Contrasting Grenville-aged tectonic histories across the Llano Uplift, Texas: New evidence for deep-seated high-temperature deformation in the western uplift

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ABSTRACT

Models of doubly vergent orogens provide an excellent proxy for the Llano Uplift of central Texas, a Grenville-aged belt that consists of two portions with different structural styles, metamorphic grades, degrees of partial melting, and opposite directions of tectonic transport. Six phases of deformation at amphibolite-facies conditions are recorded in both portions of the uplift as a result of continent-continent and arc-continent collision. However, in the western uplift, field mapping and thin section analysis of microstructures and metamorphic mineral assemblages provide evidence for temperatures above the second sillimanite isograd and partial melting during both early and late deformation. These observations correlate well with numerical models for the Alps, which have identified a prowedge and retrowedge in the crust above a subducting slab. The western uplift is coincident with the retrowedge, located at greater depth in the orogenic pile, leading to greater temperatures and more melting as well as opposite vergence from the prowedge. A lack of discrete shear zones and opposite structural stacking and vergence in the western uplift, coupled with apparently greater temperatures and more widespread partial melting, suggest that the western uplift has a different tectonic history from the eastern uplift. Most notably, this study documents widespread and pervasive partial melting during uniformly distributed deformation, as well as abundant granitic intrusions during latest deformation in the western uplift. Analogue models readily accommodate these observations from both eastern and western parts of the uplift in the form of a bivergent orogenic wedge.

INTRODUCTION

Mesoproterozoic rocks of the Llano Uplift of central Texas form the core of a Grenville-aged orogenic belt along the southern margin of Laurentia (Mosher, 1993, 1998; Mosher et al., 2008) (Fig. 1). Long-lived orogenesis along this boundary culminated in continent-continent collision at 1150–1120 Ma (Mosher, 1998) with collision of an intervening arc in the eastern portion of the uplift (Roback, 1996). Previous work in the eastern portion of the uplift has documented northeastward tectonic transport within and between domain-bounding shear zones concurrent with collision. Such kinematic information, coupled with the timing of orogenesis, has placed important constraints on Rodinia supercontinent plate reconstructions and the tectonic evolution of Precambrian Texas (Mosher, 1998; Dalziel et al., 2000; Reese and Mosher, 2004; Mosher et al., 2004, 2008).

In this paper, we present new data from the Valley Spring domain of the western Llano Uplift, the structurally highest level exposed in the western uplift, that further constrain plate kinematics and timing of deformation. We document six phases of deformation synchronous with uppermost amphibolite-facies conditions. These results correlate with previous work done in the Packsaddle domain and Lost Creek Gneiss of the western uplift (Hunt, 2000). We propose that the differences between the eastern and western uplift in structural stacking, style, and orientation, metamorphic grade, and abundance of partial melting and fluids provide evidence for different tectonic histories for these two parts of the uplift, which are likely related to differing positions within a doubly vergent orogenic wedge.

GEOLOGICAL SETTING

The Llano Uplift, located in central Texas (Fig. 1), has undergone up to six phases of deformation with an associated uppermost amphibolite-facies dynamothermal metamorphism (Mosher, 1998; Reese and Mosher, 2004; Mosher et al., 2004; Hunt, 2000; this study). The medium-pressure metamorphism was preceded by a high-pressure eclogite-facies metamorphism and postdated by a low-pressure, partially static, metamorphism associated with intrusion of late syn- to posttectonic granites (Carlson et al., 2007, and references therein). Metamorphism in the eastern and western uplifts yields similar pressure-temperature (P-T) estimates for the dynamothermal phase (~700 °C, ~0.7 GPa) and for the late low-pressure phase (525–625 °C at 0.3 GPa) (Carlson et al., 2007). For the early high-P phase, conditions differ across the uplift: in the western uplift, temperatures were ~775 °C and pressures declined from ~2.4 GPa to ~1.6 GPa, whereas in the eastern uplift, temperatures were near 650 °C and pressures increased to ~1.4 GPa (Carlson et al., 2007).

The eastern portion of the uplift is composed of three lithotectonic domains: (1) The Coal Creek domain is interpreted to be an ensimatic arc with crystallization ages for plutonic rocks of 1326–1275 Ma (all ages are U-Pb zircon dates unless otherwise noted; Roback, 1996). (2) The Packsaddle domain is interpreted to be basal supracrustal rocks deposited adjacent to a continental-margin arc (Gar- rison, 1981; Mosher, 1998) with volcanic and intrusive protolith ages ranging from 1274 to 1238 Ma (Walker, 1992; Reese et al., 2000). (3) The Valley Spring domain is interpreted to represent a continental-margin arc with some terrigenous sedimentary rocks and plutonic rock ages ranging from 1288 to 1232 Ma; one rare basement rock yielded an age of 1366 ± 3 Ma (Walker, 1992; Reese et al., 2000). Shear zones separate the three lithotectonic domains; the
high-strain Sandy Creek shear zone (Roback, 1996; Mosher, 1998; Fig. 1) separates the Coal Creek and Packsaddle domains, and a lower-strain mylonite zone separates the Packsaddle and Valley Spring domains (Mosher, 1998; Reese and Mosher, 2004; Fig. 1). The structural stacking is associated with northeast-erly tectonic transport, with the Coal Creek domain thrust over the Packsaddle domain, which in turn is thrust over the Valley Spring domain. All domains are intruded by late syn- to posttectonic granites dated at 1119–1070 Ma (Walker, 1992; Reed, 1999).

Lithotectonic domains identified in the eastern uplift, with the exception of the Coal Creek domain, can be correlated across the uplift (Hunt, 2000). The Packsaddle and Valley Spring domains are both present, although they are separated by the Lost Creek Gneiss. Most of the Lost Creek Gneiss is a deformed porphyritic granitic pluton that intruded along the Packsaddle and Valley Spring domain boundary (Hunt, 2000) and is considered part of the Valley Spring domain. In the western uplift, however, the major lithotectonic domain boundaries dip southeast-ward, making the Valley Spring domain the structurally highest exposed level and the Packsaddle domain the structurally lowest.

STRUCTURAL ANALYSIS IN THE WESTERN LLANO UPLIFT

Five phases of deformation (D₁–D₅), each defined by ductile folds (F₁–F₅) and axial planar foliations (S₁–S₅), occur on a regional to micro-sopic scale and were accompanied by dyna-mothermal medium-pressure metamorphism at uppermost amphibolite-facies conditions. No earlier structures related to the high-pressure eclogite-facies metamorphism are observed in these rocks. A detailed geologic and structural map of the area is available in Levine (2005).

The first three phases of deformation, D₁–D₃, are characterized by structures and metamorphic textures that indicate temperatures at or above the solidus (Figs. 2 and 3). S₁–S₃ foliations are similar mineralogically and in style of deformation, but they are differentiated by superposed relationships, with each foliation generation clearly folded by the next and crosscut by later foliations in fold hinges. The later phases of

Figure 1. Geologic map of the Llano Uplift (LU) of central Texas (after Barnes, 1981). Inset: location map of Grenville-age exposures in the United States and Canada. Boxes represent divisions of the eastern and western portions of the uplift; divisions are located along major Paleozoic fault systems. Structural data are from master’s theses and Ph.D. dissertations conducted by Carter (1985), Hoh (2000), Hunt (2000), Levine (2005), Nelis (1984), Reese (1995), and Zumbro (1999). For detailed maps, at scales of 1:600 to 1:10,000, see cited theses.
Figure 2. Field photographs of associations between melts and deformation. (A) F2 fold with an axial planar S2 foliation, defined by ellipsoidal pods of quartz and feldspar with biotite in the center (leucosomes). The F2 fold is folding S1 and more continuous leucosomes parallel to S1. (B) S3 foliation (red line) crosscutting an earlier S2 foliation (yellow line). Both foliations are defined by ellipsoidal quartz/feldspar segregations (leucosomes) with biotite within the leucosomes. (C) F3 folding thin segregations of quartz and feldspar parallel to S2. Thicker quartz and feldspar segregations are subparallel to S2; note segregations branch off of wider segregations parallel to S2 (inset). (D) F3 fold cut by pegmatite roughly axial planar. Inset is close-up of a parasitic F3 fold of a well-developed S3 foliation defined by quartzo-feldspathic layers alternating with biotite-rich layers, with parallel pegmatitic layers. Note weaker S2 axial planar foliation. (E) F3 folds refolding S2 and an F2 fold (right); note F2 fold is folding an S1 foliation with parallel leucosomes. Pegmatite vein cuts both fold generations. (F) Close-up of F2 fold in E. S2 foliation and parallel quartz/feldspar-rich layers that appear to have been the result of partial melting are folded; an axial planar S2 that cuts S1 also contains thin segregations of quartz/feldspar. Quarter for scale in Figures 2C–2F.
Figure 3. Microstructures and field photographs that provide evidence for high-temperature deformation; conditions are above the second sillimanite isograd and coincident with partial melting. (A) An \( F_3 \) fold with axial planar \( S_3 \) (red dashed line) defined by aligned sillimanite. \( F_3 \) folds \( S_3 \) foliation defined by aligned sillimanite as well as quartz-feldspathic layers alternating with biotite-rich layers. (B) Close-up of garnets within elongate leucosome pods. Note aligned biotites and leucosomes are parallel to \( S_3 \) foliation. (C) Photomicrograph of \( S_3 \) foliation defined by quartz ribbons and sillimanite. Note presence of aligned sillimanite parallel to \( S_3 \) included in the quartz ribbons. In boxed area, an \( F_3 \) fold of sillimanite is parallel to \( S_2 \); \( S_3 \) is axial planar to this \( F_3 \) fold, as well as larger ones within the section (not shown). Qtz—quartz. (D) Photomicrograph of fine-grained muscovite replacing coarse sillimanite, parallel to \( S_1 \), and locally overgrown by coarse-grained muscovite. Sil—sillimanite, Musc—muscovite. (E) \( F_3 \) folds of earlier folds and foliations. Sillimanite parallel to \( S_1 \) is folded by \( F_2 \) folds with \( S_2 \) sillimanite growing axial planar. \( F_3 \) folds are outlined in yellow, and the axial planes are indicated by dashed lines. \( F_3 \) folds these \( F_2 \) folds and their associated axial planar foliation, \( S_2 \). Sillimanite is also growing axial planar to \( F_3 \) folds (red dashed line), indicating \( S_3 \) defined by sillimanite. (F) Photomicrograph of \( F_3 \) folds, which are folding an \( S_2 \) foliation defined by sillimanite, and which are overgrown by a larger muscovite (orange-yellow). The tight \( F_3 \) fold (right) has an \( S_3 \) foliation also defined by sillimanite axial planar to the \( F_3 \) fold.
in length that are aligned, forming a foliation. Although they do not necessarily have contact with their nearest neighbors, in places, ellipsoidal leucosomes are attached to each other by a string of quartz-plagioclase material, forming a more planar foliation. Rarely, multiple leucosomes are amalgamated to form an agglomerate of leucosomes (Figs. 2A and 2B). Elsewhere, leucosomes defined by continuous thin layers of quartzo-feldspathic material parallel S₁ (Figs. 2C, 2E, and 2F) and appear to branch off of wider leucosome zones that have undergone partial melting apparently in situ (Fig. 2C). These leucosomes axial planar to F₁ folds could be melt material that was injected during a later deformation or could have formed in situ.

F₂ folds are class 1C–class 3, generally tight folds that fold the well-defined S₁/S₂ composite foliation and F₁ and F₂ folds (Figs. 2D, 2E, and 3A). Most F₂ folds have a well-developed axial planar S₁ foliation that is less pronounced than S₂ and can be seen to cut across S₁ in the hinges of F₂ folds (Figs. 2D and 3A). A few outcrops show evidence of progressive folding during D₁ with two sets of folds, F₁ and F₂. The F₁ generation folds F₂ folds and S₁ foliation and are folded by F₃B folds. The F₃ generation folds F₂, and are folded by F₂ folds. F₃B are generally isoclinal to tight, whereas F₂ are tight to open. F₁ folds generally plunge to the east-northeast at a moderate to shallow angle but have been subsequently folded; S₁ strikes northwest and generally dips moderately to steeply northeast (Fig. 4B).

The S₁ foliation is most commonly defined by mica-rich layers of aligned biotite alternating with quartzo-feldspathic layers. In thin section, the S₁ foliation in granitic gneisses is defined by abundant to sparsely aligned biotites; in some thin sections, quartz, plagioclase, and K-feldspar grains are somewhat elongate and aligned. In pelitic samples, three discrete generations of sillimanite can be seen in thin section: S₁ sillimanite is folded by F₂, with S₂ sillimanite forming an associated axial planar foliation; both of these are folded by F₃, and are cut by the S₁ axial planar foliation, also defined by sillimanite (Fig. 3E). Elongate quartz ribbons that contain sillimanite aligned with S₂ are folded by F₃. Typically, the S₁, sillimanite, and F₂ folds of aligned S₁ sillimanite are overgrown by large muscovite grains (Fig. 3F).

The S₂ foliation is commonly defined by leucosomes that are axial planar to F₁ folds. These leucosome foliations are identical in appearance to the ones associated with S₁, but they clearly cut across S₁ (Fig. 2B), and they are much more commonly associated with S₁ and are better defined. Generally, these leucosomes are aligned ellipsoidal lenses of plagioclase and quartz surrounding small spherical patches of biotite, or in one location, garnet (Figs. 2B and 3B). In thin section, the quartz and plagioclase are much coarser grained within the leucosome than in the surrounding material.

Abundant F₃ folds have diffuse wide bands of quartzo-feldspathic material that appears to represent crystallized melt parallel to their axial planes. Locally, thin layers of crystallized melt parallel to S₁ appear to branch off of these diffuse bands that are axial planar to S₁ (Fig. 2C). Other F₃ folds of pegmatitic layers and leucosomes are cut by pegmatites roughly parallel to F₃ axial planes (Fig. 2D).

**Late Deformation**

Two later generations of class IC open folds, F₃ and F₄, fold all earlier structures and are seen on a regional to outcrop scale. F₃ and F₄ locally have associated axial planar foliations, S₃ and S₄, which are defined by aligned metamorphic minerals, but were definitively identified only in thin section. In thin section, large open folds, gently folding the D₁–D₃ structures, are common. F₃ and F₄ open folds biotite and muscovite that are part of the S₁ and S₂ foliations. Biotite has continuous undulatory extinction around the folds and in some folds is also kinked. F₃ fold axial traces trend north-easterly, and axes plunge east-northeast. F₄ fold axial traces trend north-westerly, and axes plunge southeast (Figs. 4C and 4D).

Throughout the field area, abundant granitic and pegmatitic veins and dikes cut across and are folded by various generations of folds (Fig. 2D), including F₃ and F₄. In addition, boudinage and local shear zones affect early structures and pegmatites; both of these structures are likely associated with D₅ extension. Larger granitic plutons, more widespread than the smaller granitic and pegmatitic veins, crosscut all structures and are undeformed. These post-Grenville granites have xenoliths containing the composite S₁–S₂ foliation.

**METAMORPHIC CONDITIONS**

Although evidence for earliest high-pressure metamorphism in the western uplift is found ~15 km along strike in the Mason County eclogites and elsewhere throughout the uplift, only evidence of the medium- and low-pressure metamorphism is observed in the study area. Using structural petrology, metamorphic reactions, and assemblages, the early deformation is constrained to at least uppermost amphibolite-facies conditions, and the later deformation is constrained to slightly lower temperatures that are compatible with amphibolite-facies conditions.
Figure 4. Equal-area, lower-hemisphere stereonets of fold axes and axial planar foliations and rose diagram of axial traces. Distribution of data is caused by multiple generations of subsequent folding or formation on folded surfaces. (A) Stereonet of poles to S₁ and S₂ foliations (filled circles) and F₂ fold axes (open triangles). (B) Stereonet of poles to S₃ foliation (filled circles) and F₃ fold axes (open triangles). (C) Stereonet of F₄ (open triangles) and F₅ (open squares) fold axes. (D) Rose diagram of F₄ (light gray) and F₅ (dark gray) axial traces. Outer circle is 12% of the data.
Recovery, Recrystallization, and Melt Textures

Recovery and recrystallization processes in quartz and feldspar differ slightly depending on the rock composition, as well as specific layer compositions. Highest-temperature conditions are constrained to syn-to immediately post-D₃, whereas during D₂, recrystallization processes were similar, but are not as diagnostic of the highest-temperature conditions.

Granitic Gneisses

Within the granitic gneisses, quartz has commonly undergone grain boundary migration, as evidenced by the irregular and blurry grain boundaries, as well as some subgrain rotation recrystallization. Some large quartz grains show chessboard extinction, which is indicative of the transition from low to high quartz, with temperatures of 700 °C at 0.5 GPa (Kruhl, 1996). Feldspars dominantly show evidence of subgrain rotation recrystallization, and also less ubiquitous grain boundary migration recrystallization is observed. Recovery features are rarer in feldspars than in quartz, but they include subgrains and discontinuous undulatory extinction.

In thin section, recrystallized quartz, plagioclase, and K-feldspar are elongate parallel to S₀, indicating that recrystallization occurred during the formation of the foliation. Similar mechanical behavior of quartz, plagioclase, and K-feldspar, plus both types of feldspars undergoing subgrain rotation recrystallization, indicates that metamorphism occurred at conditions of the upper amphibolite facies or higher (Passchier and Trouw, 1996). In thin section, the S₁ foliation is defined by aligned biotites and elongate quartz, plagioclase, and K-feldspar grains that have undergone recrystallization via grain boundary migration. Thus, recrystallization in quartz, plagioclase, and K-feldspar also can be constrained to have occurred syn-D₃.

Ellipsoidal leucosomes of quartz and plagioclase ± biotite or garnet are common in the granitic gneisses (Fig. 3B). Coarse grain size is characteristic of both quartz and plagioclase, and some plagioclase is elongate parallel to the elongate leucosomes, but generally appears undeformed. Some quartz has undergone dynamic grain boundary migration, as evidenced by amoeboid shapes; quartz typically has subgrains in addition to continuous undulatory extinction and some new grains. Very coarse quartz grains typically display chessboard extinction.

Pelitic Schists

In the more pelitic, sillimanite-rich, and muscovite-rich samples, quartz forms elongate ribbons that overgrow aligned sillimanite, indicating that these ribbons formed due to easy (high-temperature) grain boundary migration (Fig. 3C) that occurred under uppermost amphibolite- to granulite-facies conditions (Hirth and Tullis, 1992; Passchier and Trouw, 1996). Because these ribbons are folded by F₃, this grain boundary migration must have occurred synchronous with or just after the formation of S₃. Subsequently, quartz has undergone recovery and rotational recrystallization, as evidenced by subgrains and new grains, typically in the hinges of F₃ folds, suggesting somewhat lower temperature conditions during D₃.

Mineral Assemblages

Textures and minerals present within granitic gneisses, particularly those associated with leucosomes, indicate supersolidus temperatures. Reactions that took place within pelitic rocks provide evidence for uppermost amphibolite-facies conditions.

Sillimanite and K-Feldspar

Sillimanite and K-feldspar in equilibrium support metamorphism above the second sillimanite isograd. An S₁ foliation, defined by quartz ribbons bounded by sillimanite (Fig. 3C), is parallel to elongate K-feldspar grains. This parallelism provides evidence that both sillimanite and K-feldspar were stable during D₂. Depending on the mole fraction of water in the system, the minimum temperature, at pressures of 0.2–0.3 GPa, is 560–610 °C for the second sillimanite isograd (Fig. 5); with increasing pressure, up to 0.8 GPa, temperatures may be as high as 750 °C. Temperatures are in the same range as those documented by recrystallization mechanisms of quartz and feldspar and support metamorphism at uppermost amphibolite facies or higher.

Sillimanite and Muscovite

Reactions between sillimanite and muscovite provide evidence for fluids in the system and temperatures staying near the second sillimanite isograd.

Coarse sillimanite is aligned parallel to S₁ and is found as inclusions within elongate quartz grains defining S₁ (Fig. 3D). Fibrolitic sillimanite is aligned parallel to S₁, S₃, and S₄ and has been folded by F₂, F₃, and F₅ or F₇ folds (Figs. 3A, 3C, 3E, and 3F). Randomly oriented fibrolitic sillimanite, which is ragged in appearance, is observed in the center of coarse-grained muscovite and on the edges, continuing into the matrix. In a few places, the sillimanite that continues into the matrix is folded by F₂, F₅, or F₇ folds.

Pelitic rocks contain fine-grained muscovite, on average 100 µm in length (Fig. 3D), and coarse-grained muscovite, up to 1 cm in length (Fig. 3F). The fine-grained muscovite defines the limbs of isoclinal F₇ folds, is aligned parallel to S₁ in the hinges of F₇, is folded by F₅ folds, or locally appears randomly oriented. The muscovite parallel to S₁ is clearly folded by late folds, as evidenced by undulatory extinction across the hinges of F₅ or F₇ folds. Coarse-grained muscovite is not aligned with any visible foliations but commonly contains sillimanite that is folded by F₂ or F₅ folds and is aligned parallel.

Contrasting Grenville-aged tectonic histories

Figure 5. Pressure-temperature (P-T) diagram of the muscovite and quartz dehydration reaction (second sillimanite isograd) and the quartz and feldspar melting reactions (modified from Kerrick, 1972; Spear, 1993). These reactions help constrain the temperature and pressure conditions for early deformation at uppermost amphibolite-facies conditions. The filled circles represent samples with sillimanite and K-feldspar in equilibrium with melt and muscovite unstable. The filled squares represent samples with muscovite stable and where partial melting of plagioclase-quartz-rich samples has not occurred. The shading of the dot/square corresponds to the activity of H₂O, with the black dot/square representing an activity of 1 and darkness of gray decreases with decreasing activity of H₂O. An influx of fluids between D₂ and D₃ leads to muscovite overgrowth of sillimanite (conditions move from the circle to the square). When sillimanite again grows post-D₃, the reaction line is crossed in the opposite direction. Early deformation takes place on both sides of the second sillimanite isograd, at uppermost amphibolite conditions. For both the second sillimanite isograd and the granite melting reaction, the series of parallel lines with numbers (i.e., 0.5, 0.6, etc.) represent different activities of H₂O. Ms—muscovite; Qtz—quartz; As—aluminosilicate; Kfs—K-feldspar; Alb—albite; L—liquid.

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to $S_2$–$S_3$ (Fig. 3F). Within some coarse-grained muscovites, there is also fine-grained muscovite mimicking the shapes of $F_2$ and/or $F_3$ folds.

**Interpretations of Sillimanite and Muscovite Replacement**

Coarse sillimanite grew during $S_1$ and was probably isoclinally folded by $F_2$ folds and later replaced by fine-grained muscovite. This interpretation is favored because coarse sillimanite is only seen parallel to $S_1$, and fine-grained muscovite is not observed folded around the hinges of $F_2$ folds, even though it is found parallel to both limbs and axial planar. Some fine-grained muscovite, defining both $F_2$ folds and $S_1$, is recrystallized and kinked in the hinges of $F_3$ folds, whereas others have no extinction change around the hinge of $F_3$ folds. This suggests that the fine-grained muscovite replaced coarse sillimanite between $D_2$ and $D_3$ and post-$D_3$.

Fibrolitic sillimanite has grown parallel to $S_2$, $S_3$, and $S_4$ foliations and was folded by $F_2$, $F_3$, and $F_4$ or $F_5$ folds (Fig. 3E). Therefore, it must have been stable during $D_1$, $D_2$, and $D_3$. Such folds are abundant in all thin sections of pelitic rocks, and a series of folds of a specific generation can be traced across thin sections. Locally, coarse-grained muscovite has overgrown both $F_2$ and $F_3$ folds; each generation of folds has limbs and axial planar foliations defined by sillimanite (Fig. 3F). It is clear in these cases where coarse-grained muscovite contains folded sillimanite that the coarse-grained muscovite has overgrown sillimanite, and not vice versa, because there is no kinking or extinction change within the coarse-grained muscovite. Here, the coarse-grained muscovite could have overgrown fibrolitic sillimanite between $D_2$ and $D_3$ and post-$D_3$, or just post-$D_3$. Fine-grained muscovite also replaces the fibrolitic sillimanite, as the fine-grained muscovite is seen parallel to $S_2$ and within coarse-grained muscovite, mimicking the shape of both $F_2$ and $F_3$ folds.

The opposite reaction, muscovite going to sillimanite, also occurs in these rocks, as evidenced by the coarse-grained muscovite with randomly oriented, generally ragged-looking, fibrolitic sillimanite on the edges of the mineral. The timing of coarse-grained muscovite growth and breakaway to fibrolitic sillimanite occurred post-$D_3$, as shown by the overgrowth of $F_3$ folds and additionally may have occurred sometime post-$D_3$ but no later than syn-$D_3$.

In summary, sillimanite is seen parallel to $S_2$, $S_3$, and $S_4$, and has been replaced by muscovite sometime between $D_2$ and $D_3$ and post-$D_3$. Sillimanite replaces muscovite post-$D_3$, and possibly between $D_2$ and $D_3$. The second sillimanite isograd must be crossed for sillimanite to be replaced by muscovite; this can occur due to an influx of aqueous fluids, a decrease in temperature, or an increase in pressure. The replacement of sillimanite between $D_2$ and $D_3$ was likely due to an influx of fluids, perhaps originating from nearby pegmatitic intrusions, whereas the post-$D_3$ replacement was either due to an influx of fluids or a decrease in temperature. This reaction occurs at temperatures ranging from 560 °C to 750 °C, depending on pressure and activity of $H_2O$, typical of uppermost amphibolite-facies metamorphism (Fig. 5).

**Leucosomes and Partial Melting**

Three types of leucosomes found in the field area provide evidence for partial melting during early ($D_1$–$D_3$) and latest ($D_6$?) deformation. Each type of leucosome is associated with $S_2$–$S_3$ foliations: ellipsoidal pods of quartz and plagioclase ± biotite and/or garnet (Figs. 2A, 2B, and 3B); more continuous, thicker layers of felsic material (Figs. 2A and 2E); and long, thin, continuous veinlets of quartz and feldspar (Fig. 2C). All three types of leucosomes provide evidence, based on mineral segregation and texture, for partial melting during deformation.

Additionally, shear zones, containing leucosomes, pegmatitic or granitic material, or biotite-rich material (melanosomes), are found at a high angle to the foliation. These zones cut across all generations of structures and do not appear to parallel any foliations. Also, pegmatitic and granitic dikes and veins are found cutting across preexisting structures and are commonly deformed as well (Figs. 2D and 2E).

Within the Lost Creek Gneiss, leucosomes also are present within shear zones likely associated with $D_2$. The quartz in these leucosomes is coarse and contains chessboard extinction similar to that in the surrounding country rock; feldspars are finer-grained than quartz and rarely show subgrains. Commonly, pseudomorphs of melt are found along grain boundaries of unlike phases, or along subgrain boundaries in quartz; also, these grain boundaries are cuspat and serrate. These textures are indicative of the former presence of melt (Sawyer, 1999; Holness and Sawyer, 2008). The leucosome material found within the shear zones is texturally similar to that in the surrounding country rock, suggesting synchronous partial melting within the shear zones and in the surrounding rock. Within the Valley Spring Gneiss, the same types of textures are seen, likely indicating that late partial melting was widespread.

**Partial Melting Interpretations**

Partial melting occurred during $D_1$–$D_3$ in situ or locally and has led to formation of various types of leucosomes, which provide evidence for metamorphism at uppermost amphibolite-facies conditions. The lack of leucosomes or any fabrics consistent with migmatization associated with $D_7$–$D_8$ suggests that temperatures were lower during this later deformation.

The leucosomes parallel to $S_3$–$S_4$ represent partial melting that occurred in situ or was injected parallel to these planes of weakness. When temperatures reach the second sillimanite isograd (Fig. 5), water is released by the breakdown of muscovite: Musc + Qtz = Sil + Kfs + $H_2O$, triggering melting of quartz and plagioclase via the reaction: Alb + Qtz + $H_2O$ = L, along $S_3$–$S_4$, foliations and producing leucosomes. Figure 5 shows these two reactions together, with various activities of $H_2O$. The intersection point gives the temperature at which partial melting occurs in these rocks; if $H_2O$ is exchanged between gneissic and pelitic rocks, a minimum temperature and pressure of 675 °C at 0.5 GPa, with $H_2O$ activities of 1, will be reached. With lower activities of $H_2O$, partial melting will occur at higher temperatures.

Ellipsoidal leucosomes of quartz and plagioclase ± garnet ± biotite (Figs. 2A, 2B, and 3B) most likely represent nucleation points for partial melting; the garnets in the center are products of this melting reaction. The coarse-grained quartz with chessboard extinction found within $S_3$ leucosomes provides evidence for syn- to post-$D_3$ deformation at temperatures of uppermost amphibolite facies or higher. Plagioclase grains are finer-grained than quartz and do not show evidence of deformation. Better-developed segregations of leucosome and melanosome occur where there are discrete layers of felsic material alternating with typical biotite-rich melanosomes. These two types of leucosomes are foliation-parallel because melting occurs in situ. The breakdown of muscovite and quartz liberates a small amount of $H_2O$, which induces melting of quartz and plagioclase (or K-feldspar, if present), which segregates parallel to the fabric because of the differential stress. Continuous veinlets of leucosome material (Figs. 2C, 2E, and 2F) may have formed the same way, but more likely formed locally and were transported along a preexisting foliation. If transported, it happened early in the deformation history, because these veinlets have been folded by $F_2$ and $F_3$ (Figs. 2E and 2F).

The late shear zones, containing either leucosomes, or pegmatitic or granitic material, or biotite-rich material (melanosome), at a high angle to the foliation, indicate that melt conditions were prevalent late in the deformational history and may be correlative with a $D_6$ extensional phase of deformation identified by Hunt (2000). Shear zones observed in the present field area and within the Lost Creek Gneiss increase in abundance near granitic plutons and are likely...
synchronous with granitic or pegmatitic intrusions, supporting a link to syn- to posttectonic plutonism that has been documented throughout the uplift (Reed, 1999; Reed and Rougvie, 2002). The similarity in textures within the shear zones and in the surrounding country rock indicates that in granitic gneisses of both Lost Creek and Valley Spring Gneiss, partial melting was widespread, synchronous with \( D_6 \) extension and granitic magmatism.

**Retrograde Assemblages**

Little evidence for retrograde overprinting of the high-temperature assemblages is observed, but a few observations confirm retrogression during late-stage deformation. Biotite in both granitic gneisses and pelitic schists has frequently retrograded to chlorite, likely 

\( \text{H}_2\text{O} \) in the system caused a series of fluid inclusions responsible for the change from sillimanite to muscovite and back again. The repeated crossing of the second sillimanite isograd gives a pressure range of 0.6–0.8 GPa and a temperature range of 700–750 °C. Perhaps \( D_3–D_4 \) occurred over a relatively short time span as the style of deformation is similar, with dominantly isoclinal folds and well-defined foliations.

Open, late-generation folds (\( F_4 \) and \( F_5 \)) refold all previous structures on both outcrop and map scale. The late folds locally have an associated weakly defined axial planar foliation. \( F_4 \) folds have northeast- to east-striking axial planes with northeast-plunging fold axes, and \( F_5 \) folds have northwest-striking axial planes with southeast-plunging fold axes. These fold generations have biotite aligned parallel to the axial planes, instead of sillimanite, leucosomes or elongate quartz, plagioclase, and K-feldspar, and thus represent lower-grade metamorphism than the earlier deformation. The growth of biotite axial planar to \( F_2 \) and \( F_4 \) folds, as well as the presence of sillimanite that has been folded by these late-stage folds, provides evidence for lower temperatures, but is still consistent with amphibolite-facies conditions.

The change in conditions that occurred between \( D_2 \) and \( D_3 \) may have been associated with exhumation, bringing the rocks up to shallower levels. The stability of sillimanite but lack of new growth during \( D_3 \) and \( D_4 \) indicate a mid-amphibolite-facies pressure range of 0.4–0.6 GPa, suggesting that the rocks had been exhumed to depths ranging from 13 to 18 km. Graines and pegmatites, some of which may be associated with \( D_3 \) are both syn- and posttectonic, as evidenced by the presence or absence of the \( S_1/S_3 \) foliation within them, folding, and boudinage. Late melt textures support fairly widespread partial melting of rocks with a granitic composition. \( D_6 \) structures, including small-scale shear zones, are likely related to continued exhumation and associated extension. Lastly, the presence of widespread syn- toposttectonic granitic plutons, many containing xenoliths with early \( D_3–D_4 \) structures, indicates extensive late melting.

**DISCUSSION**

New data from the western Llano Uplift provide evidence for a different tectonic history than in the eastern portions of the uplift. The western uplift has opposite structural stacking, a lack of discrete shear zones bounding lithotectonic domains, higher-grade assemblages, and widespread partial melting, as well as perhaps more notably the absence of the Coal Creek domain, an island arc.

In the present field area and that of Hunt (2000), the structural orientation is characterized by dominantly northwest-striking foliations that dip to the northeast and east; these directions of dip are opposite from those in the eastern uplift, where foliations dominantly strike northwest but dip to the southwest (Fig. 1). Although the structures in the eastern uplift have a similar style, fold axes also plunge more easterly in the western uplift. In the eastern uplift, the boundaries between lithotectonic domains are defined by shear zones, and the intensity of deformation increases near the domain boundaries; neither of these features are seen in the western uplift (Mosher and Hunt, 2002; this study). Also, the tectonic transport throughout the eastern uplift is to the northeast (Reese and Mosher, 2004), whereas in the western uplift, no shear zones exist to indicate tectonic transport, and evidence of vergence is generally lacking. The structural stacking in the western uplift is the opposite of that documented in the eastern uplift, and the pervasive \( S_1/S_3 \) foliation dips in opposite directions in the west and east, suggesting a different direction of transport and an opposite vergence.

Another difference between the eastern and western portions of the uplift, which is not as obvious, is the different positions within the orogenic pile. Within the eastern uplift, the metamorphic grade is postulated to increase toward the northeast (Hoh, 2000; Mosher et al., 2004), in part because of the presence of partial melting solely in the northeastern portion of the eastern uplift. Partial melting is more widespread in the western uplift, with ellipsoidal leucosomes of quartz and plagioclase ± biotite and garnet found in the structurally highest unit, the Valley Spring Gneiss, as well as in the more deeply buried Lost Creek Gneiss and Puck saddle Schist (Hunt, 2000). The Valley Spring Gneiss contains other leucosome types and quartz-feldspathic veinlets, which parallel \( S_1–S_3 \), and are folded
by \( F_2 \) and \( F_3 \), providing additional evidence for partial melting throughout early deformation. Also, the presence of sillimanite parallel to \( S_1-S_3 \), folded by \( F_2-F_4 \), supports temperatures above the second sillimanite isograd. Notably, the conditions in the structurally highest unit in the west are the same as those documented in the structurally lowest unit in the eastern uplift. This suggests the western uplift was buried at greater depth than the eastern uplift during earliest deformation.

These differences between the eastern and western Llano Uplift have generally been attributed to the presence of the Coal Creek domain island arc in the eastern uplift and lack in the western uplift (Mosher, 1998; Mosher and Hunt, 2002). The collision of this arc could have controlled the direction of tectonic transport and orientation of structures within the zone juxtaposing this domain and within the underlying rocks (Mosher and Hunt, 2002).

More recently, Mosher et al. (2008) proposed that the differences in the deformation in the two portions of the uplift are related to collision and associated retrotransport and uplift, similar to models for the Alpine collisional orogeny (Beaumont et al., 1996; Pfiffner et al., 2000). In this model, as subduction commences, shear zones develop above the subducting plate, consistent with the direction of tectonic transport expected for a south-dipping subduction zone. As subduction continues, the crust becomes more buoyant, initiating uplift and retrotransport with an opposite direction of tectonic transport; subsequent break-off of the slab further enhances uplift and retrotransport as well as resulting in the intrusion of plutons (Fig. 6A).

A comparison of the Pfiffner et al. (2000) model in more detail to new observations from the western Llano Uplift indicates that the model is a good fit for the Llano Uplift. Pfiffner et al. (2000) made a series of two-dimensional finite-element models that may approximate the Swiss Alps at the time of collision. Their basic model assumes a strong crust, a weak suture, and moderate erosion and leads to development of a crustal-scale back fold in the retrowedge and, after 300 km of convergence, several zones of distributed deformation in the procrust (Fig. 6B). Subsequent models change erosion rates, crustal strength in both the pro- and retrowedge, as well as slab break-off. The models that have pockets of weak crust localize and intensify deformation (Fig. 6C). Although Pfiffner et al. (2000) did not have a model with weak crust at depth in the retrowedge, such a model would be an excellent proxy for the Llano Uplift. The western uplift has a much greater degree of partial melting than the east, and partial melting...
is found throughout all three units in the west. The position of the western uplift within the retrowedge, which is deeper in the orogenic pile (Fig. 6A), provides a good explanation for similar style of structures across the uplift, but higher-grade mineral assemblages and partial melt fabrics in the west.

The presence of shear zones in the eastern uplift, separating lithotectonic domains, is consistent with the porosity of the subduction zone. Modeling of doubly vergent orogenic wedges has consistently shown that, in the proveudge, discrete shear zones are likely to form, as opposed to one zone of concentrated deformation, typical of the retrowedge (Willett et al., 1993; Pfiffner et al., 2000; Persson and Sokoutis, 2002). This model is consistent with features seen in the western uplift; throughout all three lithotectonic units, intensity of deformation appears to be uniform. The fabrics associated with D1–D3 are well developed and penetrative, regardless of rock type, suggesting uniform concentration of deformation throughout the west. Pfiffner et al. (2000) suggested that preexisting zones of weakness may influence the formation of these discrete shear zones in the proveudge. In the Llano region, it is possible that the island-arc collision, just prior to continent-continent collision, could have caused these preexisting zones of weakness, which later became domain-bounding shear zones. In either case, shear zones are not likely to develop in the retrowedge, and they are not seen in the western uplift. Syn- to post tectonic intrusions are more widespread and abundant, forming irregular bodies, as well as defined plutons, in the western uplift than in the east. This difference may point to late channelization of melts in the retrowedge, which has been called upon for the Alps (Pfiffner et al., 2000), or it may be more indication that the west was still at greater depth than the east, even during latest deformation. Both ideas are consistent with a rheologically weaker retrowedge than proveudge.

The Mosher et al. (2008) model explains most of the discrepancies between the two portions of the uplift. The opposite orientations of tectonic transport and structural stacking can be explained by the disparate locations of the eastern and western uplift within the orogenic wedge. As shown in Figure 6A, the dip of structures in the proveudge have an opposite orientation to those in the retrowedge, so the model correlates well with structural orientation and direction of tectonic transport. The more uniform intensity of deformation across the western uplift is consistent with its position in the retrowedge, at greater depth than most of the orogenic wedge, and with high degrees of partial melting during early deformation.

Undeformed, high-pressure eclogites found in the western uplift give pressures of up to 2.4 GPa and temperatures of 775 °C, whereas their counterparts in the eastern uplift record conditions at a maximum of 1.6 GPa and 650 °C, indicating predeformational conditions that also place the western uplift at greater depths within the orogenic belt. Absence of deformation within these eclogites indicates that they acted as coherent blocks while they were exhumed from depths of up to 70 km (Mosher et al., 2008). The greater depth of eclogites in the western uplift is consistent with the tectonic model and provides more evidence for all of the western uplift, not just this study area, being at greater depth than the eastern portion of the uplift.

CONCLUSIONS

The western Llano Uplift has undergone a different tectonic history than the eastern uplift, as evidenced by a lack of discrete shear zones, a different direction of tectonic transport, and a greater degree of partial melting distributed throughout all three lithotectonic units. These differences can be explained by tectonic models of bivergent orogens, where the western Llano Uplift is consistent with a location in the retrowedge of the model. This model explains the opposite directions of tectonic transport in the eastern and western parts of the uplift, as well as the uniform intensity of deformation in the west and the high degree of partial melting throughout D1–D3.

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