Collaborative research: Testing channel-flow models using mid-crustal rocks of North Himalayan gneiss domes

1. Introduction

A wide variety of geophysical and geological facets of the Himalayan orogen have recently been postulated to be the result of channel-flow/extrusion. Models of this process postulate that a combination of gravitational potential-energy gradients; low-viscosity, ductile middle crust bounded above and below by coeval normal- and thrust-sense shear zones; and focused, rapid erosion along the southern flank of the Himalaya results in south-directed channel flow and extrusion of the middle crust (Fig. 1) (e.g. Beaumont et al., 2001, 2004; Hodges et al., 2001; Grujic et al., 2002). These models have been derived principally from geophysical data from southern Tibet and geologic data from the Himalayan front. Absent are any geological data from north of the Himalaya, closer to the presumed fount of the flowing crust, and yet such data on the style, vorticity, strain, spatial distribution, and timing of flow are essential for testing the channel flow idea. The North Himalayan gneiss domes, southern Tibet are an ideal location for such investigations because they provide excellent exposure ~50–100 km north of the crest of the Himalaya (Fig. 2) of an originally ≥15–25 km-thick sequence of mid-crustal rocks that preserve mid-crustal fabrics that predate doming and for which PTt data are well known.

We request support for a three-year project to test the channel-flow hypothesis by: Determining whether the predicted low-viscosity channel is exposed in the North Himalayan gneiss domes and whether it shows the expected combination of southward flow and vertical thinning.

Fig. 1. Schematic diagram illustrating a flowing, hot, low viscosity middle crustal channel (gray region). Poiseuille flow dominates within the channel and Couette flow beneath the channel (see Fig. 3). Predicted locations of middle-crustal rocks exposed in the core of Mabja Dome and of the Greater Himalayan sequence exposed in the Everest region prior to exhumation are shown. Modified from Beaumont et al. (2004) and Godin et al. (2006).

Fig. 2. Regional tectonic map of the central Himalayan orogen showing location of the North Himalayan gneiss domes (italic font). ITSZ, Indus–Tsangpo suture zone; MBT, Main Boundary Thrust; MCT, Main Central Thrust; STDs, Southern Tibetan Detachment System. Modified from Burchfiel et al. (1992).

Flow within a channel can range from pure Couette flow to Poiseuille flow, or be a combination of the two (Fig. 3). Couette (or linear) flow develops between rigid plates moving relative to one another and is characterized by simple shear (high vorticity number) with the highest velocities toward the top or bottom of the channel (e.g., White, 1974). Poiseuille (or parabolic) flow develops between stationary rigid plates in which a horizontal gradient in lithostatic pressure produces the highest velocities in the center of the channel and decreasing, but opposite shear velocities toward the top and bottom of the channel. Poiseuille flow is characterized by high vorticity number (simple shear) at the top and bottom of the channel, decreasing vorticity number (mix of simple shear and pure shear or general shear) toward the center of the channel, and low vorticity number (pure shear) at the center of the channel (cf. Figs. 3 & 4).
Models that have been developed for the channel-flow paradigm for Tibet are based on general laminar flow—a combination of Couette and Poiseuille flow—in which the two types of flow can act in concert or opposition to one another (Figs. 1 & 3). While there are differences among the models that have been proposed, all predict a downward change from top-N simple shear (high vorticity number) at the top of the channel, to increasing general shear (i.e. mixed pure and simple shear) toward the center (increasing pure shear and low vorticity number), to decreasing general shear characterized by top-S simple shear toward the base of the channel (Figs. 1, 3, & 4). Thinning or thickening of the channel is implicit in general and pure-shear flow (Fig. 4). Our investigations will center on documenting the spatial patterns of deformation, vorticity of flow, deformation temperatures, and timing of flow. Integrating these results with published thermobarometric, geochronologic, and thermochronologic data (Lee et al., 2000, 2004, 2006; Lee & Whitehouse, 2007) will enable us to establish the thermal and temporal patterns of deformational flow and test whether they match the predictions of the channel-flow hypothesis.

2. Scientific Rationale

The Himalayan orogen records Eocene to Recent continental collision and convergence between India and Asia. Profound crustal shortening and thickening resulted in the formation of one of the most impressive orogenic belts Earth has enjoyed: the Himalaya, with a length of ~2500 km and 14 peaks over 8000 m in elevation, and the Tibetan plateau, the areally most extensive (>5x10^6 km²) and highest plateau, with an average elevation of ~5000 m (Fielding et al., 1994).

Extensive geologic and geophysical research over the last 15–20 years has focused on characterizing: (a) the development and outward growth of the Tibetan Plateau; (b) the development of partial melts interpreted to reside locally in the middle crust of Tibet; (c) the development of structures along the southern margin of the plateau, particularly the broadly coeval normal-slip Southern Tibetan Detachment System (STDS) and the Main Central Thrust (MCT) that bound the Greater Himalayan sequence (GHS) high-grade metamorphic rocks (Fig. 2); (d) southward extrusion of the GHS; and (e) focused erosion along the southern margin of the plateau. In aggregate, these studies led to the channel-flow hypothesis. For example, Grujic et al. (1996) used quartz microfabrics from Bhutan to demonstrate general non-coaxial flow of the GHS, and postulated that the GHS deformed as a wedge between the MCT and STDS; later they (Grujic et al., 2002) reformulated this wedge model by postulating that the GHS deformed as a
10–15 km thick channel that extends >200 km northward beneath Tibet. Vannay and others (e.g. Vannay & Grasemann, 1998; Grasemann et al., 1999, Vannay & Grasemann, 2001) used a combination of spatially varying vorticity numbers and deformation temperatures, PT conditions associated with inverted isograds, and mica Ar/Ar cooling ages from the base of the GHS in the Sutlej Valley to come to similar conclusions. These ideas have since been incorporated in thermal–mechanical models. For instance, Beaumont et al. (2001, 2004, 2006) developed a set of transient, plane-strain, finite-element models in which the GHS represents a 15–30 km thick, hot, low-viscosity middle-crust channel that extrudes southward from beneath southern Tibet toward the orogenic front during N–S convergence. Flow begins after the crust has been tectonically thickened and the middle crust experiences a reduction in viscosity due to mantle heat flux and crustal radiogenic heating. Flow and extrusion of the low-viscosity tabular body of middle crust is driven by a horizontal gravitational potential-energy gradient produced by the topographic and crustal thickness differences between the Tibetan Plateau and its margins and focused erosion along the southern flank of the high Himalaya (e.g. Beaumont et al., 2001, 2004; Hodges et al., 2001). The low-viscosity channel is bounded above and below by normal-sense (STDS) and thrust-sense (MCT) shear zones, respectively, that separate the channel from higher viscosity material above and below (Beaumont et al., 2001, 2004). (Fig. 1)

While there are reasons to discount aspects of the channel flow model (e.g., Harrison, 2006), the model has become a paradigm that focuses much current research; for this reason it deserves testing.

3. Data Constraints

A wealth of geophysical and geologic data from southern Tibet and the high Himalaya (Fig. 2) form the foundation for these channel-flow/extrusion models. Geophysical observations from heat-flow, seismic, and magnetotelluric investigations indicate that the middle crust in southern Tibet includes partial melt or aequous fluids, is hot and therefore weak compared to a stronger upper crust and lower crust/upper mantle, and is a zone of active or recently active ductile deformation characterized by vertical thinning and horizontal extensional flow (e.g. recent review by Klemperer, 2006 and references therein).

That the GHS consists of strongly deformed, moderate-temperature/pressure (peak conditions of 600–700 MPa and ~600°C) paragneisses, orthogneisses, migmatites, and variably deformed leucogranites (e.g. Hodges et al. 1988; Hubbard 1989; Murphy & Harrison 1999; Walker et al. 1999; Stephenson et al. 2000; and many others) has been used to suggest that the GHS is an exposure of early Miocene middle crust similar to the middle crust of the Tibetan Plateau today. The GHS preserves contractional structures overprinted by mylonitic fabrics and is bounded by two major, N-dipping high-strain shear zones, the STDS normal fault at the top and the MCT thrust fault at the base. Numerous geochronologic and thermochronologic studies constrain the timing of structural, metamorphic, and intrusive events recorded in these rocks. Contraction-related burial and peak Barrovian metamorphism occurred between 37 Ma and 28 Ma (Vance & Harris 1999; Walker et al. 1999; Simpson et al. 2000). Multiple generations of deformed and undeformed leucogranites were emplaced from 31.6 to 12.5 Ma (Noble & Searle 1995; Hodges et al. 1996, 1998; Edwards & Harrison 1997; Searle et al. 1997; Wu et al. 1998; Harrison et al. 1999; Murphy & Harrison 1999; Searle et al. 1999, a; b; Walker et al. 1999; Simpson et al. 2000). Cooling of paragneisses to below ~400°C between 22–20 Ma and 17–15 Ma, signifies that mylonitization ceased at this time or soon thereafter (Searle & Rex 1989; Hodges et al. 1992; Searle et al. 1992; Walker et al. 1999; Stephenson et al. 2001). The combination of field, structural, and geochronologic observations indicate that movement along the STDS and MCT shear zones was broadly simultaneous at ~22–13 Ma (see review in Godin et al., 2006 and references therein).

Because the GHS is bounded above and below by a normal fault and thrust fault, respectively, and the STDS is a stretching fault (Law et al, 2004) across which there was no net extension, the GHS must have been exhumed and extruded to the south (e.g. Burchfiel & Royden, 1985; Grujic et al., 1996; Grasemann et al., 1999; Beaumont et al., 2001, 2004; Jessup et al., 2006) (Fig. 1). Tests of this hypothesis have centered on determining the kinematics and vorticity of flow at i) the top of the GHS north of Everest (Law et al., 2004), ii) the STDS and MCT in the Everest area (Jessup et al. 2006), and iii) the base of the GHS in the Sutlej Valley of India (Grasemann et al., 1999; Vannay & Grasemann, 2001). Vorticity
studies by Law et al. (2004) and Jessup et al. (2006) showed that high-grade, structurally deep rocks at the top of the GHS record general shear (combination of pure and simple shear; 48–41% pure shear) at close to peak metamorphic conditions whereas lower-grade, structurally higher rocks record sub-simple shear (38–36% pure shear). At the base of the GHS, just above the MCT, the rocks record the highest pure-shear component (53–48% pure shear) that postdates peak metamorphism. Grasemann et al. (1999) showed a similar pure-shear dominated low-temperature deformation for the base of the section that overprinted an early high-temperature deformation with vorticity values close to simple shear. Jessup et al. (2006) concluded that flow was spatially and temporally partitioned—high-temperature samples recorded the early stages of channel-flow/extrusion at mid-crustal depths, whereas structurally higher, lower temperature samples recorded sub-simple shear at the upper margin of the channel during later stages of exhumation, and the pure-shear component just above the MCT was postulated to be the result of higher lithostatic pressures. These studies along the Himalayan front demonstrate the feasibility of collecting similar data elsewhere within the Tibetan Plateau.

4. Geology of the North Himalayan Gneiss Domes

The North Himalayan gneiss domes expose mid-crustal rocks about 50–100 km north of the high Himalaya (Fig. 2). Geologic studies have largely focused on four domes, Kangmar, Kampa, Mabja, and Malashan (e.g. Burg et al., 1984; Chen et al., 1990; Lee et al., 2000, 2002, 2004, 2006; Lee & Whitehouse, 2007; Zhang et al., 2004; Aoya et al., 2005, 2006; Quigley et al., 2006, 2008; Kawakami et al., 2007). The domes consist of a core of orthogneisses, migmatites, leucogranites, and high-grade metasedimentary rocks overlain by a progressively lower grade sedimentary cover (e.g. Burg et al., 1984; Chen et al., 1990; Lee et al., 2000, 2004, 2006; Aoya et al., 2005, 2006; Quigley et al., 2006). The domes record evidence for two major deformational events, followed by doming. D1 was characterized by N–S shortening and vertical thickening, and D2 by vertical thinning and N–S extension (Chen et al., 1990; Lee et al., 2000, 2004). In Kangmar and Mabja, progressive rotation of D1 structures into parallelism with the D2 stretching lineation and a subvertically foreshortened metamorphic pressure gradient (see below) suggest a D2 finite strain of \( \geq 16:1 \) (X:Z) (Lee et al., 2000, 2004). Reconnaissance kinematic analyses of D2 structures suggest that Mabja, and possibly Kangmar, record a change in shear sense with depth, from top-N at intermediate structural depths to top-S at the deeper structural levels (Chen et al., 1990; Lee et al., 2000, 2004), whereas Malashan, which exposes intermediate structural levels, records only top-N shear (Aoya et al., 2005). Barrovian metamorphism accompanying the D2 deformation occurred under \( P-T \) conditions that increase downward from \(-450^\circ\text{C}/370\text{ MPa}\) to \(700^\circ\text{C}/800\text{ MPa}\) (Lee et al., 2000, 2004). U/Pb zircon geochronology and \(^{40}\text{Ar}/^{39}\text{Ar} \) and apatite fission-track thermochronology indicate that D2 vertical thinning and horizontal extension, migmatization, and peak metamorphism began at \( \sim 35\text{ Ma} \), were ongoing \( 23–18\text{ Ma} \), and had ceased by \( \sim 16\text{ Ma} \), a period of \(12–19\) m.y. (Fig. 5) (Lee et al., 2000, 2006; Lee & Whitehouse, 2007; Aoya et al., 2005; Zhang et al., 2004; Quigley et al., 2006, 2008).
The range of D2 structural levels exposed in the domes is the fortuitous outcome of late-stage doming. In Kangmar and Mabja, metamorphic isograds, the S2 foliation, and mica Ar/Ar chrontours are domed, but low-temperature K-feldspar Ar/Ar and apatite fission-track chrontours are not, requiring that doming occurred between \( \sim 400^\circ \text{C} \) and \( \sim 200^\circ \text{C} \) (Lee et al., 2000, 2004, 2006). The cause of this doming not certain (but see Lee et al., 2000, 2004, 2006), but in any case, produced low strains at low temperatures after the main, D2, high-strain deformation and Barrovian metamorphism.

The structural, metamorphic, anatectic, intrusive, and geochronologic histories of the North Himalayan gneiss domes are similar to those recorded in the GHS, suggesting that during the late Oligocene to middle Miocene, high-grade mid-crustal metasedimentary and metaigneous rocks, cut by anatectic melts and leucogranites, may have extended from beneath the high Himalaya northward beneath southern Tibet. The broadly coeval middle Miocene cessation of ductile deformation within the GHS and the gneiss domes (c.f. Lee et al., 2000, 2006; Lee & Whitehouse, 2007; Searle & Rex, 1989; Hodges et al., 1992; Walker et al., 1999) provides further evidence that these discontinuous exposures of mid-crustal rocks were once linked. Because the GHS has been interpreted as the leading edge of an eroding and southward-extruding tabular channel of ductile mid-crustal rocks bounded by the STDS and MCT (e.g. Grujic et al., 1996, 2002; Nelson et al. 1996; Searle 1999a, b; Beaumont et al. 2001, 2004; Hodges et al. 2001; Vannay & Grasemann, 2001; Searle et al. 2003) and the gneiss domes record a history of vertical thinning and N–S horizontal stretching, strain compatibility implies that the hot and weak mid-crustal rocks now exposed in the North Himalayan gneiss domes represent the ductile interior of such a mid-crustal channel or block (e.g. Lee et al., 2006). Based on these interpretations and the change in shear sense recorded within the gneiss domes, we hypothesize that the gneiss domes expose a nearly complete section of a fossilized, low-viscosity mid-crustal channel or extruding ductile block that records southward flow (Fig. 1) and vertical thinning.

5. Proposed Research
5.1 Introduction

One of the most important tests of the channel-flow/extrusion models is to determine whether the channel flow/extrusion documented in the mid-crustal rocks of the GHS (e.g. Grujic et al., 1996, 2002; Grasemann et al., 1999; Law et al., 2004, Jessup et al., 2006) can be traced northward into mid-crustal rocks of southern Tibet. If we can place new constraints on the spatial–thermal–temporal distribution of finite strain, kinematics, and vorticity of ductile deformation and flow in middle crust rocks exposed in the core of the North Himalayan gneiss domes, and integrate those with existing thermobarometric, geochronologic, and thermochronologic data (Fig. 5) (Lee et al., 2000, 2006; Lee & Whitehouse, 2007; Aoya et al., 2005), we can test the channel-flow/extrusion model predictions for the style, vorticity, and
spatial distribution of flow in mid-crustal rocks and the proposed link between mid-crustal channel flow and denudation-driven extrusion (e.g. Grujic et al., 1996, 2002; Grasemann et al., 1999; Vannay & Grasemann, 2001; Beaumont et al., 2004) (Fig. 1). Because these hypotheses have had such an important impact on guiding recent thinking about collisional orogenic belts—and the Himalaya–Tibet orogenic belt is the most-studied active example of this—an accurate characterization of the deformation and flow histories of the North Himalayan gneiss domes, coupled with similar constraints from the GHS, should have major implications for understanding geodynamic processes globally (for example mid-crustal channel flow has been proposed for the Canadian Cordillera (e.g. Brown & Gibson, 2006; Williams & Jiang, 2005); the Appalachian Piedmont (Merschat & Hatcher, 2005); and the Archean of Greenland (Grocott et al., 2004)).

To address these questions, we propose a targeted, detailed investigation of the structural, finite-strain, kinematic, and vorticity histories of deformation using field structural studies, laboratory structural studies, and geochronology. We will focus on the ~10 km (present-day structural thickness) exposure of vertically thinned (by at least 50%) and horizontally stretched migmatites, orthogneisses, and high-grade metasedimentary rocks within the two best-studied domes, Mabja and Kangmar, with which we are quite familiar (e.g. Burg et al., 1984; Chen et al., 1990; Lee et al., 2000, 2002, 2004, 2006; Lee & Whitehouse, 2007) (Fig. 2). Additional advantages of working in these domes include excellent exposure, a wide range of lithologies and vorticity gauges (see below), well-preserved penetrative D2 fabrics that record ductile deformation at mid-crustal depths (e.g. Lee et al., 2000, 2004, 2006), and pre-existing detailed thermobarometric, geochronologic, and thermochronologic data (e.g. Fig. 5). By tying deformational fabrics to particular parts of a PTt path (Fig. 5) (e.g. through temperatures indicated by microstructures used to calculate the flow vorticity; see below), we can reconstruct the spatial—thermal—temporal flow patterns within these mid-crustal rocks (Fig. 1). For example, we will be able to discriminate between deformation histories such as 1) dominant simple-shear at high temperature followed by pure-shear at low temperature (e.g., Grasemann et al., 1999), or 2) dominant pure shear at high temperature followed by simple shear at low temperature when the rocks are stronger (cf. Lister and Williams, 1983).

Our strategy of incorporating our proposed studies with published and ongoing geologic investigations of the GHS in the Everest region by Drs. Rick Law and Mike Searle, and graduate students (see letter of collaboration) will provide us with an outstanding opportunity to document a spatial–thermal–temporal view of mid-crustal deformation along a transect parallel to the transport direction from the high Himalaya northward into southern Tibet.

5.2 Kinematic and Vorticity Analyses

To determine the style of flow recorded in the strongly deformed mid-crustal rocks of Mabja and Kangmar domes, we will undertake kinematic and vorticity analyses to document the kinematics and relative proportions of pure and simple shear during ductile deformation. Kinematic studies will apply both standard mesoscopic and microscopic analyses (e.g. Passchier & Trouw, 2005; see also below) to document the structural depth of, and temporal relationships among observed changes. Documenting the style (coaxial vs. non-coaxial deformation, and if the latter, the direction of non-coaxial flow), spatial distribution, and temporal relations of deformation is critical for placing these mid-crustal rocks into a spatial context for evaluating the channel-flow/extrusion models (Fig. 1).

Vorticity analysis yields quantitative information on the relative contributions of pure and simple shear, and knowing those percentages is important because a large pure-shear component indicates significant vertical thinning and horizontal extension and, relative to simple-shear, an increase in strain and extrusion rates (e.g. Pfiffner & Ramsay, 1982; Ramsay & Huber, 1987). Kinematic vorticity number, \( W_k \), measures the relative contributions of pure \( W_k = 0 \) and simple \( W_k = 1 \) shear during steady-state (instantaneous) deformation; pure- and simple-shear components of deformation are equal when \( W_k = 0.71 \) (Fig. 2) (Means et al., 1980; Law et al., 2004). However, the vorticity of flow varies both spatially and temporally in naturally deformed rocks (e.g. Fossen & Tikoff, 1998; Jiang, 1998). In such cases of non-steady state deformation, flow vorticity is better characterized by mean kinematic vorticity number, \( W_m \). The best approach to calculating \( W_m \) is to use several different vorticity gauges (e.g. rigid grains,
porphyroblast hyperbolic distribution, and oblique grain shape; see below) that re-equilibrate at different rates during the deformation history (Law et al., 2004; Passchier & Trouw, 2005). For example, oblique quartz grain shape might record only the last increment of the flow history, whereas rigid grain rotation might record a time-averaged flow history. Vorticity of flow may also be partitioned among lithologies with different rheologies (e.g. Lister & Williams, 1983). The best approach to assessing this is to collect samples from the range of lithologies exposed in a given location. Furthermore, using a variety of vorticity gauges within the same unit and over a range of lithologies, and combining those data with finite-strain, petrologic and geochronologic data (see below) will allow us to document the locus, migration, and spatial and temporal partitioning of deformation. Law et al. (2004) and Jessup et al. (2006) showed that a similar approach—without finite strain and geochronology —worked in the GHS exposed in the Everest region (see above).

The vorticity-analysis methods work by measuring planes parallel to the stretching lineation and perpendicular to foliation, and by assuming that the vorticity vector is perpendicular to the maximum and minimum principal axes of finite strain and that the deformation fabric is characterized by monoclinic symmetry; it does not assume plane strain. Nevertheless, approximate plane strain is suggested by well-developed L–S tectonites and cross-girdle fabrics recorded in quartz c-axis fabric patterns from Kangmar studied in reconnaissance (Lee et al., 2000). Vorticity-analysis techniques appropriate for use in the North Himalayan gneiss domes include:

**Rigid-grain technique.** The rigid-grain technique for measuring vorticity, developed by Wallis et al. (1993), entails measuring rigid porphyroclast aspect ratio (R) and the angle (θ) between clast long axes and matrix foliation in thin sections (Fig. 6a). For a given \( W_m < 1 \), clasts with an aspect ratio above a critical value will rotate into a stable orientation. The critical threshold \( R_c \), the transition between clasts that rotate infinitely and those that reach a stable orientation, is defined and related to \( W_m \) (Passchier, 1987) from the relationship between aspect ratio (R) and angle from foliation (θ) by the equation: \( W_m = (R_c^2 - 1)/(R_c^2 + 1) \). The rigid grain net (RGN), developed by Jessup et al. (2007), allows an estimate of \( W_m \) directly from a graph (Figs. 6b, c). Plotted on a RGN is the angle θ between clast long axis and matrix foliation vs. shape factor \( B^* \) (where \( B^* = (M_2^2 - M_0^2)/(M_2^2 + M_0^2) \) and \( M_2^2 = \) long axis; \( M_0^2 = \) short axis of clast). The rigid grain technique requires that (a) the porphyroclasts predate the dominant deformation fabric; (b) the porphyroclasts are internally undeformed; and (c) there is no mechanical interaction between adjacent clasts or the clasts and the matrix. Depending on the deformation temperature, rigid clasts that can be evaluated with this method include amphibole, apatite, chloritoid, epidote, feldspar, Fe-oxides, garnet, tourmaline, and U-bearing phases. Reconnaissance vorticity studies on two samples from Mabja Dome, one at intermediate and the other at deep structural levels, show a decrease in vorticity number with depth, indicating a downward increase in pure shear (thinning) component from 38–40% to 58–63% (Fig. 6b, c).

**Porphyroblast hyperbolic distribution (PHD) technique.** The PHD technique (Simpson & dePaor, 1993) entails recording, on a hyperbolic net, rigid porphyroclast aspect ratio (R) and the angle (θ) between clast long axis and matrix foliation, whether the clast is forward or back rotated, and the type of strain tail on the clast (σ or δ) (e.g. Passchier & Trouw, 2005). A hyperbola is defined that encloses all back-rotated σ clasts and separates them from all other clasts. The mean kinematic vorticity number, \( W_m \), is calculated from the cosine of the angle between the two limbs of the hyperbola.

**Oblique quartz grain-shape foliation.** During the last stages of deformation, newly recrystallized grains are stretched parallel to the extending instantaneous stretching axes (ISA). Measuring the maximum angle between an oblique grain shape fabric and the flow plane provides an approximate orientation of the ISA (Wallis, 1995). The orientation of the ISA is related to \( W_m \) by the equation: \( W_m = \sin 2\theta \) where θ is the acute angle between the flow apophyses (lines of zero instantaneous rotation) and the ISA (Wallis, 1995). The flow plane is defined as the perpendicular to the central quartz c-axis girdle (Fig. 7a) (Law, 1990; Wallis, 1995), which will be defined using electron-backscatter diffraction (see section 5.3 below). Strongly deformed quartz at moderate metamorphic temperatures commonly exhibits an
oblique grain-shape foliation, making it an ideal mineral for this technique (e.g., Law, 1998; Simpson, 1998).

5.3 Quartz Crystal Preferred Orientation, Microstructures, and Deformation Temperatures

We can gain important, otherwise unavailable, information about quartz crystal preferred orientations (CPO), shape preferred orientations (SPO), and deformation mechanisms using electron-backscatter diffraction (EBSD). The results from EBSD provide a wealth of information including sense of shear, orientation of flow plane, deformation temperatures associated with shear sense and vorticity number fabrics, and relative timing of events. These data are critical to elucidate the kinematics, flow vorticity, thermal, and temporal patterns of ductile deformation recorded in mid-crustal rocks. With EBSD we measure the complete crystallographic orientation of micron-scale spots; in this study, we will do this at two scales: i) the scale of the entire thin section at a 100 µm spacing, to measure the aggregate properties; and ii) over specific areas (depending on grain size, but e.g. 200 µm x 200 µm areas at a 1 µm spacing), to measure individual grain and grain–grain properties. From these data we will derive CPOs (Fig. 7) and SPOs which can be used to infer strain state and aid in slip-system determination.

In quartz, the flow type and finite strain result in distinctive CPOs that can be used to interpret the kinematics of flow. For a single slip system, pure flattening causes the slip plane normal to align parallel to the Z finite strain axis and the slip directions to be evenly distributed in the XY plane; pure constriction...
causes the slip direction to align parallel to the X axis and the pole to the slip plane to be evenly distributed in the YZ plane; plane strain leads to a unique orientation for both: the slip direction parallel to X and the slip plane normal parallel to Z. In real rocks, where the von Mises criterion ensures that multiple slip systems are active, quartz CPOs display a richer spectrum of patterns. The general upper-temperature progression in quartz slip system activation from basal-\(<a>\) to prism-\(<a>\) to rhomb-\(<a>\) to, in exceptional cases, prism-[c], means that the CPOs characteristic of strain state vary with temperature. For example, at low temperatures, where basal-\(<a>\) slip dominates, progressive pure-shear strain leads to the development of Type-I crossed girdles of [c] and symmetrically distributed, double \(<a>\) maxima in plane strain, a Type-II crossed or cleft girdle of [c] and a small circle of \(<a>\) around the X axis in constrictional strain, and a small-circle girdle of [c] and a small circle of \(<a>\) around the Z axis in flattening strain (Lister, 1977; Lister & Hobbs, 1980) (Fig. 8a). Progressive simple shear leads to the development of asymmetric monoclinic fabrics that change with increasing metamorphic grade (e.g. Carreras et al., 1977; Lister & Hobbs, 1980; Schmid & Casey, 1986; Law et al., 1986; Law & Potts, 1987; Dell’Angelo & Tullis, 1989; Kurz et al. 2001) (Fig. 8b). At low to medium metamorphic temperatures, asymmetric Type-I crossed and single girdle [c] and asymmetric single \(<a>\) maxima are inclined with respect to the foliation and lineation. At medium to high metamorphic temperatures, asymmetric [c] maxima develop parallel to Y and \(<a>\) are normal to it (e.g., Blumenfeld et al., 1986; McGrew & Casey, 1998).

Fig. 7. EBSD quartz CPO data from a low-temperature quartzite (a) and high-temperature incipient migmatite (b) from Mabja Dome. \(<0001>\), \(<1\overline{1}-20>\), and \(<10-10>\) are e-, a-, and m-axes, respectively, plotted on lower and upper hemisphere projections. Asymmetry in e- and a-axes in (a) indicates top-N shear and an opening angle of 53-58° in the e-axis fabric skeleton indicates a deformation temperature of ~370-510°C (cf. Figs. 8 & 9). Asymmetry in c-axes in (b) indicates top-S shear and development of myrmekite and incipient quartz chessboard extinction indicate deformation temperatures of ~700°C. Samples cut perpendicular to foliation and parallel to lineation.

Deformation temperatures will be compared to metamorphic temperatures and cooling histories (Fig. 5) (Lee et al., 2000, 2006; Lee & Whitehouse, 2007) to provide timing constraints on the kinematics and vorticity of flow. For example, deformation temperatures above 450–500°C would support the conclusion (Lee et al., 2006), based on a combination of structural, metamorphic petrology, geochronology, and thermochronology results, that the domes preserve mid-crustal fabrics formed within a channel prior to doming. Two methods will be used to estimate deformation temperatures: a temperature range during deformation can be extracted from quartz textures and deformation mechanisms/slip systems, and an estimate of temperature to within ±50°C can be determined from quartz c-axis CPOs. At relatively low
temperatures (200–400°C), quartz grains show irregular and patchy undulose extinction, sutured and serrated grain boundaries, inhomogeneous flattening, and deformation lamellae, produced by dislocation glide primarily on the basal-<a> slip system (Hirth et al., 2001; Stipp et al., 2002) (Fig. 8b); strain-induced grain-boundary bulging recrystallization is the dominant recrystallization mechanism, and strain is accommodated principally within the recrystallized grains (Regime 1 of Hirth & Tullis, 1992). At moderate temperatures (400–500°C), quartz textures include strongly flattened porphyroclasts, sweeping undulose extinction, core-and-mantle structures, ribbon grains, and subgrains and recrystallized grains of equivalent size, produced by dislocation creep dominated by the prism-<a> system (Hirth et al., 2001; Lloyd & Freeman, 1994; Stipp et al., 2002) (Fig. 8b); dislocation climb is sufficiently rapid to accommodate recovery and recrystallization is dominated by subgrain rotation (Regime 2 of Hirth & Tullis, 1992). At higher temperatures (500–700°C), the increased rate of grain-boundary migration means that recrystallization occurs by both grain-boundary migration and subgrain-rotation recrystallization (Regime 3 of Hirth & Tullis, 1992). At the lower temperature end of this range, grain boundaries are typically lobate, and pinning or migration microstructures are observed (Hirth et al., 2001; Jessell, 1987; Stipp et al., 2002). At the higher temperature end of this range, apparently strain-free grains and island grains are common, grain boundaries are lobate or amoeboid, and chessboard extinction or subgrains may be present (e.g. Stipp et al., 2002). Above 700°C, prism-[c] slip becomes important (Mainprice et al., 1986) (Fig. 8b).

Experimental (e.g. Tullis et al. 1973) and numerical simulation studies (e.g. Lister et al., 1978; Lister & Hobbs, 1980; Lister & Dornsiepen, 1982; Wenk et al., 1989) indicate that the opening angle of quartz c-axis fabrics increases with increasing deformation temperature and hydrolytic weakening, and decreasing strain rate. The opening angle is defined as the angle between the girdles measured in the plane perpendicular to foliation and parallel to lineation (Kruhl, 1998). There is a linear relation between this angle and deformation temperatures between ~250–650°C (Fig. 9) (Law et al., 2004). Crystal-plastic deformation of other minerals can also be used as an approximate temperature gauge. For example, orthoclase and plagioclase grains typically do not exhibit evidence for crystal-plastic deformation until temperatures exceed 450–500°C (Tullis & Yund 1992; FitzGerald & Stünitz 1993).

Preliminary thin section petrography of quartz and feldspar textures on two samples from Mabja Dome, one at intermediate and the other at deep structural levels, show an increase in deformation temperature with depth. Quartz in chloritoid-zone rocks exhibits textures compatible with Regime 2 recrystallization with subgrain rotation combined with weak grain-boundary migration. Preliminary EBSD analysis of one chloritoid-zone sample yielded asymmetric c-, a-, and m-axes with respect to the foliation (cf. Figs. 7a and 8b) indicating a top-N sense of shear. The cross-girdle c-axis CPO implies that basal <a>, rhomb <a>, and prism <a> slip systems were active (cf. Figs. 7a and 8b) and yields an opening angle of 53–58°. All these observations suggest deformation temperatures of ~400–500°C (cf. Figs. 7a, 8b, and 9). Quartz in sillimanite-zone rocks has textures consistent with Regime 3 recrystallization, including weakly developed chessboard extinction and island grains. One sillimanite-zone sample yielded weakly asymmetric c- and a-axes with respect to the foliation (cf. Figs. 7b and 8b), suggesting top-S sense of shear. The c-axis CPO suggests that prism <a> slip was dominant and rhomb <a> subordinate, (cf. Figs. 7b and 8b). These observations suggest deformation temperatures of ~600–700°C.

5.4 Finite Strain

Rotated pre-D2 linear markers and vertically collapsed isogrids in Kangmar and Mabja indicate a finite strain of ~30:1 and ≥16:1 (X:Z), respectively, associated with D2 vertical thinning and horizontal stretching (Lee et al., 2000, 2004). However, because of the absence of classic finite-strain markers (e.g. pebbles, fossils, ooids), the spatial distribution of that strain is unknown, but critical for assessing the percentage of the finite strain recorded by a vorticity number within a particular sample (e.g. documenting a spatial and/or temporal distribution of pure vs. simple shear, will allow assessment of which deformation style accommodated most of the strain). Furthermore, finite-strain studies will allow us to test the specific prediction that higher finite strains are concentrated at the boundaries of the channel in
the thermal-mechanical model (Beaumont et al., 2006) and the kinematic and geometric model (Vannay & Grasemann, 2001).

To document the spatial distribution of finite strain, we will use the Fry Method (Fry, 1979) applied to rigid metamorphic porphyroblasts, such as chloritoïd, garnet, and tourmaline, and feldspar augen (e.g. Genier & Epard, 2007; Ramsay & Huber, 1983). The Fry Method is a graphical technique for measuring finite strain in rocks that contain rigid grains. The method entails plotting the position of each rigid grain center with respect to a rigid grain placed at the origin. The origin is then placed on each grain center and the position of every other grain center is plotted. The result is a distribution of points with a vacancy field that defines the magnitude and orientation of the finite strain ellipse. The Fry Method produces an accurate estimate of the orientation and shape of the finite strain ellipse if the sample has undergone homogeneous deformation and the distribution of centers is isotropic and anticlustered (Fry, 1979).

Genier & Epard (2007) show that homogeneity can be tested using the Morishita diagram and anisotropy by plotting a cumulative histogram of angles between positions of all respective rigid grain centers. The Morishita index of dispersion (MI) (Morishita, 1959) is measured by placing a grid over the sample (e.g. a photograph of the rigid grains in the XZ plane) and is calculated by:

\[
MI = Q \frac{\sum_{i=1}^{Q} n_i(n_i-1)}{N(N-1)}
\]

where N = number of points; \(n_i\) (i = 1, 2, ... Q) is the number of points in the \(i^{th}\) cell; and Q is the number of cells. A plot of MI vs. cell size yields distinct curves for homogeneous, random, and clustered distributions. For a homogeneous distribution, MI values decrease from 1 for large cells to 0 for small cells because as cell size decreases each contains either one rigid grain or none. For a random distribution, MI values vary around 1 because regardless of the cell size it is not possible to put one rigid grain in each

Fig. 8. (a) Flinn diagram showing geometry of CPO patterns of quartz c-axes (blue) and a-axes (yellow) with respect to strain in coaxial deformation. (b) Geometry of CPO patterns of quartz c-axes (blue) and a-axes (yellow), and active slip systems with increasing deformation temperature in non-coaxial deformation. Modified from Passchier & Trouw (2005).

Fig. 9. Quartz c-axis fabric opening angle scales with deformation temperature (Kruhl, 1998). Data (boxes) from naturally deformed rocks from Kruhl (1998) with additional data from Law et al. (1992) and Nyman et al. (1995). Best fit line (dashed) between opening angle and temperatures indicated for deformation temperatures of ~250-650°C; ±50°C error shown. Modified from Law et al. (2004).
Monazite Geochronology

Monazite has long been a target of U-Pb TIMS geochronology (Parrish, 1990; Hawkins & Bowring, 1997; Foster et al., 2002) because i) it contains U but little common Pb (Overstreet, 1967), ii) undergoes Pb loss only at high temperatures (Cherniak et al., 2004), iii) does not become metamict (Seydoux-Guillaume et al., 2002), and iv) is common in pelitic and granitic rocks. Monazite appears in pelites at the expense of allanite, apatite + REE oxides, and/or silicates at the garnet, staurolite (Smith & Barreiro, 1990; Kingsbury et al., 1993; Catlos et al., 2002a; Kohn & Malloy, 2004), or aluminumsilicate (Wing et al., 2003) isograds and can persist to temperatures in excess of 1000°C (Hacker et al., 2000); in metaluminous orthogneisses it forms at the expense of allanite + apatite at >600°C (Bingen et al., 1996). This large P–T–X stability range, combined with extreme resistance to Pb loss (Cherniak et al., 2004), partial recrystallization during metamorphism, and armoring of monazite inclusions within phases such as garnet (Foster et al., 2000; Montel et al., 2000; Catlos et al., 2002b), means that a metamorphic rock may contain multiple generations of monazite. This behavior, plus incorporation of \(^{230}\)Th, has limited the applicability of bulk or single-grain TIMS dating to monazites in polymetamorphic rocks, but affords important advantages when the variability in (sub)grain ages can be exploited by microbeam techniques such as electron-probe microanalysis (Suzuki & Adachi, 1991; Montel et al., 1996; Williams et al., 1999) or SIMS (Harrison et al., 1995; Stern & Berman, 2000). Moreover, because monazite can host a range of trace elements (Heinrich et al., 1997), it is capable of recording changes in P and T or mineral assemblage through compositional variation during (re)crystallization. In principle, a combined effort to measure the compositions and ages of different generations of monazite can be used to unravel the PT evolution of a rock (Spear & Pyle, 2002; Pyle & Spear, 2003; Kohn & Malloy, 2004). At present, Y and Th are the best understood chemical tracers in monazite (Pyle et al., 2001; Pyle & Spear, 2003). Y and the HREE are strongly partitioned into garnet, such that monazite growing coevally with garnet should show a core–rim decrease in Y (Pyle & Spear, 1999); likewise, monazite that grows after garnet will have a low Y content (Zhu & O'Nions, 1999a; Foster et al., 2000). Th shows the inverse relationship as it is preferentially incorporated in monazite (Pyle & Spear, 2003; Kohn & Malloy, 2004).

For this project, we will complete ion microprobe Th/Pb dating of metamorphic monazite to determine whether the monazites ages and textures show Oligocene or Miocene deformation as predicted by the channel flow hypothesis (see Fig. 5). Ion microprobe Th/Pb dating of metamorphic monazite will be completed on six samples/dome for which we have already published a petrologic evaluation that
includes phase zoning and textural relationships in conjunction with mineral compositions. We will use this sequence of steps:

1) Make P, Sc, Y and Cr x-ray maps of garnet to relate garnet growth to accessory phase relations and melting events (Pyle & Spear, 1999; Hermann & Rubatto, 2003; Pyle & Spear, 2003).

2) Use optical microscopy, back-scattered electron imaging, and x-ray mapping to identify monazites, evaluate their textural relationships with other phases, and make a preliminary assessment of monazite genesis (Pyle et al., 2001; Williams et al., 2002). Ideally monazites can be located as inclusions in old porphyroblasts, as inclusions in other matrix phases, and as matrix grains; each of these textural types may have a unique history that can help construct the PTt path (Hawkins & Bowring, 1997; Catlos et al., 2002b).

3) Crush enough sample to produce a representative monazite separate. Make an epoxy mount of some of the grains and polish to expose their interiors. Make epoxy mounts of silicate grains expected to host monazites and polish to expose their interiors. This is an effective technique for concentrating a group of relatively rare but common monazite host grains, such as garnet.

4) Use electron-probe microanalysis to identify inclusions in monazites in thin section and in the grain mount. Of particular interest are the presence or absence of P- or T-diagnostic phases or phase assemblages and whether they occur in the cores, mantles, or rims of the monazites.

5) Make x-ray and BSE images of the monazites in thin section and in the grain mounts (Zhu & O’Nions, 1999b) and then use EPMA to analyze the compositions of the identified monazite types (Pyle et al., 2002; Williams et al., 2002) to assess when during the evolution of the rock mineralogy the monazites grew (Zhu & O’Nions, 1999a; Foster et al., 2000). Monazites exhibit a variety of zoning types similar to zircon, including oscillatory zoning from igneous growth, quasi-concentric zoning, irregular zoning from alteration along cracks, rim overgrowths from Ostwald ripening or solution-transfer creep, or patchy zoning resulting from in situ recrystallization (Ayers et al., 1999; Hermann & Rubatto, 2003).

6) Use SIMS (in collaboration with Dr. Martin Whitehouse, NORDSIM facility, Swedish Museum of Natural History; see letter of collaboration) to measure ages of various monazite compositional and textural types.

5.6 Model Limitations

The channel-flow model constitutes an important paradigm guiding research in orogenic belts, but nevertheless, like any model, it is simplistic by its very nature. For example, the channel-flow models of Beaumont et al. (2001, 2004, 2006) use a finite-element grid of 3.5 x 40 km for individual general flow domains, yet our proposed research will yield flow data (shear senses and vorticity) at a much finer scale. Furthermore, mid-crustal rocks exposed in Kangmar and Mabja have been vertically thinned by at least 50% (Lee et al., 2000, 2004), which must be balanced by stretching parallel to boundaries of this mid-crustal shear zone, and extrusion at the surface (e.g. Vannay & Grasemann, 2001). Vertical thinning, however, is not explicitly incorporated into the channel-flow models of Beaumont et al. If we show that the channel-flow model is not viable for mid-crustal rocks in the North Himalayan gneiss domes, our studies are designed to allow us to refine these models or create a new one. Hence, even if our research does not confirm these working hypotheses, our data will provide essential insights into the spatial and temporal variation in the kinematics and vorticity of ductile flow of the Oligocene–Miocene middle crust of southern Tibet.

6. Role of Personnel and Work Plan

Lee is a structural and field geologist. Hacker is a metamorphic petrologist with expertise in EBSD and chronology. Dr. Yu Wang, China University of Geosciences, China, is a structural geologist with expertise in thermochronology (see letter of collaboration). Dr. Simon Wallis, Nagoya University, Japan, is a structural geologist who pioneered two of the vorticity analysis techniques we propose to use (see letter of collaboration). Dr. Martin Whitehouse, Swedish Museum of Natural History, Sweden, is a geochronologist.
Lee, Wang, 1 CWU graduate student, and 1 CWU undergraduate student will complete a minimum 45-day field season to collect detailed mesoscopic kinematic data, ~100 oriented samples from a wide range of lithologies for kinematic, vorticity, and EBSD analyses, six samples for metamorphic monazite geochronology, and photographs for finite-strain analyses from each dome. Upon return home, samples will be prepared for kinematic, vorticity, EBSD, and monazite geochronology analyses. CWU graduate and undergraduate students, under the supervision of Lee, Hacker, and Wallis, will be responsible for kinematic, vorticity, and finite strain studies, CWU graduate students, under the supervision of Hacker and Lee, will be responsible for the EBSD studies, and a UCSB graduate student, under the supervision of Hacker, Lee, and Whitehouse, will be responsible for the metamorphic monazite geochronology.

Procedure #1 (see proposal text): CWU MS graduate and undergraduate students, under the supervision of Lee, Hacker, and Wallis, will document shear sense using standard mesoscopic and microscopic analyses, deformation temperatures using standard petrography and quartz and feldspar textures, and vorticity using orientation of rigid porphyroblasts, porphyroblast hyperbolic distribution, and oblique grain foliation.

Procedure #2 (see proposal text): CWU MS graduate students, under the supervision of Hacker and Lee, will use EBSD to measure CPOs of quartz in quartzites and gneisses from the two gneiss domes to infer strain state, determine active slip systems, and deformation temperatures.

Procedure #3 (see proposal text): CWU MS graduate and undergraduate students, under the supervision of Lee, will use the Fry Method on rigid metamorphic porphyroblasts to measure finite strain in samples for which vorticity has been calculated.

Procedure #4 (see proposal text): A UCSB graduate student, under the supervision of Hacker, Lee, and Whitehouse, will use optical microscopy, back-scattered electron imaging, and x-ray mapping to identify monazites, evaluate their textural relationships with other phases, and make a preliminary assessment of monazite genesis. Make epoxy mounts of monazites and silicate grains. Use electron probe to identify inclusions in monazites in thin section and in the grain mount. Make x-ray and BSE images of the monazites in thin section and in the grain mounts and then use EPMA to analyze the compositions of the identified monazite types to assess when during the evolution of the rock mineralogy the monazites grew. Use SIMS to measure ages of monazites.

Year 1: Complete a minimum of a 45-day field season. Begin measuring and interpreting Mabja Dome samples.
Year 3: Finish Kangmar Dome samples and compile and publish results.

The PIs, graduate, and undergraduate students will meet once a year either at CWU or UCSB and at national meetings to integrate data collected at CWU, UCSB, and the ion-probe laboratory. The results of the research will be presented at Geological Society of America, American Geophysical Union, and similar meetings, and will be submitted as manuscripts for publication in journals in the fields of structure and tectonics.

7. Broader Impacts

Our proposed research will continue a decade-long collaboration with Chinese geologists, whose involvement will include field work, data interpretation, and co-authorships on peer reviewed publications. In addition, this proposal represents a collaborative effort among Lee and students at CWU, a primarily undergraduate institution, Hacker and students at UCSB, a research-I university, Wallis at Nagoya University, Japan, and Whitehouse at the Swedish Museum of Natural History, a research institution. These international and national collaborations will foster an exchange of research and teaching ideas, which will be particularly beneficial to our students. A critical aspect of this project is involvement of graduate (2 CWU, 1 UCSB) and undergraduate students (2 CWU). To broaden the scope and breadth of their education, these students will be involved in all aspects of field, petrologic, and analytical work, and preparation of publications. Lee and Hacker have long histories of mentoring
students in research. As an example of our commitment to engaging students in independent research, Lee’s undergraduate and MS graduate students have won seven awards from CWU since 2002 and 50% of his graduate students have been female.

An important goal of this project is to involve traditionally underrepresented undergraduate students (TUURS) (e.g. Native American, Hispanic, low income, women, and first-generation college students) in our proposed research. To identify and recruit TUURS into this research project, Lee will work carefully with CWU science faculty who received NSF funding to initiate a Science Talent Expansion Program (STEP) for TUURS (see letter of collaboration from Dr. Wendy Bohrson). Key components of STEP include recruiting high-achieving students in TUURS-serving high schools and community colleges in the central Washington area and once at CWU, enhancing knowledge and skill development through programs that involve these students in research. Recruiting by STEP will focus on central Washington high school districts, whose student population is composed of 40–70% Hispanic students and 6–50% of the Native American students. Because CWU has a tradition of excellence in teaching and education, and accessibility to adjacent Native American and Hispanic communities, recruiting should be successful.

Results from this work will be integrated into upper division/graduate classes in structural petrology at CWU and UCSB. Lee and Hacker will collaborate on developing vorticity laboratory exercises and make these available on the web. Lee and Hacker integrate rock projects into their classes, thus data and the thin sections/hand samples from this project will become part of the departmental structural petrology lab collections. Therefore undergraduates will benefit from this research because these rocks and the associated data will be integrated into upper division/graduate classes in structural geology. At CWU, the impact of these new laboratory exercises on undergraduates will be significant because the number of Bachelor’s degrees granted in the Geology program has increased from 5 to 25 per year during the last 15 years, while the number of degrees granted nation wide has remained essentially flat (www.agiweb.org/workforce/).