Perspectives on the architecture of continental crust from integrated field studies of exposed isobaric sections

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ABSTRACT

Depth-dependent variations in the structure and composition of continental crust can be studied via integrated investigations of isobaric terranes. In this contribution, we summarize three isobaric terranes in Archean to Proterozoic crust. In western Canada, 35–45-km-deep lower crust is exposed over an area of more than 20,000 km². The Upper Granite Gorge of Grand Canyon, Arizona, provides a transect of 20–25-km-deep middle crust. The Proterozoic basement of central Arizona represents an isobaric exposure of 10–15 km-deep middle crust. Isobaric terranes yield a conceptual image of continental crust that can be compared to seismic images, xenolith data, and drill core data to clarify rheology, coupling/decoupling of crustal levels, and the interplay between deformation, metamorphism, and plutonism. General observations include: (1) The crust is heterogeneous at all levels and cannot be accurately modeled as a simple progression from quartz-rich to feldspar-rich lithologies or from felsic to mafic bulk compositions; (2) The crust is segmented into foliation domains that alternate between steeply dipping and shallowly dipping; (3) Magmatism is expressed differently at different depths due to different background temperatures and a general upward distillation from mafic to felsic composition, and may be the most important control on crustal architecture and rheology. The strength of continental crust (and its potential for low-viscosity flow) is not simply a function of temperature, depth, and compositional layering, but is controlled by the size and distribution of rheological domains. The rheological character of a particular layer can vary in space and time at any crustal level.

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INTRODUCTION

Knowledge of the deep structure of continental crust relies on geophysical data combined with direct observations from xenoliths and crystalline rocks that have been exhumed from deep levels. Considerable attention has been paid to exposed crustal cross sections — tilted slices of crust exposed at the surface (e.g., Salisbury and Fountain, 1990; Percival et al., 1992). These can provide cross-sectional views of depth-dependent variations in composition as well as processes. Examples such as the Ivrea-Verbano zone (Burke and Fountain, 1990; Rutter et al., 1999), Fiordland (Klepeis et al., 2004), and the Kapuskasing uplift (Percival and Card, 1983; Percival and West, 1994) have provided information on the character and degree of coupling or decoupling across crustal boundaries, including the Moho. Although important insights have arisen from these sections, by their nature, they provide less information about lateral variation in physical properties and/or rheological behavior.

In contrast to crustal cross-sections, isobaric terranes are here defined as regions that were laterally contiguous at a specific depth in the crust during part of their geologic history, commonly including the time of peak metamorphism. The degree to which a region can be defined as an isobaric terrane depends on the spatial and temporal scales over which pressures were nearly constant and on the questions being asked. Isobaric terranes provide rich field laboratories for evaluating the spatial heterogeneity of material and processes at particular crustal levels. For example, these terranes provide the most directly observable data regarding the degree of strain partitioning (e.g., Butler et al., 2002), bulk rheology (e.g., Handy and Zingg, 1991; Metzger, 1992; Rutter and Brodie, 1992; Klepeis et al., 2003, 2004), and nature of crustal flow (e.g., Williams and Jiang, 2005). Understanding such processes through map-view geologic images of particular crustal levels provides the necessary regional context to accurately interpret seismic and xenolith data.

In this contribution, we present three examples of regions we define as isobaric terranes, each representing a different crustal level; one from the western Canadian Shield and two from Proterozoic provinces in Arizona. Our purpose is to discuss and compare isobaric terranes as a tool for understanding crustal architecture, and in particular, lateral variations in characteristics such as composition, structure, and rheological properties that may be difficult to constrain from remote seismic observations or tilted two-dimensional cross-sections.

BACKGROUND—ISOBARICALLY COOLED AND ISOBARIC TERRANES

Metamorphic terranes have traditionally been classified into broad groups based on the style of \( P-T-t \) (pressure-temperature-time) history. Perhaps the most familiar are terranes characterized by near-isothermal decompression during or immediately after orogenesis. These Alpine or Himalayan-type terranes are interpreted to have evolved within a tectonically thickened (or overthickened) crust that underwent thermal relaxation due to syn-orogenic extension or orogenic collapse (England and Thompson, 1984; Platt, 1986; Harley, 1989; Percival et al., 1992). A second common type includes isobarically cooled terranes (Harley, 1989; Percival, 1989; Rudnick and Fountain, 1995). These are metamorphic terranes that record one or more tectonic events during residence at a particular level of the crust for an extended period of time (Ellis, 1987; Harley, 1989). These regions achieved isostatic stability at some time after orogenesis; i.e., they cooled to a normal geotherm and thus offer a more direct view of a particular crustal level in a stable or steady-state environment (Harley, 1989; Percival et al., 1992). Exhumation commonly involves tectonic events unrelated to those that led to initial burial and prograde metamorphism. The first type of terrane is typically described by “clockwise” \( P-T \) paths and the latter by “anti-clockwise” \( P-T \) paths, on a pressure-up \( P-T \) diagram. Both types, however, can have overall clockwise looping \( P-T \) paths (Williams and Karlstrom, 1996).

Isobaric terranes, as discussed here, are relatively large terranes that preserve nearly constant metamorphic pressures. These regions underwent peak deformation and metamorphism at some specific crustal level prior to exhumation to shallower crustal levels. It must be determined if the region was exhumed as a single block or if sub-domains or blocks were exhumed separately by similar amounts and juxtaposed again near the surface. The key point is that the components of these isobaric terranes were present at one specific depth during a tectonic event, and thus can reveal some of the variability in rock types and processes at that particular crustal level. Unlike the isobarically cooled terranes, not all isobaric terranes represent isostatically stable crust, but significant insights can still be gained. Of the three examples discussed below, two (Athabasca granulite terrane and Proterozoic basement of central Arizona) are isobarically cooled terranes. The Upper Granite Gorge of the Grand Canyon is an \(-0.7 \) GPa isobaric terrane that decompressed to 0.3–0.4 GPa late in the orogenic cycle. In addition to the isobaric terranes described here, there are numerous examples from the literature that could be considered. These include the central Sierra Nevada (0.2–0.3 GPa; Ague and Brimhall, 1988), central Otago schist terrane (\(-0.8 \) GPa; Mortimer, 2000), western Maine (0.3–0.4 GPa; Guidotti and Cheney, 1989; Guidotti and Johnson, 2002), and the Napier complex, Antarctica (\(-1.0 \) GPa; Sheraton and Black, 1983; Harley and Black, 1997). Our purpose here is to draw attention to these regions as an underutilized source of information about geologic processes and lateral heterogeneity within the crust.

EXAMPLES OF ISOBARIC TERRANES

Numerous maps and publications exist on the three isobaric terranes discussed below. A brief introduction and overview is presented here, but we refer readers to the cited literature for more detail. Our goal is to compare and contrast certain features and processes, especially those for which new insights about crustal
processes come from the isobaric nature of the terranes. We focus on: (1) the interplay between early sub-horizontal and later sub-vertical fabrics in all three regions, and (2) the character of igneous rocks and advective heating in the three regions. Admittedly, the three terranes evolved in different places at generally different times, but we suggest that important insights and conclusions may be drawn from the comparison. Ultimately, these observations provide a framework for integrating and comparing additional isobaric terranes. Generalized P-T-t paths for the three terranes summarized here are shown in Figure 1.

**Athabasca Granulite Terrane, Saskatchewan, Canada (1.0–1.2 GPa; 35–45 km Depths)**

The region northeast of Lake Athabasca, northern Saskatchewan, is underlain by more than 20,000 km² of high-pressure granulites (1.0 to >1.5 GPa). They occur at the eastern margin of the Rae domain in the western Churchill cratonic province of the Canadian Shield (Fig. 2). A fundamental characteristic of this region is its segmented architecture, with 10- to 50-km-scale, structurally bounded domains that differ from each other in composition and tectonic history. Three domains (Chipman, Northwestern, and Southern) make up the East Athabasca mylonite triangle (Fig. 3A) (Hammer et al., 1994; Williams and Hamner, 2006). The Chipman domain is dominated by the Mesoarchean (>3.0 Ga) Chipman tonalite gneiss with a voluminous 1.9 Ga mafic dike swarm that intruded at conditions of 1.0–1.2 GPa and 750–850 °C, i.e., while the rocks resided in the lower continental crust (Williams et al., 1995; Flowers et al., 2006a). The Northwestern domain is dominated by ca. 2.6 Ga plutonic rocks including the Mary batholith (granite to granodiorite) and the Bohica mafic complex (Hamner, 1997) that record metamorphic conditions of 1.0 GPa and 750–800 °C (Williams et al., 2000). The felsic to intermediate plutonic rocks were emplaced into the deep crust and preserve a record of the structural and metamorphic evolution from igneous Opx- and Hb-bearing granitoids to gneissic garnet granulites (Williams et al., 2000). The white gneiss is interlayered with meter- to kilometer-thick mafic granulite sills that may represent underplating or intraplating of mafic magma in the deep crust. The domain

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**Figure 1.** Representative P-T paths from the three isobaric terranes discussed here. (A) Paths from the East Athabasca granulite terrane. Path-1 represents igneous rocks emplaced at 1.0 GPa and cooled. Path-2 shows rocks that experienced a thermal event while resident in the deep crust. Path-3 experienced a transient period of high-P metamorphism (see Baldwin et al., 2003). Reaction Boundary shows the approximate phase boundary between intermediate-pressure (Opx) and high-pressure (Grt-Cpx) granulites (Williams et al., 2000). (B) Composite P-T path from the Upper Gorge of the Grand Canyon. Essentially all rocks experienced 0.7 GPa metamorphism during the 1.7–1.68 Ga Yavapai orogeny and then were decompressed to ~0.4 GPa (Dumond et al., 2007a). (C) Generalized P-T paths representative of much of the central Arizona (AZ) Proterozoic basement, the lower gorge of Grand Canyon, and Proterozoic basement of northern New Mexico.
also hosts minor eclogite, and all rocks record an early event of higher peak pressures of ~1.5 GPa at 750–1000 °C (Snoeyenbos et al., 1995; Baldwin et al., 2003, 2004, 2007). A large portion of the Rae domain to the west (Figs. 2 and 3) also experienced ~1.0 GPa pressure and is included with the East Athabasca mylonite triangle as the near-isobaric Athabasca granulite terrane. Two main generations of deformation have been identified throughout many portions of the Athabasca granulite terrane, although multiple periods of reactivation have been documented along the high-strain zone boundaries (e.g., Legs Lake shear zone: Mahan et al., 2003; Grease River shear zone: Lafrance and Sibbald, 1997; Dumond et al., 2008). Early structures include sub-horizontal- to northwest-striking gneissic fabrics (S1) and recumbent isoclinal folds of compositional layering (e.g., Slimmon, 1989; Kopf, 1999; Card, 2002; Mahan and Williams, 2005; Dumond et al., 2005a, 2005b; Martel, 2005). Later structures include open- to isoclinal folds of S1 and transposition of older folds and fabrics into sub-vertical, northeast-striking mylonitic foliations (Hammer et al., 1995; Hammer, 1997; Mahan et al., 2003; Mahan and Williams, 2005; Martel, 2005). The eastern boundary of the high-pressure terrane is marked by the Legs Lake shear zone, part of a ~500-km-long contractional fault system that accommodated more than 20 km of vertical displacement (Mahan et al., 2003; Mahan and Williams, 2005; Mahan et al., 2006a, 2006b) (Fig. 3B). The Hearne domain to the east records maximum peak pressures of 0.5 GPa (Mahan et al., 2003).

Rocks in the Northwestern, Southern, and Chipman domains underwent high-pressure (~1.0 GPa) metamorphism at 2.6–2.55 Ga and again at 1.9 Ga. We suggest that the rocks resided in the deep crust from 2.6 Ga to 1.9 Ga, and thus essentially constituted lower continental crust in the region during that period of time (Williams and Hamner, 2006; Flowers et al., 2008). Although the thickness of the crust at that time is unconstrained, the alternative interpretation that the rocks resided within the middle levels of an overthickened crust seems unlikely because of the long duration (700 Ma) of essentially constant pressures and the lack of geologic evidence for decompression to shallower crustal levels between ca. 2.6 Ga and 1.9 Ga. High-pressure conditions (0.8–1.0 GPa, ~900 °C) were recorded at 1.9 Ga in a large area of the Rae domain west of the Grease River shear zone (Figs. 2 and 3) (Kopf, 1999; Krikorian, 2002; Williams and Hamner, 2006). However, these rocks show little or no record of high-grade 2.6–2.55 Ga metamorphism with abundant evidence for prograde metamorphism and decompression at ca. 1.92–1.9 Ga (Kopf, 1999; Williams and Jercinovic, 2002; Martel et al., 2008). Thus, the two lower crustal domains may have been juxtaposed at

Figure 2. Location of the Snowbird tectonic zone and East Athabasca granulite terrane within northern Saskatchewan and Northwest Territories, Canada. Dashed oval shows approximate extent of East Athabasca granulite terrane.
Figure 3. (A) Simplified tectonic map showing the East Lake Athabasca granulite terrane reconstructed by removing slip on the Grease River shear zone (Mahan and Williams, 2005). (B) Simplified geologic map and cross-section (C) of East Athabasca mylonite triangle region after Gilboy (1980), Slimmon (1989), Hanmer (1994), and Mahan et al. (2003). No vertical exaggeration. See text for discussion.
1.9 Ga. Collectively, the region north of Lake Athabasca preserves a large and heterogeneous exposure of the lower continental crust at 1.9 Ga, with parts of the region having existed in the deep crust from 2.6 Ga to 1.9 Ga. Although some of the segmentation and block juxtaposition may have originated during exhumation, at least some of the domain boundaries apparently existed when the rocks resided in the deep crust (Flowers et al., 2006b).

The tectonic setting of the Athabasca granulite terrane is not completely established. It is clear that by 2.62–2.6 Ga, thick rocks resided in the deep crust (Flowers et al., 2006b). We suspect that the segmentation and block juxtaposition may have originated during exhumation, at least some of the domain boundaries apparently existed when the rocks resided in the deep crust (Flowers et al., 2006b).

Williams et al., 2000; Flowers et al., 2008). The final stages of the evolution apparently left a laterally extensive area in a state of near-isostatic equilibrium with little to no post-orogenic decompression recorded. Events at ca. 1.9 Ga are interpreted to reflect a limited amount of rifting with mafic underplating, intraplating, and diking (Flowers et al., 2006a) followed by shortening involving large-scale thrusting, smaller-scale folding, cleavage formation, and exhumation (see Baldwin et al., 2003, 2004, and Ber- man et al., 2007 for alternative interpretation). The shortening is likely related to the continental collision of the Slave province to the northwest followed by the Trans-Hudson orogeny to the southeast, both consistent with regional shortening and dextral shearing in the Athabasca area (e.g., Mahan and Williams, 2005; Dumond et al., 2008).

Proterozoic Basement of the Southwest United States (0.3–0.4 GPa, 10–15 km Depths)

Proterozoic rocks across a large area in central Arizona and northern New Mexico preserve nearly constant peak metamorphic pressures of 0.3–0.4 GPa (Fig. 4). These rocks have been interpreted to have resided within the middle crust from ca 1.7 Ga to at least 1.4 Ga and locally much longer, thus providing an exceptional view of middle crustal processes (Williams and Karlstrom, 1996; Karlstrom and Williams, 2006). The region is divided into tens-of-km-scale tectonic blocks that are bounded by sub-vertical shear zones (Karlstrom and Bowring, 1988). The blocks record markedly different temperatures of metamorphism (400–650 °C) with only moderate differences in pressure. This isobaric terrane provides a view of the 10–15-km-deep level of an accretionary orogen (Karlstrom et al., 2001; Whitmeyer and Karlstrom, 2007). The following description is taken mainly from central Arizona, specifically the Big Bug block (Karlstrom and Bowring, 1988; Karlstrom and Williams, 1995), because a number of descriptive publications are available and because this region is characteristic of much of the Proterozoic basement of central Arizona and northern New Mexico (Figs. 5A and 5B).

Supracrustal rocks in the Big Bug block are dominated by metamorphosed arc-related volcanic and immature (turbidite) sedimentary rocks, and ~50% of the exposed region is composed of granitoids. Although a number of early, arc-related plutons are present, these are intruded by a suite of ca. 1.7–1.65 Ga syn-tectonic granites. The rocks preserve evidence of two deformational events or cycles. The earliest event (D1) is marked by layer-parallel foliation with shallow dip, generally <30–40°. The later event is characterized by upright folds with sub-horizontal enveloping surfaces and sub-vertical foliation in the form of locally intense axial-planar cleavage. Although metamorphic pressures are relatively constant across the terrane, temperatures and D2 fabric intensity increase toward the margins of the syn-tectonic plutons. Three such plutons have been studied in some detail: the 1.7 Ga Crazy Basin monzogranite batholith (Karlstrom and Williams, 1995), the 1.68 Ga Horse Mountain granite, and an unnamed ca. 1.66 Ga granite to the south (Fig. 5B) (Burr, 1991). During regional metamorphism, pluton-related heating led to weakening, fabric development, and higher metamorphic temperatures (Burr, 1991; Williams, 1991; Karlstrom and Williams, 1995, 2006). Thus, the predominant upright, northeast-trending S2 foliation does not represent a single "tectonic event", but rather strain accumulated in different localities at different times over a period of contraction that lasted 50 Ma or more. We call this “pluton-enhanced tectonism” to emphasize the critical role of magmatism in localizing deformation and metamorphism.

Rocks in the Big Bug block are interpreted to have existed at 0.3 GPa to 0.35 GPa (10–15 km depths) during the period of
Figure 5. (A) General tectonic model showing tectonic blocks and high-strain D₂ shear zones in Proterozoic rocks in the central Arizona Proterozoic basement (modified from Bergh and Karlstrom, 1992). Arrows represent general motion vectors during 1.71–1.65 Ga tectonism. Inset shows area of Figure 5B. (B) Detail of the northern end of the Crazy Basin quartz monzonite and location of major metamorphic isograds (modified from Karlstrom and Williams, 1995).
syn-tectonic plutonism from 1.7 to at least 1.65 Ga (Williams, 1994; Karlstrom and Williams, 1995). Regional thermo-tectonic events between ca. 1.48 Ga and 1.35 Ga are also recorded at pressures near 0.3–0.4 GPa, and thus it is suggested that large parts of the Proterozoic terrane were in the mid-crust for the intervening 200 Ma. Exhumation occurred in some areas soon after the ca. 1.4 Ga event, but much later in other areas. Thus, as with much of the Proterozoic terrane of the southwestern United States, the Big Bug block is an isobarically cooled terrane that preserves a view of the lateral heterogeneity in the shallow mid-crust over several hundred millions of years.

Tectonic models for Proterozoic basement of the southwestern United States involve arc accretion, assembly, and stabilization that resulted in addition of new continental lithosphere along a long-lived active margin (Bowring and Karlstrom, 1990; Karlstrom et al., 2001; Whitmeyer and Karlstrom, 2007). Some of the arcs may have developed outboard of the continent, followed by assembly during arc-continent collisions. On a regional scale, these isobaric terranes generally represent first-cycle formation and stabilization of normal thickness continental crust from thinner oceanic fragments (Bowring and Karlstrom, 1990). The long-term residence of these rocks in the middle crust has been interpreted in terms of isostatically stable lithosphere and slow cooling, which was perturbed by the 1.48–1.35 Ga tectono-magmatic episode. The isobaric character of the metamorphic terrane has been interpreted in terms of an orogenic plateau with steep thermal gradients, perhaps above a ca. 1.4 Ga magma-fluid layer (at times of elevated temperature), analogous to the Tibetan plateau (Shaw et al., 2005).

**Proterozoic Basement of the Grand Canyon (0.6–0.7 GPa; 20–25 km)**

The Upper Granite Gorge of the Grand Canyon (Figs. 4 and 6) represents an important exception to the 0.3–0.4 GPa peak pressures that characterize much of the Proterozoic basement of the southwestern United States. The volcanic and immature sedimentary rocks, and the structural styles are similar to those across the region, but peak pressures of 0.6–0.7 GPa (20–25 km depths) are recorded. The region also preserves a similar block-type crustal architecture, consisting of multi-km-scale blocks characterized by relatively homogeneous deformation and metamorphism bounded by sub-vertical high-strain zones (Fig. 6B). Variations in metamorphic conditions along the Upper Granite Gorge transect are primarily thermal in nature and include differences in: the temperature of the prograde history (i.e., early andalusite versus kyanite), peak temperature, and the intensity of late-stage thermal spikes due to the local emplacement of plutons and dike swarms. High-precision $\Delta PT$ “relative” thermobarometry confirms lateral temperature variations on the order of 100–250 °C with little to no variation in pressure (Dumond et al., 2007a). The Upper Granite Gorge thus represents a sub-horizontal section of lowermost middle continental crust (~0.7 GPa) at least at the time of peak metamorphism. The entire ~70 km-long transect apparently decompressed from ~0.7 GPa to ~0.3–0.4 GPa levels (back to the regional middle crustal level) as one large coherent block immediately after tectonism in the Paleoproterozoic (Dumond et al., 2007a).

Tectonic models for this region are similar to models proposed for other parts of the southwestern United States, involving stabilization of dominantly juvenile Proterozoic crust during arc collisions. One shear zone within the Upper Gorge isobaric terrane has been interpreted as a suture zone between different arcs (Ilg et al., 1996). Hence, this isobaric terrane, like the others, probably includes juxtaposition of different crustal levels, but the peak pressures are interpreted to record stabilization of an accretionary “duplex system” at 20–25 km depths near the end of the convergent process (~1.7–1.69 Ga; Ilg et al., 1996). Exhumation of this terrane to 10–15 km levels at ca. 1.68 Ga may reflect continued syn-shortening exhumation (thrusting plus erosion or fault-removal of overlying material) of a large isobaric region within the same orogenic cycle before long-term residence (1.68–1.4 Ga) at ~10–15 km depths (Dumond et al., 2007a). Extensional collapse of overthickened crust might also have been active, but syn-decompressional extensional structures have not been identified at exposed crustal levels.

During this same time interval, rocks southeast of the Big Bug block, called the “Mazatzal block,” appear to preserve a region that was at shallow crustal levels. The Mazatzal block consists of 1.73 Ga Payson ophiolite as basement (Dann and Bowring, 1997) to ca. 1.70 Ga quartzites that were deposited at the same time as 1.70 Ga emplacement of the Crazy Basin batholith. The deformational architecture of the quartzites is that of a shallow-level fold-thrust belt (Doe and Karlstrom, 1991). Although the timing and movement history on the shear zones that juxtapose this level with the 10 km isobaric terrane of the Big Bug block remain incompletely understood (Karlstrom et al., 1987), this snapshot of upper-crustal levels can be integrated with the isobaric sections in terms of a schematic whole-crustal column, and reinforces the importance of understanding the mechanisms of juxtaposition and ultimate preservation of different crustal levels within an orogenic system (e.g., Klepeis et al., 2007).

**OBSERVATIONS AND COMPARISONS**

**Lateral Heterogeneity**

A first-order characteristic of each of the isobaric terranes discussed above is their heterogeneity of rock types, structures, fabrics, and metamorphic assemblages. Because each of the regions is interpreted to represent either a contiguous or disrupted but reconstructable isobaric terrane, it is possible to appreciate the extreme lateral variation in almost all of the characteristics that control rheology. Rocks vary from mafic to felsic, massive to layered, regular to contorted, highly strained to relatively undeformed, porphyroblast-rich to porphyroblast-poor, and relatively wet to relatively dry. Even the scale of heterogeneity is variable. It would certainly not be possible to assign a single rock type,
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Interpreting geophysical images and crustal rheology must involve understanding the nature, evolution, and implications of this heterogeneity at all scales, as emphasized (but not yet achieved) by recent thermal-mechanical numerical modeling efforts (Jamieson et al., 2007; Groome et al., 2008).

Lateral compositional heterogeneity also evolves; i.e., it can be enhanced or reduced during the orogenic process and/or during long-term residence at a given crustal level. Mafic underplating and intraplating significantly changes bulk density, rheology, and thermal structure as do metamorphic reactions (particularly those involving growth or dissolution of garnet). This is particularly relevant for interpreting geophysical data. It is important to correlate seismic images with data about the history of the particular crust being imaged (Rutter et al., 1999; Magnani et al., 2004; Karlstrom et al., 2005). This record of

![Diagram of the Upper Gorge of the Grand Canyon](image)
evolution through time can, in some cases, provide a template that may be used to help interpret the dynamics of modern orogens that are imaged geophysically (e.g., Mahan, 2006). For example, seismic images of Magnani et al. (2004) show strong deep-crustal reflections that are interpreted to be of vastly different ages. Prominent but somewhat diffuse bands of reflectivity are interpreted as a Proterozoic bivergent collisional system, and these reflections are cross-cut by sharper bright reflectors interpreted to be mafic sills emplaced during much later (either 1.1 Ga or Miocene) magmatism.

Shallow and Steep Fabrics

Despite the differences in crustal level, a number of structural characteristics are shared by each of the isobaric terranes investigated here. These include (1) the presence of early sub-horizontal or shallowly dipping deformation fabrics (S₁) and later, variably intense, upright structures and fabrics (S₂); (2) a resulting block-type architecture that reflects the distribution of zones or domains with minimal overprinting by S₂ fabric and those with intense transposition and S₃ fabric development; and (3) the scale of this structural heterogeneity, best developed on the 10 km scale, but expressed from micrometers to tens of kilometers. In each region, domains dominated by shallowly dipping and gently folded S₁ fabric are separated by domains of steep S₂ fabric. In some areas, the high-strain S₃ domains can be classified as shear zones across which rock types and structures cannot be readily correlated, but others are zones with only slight offset of earlier markers (Ilg et al., 1996). There is evidence for more than one generation of folds and cleavage within the overall shallow and steep fabric domains (i.e., S₁₁, S₁₂, S₂₁, S₂₂, etc.), but it is commonly difficult to break out regionally correlative fabric-producing sub-events. This may suggest that the transition from shallow to steep deformation reflects not just a new pulse of deformation, but a more fundamental change in rock properties and bulk strain.

Timing issues are critical and can be extremely difficult to resolve. One of the first-order questions concerning the significance of the block-type architecture is the extent to which the block geometry was established when the domains existed at depth in the crust (perhaps inheriting even earlier D₁ accretionary boundaries), if it was established during progressive crustal shortening (D₂), and/or during exhumation. A further complication is that block-bounding shear zones commonly get reactivated during even later (post-D₂) more localized movements. These later movements span the ductile and brittle strain regimes and typically coincide with exhumation of the region.

The steep fabrics, structures, and strain gradients are typically the most apparent features, especially on geologic maps and cross sections, although perhaps not on seismic images (e.g., reflection seismic transparency of vertical anisotropy: Levander et al., 1994). These might be initially taken as the dominant structures representing the dominant tectonic event(s). However, in each area, closer inspection reveals that the earlier fabrics are more penetrative and pervasive, at least outside of intense block-bounding shear zones. These early sub-horizontal fabrics represent fundamental components of the tectonic history, and are critical for understanding the crustal evolution and crustal processes in general. The less reoriented shallow fabrics would be preferentially imaged in reflection seismic profiles and thus, domains of strong versus weak reflectivity may correspond to domains where predominantly shallow versus steep fabrics are preserved. Thus, domains of weak reflectivity may correspond to vertically foliated crust rather than zones of low signal/noise in the seismic data acquisition process.

P-T Context of Steep and Shallow Fabric Development

Understanding the significance of the steep and shallow fabrics involves understanding the depth and timing of deformation events, i.e., the P-T-t-deformation history, of the block-bounding shear zones relative to the history of the blocks themselves. Recent work in the Upper Granite Gorge of the Grand Canyon (Dumond et al., 2007a) illustrates some of the methods and challenges associated with constraining these structures. Thermobarometric (P-T) calculations, and the ability to date metamorphic fabrics via geochronology, generally have large uncertainties, making it difficult to recognize pressure differences across structures. Techniques such as relative P-T calculations can provide better resolution but it can be difficult to separate thermal and baric effects (Hodges and McKenna, 1987; Worley and Powell, 2000; Dumond et al., 2007a).

All three terranes discussed here contain both earlier shallow and later steep fabrics (Fig. 7), but there are important differences with respect to the timing of fabric development and its relationship to metamorphism. In the Athabasca granulite terrane, early shallow fabrics generally coincide with the highest-temperature metamorphic conditions. Hundreds of km² areas of early fabrics are associated with intense melting followed by isobaric-cooling and L >> S to L-tectonite development (Dumond et al., 2005a, 2006). Where evidence for melting is lacking, early high-temperature (>700 °C) sub-horizontal fabric development is synchronous with lower-crustal emplacement and cooling of granitoid plutons and batholiths (Dumond et al., 2005a, 2005b) and syn-metamorphic recrystallization (Williams et al., 2000). It has generally not been possible to identify even earlier, prograde, fabric elements that may pre-date the oldest plutons because most of the rocks probably originated as deep-crustal plutonic rocks. Steep fabrics in the deep crustal rocks are generally more discrete and commonly indicate lower temperature conditions.

In contrast, the shallowly dipping S₁ fabric in both Proterozoic terranes of the southwestern United States apparently initiated at lower grade and is interpreted to reflect prograde metamorphic conditions. Peak conditions were generally associated with the D₃ events (Williams, 1994; Karlstrom and Williams, 1995; Ilg et al., 1996; Dumond et al., 2007a). However, an additional characteristic of the Upper Granite Gorge is that the shallow fabrics locally preserve composite histories where
Figure 7. Photos showing early shallow ($S_1$) and steep ($S_2$) fabrics in East Lake Athabasca and the Grand Canyon. (A) Shallow $S_1$ surface, northwestern domain, East Lake Athabasca with strong lineation parallel to hammer. (B) $S_1$-overprinting of $S_1$ migmatite/restite “white gneiss”, Southern domain, East Lake Athabasca granulite terrane. (C) $S_2$-overprinting of $S_1$ in tonalitic orthogneiss, eastern Chipman domain, East Athabasca. Dike is interpreted to represent a syn-$S_2$ Chipman dike. Camera lens cap for scale. (D) Large early intrafolial fold with $S_1$ defining the axial plane, mile-102, Grand Canyon. (E) Andalusite crystal with $S_1$ inclusion trails and $S_2$ external foliation, mile 84, Grand Canyon. (F) $S_1$-$S_2$ relationships from sub-vertical thin section cut normal to $S_1$ foliation and parallel to lineation in Grt-Bt-Ms-Pl-Qtz schist from Upper Granite Gorge, Grand Canyon (scale ~3 mm: modified from Ilg and Karlstrom, 2000).
the \( S_1 \) fabric initially formed during prograde metamorphism but was reactivated and modified during peak metamorphism and \( D_2 \) deformation (Fig. 8).

Differences in timing relationships, especially with respect to metamorphism, have led to differences in the interpretations and implications of the fabric transition in the deep and mid-crustal examples. In the former (Athabasca), the shallow fabrics are interpreted to result from weakness of the deep crust possibly involving flow due to differential lithostatic load (e.g., Royden, 1996) or mantle dynamics (Tikoff et al., 2004). In the middle crust of the southwestern United States, the early fabrics are interpreted to represent collisional fabrics, perhaps shear zones related to early thrust sheets that formed during crustal thickening (Karlstrom and Williams, 2006). These differences have important implications (discussed below) for interpretations of rheology and seismic images.

**Absolute Timing Constraints**

Absolute timing constraints are critical for interpreting and correlating fabrics in any poly-metamorphic and poly-deformed terrane, particularly the isobaric terranes discussed here. By their nature, these terranes spend an extended period of time at some level of crust, and thus generally preserve fabrics that formed during several, perhaps widely separated, deformation events. Traditionally, ages of fabrics are constrained by dating igneous units that either contain or cross-cut the fabric. However, it can be difficult to isolate the age of one particular fabric in a multiply deformed rock, strain-partitioning can lead to difficulty in determining if an igneous rock was or was not affected by a particular deformation, and fabrics commonly develop diachronously and hence cannot be used as time markers.

U-Pb monazite geochronology, and especially in situ monazite dating, is proving to be a useful tool for constraining the age of metamorphic reactions and deformational fabrics (Hawkins and Bowring, 1999; Williams and Jercinovic, 2002). Under the right circumstances, monazite itself can be a fabric-forming mineral, an inclusion in a fabric-forming mineral, or a product in a reaction that occurred during a fabric-forming event (Shaw et al., 2001; Williams and Jercinovic, 2002; Mahan et al., 2006b). Monazite overgrowths can grow on earlier grains in the extensional quadrants of the strain ellipsoid; and as such can be used to directly date a deformation event or pulse (Dumond et al., 2008). Further, monazite trace elements can be balanced into metamorphic reactions and thus used to directly date a segment of the \( P-T \) path and associated deformation (e.g., Pyle and Spear, 2003; Foster et al., 2004; McFarlane et al., 2005; Mahan et al., 2006b).

Careful in situ work can produce precise constraints (with \( 2\sigma \) errors on the order of 10 Ma) on the age of particular fabrics or reactions, and thus allow the history of isobaric terranes to be interpreted and correlated.

Mahan et al. (2006b) used in situ monazite analysis to constrain the age of the major thrust zone that forms the eastern boundary of the Athabasca granulite terrane, and that played a key role in the ultimate exposure of the region. Dumond et al. (2005a, 2005b, 2007b) used similar techniques to show that early shallow fabrics in the terrane are Neoarchean and correspond with high-temperature peak metamorphism in the region (Fig. 9A). Multiple generations of steep, cross-cutting fabrics are generally Proterozoic in age and are associated with rheologically stronger conditions. The earliest \( S_1 \) fabrics were synchronous with ca. 1.9 Ga granulite-grade metamorphism and were subsequently cross-cut by younger retrograde steep fabrics (Dumond et al., 2006) (Fig. 9B). In contrast, shallow and steep fabrics overlap in age in the Upper Granite Gorge (\( D_2/S_1 \) at 1730–1698 Ma and \( D_2/S_2 \) at 1713–1685 Ma; Hawkins et al., 1996; Karlstrom et al., 2003), with the steep fabrics corresponding to peak metamorphism. In central Arizona (Big Bug block), peak metamorphism was also synkinematic with \( D_1 \) steep fabrics and both fabric intensity and metamorphic temperatures increase with proximity to 1.70–1.68 Ga granitoids (Karlstrom and Williams, 1995). These absolute constraints help to establish a fundamental difference in
Figure 9. In situ monazite geochronologic constraints on timing of deformation and metamorphism in rocks from East Lake Athabasca. Monazite overgrowths tend to grow in extensional quadrants of strain and are linked chemically to syn-tectonic reactions (Dumond et al., 2006, 2007b). (A) syn-tectonic monazite associated with Archean $S_1$ deformation. Small Paleoproterozoic overgrowths on top and bottom edges of grains reflect $S_2$ reactivation. (B) syn-tectonic $S_2$ monazite associated with the upright Grease River shear zone.
the shallow-to-steep fabric transition between the deep and mid-crustal terranes.

Character and History of Plutonism

The nature of igneous rocks and especially their role in deformation/metamorphic processes is another characteristic that can be evaluated using isobaric terranes. Variation in size, geometry, distribution, age, and composition of the igneous bodies is common, and thus the lateral exposure of isobaric terranes allows a more robust assessment of these characteristics through time. Particularly valuable is the ability to evaluate the nature of deformation and metamorphic gradients between domains with a relatively more and relatively less abundant igneous component. These gradients have an important connection to the issues of deep and shallow fabric evolution and the regional block-type architecture.

Athabasca granulite terrane. The deep-crustal rocks of the eastern portion of the Athabasca granulite terrane are dominated by plutonic rocks (tonalite, granodiorite, charnockite, mafic dikes and sills). Although several workers have suggested that some migmatitic rocks were derived from sedimentary protoliths, it is likely that some or many of these rocks were also igneous in origin (i.e., derived from anatexis of orthogneisses: Baldwin et al., 2006). In nearly all areas, the emplacement of new igneous rocks was intimately involved with metamorphism, deformation, or recrystallization of preexisting rocks; that is, an important component of the thermal budget for generating these deep-seated rocks involves advective heating from plutonic rocks. For example, one of the most striking features of the southern domain is the interlayering of thick mafic granulite sheets with garnet-rich felsic granulite (Figs. 10A, 10B) (diatectite of Hanmer, 1994; “white gneiss” of Snoeyenbos et al., 1995; Baldwin et al., 2006). The mafic sheets are interpreted to have been emplaced as thick sills or “intralaves” of mafic magma (Baldwin et al., 2006). The felsic granulites are interpreted to be the restitic products of large mafic dikes, one to ten meters in thickness, were emplaced into the Mesoproterozoic Chipman tonalite. Earlier dikes were metamorphosed and locally melted by heating interpreted to result from later Chipman magmatism, involving dikes and underplating. D2 deformation was localized in the region of the dike swarm (Figs. 10C, 10D).

Upper Granite Gorge. Mafic, intermediate, and felsic arc-related plutonic and volcanic rocks are volumetrically significant components of the lowermost middle continental crust represented by Upper Granite Gorge of the Grand Canyon. Unlike the Athabasca region, the metasedimentary and metavolcanic rocks of most blocks were at or near the surface prior to the onset of tectonism. Thus, the nearly contemporaneous arc-related plutons were intruded relatively shallowly into low-temperature metasedimentary rocks, and had minimal effect on the rheology or metamorphic grade. In contrast, magmatism associated with peak metamorphism and D2 deformation was dominantly 1.7–1.68 Ga granite and pegmatite (Hawkins et al., 1996). These felsic rocks occur primarily as swarms of meter-scale dikes associated with three large complexes (Figs. 10E, 10F). The dike swarms are spatially associated with increased metamorphic temperatures (Ilg et al., 1996; Dumond et al., 2007a). Although there is evidence for some in situ melting, a significant proportion of the melt is interpreted to have been introduced from below Fig. 10G. Karlstrom and Williams (2006) suggested that these dike complexes represent granite melts that were in transition to higher structural levels; they acted as conduits feeding shallow-level ca. 1.7–1.68 granite plutons. In addition, some zones of intense S2 fabric are strongly migmatitic with abundant cm-scale leucosomes, suggesting that they may have been conduits through which magma has passed in a more dispersed fashion. We suggest that the Upper Granite Gorge isobaric section captures a zone of transport through which felsic melts were passing to higher levels of the crust. This through-put certainly contributed advective heat and dramatically modified the bulk composition of the crust at this level, by the addition

\[ \text{Opx} + \text{Ca-Pl} = \text{Grt} + \text{Na-Pl} \]  

occurred during D2 deformation and recrystallization may have been aided by the emplacement of later stages (sheets) of the same pluton (Williams et al., 2000). In this case, heat associated with younger phases may have contributed to metamorphism of older phases during the progressive development of the S2 gneissic fabric. Thus, intrusion, heating, and migmatization of deep-crustal rocks are associated with crustal flow (S1) and a relatively weak stage in the history of these rocks. Dehydration from melt loss and cooling may lead to strengthening and the shift from shallow to upright deformation phases (Dumond et al., 2005b; e.g., Klepeis et al., 2003, 2004). Igneous rocks are also associated with the younger S2 fabrics. Granitoids and pegmatitic dikes were emplaced along zones of intense D2 fabric and most also contain the D2 fabric. The D2 high-strain domains may have facilitated emplacement of these granitoid bodies, but the presence of magma may have equally played a role in weakening the crust and allowing localization of the D2 strain (e.g., Hollister and Crawford, 1986). The Chipman dike swarm (Williams et al., 1995; Flowers et al., 2006a) is one noteworthy example. Near vertical, northeast-striking mafic dikes, one to ten meters in thickness, were emplaced into the Mesoarchean Chipman tonalite. Earlier dikes were metamorphosed and locally melted by heating interpreted to result from later Chipman magmatism, involving dikes and underplating. D2 deformation was localized in the region of the dike swarm (Figs. 10C, 10D).
Figure 10. Photos showing relationships with igneous rocks. Photo (A) and line drawing (B) of sub-horizontal mafic granulite sills, derived from gabboic protolith, emplaced in southern domain, East Athabasca. The sill-like bodies of mafic granulite (and locally eclogite) contain the penetrative $S_1$ fabric, and are intensely veined internally and at their contacts by felsic leucosomes. Relatively undeformed (C) and strongly deformed within $S_2$ fabric (D) sub-vertical mafic dikes, Chipman dike swarm, East Athabasca (E) Large granitic-pegmatitic dike and strong $S_2$ fabric, Upper Gorge, Grand Canyon. (F) Pegmatite swarm near Phantom Ranch, Upper Gorge, Grand Canyon. (G) Close-up of migmatitic schist, Upper Gorge, Grand Canyon, interpreted to be injection migmatites in a zone of magma through-put.
of abundant granitoid materials. The Upper Granite Gorge iso-
baric terrane provides a excellent opportunity to compare the
cooler less-modified rocks, away from the dike complexes, to
those that have served as conduits for magma transfer. Unlike
the Canadian deep crustal terrane, there is a notable lack of syn-S2
mafic dikes in most of the exposed Upper Granite Gorge terrane.
We speculate that such magmas may not have been able to pass
through a zone of more abundant migmatite somewhat below the
20–25 km exposure level (Vernon et al., 1990). This level might
be associated with the 16–21-km-deep “bright” seismic reflectors
imaged by the INDEPTH profile beneath the Tibetan Plateau that
are interpreted to mark the top of a mid-crustal partial melt zone
(Nelson et al., 1996).

**Central Arizona basement.** At least 50% of the ~0.3–
0.4 GPa basement rocks in central Arizona consists of grano-
diorite to granite. Similar to the Upper Granite Gorge, the gra-
ndiorite complexes tend to be older and more mafic, and are
interpreted to be portions of arc batholiths. However, some of
the largest plutons are ca 1.7–1.65 Ga syn-tectonic granite bath-
liths. For example, the Crazy Basin quartz monzonite (Fig. 5) is
composed of a large number of sheets and dikes (Karlstrom and
Williams, 1995), and screens of country rocks are present, but
plutons can be readily distinguished. It is likely that some magma
passed to higher levels of the crust and perhaps erupted, but exposures at 0.3–0.4 GPa (10–15 km paleodepth) level are
distinctly different from those of the Upper Granite Gorge. This
crustal level is considered to be one of pluton-building rather than
the site of through-put as inferred for the Upper Granite Gorge.
This is compatible with the hypothesis that the 10–15-km-deep level represents a zone of neutral buoyancy for granitoid magmas
(Vernon et al., 1990).

Calculated metamorphic pressures are relatively constant
across much of the central Arizona basement terrane (0.3–0.4 GPa),
but temperatures, and D2 fabric intensity, increase toward the mar-
gins of syn-tectonic plutons. Regional (“background”) metamor-
phic temperatures may have been as low as 350–400 °C, but tem-
peratures reached 600 °C or higher near plutons (Karlstrom and
Williams, 1995). As in the Upper Granite Gorge, ca. 1.7–1.65 Ga
crustal rocks are an important component of the thermal
budget, but the effects seem to be much more localized around
the large plutons. Even though plutonic rocks are extremely abun-
dant, migmatites are rare. They occur primarily near igneous bod-
ies and in country rock screens where temperatures were highest.
These migmatites are mainly injection migmatites and may have
had the least affect on the overall rheology of this crustal level,
although they may have helped accommodate pluton expansion
during progressive syn-tectonic emplacement.

**Summary of plutonic observations.** All three isobaric ter-
ranes record different expressions of the transfer of heat and
magma through the crust. At deep crustal levels, heat and mass
are transferred from the upper mantle to the deepest levels of con-
tinental crust via basaltic magmas. In appropriate bulk composi-
tions, the mafic rocks can lead to extensive melting with weak-
ening, flow, and ultimately, upward transfer of tonalitic magma.
The 20–25 km level preserved in the Grand Canyon may be pri-
marily a zone of granite magma transfer with local partial melt-
ing. The heat and fluids play an important role in partitioning S2
fabric and localizing high-strain zones that serve to maintain and
enhance the segmented block-type architecture. The relative lack
of mafic igneous rocks in the Grand Canyon, in contrast to the
abundant mafic dikes and sills in the Athabasca terrane, may sug-
gest that mafic magmas were “filtered” at some deeper level. The
10–15-km levels preserved across the southwestern United States
host abundant granitic plutonic rocks with many plutons built by
successive dike emplacement. This is clearly a level of accumula-
tion of felsic igneous rock and the locus for a dramatic feedback
between magmatism and deformation: the D2/S2 strain/fabric
accommodated magma emplacement and in turn, the magma led
to weakening and strain localization.

The combined picture is one of crustal differentiation and
evolution driven by heat and magma and with important
feedbacks with deformation and metamorphism at each level
(Fig. 11) (see also Buddington, 1959; Pitcher, 1979; and Holm
et al., 2005). Important questions in interpreting seismic sections
include: (1) At what stage in this type of evolution is the particu-
lar seismic image? (2) Can we recognize the transient magma-
related weakening within the overall image, as was postulated
from the INDEPTH sections (Nelson et al., 1996) and from
“bright spots” in rift zones (Brown et al., 1980); and (3) How is
the history of the rock material presently captured in these images
likely to influence its rheology and geophysical properties? Iso-
baric terranes can provide data on the distribution, abundance,
and deformational character of igneous rocks through time that
two-dimensional sections may miss or misrepresent.

**IMPLICATIONS**

Figure 11 is a composite section showing characteristics of
the isobaric terranes highlighted here on a single depth axis. The
sections that come from southwestern Laurentia involve rocks
and tectonic events of similar age and thus might be reasonably
stacked into a theoretical crust (e.g., central Arizona above the
Upper Gorge section). However, our main purpose in stacking
the sections is not to build a theoretical (model) crust, but to pro-
vide a comparison of tectonic style and process at specific depths
as a general illustration for the following discussion.

**Implications for Crustal Flow Models**

Lateral flow in the crust at one crustal level or another has
been increasingly invoked in models for the evolution of orogenic
belts (Beaumont et al., 2001; Hodges et al., 2001; Grujic et al.,
2002; Williams and Jiang, 2005; cf. Law et al., 2006). From fluid
mechanics, two broad types of flow have been distinguished:
Couette flow, in which vergence remains constant across the zone
of flow and Poiseuille flow, in which vergence reverses within
the zone of flow (e.g., Turcotte and Schubert, 2002: cf. Ch. 6).
Couette flows are essentially shear zones and might be considered
as detachment or partial detachment zones (Tikoff et al., 2002; Williams and Jiang, 2005). Poiseuille (pipe) flows are generally called channel flows and involve the ductile extrusion of crustal material between two rheologically stronger crustal layers. Both types of flow, but particularly channel flow, have been used to reconcile surface topography and surface processes with crustal and/or mantle dynamics (e.g., Clark and Royden, 2000). Many interpretations have been based on numerical or analog models or on interpretations of surface topography and structure (Royden et al., 1997; Beaumont et al., 2001, 2004), and most models are fundamentally two-dimensional. Isobaric terranes offer an opportunity to evaluate the 3rd dimension, the aerial extent, of lateral flow processes and the conditions that promote or inhibit crustal flow. This is particularly important for interpretations based on seismic images that might preferentially image gently-dipping fabrics (TRANSLP Working Group, 2002; Meissner et al., 2006).

Sub-horizontal transposition fabrics are extremely common, both in the examples described here and in many other metamorphic terrains (e.g., Williams and Jiang, 2005). One obvious first-order question involves evaluating the nature of the strain or flow that is implied by these fabrics. Distinguishing channel flow from a shear zone or detachment involves the recognition of a reversal in vergence within the zone. At first, this seems less feasible in an isobaric (i.e., horizontal) crustal exposure rather than in a crustal cross-section. However, many isobaric terranes are at least somewhat oblique allowing some depth perspective, or as in all of the isobaric terranes described here, internal faults and shear zones within the terrane can allow evaluation of slightly different depths. The isobaric terranes thus allow a broader survey of kinematics, deformational style, and strain partitioning at each level.

A comprehensive evaluation is not yet complete in any of these terranes. However, the sense of shear on the early (S1) fabric in the Athabasca granulite terrane is remarkably consistent (i.e., top to the southeast), suggesting that this may be more a zone of partial detachment (Couette-type flow). If the fabrics do represent a channel flow, consistent with crustal extrusion, then only part of the channel is exposed. In the southwestern U.S. terranes, vergence on S1 is more difficult to determine because of the intensity of syn-D2 peak metamorphism. The orientation of the early mineral lineation is regionally consistent and the northwesterly orientation is also compatible with paleogeography and regional tectonic constraints (Jessup et al., 2005; Whitmeyer and Karlstrom, 2007). These may favor a detachment style of flow that is more typical of processes in accretionary orogens (e.g., Platt, 1986) rather than a channel flow.

Tikoff et al. (2002) identified three general levels of partial detachment or “clutch zones” that might be expected within continental lithosphere: one in the mid-crust, one in the deep crust above the Moho, and one near the lithosphere-asthenosphere boundary. The isobaric terranes described here provide exposures of the upper two of these levels. The following comments are meant to illustrate that isobaric terranes provide an important perspective and data source for understanding crustal flow, especially the conditions under which flow may occur or be inhibited, particularly in a typical heterogeneous crust.

Figure 11. Schematic diagram showing isobaric sections from East Lake Athabasca and Proterozoic terranes from southwestern North America. This is not meant to depict a model crust but instead to provide a means of comparing tectonic styles and processes at different crustal levels. See text for discussion.
The basic structural geometry of the three crustal levels described here is extremely similar, i.e., early penetrative S1 fabric overprinted by upright S2 fabric domains and shear zones. The details of the P-T history and the timing of fabric development relative to metamorphism, however, are different, and in turn, the interpretation of the S1 fabric in terms of crustal flow is different. In the Athabasca granulite terrane, the early shallow fabric was developed regionally near peak metamorphic temperatures, with extensive migmatite development. This event represents the principal dehydration of biotite- and amphibole-bearing rocks. The rocks were apparently at their weakest during the development of the sub-horizontal S1 fabric (Dumond et al., 2005b). If the S1 fabric was the product of channel flow it might be interpreted to be analogous to the hypothesized flow of the deepest crust eastward away from the Tibetan Plateau today (Royden, 1996; Royden et al., 1997; Clark and Royden, 2000; Shapiro et al., 2004). However, indicators associated with S1 are abundant and suggest a uniform vergence, as would be expected for a deep crustal partial detachment zone immediately above the Moho. Thus far, there is no evidence for a reversal of vergence or extrusion of material. Without such a reversal, the early fabric more likely represents a large scale shear zone or “clutch zone” (Tikoff et al., 2002). Without such a reversal, the early fabric more likely represents a detachment or “clutch zone,” and the early S1 time, even though metamorphic temperatures were locally as hot or hotter than during S1 formation, the lower crust was rheologically stronger, mainly because it was drier and less fertile for melting. The lower crust was segmented by steep S1 domains and shear zones and there was little tendency for lateral flow.

In both of the mid-crustal examples, the 20–25 km Grand Canyon and the 10–15 km central Arizona terrane, S1 apparently developed during prograde metamorphism. During S1 strain, the rocks were generally cooler with little or no melt component. The gently dipping deformational fabric may represent the deeper parts of thrust faults or shear zones associated with thickening and prograde metamorphism in an accretionary system. The metamorphic peak, amphibolite or lower granulite facies, occurred during the formation of the upright (S2) fabrics, with local reactivation of S1. The crust was highly partitioned into the S1 and S2 domains, despite the abundance of water (hydrated phases like muscovite and biotite) and locally significant melting. There is little evidence for crustal flow at this time, even though the rheology of the hot domains was probably appropriate for flow. A regional channel flow would require a dramatic disruption of the segmented crustal architecture that seems to characterize this crustal level through a vast region and thus would probably require significant additional weakening and/or melting (see also Beaumont et al., 2004). The earlier S1 fabric may represent a detachment or “clutch zone,” and the prograde character of the P-T-t history indicates a connection with thrusting and crustal thickening.

An important characteristic of the S1 fabric in each isobaric terrane is its lateral heterogeneity. The fabrics change from L-tectonite to L-S tectonite to S-tectonite, and some blocks within the overall flow regime were apparently characterized by a much stronger rheology such that the early flow fabric was partitioned against or around these blocks (Karlstrom and Bowring, 1988). The fabric heterogeneity may be one strong control on the degree of coupling or detachment across the zone. Ultimately, it will be possible to build a map of strain and flow at specific crustal levels and specific times and even possibly stack appropriate maps to build a three-dimensional image of flow and coupling. These can be compared and combined with interpretations from tilted crustal sections (e.g., Klepeis et al., 2003, 2004, 2007).

Channel flows and/or detachment zones appear to require a special combination of conditions. Particular crustal levels may only exhibit a susceptibility to flow or detachment (i.e., weak behavior) during specific times in their P-T-deformation evolution, a behavior that is also heavily dependent upon bulk composition and fertility for melting. These interpretations can only come from accurately timing the P-T history and linking it to the tectonometamorphic history and thus are virtually impossible to infer from remote seismic images. More speculatively, regionally extensive isobaric domains may be a record of orogenic plateau development, such that flowing crust at some level precluded the maintenance of large topographic relief at the surface. Multi-disciplinary study of the distribution of isobaric terranes consequently provides important data for reconstructing aspects of the topography of ancient mountain belts, in addition to providing a window into orogenic plateaus at depth.

Implications for Interpretation of Seismic Data

The block-type architecture and common presence of steep and shallow fabric domains is important for the interpretation of seismic images of the middle and deep crust. Steep shear zones may not be directly recognized in many seismic reflection surveys and may show up as velocity transitions between disparate blocks, seismically slow and seismically fast domains, in teleseismic images (e.g., Levander et al., 1994). The presence of steep deformation zones may be inferred based on differences in the character of juxtaposed domains, but the presence of a relatively intense S1 in most domains may lead to similar reflectivity, especially because similar metamorphic grades are generally involved in each case. Consequently, the relative invisibility of the steep domains may lead to erroneous interpretations about the continuity of domains and the associated subhorizontal structures. One might get the impression that lateral transport of material may have occurred or at least that there is a relatively homogeneous structural character. In addition, the significance of fabric domains for the overall tectonic history is difficult to deduce remotely, and similar architectures may represent very different tectonic histories.

The metamorphic grade and P-T history of a region are also very important for interpreting seismic data. Each terrane shows evidence for dramatic changes in mineralogy, fluid content, and texture during orogenic events and also after, perhaps long after, tectonism. For example the progressive growth of garnet at the expense of orthopyroxene or hornblende in deep-crustal charnockites may contribute to densification of the deep crust long after tectonism (Williams et al., 2000). Isobaric terranes can be
used to recognize styles of syn-tectonic and post-tectonic metamorphic reactions at different crustal levels and ultimately allow these effects to be identified in remote, geophysically derived sections (e.g., Fischer, 2002). Another consideration is the fabric anisotropy that is characteristic of each of the examples presented. Even in seismic studies that are focused on the mantle, the effects of crustal anisotropy and the extreme compositional heterogeneity are significant and must be considered (Mahan, 2006).

After decades of work, it is surprising how difficult it has been to recognize original microplate boundaries and/or sutures in each of these isobaric terranes. Each terrane is likely to include boundaries and paleosutures, and several have been proposed in the literature (Ilg et al., 1996; Tyson et al., 2002; Magnani et al., 2004; Duebendorfer et al., 2006). The lack of ophiolites, blueschists, and accretionary prisms has long been used to argue against plate tectonics (Stern, 2005; Ernst, 2006), but the counter to that argument is the recognition that the preservation potential of these rocks is very low at all levels in orogens. The isobaric sections discussed here reinforce this observation, but also make the point that such rocks may not be easily recognized at deep crustal levels. For example, the Grand Canyon transect (Fig. 6) is interpreted as a deep mid-crustal duplex in part because of the interleaving, at the 20–25 km levels, of pillow lavas and turbidites from the surface with ultramafic rocks of possible deep-crustal arc cumulate origin. This necessitates that these rocks have been translated >10’s of km relative to each other before becoming stabilized at 20–25 km depths prior to final exhumation. This is another issue that would not be clear from remote images or xenolith data alone.

Implications for Understanding Stabilization of Continental Lithosphere

Each of the three isobaric terranes preserves a view of the transition from a less stable or less mature crust to a more stable or more mature one, but the transition is dramatically different in each case. The deep-crustal level records the progressive dehydration, local melt loss, and subsequent isobaric cooling and extensive garnet growth that reflects a temporal transition from a relatively weak and melt-laden crust to a strong, anhydrous deep crust (see also Klepeis et al., 2003, 2004). The mid-crustal examples document the establishment of a dynamic block-type architecture that fundamentally controls the crustal rheology. The overall strength may be high despite local zones of high strain and magma flow (Karlstrom and Williams, 1998). Igneous rocks play a number of critical roles at each level, contributing to the progressive evolution of lithologic and rheologic properties through time. They introduce significant heat and serve to localize deformation in a feedback relationship at each crustal level.

In terms of crustal evolution and stabilization, the deep flux of basaltic magma from mantle depths to the base of the crust has the tectonic effects of destabilizing the lithosphere (e.g., Flowers et al., 2006a), enhancing deformation and crustal flow through melt-weakening at appropriate levels, and instigating upward flux of heat and melt that variably affect all higher levels. The longer-term effect involves depleting (and de-densifying) the upper mantle in terms of its Fe-component and causing thickening (and stabilization) of the lower crust (cf. Dufek and Bergantz, 2005). During isobaric cooling of deep-crustal rocks, there is a general tendency for rocks to grow garnet at the expense of pyroxene and plagioclase (Williams et al., 2000). This occurs because the rocks progressively cool into the field of high-pressure granulites (Fig. 1A). Although there may be kinetic barriers and small amounts of deformation and/or heating may be required, the general tendency is toward increasing density and probably strength (Williams et al., 2000). There is little evidence that surface-derived fluids reach this level to a significant degree. The lack of water may at least partly explain the general lack of coarsening and annealing of many rock types. The post-metamorphic evolution of the mid crust at both levels (0.7 GPa, Grand Canyon and 0.4 GPa, central Arizona) seems to involve local rehydration and annealing. Water may reach these levels from the surface but additional water may come from cooling plutonic rocks. This would be associated with a general decrease in density and strength and a general fertility for further deformation and or metamorphism in the future. It is interesting to note that both deep- and mid-crustal regions preserve evidence for major reactivation at a much later time, 1.9 Ga in Canada and 1.4 Ga in the southwestern United States. Both of these events are associated with new igneous activity. It seems likely that these igneous rocks provide not only heat but also new fertility for deformation and metamorphism especially at deeper levels, but the deformation was entirely different, as it was marked by development of steep cleavage rather than lateral flow. These observations emphasize that fundamental modifications may occur in a crustal terrane after the predominant tectonic event. This is an important consideration for interpretation of seismic images of ancient orogens.

SUMMARY

Isobaric terranes provide laboratories within which to evaluate the nature of continental crust and especially the nature and interplay of tectonic process through time at specific crustal levels. They provide a set of observable features against which geophysical models can be compared, and possible source rocks for igneous rocks and xenoliths. Clearly, each isobaric terrane is special and to some degree unique. We would not suggest that they can be simply stacked to make a model crust. Our hope is that these regions and interpretations can motivate studies of, and comparisons with, other isobaric terranes in order to better illuminate the common characteristics and processes that underlie the behavior of continental crust.

Numerous insights may emerge from integrating observations of isobaric terranes with crustal seismic cross sections and with data from other geophysical surveys, igneous petrology and xenolith studies. The most provocative commonalities of the 10, 20, and 35 km crustal levels discussed here are as follows.
(1) All crustal levels in continents can have quartz-rich lithologies (including metasedimentary rocks) such that rheological models for compositionally zoned mafic lower crust to felsic middle and upper crust are oversimplified, as are rheological models based on such zonation. (2) Foliation patterns of steep fabric domains segmenting earlier shallow fabric domains are common at all crustal levels; this heterogeneity of finite strain fabric may be as important as heterogeneity of composition in controlling crustal rheology, as shown in the Grand Canyon example where the rheology of hot domains would be consistent with crustal flow, but the intervening colder domains and shear zone architecture may have inhibited flow. (3) Advection heat transfer via melt flux is common to all crustal levels as are the resulting lateral temperature gradients due to melt partitioning, but the expression of melt-enhanced tectonism varies: basalt underplating and temperature gradients due to melt partitioning, but the expres-


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