Dike propagation, and magma flow in a glassy rhyolite dike: a structural and kinematic analysis

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Abstract

Exhumed magma conduits provide important evidence of the development and evolution of subvolcanic plumbing systems. We use a 5-14 m thick flow-banded rhyolite dike in Arran (Scotland) to present the first reconstruction of the directions and styles of initial propagation and subsequent magma flow, based on mesoscale kinematic indicators. The dike has concave-inward dike margin segments with plumose-like structures that record vertical and horizontal propagation of lobes, which inflated and linked to form a through-going sheet. Devitrified rhyolite zones at the dike margins show gentle to open folds. In contrast, glassy central parts of the dike are flow-laminated and preserve folded and refolded isoclinal, curvilinear folds and sheath folds that record sustained progressive deformation. The inner interface between the glassy and lithoidal facies is abrupt and marked by elongation lineations and mullions. In the dike center, fold axes plunge 27° NE along the dike, and parallel to elongation lineations. Combined with shear sense indicators (σ - and δ -objects, sheared vesicles, and asymmetric folds) these indicate magma flow was obliquely upwards, to the southwest; locally $\leq 60^\circ$ to the propagation direction of the dike. The distribution of structures within the rhyolite indicates local accretion of the (now) devitrified material to the margins, with localization of flow into the center of the dike. We find that the initial magma flow direction was controlled by fracture propagation and interaction, with the subsequent flow record controlled by accretion

and flow localization in the conduit. The study demonstrates that analysis of mesoscopic structural and kinematic features (several of which have not previously been reported from dikes) is a powerful tool to reconstruct the complex evolution of conduit initiation and magma flow processes.

1 . Introduction

Exposed igneous dikes are exhumed frozen conduits along which magma flowed, and in some cases, fed explosive or effusive volcanic eruptions. Dike propagation, and the behavior of magma flowing along a conduit, influence whether material will reach the surface or arrest underground (Rubin, 1995) and understanding is needed to better inform models of subsurface magma movement used to forecast eruptive activity. Recent interest has focused on exhumed conduits to provide information about the nature of conduit flow and extrusive eruptions. For example, textural observations have been used to constrain shallow magma flow, fracture, and vesiculation processes during silicic eruptions. (Tuffen and Dingwell, 2005; Tuffen and Castro, 2009; Wadsworth et al., 2016). A useful additional approach has been to deduce flow directions from anisotropy of magnetic susceptibility (AMS) (Callot et al., 2001; Geoffroy et al., 2002; Roni et al. 2014). The major axis of the AMS ellipsoid is commonly inferred to lie parallel to the direction of magma flow, whereas studies of microfabrics, such as oriented phenocrysts and vesicles, can give flow directions orthogonal to this (e.g., Poland et al., 2004; Philpotts and Philpotts, 2007). Magma flow directions may vary and even reverse with time, as magmatic pressures build up and relax, and as the magmatic plumbing system propagates, dilates, locally solidifies or becomes too viscous to flow, leading to channelized magma flow (e.g., Bruce and Huppert, 1990; Lister, 1995; Platten, 2000; Holness and Humphreys, 2003). Measured fabrics record the last increments of strain as magma shear at that local site ceased, whereas initial conduit propagation by hydrofracturing (e.g., Baer 1995) ahead of the dike tip may be independent of magma transport direction in the formed conduit, hence it is important to constrain the temporal evolution of flow within the system.

Here, we present new mesoscopic evidence of dike propagation, magma transport and lateral accretion, with channelization of subsequent magma flow. Exceptionally preserved mesoscopic structures and fabrics highlight how changes to magma properties cause the conduit geometry to evolve during transport and arrest. Mesoscopic features, such as dike-wall plumose structures, σ - and δ -objects, verging folds, sheath folds, sheared vesicles, and

mullion structures, are common in silicic plumbing systems, and can readily be employed to reconstruct processes in intrusions, eruptive conduits, and lavas.

2 Geological setting

The Paleogene rhyolite dike documented here is part of the basalt-rhyolite North Atlantic Igneous Province. It is named ‘Judd I’ after Judd (1893) and is one of four composite dikes at Tormore, on the west coast of the Isle of Arran, western Scotland (**Fig. 1A,B**), which intrude thin-bedded Triassic sandstones of the Palaeozoic Midland Valley Terrane (Meade et al., 2009). Arran hosts two Paleogene igneous centres (Emeleus, 1982; **Fig. 1A**) emplaced at ~60 Ma (Dickin et al., 1981): (1) the Northern Granite, which intrudes Neoproterozoic (Dalradian) metasedimentary basement and probably straddles the Caledonian Highland Boundary Fault; and (2) the slightly younger Central Complex, which is an exhumed caldera volcano (King, 1955). Sr and Pb isotopes show that Judd I is related to the Central Complex (Dickin et al. 1981). The host sandstone around Judd I generally dips at 092°/23° S (here we use the convention strike, dip, and dip direction) (**Fig. 1C,E**) and is cut by deformation band sets that strike N-S and dip steeply west (**Fig. 1D**). The deformation bands pre-date the dike and are related to the emplacement of the earlier Northern Granite (Woodcock and Underhill, 1987). The present-day dike outcrops were exhumed from a depth of about 1 km, but otherwise exhibit no evidence for post-emplacement deformation.

3 Dike geometry and internal structure

3.1 Dike geometry

Judd I dike is exposed intermittently along strike for a distance of 450 m on the coastal foreshore (**Fig. 1**). The dike generally dips steeply eastward towards its inferred source (the Central Complex: Dickin et al., 1981) and has the geometry of a cone sheet. Contact attitude varies locally from north to south. The northernmost part strikes NE-SW (**Fig. 1B,C**) as a bifurcation from the longer N to NNE-striking part (**Fig. 1C**). The outcrop width of the dike varies from <6.5 m to 12 m and reflects a change from sub-vertical to inclined margins. The dike is rhyolitic and mostly glassy, ranging from massive to flow-banded and locally has devitrified (lithoidal) flow bands, 0.02-2.00 m thick. The strike of the dike gradually changes from N-S to NNE-SSW southwards over a distance of 330 m. Along that length the dike is up to 3 m thick with a steeply-dipping flow lamination that grades eastwards into a massive vitrophyre.

The dip of the main exposed region of the dike varies (**Fig 1E**). The northern parts dip steeply west ($\sim 78^\circ$) but the central and southern part generally dips steeply east ($60\text{--}80^\circ$), and includes a more gently dipping ($0\text{--}30^\circ$ E) medial segment, expressed as a 40-m long section of broader (10-12 m) outcrop width (**Fig. 1E**). This bend or step-like geometry is also reflected in a staggered offset of the dike margins, in contrasting dips of flow banding, and of opposing dike walls at outcrop (**Fig. 1E**). The dike geometry is best considered in cross-section (**Fig. 1F,G**), showing the form of a dilational jog that would be consistent with a relative upward movement of the eastern side and/or dextral shear in plan view. A west-striking basalt dike, 0.1-0.3 m wide, joins the rhyolite dike and south of this a ≤ 30 cm thick basalt dike lines both margins (**Fig. 1E**), but there is no clear evidence of the relative timing between basalt and rhyolite intrusion.

3.2 Flow lamination and bands

The rhyolite exhibits laminations (< 3 mm thick) that are near continuous at the outcrop scale, and mostly sub-parallel to the nearest dike margin. They are particularly clear in the vitrophyre (e.g., **Fig. 1E** and **Fig. 2**) and are defined by differences in glass colour, states of devitrification, and the abundance of microlites (**Fig. 2B**; see e.g., Castro et al. 2005). Lateral terminations and folding of laminations indicate ductile deformation during flow, i.e., they represent a flow banding. The devitrified rhyolite ('felsite') is also flow-laminated, with laminations 1-15 mm thick, of variably microcrystalline and spherulitic material. Centimetre- to metre-thick bands of devitrified material with sharp contacts are also observed locally within zones of laminated glassy rhyolite (e.g., **Fig. 2A**). Where folded, these bands are harmonious to folds in the glass, and thicken in the hinges (e.g., **Fig. 2**). Thickening of this kind is associated with passive folding (e.g., Class 2 folds: *cf.* Ramsay and Huber, 1987), which tend to form in cases where the layering has no mechanical control on fold geometry. The apparent Class 2 folds shown in Figure 2 are therefore probably a type of shear fold, related to a component of simple shear; to be expected during parabolic conduit flow. Notably, devitrification would have been accompanied by a very marked change in rheology between the felsite bands and the enclosing glassy rhyolite, hence devitrification probably post-dated the folding and attenuation, as inferred in rhyolites elsewhere (Andrews and Branney, 2011).

3.3 Elongation lineations

Lineations in the form of straight slickenline-like grooves with micro-pitted surfaces (**Fig. 3A**) and more irregular micro-corrugations (**Fig. 3B**) occur on foliation surfaces

between vitrophyre and devitrified rhyolite. Individual grooves are 1.0-2.5 mm wide, sub-parallel to each other, and range in length from discontinuous to ≥ 2 m (limited only by the size of exposure). The lineated surfaces exhibit abundant equant micropits that may represent late-stage vesiculation or mineral growth along a parting, as commonly seen on partings in rhyolite lavas and rheomorphic ignimbrites (Branney and Kokelaar 1992; Andrews and Branney 2011). In the southernmost part of the dike (bottom left in **Fig. 1E**) the lineations trend $27^\circ/046^\circ$ (NE) (in the format: plunge / down-plunge azimuth). Further north, in the wider part of the dike (top right in **Fig. 1E**) they trend $38^\circ/227^\circ$ (i.e. at over 90° to that in the south). These types of lineation are characteristic of elongation (stretching) lineations, which form in the major (x -) axis of the finite strain ellipsoid, though it should be noted that such features could be rotated by subsequent flow events. In either case, the lineations are inferred to be parallel to a magma transport direction, indicating local variations along the exposed length of the dike.

3.4 Open folds, curvilinear folds, curtain folds, sheath folds, and refolded folds

Laminations within the rhyolite are folded into abundant gentle to isoclinal intrafolial folds. Fold class varies with distance from the dike margins, with gentle folds (interlimb angle (α) = 180° to 120°) and open folds (α = 120° to 70°) near the margins (**Fig. 4**), and close folds (α = 70° to 30°) to isoclinal folds (α = 0°) in the centre (**Fig. 5**). The dike centre also exhibits numerous refolded folds (e.g., the ‘Type 3’ coaxial refolded sheath folds shown in **Fig. 5B**). This increase in fold tightness and complexity from single- to multi-phase folds and refolds coincides with the change from felsite near the margins, to laminated vitrophyre near the dike centre (**Fig. 1E,G**). Shear strain recorded by folds – which accounts for a fraction of the total shear strain associated with emplacement of the flow banded magma – is significantly higher towards the centre of the dike than near the margins. This is in contrast to a simple laminar flow model, in which the greatest shear would be at the margins between the magma and the stationary host rock (Gonnermann and Manga, 2003). The highest fold-related shear strain occurs at the interface between the laminated vitrophyre and devitrified rhyolite. This may reflect initial accretion of what is now devitrified rhyolite onto the dike margins, resulting in a narrowing of the conduit and the localization of flow within the dike centre.

Gentle to open fold hinges are variably orientated, and commonly, are at high angles to the dominant elongation lineation (**Fig. 1E**, and **Fig. 4C,D**), whereas close to isoclinal curtain folds (‘oblique’ folds of Passchier and Trouw, 1996), sheath folds, and refolds have hinge lines sub-parallel to the local lineation direction (i.e., $27^\circ/046^\circ$; **Fig. 1E**). When viewed along

the lineation, sub-perpendicular surfaces within the glassy rhyolite reveal the distinctive elliptical *eye-structure* patterns that are cross-sections through sheath fold culminations or saddles (**Fig. 5**). The geometric x , y and z axes (i.e., the long, intermediate, and short axes) of sheath folds are generally accepted as lying sub-parallel to the x , y , and z axes of the finite strain ellipsoid (Alsop and Holdsworth, 2006). The x -axis of a sheath fold is therefore sub-parallel to the transport direction; the y -axis lies in the plane of the foliation - in this case the flow-lamination - with the z -axis forming the normal to that plane (**Fig. 5A,B**). In Judd I, the sheath fold x -axes lie parallel to the local elongation lineation, and the x - y planes are sub-parallel to the dike margins.

The presence of sheath folds and oblique folds indicates non-coaxial, simple (or sub-simple) shear as the magma flowed along the dike. The fold axes probably formed at various orientations, and then rotated or transposed into sub-parallelism with the foliation and stretching lineations. Sheath folds, curtain folds, and refolds are a characteristic feature of silicic rheomorphic ignimbrites worldwide (Branney et al. 2004), and are ascribed to protracted non-coaxial flow prior to quenching. For the same reasons we speculate they are probably a common feature of silicic dikes and lavas, even though they have not been widely reported hitherto.

3.5 Mullion Structures

Mullion structures are well developed in the dike (**Fig. 6**) and range in size, from cm-scale amplitudes and wavelengths on felsite-basalt contacts (**Fig. 6B**), to centimeter to meter-scale amplitudes and wavelengths on contacts between the felsite and the glassy rhyolite (**Fig. 6A**). In all cases the sharp cusps of the mullions point into the felsite, indicating that the felsite had a higher shear viscosity than the vitrophyre at the time of deformation. The mullions are curvilinear, elongate parallel to stretched vesicle lineations, and lie parallel to local fold hinge lines in the glassy rhyolite, at $27^{\circ}/046^{\circ}$ (NE). Exceptions to this occur towards the eastern margin on the felsite-basalt contact, where the mullions exhibit a near 90° bend within the magmatic foliation, from lineation-parallel, to sub-horizontal. In the southernmost part of the dike (bottom left in **Fig. 1E**), mullions and stretching lineations generally trend at $27^{\circ}/046^{\circ}$ (NE), with the aforementioned exceptions.

Mullions have not been reported hitherto in dikes. Elsewhere they have been described with respect to pure shear (e.g., Ramsay and Huber, 1987) and simple shear deformations (e.g., Harigane et al., 2008). In pure shear (layer-parallel shortening) the mullion crest lines form at a high angle to the shortening axis, as in the early stages of fold development. During

simple shear (i.e. stretching- and fault-related), as would be expected for a parabolic conduit-flow model (e.g., Rust et al., 2003; Manga, 2005), mullion crest-lines align in the transport direction. Whether the mullions formed in their present orientation, or were rotated into it, is not clear, but in either case the finite strain ellipsoid is strongly prolate, with the x -axis aligned with the mullion crest lines. Mullions related to pure shear have their z -axis in plane with the dike margin and at 90° to the crests whereas mullions related to simple shear have their z -axis at 90° to the crests but normal to the margin. The latter is consistent with the inferred strain ellipsoid for sheath folds in the central part of the dike, but as noted earlier, the interface marked by the mullions also represents a strain zone boundary. In all cases, the mullion crest lines are curved, which we infer as representing modification by later flow: i.e. that the mullions are refolded rather than representing primary non-cylindrical folds.

3.6 Shear sense indicators

The dike contains abundant asymmetric features that record the local sense of simple shear within the rhyolite. Combined with the local elongation lineations, they allow the magma transport direction to be deduced. Similar features have been used to indicate shear sense in rhyolitic rheomorphic ignimbrites (e.g. Knott et al., 2016) and lavas (Smith, 1996). They include small asymmetric folds (**Fig. 7A**), rotated laminations (**Fig. 7B,C**), and rotated objects (**Fig. 7B**), and are variously exposed in plan and section view across the thickness of the dike. This distribution is broadly consistent with a laminar flow regime at the dike-thickness scale, though it is worth noting that high shear-strain boundaries within the dike (see *Mullion Structures* above) suggest an evolution to localized laminar flow. Parasitic folds indicate flexural slip on lamination surfaces, which is supported by slickenline-like lineations noted earlier (**Fig. 3A**). In the vertical plane, all shear sense indicators show that the centre of the dike has moved upward relative to material nearer the dike margins (e.g., **Fig. 7A**), suggesting that the primary ascent direction is preserved rather than backflow kinematics. Combined with observations in the horizontal plane, those shear sense indicators record upward flow toward the south (**Fig. 7B**), parallel to elongation lineations (**Fig. 1E and Fig. 3**) and to the dominant mullion crest line (**Fig. 6**).

3.7 An axial planar fabric

A mm-spaced fabric cuts the flow lamination in the hinge-zone of folds in the central part of the glassy rhyolite (**Fig. 8**). It lies parallel to the axial plane of close to tight refolded sheath folds, and is sub-parallel to the x - y plane of nearby sheath folds that are not refolded. Thin sections (cut parallel to the x - z plane of the folds) highlight several axial planar fabrics,

including preferential microlite alignment (**Fig. 8D**), and two scales of fractures: (1) mm-length and μm -aperture fractures that are preferentially developed in devitrified laminations (**Fig. 8B**) and form close-spaced networks; and (2) cm-length fractures with apertures up to ~ 0.1 mm (**Fig. 8b,C**) that are spaced at the mm-scale. Both fracture scales are observed cutting phenocrysts (**Fig. 8C**). The smaller-scale fractures are also observed wrapping phenocrysts, and commonly appear to correspond to microlite crystal boundaries. Where larger fractures cut phenocrysts (**Fig. 8C**), there is no observed offset of the crystal surface, which is inferred here to represent opening mode. Both scales of fracture are segmented along their length, though no preferred stepping direction is noted here.

The fabric presented by microlite alignment (**Fig. 8D**) is relatively weak compared to the fracture sets; the fractures are inferred to be responsible for the mesoscale fabric observed in the field. Microlite alignment may however impose a mechanical control on fracture nucleation and growth, particularly in the case of the smaller-scale and closer-spaced fracture set. We infer here that the alignment of microlites is related to flow in the conduit, in which elongate microlites are generally rotated into parallelism with the flow direction. Microlites that are at a high angle to the preferred orientation may represent an impinged crystal population, such that adjacent microlites inhibit rotation (cf. Manga, 1998). This effect is known to occur even in relatively dilute suspensions subjected to simple shear flow (Stover et al., 1992; Manga, 1998).

The fractures are similar in appearance to close-spaced ‘sheeting fractures’ that develop during cooling preferentially along strongly anisotropic flow banded rhyolitic lavas and lava-like ignimbrites (Bonnichsen and Kauffman, 1987; Andrews and Branney 2011). Such fractures are late-stage and cross-cut tight hinge zones of near-isoclinal folds and show negligible offset of the laminations they cut (Andrews and Branney 2011). The fabric in Judd 1 dike shows gentle fanning in the hinge zone (**Fig. 8**), and on limbs seems to gently refract as it cuts through the different flow laminations. Mechanical rotation of microlites, into the x -axis of the finite strain ellipsoid, during folding may be involved, with the hinge zone pattern suggesting pre-folding layer-parallel shortening, but this requires further study. Axial planar fabrics have not been reported in glassy dikes hitherto, although an example has been reported in a glassy rhyolite lava and inferred to have formed by mechanical rotation of phenocrysts (cf. Cioni and Funedda, 2005).

4 Dike margins

The sandstone country rock near the margins of the dike has been indurated with porosity reduction (**Fig. 9B**). These <1 m thick baked margins are resistant to erosion and locally form walls enclosing the dike. These country rock dike-margin surfaces preserve a number of intriguing features that can be viewed facing outwards from the dike (**Fig. 9A,C**). In the north (see **Fig. 1C**), the contact has concave-to-dike curved surfaces, which meet at ridges, spaced ~20 cm apart, that project 1-4 cm into the dike (**Fig. 9A**). The curved surfaces are marked by a pattern of sub-parallel, discontinuous open folds that have a range of forms, from sinusoidal to cusp-and-lobe geometry (**Fig. 4A,B and Fig. 9A,C**). These folds have a wavelength of 10 mm, and amplitudes in the range 2-5 mm, and in some cases occur en echelon (**Fig. 9C**). The fold crests are curvilinear and mostly orientated at a high angle to the larger ridges (**Fig. 9A,C**). Most of the crests are not aligned with the bedding of the country rock and so cannot relate to original host layering: this is confirmed in thin sections of the country rock margin (**Fig. 9B**), which show pore space reduction related to quartz recrystallization, as well as 1-3 mm thick vestiges of rhyolite. Notably, pockets of rhyolite are found up to 2 cm from the margin, which are closely associated with 0.2-1.5 mm diameter pores. As these pores are significantly larger than the country rock porosity, and in the absence of evidence for clast removal during sectioning, we infer that these are vesicles, formed during heating by the dike, and hence that the narrow contact zone represents a peperite (*cf.* Kokelaar 1982; Skilling et al., 2002).

Linear ridges such as those shown in **Figure 9A** are similar in appearance to plumose structures on joints (*cf.* Suppe, 1985), or radial fast fracture surfaces (*cf.* Liu, 2005). Examples of joint surface plumose structures comprise a central lineation, from which curvilinear lineations (hackles) radiate (Suppe, 1985; Van der Pluijm and Marshak, 2004; Pollard et al., 2004); the central lineation originates at a point of weakness, and are sometimes associated with a smooth “mirror zone” (see **Fig. 9A,C**). Hackle marks radiate from the central lineation and are convex in the direction of fracture propagation. In some cases, plumose structures are accompanied by ridges, which record joint arrest. The Judd I margins host linear ridges of this type at various attitudes, from sub-vertical to sub-horizontal, implying that fracture propagation directions varied locally along the margin. Open fold hinge attitudes are broadly consistent between plumose structures, but are variably oriented along the margin, with hinge plunges also ranging from sub-vertical to sub-horizontal. We infer that these open folds formed at a high angle to the magma transport direction (see section above on *Open folds*),

and have not been modified by later flow. In which case, variability in attitudes suggests that initial magma transport was controlled by local fracture propagation, and not *vice versa*.

5 Discussion

5.1 Dike propagation inferred

Segmentation of the dike margins (**Fig. 1A,B** and **Fig. 9A,C**) indicates that the initial propagation of the dike occurred as a series of fractures ahead of the magma front, which were infilled and inflated by magma following phases of fracture linkage, to form a through-going sheet. Similar propagation and infilling has been proposed for basaltic dikes (Baer, 1995) and for basaltic sills (Walker, 2016). Stepped linkage of dike lobes is generally taken to indicate the direction of propagation (e.g., Baer and Reches, 1987). In the northern section (see **Fig. 1C,D** and **Fig. 9A,C**), local segments indicate upward propagation as lobes (e.g., **Fig. 9A**) but they do not show a consistent step orientation. This is supported by plumose structures and folds preserved in the country rock margins, which indicate highly irregular local propagation directions, including abrupt changes of as much as 90° in direction along individual segments (**Fig. 9C**). These structures show that although the geometry of the segments may indicate upward propagation, much of the fracture propagation was lateral, and served to link isolated fracture surfaces (**Fig. 10A**). The local lateral propagation directions may have been controlled by the interaction of zones of induced tension within the host rock ahead of adjacent propagating dike tips (see e.g., Rogers and Bird, 1987), or by mechanical heterogeneity within the host rock, leading to local stress perturbations. In such cases, the initial magma flow direction is therefore dictated by the propagation and interaction of cracks ahead of the magma front, hence we should not expect a consistent flow direction at this stage in conduit evolution. The contrast between the multiple propagation and flow directions recorded on the margins of Judd I, and the relatively consistent upward and southwest directed interior flow record, suggests that the incipient and established conduit flow records are relatively independent of one another.

5.1 Migration of the conduit margin by accretion

Rhyolite dikes commonly have quenched glassy margins, with devitrified centres recording slower cooling (e.g., Stasiuk et al., 1996; Tuffen and Dingwell, 2005). Judd I does not fit this configuration in that the devitrified parts occur locally at the dike margins. Devitrification can be enhanced by uptake of Na- and/or K-rich solutions, which may have been present in the sandstone country rock (Lofgren, 1970). Contacts with adjacent glassy

rhyolite are sharp, and in some cases, devitrified material is intercalated with vitrophyre as flow bands (e.g., **Fig. 2**). Mullion structures indicate that what is now devitrified rhyolite was more viscous than the material that remains as glassy rhyolite. The mullion structure surface also serves as a shear strain boundary that separates relatively low strain marginal features from the higher strain centralized structures within the adjacent vitrophyre. These observations point to viscosity variability in the conduit during emplacement, and therefore cannot be related to post emplacement devitrification. The viscosity of silicic magmas is strongly affected by temperature and volatile content (e.g., Hess and Dingwell, 1996; Giordano et al., 2008). A decrease in H₂O concentration of 1 wt. % can result in a shear viscosity increase of 1-4 orders of magnitude (Hess and Dingwell 1996), and a drop in temperature of 100°C could increase viscosity by an order of magnitude (Giordano et al., 2004). Minor cooling and/or degassing in Judd I may have resulted in significant increases in shear viscosity. We infer that initial intrusion into thin cracks resulted in minor reduction in magma temperature (due to conduction into the country rock, and/or convection during host-rock fluidization resulting in peperite), and/or volatile escape into the propagating crack tip (e.g., Rubin, 1995). This material became increasingly viscous and accreted locally at the conduit margins (**Fig. 10Bi-ii**). Continued flow within the accretion-modified conduit would be insulated from losses in temperature and/or volatiles (as a result of reduced permeability at the dike wall due to accretion) to the country rock; instead heat (and/or volatile) transfer would be into the accreted material, probably leading to a reduction in viscosity, and re-entrainment of some of the formerly accreted rhyolite at the conduit margins (**Fig. 10Biii**). Progressive shear within the margins led to the development of mullion structures. With prolonged conduit flow the dike margins may have had a complex thermal history of cooling and accretion followed by partial heating, re-entrainment, and cooling, and this may partly account for the post-magmatic textural variation (i.e., the present distribution of devitrified rhyolite).

5.2 Magma emplacement history summarized

The distribution and style of mesoscopic folds, fabrics, and kinematic indicators, together with the spatial distribution of textural variations within the rhyolite dike, record progressive deformation associated with sustained magma flow and accretion within the conduit. Margin-hosted folds record magma flow associated with initial fracture propagation and infilling. The variable attitudes of these folds indicates initial flow directions were highly variable, and dictated by fracture propagation direction. We infer that the now devitrified

rhyolite lenses preserved near corners at the dike margins represent the early stages of dike filling and accretion. Marginal devitrified material displays open folds (**Fig. 4A,B**) with hinge lines at high angles to local elongation lineations. Those open folds could represent either late-nucleated folds, or material that was accreted to the dike margin and has been subjected to only minor transposition prior to that accretion. Elongation lineations lie parallel to mullion structures developed on the interface between material that is now devitrified rhyolite, and the laminated rhyolite, and parallel to the x -axis of sheath folds within the laminated vitrophyre. Sheath folds represent early-nucleated folds that were progressively transposed (with rotation of the fold hinges from near flow-perpendicular to flow-parallel) within less rapidly chilled, central parts of the conduit within the dike. The cusp and lobe shape of mullion structures is indicative of a minor contrasts in shear viscosity during deformation, and here, demarcate the boundary between high and low strain zones. We therefore infer that the mullions preserve the interface between a relatively immobile material at the dike margins (material that is now devitrified rhyolite), and localized flowing magma within the dike centre (i.e., the laminated vitrophyre) (**Fig. 10Bi-iii**). By combining the trend of mullion structures and elongation lineations, with shear sense indicators across the dike, we infer that the local flow direction *recorded* in the laminated vitrophyre was generally upward to the south at 27° from horizontal. This flow direction relates to emplacement of the laminated vitrophyre only; earlier flow directions are not generally preserved. Elongation lineations in the wider part of the dike (e.g. the northernmost part of **Fig. 1E**) trend $38^\circ/227^\circ$, at nearly 90° to this direction, and are associated with an interface between massive rhyolite glass, and devitrified material. On the basis that lineations prior to sheath fold formation would be rotated into a southward flow direction, we infer that this $38^\circ/227^\circ$ trend relates to a late-stage flow, presumably during emplacement of the massive glass, again through the central part of the dike (**Fig. 10Biv**). Given the tendency for modification of the early flow record by later stages of *localized* flow, it is therefore critically important to consider the context of kinematic indicators during studies of magma flow.

6 Conclusions

Detailed analysis of a glassy rhyolite dike has demonstrated that useful insights into the propagation and magma flow history of volcanic conduits can be reconstructed from dike margin structures, and the combination of mesoscopic magmatic fabrics, and folds, together with kinematic indicators, and the spatial disposition of contrasting textural variations. Several of these features (e.g. sheath folds, curtain folds, mullion structures, δ -objects) have

not been documented hitherto in silicic dikes, although some (δ -objects and elongation lineations) have been documented in silicic lavas. Structural analysis of Judd I dike reveals highly variable flow directions during the early stages of dike propagation, followed by dike segment inflation, and fracture linkage to form a through-going sheet. Early-intruded material accreted to the margins, narrowing the conduit. Subsequent and prolonged flow is recorded as upward and to south, at about 27° from horizontal. Accreted marginal rhyolite is locally overprinted by structures formed during the later stages of magmatic flow, hence the flow record is notably incomplete. Nevertheless, the study highlights the complexity of flow histories in dikes, and the importance of context in distinguishing between dike propagation directions and the various subsequent flow phases within the conduit.

Acknowledgements

The authors thank Hugh Tuffen and Ze'ev Reches for thorough and insightful reviews that improved this paper, and constructive comments from the Associate Editor Gerardo Carrasco Núñez. The field area falls within the Drumadoon-Tormore Site of Special Scientific Interest (SSSI; site code 537): samples were collected in 2005 with permission of Scottish National Heritage.

References

- Alsop, G.I., Holdsworth, R.E., 2006. Sheath folds as discriminators of bulk strain type. *Journal of Structural Geology* 28, 1588-1606.
- Andrews, G.D. M., Branney, M.J. 2011. Emplacement and rheomorphic deformation of a large rhyolitic ignimbrite: Grey's Landing, southern Idaho. *Geological Society of America, Bulletin* 123: 725-743. doi: 10.1130/B30167.1
- Baer, G. 1995. Fracture propagation and magma flow in segmented dykes: field evidence and fabric analysis, Makhtesh Ramon, Israel. In: Baer, G. & Heinmann, A.A. (eds) *Physics and Chemistry of Dykes*. Balkema, Rotterdam, 125-140.
- Baer, G., Reches, Z. 1987. Flow patterns of magma dikes, Makhtesh Ramon, Israel. *Geology*, 15;569-572. doi: 10.1130/0091-7613
- Bonnichsen, B., and Kauffman, D.F., 1987, Physical features of rhyolite lava flows in the Snake River Plain volcanic province, southwestern Idaho, in Fink, J.H., eds., *The Emplacement of Silicic Domes and Lava Flows*: Geological Society of America Special Paper 212, p. 119-145.
- Branney, M.J., and Kokelaar, B.P., 1992, A reappraisal of ignimbrite emplacement: Progressive aggradation and changes from particulate to non-particulate flow during emplacement of high-grade ignimbrite: *Bulletin of Volcanology*, v. 54, p. 504-520, doi: 10.1007/BF00301396.
- Branney, M.J., Barry T.L., Godcheaux, M. 2004. Sheathfolds in rheomorphic ignimbrites. *Bulletin of Volcanology*, 66: 485-491
- Bruce, P. M. & Huppert, H. E. (1990). Solidification and melting along dykes by the laminar flow of basaltic magma. In: Ryan, M. P. (ed.) *Magma Transport and Storage*. Chichester: John Wiley, pp. 87-101
- Callot JP, Geoffroy L, Aubourg C, Pozzi JP, Mege D (2001) Magma flow directions of shallow dykes from the East Greenland volcanic margin inferred from magnetic fabric studies. *Tectonophysics* 335: 313-329

- 433 Castro JM, Dingwell DB, Nichols ARL, Gardner JE (2005) New insights on the origin of flow bands in obsidian,
434 In Manga M, and Ventura G, eds., Kinematics and dynamics of lava flows : Geol Soc Am special paper
435 396: 1-14
- 436 Cioni, R., Funedda A. (2005) Structural geology of crystal-rich, silicic lava flows: a case study from San Pietro
437 Island (Sardinia, Italy) In: Manga M., and Ventura G (eds) Kinematics and dynamics of lava
438 flows. Geological Society of America Special Paper 396, 1-14 doi: 10.1130/2005.2396(01)
- 439 Dickin AP, Moorbath S, Welke HJ (1981) Isotope, trace element and major element geochemistry of Tertiary
440 igneous rocks, Isle of Arran, Scotland. Royal Society of Edinburgh Transactions, 72: 159-170
- 441 Emeleus, C.H. 1982. The British Tertiary Province: the central complexes. In: SUTHERLAND, DS. (ed.)
442 Igneous rocks of the British Isles. John Wiley and Sons, Chichester, 369-414.
- 443 Geoffroy L, Callot JP, Aubourg C, Moreira M (2002) Magnetic and plagioclase linear fabric discrepancy in
444 dykes: a new way to define the flow vector using magnetic foliation. Terra Nova 14: 183-190
- 445 Giordano D, Romano C, Papale P, Dingwell DB (2004) The viscosity of trachytes, and comparison with basalts,
446 phonolites, and rhyolites. Chem Geol 213 1-3:49-61
- 447 Giordano, D., Russell, J.K. and Dingwell, D.B., 2008. Viscosity of magmatic liquids: a model. Earth and
448 Planetary Science Letters, 271(1), pp.123-134.
- 449 Gonnermann, H.M. and Manga, M., 2003. Explosive volcanism may not be an inevitable consequence of magma
450 fragmentation. *Nature*, 426(6965), pp.432-435.
- 451 Harigane, Y., Michibayashi, K., Ohara, Y. 2008. Shearing within lower crust during progressive retrogression:
452 Structural analysis of gabbroic rocks from the Godzilla Mullion, an oceanic core complex in the Parece
453 Vela backarc basin. Tectonophysics, 457, 183-196
- 454 Hess, K.U., Dingwell, D.B. 1996. Viscosities of hydrous leucogranitic melts: a non-Arrhenian model. Am.
455 Mineral 81:1297-300
- 456 Holness, M.B., Humphreys, M.C.S. 2003. The Traigh Bhaán na Sgúrra Sill, Isle of Mull: Flow Localization in a
457 Major Magma Conduit. Journal of Petrology, 44, 1961-1976
- 458 Judd, J.W. 1893. On composite dykes in Arran. Quarterly Journal of the Geological Society London 49: 536-564
- 459 King, B.C. 1955. The Ard Bheinn area of the Central Igneous Complex of Arran. Quarterly Journal of the
460 Geological Society of London, 110, 323-356.
- 461 Kokelaar, B.P. 1982. Fluidisation of wet sediment during the emplacement and cooling of various igneous
462 bodies. Journal of the Geological Society, London, 16, 139, 21-33.
- 463 Knott, T.R., Reichow, M.K., Branney, M.J., Finn, D.R., Coe, R.S., Storey, M., Bonnicksen, B. 2016.
464 Rheomorphic ignimbrites of the Rogerson Formation, central Snake River Plain, USA: record of mid-
465 Miocene rhyolitic explosive eruptions and associated crustal subsidence along the Yellowstone hotspot
466 track. Bulletin of Volcanology 78, 4, doi 10.1007/s00445-016-1003-x
- 467 Lister, J.R. 1995. Fluid-mechanical models of the interaction between solidification and flow in dykes. In
468 Physics and chemistry of dykes, G. Baer & A. Heimann (eds.), 115—124, Balkema.
- 469 Liu, A. 2005. Mechanics and mechanisms of fracture: An introduction. ASM International
- 470 Lofgren, G., 1970. Experimental devitrification rate of rhyolite glass. *Geological Society of America Bulletin*,
471 81(2), pp.553-560.
- 472 Manga, M., 1998. Orientation distribution of microlites in obsidian. Journal of Volcanology and Geothermal
473 Research, 86(1), pp.107-115.
- 474 Manga, M., 2005. Deformation of flow bands by bubbles and crystals. *Geological Society of America Special
475 Papers*, 396, pp.47-53.
- 476 Meade, F.C., Chew, D.M., Troll, V.M., Ellam, R.M., Page, L.M. 2009. Magma ascent along a major terrane
477 boundary: crustal contamination and magma mixing at the Drumadoon Intrusive Complex, Isle of Arran,
478 Scotland. Journal of Petrology, 50, 2345-2374
- 479 Passchier, C.W., Trouw, R.A.J. 1996. Microtectonics. Springer, Berlin. pp 289
- 480 Philpotts, A.R., and Philpotts, D.E., 2007, Upward and downward flow in a camptonite dike as recorded by
481 deformed vesicles and the anisotropy of magnetic susceptibility (AMS). Journal of Volcanology and
482 Geothermal Research, 161, 81-94 doi: 10.1016/j.jvolgeores.2006.11.006

- 483 Platten, I.M. 2000. Incremental dilation of magma filled fractures: evidence from dykes on the Isle of Skye,
484 Scotland. *Journal of Structural Geology*, 22, 1153-1164
- 485 Poland MP, Fink JH, Tauxe L (2004) Patterns of magma flow in segmented silicic dikes at Summer Coon
486 volcano, Colorado: AMS and thin section analysis. *Earth Planet Sci Lett* 219:155–169
- 487 Pollard, D.D., Bergbauer, S. and Mynatt, I., 2004. Using differential geometry to characterize and analyse the
488 morphology of joints. Geological Society, London, Special Publications, 231(1), pp.153-182.
- 489 Ramsay JG, Huber MI (1987) The techniques of modern structural geology Volume 2: Folds and fractures.
490 Academic Press Ltd., London.
- 491 Rogers, R.D., Bird, D.K. 1987. Fracture propagation associated with dike emplacement at the Skaergaard
492 intrusion, East Greenland. *Journal of Structural Geology*, 9, 71-86
- 493 Roni, E., Westerman, D.S., Dini, A., Stevenson, C., Rocchi, S., 2014. Feeding and growth of a dyke-laccolith
494 system (Elba Island, Italy) from AMS and mineral fabric data. *Journal of the Geological Society, London*.
495 171. 413-424
- 496 Rubin AM (1995) Propagation of magma filled cracks. *Annu. Rev. Earth Planet Sci.* 23: 287-336
- 497 Rust, A.C., Manga, M. and Cashman, K.V., 2003. Determining flow type, shear rate and shear stress in magmas
498 from bubble shapes and orientations. *Journal of Volcanology and Geothermal Research*, 122(1), pp.111-
499 132.
- 500 Skilling, I.P., White, J.D.L. and McPhie, J., 2002. Peperite: a review of magma–sediment mingling. *Journal of*
501 *Volcanology and Geothermal Research*, 114(1), pp.1-17.
- 502 Smith, J.V., 1996. Ductile-brittle transition structures in the basal shear zone of a rhyolite lava flow, eastern
503 Australia. *Journal of Volcanology and Geothermal Research*, 72(3), pp.217-223.
- 504 Stasiuk, M.V., Barclay, J., Carroll, M.R., Jaupart, C., Ratté, J.C., Sparks, R.S.J. and Tait, S.R., 1996. Degassing
505 during magma ascent in the Mule Creek vent (USA). *Bulletin of Volcanology*, 58(2-3), pp.117-130.
- 506 Stover, C.A., Koch, D.L. and Cohen, C., 1992. Observations of fibre orientation in simple shear flow of semi-
507 dilute suspensions. *Journal of Fluid Mechanics*, 238, pp.277-296.
- 508 Suppe, J. 1985. *Principles of Structural Geology*. Prentice Hall, New Jersey.
- 509 Tuffen, H. and Castro, J.M., 2009. The emplacement of an obsidian dyke through thin ice: Hrafninnuhryggur,
510 Krafla Iceland. *Journal of Volcanology and Geothermal Research*, 185(4), pp.352-366.
- 511 Tuffen, H., Dingwell, D.B. (2005) Fault textures in volcanic conduits: evidence for seismic trigger mechanisms
512 during silicic eruptions. *Bulletin of Volcanology*, 67, 370–387.
- 513 van der Pluijm, B.A. and Marshak, S., 2003. *Earth Structure: An Introduction to Structural Geology and*
514 *Tectonics* (2nd edition)
- 515 Wadsworth, F.B., Kennedy, B., Branney, M.J., von Aulock, F.W., Lavallée, Y., Menendez, A. 2016. Exhumed
516 conduit records magma ascent and drain-back during a Strombolian eruption. *Bulletin of Volcanology*
- 517 Walker, R.J. 2016. Controls on transgressive sill growth. *GEOLOGY*, 44, 99-102. doi: 10.1130/G37144.1
- 518 Woodcock, N.H., Underhill, J.R. 1987. Emplacement-related fault patterns around the Northern Granite, Arran,
519 Scotland. *Geological Society of America Bulletin*, 5, 515-527. doi: 10.1130/0016-7606

520 **Figures**

521 **Fig. 1.** (A) Map of the Isle of Arran showing the distribution of the major Cenozoic igneous
522 bodies (Northern Granite, Central Complex, and Tighvein Complex). (B) Aerial image of the
523 Tormore coastline (GoogleEarth image 2009, © Getmapping PLC), with the locations and
524 mapped extents of Judd's composite dikes (I-IV). (C) Map of the northernmost exposures of
525 Judd I dike with cross-section (D) to show contact attitudes and inferred dike geometry in the
526 subsurface. Equal area lower hemisphere stereographic projection shows data relating to the

mapped area in C. **(E)** Detail map of the southern part of Judd I, with equal area lower hemisphere stereographic projections for data collected locally (as indicated in figure). **(F and G)** Cross-sections corresponding to the labeled traces Y-Y' and Z-Z' in E, showing contact attitude changes along strike, and the inferred sub-surface geometry of the dike.

Fig. 2. Folded flow laminations within rhyolite vitrophyre. **(A)** Devitrified bands are folded with the glass, and thicken into fold hinges. Stereographic projection shows structural data for the rhyolite vitrophyre shown in A. Flow is interpreted to be from left to right (upwards toward the southwest, sub-parallel to the fold hinges). **(B)** Plane polarized light photomicrograph showing folded flow-laminations of glassy and devitrified rhyolite.

Fig. 3. **(A)** Parallel lineations on the interface between laminated vitrophyre and devitrified rhyolite. **(B)** Lineations on the interface between devitrified rhyolite and the basalt margin of the dike. Stereographic projection shows lineations in A and B; average dike margin refers to the nearest (western) margin. Flow direction is interpreted to have been from bottom right to top left (upwards toward the southwest, parallel to lineation).

Fig. 4. Fold styles and phases associated with the lithoidal rhyolite of Judd I dike. **(A)** Margin-hosted open folds ($\alpha = 120^\circ$ to 70°) in the northern part of the dike. **(B)** Margin-hosted folds show ductile deformation across minor normal faults cutting the host rock and margin. **(C)** Open folds in lithoidal rhyolite (felsite) adjacent to the eastern margin in the southern part of the dike (sample no longer *in situ*). **(D)** Gentle folds ($\alpha = 180^\circ$ to 120°) on a sub-horizontal felsitic rhyolite band on the contact between rhyolite and basalt, and **(E)** open folds on a steeply-dipping felsite band on the contact between felsite and vitrophyre, in the outer 'marginal' zone of the dike. **(F)** Buckle folds in flow laminations within the marginal felsitic rhyolite.

Fig. 5. Fold styles and phases associated with centralized vitrophyres of Judd I dyke. **(A)** sheath fold and **(B)** refolded sheath fold in laminated rhyolite vitrophyre. **(C)** Curtain folds. Ruler in each photo is held parallel to the fold hinge (i.e., plunging $\sim 20\text{-}30^\circ$ NNE): the inferred x-axis of the finite strain ellipsoid. Flow direction with respect to the image is interpreted to be toward the viewer, parallel to the ruler shown in each image: this is upward to the southwest, sub-parallel to the fold hinges.

Fig. 6. Mullion-structures in the devitrified rhyolite showing cusate grooves separated by convex-up lobes, forming the interface with **(A)** the laminated vitrophyre, and **(B)** the basalt

margin. Flow direction is interpreted to be parallel to the mullions, upward to the southwest (i.e. from right to left, and toward the viewer).

Fig. 7. Shear sense indicators preserved within the rhyolite dike. (A) Asymmetric fold (left) and parasitic folds (right). (B) Rotated foliation and enclave-wrapping foliation in laminated vitrophyre. (C) Rotated δ -object in marginal felsite.

Fig. 8. (A) Axial planar fabric in the fold hinge zone of a tight fold in the central vitrophyre. Flow direction is interpreted to have been from left to right (upward to the southwest). (B) Plane polarized light photomicrograph of folded glassy and devitrified flow laminations. Inset shows lamination boundaries and main fracture set. (C) Major fracture sets cut phenocrysts, and show opening mode offset of crystal boundaries. (D) Fold hinge zone in devitrified lamination, showing minor fracture set, and microlite alignment.

Fig. 9. Dike margin textures: (A) Plumose structures with hackle ornament within the sandstone margin. Viewed from within the dike, toward the country rock. (B) Plane polarized light photomicrograph of the sandstone margin, cut normal to the dike margin (location shown in A; light grey area at the top of the photo was occupied by the dike). (C) Sandstone-hosted folds, viewed toward the country rock from within the dike. Hinge attitudes are highly variable on each fracture surface.

Fig. 10. (A) Simplified schematic dike segment propagation model for Judd I, based on dike margin textures in the northernmost part of the exposed dike (see **Fig. 1C-D** and **Fig. 9**). Stages 1 and 2: initial fracture propagation as segmented dike fingers. Stage 3: overlapping zones of induced tensile stress ahead of the crack tip favors lateral propagation of fingers. Inset shows broader context of the crack propagation, with idealized principal stress orientations related to a spherical magma chamber at depth (where compressive $\sigma_1 > \sigma_2 > \sigma_3$). Initial formation of the Judd I dike is as isolated extension or extensional-shear fractures, respectively parallel or sub-parallel to the σ_1 / σ_2 plane. The dike is segmented in plan and section view, i.e. a non-plane-strain. Note there is no consistent stepping direction observed. (B) Linkage of dike segments to form a through-going conduit. (i) Initial laminar flow regime as material first enters the conduit. (ii) A decrease in magmatic temperature and/or volatile content at the margins results in an increase in shear viscosity, leading to accretion onto the dike margins; the conduit margin moves toward the dike centre. (iii) Continued flow through the conduit results in heat transfer into the accreted margins, reducing viscosity, allowing some shear remobilization of accreted material. Mullion structures and slickenline-like

590 striations preserved on this contact indicate a step in flow velocity across the boundary, as
591 illustrated by the flow profile. Parts of the accreted margin may have become entrained into
592 the dike centre through flow perturbation folding (see inset). (iv) A final pulse of material
593 intruded the central part of the dike leading to further deformation of accreted material. The
594 localized nature of this intrusion results in preserved shear couples that do not conform to an
595 otherwise laminar flow regime across the dike thickness, and kinematic indicators that are not
596 aligned with the dominant, up-to-the-SW flow direction (see the northernmost part of **Fig.**
597 **1E**). Figure 10 is not drawn to scale. The flow velocity profiles are illustrative only.

Figure 1
w: 185 mm
h: 175 mm
(full page width)

Fig. 1. (A) Map of the Isle of Arran showing the distribution of the major Cenozoic igneous bodies (Northern Granite, Central Complex, and Tighvein Complex). (B) Aerial image of the Tormore coastline (GoogleEarth image 2009, © Getmapping PLC), with the locations and mapped extents of Judd's composite dikes (I-IV). (C) Map of the northernmost exposures of Judd I with cross-section (D) to show contact attitudes and inferred dike geometry in the subsurface. Equal area lower hemisphere stereographic projection shows data relating to the mapped area in C. (E) Detail map of the southern part of Judd I, with equal area lower hemisphere stereographic projections for data collected locally (as indicated in figure). (F and G) Cross-sections corresponding to the labeled traces Y-Y' and Z-Z' in E, showing contact attitude changes along strike, and the inferred sub-surface geometry of the dike.

Figure 2
w: 185 mm
h: 91 mm
(full page width)

Fig. 2. Folded laminations within rhyolite vitrophyre. (A) Devitrified bands are folded with the glass, and thicken into fold hinges. Stereographic projection shows structural data for the rhyolite vitrophyre shown in A. Flow is interpreted to be from left to right (upwards toward the southwest, sub-parallel to the fold hinges). **(B)** Plane polarized light photomicrograph showing folded laminations of glassy and devitrified rhyolite.

Figure 3
w: 123 mm
h: 165 mm
(2-column width)

Fig. 3. (A) Parallel lineations on the interface between laminated vitrophyre and devitrified rhyolite. (B) Lineations on the interface between devitrified rhyolite and the basalt margin of the dike. Stereographic projection shows lineations in A and B; average dike margin refers to the nearest (western) margin. Flow direction is interpreted to be from bottom right to top left (upwards toward the southwest, parallel to lineation).

Figure 4
w: 185 mm
h: 153 mm
(full page width)

Fig. 4. Fold styles and phases associated with the lithoidal rhyolite of Judd I. (A) Margin hosted open folds ($\alpha = 120^\circ$ to 70°) in the northern part of the dike. (B) Margin-hosted folds show ductile deformation across minor normal faults cutting the host rock and margin (C) Open folds in felsite adjacent to the eastern margin in the southern part of the dike (sample no longer in situ). (D) Gentle folds ($\alpha = 180^\circ$ to 120°) on a sub-horizontal felsite band on the contact between felsite and basalt, and (E) open folds on a steeply-dipping felsite band on the contact between felsite and vitrophyre, in the outer 'marginal' zone of the dike. (F) Buckle folds within the marginal felsite.

Figure 5
w: 123 mm
h: 181 mm
(2-column width)

Fig. 5. Fold styles and phases associated with centralized vitrophyres of Judd I. (A) sheath fold and (B) refolded sheath fold in laminated rhyolite vitrophyre. (C) Curtain folds. Ruler in each photo is held parallel to the fold hinge (i.e., plunging ~20-30°NNE): the inferred x-axis of the finite strain ellipsoid. Flow direction is interpreted to be from left to right, parallel to the ruler shown in each image: upward to the southwest, sub-parallel to the fold hinges.

Figure 6
w: 123 mm
h: 138 mm
(2-column width)

Fig. 6. Mullion-structures in the devitrified rhyolite showing cusate grooves separated by convex-up lobes, forming the interface with (A) the laminated vitrophyre, and (B) the basalt margin. Flow direction is interpreted to be parallel to the mullions, upward to the southwest.

Figure 7
w: 185 mm
h: 124 mm
(2-column width)

Fig. 7. Shear sense indicators preserved within the rhyolite dike. (A) Asymmetric fold (left) and parasitic folds (right). (B) Rotated foliation and enclave-wrapping foliation in laminated vitrophyre. (C) Rotated δ -object in marginal felsite.

Figure 8
w: 123 mm
h: 82 mm
(two column width)

Fig. 8. (A) Axial planar fabric in the fold hinge zone of a tight fold in the central vitrophyre. Flow direction is interpreted to be from left to right (upward to the southwest). **(B)** Plane polarized light photomicrograph of folded glassy and devitrified laminations. Inset shows lamination boundaries and main fracture set. **(C)** Major fracture sets cut phenocrysts, and show opening mode offset of crystal boundaries. **(D)** Fold hinge zone in devitrified lamination, showing minor fracture set, and microlite alignment.

Figure 9
w: 120 mm
h: 189 mm
(2-column width)

Fig. 9. Dike margin textures: (A) Plumose structures with hackle ornament within the sandstone margin. (B) Plane polarized light photomicrograph of the sandstone margin, cut normal to the dike margin (location shown in A). (C) Sandstone-hosted folds. Hinge attitudes are highly variable on each fracture surface.

Figure 10
w: 185 mm
h: 160 mm
(full page width)

Fig. 10. (A) Simplified schematic dike segment propagation model for Judd I, based on dike margin textures in the northernmost part of the exposed dike (see Fig. 1C-D and Fig. 9). Stages 1 and 2: initial fracture propagation as segmented dike fingers. Stage 3: overlapping zones of induced tensile stress ahead of the crack tip favors lateral propagation of fingers. Inset shows broader context of the crack propagation, with idealized principal stress orientations related to a spherical magma chamber at depth (where compressive $\sigma_1 > \sigma_2 > \sigma_3$). Initial formation of the Judd I dike is as isolated fractures aligned with the σ_1 / σ_2 plane. The dike is segmented in plan and section view, i.e. a non-plane-strain. Note there is no consistent stepping direction observed. (B) Linkage of dike segments to form a through-going conduit. (i) Initial laminar flow regime as material first enters the conduit. (ii) A decrease in magmatic temperature and/or volatile content at the margins results in an increase in shear viscosity, leading to accretion onto the dike margins; the conduit margin moves toward the dike centre. (iii) Continued flow through the conduit results in heat transfer into the accreted margins, reducing viscosity, allowing some shear remobilization of accreted material. Mullion structures and slickenline-like striations preserved on this contact indicate a step in flow velocity across the boundary, as illustrated by the flow profile. Parts of the accreted margin may have become entrained into the dike centre through flow perturbation folding (see inset). (iv) A final pulse of material intruded the central part of the dike leading to further deformation of accreted material. The localized nature of this intrusion results in preserved shear couples that do not conform to an otherwise laminar flow regime across the dike thickness, and kinematic indicators that are not aligned with the dominant, up-to-the-SW flow direction (see the northernmost part of Fig. 1E). Figure 10 is not drawn to scale. The flow velocity profiles are illustrative only.