Emplacement and rheomorphic deformation of a large, lava-like rhyolitic ignimbrite: Grey’s Landing, southern Idaho

Graham D.M. Andrews† and Michael J. Branney
Department of Geology, University of Leicester, Leicester, LE1 7RH, UK

ABSTRACT

The Miocene Grey’s Landing ignimbrite reaches 70 m thick and covers at least 400 km² in the central region of the Snake River Plain. It shows particularly intense welding and rheomorphic deformation, and although parts are eutaxitic, most is lava-like with flow-banding and no flame. A near-ubiquitous penetrative flow lamination, associated with a well-developed elongation lineation, is folded into small intrafolial tight to isoclinal oblique and sheath folds, which are refolded by larger folds in the upper parts. Structural and kinematic analysis reveals that welding and early deformation occurred rapidly during deposition from a very hot (<1000 °C), high-mass-flux pyroclastic density current that flowed westward across a graben-faulted landscape. As hot particles were deposited, they rapidly agglutinated and coalesced, and underwent noncoaxial shear in a subhorizontal ductile shear zone close to the current-deposit interface. The shear zone is interpreted to have been less than 2 m thick. It produced and deformed the rheomorphic fabric, and it migrated upward with the rising current-deposit interface during aggradation, so that it transiently affected all levels of the resultant thick ignimbrite. Deformation was progressive, and after the density current had dissipated, viscous spreading and downdipe flow continued and involved an increasingly thick portion of the sheet. This folded the flow banding and F1 intrafolial isoclines into larger shear folds, and into more upright periclines near the top of the ignimbrite. We demonstrate that structural and kinematic analysis can elucidate the emplacement history of rheomorphic ignimbrites.

INTRODUCTION

Rheomorphic ignimbrites are the deposits of very hot (>900 °C) pyroclastic density currents that have undergone intense welding and ductile flow prior to cooling through the brittle-ductile transition (Schmincke and Swanson, 1967). They occur abundantly in parts of the geologic record, from Precambrian to Holocene (Branney et al., 2008), and they record a particularly devastating style of high-temperature explosive volcanic eruption in which a sheet of hot volcanic glass is fused regionally across a landscape. The hot glass may continue to spread and flow downslope for weeks to months after the eruption. Eruptions of this type have not been witnessed historically, and our understanding derives primarily from the geological record.

Rheomorphic ignimbrites occur in diverse settings ranging from continental arcs (e.g., Branney et al., 1992; Kokeelaar and Moore, 2006) and rifts (e.g., Mahood and Hildreth, 1986) to intraplate ocean-island volcanoes (e.g., Schmincke, 1974). They are particularly abundant in provinces that record “Snake River–type” volcanism, a distinctive type of continental rhyolitic volcanism characterized by: (1) extensive layers of stratified ash rather than typical Plinian pumice-fall layers; (2) unusually long and large-volume rhyolite lavas rather than typical small rhyolite domes and coulees; (3) ignimbrites that are better sorted and contain less pumice lapilli (or flame) than is typical of ignimbrites elsewhere; and (4) ignimbrites that are more intensely rheomorphic and predominantly lava-like than is typical elsewhere, where most meta-luminous rhyolite ignimbrites are only partly welded and eutaxitic and where rheomorphism is often associated with strongly peralkaline compositions (Mahood, 1984). The importance of Snake River–type volcanism, with its large lavas and lava-like ignimbrites, is perhaps not yet fully appreciated: features of Snake River–type volcanism also occur within volcanic successions of various ages and settings elsewhere (e.g., Trans-Pecos Texas—Henry et al., 1988; Etendeka-Parana—Müller et al., 1995; Lebombo Rift, Karoo—Cleaver, 1979; Keweenawan volcanics of Minnesota—Green, 1989; Gawler Range volcanics of Australia—Allen et al., 2003) and probably reflect eruption processes that differ from those of more conventional, better-understood rhyolitic eruptions.

We present the first structural and kinematic analysis of a Snake River–type rheomorphic ignimbrite, and we selected an example from the type area, the Miocene succession of the central Snake River Plain on the Yellowstone hotspot track, United States (Malde and Powellers, 1962; Branney et al., 2008). It preserves an incredible suite of rheomorphic fabrics, including elongation lineations, folds, shear folds, and kinematic criteria that form the basis of our structural analysis. Our analysis reveals that welding in this ignimbrite began very rapidly, and that the deformation was progressive, initially partitioned into a thin, upward-migrating shear zone, and later across a much thicker portion of the cooling body.

Structurally, the deposit provides an excellent opportunity to examine the effects of noncoaxial ductile deformation, and rapid development of transposition and shear folding. Unlike most tectonic ductile shear zones, in which strain becomes increasingly localized (Ramsay, 1980), we infer the migration of a shear zone through a thickening body of rock. Another factor of interest is that rheomorphic ignimbrites differ from tectonic shear zones because they are noncrystalline at the time of deformation, and they do not develop conventional cleavage. However, their rheology is in some ways easier to constrain than crustal shear zones (cf. Talbot, 1999); they occur on Earth’s surface, form very quickly, have simple compositions, and do not undergo complex crystal-boundary processes (e.g., diffusion creep, grain-boundary sliding; Passchier and Trouw, 1998).
**Welding and Rheomorphism**

Welding is the adhesion and compaction of hot pyroclasts, and it ranges from slow to moderate “load welding” (Freundt, 1998; Quane et al., 2009) during the cooling and compaction of a hot, thick ignimbrite sheet, to much quicker (almost instantaneous) agglutination of more fluidal pyroclasts during deposition and prior to significant loading (Mahood, 1984; Branney and Kokelaar, 1992). Rheomorphism is manifested by the presence of ductile folds and elongation lineations. Most welded ignimbrites are not rheomorphic (Ross and Smith, 1961), some are rheomorphic only locally, such as at fault scarps (e.g., Long Top Tuffs; Branney and Kokelaar, 1994) or in a particular welding zone (e.g., Rattlesnake Tuff; Streck and Grunder, 1995), whereas others, including most Snake River–type ignimbrites, are intensely and pervasively rheomorphic (e.g., Bad Step Tuff; Branney et al., 1992).

Welding and rheomorphism are favored by low-viscosity “soft” pyroclasts, resulting from some combination of (1) high emplacement temperature (above the glass transition; Freundt, 1998; Russell et al., 2003); (2) an eruption style that minimizes cooling during emplacement (e.g., high mass flux, low pyroclastic fountaining with minimal air ingestion) and rapid deposition (Ekren et al., 1984; Branney et al., 1992; Bachmann et al., 2000); (3) high residual contents of water or halogens in the pyroclasts at the time of deposition (Henry et al., 1988; Duffield and Dalrymple, 1990); (4) strongly peralkaline compositions (Mahood, 1984); and (5) topographic slopes. In some cases, onset of rheomorphism occurs rapidly during deposition (Branney and Kokelaar, 1992), but in others, it occurs predominantly as late-stage creep (as envisaged by Wolff and Wright, 1981). Rheomorphic folds are characteristically similar in style, curvilinear, have abundant oblique folds and sheath folds (Branney et al., 2004), and are associated with prolate, elongated fiamme and vesicles, local boudinage, and tension gashes in fiamme or matrix, thrusts, and local auto-brecciation of degassed upper surfaces (Schmincke and Swanson, 1967; Chapin and Lowell, 1979; Wolff and Wright, 1981; Branney et al., 1992; Summer and Branney, 2002; Pioli and Rossi, 2005). In extreme cases, pyroclast coalescence and intense rheomorphic transposition can obliterate original clast outlines, and result in the creation of massive or flow-banded welded lithofacies in which the pyroclastic origin is obscured: Such lithofacies are commonly termed “lava-like” (Branney and Kokelaar, 1992). There have been few structural studies of rheomorphism (e.g., Chapin and Lowell, 1979; Branney et al., 2004), and most have been of strongly peralkaline ignimbrites (e.g., Schmincke and Swanson, 1967; Wolff and Wright, 1981; Kobberger and Schmincke, 1999; Sumner and Branney, 2002).

This paper documents the architecture of lithologies, fabrics, and structures in a lava-like, rheomorphic ignimbrite, and uses them to reconstruct the history of emplacement and ductile deformation. We infer that pyroclast deposition, welding, and ductile flow were rapid and initially occurred within a thin, upward-migrating ductile shear zone; however, the deformation was progressive, and later gravitational spreading or downslope flow affected a much thicker portion of the deposit, ultimately becoming brittle locally and ceasing as the temperature dropped beneath the glass transition.

**Geological Setting**

Voluminous bimodal (basalt-rhyolite) volcanism associated with the Yellowstone hotspot has dominated the interior northwestern United States since the mid-Miocene (Pierce and Morgan, 1992), giving rise to the Yellowstone–Snake River Plain volcanic province. This developed contemporaneously with Basin and Range crustal extension, and rhyolitic volcanic rocks are variously cut by, and partly fill, half-graben basins along the margins of the Snake River Plain (e.g., Rodgers et al., 2002). Rhyolitic eruptive centers, now buried by basalt in the interior of the plain generally get younger toward the east, and the most voluminous volcanism was contemporaneous with periods of greatest regional extension (Bonnichsen et al., 2008). Rheomorphic ignimbrites with abundant lava-like facies are abundant within successions exposed around the central Snake River Plain and include Cougar Point Tuffs 1–15 (Bonnichsen and Citron, 1982) and ignimbrites of the Mount Bennett Hills (Honjo et al., 1992) and Cassia Mountains (McCurry et al., 1996; Wright et al., 2002; Fig. 1).

The Grey’s Landing ignimbrite is part of the ca. 10.6–8 Ma Miocene Rogerson Formation (Andrews and Branney, 2005) in the Rogerson graben, a north-south–trending half-graben on the south side of the central Snake River Plain (Fig. 1). The formation consists of eight extensive rhyolitic ignimbrites (total volume ≥40 km³) intercalated with paleosols, ash-fall layers, and volcaniclastic sediments emplaced during episodic subsidence of the graben (Andrews et al., 2008). Vent locations are not seen and may lie to the north between the inferred Bruneau-Jarbridge and Twin Falls eruptive centers (ca. 15–8 Ma; Fig. 1; Bonnichsen et al., 2008). The Rogerson Formation may, in part, correlate with contemporaneous and compositionally similar rheomorphic and predominantly lava-like ignimbrites in the adjacent Cassia...
Mountains (Tuff of McMullan Creek; Wright et al., 2002); both successions are of Snake River type (Branney et al., 2008).

**Grey’s Landing Member**

The ca. 8 Ma Grey’s Landing Member (Andrews et al., 2008) is a paleosol-bounded rhyolitic eruption unit of at least 13 km$^3$ volume. It is composed of an ~2-m-thick, parallel-beded ash-fall layer overlain by a welded ignimbrite (Figs. 2A and 2B) that varies in thickness from less than 2 m over palaeotopographic highs to 70 m thick in the deeper, western part of Rogerson graben and northward toward the Snake River Plain (Fig. 2). The ignimbrite has a basal vitrophyre, a thick central lithoidal (crystalline) zone, and an upper, locally pumiceous, vitrophyre, overlain by nonwelded ash (Fig. 2). It is the subject of this study because it is spectacularly rheomorphic and excellently exposed in a series of easily accessible, perpendicular canyons.

It is a compositionally zoned, metaluminous, high-silica rhyolite ($\text{SiO}_2 = 68$–75 wt%) with an A-type granitic composition and elevated FeO/($\text{FeO} + \text{MgO}$) ratio, Zr, and high field strength elements (GSA Data Repository Appendix DR1$^1$). It contains a sparse phenocryst assemblage of plagioclase-augite-pigeonite glomerocrysts, Fe-Ti oxides, and rare hypersthene. Quartz and sanidine are absent. Vertical zoning in the ignimbrite is marked by a changing distribution of pyroxene phenocrysts: augite-pigeonite-hypersthene at the base, and pigeonite only at the top. Two-pyroxene geothermometry obtains magmatic equilibrium temperatures of 925–975 °C at the base and 950–1025 °C at the top (Appendix DR1 [see footnote 1]; Andrews et al., 2008).

**Basal Parallel-Beded Ash**

Along the eastern margin of the Rogerson Graben (Fig. 3) a thin-beded to parallel-laminated rhyolitic ash (Fig. 4) overlies ripple-laminated volcaniclastic sands and a weak paleosol developed on the underlying ignimbrite (Backwaters Member; Andrews et al., 2008). The deposit is predominantly medium to coarse ash with cuspatve shards, and rare layers of small- to medium-sized (2–20 mm diameter) angular pumice clasts (Fig. 4). It is interpreted as an ash-fall deposit on the basis of its good

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$^1$GSA Data Repository item 2010185, Appendix DR1, whole-rock geochemistry table; Appendix DR2, glass transition temperature, is available at http://www.geosociety.org/pubs/ft2010.htm or by request to editing@geosociety.org.

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Figure 2. Graphic logs of typical sections through (A) thick and (B) thin Grey’s Landing ignimbrite, showing internal zones A–E (see text). Two structural domains are developed in thick sections: a lower, flat domain with subhorizontal planar fabrics and intrafolial isoclinal folds, and an upper, contorted zone, in which early isoclinal folds are refolded and the fabrics vary from horizontal to steep.
Figure 3 (continued on following page). Simplified geologic maps showing the distribution of the Grey’s Landing ignimbrite in the north part of Rogerson graben and the southern margin of the Snake River Plain. (A) Structural data from the lower, flat domain: stereonets present elongation lineations, fold axes, poles to S₀–1, and poles to fold axial planes. Inferred transport directions from kinematic indicators are given at discrete sampling sites. (B) Trends of elongation lineations and fold axes plotted against height within the lower, flat domain of the Grey’s Landing ignimbrite at individual locations.
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Figure 3 (continued). (C) Structural data from the upper, contorted domain. (D) Cross section (line Z-Z’ in C; note vertical exaggeration) showing thickening of the ignimbrite westward into the asymmetric Rogerson graben.
Contact-Thermal Effects

Where the Grey’s Landing ignimbrite is thickest, the substrate below shows remarkable contact-thermal effects (Andrews et al., 2008). The entire basal ash-fall layer is fused (sensu Christiansen and Lipman, 1966) and compacted into a 50–75-cm-thick, dark vitrophyre with parallel, black and dark-gray laminations of constant thickness (Fig. 5A), and a bedding-parallel eutaxitic fabric of strongly flattened, <0.5 mm ash shards. Megascopic fiamme are not seen, and there is no evidence of rheomorphism. Flattened grass-like leaves are preserved as soil casts against the glass at the basal contact (Fig. 5B), and the underlying <90-cm-thick paleosol has been baked to terracotta and exhibits columnar cooling joints spaced ~6 cm apart (Fig. 5C). Even beneath the baked paleosol, the upper ≤1.25 m of the underlying (normally nonwelded) Backwaters ignimbrite has been fused to black vitrophyre (Andrews et al., 2008, their fig. 8A); animal burrows preserved in this vitrophyre indicate that the fusion was imposed significantly later than its emplacement. The overlying Grey’s Landing ignimbrite was extremely hot when emplaced, probably only slightly below its magmatic temperature (≤1000 °C; Appendix DR1 [see footnote 1]; Andrews et al., 2008), and conduction of this heat into the substrate led to fusing.

RHEOMORPHIC IGNIMBRITE

The rhyolitic sheet above the basal fallout layer is predominantly flow banded and flow folded, and it has upper and lower vitrophyres and a lithoidal (crystallized) center. It is distinguished as an ignimbrite (similar to lava-like ignimbrites elsewhere; Bonnichsen and Citron, 1982; Ekren et al., 1984; Henry et al., 1988; Branney et al., 1992; Sumner and Branney, 2002) on the basis of a combination of features: (1) it lacks a basal autobreccia; (2) abundant vitric ash shards are preserved in the lower and upper vitrophyres (Fig. 6A); (3) it lacks the lobate morphology with steep terminations typical of long rhyolite lavas in this province (Bonnichsen and Kauffman, 1987); (4) where the sheet is thin, matrix-supported fiamme are locally present in the basal vitrophyre (Fig. 6B); (5) it is considerably thinner (≤70 m thick) than long rhyolite lavas in the Snake River Plain and elsewhere, and it has a lower aspect ratio; and (6) it is extensive (>400 km²) and thins to less than 2 m thick over topographic highs. The preserved shards are unequivocally pyroclastic bubble-wall or cuspate morphologies (Fig. 6A), grade seamlessly into flow-banding, and do not have the characteristics of sheared clasts in lava autobreccias, as described elsewhere (Pichler, 1981; Sparks et al., 1993; Manley, 1996). Pumice lapilli, fiamme, and lithic lapilli are scarce, typical of Snake River-type ignimbrites, many of which are dominated by lava-like facies (Bonnichsen and Citron, 1982; Branney et al., 2008). The absence of bedding and the simple welding profile, with a thick welded zone and basal and upper vitrophyres, suggest that the ignimbrite is a simple cooling unit (Smith, 1960). There is no evidence for deposition from more than one pyroclastic density current despite superb continuous exposure: however, the possibility that flow-unit boundaries have been obliterated everywhere by rheomorphism cannot be discounted.

We divide the ignimbrite into five distinct but intergradational zones, described next, distinguished on the basis of lithofacies (welding fabrics, crystallization, and structural features; Fig. 2). After briefly interpreting each zone, we describe the deformation and present an interpretation of the ignimbrite’s emplacement and rheomorphic evolution.

Zone A: Eutaxitic Basal Vitrophyre

Where the Grey’s Landing ignimbrite is thinner than 10 m (Fig. 2B), the lowermost few decimeters (≤1 m) constitute massive vitrophyric tuff with a weak eutaxitic fabric (Sₐ) defined by flattened ash shards (Fig. 6A). Matrix-supported, macroscopic fiamme (<10% modal abundance) are visible in the lower 50 cm at one location (Fig. 6B). Zone A is distinguished from the underlying fused ash-fall deposit by the absence of parallel stratification. The fabric at the base
of the ignimbrite is predominantly oblate (vertical compaction), but asymmetrical crenulation of shards and rare highly attenuated vesicles indicate a component of non coaxial rheomorphic shear, the intensity of which increases markedly with height, such that zone A grades upward into flow-banded, rheomorphic vitrophyre (zone B). Where the ignimbrite thickens, zone A also grades laterally into zone B.

Interpretation

Zone A is formed of recognizable vitric ash shards and rare pumice lapilli thought to represent the first-deposited material from a voluminous pyroclastic density current that covered at least 400 km². The eutaxitic fabric records welding: eutaxitic vitrophyres are a common chilled basal facies of rheomorphic ignimbrites worldwide (Streck and Grunder, 1995; Kobberger and Schmincke, 1999; Sumner and Branney, 2002). The stretched vesicles record limited rheomorphic flow at this level compared to higher levels. Therefore, zone A records rapid quenching (cooling through $T_g$ and increase in effective viscosity), possibly due to the deposition of lower-temperature material compared to that in zones B–D. Slightly cooler conditions may have yielded lower-temperature material compared to that in zone C, where the contact is sharp and defined by a surface of close-packed and intergrown spherulites.

Zone B: Flow-Banded Basal Vitrophyre

Zone B is dark gray and black, sparsely vesicular, and strongly rheomorphic, with near-continuous flow-banding ($S_{w1}$), rotated crystals (Fig. 6C), asymmetric microscopic folds, and stretched (prolate) vesicles (Fig. 6D). Where the ignimbrite exceeds 10 m thick, zone B reaches 2 m thick and zone A is absent (Fig. 2A); however, on topographic highs, where the ignimbrite is thinner, zone B overlies zone A such that the basal vitrophyre (zones A plus B) reaches ~3 m (Fig. 2B). In thin section, zone B is flow-laminated, but welded former cuspat e shards are preserved in some low-strain zones adjacent to rotated crystals (Fig. 6C; Andrews, 2006). Sparse spherulites ($\leq$ 3 cm diameter) and lithophysal cavities ($\leq$ 3 cm) overprint the flow banding. The vitrophyre passes up into zone C, where the contact is sharp and defined by a surface of close-packed and intergrown spherulites.

Interpretation

The near-continuous flow banding, vitroclastic fabrics preserved in strain shadows, rotated crystals, prolate vesicles, and gradations into zone A indicate that this is a rheomorphic tuff, similar to flow-banded facies of extremely high-grade ignimbrites elsewhere (e.g., Bar Step Tuff—Branney et al., 1992; Gomez Tuff—Henry et al., 1988; House Creek ignimbrite—Branney et al., 2004). The gradational contact suggests that zones A and B derive from the same current, and the contrast in welding fabric reflects differing thermal and deformation histories: residual temperatures in zone B must have been higher than in zone A, and rheomorphism transposed and obliterated a former eutaxitic fabric, which in zone A was preserved by rapid chilling.

Zone C: Lithoidal Lava-Like Rhyolite

The central zone (C) of the Grey’s Landing ignimbrite is flow-banded lithoidal lava-like rhyolite, up to 55 m thick. It is composed of granophyric intergrowths of quartz and alkali feldspar crystals ($0.3–1$ mm in size) with no devitrification features other than spherulitic upper and lower margins (Fig. 2A). Laterally continuous, millimeter- to decimeter-thick flow banding is ubiquitous and marked by variations in matrix color (Fig. 7A), crystallinity, and, to a lesser degree, vesicle abundance. It is locally picked out by subparallel discontinuous partings (or “platey joints”), spaced 5–25 mm apart, some coated with vapor-phase minerals. Abundant flow folds (e.g., Figs. 2 and 7A), rotated crystals, and lineated, highly attenuated vesicles (Fig. 7B) lie along the foliation. Fabrics in basal parts lie subparallel to those in the basal vitrophyre. Higher in zone C, the fabrics are locally deformed by, and also crosscut by, large ($\leq 15$ cm diameter) spheroidal vesicles (Fig. 7C). The fabric is cut by crude subvertical cooling joints that define polygonal columns, 2–4 m wide. Zone C is directly overlain by an upper vitrophyre (zone D; Fig. 8). It is absent where the ignimbrite is less than 8 m thick.

Interpretation

Zone C is the crystalline center of an intensely welded rheomorphic ignimbrite. Although it lacks vitroclastic textures, it grades downward, upward, and laterally (where the ignimbrite thins) into demonstrably pyroclastic vitrophyre with no intervening autobreccia. Given the intense folding, we infer that primary vitroclastic and welding textures were obliterated during pervasive rheomorphism and crystallization, as has been invoked for extremely high-grade tuffs elsewhere (e.g., Cougar Point Tuffs—Bonnichsen and Citron, 1982; Bar Step Tuff—Branney et al., 1992). The overall degree of crystallization and absence of remnant devitrification features suggest that the majority of zone C cooled through the solids without first intersecting the glass transition ($T_g$) and quench-
ing to form a glassy vitrophyre. \( T \) is estimated to vary from \(-525 \, ^\circ C \) (1 wt% H\(_2\)O and 1 wt% F\(_2\)O\(_3\)) to \(-725 \, ^\circ C \) (volatile-free, Appendix DR2 [see footnote 1]; Giordano et al., 2008). This is consistent with the other evidence for high emplacement and residual temperatures (e.g., the rheomorphism, lava-like lithofacies, the \( \leq 1000 \, ^\circ C \) magmatic temperature estimates, and the 3-m-thick zone of contact-fusing and baking of underlying units).

**Zone D: Upper Vitrophyre**

Zone D is a black flow-banded upper vitrophyre (Fig. 2) that reaches 5 m thick. It becomes increasingly vesicular upward (0%–70%), locally forming partly brecciated zones of microvesicular pumiceous rhyolite. It is pervasively rheomorphic, with rotated crystals, stretched vesicles, and millimeter- to meter-scale folds, and it locally shows autobrecciation (Figs. 8 and 9A), which varies from jigsaw-fit to more disturbed. Eutaxitic textures are locally visible in thin section, and there is some evidence of vapor-phase crystallization and oxidation. There are occasional bands of spherulites, \(< 5 \, \text{cm diameter} \) (Fig. 9B). The upper vitrophyre grades down into lithoidal zone C via a transitional zone of interbanded decimeter-thick lithoidal and vitrophyric layers (Figs. 8 and 9B), some of which host close-packed, partly merged, and sometimes lithophysal spherulites, \(< 5 \, \text{cm in diameter} \) (Fig. 2B). It is locally overlain by an orange nonwelded tuff (zone E; Fig. 8).

**Interpretation**

The vitroclastic textures transition into flow banding, the pervasive stretching lineation, and the gradational downward transition into zone C indicate that zone D represents the chilled upper part of the rheomorphic Grey’s Landing ignimbrite, where rapid chilling and degassing prohibited crystallization and pervasive devitrification.

**Zone E: Upper Nonwelded Ash**

A gray to orange-oxidized, nonwelded loose ash locally overlies the upper vitrophyre and is preserved predominantly in rheomorphic synformal clefs in the vitrophyre (e.g., Fig. 8). It is very well sorted and composed of macroscopic, coarse (0.5–2 mm) ash shards. It reaches 7 m thick, but it is considerably disturbed, so that its original thickness is unknown. Most is massive, but basal parts locally show faint stratification. At the contact with the upper vitrophyre, the basal 25 cm of ash are incipiently (sintered) to strongly welded (dense dark-red obsidian; Fig. 9C). These lower parts are typically brecciated into clasts as much as 25 cm in size, with a matrix of orange ash containing angular clasts of vitrophyre, mostly derived from zone D (Fig. 9D). Being largely nonlithified, zone E has widely been stripped by erosion, but remnants of orange ash are a ubiquitous component of regolith seen on the eroded upper surface of the ignimbrite. The top of the zone E is a bioturbated paleosol.

**Interpretation**

The orange ash forms part of the same cooling unit as zones A–D, i.e., it has fused locally at the contact with underlying hot upper vitrophyre. The unusually narrow and abrupt welding profile suggests a marked thermal discontinuity, with the orange ash being deposited soon after, but at a lower temperature (\(< T \) ) than, zone D. Other than this, there is no evidence for a significant hiatus here: the ignimbrite vitrophyre (zone D) was still hot and undergoing rheomorphic deformation when the orange ash was deposited, because late folds and related brittle deformation affect both the orange ash and zone D (Fig. 8), and because where the orange ash fills rheomorphic fissures in zone D, it was locally fused and thermally oxidized at the contact with hot ignimbrite.

The nonwelded orange ash probably exceeded 60 cm thickness prior to erosion. It may be an ash-fall deposit, or late nonwelded ignimbrite flow units deposited from smaller and cooler postclimactic density currents. There are no diagnostic sedimentary structures (partly due to disaggregation), and grain-size sorting is better (\( \sigma_v = 0.7 \) ) than is typical for ignimbrites (\( \sigma_v \) values of 2–5; Walker, 1971), even for Snake River-type ignimbrites, which are generally better sorted (\( \sigma_v = 0.8–2.5 \) ) than non–Snake River-type ignimbrites (Brannen et al., 2008, their fig. 4). Based on the good sorting, and general absence of crystals and welding, we tentatively interpret it as a disturbed ash-fall deposit. Given its position upon a large ignimbrite, it may be a coignimbrite ash, although it is thicker and its grain size is coarser (\( \phi = 3 \) ) than most “conventional” (non–Snake River type) coignimbrite ash-fall deposits (Sparks and Walker, 1977); however, this could reflect what was probably a very high thermal flux of the Snake River-type eruption.

**RHEOMORPHIC STRUCTURE**

The Grey’s Landing ignimbrite is pervasively rheomorphic, and it divides broadly into two structural domains, one overlying the other: a lower “flat domain” and a thicker, upper “contorted domain” (Fig. 2).

**Lower, Flat Domain**

The flat domain is \( \leq 23 \, \text{m thick} \) and constitutes the lowest third of the ignimbrite (zones A, B, and lower part of C). It is mostly an L > S tectonite with a well-developed subhorizontal planar foliation and a pervasive elongation lineation. The foliation \( (S_{0,1}) \) is a near-continuous flow laminating (banding) that represents a transposed and attenuated welding fabric \( (S_\beta) \). The elongation lineation \( (L_\beta) \) lies in the plane of the foliation and is defined predominantly by strongly prolate (cigar- to surfboard-shaped) vesicles with length to width ratios \( \leq 100:1 \) distributed along certain flow laminations (Fig. 7B).

Along the northern flank of the Browns Bench Massif and in the center and eastern margin of Rogerson graben (Fig. 3A), the lineation trends E-W and plunges \(-5^\circ \) eastward (Fig. 3), and kinematic indicators, such as rotated crystals (e.g.,
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Fig. 6C) and verging folds, indicate top-to-west rheomorphic transport. Along the southern margin of the ignimbrite, the lineations trend is NE-SW, with an ~5°NE plunge, and kinematic indicators show a top-to-southwest transport direction (Fig. 3A).

Throughout the flat domain, apart from in zone A, the fabric is folded into abundant intrafolial, recumbent isoclinal folds (F1) with oblique and curvilinear hinges (Figs. 10A and 10B). Many are sheath folds (Fig. 10C). They vary from microscopic to mesoscopic, with wavelengths less than 1 m. Individual folds do not deform layers ≥1 m thick. They are predominantly similar in style (class 2 of Ramsay, 1967), although locally disguised as chevron-style folds by the later development of subhorizontal curvy-planar platy jointing that is mimetic of the foliation but has propagated to crosscut the foliation at hinge zones (e.g., Fig. 10D). Rootless hinges are common (Fig. 10B), and multiple generations of F1 folds are refolded folds, with complex hinge zones (Fig. 10A). However, their orientations are generally consistent. The isoclinal, recumbent nature of the F1 folds is demonstrated by the characteristic subhorizontal (“flat”) attitude of the flow banding (stereonets in Fig. 3A) within the flat domain. The folds have no associated axial planar cleavage.

Figure 8. Sketch of the folded top of thick Grey’s Landing ignimbrite at Salmon Dam (inset as Fig. 3). Zone E (nonwelded orange ash, locally fused to vitrophyre) is preserved in synforms of F2 folds in zones C (crystalline) and D (vitrophyre, locally pumiceous and autobrecciated). Plant symbol depicts float.

Figure 9. Zones D and E of the Grey’s Landing ignimbrite. (A) Autobrecciated revesiculated vitrophyre (zone D) infilled by ash (zone E) at Salmon Dam (Fig. 8 inset). (B) Rheomorphic folds in zones C (lithoidal) and D (spherulitic vitrophyre) of the Grey’s Landing ignimbrite. (C) Schematic log showing relationships of upper vitrophyre and fused tuff. Clasts were formed by rheomorphic autobrecciation, and disruption of upper nonwelded ash. (D) Autobreccia clasts of upper vitrophyre (black) surrounded by fused orange ash (gray) (zone E) at Salmon Dam.
L1 lineations lie parallel to F1 sheath fold hinges (Fig. 3A), and the trends are consistent (typically E-W) at most locations. However, at some locations (e.g., Grey’s Landing and Monument Creek), both L1 and F1 azimuths change gradually and systematically with height through the ignimbrite (e.g., E-W to NE-SW and then back to E-W) through as much as 90° (see Figs. 3A and 11).

**Interpretation**

We infer that the foliation in the flat domain of the ignimbrite was formed in a flat-lying zone of intense noncoaxial ductile shear (Fig. 12). Because the welding fabric (referred to as S0) in the chilled base of zone A grades seamlessly into the more ubiquitous flow banding, we infer that the flow banding is the welding fabric, S0, transposed during D1 (we therefore refer to the flow banding as S0–1). This interpretation is consistent with interpretations of intensely sheared parts of rheomorphic ignimbrites elsewhere (Sumner and Branney, 2002; Branney et al., 2004). The ductile shear zone was subhorizontal, as indicated by the horizontal attitude of S0–1 flow banding and F1 fold axial planes and the consistently low plunge (≤5°) of the L1 stretching lineation (Fig. 3A). Its vertical thickness is constrained to have been less than 2 m, based on the maximum scale (wavelength ≤1 m) of F1 folds observed, the maximum thickness (≤1 m) of the folded layers, and conservatively taking into account likely subsequent flattening (welding compaction) due to loading. Two meters is much thinner than the (≤23 m thick) flat domain, and we infer that as the hot agglutinate aggraded by addition from above, the ductile shear zone migrated upward (Branney and Kokelaar, 1992), such that all the ignimbrite was sheared in a diachronous wave of noncoaxial deformation. In this way, the intensity of strain recorded at any particular height in the flat domain is a function of both the strain rate and the duration of the transient period in which that particular level resided within the shear zone. If the rheomorphic deformation occurred only after the ignimbrite had been deposited, one would expect to see individual folds affecting the entire thickness of the ignimbrite.

The gradual changes in azimuth of L1 lineations and F1 fold hinges (Fig. 3B) with height in the lower flat domain of the deposit are consistent with an upwards-migrating shear zone, and they probably record temporal changes in the transport direction of rheomorphic shear while the ignimbrite gradually aggraded. This phenomenon has been described in the Grey’s Landing ignimbrite and in rheomorphic ignimbrites elsewhere (Branney et al., 2004). Even in ignimbrites that are not rheomorphic, the

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**Figure 10.** F1 folds in zone C (lithoidal ignimbrite). (A) Intrafolial F1 folds and of flow banding, with isoclinal refolding. (B) Rootless intrafolial F1 fold hinge. (C) Eye structure of F1 sheath fold (elongation lineation, L1, not visible on photo, trends toward viewer). (D) F1 similar folding of S0–1 flow banding (form-line) mimicked by later platy-jointing giving appearance of chevron-style folding.
azimuth trends of grain fabrics can change with height (MacDonald and Palmer, 1990; Hughes and Druitt, 1998) and have been interpreted as indicating changes in the direction of granular shear during incremental sedimentation (Branney and Kokelaar, 1992; Hughes and Druitt, 1998). In rheomorphic ignimbrites, this effect is similar but enhanced by agglutination and ductile shear around the rising depositional interface (Fig. 12). In the Grey’s Landing ignimbrite, these vertical variations in the trend of L1 are substantial (≤70°) at some locations but weak to absent at others (Fig. 3B). This suggests that the temporal changes in transport direction were localized, and probably resulted from small topographic irregularities that developed and were modified (for example, by nonuniform deposition) at certain locations during the sustained passage of the density current.

The cigar-shaped L1 stretched vesicles indicate (monoclinic) uniaxial prolate strain in which the component of noncoaxial shear strain was significantly greater than that of coaxial shear strain, which was close to zero. However, surfboard-shaped L1 vesicles (e.g., Fig. 6D) are more abundant and indicate a weakly triclinic regime with a minor component of coaxial shear strain. Intuitively, one might expect more coaxial shear strain in the lower parts of the ignimbrite, in response to the greater load there. However, no simple relationship between vesicle shape and height within the deposit was observed, and at several locations, surfboard-shaped vesicles occur near the top of the ignimbrite. This is consistent with the vesicle shape reflecting the transient strain within a migrating shear zone.
prior to the assembly of the entire thickness of the ignimbrite. In addition, vesicles that grew in the welded tuff late would only record the later increments of deformation.

**Upper, Contorted Domain**

The upper, contorted domain constitutes the upper third to two-thirds of the ignimbrite, including the upper part of zone C and all of zone D, and it has disturbed zone E. It exhibits all the structural elements (S₀₋₁, L₁, and F₁) seen in the lower flat domain (i.e., planar foliation, elongation lineation, rotated crystals, stretched vesicles, intrafolial isoclinal oblique folds, and sheath folds), but here they are refolded into larger-scale (1 to >20 m) and, in some cases, more open and steep, curvilinear folds (F₂), such that at most locations, the flow banding (S₀₋₁) varies widely from 0° to 90° (Fig. 2A). The F₂ folds are polyphase, noncylindrical, and include sheath folds (Fig. 13A) that fold the S₀₋₁ fabric and refold F₁ folds (Fig. 13B). In contrast to F₁, they are not intrafolial. They are morphologically heterogeneous and vary between two end-member forms: (1) large recumbent sheath folds in zone C, and (2) large upright to overturned noncylindrical folds (affecting zones C–E), commonly in the form of elongate structural domes (culminations) and basins (saddles) (Figs. 8, 13C, and 14). The large recumbent F₂ sheath folds (Fig. 13A) in the interior of the ignimbrite sheet are abundant, with subhorizontal fold axial planes and generally E-W–trending fold hinges (Fig. 3C). They exhibit eye structures larger than 5 m across (Fig. 13A) and can be traced for more than 50 m in the plunge direction. Most trend parallel to S₀₋₁, F₁, and L₁ in the flat domain; in the Rogerson graben, this is parallel to the dip of the underlying paleoslope (Fig. 3C).

The upright to overturned, elongated F₂ domes and basins (Fig. 14) are open to isoclinal and have curvilinear fold axes (Fig. 3C) with highly variable plunges (0°–90°), even across distances less than 2 m. Their axes commonly trend subparallel to the trend of elongation lineations. They have composite hinge zones that refold smaller F₁ and early formed F₂ folds. At some locations, the flow banding in upper portions of the contorted domain is predominantly subhorizontal because it forms limbs of recumbent and doubly vergent, anvil-shaped isoclinal folds (Fig. 13D), or broad hinges of upright open folds (Fig. 13C). Wavelengths of F₂ range from less than 5 m to over 30 m, and amplitudes range from less than 5 m to 15 m. A weak, stretched vesicle lineation (L₂) is locally developed perpendicular to the hinges of some upright F₂ folds and oblique to the layering (Fig. 14). In some cases, this may be due to late-stage outer-arc extension across a competent fold hinge zone; in others, it reflects rotation of the flow banding (S₀₋₁) during D₂. Where L₁ is subparallel to L₂, the flow direction rotated porphyroclasts (dextral) eye structure (sheath fold) doubly vergent fold

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**Figure 12.** Deposition of hot pyroclasts, agglutination, and ductile shear within a rising subhorizontal shear zone at the base of a granular-fluid–based pyroclastic density current (adapted from Branney and Kokelaar, 1992; Pioli and Rossi, 2005). Hot, soft pyroclasts and rigid crystals are deposited from a high-concentration shearing dispersion, where they agglutinate and undergo ductile deformation in the shearing, uppermost parts of the hot deposit, producing flow banding (S₀₋₁), planar exsolution partings, recumbent F₁ folds and sheath folds, rotating the crystals, and elongating still-growing vesicles (L₁). With continued rapid deposition, the top of the agglutinate rises, and the shear zone migrates up with it. The later, larger-scale deformation (D₂) is not depicted here.
Emplacement and rheomorphic deformation of a large, lava-like rhyolitic ignimbrite: Grey’s Landing, southern Idaho

it is difficult to distinguish between these two extensional lineations.

Competency contrasts between welded and nonwelded ignimbrite (zones C and D vs. E) are manifested in cuspsate and lobe-style F2 folds; cuspsate synforms cored by nonwelded ignimbrite have angular closures, whereas lobate antiforms cored by welded ignimbrite have rounded, and often bulbous, closures (Fig. 8).

The boundary between the flat and contorted structural domains is a cryptic horizon within the ignimbrite marked by the first appearance of steep flow banding and fold axial planes (Fig. 2A). It is gradational, typically subhorizontal, 10–23 m above the base of ignimbrite, and can be traced laterally for many hundreds of meters. Despite good exposure, we have found no corresponding change in lithology, crystalinity, grain size, brecciation or brittle décollement surface at this horizon.

**Interpretation**

The presence throughout the contorted domain of flow banding (S0–1) and small intrafolial F1 isoclines closely similar to those in the flat domain suggests that the contorted domain initially underwent similar D1 deformation to the flat domain. We infer that the subhorizontal D1 shear zone that rose through the flat domain continued to rise through the central and upper parts of the hot agglutinate as the deposit continued to aggrade. This produced more flow banding, elongation lineations, and sheath folds (F1) almost to the top of the ignimbrite.

In the contorted domain, the F1 folds and S0–1 were then refolded by larger folds, which we term “F2” because they postdate F1. However, absolute timings are not constrained, and there was no cleavage development during D1, as would, for example, be used to distinguish between fold phases in tectonic deformation. Moreover, D1 was itself progressive and produced refolded folds. Fold size apparently increased with time, because as in addition to folding both F1 and S0–1, the large F2 folds refold smaller, earlier F1 folds. Since fold size is controlled by the thickness of the layer being folded (Ramsay, 1967), the folding also indicates that the thickness of the actively deforming layer within the ignimbrite increased with time.

After the D1 shear zone migrated to the top of zone D, rheomorphic shear continued (D2) and affected increasingly thick proportions of the deposit, ultimately with almost the entire 50 m thickness of the contorted domain being folded at the same time, as demonstrated by the largest F2 folds (Fig. 3C). The large F2 recumbent isoclines and sheath folds, together with the clear predominance of <10° plunges of elongation lineations, coupled with a paucity of L1 lineations steepened during D2, indicate that...
D₂ predominantly involved noncoaxial subhorizontal ductile shear with a similar, commonly westward transport direction to D₁. It is likely that the entire deformation was progressive: we found no evidence for a hiatus in deformation prior to D₂ remobilization, although this possibility cannot be excluded.

The more upright F₂ folds in uppermost parts of the contorted domain seem to have developed where subvertical compactional stress (loading) was minimal, close to the upper free surface of the deposit. The surface folds probably formed broadly at the same time as the F₂ sheath folds beneath, and with progressive deformation, fold axial trends tended to rotate and transpose into curvilinear festoon-type folds. Because of this tendency, transport direction is not readily inferred from the fold-axis orientations, which may lie at various angles to transport direction.

The D₂ pattern, in which steep surface folding toward the top of the contorted domain passes down to the flat domain, is similar to other types of nonparticulate gravity-driven flows, such as silicic lavas (e.g., Fink, 1980, 1983; Smith, 2002), soft-sediment slumps (Bradley and Hanson, 1998), and ice and salt glaciers (e.g., Hambrey and Lawson, 2000; Talbot and Aftabi, 2004). Typically, folds formed by near-coaxial shortening at the free upper surface are upright or inclined, and root downward into a region of noncoaxial shear strain (e.g., Merle, 1998). The trends of interior sheath folds and the shortening direction across upright folds before transposition are parallel to the (commonly downslope) transport direction. However, in the case of the Grey’s Landing ignimbrite, this deformation (D₂) postdated the earlier phase of ductile shear that produced the flow banding (S₀–1), L₁, and F₁ intrafolial isoclines throughout both domains. In the upper contorted domain, folding is likely to have reoriented L₁ lineations.

We did not identify a physical manifestation of the lower limit of D₂ deformation. Shortening evident from the large upright to overturned F₂ folds was probably accommodated by ductile shear in a (D₁) basal shear zone as in other types of nonparticulate gravity flow. As this would have resided within the lower flat domain, its limits cannot readily be distinguished amongst the already strongly attenuated and transposed (S₀–1) fabric with similar trend.

At most locations, the pyroclastic density current that deposited the Grey’s Landing ignimbrite flowed downslope and the D₁ and D₂ transport directions were similar. However, at one location (Monument Springs Creek), the paleoslope dipped gently northward, at a high angle to the initially west-flowing pyroclastic density current (Fig. 11), as recorded by

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Figure 14. Detailed map and cross section of a part of an elongate F₂ culmination at Cedar Creek Reservoir (Fig. 3B).
E-W–trending L₁ elongation lineation in the lower flat domain of the ignimbrite. Thin Grey’s Landing ignimbrite forms a perched ramp on this north-facing slope, and it thins and ultimately pinches out upslope to the south. L₁ progressively rotates with height in the ignimbrite, becoming more northward trending (Fig. 11). This suggests that the pyroclastic density current initially flowed westward along the south margin of the Snake River Plain, as recorded by the D₁ structures, fabrics, and kinematic indicators. The rheomorphic flow direction then changed with time, increasingly responding to the local substrate slope, first flowing to the northwest and finally to the north (see Branney et al., 2004). It seems that the thin perched agglutinated layer flowed downslope almost perpendicular to the original current direction, and is recorded by north-trending stretched vesicles and E-W–trending buckles in the top of the ignimbrite. The change in transport direction may be analogous to individual basin-margin turbidites that record two contrasting paleocurrent directions: sole structures recording the initial basin-axis–parallel rapid transport, and ripples analogous to individual basin-margin turbidites that record two contrasting paleocurrent directions: sole structures recording the initial basin-axis–parallel rapid transport, and ripples (Bouma C division) recording subsequent local, waning flow down the local basin-margin slope (Kneller et al., 1991). Alternatively, the northward, downslope flow may have occurred in the uppermost agglutinate only after the pyroclastic current had dissipated.

Cooling and Late Brittle Deformation

The locally abundant dilational fissures, prismatic jointing, and autobrecciation within the top of the upper vitrophyre indicate a change to brittle behavior during the latter stages of D₂, as the top surface chilled and degassed. The intense welding limited the permeability of the tuff to escaping volatiles, recorded by late growth of large vesicles (Fig. 7C) and lenses of coarse pumiceous vitrophyre. This would have contributed to marked competence heterogeneity between a relatively degassed crust and more ductile tuff beneath. Steep gradients in mechanical strength at the top of the ignimbrite also arose by fusing of the upper orange ash (Fig. 9C), which during late deformation acted as a relatively competent layer of brittle tuff enclosed by less competent layers (loose orangish ash above and hotter, more ductile ignimbrite below). Surfaces of hot ignimbrite exposed by dilatational fissuring developed prismatic jointed chills and sometimes bread-crusting (Figs. 15A and 15B), which probably contributed to autobrecciation.

The duration of rheomorphism is not known, but a simple one-dimensional conductive cooling model (Manley, 1992) allowing for conduction into and thermal metamorphism of the substrate suggests that the entire ignimbrite cooled below Tₐ (~525–750 °C; Appendix DR2 [see footnote 1]) in much less than 2 yr (Manley and Andrews, 2004; Andrews, 2006). Rheomorphism evidently ceased prior to quenching, devitrification, and crystalization, because the folds and fabrics do not record competence changes associated with spherulites (which overprint the flow banding; Fig. 9B) or crystalline (lithoidal) versus vitrophyric layers. It also ceased prior to cessation of volatile exsolution, as recorded by little-deformed subspherical vesicles, some of which crosscut and deform the flow banding and D₂ structures (e.g., Fig. 7C). It may be that the gravitational head required to drive spreading or downslope rheomorphic flow was gradually removed by rheomorphic flow until the ignimbrite sheet achieved gravitational equilibrium and became static. Platy joints exploited weak vesiculating layers in the flow banding (Pioli and Rossi, 2005), and their development probably accounts for the rather crude nature of vertical columnar cooling-jointing.

Summary of Emplacement History

The emplacement history of the Grey’s Landing ignimbrite is inferred to be as follows. (1) An explosive eruption of high-temperature, crystal-poor rhyolite initially generated an extensive umbrella cloud that gave rise to unsteady ash fallout, which blanketed the south side of the central Snake River Plain with parallel-stratified ash. (2) A change at the source (possibly to the west of Twin Falls; Fig. 1) to low pyroclastic fountaining (Branney and Kokelaar, 2002) generated a very hot, sustained pyroclastic density current that spread westward along the southern margin of the Snake River Plain and across the Rogerson graben. (3) The first deposited pyroclasts (zone A) were chilled by atmospheric cooling in the leading parts of the current (e.g., Branney and Kokelaar, 2002), and by contact with the cool ash substrate, but later pyroclasts were sufficiently hot and soft to agglutinate upon deposition (Branney and Kokelaar, 1992; Freundt, 1998) and undergo initial ductile flow during deposition (zones B–D).

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Folds affected the entire thickness of the contorted domain. Rheomorphic deformation seems to have been progressive: both phases involved refolding and attenuation, and show a westward transport direction. However, locally, on the south side of the plain, late rheomorphism of a perched feather-edge of the ignimbrite resulted from gravitationally driven flow to the north. The upper part of the ignimbrite chilled and degassed, locally frothed, and then rapidly became competent and brittle with the development of faults, dilational fissures, and autobreccias, locally enhanced by broad-crusting and prismatic jointing at exposed surfaces. Zone E disaggregated and oxidized while being transported and disturbed by rheomorphism in the hotter interior. (11) As it cooled, the interior of the ignimbrite began to crystallize, first with spherulitic devitrification near the margins of the two vitrophyses, and subsequently with extensive and thorough granophyric crystallization (zone C). Rheomorphism had ceased by the onset of quenching and devitrification, but minor degassing continued, as indicated by the presence of late subspherical vesicles. During cooling, late-stage platy joints propagated along the flow lamination, exploiting spaced, weak revesiculated laminations, but commonly crosscutting tighter fold hinges. Crude vertical columnar cooling joints developed. The upper nonwelded ash was widely stripped by erosion after the eruption, except where preserved in tight synformal clefts.

**DISCUSSION**

**Comparison with Other Snake River–Type Ignimbrites**

This study of the Grey’s Landing ignimbrite was undertaken to better understand rheomorphism in extremely high-grade tuffs, including those in Snake River–type volcanic successions. Most known Snake River–type ignimbrites (Bonnichsen and Citron, 1982; McCurry et al., 1996; Andrews et al., 2008) have many features in common with the Grey’s Landing ignimbrite, including (1) dense welding, with very little remnant porosity; (2) mostly sparse postwelding vesiculation, other than in uppermost zones; (3) curvilinear rheomorphic folds, including sheath folds, oblique folds, intrafolial isoclines, and refolded folds; (4) well-developed flow banding or lamination throughout; and (5) an intense elongation lineation including highly attenuated surfboard-shaped vesicles along spaced partings in the flow banding. As with the Grey’s Landing ignimbrite, most also have well-developed upper and lower vitrophyres, and zones of spherulites. Most overlie parallel-laminated ash-fall tuffs, which are partially fused to black vitrophyre, and they lack any basal autobreccia. Most have folded and locally autobrecciated upper surfaces with occasionally fused, nonwelded tuff preserved in the cores of synformal clefts. Chemically, Snake River–type ignimbrites are metaluminous high-silica rhyolites with sparse crystals and high magmatic temperatures. However, there is variation within Snake River type ignimbrites: for example, many are less intensely folded than the Grey’s Landing ignimbrite, and some are more pervasively eutaxitic and not rheomorphic (e.g., Rabbit Springs ignimbrite; Andrews et al., 2008).

Most Snake River–type ignimbrites that we have examined are predominantly flow banded throughout both flat and contorted zones, not just in a basal shear zone (e.g., Tuff of McMullan Creek—Wright et al., 2002; Castleford Crossing ignimbrite—Bonnichsen et al., 2008). This, together with the occurrence of small intrafolial isoclinal sheath sheaths throughout, suggests that much of the ignimbrite has undergone intense noncoaxial shear strain, and the formation of banding or lamination by the transposition and attenuation of welding fabrics in a subhorizontal ductile shear zone. In cases where the early folds are intrafolial, and where fabrics remain intense to higher levels in the ignimbrite, we would suggest that this shear zone rose through the deposit while the ignimbrite aggraded, as documented herein.

**Comparison with Peralkaline Rheomorphic Ignimbrites**

Most studied rheomorphic ignimbrites are strongly peralkaline (Mahood, 1984); these are not of Snake River type, and their rheomorphism is attributed to presence of abundant alkali-lowering particle viscosities by disrupting silicate polymerization. Despite differences in scale, appearance, and physical chemistry, most strongly peralkaline ignimbrites share several of the structural features of Snake River–type ignimbrites listed here, including a well-developed planar foliation; an elongation lineation; curvilinear, predominantly sheath folding; postwelding vesiculation; lower flat and upper more contorted domains; chilled vitrophyric zones; and local upper autobreccias. We therefore propose that their emplacement and rheomorphism are broadly similar, with the highest-grade examples having undergone early ductile shear during deposition and agglutination (Brannen and Kokeela, 1992; Freundt, 1998; Pioli and Rossi, 2005) followed by different amounts of post depositional spreading or downslope sliding of the hot deposit. However, there are significant differences in vesiculation, shear strain, and scale. (1) Snake River–type ignimbrites tend to be denser and less pervasively vesicular than most well-studied strongly peralkaline rheomorphic ignimbrites. Most vesiculation is confined to sheared partings along certain flow bands, a few isolated late vesicles, and restricted frothy zones in the uppermost part, whereas many strongly peralkaline ignimbrites, such as the pantelleritic Green Tuff of Pantelleria (Mahood and Hildreth, 1986) and comenditic to trachytic ignimbrites “D” and “TL” of Gran Canaria (Kobberger and Schmincke, 1999; Sumner and Brannen, 2002) are mostly highly porous and record sustained pervasive exsolution and revesiculation of fiamme and of welded matrix both during and after rheomorphism, recorded, respectively, by attenuated (prolate) and subspherical (or oblate) vesicles (Schmincke, 1974; Mahood, 1984). Although we do not know the species of gas that causes the late-stage vesiculation in either type of ignimbrite, their retention within the glass during eruption, transport, and emplacement may have helped maintain low viscosities. (2) Most studied ignimbrites of strongly peralkaline composition are predominantly eutaxitic, with large fiamme that record former hot blocks, lumps, and lapilli of pumice, scoria, or spatter, or they may record lenses of welded matrix that vesiculated. In contrast, most Snake River–type ignimbrites are predominantly flow banded and lava-like, with restricted eutaxitic zones. This partly reflects a relative paucity of large juvenile pumice or spatter in the Snake River–type density currents (Brannen et al., 2008), but it mainly records a higher shear strain, with higher intensities of rheomorphic attenuation and transposition. (3) Snake River ignimbrites tend to be larger in volume (tens to thousands of cubic kilometers DRE [dry rock equivalent]; Bonnichsen et al., 2008) than most well-studied strongly peralkaline ignimbrites (singe to tens of cubic kilometers DRE; Mahood, 1984), possibly reflecting the larger size of Snake River–type subvolcanic magma chambers. Given the wider extent of Snake River–type ignimbrites, we expect the eruptions were characterized by higher mass flux. There are, of course, exceptions to these generalizations in that strongly peralkaline ignimbrites exhibit a wide range in “grade,” from highly complex-folded ignimbrites with lava-like facies, such as ignimbrite “WTL” (Sumner and Brannen, 2002) on Gran Canaria, to less intensely rheomorphic examples where load-welding followed by post depositional creep may be more dominant. There are also rare examples of nonperalkaline rhyolitic ignimbrites that have pervasive fiamme, eutaxitic fabrics,
and syn- and postwelding vesiculation (e.g., Wall Mountain Tuff; Chapin and Lowell, 1979).

**Comparison with Silicic Lavas**

In some volcanic fields, it can be difficult to determine whether an extensive lava-like silicic sheet is a true lava or an unusually high-grade rheomorphic ignimbrite (e.g., Branney, 1986), particularly in cases where vitroclastic textures are not preserved. The difficulty arises because the physical characteristics of lavas and ignimbrites start to converge with increasing emplacement temperatures (Branney and Kokelaar, 1992); for example, coalescence of hot particles during deposition may in some cases obliterate original pyroclastic textures (Branney et al., 1992). For silicic bodies of unclear origin, structural and kinematic analysis has the potential to help distinguish between long lavas (e.g., Bonnichsen and Kauffman, 1987) and unusually high-grade ignimbrites. However, this distinction was not the aim of the present study, which was of an unequivocal rheomorphic ignimbrite, and to distinguish between lava and ignimbrite by structural and kinematic analysis would first require a comparable detailed study of an unequivocal long silicic lava.

In the Grey’s Landing ignimbrite, the D-deformation was inferred to occur in a subhorizontal ductile shear zone that migrated upward during deposition, ultimately to affect most of the thickness of the ignimbrite: there is no direct equivalent in silicic lava emplacement, although we cannot assume that all rheomorphic ignimbrites undergo this type of deformation. On the other hand, a silicic lava may inherit some flow bands, flow folds, and elongation lineations from shear adjacent to conduit margins (Gonnermann and Manga, 2005), whereas an ignimbrite cannot directly record conduit-margin shear, because the rock was fragmental on exiting the conduit, and the ductile structures do not predate assembly, at the site of deposition. Also, whereas lava travels most of the distance from source as a ductile to brittle flow, an ignimbrite flows in such a manner only for the last few meters, after initial transport to the site as a particulate density current. The ways in which such differences are recorded by structural, textural, and kinematic features would richly deserve further investigation. Until then, the best way to distinguish between long silicic lavas and predominantly lava-like rheomorphic ignimbrites remains the combination of features proposed by Bonnichsen and Kauffman (1987) and Henry and Wolff (1992); for example, silicic lavas generally have widespread, basal, clast-supported autobreccia, whereas rheomorphic ignimbrites lack extensive basal autobreccias (see Summer and Branney, 2002) and instead have sharp basal contacts between the substrate and coherent, chilled basal vitrophyre (or crystallized equivalent).

**Can Rheomorphic Structures Reveal Ignimbrite Sources?**

As with most ignimbrites in central Idaho, the source location of the Grey’s Landing ignimbrite is not known because the inferred calderas are buried beneath basalts lavas of the Snake River Plain, unequivocal proximal facies (e.g., proximal lithic breccias in ignimbrites associated with thick, coarse fall deposits) are not seen, and the regional stratigraphy is incompletely resolved (Branney et al., 2008). Identification of ignimbrite sources is problematic in many volcanic provinces, and it has traditionally relied on the identification of thick calderas fills and typical near-vent facies (e.g., Druitt, 1985), coupled with maps of ignimbrite paleocurrent indicators, such as imbricate fabrics in pumice or lithic lapilli (e.g., Hughes and Druitt, 1998) or studies of anisotropy of magnetic susceptibility (e.g., Le Pennec, 2000; Pioli et al., 2008) and fabric image analysis (Valentini et al., 2008). In rheomorphic ignimbrites, mapped distributions of elongation lineation trends have been used to infer transport direction (Elston and Smith, 1970; Schmincke and Swanson, 1967), including in the Snake River Plain (Bonnichsen and Citron, 1982; Bonnichsen et al., 1989; McCurry et al., 1996). In addition, trends of rheomorphic fold axes of oblique and sheath folds, and of prolate spherulites that grew in anisotropic rheomorphic tuff, can be used together with kinematic indicators to infer rheomorphic transport directions (Branney et al., 2004). These methods serve best to locate an erosion source where the substrate along which the pyroclastic density currents traveled sloped systematically away from source, such as on a broad shield. However, where the topography is more complex, local slopes do not everywhere dip away from source. Also, rheomorphic transport may not have been in the same direction as the overall transport direction of the density current. In this case, distinguishing between syndepositional and postdepositional components of rheomorphism may be crucial.

The transport direction inferred for the Grey’s Landing ignimbrite is westward (Fig. 3). No source is known to the east (Fig. 1), and we consider the possibility that the pyroclastic density current traveled southward from the inferred Twin Falls center in the central Snake River Plain (Fig. 1). In this case, the E-W-trending lineations in Rogerson graben (Fig. 3) would record westward rheomorphic transport as a consequence of the west-dipping graben floor. As these lineations are syndepositional (Lw), this interpretation would require partial decoupling of a southward-flowing particulate transport system from a depositional/agglutination system shearing westward down the paleoslope. However, this scenario makes it more difficult to account for the E-W-trending Lw outside the graben at the southern margin of the Snake River Plain (Monument Canyon; Figs. 3A and 11), where we have no independent reason to invoke a westerly paleoslope. We tentatively infer that the Grey’s Landing pyroclastic density current originated along the southern margin of the Twin Falls eruptive center (Fig. 1), but it ultimately spread westward across the southern Snake River Plain and Rogerson graben. We conclude that although lineation trends and kinematic indicators combine to give an accurate local transport direction (cf. McPhie et al., 2008), they do not always provide a straightforward indication of source location, particularly without independent constraints on the substrate paleotopography.

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