SiO$_2$–rich lava complexes: Textures in SiO$_2$-rich lava complexes

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Abstract. Silicic lava flows and domes show very specific internal zonation of lava textures. The flow stratigraphy consists of 1) finely vesicular pumice carapace underlain by 2) obsidian layer, 3) coarsely vesicular pumice and 4) a second obsidian layer with lithophysae and spherulites. 5) The central part is represented by the crystalline rhyolite, which in turn may be underlain by a basal obsidian layer with lithophysae and spherulites. Beneath these layers sometimes occurs coarse vesicular pumice followed by the basal breccia at the bottom of the flow stratigraphy. This textural zonation in SiO$_2$-rich lava complexes results from the cooling rate, migration of water vapor, lava rheology, devitrification, degassing, autobrecciation.

Introduction

The study of textures of voluminous silicic units involves understanding the development of magmatic bodies. Many different processes influence textures. Thus textures reflect the history of expansion and evolution of silicic lava flows and domes and show under which conditions these bodies were formed. Textures in volcanic silicic units show very complex structure. Since in ancient volcanic zones the distinction between SiO2-rich lavas and high grade ignimbrites is difficult, textures are of particular interest. They help to reconstruct the evolution of old volcanic complexes even so big parts of them often have been altered.
Textures

Textures are distinguished on the bases of crystallisation, vesiculation, foliation, brecciation, color, specific gravity and their relation to each other. One of the main factors determining the texture of an igneous rock is the cooling rate, which describes the decrease of temperature per time unit. The viscosity of silicic melts changes with temperature and the viscosity defines the diffusion rate, which is based upon the time in which atoms or molecules diffuse through the liquid. Other factors influencing textures are the rate of nucleation of new crystals which represents the formation of an embryonic crystallite from a melt and the rate of growth of crystals, which is the rate at which new constituents can arrive at the surface of the growing crystal. This depends largely on the diffusion rate of the molecules.

Fig. 1. (A) Schematic cross-section through a subaerial silicic lava flow. The left side shows the internal textural variations arising from vesiculation, devitrification and flow fragmentation. The right side shows the orientation of internal flow foliations, and crude layering in flow margin talus breccia. (B) Vertical section through the flow at the position indicated in (A), showing the major textural zones. (modified from Pink and Manley 1987; Duffield and Dalrymple 1990; after McPhee et al. 1993)
There is a large variety of textures to be found in SiO$_2$-rich volcanic rocks.

After MANLEY and FINK (1987) the uppermost part of the SiO$_2$-rich lava succession is represented by the carapace breccia (the top breccia) (Fig. 1) which is formed due to autobrecciation of the rigid lava top (auto-fragmentation due to movement of liquid lava breaking up solidified lava). This breccia is made up primarily by fine vesicular pumice, although debris of the upper obsidian layer and coarsely vesicular pumice layers is locally included. As the flow advances, this breccia cascades from the upper surface and mixes with debris from the fragmenting flow front. This heterogeneous talus breccia is then overridden by the body of the flow, becoming a basal breccia layer (MANLEY and FINK 1987).

The carapace breccia is underlain by finely vesicular pumice (FVP) (Fig. 1) which develops due to the “second boiling” process (MANLEY and FINK 1987). After TAYLOR ET AL. (1983), silicic magma looses most, but not all of its dissolved water vapor at a depth of 1 km or less (“first boiling”) retaining only 0.1 to 0.2 wt. % H$_2$O. During effusion, degassing occurs again in the uppermost part of the flow, where overburden pressure is insufficient to overcome the vapor pressure of the remaining dissolved water creating finely vesicular pumice with vesicles, whose size decreases downward toward a glassy obsidian layer (MANLEY and FINK 1987). The FVP layer ends where the load pressure, exerted by overlying material, is bigger than the vapor pressure within the bubbles. Cooling of the upper surface also inhibits crystallization and prevents bubbles from growing very large. Continued cooling at the surface and movement of the flow may cause large fractures to propagate down through this brittle crust to the depth, at which the flow remains ductile (FINK and MANLEY 1987).

Towards the center of the lava flow the FVP layer is followed by the obsidian layer (Fig. 1). Obsidian is a dense non-vesicular acid volcanic rock with a glass content >80 % and less than 1 % water (SHELLEY 1993; BREITKREUZ 2001). The obsidian shows varying abundance of microlites, which are tiny crystals of Fe- and Ti-rich mafic or felsic silicates that form during the early stage of cool-
ing at temperatures of 800 – 600 °C (Fig. 2), above the glass transition temperature \(T_g\). (On the \(T_g\) melts changes from a viscous to hard and relatively brittle condition). (Breitkreuz 2001).

**Fig. 2.** Processes of texture formation in SiO\(_2\)-rich volcanics during ascent, eruption, emplacement and cooling. Glass in the first instance displays ductile deformation and reacts on stress with brittle deformation (fragmentation) below the glass transition temperature \(T_g\). However, brittle vs. ductile deformation is also dependent on deformation rate, e.g. glass with a temperature above \(T_g\) can break during swift deformation. \(T_g\) approximates two thirds of the liquidus temperature \(T_m\) (modified after Lofgren 1971; Eichelberger et al. 1986; Swanson et al. 1989; Davies and McPhee et al. 1996; Manley 1996; McArthur et al. 1998; taken from Breitkreuz 2001).

Very often a highly vesiculated zone can be found in the middle of the obsidian layer. Crystallization of the underlying rhyolite causes the release of dissolved magmatic volatiles, which migrate upward due to microcracks at the bottom of the obsidian layer and accumulate below the already rigid upper crust. This region is characterized by an alternation of centimeter-scale layers of obsidian with centimeter to meter-scale layers of coarsely vesicular pumice (CVP). The vesicles range from angular and very contorted to spherical shape from mm to meter in diameter. (Manley and Fink 1987). Continued crystallization of the flow interior leads to a thickening of the gas-rich layer as well as to a lowering of its density. The density inversion between the obsidian and
CVP layers (Fig. 3) provides the basis for a gravitational instability, in some cases allowing it to rise buoyantly through the overlying obsidian and FVP layers. If the rate of ascent is high enough, relative to the forward velocity of the lava and the cooling rate, then these diapirs could reach the flow surface (Fig. 4) (Fink 1980a, 1983; Fink and Manley 1987).

Fig. 3. Rhyolitic obsidian flow profiles. (A) Stratigraphy for a typical 35-m-thick rhyolitic obsidian flow. (B) Density profile, based on measured densities of samples of coarse pumice, obsidian, and fine pumice. Note density inversion at contact between obsidian and coarse pumice. (C) Temperature profile, assuming constant internal temperature. Note steep surface-temperature gradient. (D) Viscosity profile, based on (B), (C), and laboratory viscosity measurements. Note rapid decrease of viscosity with depth near the upper surface (after Friedman and others 1963, from Fink 1983).
Further cooling can lead to partial or complete devitrification of the glass. Quench crystals (mainly tridymite, cristobalite and/or feldspar) grow to spherulites, often using a crystal, a bubble or a crack as a nucleation surface (Breitkreuz 2001) (Fig. 5). After McPhie et al. (1993) spherulites are a...
characteristic product of the high-temperature (700 – 450 °C) (Fig. 2) devitrification of natural glass consisting of radiating arrays of crystal fibres. Each fibre is a single crystal that has only slightly different crystallographic orientation from adjacent crystals. Spherulite morphologies depend upon their formation temperature and the focus of nucleation typically have diameters of 0.1 to 2 cm. Outlines of spherulites are often irregular, while bow-tie spherulites consist of two conical bundles of fibres joined at their apices. Plumose spherulites are open, coarse and commonly fan shaped. Fibres in axiolitic spherulites radiate from a line (Fig 5). Lithophysae are special spherulites forming at an early stage in the cooling history that have a central vug (Fig. 5). They can reach a few tens of centimeters (McPHEIE ET AL. 1993). Generally the amount of spherulites increases toward the devitrified rhyolite zone. In the lowermost part of the obsidian layer, internal breccia of monolithologic composition, characterized by fragmented and rotated obsidian blocks up to 10 cm in diameter, set in a pulverized, dusty matrix can occur (MANLEY and FINK 1987).

Fig. 5. (A) Axiolitic structure; spherulitic fibres radiate from a plane. (B) Fan spherulite with fibres radiating from a point. (C) Bowie (or “wheat sheaf”) spherulite. Two fan-like arrays are joined at their apices. (D) Plumose spherulite showing extensive side branching. Unlike dendrites, branching does not occur on crystallographic axes. (E) Spherical spherulite. (F) Pectinate texture defined by fine axiolites growing inward from the walls of a juvenile pyroclast, in this case a tricuspate shard. Scale will vary according to the size of the fragment. (G) Lithophysae structure with fibres radiating outward from a central hollow. (H) Lithophysae with concentric hollows arranged parallel to the crystallization front (modified from LOFGREN 1974; after McARTHUR ET AL. 1998 taken from BREITKREUZ 2001).
If external water has access to remaining glass during final cooling and/ or early diagenesis, hydration takes place. Volcanic glass incorporates up to about 3% of water into the amorph tetraeder framework. If this process is accompanied by rapid cooling, characteristic cracks often marked by clay minerals can develop (perlite cracks) (Breitkreuz 2001). In classical perlite, the cracks are distinctly arcuate and concentrically arranged around spherical, non-hydrated cores, while strongly flow-banded glassy lava forms a roughly rectilinear network of perlitic fractures, comprising cracks that are subparallel and strongly oblique to the banding (banded perlite) (Fig. 6, 7).

**Fig. 6.** Classical perlite in thin-section. The glassy groundmass of this rhyolite shows classical perlitic fractures, comprising arcuate, overlapping and intersecting cracks (arrow). Sparse phenocrysts of clear plagioclase and brown biotite, and faint flecks in the glass (feldspar microlites) define a subtle flow lamination, which is overprinted by the perlitic fractures. Plane polarized light. *Glassy rhyolite, Pleistocene; Cala de Gaetano, Ponza, Italy (McPhie et al. 1993)*

**Fig. 7.** Banded perlite in thin-section. The formerly glassy groundmass of this strongly flow-foliated dacite contains rectilinear fractures (arrow) typical of banded perlite. The fractures are preserved by a thin, infilling layer of sericite and chlorite. Most of the glass has recrystallized to a mosaic of quartz and feldspar; some has been replaced by sericite and chlorite. Plane polarized light. *Mount Read Volcanics, Cambrian; Pieman Road, western Tasmania (McPhie et al. 1993)*

In the center of the profile develops the crystalline rhyolite (RHY) which shows well-developed flow foliation, consisting of alternating submillimetre-scale layers of crystal-rich and crystal-free lava. The crystal-rich bands contain
quartz and feldspar as well as numerous angular vesicles which are absent from the aphyric bands. The vesicles constitute up to 20% of the volume of the layers (Fink and Manley 1987). After Manley and Fink (1987) the vesicularity in the RHY layer develops by the local increase of water vapor pressure around spherulites crystallizing preferentially in specific flow layers. The quartz and feldspar that make up the spherulites are anhydrous. As a spherulite crystallizes, $H_2O$ diffuses away and becomes concentrated in a halo around the spherulite (Friedman and Long 1984; after Manley and Fink 1987) (Fig. 3 (H)). Crystallization occurs preferentially in the central part of the flow, because the ability of ions to migrate is higher in this zone of higher temperature and thus lower viscosity (Fink and Manley 1987). The shear stress associated with flow advance leads to a flow foliation of the crystal-rich regions. Crystallized rhyolite shows spherulitic, micropoikilitic and granophyric textures. Micropoikilitic (snowflake) texture consists of small (<1mm), commonly irregular crystals of one mineral that completely enclose even smaller crystals of another mineral. Micropoikilitic texture, comprising optically continuous quartz that encloses laths or speruliths of feldspar, is especially common in rhyolites. It results from initial devitrification of cooling glass (McPhie et al. 1993 or after Lofgren 1971b) through primary devitrification, especially in glasses that have relatively high water contents or are cooled (or reheated) slowly (Fig. 8). The granophyric texture consists of fine equigranular quartz and feldspar. The presence of an internal breccia in the RHY zone implies the exceed of the lava’s tensile strength through shear stress during the flow, causing it to fracture into blocks. These blocks can be subsequently rotated further and fragmented. The fracturing may occur due to the concentration of the shear stress in specific horizons or lowering of the tensile strength by devitrification. Internal breccias should show a very restricted lithological character (Manley and Fink 1987).
Under the RHY layer, the flow stratigraphy consists of basal obsidian with lithophysae, whose content decreases downward. The obsidian layer may be in turn underlain by coarsely vesicular pumiceous lava, which may rise in response to temperature increase, due to the shear stress in the moving flow (NELSON 1981; after McPHIE ET AL. 1993).

The deepest part of a flow is represented by the basal breccia, which in comparison with the surficial breccia, primarily composed of the FVP unit, the basal breccia will show more interstitial pulverized ash and a greater range of clast textures, possibly including crystalline lava (MANLEY and FINK 1987).

**Flow layering**

The flow banding in a silicic dome or flow reflects how the lava moved during its emplacement. Lava flows commonly contain a zone of subhorizontal foliations extending from the base to within several meters of the upper surface (FINK and MANLEY 1987). Flow layering is different in terms of its dependence on the distance from the vent. Near the vent flow layering is subvertical throughout the entire cross section. FINK (1983) (Fig. 9) has interpreted these relations as being the combined results of flow through a conduit and along the ground. In response to shear stress along the walls during magma rising through a conduit, vertical foliation develops. Emerging lava rises vertically above the vent, rifts, and then spreads laterally, inheriting nearly vertical layering. As the lava moves outward, shear stresses at the flow base can cause the flow layering to rotate to a horizontal state which propagates upward through the flow as it
advances. This rotation can only occur in those parts of the flow that behave plastically. The cooler, brittle crust retains its original vertical foliation and is passively transported by the flowing material beneath. The transition from horizontal to vertical layering occurs at the top of the CVP layer (after Fink 1983; Fink and Manley 1987). The foliation patterns on a flow can be exceedingly complex due to the interaction of diapirism, folding, and fracture. Foliation patterns are one of the most useful criteria for identifying pumice diapirs which generally define the domal or anticlinal shape of the outcrops. The preferred orientation of foliations in the domes lies parallel to those directions that correspond to either a local or regional structural trend (Fink 1983).

Fig. 9. Schematic diagram showing development of foliation attitudes in vent area of dome. (1) Viscous dome emplaced. Shallow surface fractures develop. (2) Fractures nearest center deepen preferentially. (3) Fractures propagate inward as lava spreads laterally, causing most of upper surface to become a fracture surface. (4) Later stage of growth. Flows have developed. Most of flow still capped by fracture surface. Compression during flow forms surface folds. Flow stratigraphy not indicated. (5) Detail of vent area showing uplift and outward rotation of blocks as lava continues to rise (after Fink 1983).
SUMMARY

The model presented above describes conditions and processes that may form the various textures in silicic lava flows and domes. The principal processes involved are: vesiculation, crystallization, fracturing, migration of exsolved gases and lava movement. The sequence of lava textures is almost the same in all flows, but differences in the specific geometry or physical properties of a flow may lead to a variation of the thicknesses of the variously textured layers or to the absence of certain layers. The surface structure in turn, results from three deformational processes: a) compression during advance of the lava, b) fracturing due to cooling and radial expansion of the flow, and c) diapirism caused by the presence of a light pumiceous base beneath the non-vesicular core of the flow.

References

Breitkreuz C. (2001) Introduction to physical volcanology and volcanic textures. Short course