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

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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?

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- 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

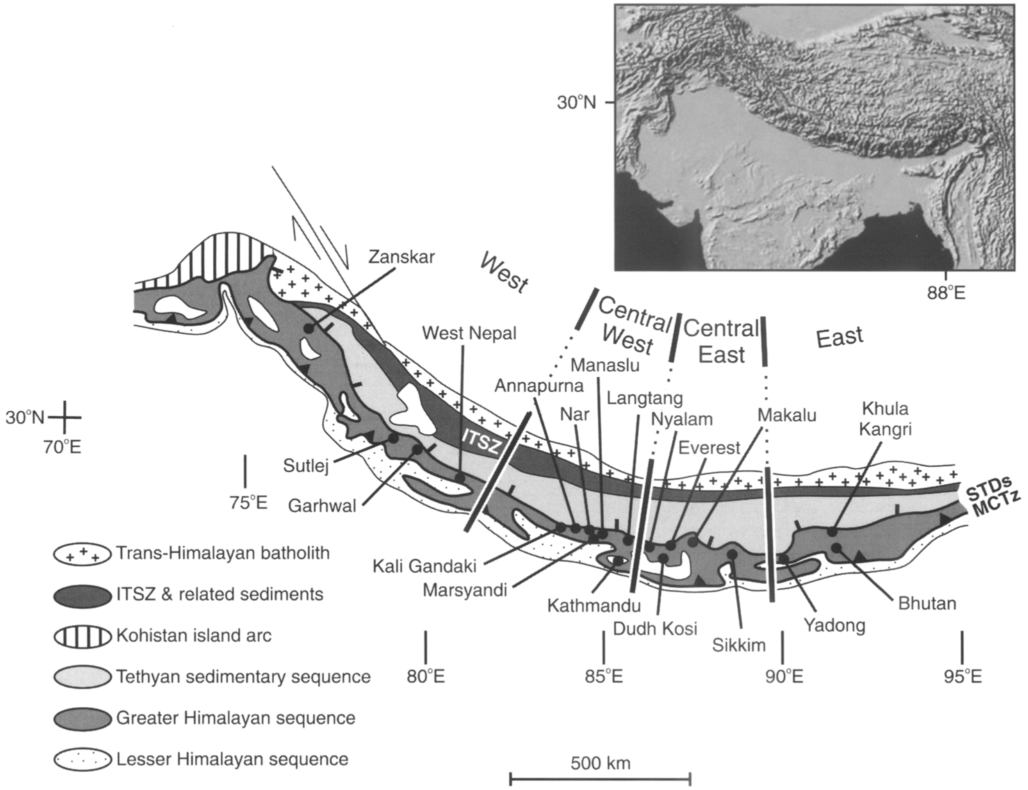

Fig. 1. Simplified geological map of the Himalayan orogen, with general physiographic features of the Himalaya–Tibet system (inset). The Greater Himalayan sequence is bounded above by the north-dipping top-to-the-north South Tibetan detachment system (STDs), and below by the north-dipping top-to-the-south 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.

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 (Kusznir & 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

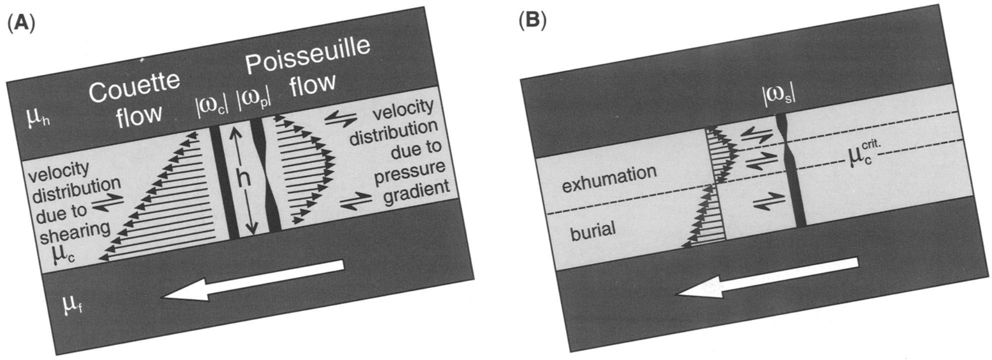

Fig. 2. Schematic diagram of the flow pattern in a viscous channel of width h. The viscosity of channel material is lower than the viscosity of rocks in the hanging wall and in the footwall ($\mu_h > \mu_c < \mu_f$). Velocity distributions are shown relative to a reference frame attached to the hanging 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 (ω_c: vorticity in pure Couette flow); right, Poiseuille flow with velocity profile caused by pressure gradient within the channel (ω_p: 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 (ω_s: vorticity in a hybrid channel flow). From Grujic *et al.* (2002), after Mancktelow (1995) and Turcotte & Schubert (2002).

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 (Fig. 3A, no. 7) traverse the rheological boundary at the tip of the channel (Fig. 3A, no. 9), and will continue their exhumation by extrusion of the palaeo channel (Fig. 3A, no. 10). It follows that the ductile outward channel flow converts to ductile extrusion where rock motion is balanced by surface erosion (Fig. 3A, no. 13). The extrusion is aided by focused erosion along the mountain front (e.g. Grujic *et al.* 2002; Vannay *et al.* 2004), the rate of extrusion/exhumation being proportional to the rate of flow in the mid-crustal channel. Extrusion of the crustal layer is assumed to occur along two bounding shear zones (Fig. 3A, nos 11–12). The bounding shear zones 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

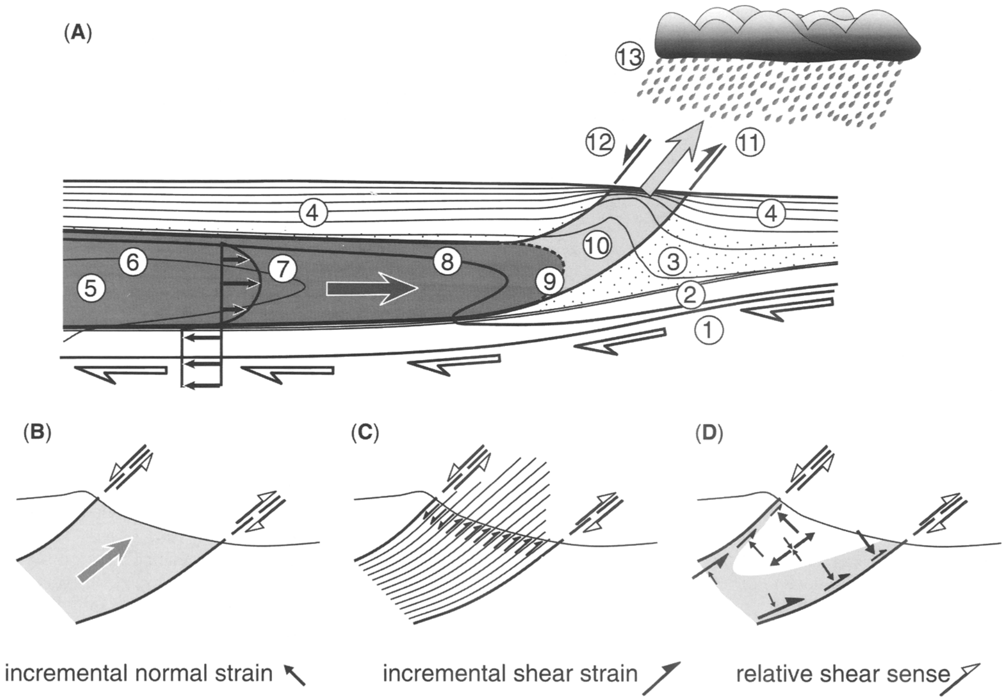

Fig. 3. (**A**) Schematic diagram of kinematic relationship between channel flow and extrusion of a palaeo-channel. All material depicted belongs to the underthrusting plate. 1, Lithospheric mantle; 2, lower crust; 3, mid-crust; 4, upper crust; 5, weak crustal channel; 6, isotherms (taken from Beaumont *et al.* 2004), 7, schematic velocity profile during return channel flow; 8, 750° C isotherm structurally below which partial melt starts; 9, rheological tip of the channel: at lower temperatures (for a given channel width and pressure gradient) Couette flow will dominate and all the material will be underthrust; 10, extruding crustal block (palaeo-channel): if the rheological tip is at steady state, material points may move through this tip and pass from the weak crustal channel into the extruding block; 11, lower shear zone of the extruding crustal block (dominantly thrust-sense kinematics); 12, upper shear zone of the extruding crustal block (dominantly normal-sense kinematics); 13, focused surface denudation (controlled by surface slope and, as in the Himalaya, by orographic precipitation at the topographic front). (**B**)–(**D**) Possible strain distribution in an extruding crustal block; possible end-members (inspired by Grujic *et al.* 1996; Grasemann *et al.* 1999). (**B**) Rigid block with high concentration of strain along the boundaries (Hodges *et al.* 1996). (**C**) Ductile block deforming by pervasive simple shear. (**D**) Ductile block deforming by general shear with a pure shear component increasing towards the bottom of the wedge, as well as with time following a 'decelerating strain path' (Grasemann *et al.* 1999; Vannay & Grasemann 2001).

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 syncollisional 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 channelling, 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 ($< c.\ 300^{\circ}C$). 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).

Plateau 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 buttresses 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 buttresses 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_l) and upper MCT (MCT_u), and lower STD (STD_l) and upper STD (STD_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_u and STD_l were mostly active between 25–14 Ma and 24–12 Ma, respectively (Fig. 4). The activity along the higher STD_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_l. The structurally lowest MCT_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_u and STD_l), 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_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.* 1999*a*, 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.* 1999*a*; 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

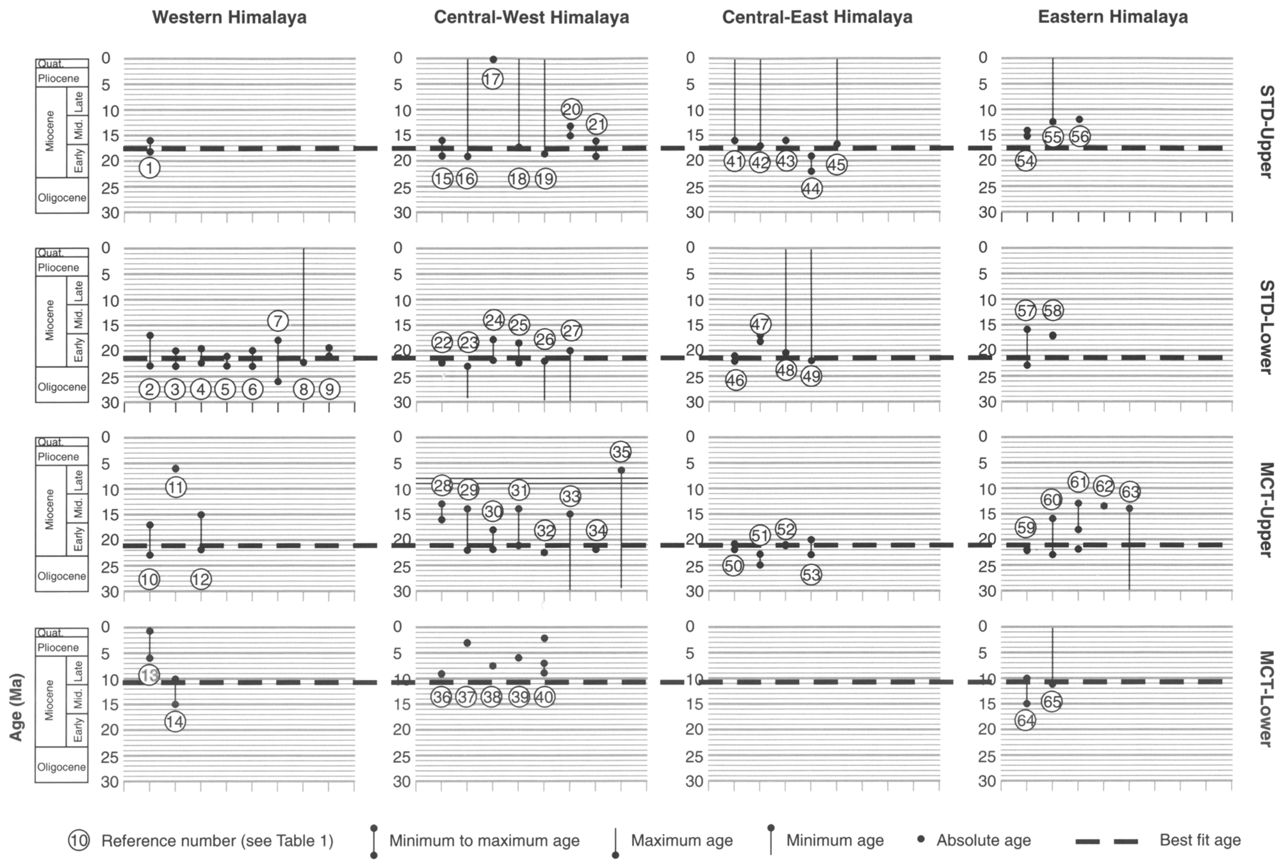

Fig. 4. Compilation of interpreted ages of motion on the Main Central thrust (MCT) and South Tibetan detachment (STD) systems. MCT-Lower refers to the mostly brittle, structurally lower fault in the MCT zone. Local names include MCT-1, Ramgarh thrust, and Munsiari thrust. MCT-Upper refers to the mostly 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, Chame detachment, Lhotse detachment and Zherger La detachment. STD-Upper refers to the mostly brittle, post-metamorphic, structurally higher fault in the STD system. Local names include Jhala detachment, Macchapuchhare 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*

Structure[1]	Location[2]	Age[3]	Minerals[4]	System[5]	Reference[6]
Western Himalaya					
STD_u	Zanskar	<18 Ma to 16 Ma	Ms, Bt	Rb-Sr	1. Inger (1998)
STD_l	Sutlej	23 Ma to 17 Ma	Mz, Ms	Th-Pb, Ar	2. Vannay *et al.* (2004)
STD_l	Zanskar	23 Ma to 20 Ma	Ms, Bt, Xe, Mz, Zr	Ar, U-Pb	3. Walker *et al.* (1999)
STD_l	Zanskar	*c.* 22.2 Ma to 19.8 Ma	Mz, Ms	U-Pb, Ar	4. Dèzes *et al.* (1999)
STD_l	Garhwal	23 to 21 Ma	Mz, Ms	U-Pb, Ar	5. Searle *et al.* (1999)
STD_l	Zanskar	*c.* 23 Ma to 20 Ma	Ms, Bt	Ar	6. Vance *et al.* (1998)
STD_l	Zanskar	*c.* 26 Ma to 18 Ma	Ms, Bt	Rb-Sr	7. Inger (1998)
STD_l	Garhwal	<21.9 Ma	Mz	Th-Pb	8. Harrison *et al.* (1997*a*)
STD_l	Zanskar	21 Ma to 19.5 Ma	Mz	U-Pb	9. Noble & Searle (1995)
MCT_u	Sutlej	23 Ma to 17 Ma	Mz, Ms	Th-Pb, Ar	10. Vannay *et al.* (2004)
MCT_u	Garhwal	*c.* 5.9 Ma	Mz	Th-Pb	11. Catlos *et al.* (2002)
MCT_u	West Nepal	22 Ma to 15 Ma	Ms	Ar	12. DeCelles *et al.* (2001)
MCT_l	Sutlej	*c.* 6 to 0.7 Ma	Ms, Ap, Zr	Ar, FT	13. Vannay *et al.* (2004)
MCT_l	West Nepal	15 Ma to 10 Ma	Ms	Ar	14. DeCelles *et al.* (2001)
Central-West Himalaya					
STD_u	Nar	19 Ma to 16 Ma	Ms, Bt, Hbl	Ar	15. Godin *et al.* (2006)
STD_u	Nar	<19 Ma	Mz	Th-Pb	16. Searle & Godin (2003)
STD_u	Kali Gandaki	<17.2 ka	Terr	^{14}C	17. Hurtado *et al.* (2001)
STD_u	Langtang	<17.3 Ma	Mz, Xe	U-Pb	18. Searle *et al.* (1997)
STD_u	Annapurna	*c.* 18.5 Ma	Zr	U-Pb	19. Hodges *et al.* (1996)
STD_u	Kali Gandaki	15 Ma to 13 Ma	Ms	Ar	20. Vannay & Hodges (1996)
STD_u	Manaslu	19 Ma to 16 Ma	Bt, Ms	Ar	21. Guillot *et al.* (1994)
STD_l	Kali Gandaki	*c.* 22.5 Ma	Mz	U-Pb	22. Godin *et al.* (2001)
STD_l	Manaslu	>22.9 Ma	Mz	Th-Pb	23. Harrison *et al.* (1999*b*)
STD_l	Marsyandi	22 Ma to 18 Ma	Mz	U-Pb	24. Coleman (1998), Coleman & Hodges (1998)
STD_l	Annapurna	22.5 Ma to 18.5 Ma	Zr, Mz, Xy	U-Pb	25. Hodges *et al.* (1996)
STD_l	Manaslu	>22 Ma	Hbl	Ar	26. Guillot *et al.* (1994)
STD_l	Manaslu	>20 Ma	Ms	Ar	27. Copeland *et al.* (1990)
MCT_u	Langtang	16 Ma to 13 Ma	Mz	Th-Pb	28. Kohn *et al.* (2004)
MCT_u	Kathmandu	22 Ma to 14 Ma	Mz, Zr	U-Pb	29. Johnson *et al.* (2001)
MCT_u	Marsyandi	22 to 18 Ma	Mz	U-Pb	30. Coleman (1998)
MCT_u	Kathmandu	21 Ma to 14 Ma	Ms, Bt	Rb-Sr	31. Johnson & Rogers (1997)
MCT_u	Annapurna	*c.* 22.5 Ma	Mz, Zr	U-Pb	32. Hodges *et al.* (1996)
MCT_u	Kali Gandaki	>15 Ma	Ms	Ar	33. Vannay & Hodges (1996)
MCT_u	Kali Gandaki	*c.* 22 Ma	Mz, Th	U-Pb	34. Nazarchuk (1993)
MCT_u	Langtang	>5.8 Ma	Ms	Ar	35. Macfarlane (1993)
MCT_l	Langtang	*c.* 9 Ma	Ms	Ar	36. Kohn *et al.* (2004)
MCT_l	Marsyandi	*c.* 13.3 Ma	Mz	U-Pb	37. Catlos *et al.* (2001)
MCT_l	Kathmandu	*c.* 17.5 Ma	Ms, Bt	Rb-Sr	38. Johnson & Rogers (1997)
MCT_l	Marsyandi	*c.* 16 Ma	Mz	Th-Pb	39. Harrison *et al.* (1997*b*)
MCT_l	Langtang	<9 to 7 Ma; *c.* 2.3 Ma	Ms	Ar	40. Macfarlane *et al.* (1992, Macfarlane (1993)
Central-East Himalaya					
STD_u	Everest	<16 Ma	Mz	U-Pb	41. Searle *et al.* (2003)
STD_u	Everest	*c.* 17 Ma	Mz	Th-Pb	42. Murphy & Harrison (1999)
STD_u	Everest	*c.* 16 Ma	Xe, Mz, Zr	U-Pb	43. Hodges *et al.* (1998)
STD_u	Everest	22 Ma to 19 Ma	Ti, Xe, Hbl	U-Pb, Ar	44. Hodges *et al.* (1992)

(*Continued*)

Table 1. *Continued*

Structure[1]	Location[2]	Age[3]	Minerals[4]	System[5]	Reference[6]
STD$_u$	Nyalam	<16.8 Ma	Mz	U-Pb	45. Schärer *et al.* (1986)
STD$_l$	Everest	*c.* 21$_u$2 Ma	Mz, Xe	U-Pb	46. Viskupic *et al.* (2005)
STD$_l$	Everest	18 Ma to 17 Ma	Mz, Xe	U-Pb	47. Searle *et al.* (2003)
STD$_l$	Everest	<20.5 Ma	Mz, Xe, U	U-Pb	48. Simpson *et al.* (2000)
STD$_u$	Makalu	<21.9 Ma	Mz	U-Pb	49. Schärer (1984)
MCT$_u$	Everest	*c.* 21$_u$2 Ma	Mz, Xe	U-Pb	50. Viskupic *et al.* (2005)
MCT$_u$	Dudh Kosi	*c.* 25 to 23 Ma	Mz	Th-Pb	51. Catlos *et al.* (2002)
MCT$_u$	Everest	*c.* 21 Ma	Hbl	Ar	52. Hubbard & Harrison (1989)
MCT$_u$	Everest	23 Ma to 20 Ma	Hbl, Bt	Ar	53. Hubbard (1989)
Eastern Himalaya					
STD$_u$	Sikkim	*c.* 14$_l$5 Ma	Mz, Zr	Th-Pb	54. Catlos *et al.* (2004)
STD$_u$	Khula Kangri	<12.5 Ma	Mz	Th-Pb	55. Edwards & Harrison (1997)
STD$_l$	Wagye La	*c.* 12 Ma	Mz	U-Pb	56. Wu *et al.* (1998)
STD$_l$	Sikkim	23 Ma to 16 Ma	Grt	Sm-Nd	57. Harris *et al.* (2004)
STD$_l$	Sikkim	*c.* 17 Ma	Mz, Zr	Th-Pb	58. Catlos *et al.* (2004)
MCT$_u$	Sikkim	*c.* 22 Ma	Mz	Th-Pb	59. Catlos *et al.* (2004)
MCT$_u$	Sikkim	23 Ma to 16 Ma	Grt	Sm-Nd	60. Harris *et al.* (2004)
MCT$_u$	Bhutan	*c.* 22 Ma; 18 Ma to 13 Ma	Mz, Xe	U-Pb	61. Daniel *et al.* (2003)
MCT$_u$	Bhutan	*c.* 113.5 Ma	Mz	U-Pb	62. Grujic *et al.* (2002)
MCT$_u$	Bhutan	>14 Ma	Ms	Ar	63. Stüwe & Foster (2001)
MCT$_l$	Sikkim	15 Ma to 10 Ma	Mz	Th-Pb	64. Catlos *et al.* (2004)
MCT$_l$	Bhutan	<11 Ma	Ms	Ar	65. Stüwe & Foster (2001)

[1]MCT$_l$, Lower MCT; (and/or) mostly brittle, post-metamorphic; local names include MCT-1, Ramgarh, Munsiari. MCT$_u$, Upper MCT; (and/or) ductile, synmetamorphic, synmagmatic; local names include MCT-2, Vaikrita, Mahabharat, Chomrong. STD$_l$, Lower STD; (and/or) ductile, synmetamorphic, synmagmatic; local names include Zanskar, Sangla, Annapurna, Deurali, Chame, Lhotse, Zherger La. STD$_u$, Upper STD; (and/or) mostly brittle, post-metamorphic; local names include Jhala, Macchupuchare, Phu, Qomolangma
[2]See Figure 1 for location.
[3]Compilation of direct geochronological results only; includes age constraints based on cross-cutting structures/intrusion relationships.
[4]Ap, apatite; Bt, biotite; Grt, garnet; Hbl, hornblende; Ms, muscovite; Mz, monazite; Terr, terraces; Th, thorite; Ti, titanite; U, uraninite; Xe, xenotime; Zr, zircon.
[5]Ar, $^{40}Ar/^{39}Ar$ thermochronology; ^{14}C, carbon 14; FT; fission track geochronology; Rb-Sr, Sm-Nd, Th-Pb: thorium-lead ion microprobe ($^{208}Pb/^{232}Th$ age); U-Pb, U-(Th)-Pb geochronology.
[6]Reference 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 & Pêcher 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.* 1999*a*, *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 crenulation 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.* 1999*a*, *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.* 1997*b*, 1998, 1999*a*); (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 (Schärer *et al.* 1986; Harrison *et al.* 1997*a*; 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.* 1999*b*, 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 (Eohimalayan phase; Hodges 2000), associated with early burial of the GHS beneath a thickening overlying Tethyan sedimentary sequence (Godin *et al.* 1999*b*, 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 the Kalopani shear zone (Grujic *et al.* 1996, 2002; Vannay & Hodges 1996), and large amplitude folding of the GHS (Johnson *et al.* 2001; Gleeson & Godin 2006; Godin *et al.* 2006) may account for some of the observed thickness variation. Alternatively, various instabilities and failure of the upper crust may induce local accretion of channel material and sequential development of domes (i.e. spatio-temporal variations of channel thickness), both along-strike and down-dip (e.g. Beaumont *et al.* 2004). Some out-of-sequence thrusting may be the result of such pulsed channel flow and related doming and extrusion (Grujic *et al.* 2004; Hollister & Grujic 2006).

In the Kali Gandaki (Annapurna area) and eastern Bhutan, the GHS is as thin as 3 km, while preserving its typical apparent internal 'stratigraphy' and metamorphic zoning. Whether this is a reflection of a lateral variation in the component of coaxial (pure shear) deformation and thinning during the channelling and/or extrusion phase remains unclear. Thin segments of the GHS may represent the most proximal parts of the channel, while the thicker segments are more distal parts of the palaeo-channel. In this alternate interpretation, variation 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. The INDEPTH program laid the foundation for imaging critical mid- to lower crustal features that the channel flow hypothesis relies on. Unfortunately, data for the Himalaya are limited to a single transect. Higher resolution and more extensive seismic surveys may help resolve long-standing criticism of the channel flow models. For example, Harrison (2006) highlights the risk of generalizing the 'bright spots' low velocity zones to the entire southern Tibetan Plateau, because the seismic line was run within an active graben, where intrusions might be locally controlled by extension, and not crustal-scale melting.

Assessing the applicability of the model to older, more deeply exhumed orogens will also prove to be challenging because of overprinting deformation and thermal events common to these systems. In older orogenic systems, probably the most important limitation to the application of the channel flow concept is the absence of control on palaeo-horizontal and palaeo-vertical directions. The channel flow model in the Himalaya–Tibet system is based on the concept of a lateral lithostatic pressure gradient acting as the driver for flow. Lateral variations in crustal thicknesses at the time of orogenesis in older systems are simply unknown. Furthermore, as pointed out by Jones *et al.* (2006), many of the structural/kinematic indicators of a channelized flow could also be compatible with tectonically rather than gravitationally driven systems (e.g. transpressional strike-slip tectonics).

Potentially the most important limitation to testing the channel flow model is the fact that many of the initial parameters used in numerical models such as those presented by Beaumont *et al.* (2001, 2004) are taken directly from field and geochronological data collected in the Himalaya–Tibet system. Some participants at the Burlington House conference argued that it therefore follows that assessing the channel flow model, and testing its applicability against field constraints, becomes an inherently circular argument. In contrast, other participants (e.g. D. Grujic pers. comm. 2004) argued that one of the major strengths of the rigorously constructed thermal–mechanical generic models is their ability to test a wide range of potentially important parameters, and thereby identify the combinations of parameters that produce results which most closely resemble a given orogenic system. Yet other participants argued that where model results are sensitive to a wide range of potential boundary conditions and input parameters, it is inherently difficult to determine which are the most important parameters. In contrast, Jones *et al.* (2006) have argued that choosing geologically realistic boundary conditions and input parameters for 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 over-thickened crust (such as the Tibetan Plateau) towards the topographic surface at the plateau margins remains controversial, however, both with respect to the Himalaya–Tibet 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.

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