Examining rhyolite lava flow dynamics through photo-based 3-D reconstructions of the 2011-2012 lava flowfield at Cordón-Caulle, Chile.

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Abstract

During the 2011-2012 eruption at Cordón-Caulle, Chile, an extensive rhyolitic flowfield was created (in excess of 0.5 km^3 in volume), affording a unique opportunity to characterise rhyolitic lava advance. In 2012 and 2013, we acquired approximately 2500 digital photographs of active flowfronts on the north and east of the flowfield. These images were processed into three-dimensional point clouds using Structure-from-Motion Multi-view Stereo (SfM-MVS) freeware, from which digital elevation models were derived. Sequential elevation models—separated by intervals of three hours, six days, and one year—were used to reconstruct spatial distributions of lava velocity and depth, and estimate rheological parameters. Three-dimensional reconstructions of flowfronts indicate that lateral extension of the rubbly, 'a'ā-like flowfield was accompanied by vertical inflation, which differed both spatially and temporally as a function of the underlying topography and localised supply of lava beneath the cooled upper carapace. Compressive processes also drove the formation of extensive surface ridges across the flowfield. Continued evolution of the flowfield resulted in the development of a compound flowfield morphology fed by iterative emplacement of breakout lobes. The thermal evolution of flow units was modelled using a one-dimensional finite difference method, which indicated prolonged residence of magma above its glass transition across the flowfield. We compare the estimated apparent viscosity ($1.21–4.03 \times 10^{10}$ Pa.s) of a breakout lobe, based on its advance rate over a known slope, with plausible lava viscosities from published non-Arrhenian temperature-viscosity models and accounting for crystallinity (~50 vol. %). There is an excellent correspondence between viscosity estimates when the lava temperature is taken to be magmatic, despite the breakout being located >3km from the vent, and advancing approximately nine months after vent effusion ceased. This indicates the remarkably effective insulation of the lava flow interior, providing scope for significant evolution of rhyolitic flow fields long after effusive activity has ceased.

1. Introduction

Lava flows constitute the primary emplacement mechanism for erupting magmatic products at the surface of Earth and other planetary bodies. As well as providing valuable information regarding planetary evolution and crust formation, their study is vital for understanding the associated hazard posed to settlements or developments in their proximity (Harris and Rowland, 2001). Lava advance is governed by its rheology, and lava rheology is in turn determined by magma composition, temperature, pressure, crystallinity, and vesicularity, which can differ spatially and temporally during an eruption (*e.g.* Griffiths, 2000). Constraining rheological properties and emplacement behaviour is thus of use both in the interpretation of extant flows and the forecasting of actively emplacing or future flows.

Processes, timescales, and sequence of lava flow emplacement have been inferred from interpretation of solidified flows (*e.g.* Fink, 1983; Anderson and Fink, 1992; Anderson *et al.*, 1998; Applegarth *et al.*, 2010a, b), or estimated using numerical (*e.g.* Young and Wadge, 1990; Favalli *et al.*, 2006; Vicari *et al.*, 2007; Ganci *et al.*, 2012; Spataro *et al.*, 2012), thermo-rheological (*e.g.* Manley, 1992; Stevenson *et al.*, 2001; Wright *et al.*, 2008), or mechanical (*e.g.* Christiansen and Lipman, 1966; Ventura, 2001) models. Here we constrain the evolving flow characteristics of an active rhyolitic lava using ground-based remote sensing and emergent image analysis techniques. Remote sensing (RS) methods have often been used in order to observe and monitor flows either to directly study structures and processes (*e.g.* Fink, *et al.*, 1983; Anderson and Fink, 1992; Guest and Stofan, 2005; Applegarth *et al.*, 2010a) or to derive digital elevation data subsequently used in analysis or modelling (*e.g.* James *et al.*, 2006; James *et al.*, 2007; Tarquini and Favalli, 2011; Dietterich *et al.*, 2012; Ebmeier *et al.*, 2012). The ability to construct digital elevation models (DEMs) of sufficient quality over relevant timescales depends in turn on having a suitable RS acquisition strategy (Ebmeier *et al.*, 2012). Recent progress has been made in extracting data from RS images or image sets in order to estimate key dynamic parameters governing lava emplacement (*e.g.* Harris *et al.*, 2004; James et al., 2007). The capacity to derive rheological data from field-based RS images has a number of advantages over traditional field methods such as penetrometers or shear vanes, which are challenging to operate and provide spatially and temporally limited data due to methodological difficulty or issues with site accessibility (Pinkerton and Sparks, 1978).

The approach used in this study involves a combination of structure-from-motion and multi-view stereo (SfM-MVS) computer vision techniques, which allow the development of three-dimensional (3D) spatial data from photographs collected in the field (*e.g.* James and Robson, 2012). SfM-MVS has been previously used to analyse lava flow (James *et al.*, 2012; Tuffen *et al.*, 2013; James and Robson 2014) and dome (James and Varley, 2012) processes, and offers significant potential for measuring active volcanic processes. Ground-based imaging provides straightforward acquisition with greater spatial and temporal resolution than most satellite or airborne platforms, and is thus well suited for measurement of rapid surface changes associated with ongoing lava emplacement. RS-derived results may then be used in order to obtain basic rheological data regarding lava flows (such as surface velocities or viscosity), for example using the equation of Jeffreys (1925), which relates flow rate (velocity) of a fluid to its intrinsic properties (*e.g.* viscosity, density) and external forces acting on the flow (*e.g.* gravity). Despite being developed to model the two-dimensional laminar flow of water on an incline—requiring the assumption of Newtonian behaviour and well-constrained channel dimensions—the Jeffreys (1925) equation has been commonly used to provide first-order estimates of lava viscosity (among others, Hulme, 1974; Gregg and Fink, 1996, 2000; Hiesinger et al., 2007; Castruccio et al., 2010; Takagi and Huppert, 2010; Chevrel et al., 2013) since first being applied to volcanic processes by Nichols (1939).

The use of Jeffrey's equation—and other models based on Newtonian rheology—implies that there is negligible shear stress acting on a flow if it is not in motion. However, the propensity for cooling lava flows to form a solidified crust overlying viscous lava means that this premise is not necessarily appropriate. Accordingly, rheological models such as the constitutive Herschel-Bulkley relation have been similarly applied to lavas and lava flowfields (*e.g.* Balmforth *et al.*, 2000; Castruccio *et al.*, 2013; 2014) to account for the potential for nonzero shear stresses (corresponding to a yield strength of the crust or core of a lava). These end-member regimes highlight the contrasting theories of "crust-dominated" or "core-dominated" flow (that is, whether flow advance is governed by the rheology of the interior lava or by a thickening overlying crust). Rhyolitic lavas are often posited to have high-yield strength crusts of significant thickness (*e.g.* Fink and Fletcher, 1978; Fink, 1980), serving to retard flow rates by imparting shear on the internal lava. By analysing photo-based reconstructions of an advancing rhyolitic lava, in concert with simple rheological and thermal models, we seek to explore the properties governing the emplacement dynamics of a compound high-silica flowfield.

1.1 Puyehue Cordón-Caulle

The Puyehue Cordón-Caulle Volcanic Complex (PCCVC) comprises the coalesced edifices of Volcàn Puyehue and the Cordón-Caulle fissure system, located at 40.5°S in the Andean Southern Volcanic Zone (SVZ) (Figure 1a). PCCVC is notable in its production of rhyolitic domes and lavas, particularly within the last 100 ka, with significant lava production in the 1921-22 and 1960-61 eruptions (Lara *et al.*, 2006, Singer *et al.*, 2008). For details on the geological history of PCCVC, and a more comprehensive background to the 2011-12 eruption, the reader is referred to Lara *et al*. (2006), Silva-Parejas *et al*. (2012), and Castro *et al*. (2013).

The 2011-12 eruption at Puyehue Cordón-Caulle (PCC) allowed, for the first time, the detailed scientific study of an actively evolving rhyolite flow (Tuffen *et al.*, 2013). A moderate explosive eruption (VEI 4: Silva Parejas *et al.*, 2012) commenced on 4 June 2011, characterised by an initial Plinian column, ballistic explosions, and pyroclastic jetting (Castro *et al.*, 2013). Lava extrusion was observed from 15 June 2011, emanating from the same vents from which the eruption began, initially at a high flux rate (30-80 m^3s^{-1}: Silva Parejas *et al.*, 2012). The source vent, at 40°32' S, 72°08' W, fed an extensive flowfield of volume >0.5 km^3, shown in Figure 1b, which continued to grow even after effusion ceased in April 2012 (Tuffen *et al.*, 2013).

1.2 Flow facies

Two main flow surface types can be identified across the flowfield, the first of which comprises rubbly, 'a'ā-like lava; approximately 30-45 m thick throughout the areas studied, and typical of most of the flowfield (Tuffen *et al.*, 2013). The upper surface is covered in decimetre- to metre-scale blocks, most of which are roughly equant or subangular (Figure 2a). The rubbly lava is generally a light grey colour owing to ashfall and vapour-phase precipitates (Tuffen *et al.*, 2013), and a discontinuous pumiceous veneer approximately 0.5 m in thickness (Figure 2a), with variably oxidised denser lava visible beneath. The margins are bounded by talus, giving the flow a discernible edge between the top and side faces, which assume a generally consistent angle of repose (35-45°). The second surface type is dark grey, brown or black with red oxidised surfaces (Figure 2b), and formed of larger coherent slabs, spines and tongues of lava, with localised torsion and *en echelon* tensional fractures evident. These two different flow facies are hereafter referred to as rubbly and breakout lava, respectively.

1.3 Study areas

During the course of the study, two main regions of the flowfield were investigated, highlighted in Figure 1b as "northern" and "eastern" flowfronts. The former comprises a number of flow units—both rubbly lobes and breakout units—creating a "scalloped" flow margin (Figures 1b; 3a). Constrained by the underlying topography, the majority of this flowfront is abutting against an inward-dipping slope. The eastern flowfront consists of a single unit dominated by a rubbly surface (Figure 1b, 3b). Crease structures, as described by Anderson and Fink (1992), can be seen at both sites, characterised by metre-scale valleys perpendicular to the flow edge, with convexly sloping walls and an apical angle of between 30 and 90° (*e.g.* Figure 3c). On the eastern flowfront, spiny and ensiform structures dominate the upper surface of the flow, as well as large, variably contorted slabs (Figures 3b, d, e). Endogenous features such as tumuli cannot be discerned at either site.

2. Image analysis and 3D reconstruction

Digital field photographs were acquired during 2012 and 2013 during two field campaigns. Initial results using 3D models from the 2012 data (Tuffen *et al.* 2013) are extended here by modelling thermal structure and rheological properties of the flow from the wider dataset. The northern flowfront was imaged on 04 and 10 January 2012 with a Canon EOS450d digital SLR camera and 28 mm fixed focus lens, from a traverse approximately parallel to the edge of the flowfield, with simultaneous handheld GPS logging of photographer position conducted on the later date. The same site was revisited in January 2013 and a comparable image set collected. The eastern flowfront was imaged twice on 11 January 2013, offset by around three hours, also with synchronous GPS logging. The image suites were processed into 3D point clouds using a SfM-MVS freeware package, as described in James and Robson (2012). SfM-MVS reconstructions require image suites of a given object or scene, with different acquisition positions. Feature-matching algorithms identify prominent features of the scene or object (*e.g.* Figure 4a-c) and constructs a sparse (SfM) or dense (MVS) point cloud (*e.g.* Figure 4d), with an arbitrary orientation, scale, and geolocation. In this study, derived clouds were filtered using MeshLab processing software, so as to reduce the amount of noise associated with the SfM-MVS approach (*e.g.* the inclusion of patches of sky).

With the camera clock synchronised to the GPS time, combining the image time-stamps with interpolated GPS logs for the surveys on 10 January 2012 and 11 January 2013 enabled real-world camera coordinates to be estimated for each image. A Matlab tool, sfm_georef (James and Robson, 2012), was then used to determine a scalar, rotational, and translational transform for the corresponding point clouds using the interpolated camera positions as control data. To register the other datasets to these georeferenced models, sfm_georef allows the calculation of 3D coordinates of features identified in images. Thus, static features identified in the georeferenced image suites (such as rocks distant from the flow margin) were matched in the unreferenced sets and used as control points to calculate the scaling and georeferencing transform. Data for 04 and 10 January 2012, and January 2013 for the northern study area are hereafter referred to as N_1, N_2 and N_3, respectively; data from the first and second traverses of the eastern study area in January 2013 are referred to as E_1 and E_2. The root-mean-square error (RMSE) between the GPS-derived camera positions and those in the transformed models, N_2 and E_2, were 4.56 m and 2.06 m respectively. Such values are in line with the expected positional error of the GPS coordinates (Tuffen *et al.*, 2013) and represent the overall uncertainty in absolute geo-referencing. In contrast, the relative registration between sequential 3D models is characterised by RMSE values of 0.22 m (between surveys N_1 and N_2), 0.21 m (N_3–N_2) and 0.09 m (E_1–E_2) These relative errors indicate how well the different surveys within the sequences are registered with respect to each other, with their values indicating that sub-metre changes can be detected with confidence.

Flow movement between image acquisition dates was determined by selecting a number of corresponding points on the surface of the lava that are distinguishable in different datasets, enabling 3D displacement vectors to be calculated. Furthermore, selected regions of the point clouds were interpolated using the kriging method to create DEMs for each of the study areas (*e.g.* Figure 4e). Difference maps were calculated for $N_{1\rightarrow2}$, $N_{2\rightarrow3}$, and $E_{1\rightarrow2}$, by subtracting the earlier surface from the later one in each case.

3. Northern flowfront: Advance, inflation and breakouts

Effusion rate variations, combined with irregular topography, cooling, and crystallisation, mean that the emplacement of a compound flowfield can be expected to be both spatially and temporally heterogeneous. The difference maps of the northern flowfront (Figure 5a, b) highlight this: we observe some regions with relatively more inflation than others, and the areas of maximum inflation or advance are not necessarily the same between image suites.

The flow lobes for which a reasonable number of features (twenty or more features, identified in at least nine images) could be matched between 04 and 10 January 2012 are indicated on Figure 5a. Notably, features associated with a breakout lobe (B1 in Figure 3a) were estimated to have moved a mean distance of 11.72 m in the six-day interval, a surface velocity of approximately 1.95 m day^{-1}. This is significantly different to that of the rubbly units, which moved at 0.65 and 1.37 m day^{-1} (R1 and R2, respectively, in Figure 3a); mean distances of approximately 3.90 and 8.19 m over the same time period. Although the same flow units could be identified in the 2013 dataset (N_3), displacement during this time (12 months) had been too great to reliably identify any of the individual features.

The scalloped margin (Figure 5a, b) is characteristic of compound flowfields: similar morphology can be observed in aerial views of, for example, SP crater in the US, Parícutin (Mexico), or Mt Etna (Italy), indicating iterative emplacement of multiple flow units. Further, the development of surface ridges can be observed, in both time intervals, across the flowfield surface (characterised by inflated arcuate structures and corresponding troughs transverse to the primary flow direction). These ridges are suggestive of compressional processes driven by a thermo-rheological contrast between the relatively hot and viscous lava, and cooler overlying crustal material (Fink and Fletcher, 1978; Fink, 1980). This contrast increases with distance from the vent, causing the upper surface to compress and ruck towards the vent (Lescinsky and Merle, 2005). From a birds-eye-view, these ridges develop a parabolic form; a result of the flowfield spreading laterally away from the vent, and from a velocity differential between the flow margins and the centre (*e.g.* Lescinsky and Merle, 2005). The maximum vertical displacement between N_1 and N_2 (from five metres downwards to ten metres upwards) is probably due to the horizontal translation of these features (Figure 5a).

Apart from the compressional ridges, inflation is mainly discernible at the front of the rubbly and breakout lobes (Figure 5a, b), indicating ongoing lava transport to these areas. Between N_2 and N_3 there was no apparent deflation, although some patches remained at a constant thickness over this time. Maximum vertical difference exceeds 40 m over the year; notably, some of the marginal regions that were actively inflating in January 2012 (Figure 5a) had developed into discrete new breakout units in January 2013 (Figure 5b). Significant flow inflation and formation of new breakouts in the northern flowfront is clearly evident from comparing images from 2012 and 2013 (Figure 6).

The compound morphology of the northern flowfronts indicates that the emplacement was not limited by the supply of lava; an observation supported by ample evidence of flow inflation (Figure 5). However, whether the advance was retarded by cooling-induced viscosity decrease, or as a result of the variable underlying topography, cannot be determined from the difference maps alone. Existence of multiple lobes means that emplacement was spatially and temporally heterogeneous during and after effusion (nonzero values of surface velocity between 2012 and 2013 datasets prove that flow continued after effusion ceased in April 2012). On a broad scale, the evolution of flow type follows the classification of Lipman and Banks (1987), which categorises flows into a channelised zone, well defined by levees approaching the vent; a dispersed zone, where the flow spreads laterally, and a frontal sector where advance is dominated by "rollover". This classification has been used to describe active basaltic (e.g. Kilburn and Guest, 1993; Bailey et al., 2006; Favalli et al., 2010), trachybasaltic (e.g. Loock et al.,2010), and dacitic (e.g. Harris et al., 2004) flows.

4. Eastern flowfront: estimating rheological parameters

Naturally, less deformation is observable at the eastern flowfront, due to the much shorter time interval between data acquisition (approximately three hours, rather than days or months). Between image sets E_1 and E_2, flow features (see Figure 7a) were displaced a mean distance of 0.26 m over the course of approximately three hours, yielding an average advance rate of 2.10 m day^{-1}.

However, closer analysis shows that the identified features in the uppermost third of the flowfront generally moved faster and further than those on the lower two thirds (3.08 versus 1.51 m day^{-1}, respectively). Figure 7a distinguishes between the uppermost and lower flow features. Vertical displacement ranges from approximately seven metres upwards to five metres downwards (Figure 7b), resulting from flow advance or inflation coupled with rockfall from the top of the flowfront (discernible in the relevant image sets, see Figure 3e). These observations are consistent with "caterpillar track" or "rollover" advance typically assumed for 'a'ā lava flows, whereby cooled and fractured surface material moves to the front of the flow before cascading down the frontal face, eventually forming a contiguous rubble or breccia envelope (*e.g.* Rowland and Walker, 1987; Kilburn and Guest, 1993; Harris *et al.*, 2004; Lescinsky and Merle, 2005). Analysis of sequential ALI (Advanced Land Imager) images from the NASA EO-1 satellite indicates that the initial advance rate of the eastern flowfront was around 5 m day^{-1} (Tuffen *et al.*, 2013); the disparity between these rates is probably due to a combination of topography (discussed and shown in Appendix A, the slope decreases notably in this region), and an overall decrease in volumetric flux supplied to this flowfront over time.

4.1 Estimating viscosity from RS data

As the eastern flowfront approximates a channelised flow moving down an incline, we can complement the RS-derived observations with an estimation of the bulk apparent viscosity η_A, using the Jeffreys (1925) equation:

$$\eta_A = \frac{\rho g d^2 \sin\theta}{nU} \quad (1).$$

Lava density ρ is taken as 2300 kg m^{-3} (Castro *et al.*, 2013), g is acceleration due to gravity (9.81 m s^{-2}), U is the maximum surface velocity (3.57×10^{-5} m s^{-1}), and n is an empirical constant thus equal to 2 for flow in wide channel. Slope angles θ between 2.9 and 7.4° are used, and corresponding flow depths d between 31.5 and 27.5 m (the derivation of these values is described in Appendix A). For the purpose of this study, viscosity η is considered equivalent to η_A (as in Hulme, 1974; Stevenson *et al.*, 2001; Harris *et al.*, 2004). The derived range of viscosities is between 1.21×10^{10} and 4.03×10^{10} Pa.s.

4.2 Post-emplacement flow-cooling

Once emplaced, a lava flow will primarily lose heat to the atmosphere by radiation and convection (Griffiths, 2000), whereas heat transport within the flow is dominated by conduction

(Manley, 1992). The Péclet (*Pe*) number defines the ratio of conductive and convective heat transport within the system—*i.e.* the thermal energy conducted within the lava unit versus the convective transport of heat away from the unit—and is determined by $Pe = U/\sqrt{dg}$. The calculated Péclet value for the eastern flowfront is much greater than one (Pe = 1113); as such we may reasonably adopt a simple one-dimensional finite cooling model in order to constrain post-emplacement temperature profiles (Patanka, 1980). The model assumes a flow depth of 30 m, and an initial basal temperature equal to the mean of the eruption and basement temperatures (as in Manley, 1992; Stevenson *et al.*, 2001). Eruption temperature is assumed to be 900°C (Castro *et al.*, 2013). Neglecting the contrasting effects of heat radiation and rainfall-driven advective cooling versus viscous heating, we obtain a first-order estimate of flow cooling over time due to conduction alone. For each timestep, temperature is calculated at nodes every metre into the flow and the underlying basement rock. Boundary conditions are constant, in that the interface between the lava surface and air is 0°C (consistent with local atmospheric temperatures, given the altitude ~1500 m a.s.l.), as is an arbitrary depth in the basement, which represents an unknown depth at which heat will leave the system (*i.e.* due to advection due to groundwater). Lava cools by heat conduction over time (*e.g.* Manley 1996; Gottsman and Dingwell, 2001). The model is of the form:

$$T_{(i)} = \frac{\kappa\, \delta t \dfrac{T_{\mathrm{O}(i+1)} - 2T_{\mathrm{O}(i)} + T_{\mathrm{O}(i-1)}}{\delta z^2}}{1 - \dfrac{Lh}{C_{\mathrm{P}}\rho}} + T_{\mathrm{O}(i)}\ ;\ C_{\mathrm{P}} = \frac{k}{\kappa\rho} \quad (2),$$

where $T_{(i)}$ is calculated temperature at each vertical node i, T_{O} is the temperature at the previous timestep, and δt and δz are the intervals for the timestep and vertical node spacing, respectively. Table 1, below, shows the definition and values of the model parameters.

Consecutive satellite images of the advancing eastern flow (Tuffen et al., 2013) show that the advancing lava in the region was emplaced after 01 November 2013, *i.e.* in a timeframe ≤74 days prior to data acquisition. Accordingly, Figure 8a-g shows the likely temperature profiles through the flow, over time since emplacement. If we assume that the glass transition Tg of the melt phase occurs at 10^{12} Pa.s (Giordano et al., 2008; Hui et al., 2009) then a temperature of around 710°C can be taken as an approximate threshold for solidification, according to the models of Hess and Dingwell (1996) and Zhang *et al.*, (2003), using glass oxide fractions derived from the eastern flowfront (Schipper *et al.*, 2015). Thus the thickness of the solidified crust of the flow increases with time (shown in Figure 8). Our model indicates that after a cooling period of two and a half months (Figure 8b), cooling-induced solidification of the flow has only penetrated the uppermost two - three metres of the flow at the eastern flowfront. Within the flow, the majority of the rest of the lava remains close to the initial eruption temperature, being around 830°C at the base of the flow profile, and near 900°C in its centre. Thus the solid three metre crust is overlying approximately 27 m of lava still nominally above its glass transition (*i.e.* able to flow). Our model further indicates that after four years, a 30 m thick rhyolitic lava flow will be entirely below Tg, and thus completely stalled. Despite this simplified model of flow cooling, other factors can prolong the mobility of the lava (*i.e.* longer than four years), such as flow down an incline, reactivation of the flow units due to subsurface supply of relatively hotter lava, or reactivation due to flow unit superposition (*e.g.* Applegarth *et al.*, 2010b).

4.3 Comparing RS-derived viscosities to non-Arrhenian models

In this section, we compare our RS-derived values to those of three published non-Arrhenian temperature-viscosity models (Hess and Dingwell, 1996; Zhang *et al.*, 2003, and Giordano *et al.*, 2008). These models assume a single-phase medium (*i.e.* melt viscosity only). However, recent work (Schipper *et al.*, 2015) indicates that the crystal fraction of lava from the eastern flowfront is approximately 50 vol. %. Using the modified Einstein-Roscoe equation (*e.g.* Pinkerton and Stevenson, 1992; Crisp et al., 1994) we can therefore estimate the influence of the crystal fraction ϕ on the effective viscosity of the lava η, whereby

$$\eta = \eta_0 (1 - R\phi)^{-q} \quad (3),$$

where η_0 is the calculated viscosity of the melt (Hess and Dingwell, 1996; Zhang *et al.*, 2003; Giordano *et al.*, 2008), and R and q are constants equal to 1.67 and 2.5, respectively. We acknowledge that the Einstein-Roscoe equation is underpinned by some basic assumptions that inherently simplify the influence of crystallisation on lava viscosity. Chief among these is the supposition that crystal growth is isotropic (*i.e.* spherical), which governs the R term (Marsh, 1981). The intricacies of the crystal cargo of the PCC lavas—such as the mean aspect ratio and the maximum packing fraction (e.g. Mueller et al., 2010, 2011; Mader et al., 2013; Le Losq et al., 2015)—remain open to a systematic petrographic study. Nonetheless, we observe an excellent coincidence between our estimated range of viscosities (from 1.21×10^{10} to 4.03×10^{10} Pa.s) and the modelled ranges (shown in Figure 9), suggesting that the assumptions are not disproportionate.

5. Emplacement summary and implications of study

Apparent viscosities calculated from emplacement dynamics of the eastern flowfront correspond well with those derived from the models of Hess and Dingwell (1996), Zhang *et al.* (2003), and Giordano *et al.* (2008), falling within uncertainty (~0.3 log units of viscosity) in the same T-η space after accounting for the influence of the crystallinity of the PCC lavas (Figure 9). This excellent correlation between the RS-derived and modelled viscosities suggests that—despite their simplicity and attendant assumptions—equations 1 and 2 may be used in conjunction to determine a first-order estimate of thermo-rheological properties of advancing silicic lava. Significantly, this implies that, at least in the initial stages of emplacement of any given flow lobe, the advance rate is not notably influenced by an overlying cooled crust. At the time of data acquisition on the eastern flowfront, the degree of cooling had been insufficient to form a surface crust capable of significantly impeding flow advance. This observation agrees with flow textures and breakout emplacement processes modelled using analogues (*e.g.* by Lescinsky and Merle, 2005).

Crustal control is favoured by long-lived eruptions with relatively low effusion rates, and prolonged cooling of thick lava units (Castruccio *et al.*, 2013). With a longer cooling interval, a high yield-strength crust can develop, increasing in thickness in line with $\sqrt{\kappa t}$ (Figure 8). The existence of compressional flow ridges across the northern flowfront attests to this: although the flow interior can retain heat and flow viscously, advance rates are retarded by the thickening crust (Castruccio *et al.*, 2013). The implication that the eastern flowfront initiated as a breakout at (or very close to) the estimated eruption temperature highlights the remarkable insulation of subsurface lava throughout the flowfield.

Consistent with Walker's (1971) definition of compound flows, the PCC flowfield is divisible into individual units, with breakout development appearing to be an iterative process whereby new lobes are extruded viscously and limited in volume by topography and cooling. Those that do persist evolve towards rubbly facies, as the propagation of tensile fractures creates a nascent talus layer (Tuffen *et al.*, 2013). The cooling-driven viscosity increase in the uppermost portion of the flow is reflected in Figure 8, as predicted by Equation 2.

The features and inferred emplacement of lava breakouts at PCC have many parallels with those observed at basaltic-intermediate flowfields. For example, blade-like and spiny structures are reminiscent of late-stage lava extrusion in low-silica compound flows, termed "squeeze ups" (Applegarth *et al.* 2010a). Although transport time between the main and ephemeral vents (*i.e.* the breakout points) increases as effusion rate dwindles and the flowfield expands, we do not observe a notable increase of cooling and crystallisation of lava in later breakouts (samples from breakouts in 2012 and 2013 both yielded a crystal content of approximately 50 vol. %: Schipper *et al.*, 2015). The observed features generally attributed to significantly higher yield-strength lavas—such as slabby lava (*e.g.* Guest and Stofan, 2005)—are therefore not necessarily primarily induced by cooling. Rather, it is likely that many of these features arise because of flow stagnation due to the pre-eruption topography of the flowfield, thus increasing the ratio of effusion to advance rate (Guest and Stofan, 2005).

Similarly, the abundance of breakouts at the northern flowfield may be explained by the underlying topography: as the flowfronts in this region abutted against a topographic barrier, the advance rate decreased. Continued supply of lava through subsurface thermal pathways has been discussed with respect to basaltic flowfields (*e.g.* Calvari and Pinkerton, 1998; Anderson et al., 1999; Guest and Stofan, 2005) and modelled using wax analogues (*e.g.* Anderson *et al.*, 2005). Interior thermal pathways can apply volumetric stress over large areas of a flow, resulting in spatially extensive inflation and deformation by delivering relatively hotter, less viscous lava to the flowfront or margins (Anderson *et al.*, 2005). A transient or sustained subsurface lava supply to a stagnant lobe can result in overpressure, inflation, and consequent breaching of the solidified crust as a breakout from an ephemeral vent (Hon *et al.*, 1994); indeed, DEM difference maps (Figure 5) indicate that breakout emission (*e.g.* Figure 6) is typically preceded by a period of inflation. Usually, the precise location of a breakout cannot be predicted, though it is empirically evident that it will be at a point of relatively greater stress: here, difference mapping provides a tool for identifying potential breakout areas. Lava extruded at such a breach will do so initially without a thick crust, as discussed. Until a cooling-induced crust develops on these flow units, breakout lava will be subject to distinct shear regimes to the rubbly lava, reflected in the contrasting surface structures observed in the 'a'ā and breakout flow facies (Figure 2). Thus, the governing rheology of silicic lavas may transition from being core-dominated, as inferred for the breakout lobes at PCC, to being controlled by the thickening crust, as we can infer from the compressional processes evident across the flowfield, particularly in the northern study area. This observation is not dissimilar to the frequently observed transition from pāhoehoe to 'a'ā-type lavas in basaltic systems (*e.g.* Cashman *et al.*, 1999; Soule *et al.*, 2004). In turn, this supports the inference that flow morphology may be described in a cross-compositional continuum, whereby the evolution of a lava flow or flowfield is a function of the competing influences of internal viscosity (governed by cooling rate, crustal growth, and crystallisation) and advance rate (governed by effusion rate and underlying topography).

Furthermore, as "squeeze-ups" are thought to develop on halted flow units (Applegarth *et al.*, 2010a), the existence of these features on the eastern flowfront highlights that the breakout occurred from a flowfront or lobe that was halted for a time, before being reactivated. We attribute the remarkable mobility calculated for the eastern flowfront to the efficient thermal insulation between the primary vent and the ephemeral breakout vent, after which point it flowed down an incline. This shows that despite low inferred effusion rates and high apparent viscosities, rhyolitic lavas can evolve considerably after initial stagnation, in agreement with Tuffen et al. (2013). This process is facilitated by highly effective heat retention by the brecciated material of the flow surface insulating the hotter and less viscous lava beneath: indeed, our cooling model—though simple—indicates that the innermost portions of the flow could comprise lava hot enough to flow (*i.e.* above T_g), even three years after effusion. In regions of the flowfield where the lava is thicker, this timescale is greatly increased; for example, the model predicts that a lava flow 40 m thick could retain sufficient heat that there would be lava still nominally above the glass transition of its melt phase up to six years after emplacement. Given the degree of displacement we observe in the northern flowfront (Figures 5a and b) there is ample evidence of lava in the flowfield greater than 40 m in thickness. Thus there remains potential for significant spatial evolution of the flowfield, even years after emplacement.

Many of the emplacement processes observed at PCC bear similarity to those described for andesitic, dacitic, and basaltic lava flowfields; for example Mt Etna, Italy (*e.g.* Kilburn and Guest, 1993; Bailey *et al.*, 2006), and Santiaguito, Guatemala (*e.g.* Harris *et al.*, 2004). The existence of cross-compositional features such as crease structures, slabby lava, and breakouts further indicates that compound flow morphology may be described by flow models that encompass rheological differences of many orders of magnitude and suggests the universality of flow models such as those of Walker (1971) or Lipman and Banks (1987). This interpretation is supported by the analogue experiments of Fink and Griffiths (1998). These authors conclude that lava flow morphology evolves sequentially, in a manner dictated by the ratio of cooling and advance rates rather than discrete compositional differences.

We suggest that the SfM-MVS techniques could be used to improve flow prediction models by facilitating targeted DEM generation and thus highlighting regions of subsurface supply, inflation and potential hazards. SfM-MVS was found to yield valuable spatiotemporal information over an interval of days to weeks, although useful data were also gained over longer (months) and shorter (hours) timescales. Furthermore, the effects of crystal fraction and surface crust on the apparent viscosity is an area that entreats future research, which may be undertaken by way of scaled analogue models as well as field observation and high-temperature rheological experimentation on lavas.

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6. Conclusions

Rhyolitic lava flows from the 2011-2012 Cordón-Caulle eruption were found to emplace by processes comparable to those observed in compound flows of less silicic lavas. After an initial period of simple channelised rubbly flow, the lava progressively stagnated, probably primarily due to topographic barriers to flow advance. Lateral extension of the rubbly flowfield was accompanied by spatially and temporally heterogeneous vertical inflation, determined by topography and localised subsurface supply, plus compression and the formation of surface ridges. Continued effusion fed a compound flowfield defined by breakout lobes, some of which matured over time to resemble nascent rubbly units. The apparent viscosity of the last-advancing breakout lobe, as estimated from a simple

Newtonian flow model (1.21 – 4.03 × 10^{10} Pa.s), tallies closely with viscosity estimates based on breakout composition. This suggests that, despite advancing nine months after effusion ceased, and >3 km from the vent, this breakout lava remained close to eruption temperatures and was initially governed by internal viscosity, rather than crustal retardation. The highly effective thermal insulation of this rhyolitic lava yields the potential for significant flowfield evolution—for example breakout initiation, compound flow development, and lateral spreading—even years after the cessation of effusion at the vent. Marked parallels between inferred low- and high-silica processes suggest that compound flow emplacement may be described by universal, cross-compositional models.

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Table 1: Definition and units of parameters used in Equation 2 and throughout text, as well as sources for values pertaining to rhyolitic lava.

Term	**Definition**	**Units**	**Value and source**	
C_P	Heat capacity	$J\ kg^{-1}\ K^{-1}$	1185.77	(Equation 2)
k	Thermal conductivity	$W\ m^{-1}\ K^{-1}$	1.5	(Romine et al., 2012)
Lh	Latent heat	$J\ kg^{-1}$	$5x10^5$	(Fagents and Greeley, 2001)
κ	Thermal diffusivity	$m^2 s^{-1}$	$5.5\ x\ 10^{-7}$	(Romine et al., 2012)
ρ	Bulk density	$kg\ m^{-3}$	2300	(Castro *et al.*, 2013)

Appendix A: Model parameter estimation

1. Slope

As with any models, the reliability of Jeffreys equation (Equation 1) and the cooling model (Equation 2) depend on the quality of the input parameters.

In order to constrain the incline angle of the underlying topography at the eastern flowfront, elevation data from prior to the eruption (April 2011) was used. These data were obtained from Google Earth, a free geographical information program which comprises an amalgamation of elevation data, primarily collected by NASA's Shuttle Radar Topography Mission (SRTM). Figure A1 (a) shows the eastern site pre-eruption. Slope profiles were then extracted with reference to six transects running the length of the eastern flowfront (Figure A1, c). The elevations corresponding to the start (h_{MAX}) and finish (h_{MIN}) of each transect are given in Table A1, as are the length of each transect and the corresponding slope value, determined by $\theta = \tan^{-1}\left((h_{\mathrm{MAX}} - h_{\mathrm{MIN}})/l_{\mathrm{T}}\right)$.

Table A1: slope profile data for the eastern flowfront (pre-eruption).

Path	**Maximum elevation h_{MAX} [m]**	**Minimum elevation h_{MIN} [m]**	**Distance l_{T} [m]**	**Slope angle θ [°]**
1	1340	1330	200	2.9
2	1343	1325	250	4.1
3	1348	1327	250	4.8
4	1349	1325	250	5.5
5	1354	1327	250	6.2
6	1360	1334	200	7.4

2. Lava thickness

Accounting for the basal slope and the distance l between the two points (a and b, Figure A2) gives us an estimate of the flow depth d, approximated by $d = d_{\mathrm{T}} - (\tan\theta\, l)$, where d_{bc} is the total difference between the top and base of the flow (the difference between *b* and *c* in Figure A2). The determined range of flow depths (from 27.5 to 31.5 m) has been incorporated into the thermal and rheological model estimations in the main body of the text (*i.e.* Equations 1, and 2).