture toughness, elasticity, and strength (cohesion) of the interface. Each mechanism is modelled typically in a hydrostatic stress state. Each mechanism is partly dependent on the dike being opening mode, rather than accommodating shear. Although this is likely for a significant proportion of natural dikes and sills, it is unreasonable to assume it in all cases (Walker, 1993). Analogue model results suggest sills require mechanical layering, and this gains some traction from the preponderance of sills in layered sedimentary basin settings. However, if layering is the primary control on sill formation in the SRSVF, then all other parameters being equal, all dikes should rotate when they reach the same material interfaces. It is well know that propagation at an interface may not be possible if the driving pressure is insufficient, or if the principal stress axes are not oriented favourably (e.g., Gudmundsson, 2011). Despite this, the inference that dikes feed sills has become so embedded in the literature, that it is not required for studies to show that material interfaces have low cohesion, or that materials on either side of the interface differ significantly in their properties. Clear examples of dike to sill transitions are generally small scale, with meter-thickness dikes deflecting into sills for a few meters before returning to the original dike geometry (Gudmundsson, 2011). If this mechanism of instantaneous deflection is to operate at larger scales – to feed sills that are laterally continuous for many kilometers – there must be a favourable stress state: either σ_3 is regionally vertical, or intrusion is during a low deviatoric ambient stress state. We suggest that the sill geometry and position within the stratigraphy is therefore an indication of the ambient stress: sills that cut layering may relate to phases of horizontal shortening, whereas strictly layer-parallel sills may reflect the dominance of material anisotropy presumably during periods of low deviatoric regional (tectonic) stress.

In our model, sills in the SRSVF were intruded during horizontal shortening, associated with a vertical σ_3 axis. It is not clear from field observations how the sills are fed from below; whether this is via a complex arrangement of gently transgressive to inclined sheets, or via unobserved dikes. If the latter, a stress rotation is required, though notably this could involve switching the σ_2 and σ_3 axes, with a constant horizontal σ_1 axis. Based on the numerical models by Maccaferri et al., (2011) and the analogue models of Menand et al. (2010), we infer that this rotation would be gradual, occurring vertically over the hundred-meter scale or greater. In either case, we should not expect to see the vertical feeder dike at the same stratigraphic level; rather we would require a larger cross section through the system, in which we would probably observe sills fed by inclined sheets, in turn fed by dikes at depth.

5.2. Mechanisms for shortening during sill emplacement

Our field study and poroelastic models show that tectonic shortening during magmatism could facilitate sill intrusion, with new failure planes predicted to form at a low angle to the horizontal σ_1 axis, and reactivation of fractures at a broad range of angles from the horizontal σ_1 axis. The angular range of these reactivated fractures is dependent on magma pressure having sufficient effect to counter the normal stress on the plane. The models do not *require* stress rotation due to mechanical layering, but would be *aided* by variations in elastic properties through the sequence. By increasing the differential stress, intrusion at low angles to σ_1 is possible even at very low magma overpressures (i.e. approaching $\lambda_v = 1$). We infer here that the sills in the San Rafael Sub-Volcanic Field are representative of large sill complexes in which the primary control for their emplacement is horizontal shortening. This is distinctly at odds with existing conceptual models for sill emplacement, in particular because horizontal shortening is not recorded in association with regional sill complexes. Shortening in the SRSVF is minor, and reverse faults probably account for <1% horizontal shortening at the scale of the study area; our models are elastic only, and although they do not quantify strains, the elastic range would probably involve shortening on the order of <1%. Many studies of sill geometry are based on 3-D seismic data (e.g. Magee et al., 2016), and such minor shortening may not be visible owing to the resolution limits of seismic imaging. However, our model can account for sub-horizontal intrusion into homogenous or even vertically-layered materials, in that the geometry of the sills is predicted to be controlled primarily by the stress state. For instance, the gently-dipping basaltic sills at Loch Scridain on Mull, Scotland, intrude vertically bedded and foliated, metamorphosed sandstones and mudstones, as well as horizontally bedded lavas (Preston, 2006; Holness and Humphreys, 2003). Such examples cannot be explained by stress rotation due to layering, nor neutral buoyancy, as the sills are observed in both the basement and cover sequences. Models for discontinuity reactivation would predict dike emplacement in the vertically foliated basement rocks, and sill emplacement in the horizontally-layered cover. In stark contrast, dikes are present in the cover and basement, and sills gently climb with respect to the paleo-horizontal regardless of the host rock foliation.

The cause of horizontal shortening in the SRSVF remains unclear, and previous studies of the intrusions have considered the region as generally tectonically inactive since the late Cenozoic. The SRSVF is located towards the western margin of the Colorado Plateau, adjacent to the Basin and Range province (**Fig. 1**). The region is host to several Laramide-age folds, which predate intrusion by >30 million years. Laramide folding relates to northeast-southwest shortening in the Colorado Plateau (Davis, 1978). Imbricate thrust faults have been identified previously in the Cedar Mountain area of the San Rafael Swell (north of the SRSVF; **Fig. 1**), but these are associated with Laramide shortening also (Neuhauser, 1988). Cenozoic deformation in the region is dominated by Basin and Range extension, though this is largely outwith the Colorado Plateau, which has seen little structural reorganization since ~8 Ma (Burchfiel et al., 1992; Faulds et al., 2008). The Plateau has been subject to considerable uplift since the Late Cretaceous (Liu and

Gurnis, 2010), with numerous mechanisms proposed as to the cause. The plateau currently stands at ~2 km, but with a notable bowl-shape such that the margins are elevated ~400 m above the plateau interior (Hunt, 1956; van Wijk et al., 2010). Again, the cause of this uplift geometry is debated, but probably relates to edge-driven convection following lithospheric rehydration (van Wijk et al., 2010) and lithospheric down-warping (Levander et al., 2011) particularly during the late Cenozoic. The distribution of Pliocene volcanism, and incision rates in the Grand Canyon, suggest this style of differential uplift has been active since ~ 6 Ma. Late Cenozoic lower crustal delamination and crustal thinning was focused south of the SRSVF, within the Grand Canyon Section of the plateau, which coincides with active normal faults that have accommodated ~100 m Myr^{-1} differential uplift of the plateau relative to the Basin and Range (Lavender et al., 2011). The SRSVF coincides with the margin of a down-welling body at ~200 km depth (**Fig. 1**; Levander et al., 2011). Differential uplift of a region that is host to numerous pre-existing major structures (i.e., the Laramide-age fault systems) could result in a complex stress state and local/regional geometric reactivation. We speculate that such differential uplift could provide a mechanism for upper crustal horizontal compression. In this case, the direction of maximum horizontal shortening would be oriented with respect to the major structures that are reactivated. Sills in the northern SRSVF dip northwest and southeast, perhaps corresponding to reactivation of the NW-dipping San Rafael Swell fault system. In the southern part of the SRSVF, the sills dip northeast and southwest, normal to the crest of Waterpocket monocline. To our knowledge, the examples of thrust faults presented here are the first recorded for this area. It is clear however that further work is needed to relate these structures to specific events.

Horizontal shortening is not typically recorded in association with sill emplacement in rift basins, or passive margins (see Sundvoll et al. (1992) and Walker (2016) for rare exceptions). However, it should be noted that tectonic compression in the sense of a horizontal σ_1 axis, is recorded in rift basins by variably-oriented folds, strike-slip faults that are oblique to basin bounding normal faults, and recent or present day focal mechanisms. For instance, Walker (2016) showed that sills cutting Paleocene lavas in the Faroe Islands were emplaced during horizontal shortening on the Atlantic margin. Although not directly dated, the sills are cut by fault sets dated by Roberts and Walker (2016), which bracket the age of the sills to ~54-46 Ma. Pre-, syn- and post-breakup mild contractional folds are observed along the Atlantic margins, including along the NE Atlantic margin from Ireland, past the Faroes and UK, and through to Norway (e.g., Doré et al., 2008). The timing of tectonic compression along the margin therefore overlaps the timing of sill emplacement in those areas (e.g., Magee et al., 2014). Pre- and syn-break-up contraction may be accounted for by various rift propagation models (e.g., Hey et al., 1980, and references therein). Post break-up shortening on the margin is typically inferred to reflect ridge push effects, or elevated gravitational potential energy induced by the combination of an upstanding continental interior, and the large volume represented by Iceland (e.g., Cloetingh et al., 2008). Syn- to post-break-up conjugate strike-slip faults in the Faroe Islands and Faroe-Shetland basin accommodated crustal extension at a high angle to the developing continental margin, but also a horizontal shortening sub-parallel to the margin (Walker et al., 2011). Dikes that are parallel to those conjugate strike slip faults also record crustal extension, and with minor shear offset across the dike planes; as extensional shear structures, they are interpreted to represent conjugate intrusions, generated by a horizontal σ_1 and σ_3. Walker et al. (2011) and Walker (2016) showed that the horizontal shortening direction and inferred σ_1 axis for conjugate dikes and sills in the Faroe Islands was coaxial: E-W. Dikes recorded N-S horizontal extension (parallel to the inferred σ_3 axis), with a vertical inferred σ_2, and sills record vertical extension (parallel to the inferred σ_3 axis), with a horizontal and N-S inferred σ_2. Focal mechanisms for recent and present day earthquakes in active oceanic and continental rifts, volcanic flank rift systems, and passive margin settings show that stress orientations can be highly variable spatially, and temporally, across or along the rift axis, recording combinations of extensional, contractional and strike-slip events (e.g. Stein et al., 1979; Ebinger et al., 2008; Green et al., 2014; Lin and Okubo, 2016). In summary, we suggest that applying our tectonic shortening model to sills in rift systems - considering sill complexes as a record of the regional stress - will lead to a better understanding of the intrinsically fluctuating nature of stress in such systems.

6. Conclusions

Mutual cross-cutting relationships between thrust faults and igneous sills in the San Rafael Sub-Volcanic Field in Utah, provide evidence for sill emplacement during horizontal shortening in a tectonically inert or extensional system. We infer that horizontal compression due to tectonic shortening may be a requirement for some other regional-scale horizontal intrusions, even in regions otherwise considered dominantly extensional. As a record of the stress state, igneous sills could be used as a tool to constrain regional tectonics, such as phases of compression within basins or along passive continental margins. Mechanical models show that sill emplacement can be aided by the development of oriented microcracks related to the compressional stress state, particularly at a local scale, around pre-existing faults where the high density of existing microcracks will facilitate failure at lower magnitudes of fluid overpressure. Our model for horizontal intrusion does not require host rock mechanical layering, and can be applied to horizontal intrusions within non-layered, or vertically-layered media.

Acknowledgements

The authors would like to thank DONG E&P UK for funding in support of this study. The authors gratefully acknowledge helpful reviews by Nick Schofield and Graham Andrews, and additional commentary from Associate Editor Bernhard Grasemann. Many thanks to Craig Magee and Agust Gudmundsson, as well as two anonymous reviewers, for their extensive and helpful reviews at an earlier stage in preparation of this paper.

Figures

Fig. 1. Location maps for the San Rafael Sub-Volcanic Field in Utah. (**A**) Digital elevation Model for Utah, showing major structural and depositional areas of the Colorado Plateau. Solid black line shows province boundaries. Dashed black line is a region of lower-crustal delamination and crustal thinning detailed in Levander et al. (2011); dashed white line is their outline of a downwelling body at 200 km depth, estimated from body wave tomography. (**B**) Aerial imagery for the San Rafael Sub-Volcanic Field (SRSVF) highlighting location and distribution of intrusive bodies. (**C**) Geological map of the region of interest, showing relative positions of the Northern (N.) and Southern (S.) SRSVF. Cen.: Cenozoic. Cret.: Cretaceous. FM: Formation. Mbr: Member.

Fig. 2. Dike orientations in the SRSVF. (**A**) Hill shaded digital elevation model of the SRSVF showing dikes identified from aerial images. (**B**) Rose plots show dike orientations, separated by geographic location, and combined. Interpreted aerial image of dikes in (**C**) the eastern and (**D**) the western SRSVF showing the acute angular relationship between dike segments.

Fig. 3. Field photographs of dikes hosted in the Entrada Formation, within the SRSVF. (**A**) 1 m thick dike cuts sandstone-siltstone units, and shows minor angular deflection from vertical through the siltstone. (**B**) Segmented dikes show acute angular relationship (~23°) along strike. Segments both cut and abut a thin (10-30 cm thick) mudstone that separates siltstones above and below. (**C**) Dike segment abuts upper contact of a mudstone. Dike appears to be continuous across the mudstone, but shows a pronounced thinning above the contact, and ~1 m lateral offset. (**D**) Steeply-dipping dikes butting sandstones and siltstones. (**D**) Steeply-dipping dike segments show segmentation in plan, and section view. Segment tips correspond to unit boundaries in section view, but no pre-existing discontinuity is noted in plan view.

Fig. 4. Geometric analysis of thick sills in the SRSVF. Hillshaded digital elevation models for (**A**) the northern SRSVF and (**B**) the southern SRSVF. Models show extrapolated elevation data for

sill top contacts. Lower hemisphere stereographic projections show sill top contact polygon attitudes as great circles, and contoured poles to planes for each sill system named in A and B. See text for details.

Fig. 5. Examples of cross-cutting relationships between sills and dikes. (**A**) Segmented dikes cut sills in the southern SRSVF. (**B-C**) Dikes cut sills in the northern SRSVF. Note that the thick sills in B and C are not parallel to bedding. Note that in C, dike segments (outlined with white dashes) cut the sill upper and lower contacts, but appear to abut internal sill contacts. (D-F) “Co-magmatic conduit of Richardson et al., (2015). Dikes within the volcanic breccia body (dark grey) cut thin sills below the main thick sill (light grey) shown in D. Chilled margin surfaces are observed at the same leval as the thick sill, though no direct contact is observed. Black star represents a marker to tie images D, E, and F.

Fig. 6. The Cedar mountain sills, northern SRSVF. (**A**) Photo panorama showing the Lower, Central, and Upper Cedar Mountain sills. (**B-D**) Dike and volcanic breccia body cut the Central Cedar Mountain sill. (**C**) Breccia body is developed along vertical joints in the sill. (**D**) Dike cuts volcanic breccia body, and shows chilled margin contacts with the Central Cedar Mountain sill. (**E**) Central Cedar Mountain sill is segmented across an apparent relay structure. Relay structure is brecciated, and hosts minor (cm-thick) sills that are inclined relative to the main sill.

Fig. 7. Examples of sills in the SRSVF. (**A**) 30 m thick sills in the southern SRSVF, are gently inclined relative to the host stratigraphy (at ~3°) such that the upper sill is observed intersecting the Entrada-Carmel Formation boundary ~700 m to the SW of the photograph. Note right hand edge of A is oriented N-S; black star indicates a marker point linking A and E. Breached relay structures (*cf.* broken bridges, e.g. Hutton, 2009), which record early sill segments, consistently strike NW-SE. Star shows reference position for view shown in E. (**B**) Thin sills (10 cm to 1 m thick) occur in close proximity to thick sills. Some thin sills are parallel to deformation bands, whereas some are horizontal. (**C**) Intrusions range in dip, from horizontal to ~60°; here, steeply

inclined sheets are cut and offset by shallowly-dipping sills. (**D**) Thin and thick sills cut vertical dike. (**E**) Lower thick sill shows abrupt vertical steps along exposure, whereas upper thick sill does not, suggesting the lower sill may predate the upper. Note the position of the lower sill base contact relative to the Entrada-Carmel Formation boundary.

Fig. 8. Examples of thin sills in the SRSVF. (**A**) Multiple sill network comprising cm-thick sills. Sills are generally bedding parallel but display local ramp sections that dip NE and SW. (**B**) Some thin sills are segmented, and separated by apparent relay structures that are intruded by inclined sheets. (**C**) Relay structures occur at a range of scales, up to ~2 m separation. Individual sills are stacked to form a multiple sill. (**D**) Locally, sills cross-cut each other, indicating staged intrusion.

Fig. 9. Relationship between thin and thick sills. (**A**) Thin sills are stacked to form multiple sills. (**B**) Locally, thick sill lobes cut thin sill contacts, forming breccia of thin sills. Long-dash line marks the contact between the thick sill and thin sills. Short-dash line marks the boundary between dominantly intact thin sills, and brecciated thin sills. (**C**) The volume of thick sills appears to be accommodated by folding of the country rock, including the thin sills.

Fig. 10. Gently-dipping sills in the southern SRSVF. (**A**) Thick sills are locally parallel to host bedding, but otherwise gently climb through the stratigraphy. Sills dips are dominantly NE and SW, and form an acute angle about the horizontal plane. (**B**) Thick and thin sills show NE and SW dips. (**C-D**) Thin sills range in attitude from horizontal, to inclined (~20-25°). Lower hemisphere stereographic projection shows deformation bands and thrusts in the southern SRSVF. Sills are locally parallel to (**D**) bedding, and (**E**) deformation bands and thrust faults. Thick arrows in E show sill opening direction. Lower hemisphere stereographic projection shows poles to planes for thrust and deformation band data collected in the southern SRSVF, at localities shown in **Fig. 10** and **Fig. 11**. Deformation band data is contoured in grey.

Fig. 11. Relationship between sills and reverse faults. (**A-C**) Sills cut and are cut by a thrust fault. (**B**) A multiple sill is cut by an E-dipping thrust. In the upper right of the image, a separate thin

sill is observed along the fault plane, inferred as representing post fault intrusion. (**C**) View from the other side of the crag shown in A and B. Minor fractures parallel to the thrust are observed in the multiple sill. Breccia of the sill is developed along the main thrust, and along minor faults that are sub-parallel to it. (**D-F**) A thick sill that shows a ramp-flat-ramp geometry, parallel to reverse faults (dipping 25-45° NE) within the country rock. Thick arrows in D show sill opening direction. (**E-F**) Inclined sills appear to have intruded parallel to thrusts, suggesting they reactivate existing structure, but are also locally cut by thrusts. (**F**) Bedding-parallel sill is dragged into a reverse fault. The sill hosts gypsum-mineralized fractures. Fault rock along the reverse fault comprises breccia of the country rock and the sill. (G) Multiple sill appears to be offset across a thrust fault (dipping ~10°E). Note that the country rock in contact with the sill displays thermal alteration, with the exception of the zone along the thrust plane. (H) Along the fault plane, the sill displays mineralized dip slip fault surfaces, and a 5-10 cm thick zone of altered fault rock.

Fig. 12. Conceptual model for sill emplacement during compression. (**A**) Horizontal shortening produces a fault and fracture system comprising isolated inclined and flat segments. (**B**) Existing fractures are in-filled and inflated by magma and propagate as extension and extensional shear veins. (**C**) Adjacent sheets link to form a through-going sill. New fractures and faults continue to form during on-going compression. (**D**) Minor sills are abandoned in favour of the more thermally efficient main sill. Note that, as this process may operate across scales, the illustrated box widths may represent centimetres to hundreds of metres, provided there is fault/fracture connectivity in or out of the page. We purposefully do not show the model *feeder* system, as this is not observed in the field.

Fig. 13. Mohr diagrams depicting the poroelastic response to isotropic pores and oriented cracks, and to different values of host rock Poisson's ratio. (A) Example Mohr diagram (shear stress, τ, against normal stress, σ) showing the composite failure envelope for intact rock (solid

black line) plus the re-shear condition for a cohesion-less fault (dashed black line), and critical stress circles for the three mesoscopic modes of failure. θ represents the angle between the failure plane and the σ_1 axis; θ_s denotes the angular range where reactivation is possible; μ is the coefficient of friction; ϕ_i is the angle of internal friction for intact rock; ϕ_s is the angle of internal friction for re-shear of a cohesion-less fault. Values are idealised based on the Berea sandstone (Healy, 2012). (B) The classical model for the application of fluid pressure (P_f) (after Hubbert and Rubey, 1959). The model involves idealised values for rock compressibility (i.e. Poisson's ratio) and the Biot coefficient, so that the applied fluid pressure has a 1:1 influence on the normal stress.

Fig. 14. Mohr diagrams illustrating the poroelastic effect of crack density, and Poisson's ratio at (A) 0.11, and (B) 0.4. Cracks in the model are randomly oriented (i.e. isotropic). Black circles are the normal stress before fluid pressure is applied; grey short-dashed circle shows the effect of an increase of 25 MPa fluid pressure where ρ = 0.1; grey long dashed lines circle shows the effect of 25 MPa fluid pressure where ρ = 0.4; red circle shows failure condition.

Fig. 15. Mohr diagrams illustrating the poroelastic effect of anisotropic crack density, and Poisson's ratio at (A) 0.11, and (B) 0.4. Cracks in the models are horizontal. Black circles are the normal stress before fluid pressure is applied; grey short-dashed circle shows the effect of +25 MPa fluid pressure where ρ = 0.1; grey long dashed lines circle shows the effect of +25 MPa fluid pressure where ρ = 0.4; red circle shows failure condition. (C) Photograph shows example of sills in the southern SRSVF, highlighting the range of sill attitudes observed in the field. Notably the extension direction is ubiquitously vertical, parallel to our inferred σ3 axis.

TABLES

Table 1. Mechanical model parameters and results corresponding to Figures 13, 14, and 15. ν, Poisson's ratio; ρ, crack density; α, Biot coefficient; σD, differential stress; θs, reshear angle; P, fluid pressure; θ, failure plane of intact rock; λv, pore fluid factor.

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