Channel flow, ductile extrusion and exhumation in continental collision zones: an introduction

L. GODIN1, D. GRUJIC2, R. D. LAW3 & M. P. SEARLE4

1Department of Geological Sciences & Geological Engineering, Queen’s University, Kingston, Ontario, K7L 3N6, Canada (e-mail: godin@geol.queensu.ca)
2Department of Earth Sciences, Dalhousie University, Halifax, Nova Scotia, B3H 4J1, Canada
3Department of Geological Sciences, Virginia Tech., Blacksburg, VA 24061, USA
4Department of Earth Sciences, Oxford University, Oxford, OX1 3PR, UK

Abstract: The channel flow model aims to explain features common to metamorphic hinterlands of some collisional orogens, notably along the Himalaya–Tibet system. Channel flow describes a protracted flow of a weak, viscous crustal layer between relatively rigid yet deformable bounding crustal slabs. Once a critical low viscosity is attained (due to partial melting), the weak layer flows laterally due to a horizontal gradient in lithostatic pressure. In the Himalaya–Tibet system, this lithostatic pressure gradient is created by the high crustal thicknesses beneath the Tibetan Plateau and ‘normal’ crustal thickness in the foreland. Focused denudation can result in exhumation of the channel material within a narrow, nearly symmetric zone. If channel flow is operating at the same time as focused denudation, this can result in extrusion of the mid-crust between an upper normal-sense boundary and a lower thrust-sense boundary. The bounding shear zones of the extruding channel may have opposite shear sense; the sole shear zone is always a thrust, while the roof shear zone may display normal or thrust sense, depending on the relative velocity between the upper crust and the underlying extruding material. This introductory chapter addresses the historical, theoretical, geological and modelling aspects of channel flow, emphasizing its applicability to the Himalaya–Tibet orogen. Critical tests for channel flow in the Himalaya, and possible applications to other orogenic belts, are also presented.

The hinterlands of collisional orogens are often characterized by highly strained, high-grade metamorphic rocks that commonly display features consistent with lateral crustal flow and extrusion of material from mid-crustal depths towards the orogenic foreland. A recent model for lateral flow of such weak mid-crustal layers has become widely known as the ‘channel flow’ model. The channel flow model has matured through efforts by several research groups and has also been applied to a variety of geodynamic settings. Thermal-mechanical modelling of collision zones, including the Himalayan–Tibetan system, has brought the concept of channel flow to the forefront of orogenic studies. Original contributors to the concept of channel flow initiated an important paradigm shift (Kuhn 1979), from geodynamic models of continental crust with finite rheological layering to the more encompassing channel flow model. This time-dependent mid- to lower crustal flow process, which will be reviewed in this chapter, may progress into foreland fold-and-thrust tectonics in the upper crust, thereby providing a spatial and temporal link between the early development of a metamorphic core in the hinterland and the foreland fold-and-thrust belt at shallower structural levels. Outcomes and implications of such a viscous flowing middle to lower crust include a dynamic coupling between mid-crustal and surface processes, and limitations to accurate retro-deformation of orogens (non-restorable orogens, e.g. Jamieson et al. 2006).

This Special Publication contains a selection of papers that were presented at the conference ‘Channel flow, extrusion, and exhumation of lower to mid-crust in continental collision zones’ hosted by the Geological Society of London at Burlington House, in December 2004. Because most of the ongoing debate on crustal flow focuses on the Cenozoic age Himalaya–Tibet collisional system, some of the key questions that are addressed in this volume include the following.

- Does the model for channel flow in the Himalaya–Tibet system concur with all available geological and geochronological data?
How do the pressure–temperature-time (P-T-t) data across the crystalline core of the Himalaya fit with the proposed channel flow?

Are the microstructural fabric data (pure shear and simple shear components) compatible with crustal extrusion (thickening or thinning of the slab)?

If the channel flow model is viable for the Himalaya–Tibet system, what may have initiated channel flow and ductile extrusion?

Why did the extrusion phase of the Himalayan metamorphic core apparently cease during the late Miocene–Pliocene?

Are some of the bounding faults of the potential channel still active, or were they recently active?

Is the Himalayan channel flow model exportable to other mountain ranges?

This introductory paper addresses the historical, theoretical, geological and modelling aspects of crustal flow in the Himalaya–Tibet orogen. Critical tests for crustal flow in the Himalaya, and possible applications to other orogenic belts, are presented and difficulties associated with applying these tests are discussed. Personal communication citations (pers. comm. 2004) identify comments expressed during the conference.

The Himalaya–Tibetan plateau system

The Himalaya–Tibet system initiated in Early Eocene times, following collision of the Indian and Eurasian plates (see Hodges (2000) and Yin & Harrison (2000) for reviews). The collision resulted in closure of the Tethyan Ocean, southward imbrication of the Indian crust, and northward continental subduction of Indian lower crust and mantle beneath Asia. The collision thickened the southern edge of the Asian crust to 70 km, and created the Tibetan Plateau, the largest uplifted part of the Earth’s surface with an average elevation of 5000 m (Fielding et al. 1994).

The Himalayan orogen coincides with the 2500-km-long topographic front at the southern limit of the Tibetan Plateau. It consists of five broadly parallel lithotectonic belts, separated by mostly north-dipping faults (Fig. 1). The Himalayan metamorphic core, termed the Greater Himalayan sequence (GHS), is bounded by two parallel and opposite-sense shear zones that were both broadly active during the Miocene (Hubbard & Harrison 1989; Searle & Rex 1989; Hodges et al. 1992, 1996). The Main Central thrust (MCT) zone marks the lower boundary of the GHS, juxtaposing the metamorphic core above the underlying Lesser Himalayan sequence. The South Tibetan detachment (STD) system defines the upper boundary roof fault of the GHS, marking the contact with the overlying unmetamorphosed Tethyan sedimentary sequence.

The apparent coeval movement of the MCT and STD, combined with the presence of highly sheared rocks and high grade to migmatitic rocks within the GHS, has led many workers to view the GHS as a north-dipping, southward-extruding slab of mid-crustal material flowing away from the thick southern edge of the Tibetan Plateau, towards the thinner foreland fold-thrust belt.

Dynamics of channel flow

The concepts of crustal extrusion and channel flow originated in the continental tectonics literature in the early 1990s. Unfortunately, these two processes are often referred to interchangeably without justification. One of the main points that emerged from the Burlington House conference was that a distinction between channel flow and crustal extrusion must be made. Parallel versus tapering bounding walls on channel flow and/or extrusion processes, and how these processes may replenish over time, are two resolvable parameters that are critical for distinguishing channel flow from extrusion. Brief definitions and overviews of the two processes are presented below. A more detailed overview of the mechanics of the related processes is provided by Grujic (2006).

Channel flow

Channel flow involves a viscous fluid-filled channel lying between two rigid sheets. The viscous fluid between the sheets is deformed through induced shear and pressure (or mean stress) gradients within the fluid channel (Fig. 2; e.g. Batchelor 2000; Turcotte & Schubert 2002). The weak layer flows laterally due to a horizontal gradient in lithostatic pressure; gravity is therefore the driving force. The finite displacement depends on the geometry of the channel, viscosity, and displacement rate of the bounding plates. In situations where the channel walls are non-parallel, the (non-lithostatic) pressure gradient may cause high rates of buoyant return flow of the channel material, provided that the viscosity is low enough (Mancktelow 1995; Gerya & Stöckhert 2002). The simplest qualitative characteristic of the channel flow model is that the velocity field consists of a hybrid between two end-members: (1) Couette flow (simple shear) between moving plates where the induced shear across the channel produces a uniform vorticity across the channel (Fig. 2A, left); and (2) Poiseuille flow (also known as ‘pipe-flow’ effect) between
stationary plates in which the induced pressure gradient produces highest velocities in the centre of the channel and opposite shear senses for the top and bottom of the channel (e.g. Mancktelow 1995; Fig. 2A, right).

Quantitative analyses of channel flow between moving or stationary boundaries (e.g. Batchelor 2000; Turcotte & Schubert 2002) have been applied to a wide range of geodynamic processes including: (1) asthenospheric counterflow (Chase 1979; Turcotte & Schubert 2002); (2) mechanics of continental extension (Kuszni & Matthews 1988; Block & Royden 1990; Birger 1991; Kruse et al. 1991; McKenzie et al. 2000; McKenzie & Jackson 2002); (3) continental plateau formation and evolution, both in extension and compression (Zhao & Morgan 1987; Block & Royden 1990; Wernicke 1990; Bird 1991; Fielding et al. 1994; Clark & Royden 2000; McQuarrie & Chase 2000; Hodges et al. 2001; Shen et al. 2001; Husson & Sempere 2003; Clark et al. 2005; Gerbault et al. 2005; Medvedev & Beaumont 2006); (4) tectonics of large continent–continent collision orogens (Johnston et al. 2000; Beaumont et al. 2001, 2004, 2006; Grujic et al. 2002, 2004; Williams & Jiang 2005; (5) metamorphic histories in large, hot, collisional orogens (Jamieson et al. 2002, 2004, 2006); (6) subduction zone flow regimes under both lithostatic and overpressured conditions (Bird 1978; England & Holland 1979; Shreve & Cloos 1986; Peacock 1992; Mancktelow 1995; Gerya et al. 2002; Gerya & Stöckhert 2002); and (7) deformation along passive continental margins in the presence of salt layers (Gemmer et al. 2004; Ings et al. 2004). For all these examples, with the exception of salt tectonics, the most likely cause for weakening in the channel is partial melting. In the case of salt tectonics, flow

Fig. 1. Simplified geological map of the Himalayan orogen, with general physiographic features of the Himalaya–Tibet system (inset). The Greater Himalayan sequence is bounded above by the north-dipping top-to-the-north South Tibetan detachment system (STDs), and below by the north-dipping top-to-the-south Main Central thrust zone (MCTz). The Himalaya is arbitrarily divided into four sections to facilitate age compilations (see Fig. 4). ITSZ: Indus–Tsangpo Suture Zone.
is due to the inherent low viscosity of salt under upper crustal conditions.

The application and significance of channel flow in continent–continent collisional settings is becoming progressively more refined, yet remains controversial. Evidence for channel flow and/or ductile extrusion of mid-crustal rocks from the geologically recent Himalaya–Tibet orogen (Grujic et al. 1996, 2002; Searle & Szulc 2005; Carosi et al. 2006; Godin et al. 2006; Hollister & Grujic 2006; Jessup et al. 2006; Searle et al. 2006) and related geodynamic models (Beaumont et al. 2001, 2004, 2006; Jamieson et al. 2004, 2006) are vigorously disputed (e.g. Hilley et al. 2005; Harrison 2006; Williams et al. 2006), while there are still few documented examples from older orogens (Jamieson et al. 2004; White et al. 2004; Williams & Jiang 2005; Brown & Gibson 2006; Carr & Simony 2006; Hatcher & Merschat 2006; Kuiper et al. 2006; Xypolias & Kokkalas 2006).

Extrusion

Extrusion is defined as the exhumation process of a channel (or the shallower part of it) operating at a localized denudation front. Channel flow and extrusion can operate simultaneously, with lateral tunnelling occurring at depth, while extrusion occurs at the front of the system, at progressively shallower crustal levels (Fig. 3). The focused denudation results in exhumation of the channel material within a narrow, nearly symmetric zone; the extruded channel is characterized by an upper normal-sense boundary, and a lower thrust-sense boundary. We present brief reviews of the four major channel flow and extrusion models.

Figure 3A presents a schematic overview of the kinematic relationships between channel flow and extrusion processes. The weak crustal channel flow (Fig. 3A, no. 5) is localized structurally below the 750°C isotherm, where melting starts (Fig. 3A, no. 8). Material points affected by a Poiseuille flow within the channel (ωp; vorticity in pure Poiseuille flow). For a given velocity of the subducting plate and channel width there is a critical viscosity of the channel material below which the Poiseuille flow will counteract the shear forces and cause return flow (negative velocity) and therefore exhumation of that part of the channel material. The part of the channel that remains dominated by the induced shear (positive velocities) will continue being underplated (ωs; vorticity in a hybrid channel flow). From Grujic et al. (2002), after Mancktelow (1995) and Turcotte & Schubert (2002).

Fig. 2. Schematic diagram of the flow pattern in a viscous channel of width h. The viscosity of channel material is lower than the viscosity of rocks in the hanging wall and in the footwall (μh > μc < μf). Velocity distributions are shown relative to a reference frame attached to the hanging wall. The vorticity values (rotational component of the flow profile) are schematically indicated by the width of the black bar: the wide bar segment indicates a high simple shear component; the narrow bar segment indicates a high pure shear component. Only the absolute value of the vorticity is indicated regardless of whether it is positive (sinistral simple shear) or negative (dextral simple shear). (A) End-members of flow in a channel; left, Couette flow with velocity profile caused by shearing (ωc; vorticity in pure Couette flow); right, Poiseuille flow with velocity profile caused by pressure gradient within the channel (ωp; vorticity in pure Poiseuille flow). (B) For a given velocity of the subducting plate and channel width there is a critical viscosity of the channel material below which the Poiseuille flow will counteract the shear forces and cause return flow (negative velocity) and therefore exhumation of that part of the channel material. The part of the channel that remains dominated by the induced shear (positive velocities) will continue being underplated (ωs; vorticity in a hybrid channel flow). From Grujic et al. (2002), after Mancktelow (1995) and Turcotte & Schubert (2002).
between the upper crust (Fig. 3A, no. 4) and the underlying extruding material. This is similar to the asymmetric thrust exhumation/extrusion mode described by Beaumont et al. (2004).

The extruding mid-crustal layer can be slab- or wedge-shaped, depending on the parallelism of the bounding shear zones, and advances towards the foreland. As the channel material is extruded, deformation is pervasively distributed within, or at the boundaries of the crustal layer. A concentration of deformation along the boundaries results in extrusion of a rigid crustal wedge (Fig. 3B; e.g. Burchfiel & Royden 1985; Hodges et al. 1992). This type of extrusion cannot be a long-lived geological process, but rather is probably a transient event (cf. Williams et al. 2006). Alternatively, deformation that is distributed throughout the wedge results in ductile extrusion (Fig. 3C; Grujic et al. 1996). The vorticity of flow within the extruded crust may be a perfect simple shear (Fig. 3C), or more likely a general shear combining components of simple shear and pure shear (Fig. 3D; Grujic et al. 1996; Grasemann et al. 1999; Vannay & Grasemann 2001; Law et al. 2004; Jessup et al. 2006; but cf. Williams et al. 2006).

During extrusion, the crustal slab or wedge cools from mid-crustal ductile flow to upper crustal brittle
conditions where deformation is partitioned into discrete faults. This concept resembles the buoyancy-driven extrusion of a crustal slab within a subduction zone (Chemenda et al. 1995). The differences between these two models include the size of the extruding wedge and nature of the primary driving forces. Analogue models of Chemenda et al. (1995) demonstrate that synollisional exhumation of previously subducted (underthrust) crustal material can occur due to failure of the subducting slab. In this model, erosional unloading causes the buoyant upper crust to be exhumed (at a rate comparable to the subduction rate), producing a normal-sense movement along the upper surface of the slab. This model, however, regards the exhumed crustal slice as a rigid slab bounded below and above by well-defined thrust and normal faults.

One of the key Himalayan problems is whether the GHS represents extrusion of a complete section of the mid-crust, with the STD–MCT surfaces representing potential channel-bounding structures, or whether the GHS is simply an extruded segment of a cooling channel, with the STD–MCT surfaces being more akin to roof and sole faults bounding a thrust duplex (Yin 2002). Furthermore, the GHS-bounding faults exposed at the topographic surface could be associated with late-stage exhumation of the GHS, rather than the original channel formed at depth beneath the Tibetan Plateau (Jessup et al. 2006). Related problems also concern the origin of fabrics within the GHS; are they related to flow during channeling, extrusion, or could they pre-date the Himalayan event? Another unresolved question is whether ‘the currently exposed GHS more closely resembles an exhumed plugged channel, with little extrusion during exhumation, or whether there was (and perhaps still is) active extrusion at the surface (e.g. Hodges et al. 2001; Wobus et al. 2003)’ (Beaumont et al. 2004, p. 26).

Exhumation

Exhumation is defined as the displacement of rocks with respect to the topographic surface (England & Molnar 1990), and requires either removal of the overburden (e.g. by erosion, normal faulting, vertical lithospheric thinning) or transport of material through the overburden (e.g. by diapirism, buoyancy-driven return flow in subduction zones) (see reviews by Platt 1993; Ring et al. 1999).

In the context of channel flow, exhumation of the channel occurs by a balance between orographically and topographically enhanced focused erosion and extrusion on the southern slopes of the Himalaya. The channel’s tunnelling capacity may be dramatically reduced as it is deflected upward during exhumation and cooling (Fig. 3A; Beaumont et al. 2004). Although exhumation of the GHS in the Himalaya may be associated with southward extrusion along coeval STD–MCT bounding faults, it may also be locally enhanced by post-MCT warping of the GHS and localized erosion following cessation of extrusion (Thiede et al. 2004; Vannay et al. 2004; Godin et al. 2006), and growth of duplexes in the footwall of the MCT during the middle Miocene (Robinson & Pearson 2006).

Exhumation of tunnelling material (not to be confused with extruded palaeo-channel material) will only occur if the active channel flow breaks through to the topographic surface. In the numerical models of channel flow, this is determined by the rheological properties of the upper-middle crust, frictional strength, and degree of advective thinning of the surface boundary layer (Beaumont et al. 2004). Conceptually, a threshold exhumation rate must be achieved to keep the channel sufficiently hot so that the material does not ‘freeze’ until it is close to the topographic surface (Beaumont et al. 2004). This situation may have been achieved in the two Himalayan ‘syntaxes’: the Nanga Parbat–Haramosh massif (e.g. Craw et al. 1994; Zeitler et al. 2001; Butler et al. 2002; Koons et al. 2002; Jones et al. 2006) and the Namche Barwa massif (e.g. Burg et al. 1997, 1998; Burg 2001; Ding et al. 2001).

Requirements and characteristics of channel flow

The following is a list of geological characteristics of channel flow, based on field observation and geodynamic modelling, and field criteria for recognizing an exhumed channel from the geological past (Searle & Szulc 2005; Searle et al. 2003, 2006). The criteria used to identify an active channel are based on geological and geophysical data (Nelson et al. 1996; S. Klemperer pers. comm. 2004; Klemperer 2006) and landscape analyses (Fielding et al. 1994; Clark et al. 2005).

1. A crustal package of lower viscosity material bounded by higher viscosity rocks.
2. A plateau with well-defined margins (or a significant contrast in crustal thickness) to produce a horizontal gradient in lithostatic pressure.
3. Coeval movement on shear zones with thrust and normal-fault geometry that bound the channel flow zone.
4. Kinematic inversion along the roof shear zone: earlier reverse-sense motion resulting from underthrusting (Couette flow) followed by normal-sense shearing resulting from back flow (by dominant Poiseuille flow) in
the channel, and/or by normal-sense motion on shear zones and brittle faults during extrusion and exhumation of the palaeo-channel.

(5) Pervasive shearing throughout the channel and extruded crustal block, although strain is predicted to be concentrated along its boundaries due to the flow geometry and deformation history.

(6) Inverted and right-way-up metamorphic sequences at the base and top of the extruding channel, respectively.

Modelling of the channel flow predicts the following tectonic consequences.

(1) The incubation period necessary for mid-crustal temperatures to rise, thereby increasing the melt content for commencement of channel flow, is typically between 10 and 20 million years from the time of onset of crustal thickening. This incubation period is judged necessary to increase the mid-crustal temperature sufficiently to produce the low viscosity necessary for initiation of channel flow.

(2) Melts (leucosomes) coeval with ductile channel flow must be younger than shortening structures in overlying rocks (upper crust).

(3) When active, the channel is predicted to be 10–20 km thick (Royden et al. 1997; Clark & Royden 2000; Beaumont et al. 2004; Jamieson et al. 2004, 2006).

(4) There is more lateral transport of material in the channel than vertical.

(5) Pre-existing structures cannot be traced through the channel (from the upper crust, through the channel and into channel footwall rocks). This has direct consequences on correlation possibilities between rock units and structures from the upper crust to the lower crust.

These conditions and consequences are reviewed for the Himalayan belt, and compared with available field and geochronological data.

**Viscosity**

A small percentage of partial melt significantly reduces the effective viscosity of rocks (Rosenberg & Handy 2005, and references therein). The GHS includes a significant percentage of migmatites and synorogenic leucogranites, while evidence for partial melting is absent in both the Lesser Himalayan sequence and the Tethyan sedimentary sequence. At the time of protracted peak temperature metamorphism (up to granulite facies) and melt generation, the GHS was therefore weaker than the overlying and underlying rocks by at least one order of magnitude (Beaumont et al. 2004, 2006; Hollister & Grujic 2006; Medvedev & Beaumont 2006).

The Lesser Himalayan sequence consists of a thick package of metasediments that were deformed under greenschist facies or lower conditions (< c. 300°C). Although these temperatures allow for ductile flow of quartz-dominated rocks, the expected viscosities are higher than for rocks with partial melt (see Medvedev & Beaumont 2006). The Tethyan sedimentary sequence is generally unmetamorphosed and only experienced greenschist-facies metamorphism in a narrow zone at its base (Garzanti et al. 1994; Godin 2003), although contact metamorphic aureoles have been reported associated with young granites emplaced in the Tethyan sedimentary sequence of southern Tibet (Lee et al. 2000, 2006).

**Platoe formation**

The Himalayan orogen is genetically linked to growth of the Tibetan Plateau (Hodges 2000; Yin & Harrison 2000). Various palaeo-elevation data suggest that the southern Tibetan Plateau has existed since at least the mid-Miocene (Blisniuk et al. 2001; Rowley et al. 2001; Williams et al. 2001; Spicer et al. 2003), attaining high elevations similar to the present day by 18–12 Ma, and possibly by 35 Ma (Rowley & Currie 2006). Geochronological and structural data suggest that east–west extension in the Tibetan Plateau – which is believed to be linked to crustal overthickening – was well underway by 14 Ma (Coleman & Hodges 1995; Williams et al. 2001). Contrasts in crustal thicknesses (between the Indian foreland and the Tibetan Plateau) that produced the necessary gravitational potential energy for channel flow (Bird 1991) may have existed at least since the Miocene, and perhaps earlier.

Creation of high topography by ‘inflational’ thickening is a potential consequence of crustal flow (Royden 1996; Burchfiel 2004). In the Longmen Shan belt of eastern Tibet, Clark & Royden (2000) and Clark et al. (2005) suggest that topography is generated when lower crustal flow butresses against cold, stronger crust (e.g. Sichuan basin). The resistance to lateral flow inflates the lower weak crustal zone, and supports a high topography above, and possibly generates ramping up (extrusion) and eventual exhumation of lower crust. This process could partly address concerns about the ‘support’ of the plateau’s high elevation, if Tibet is underlain by a weak, low-viscosity mid-crust (S. Lamb, pers. comm. 2004). A similar lithospheric strength contrast to the Longmen Shan could exist on the southern edge of the Himalaya, where the south-flowing weak mid-crust butresses against cold, strong Indian lithosphere, favouring extrusion and ‘inflational’ support of the southern Tibetan Plateau (e.g. Hodges et al. 2001).
Coeval channel-bounding structures

The MCT and STD zones include multiple fault strands that operated at different times and under different mechanical conditions (ductile to brittle). The broadly coeval activity of the MCT and the STD over extended geological time (from c. 25 Ma to 5 Ma) is documented by various sets of geochronological data. Figure 4 presents a compilation of interpreted age(s) of motion on the various strands of the MCT and STD, along the length of the Himalayan belt as taken directly from the available literature. We refer to these various strands as lower MCT (MCT\textsubscript{l}), upper MCT (MCT\textsubscript{u}), lower STD (STD\textsubscript{l}) and upper STD (STD\textsubscript{u}) to avoid confusion with past terminology. A major limitation to compiling such diverse data (Table 1) is the range of different approaches utilized by different authors to constrain either a maximum or minimum age of motion, on either the upper or lower strand of each fault system (from indirect geochronological tools – monazite crystal ages or peak metamorphic ages – to field relationships, e.g. pre- or post-kinematic intrusions). We emphasize that no attempt has been made in this compilation to critically assess the validity of different approaches taken by different authors in different areas.

The compiled data indicate that the MCT\textsubscript{u} and STD\textsubscript{u} were mostly active between 25–14 Ma and 24–12 Ma, respectively (Fig. 4). The activity along the higher STD\textsubscript{u} apparently started later and lasted longer (c. 19 Ma to 14 Ma, and perhaps is still active today, e.g. Hurtado et al. 2001) than along the more ductile lower STD\textsubscript{l}. The structurally lowest MCT\textsubscript{l} appears as the youngest structure (c. 15 Ma to 0.7 Ma). Combined, the available data indicate simultaneous or overlapping periods of thrust- and normal-sense ductile shearing between c. 24 Ma and 12 Ma. However, with time, the position of the active faults moved towards upper and lower structural levels, and became more diachronous and possibly less dynamically linked (e.g. Godin et al. 2006). Since the early recognition of the STD, the coeval activity on the two bounding (and innermost) shear zones (MCT\textsubscript{u} and STD\textsubscript{u}), and its implication for exhumation of the metamorphic core of the Himalaya, has been suggested (Burchfiel & Royden, 1985; Hubbard & Harrison 1989; Searle & Rex, 1989; Burchfiel et al. 1992; Hodges et al. 1992, 1996; Grujic et al. 1996; Grasemann et al. 1999); however, the proposed driving forces and kinematic details vary between authors.

The late Miocene to recent activity along the MCT\textsubscript{l} overlaps with activity along the two in-sequence external thrust faults, the Main Boundary thrust (MBT) and Main Frontal thrust (MFT), and may represent on-going exhumation of the modern cryptic (hypothetical) channel (Hodges et al. 2004). In the context of proposed exhumation by combined channel flow and extrusion, a corresponding active zone of normal faulting at a higher structural level is required. The data are scarce but there are indications of neotectonic faulting along the northern boundary of the GHS (Hodges et al. 2001, 2004; Hurtado et al. 2001; Wiesmayr et al. 2002). The younging of structures away from the core of the orogen may suggest progressive widening of the channel as it passes from the channel flow to extrusion mode of exhumation (Searle & Godin 2003; Searle et al. 2003, 2006).

Kinematic inversions

Studies have shown that deformation along the STD is distributed in the adjacent footwall and/or hanging wall for up to 3–4 km, rather than being restricted to a single fault plane. Most of these studies indicate an overprint of top-to-the-north (normal sense) shearing on an older top-to-the-south thrusting within the STD system (Burg et al. 1984; Brun et al. 1985; Kündig 1989; Burchfiel et al. 1992; Vannay & Hodges 1996; Carosi et al. 1998; Godin et al. 1999a, 2001; Grujic et al. 2002; Wiesmayr & Grasemann 2002). Based on field evidence for a reversal in shear sense during motion along the STD, it seems likely that the return flow of the metamorphic core (relative to the underthrusting Indian plate) developed late in the channel flow history. The kinematic history of the STD is further complicated by overprinting top-to-the-south shearing (e.g. Godin et al. 1999a; Godin 2003). Some of this late stage overprinting may relate to north-dipping thrust faults in the Tethyan sedimentary sequence, between the suture zone and the STD (e.g. Ratschbacher et al. 1994). Geodynamic modelling (Beaumont et al. 2004) supports this possibility if the weak channel overburden fails and glides towards the foreland causing relative thrusting along the upper boundary of the channel.

Internal deformation within the channel

On regional cross-sections and maps, the MCT and STD are often depicted as sharp boundaries; however, both are broad ductile shear zones. Although most field- and laboratory-based investigations agree that there is a broad zone of deformation adjacent to the MCT and STD, pervasively distributed ductile shear throughout the GHS is also documented (e.g. Jain & Manickavasagam 1993; Grujic et al. 1996; Grasemann et al. 1999; Jessup et al. 2006). The kinematics of deformation consistently indicate top-to-the-south shearing in the Lesser Himalayan sequence and in most of the
Fig. 4. Compilation of interpreted ages of motion on the Main Central thrust (MCT) and South Tibetan detachment (STD) systems. MCT-Lower refers to the mostly brittle, structurally lower fault in the MCT zone. Local names include MCT-1, Ramgarh thrust, and Munsiari thrust. MCT-Upper refers to the mostly ductile, synmetamorphic, structurally higher fault in the MCT zone. Local names include MCT-2, Vaikrita thrust, Mahabharat thrust and Chomrong thrust. STD-Lower refers to the mostly ductile, synmetamorphic, structurally lower fault in the STD system. Local names include Zanskar detachment, Sangla detachment, Annapurna detachment, Deurali detachment, Chomrong detachment, and Lhotse detachment. STD-Lower refers to the mostly ductile, post-metamorphic, structurally higher fault in the STD system. Local names include Jhalta detachment, Machapuchhare detachment, Phu detachment, and Qomolangma detachment. See Table 1 for the complete list of data and references. The four geographical areas refer to the subdivisions presented in Figure 1. The thick dashed lines represent best-fit ages for motion on the faults.
Table 1. Compilation of interpreted ages of fault motion on the Main Central thrust system and South Tibetan detachment system, based on geochronological data

<table>
<thead>
<tr>
<th>Structure</th>
<th>Location</th>
<th>Age</th>
<th>Minerals</th>
<th>System</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Western Himalaya</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>STD&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Zanskar</td>
<td>&lt;18 Ma to 16 Ma</td>
<td>Ms, Bt</td>
<td>Rb-Sr</td>
<td>1. Inger (1998)</td>
</tr>
<tr>
<td>STD&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Sutlej</td>
<td>23 Ma to 17 Ma</td>
<td>Mz, Ms</td>
<td>Th-Pb, Ar</td>
<td>2. Vannay et al. (2004)</td>
</tr>
<tr>
<td>STD&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Zanskar</td>
<td>23 Ma to 20 Ma</td>
<td>Ms, Bt, Ar, U-Pb</td>
<td>Xe, Mz, Zr</td>
<td>3. Walker et al. (1999)</td>
</tr>
<tr>
<td>STD&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Zanskar</td>
<td>c. 22.2 Ma to 19.8 Ma</td>
<td>Mz, Ms</td>
<td>U-Pb, Ar</td>
<td>4. Dézes et al. (1999)</td>
</tr>
<tr>
<td>STD&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Garhwal</td>
<td>23 to 21 Ma</td>
<td>Mz, Ms</td>
<td>U-Pb, Ar</td>
<td>5. Searle et al. (1999)</td>
</tr>
<tr>
<td>STD&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Zanskar</td>
<td>c. 23 Ma to 20 Ma</td>
<td>Ms, Bt</td>
<td>Ar</td>
<td>6. Vance et al. (1998)</td>
</tr>
<tr>
<td>STD&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Zanskar</td>
<td>c. 26 Ma to 18 Ma</td>
<td>Ms, Bt</td>
<td>Rb-Sr</td>
<td>7. Inger (1998)</td>
</tr>
<tr>
<td>STD&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Garhwal</td>
<td>&lt;21.9 Ma</td>
<td>Mz</td>
<td>Th-Pb</td>
<td>8. Harrison et al. (1997a)</td>
</tr>
<tr>
<td>STD&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Zanskar</td>
<td>21 Ma to 19.5 Ma</td>
<td>Mz</td>
<td>U-Pb</td>
<td>9. Noble &amp; Searle (1995)</td>
</tr>
<tr>
<td>MCT&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Sutlej</td>
<td>23 Ma to 17 Ma</td>
<td>Mz, Ms</td>
<td>Th-Pb, Ar</td>
<td>10. Vannay et al. (2004)</td>
</tr>
<tr>
<td>MCT&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Garhwal</td>
<td>c. 5.9 Ma</td>
<td>Mz</td>
<td>Th-Pb</td>
<td>11. Catlos et al. (2002)</td>
</tr>
<tr>
<td>MCT&lt;sub&gt;1&lt;/sub&gt;</td>
<td>West Nepal</td>
<td>22 Ma to 15 Ma</td>
<td>Ms</td>
<td>Ar</td>
<td>12. DeCelles et al. (2001)</td>
</tr>
<tr>
<td>MCT&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Garhwal</td>
<td>c. 6 to 0.7 Ma</td>
<td>Ms, Ap, Zr</td>
<td>Ar, FT</td>
<td>13. Vannay et al. (2004)</td>
</tr>
<tr>
<td>MCT&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Garhwal</td>
<td>c. 22.5 Ma</td>
<td>Mz</td>
<td>U-Pb</td>
<td>14. DeCelles et al. (2001)</td>
</tr>
<tr>
<td><strong>Central-West Himalaya</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>STD&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Nar</td>
<td>19 Ma to 16 Ma</td>
<td>Ms, Bt, Hbl</td>
<td>Ar</td>
<td>15. Godin et al. (2006)</td>
</tr>
<tr>
<td>STD&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Kali Gandaki</td>
<td>&lt;17.2 ka</td>
<td>Terr</td>
<td>14C</td>
<td>17. Hurtado et al. (2001)</td>
</tr>
<tr>
<td>STD&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Langtang</td>
<td>&lt;17.3 Ma</td>
<td>Mz, Xe</td>
<td>U-Pb</td>
<td>18. Searle et al. (1997)</td>
</tr>
<tr>
<td>STD&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Annapurna</td>
<td>c. 18.5 Ma</td>
<td>Zr</td>
<td>U-Pb</td>
<td>19. Hodges et al. (1996)</td>
</tr>
<tr>
<td>STD&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Kali Gandaki</td>
<td>15 Ma to 13 Ma</td>
<td>Mz</td>
<td>Ar</td>
<td>20. Vannay &amp; Hodges (1996)</td>
</tr>
<tr>
<td>STD&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Manaslu</td>
<td>19 Ma to 16 Ma</td>
<td>Bt, Ms</td>
<td>Ar</td>
<td>21. Guillot et al. (1994)</td>
</tr>
<tr>
<td>STD&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Kali Gandaki</td>
<td>c. 22.5 Ma</td>
<td>Mz</td>
<td>U-Pb</td>
<td>22. Godin et al. (2001)</td>
</tr>
<tr>
<td>STD&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Manaslu</td>
<td>&gt;22.9 Ma</td>
<td>Mz</td>
<td>Th-Pb</td>
<td>23. Harrison et al. (1999b)</td>
</tr>
<tr>
<td>STD&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Annapurna</td>
<td>22.5 Ma to 18.5 Ma</td>
<td>Zr, Mz, Xy</td>
<td>U-Pb</td>
<td>25. Hodges et al. (1996)</td>
</tr>
<tr>
<td>STD&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Manaslu</td>
<td>&gt;22 Ma</td>
<td>Hbl</td>
<td>Ar</td>
<td>26. Guillot et al. (1994)</td>
</tr>
<tr>
<td>STD&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Manaslu</td>
<td>&lt;20 Ma</td>
<td>Mz</td>
<td>Th-Pb</td>
<td>27. Copeland et al. (1990)</td>
</tr>
<tr>
<td>MCT&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Langtang</td>
<td>16 Ma to 13 Ma</td>
<td>Mz</td>
<td>Th-Pb</td>
<td>28. Kohn et al. (2004)</td>
</tr>
<tr>
<td>MCT&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Kathmandu</td>
<td>22 Ma to 14 Ma</td>
<td>Mz, Zr</td>
<td>U-Pb</td>
<td>29. Johnson et al. (2001)</td>
</tr>
<tr>
<td>MCT&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Marsyandi</td>
<td>22 to 18 Ma</td>
<td>Mz</td>
<td>U-Pb</td>
<td>30. Coleman (1998)</td>
</tr>
<tr>
<td>MCT&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Kathmandu</td>
<td>21 Ma to 14 Ma</td>
<td>Ms, Bt</td>
<td>Rb-Sr</td>
<td>31. Johnson &amp; Rogers (1997)</td>
</tr>
<tr>
<td>MCT&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Annapurna</td>
<td>c. 22.5 Ma</td>
<td>Mz, Zr</td>
<td>U-Pb</td>
<td>32. Hodges et al. (1996)</td>
</tr>
<tr>
<td>MCT&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Kali Gandaki</td>
<td>&gt;15 Ma</td>
<td>Ms</td>
<td>Ar</td>
<td>33. Vannay &amp; Hodges (1996)</td>
</tr>
<tr>
<td>MCT&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Kali Gandaki</td>
<td>c. 22 Ma</td>
<td>Mz, Th</td>
<td>U-Pb</td>
<td>34. Nazarchuk (1993)</td>
</tr>
<tr>
<td>MCT&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Langtang</td>
<td>&gt;5.8 Ma</td>
<td>Ms</td>
<td>Ar</td>
<td>35. Macfarlane (1993)</td>
</tr>
<tr>
<td>MCT&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Langtang</td>
<td>c. 9 Ma</td>
<td>Ms</td>
<td>Ar</td>
<td>36. Kohn et al. (2004)</td>
</tr>
<tr>
<td>MCT&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Marsyandi</td>
<td>c. 13.3 Ma</td>
<td>Mz</td>
<td>U-Pb</td>
<td>37. Catlos et al. (2001)</td>
</tr>
<tr>
<td>MCT&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Kathmandu</td>
<td>c. 17.5 Ma</td>
<td>Ms, Bt</td>
<td>Rb-Sr</td>
<td>38. Johnson &amp; Rogers (1997)</td>
</tr>
<tr>
<td>MCT&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Marsyandi</td>
<td>c. 16 Ma</td>
<td>Mz</td>
<td>Th-Pb</td>
<td>39. Harrison et al. (1997b)</td>
</tr>
<tr>
<td>MCT&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Langtang</td>
<td>&lt;9 to 7 Ma; c. 2.3 Ma</td>
<td>Mz</td>
<td>Th-Pb</td>
<td>40. Macfarlane et al. (1992, Macfarlane (1993)</td>
</tr>
<tr>
<td><strong>Central-East Himalaya</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>STD&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Everest</td>
<td>&lt;16 Ma</td>
<td>Mz</td>
<td>U-Pb</td>
<td>41. Searle et al. (2003)</td>
</tr>
<tr>
<td>STD&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Everest</td>
<td>c. 17 Ma</td>
<td>Mz</td>
<td>Th-Pb</td>
<td>42. Murphy &amp; Harrison (1999)</td>
</tr>
<tr>
<td>STD&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Everest</td>
<td>c. 16 Ma</td>
<td>Xe, Mz, Zr</td>
<td>U-Pb</td>
<td>43. Hodges et al. (1998)</td>
</tr>
<tr>
<td>STD&lt;sub&gt;1&lt;/sub&gt;</td>
<td>Everest</td>
<td>22 Ma to 19 Ma</td>
<td>Ti, Xe, Hbl</td>
<td>U-Pb, Ar</td>
<td>44. Hodges et al. (1992)</td>
</tr>
</tbody>
</table>

(Continued)
INTRODUCTION

Table 1. Continued

<table>
<thead>
<tr>
<th>Structure¹</th>
<th>Location²</th>
<th>Age³</th>
<th>Minerals⁴</th>
<th>System⁵</th>
<th>Reference⁶</th>
</tr>
</thead>
<tbody>
<tr>
<td>STD₁ Everest</td>
<td>&lt;16.8 Ma</td>
<td>Mz</td>
<td>U-Pb</td>
<td>45. Schärer et al. (1986)</td>
<td></td>
</tr>
<tr>
<td>STD₁ Everest</td>
<td>c. 21,2 Ma</td>
<td>Mz, Xe</td>
<td>U-Pb</td>
<td>46. Viskupic et al. (2005)</td>
<td></td>
</tr>
<tr>
<td>STD₁ Everest</td>
<td>18 Ma to 17 Ma</td>
<td>Mz, Xe</td>
<td>U-Pb</td>
<td>47. Searle et al. (2003)</td>
<td></td>
</tr>
<tr>
<td>STD₁ Everest</td>
<td>&lt;20.5 Ma</td>
<td>Mz, Xe, U</td>
<td>U-Pb</td>
<td>48. Simpson et al. (2000)</td>
<td></td>
</tr>
<tr>
<td>STD₁ Makalu</td>
<td>&lt;21.9 Ma</td>
<td>Mz</td>
<td>U-Pb</td>
<td>49. Schärer (1984)</td>
<td></td>
</tr>
<tr>
<td>MCT₁ Everest</td>
<td>c. 21,2 Ma</td>
<td>Mz, Xe</td>
<td>U-Pb</td>
<td>50. Viskupic et al. (2005)</td>
<td></td>
</tr>
<tr>
<td>MCT₁ Everest</td>
<td>c. 25 to 23 Ma</td>
<td>Mz</td>
<td>Th-Pb</td>
<td>51. Catlos et al. (2002)</td>
<td></td>
</tr>
<tr>
<td>MCT₁ Everest</td>
<td>c. 21 Ma</td>
<td>Hbl</td>
<td>Ar</td>
<td>52. Hubbard &amp; Harrison (1989)</td>
<td></td>
</tr>
<tr>
<td>MCT₁ Everest</td>
<td>23 Ma to 20 Ma</td>
<td>Hbl, Bt</td>
<td>Ar</td>
<td>53. Hubbard (1989)</td>
<td></td>
</tr>
</tbody>
</table>

Eastern Himalaya

<table>
<thead>
<tr>
<th>Structure¹</th>
<th>Location²</th>
<th>Age³</th>
<th>Minerals⁴</th>
<th>System⁵</th>
<th>Reference⁶</th>
</tr>
</thead>
<tbody>
<tr>
<td>STD₁ Sikkim</td>
<td>c. 14,5 Ma</td>
<td>Mz, Zr</td>
<td>Th-Pb</td>
<td>54. Catlos et al. (2004)</td>
<td></td>
</tr>
<tr>
<td>STD₁ Khula Kangri</td>
<td>&lt;12.5 Ma</td>
<td>Mz</td>
<td>Th-Pb</td>
<td>55. Edwards &amp; Harrison (1997)</td>
<td></td>
</tr>
<tr>
<td>STD₁ Wagye La</td>
<td>c. 12 Ma</td>
<td>Mz</td>
<td>U-Pb</td>
<td>56. Wu et al. (1998)</td>
<td></td>
</tr>
<tr>
<td>STD₁ Sikkim</td>
<td>23 Ma to 16 Ma</td>
<td>Grt</td>
<td>Sm-Nd</td>
<td>57. Harris et al. (2004)</td>
<td></td>
</tr>
<tr>
<td>STD₁ Sikkim</td>
<td>c. 17 Ma</td>
<td>Mz, Zr</td>
<td>Th-Pb</td>
<td>58. Catlos et al. (2004)</td>
<td></td>
</tr>
<tr>
<td>MCT₁ Sikkim</td>
<td>c. 22 Ma</td>
<td>Mz</td>
<td>Th-Pb</td>
<td>59. Catlos et al. (2004)</td>
<td></td>
</tr>
<tr>
<td>MCT₁ Sikkim</td>
<td>23 Ma to 16 Ma</td>
<td>Grt</td>
<td>Sm-Nd</td>
<td>60. Harris et al. (2004)</td>
<td></td>
</tr>
<tr>
<td>MCT₁ Bhutan</td>
<td>c. 22 Ma; 18 Ma to 13 Ma</td>
<td>Mz, Xe</td>
<td>U-Pb</td>
<td>61. Daniel et al. (2003)</td>
<td></td>
</tr>
<tr>
<td>MCT₁ Bhutan</td>
<td>c. 113.5 Ma</td>
<td>Mz</td>
<td>U-Pb</td>
<td>62. Grujic et al. (2002)</td>
<td></td>
</tr>
<tr>
<td>MCT₁ Bhutan</td>
<td>&gt;14 Ma</td>
<td>Ms</td>
<td>Ar</td>
<td>63. Stüwe &amp; Foster (2001)</td>
<td></td>
</tr>
<tr>
<td>MCT₁ Sikkim</td>
<td>15 Ma to 10 Ma</td>
<td>Mz</td>
<td>Th-Pb</td>
<td>64. Catlos et al. (2004)</td>
<td></td>
</tr>
<tr>
<td>MCT₁ Bhutan</td>
<td>&lt;11 Ma</td>
<td>Ms</td>
<td>Ar</td>
<td>65. Stüwe &amp; Foster (2001)</td>
<td></td>
</tr>
</tbody>
</table>

¹MCT₁, Lower MCT; (and/or) mostly brittle, post-metamorphic; local names include MCT-1, Ramgarh, Munsiari. MCT₂, Upper MCT; (and/or) ductile, synmetamorphic, synmagmatic; local names include MCT-2, Vanka, Mahabharat, Chomrong. STD₁, Lower STD; (and/or) mostly brittle, post-metamorphic; local names include Zanskar, Sangla, Annapurna, Deurali, Chame, Lhotse, Zherger La, STD₂, Upper STD; (and/or) mostly brittle, post-metamorphic; local names include Jhala, Macchupuchare, Phu, Qomolangma.
²See Figure 1 for location.
³Compilation of direct geochronological results only; includes age constraints based on cross-cutting structures/intrusion relationships.
⁴Ar, apatite; Bt, biotite; Grt, garnet; Hbl, hornblende; Ms, muscovite; Mz, monazite; Terr, terraces; Th, thorite; Ti, titanite; U, uraninite; Xe, xenotime; Zr, zircon.
⁵¹⁴⁷Sm/⁹²⁷Nd, ⁴⁰Ar/³⁹Ar thermochronology; ¹⁴C, carbon 14; FT; fission track geochronology; Rb-Sr, Sm-Nd, Th-Pb: thorium-lead ion microprobe (208Pb/²³²Th age); U-Pb, U-(Th)-Pb geochronology.
⁶Reference numbers refer to respective ‘age range bar’ on Figure 4.

GHS, with top-to-the-north shearing only appearing in the top-most part of the GHS, near and within the STD system. The location of this transition in shear sense has yet to be documented. Field and microstructural data indicate that this pervasive ductile deformation is characterized by heterogeneous general non-coaxial flow (components of both simple and pure shear) rather than by ideal simple shear. For example, quartz petrofabric data (Boullier & Bouchez 1978; Brunel 1980, 1983; Bouchez & Pécher 1981; Burg et al. 1984; Greco 1989; Grujic et al. 1996; Grasemann et al. 1999; Bhattacharya & Weber 2004; Law et al. 2004) consistently indicate a component of pure shear. Williams et al. (2006), however, present an opposite interpretation of these data based on strain compatibility and mechanics theory.

Quantitative vorticity analyses within the GHS document a progressively increasing component of simple shear traced upward towards the STD and overlying sheared Tethyan sedimentary rocks (Law et al. 2004; Jessup et al. 2006), a general shear deformation within the core of the GHS (Carosi et al. 1999a, b, 2006; Grujic et al. 2002; Law et al. 2004; Vannay et al. 2004; see also Fig. 3C), and an increasing pure shear component traced downward towards the underlying MCT zone (Grasemann et al. 1999; Jessup et al. 2006; but cf. Bhattacharya & Weber 2004). Macro- and microstructural fabric data (especially conjugate shear bands, porphyroclast inclusion trails, and culmination cleavage at various stages of development) also suggest a strong component of shortening across the foliation in addition to foliation-parallel shearing (e.g. Carosi et al. 1999a, b, 2006; Grujic et al. 2002; Law et al. 2004; Vannay et al. 2004). The structural data indicate that ductile deformation is pervasively distributed through the entire GHS, in the top part of the Lesser Himalayan sequence, and at the base of the Tethyan sedimentary sequence. A direct implication of the general flow model is that the bounding surfaces of the
crystalline core (i.e. MCT and STD shear zones) must therefore be ‘stretching faults’ (Means 1989) accommodating transport-parallel pervasive stretching of the crystalline core during internal flow (Grasemann et al. 1999; Vannay & Grasemann 2001; Law et al. 2004).

**Metamorphic characteristics**

One of the most intriguing phenomena of the Himalaya is the inverted metamorphic sequence present in both the Lesser Himalayan sequence and GHS (see reviews by Hodges 2000). At the top of the GHS and at the base of the Tethyan sedimentary sequence, a strongly attenuated, right-way-up decrease in metamorphic grade is present. Models for inverted metamorphism include: (1) overthrusting of hot material (‘hot iron effect’; Le Fort 1975); (2) imbricate thrusting (Brunel & Kienast 1986; Harrison et al. 1997b, 1998, 1999a); (3) folding of isograds (Searle & Rex 1989); (4) transposition of a normally zoned metamorphic sequence due to either localized simple shear along the base of the GHS (Jain & Manickavasagam 1993; Hubbard 1996), heterogeneous simple shear distributed across the Lesser Himalayan sequence and GHS (Grujic et al. 1996; Jamieson et al. 1996; Searle et al. 1999) or general shear of previously foreland-dipping isograds (Vannay & Grasemann 2001); and (5) shear heating (England et al. 1992; Harrison et al. 1998; Catlos et al. 2004). The metamorphic isograds can be deformed passively according to various kinematic models that are compatible with either extrusion or channel flow, or both (e.g. Searle et al. 1988, 1999; Searle & Rex 1989; Jain & Manickavasagam 1993; Grujic et al. 1996, 2002; Hubbard 1996; Jamieson et al. 1996; Davidson et al. 1997; Daniel et al. 2003). Coupled thermal mechanical finite element modelling (Jamieson et al. 2004) has been successful in replicating the distribution of the metamorphic isograds and P-T-t data obtained through field and laboratory studies, although it failed to predict the timing of the low temperature metamorphic overprint. Other models propose a specific style of thrusting along the base of the GHS as an alternative model to explain both the distribution of metamorphic zones and the timing of metamorphism (e.g. Harrison et al. 1998; Catlos et al. 2004).

**Lateral versus vertical transport of material**

Integration of geobarometry and thermochronology can deduce the amount and timing of exhumation: more specifically, the rate of vertical displacement of rocks within the crust. Only the vertical component of exhumation can be estimated using these techniques. Along low-angle shear zones like the STD and MCT, however, the horizontal component of displacement is predominant. Some investigations use the jump in pressures, estimated by metamorphic assemblages across the STD, to estimate the horizontal component of displacement (Searle et al. 2002, 2003). Displacement estimates based on temperatures inferred from metamorphic assemblages, however, involve assumptions about the shape of the isotherms, which may change during the exhumation process. Simplified restoration of the GHS (e.g. INDEPTH data; Nelson et al. 1996; Hauck et al. 1998) indicate that the GHS may extend down-dip for at least 200 km, and possibly up to 400 km (Grujic et al. 2002). Exhumation from mid-crustal levels at 35–40 km (as suggested by pressures at peak T; see Hodges (2000) for summary of data, and Hollister & Grujic (2006) for interpretation) indicates that lateral displacement rates in the GHS are five to ten times larger than the vertical displacement rates. These values ought to be compared with inferred surface denudation rates (e.g. Thiede et al. 2004; Vannay et al. 2004; Grujic et al. 2005), and estimation of shortening or displacements across the MCT and STD. Conventional cross-section (usually line-length) restoration techniques are used to estimate these values (e.g. Schelling & Arita 1991; DeCelles et al. 2002; Searle et al. 2003). However, if deformation is pervasive through the GHS and there is an inversion of the displacement along the STD, no single value can fully describe the displacement along the shear zone. Displacements across the GHS relative to the Lesser Himalayan sequence are also expected to progressively increase towards the core, and progressively decrease upward towards the STD, which is compatible with the calculations of particle displacement paths for various points within a model GHS (Jamieson et al. 2004, 2006).

**Discontinuity of protoliths across the channel**

According to the above discussion, the largest rate of particle displacement change occurs across the STD and MCT (e.g. Davidson et al. 1997). Most detrital zircon and isotopic studies suggest that the Lesser Himalayan sequence and GHS metasediments may have different protolith ages. Zircon and Nd model ages and the εNd values suggest a Late Archean to Palaeoproterozoic source for the metasediments of the Lesser Himalayan sequence versus a Meso- to Neoproterozoic source for
The GHS (Parrish & Hodges 1996; Whittington et al. 1999; Ahmad et al. 2000; DeCelles et al. 2000, 2004; Miller et al. 2001; Robinson et al. 2001; Argles et al. 2003; Martin et al. 2005; Richards et al. 2005; but cf. Myrow et al. 2003), although structural restoration suggests otherwise (e.g. Walker et al. 2001). The lithotectonic units, separated by the first-order shear zones, may have distinct palaeo-geographic origins; however, this does not necessarily mean they belong to different tectonic plates. Similar results are obtained by numerical modelling and particle tracking (Jamieson et al. 2006), which suggest that from the base to the top of the GHS, the protoliths should have a progressively more distal origin (with respect to the pre-collision plate margin), while the opposite situation is predicted for the Lesser Himalayan sequence.

Although different protolith origins for the GHS and the Lesser Himalayan sequence might exist, a similar interpretation cannot be applied to the GHS and the Tethyan sedimentary sequence. Channel flow models predict that the STD should be the locus for large relative particle displacement, implying a different origin for the GHS and the Tethyan sedimentary sequence (Jamieson et al. 2006). Recent structural restorations and isotopic studies, however, propose the lower Tethyan sedimentary sequence as a potential protolith for some of the GHS (Vannay & Grasemann 2001; Argles et al. 2003; Gehrels et al. 2003; Searle & Godin 2003; Gleeson & Godin 2006; Richards et al. 2005). The STD is generally interpreted as either a décollement surface (stretching fault), where the thick pile of continental margin rocks (Tethyan sedimentary sequence) has been decoupled without much internal disturbance to the stratigraphy, or a passive roof thrust within the MCT system, with a hanging-wall flat–footwall flat geometry (Searle et al. 1988; Yin 2002).

The GHS is dominated by three lithologic units, which maintain their respective structural positions for over a thousand kilometres along-strike (Gansser 1964; Le Fort 1975). Recent detailed mapping across the GHS locally reveals a more complex distribution of, and variation within, these units (Searle & Godin 2003; Searle et al. 2003; Gleeson & Godin 2006). Nonetheless, the first-order lateral continuity of the GHS units indicates an apparent lack of internal stratigraphic disturbance. This has been highlighted as a possible pitfall for the channel flow model (Harrison 2006). Model results indicate, however, that the channel may very well maintain internal ‘stratigraphy’, as long as the deformation is concentrated along the boundaries and flow is planar along the length of the channel (Jamieson et al. 2006).

**Timing of melting and shortening structures**

The channel flow model assumes that melts (leucosomes and granites) will substantially reduce the viscosity of a crustal layer (i.e. channel). It also predicts that these melts should be younger than shortening structures found in the upper plate–shortening structures that would have created the necessary crustal thickening and ensuing heating to partially melt and lower the viscosity of the underlying mid-crust. The Tethyan sedimentary sequence is the upper plate in the Himalaya.

Leucosome and leucogranite bodies occur within all units of the GHS (Dietrich & Gansser 1981; Le Fort et al. 1987; Burchﬁel et al. 1992; Guillot et al. 1993; Hodges et al. 1996; Hollister & Grujic 2006). Most U–Th–Pb ages for the melts in the central Himalaya range from 23–22 Ma (Harrison et al. 1995; Hodges et al. 1996; Coleman 1998; Searle et al. 1999; Godin et al. 2001; Daniel et al. 2003; Harris et al. 2004) to 13–12 Ma (Edwards & Harrison 1997; Wu et al. 1998; Zhang et al. 2004). However, evidence for leucosome melt production during the Oligocene also exists (Coleman 1998; Thimm et al. 1999; Godin et al. 2001). North Himalayan granites found in southern Tibet range in crystallization age between 28 Ma and 9 Ma (Scharer et al. 1986; Harrison et al. 1997a; Zhang et al. 2004; Aoya et al. 2005). Syntectonic (synchannel?) granites yield ages of 23.1 ± 0.8 (Lee et al. 2006). Some North Himalayan granites, however, yield zircon and monazite crystallization ages of 14.2 ± 0.2 Ma and 14.5 ± 0.1 Ma, respectively, indicating that vertical thinning and subhorizontal stretching had ceased by the middle Miocene (Aoya et al. 2005; Lee et al. 2006).

Several phases of deformation are recorded by the overlying Tethyan sedimentary sequence (Steck et al. 1993; Wiesmayr & Grasemann 2002; Godin 2003). Although the absolute age(s) of the dominant shortening structures is disputed, most authors agree that significant thickening of the Tethyan sedimentary sequence occurred prior to the Miocene, most likely in the Oligocene or even before (Hodges et al. 1996; Vannay & Hodges 1996; Godin et al. 1999b, 2001; Wiesmayr & Grasemann 2002; Godin 2003; Searle & Godin 2003). Some of these shortening features are interpreted to be coeval with high-pressure metamorphism in the GHS (Eo-Himalayan phase; Hodges 2000), associated with early burial of the GHS beneath a thickening overlying Tethyan sedimentary sequence (Godin et al. 1999b, 2001; Godin 2003).
Channel thickness and late-stage modifications

During periods of active channel flow, models predict that the channel should be 10 to 20 km thick (Royden et al. 1997; Clark & Royden 2000; Beaumont et al. 2004; Jamieson et al. 2004, 2006). The structural thickness of the GHS varies considerably, from 2–3 km in the Annapurna area (Searle & Godin 2003; Godin et al. 2006), up to 30 km in the Everest area (Searle et al. 2003, 2006; Jessup et al. 2006), and even more in the Bhutan Himalaya (Grujic et al. 2002). Substantial post-channel, post-extrusion modifications have altered the original geometry of the channel. Out-of-sequence thrusts such as the Kakhtang thrust or altered the original geometry of the channel. Outpost-channel, post-extrusion modifications have been attributed to variations in thickness of the GHS at the present-day topographic front could reflect along-strike variation in the foreland-directed advance of a channel flow regime.

Challenges and unresolved issues

The challenge to testing the applicability of the channel flow model in the Himalaya–Tibet system lies within the Earth scientist's ability to accurately interpret deformation paths and palaeo-isothermal structures recorded by exhumed metamorphic rocks that exhibit finite strain and metamorphic field gradients. Limited subsurface geophysical coverage of the Himalaya–Tibet system makes correlation of surficial data with a putative channel at mid-crustal depths tentative.
thermal–mechanical models is a critically important first step in ensuring that models are well-calibrated to a specific orogen. They further argue that ‘tuning’ a model to match a specific orogen should not be regarded as a weakness of the modelling method, but is the basis for a better understanding of which factors are likely to have the most influence on orogenic processes such as channel flow and crustal extrusion.

Concluding remarks

The proposed channel flow model explains many features pertaining to the geodynamic evolution of the Himalaya–Tibetan Plateau system, as well as other older orogenic systems. It reconciles the apparent coeval nature of the MCT and STD faults and kinematic inversions at the top of the GHS, leading to southward extrusion and exhumation of the crystalline core of the Himalaya from beneath the Tibetan Plateau. In addition, it provides an alternative and quantitative explanation of the inverted metamorphic sequence at the orogen scale, and effectively couples the tectonic and surface processes. The proposal that the middle or lower crust acts as a ductile, partially molten channel flowing out from beneath areas of overthickened crust (such as the Tibetan Plateau) towards the topographic surface at the plateau margins remains controversial, however, both with respect to the Himalaya–Tibetan system and particularly older, less well documented orogenic systems. The channel flow model nonetheless presents an exciting new conceptual framework for understanding the geodynamic evolution of crystalline cores of orogenic belts, and may become the source for a paradigm shift in continental tectonics studies.

The following 26 papers in this Special Publication are arranged into four main groups. In the first group of papers this brief introduction to channel flow and ductile extrusion processes is paired with a more in-depth review by Grujic of channel flow processes associated with continental collisional tectonics. In the second group of papers detailed overviews are given by Klemperer and Hodges of the geophysical and geological databases from which the concepts of channel flow and ductile extrusion in the Himalaya–Tibetan Plateau system originally developed.

Different aspects of the modelling of channel flow and ductile extrusion processes are covered in the third group of papers. Coupled thermal–mechanical finite element models are presented in papers by Beaumont et al., Medvedev & Beaumont and Jamieson et al., while the effects of volume change on orogenic extrusion are considered by Grasemann et al. In the last two papers in this group, problems associated with identifying channel flow and ductile extrusion in older orogens are discussed by Jones et al., while linkages between flow at different crustal levels (infrastructure and suprastructure) and constraints on the efficiency of ductile extrusion processes are explored by Williams et al.

The fourth and largest group of papers is composed of a series of predominantly field-based case studies providing geological constraints on channel flow and ductile extrusion as an orogenic process. This last group of papers is divided into subsections on the Himalaya–Tibetan Plateau system, the Hellenic and Appalachian orogenic belts, and the Canadian Cordillera. The Himalaya subsection begins with a wide-ranging critique by Harrison of the applicability of channel flow models to the Himalaya–Tibetan Plateau system. Subsequent papers in the Himalaya subsection focus dominantly on specific field areas within the Lesser and Greater Himalaya and are arranged in order of geographic location starting with western Nepal (Robinson & Pearson) and then progressing eastwards through the Annapurna region of central Nepal (Godin et al., Scaillet & Searle, Annen & Scaillet) and the Nyalam–Everest regions of Tibet and eastern Nepal (Wang et al., Searle et al., Jessup et al.) to the Bhutan Himalaya (Hollister & Grujic, Carosi et al.). The Himalaya–Tibetan Plateau subsection concludes with papers by Lee et al. and Aoya et al. on gneiss domes exposed to the north of the Himalaya and their implications for mid-crustal flow beneath southern Tibet.

Geological evidence for and against channel flow and ductile extrusion in older orogenic systems is discussed in the remaining two subsections of this volume. Xypolias & Kokkalas present integrated strain and vorticity data indicating ductile extrusion of mid-crustal quartz-rich units in the Hellenides of Greece, while Hatcher & Merschat present field evidence in support of channel flow operating parallel to orogenic strike in the Appalachian Inner Piedmont, USA. Arguments for (Brown & Gibson, Kuiper et al.) and against (Carr & Simony) channel flow in the crystalline interior of the Canadian Cordillera are presented in the final three papers of the volume.

The authors wish to warmly thank all participants of the 2004 Burlington House conference for fruitful discussions during and after the meeting. We thank C. Beaumont, R. L. Brown, R. A. Jamieson, M. Jessup, K. Larson, S. Medvedev, R. A. Price and P. F. Williams for discussion, and K. Larson, M. Jessup, and Chief Editor J. P. Turner for their detailed reviews of earlier versions of this manuscript. D.G. acknowledges support from the Canadian Institute for Advanced Research (CIAR). L.G., R.D.L. and M.S. are funded by the Natural Sciences and
Engineering Research Council of Canada (NSERC), the National Science Foundation of the United States (NSF), and the National Environment Research Council of the United Kingdom (NERC), respectively.

References


INTRODUCTION


Hubbard, M. S. & Harrison, T. M. 1989. 40Ar/39Ar age constraints on deformation and metamorphism.


INTRODUCTION


Murphy, M. A. & Harrison, T. M. 1999. Relationship between leucogranites and the Qomolangma detachment in the Rongbuk valley, South Tibet. Geology, 27, 831–834.


INTRODUCTION


