Channel flow and continental collision tectonics: an overview

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Abstract: The principle of channel flow as defined in fluid dynamics has been used in continental geodynamics since the 1980s. The basic equations for one-dimensional flow introduced to geologists by Turcotte and Schubert were further developed by several research groups to meet the needs of specific studies. The most substantive differences among numerical models are results of different solutions for flow in crust, developed for different boundary conditions. The concept of channel flow has met with strong opposition and criticism from geophysicists and modellers. Although it is difficult to prove unambiguously that there is an active weak channel, it is still the most successful model to explain and predict the tectonics, metamorphism and exhumation of high-grade terranes in some orogens. Moreover, the concept of channel flow has stimulated novel approaches to the study of both the tectonics and metamorphism of large, hot orogens and the interaction between tectonic and surface processes.

The concept of channelized flow of a weak crustal layer has been applied to various tectonic settings (Godin *et al.* 2006): (a) asthenospheric counterflow; (b) lower crustal channels; (c) intra-crustal channels; (d) subduction channels; (e) salt tectonics. In this overview only the concept of intra-crustal channels in collision orogens will be discussed, while detailed discussion of the range of proposed geologically relevant solutions to Stoke's equation will be presented elsewhere. In active collision orogens the channel flow model has been used to explain the coupling between crust and mantle, strain in the crust, metamorphism, synorogenic exhumation of high-grade terranes and landscape evolution. Some numerical modelling strongly supports existence of a weak crustal layer (e.g. Royden 1996; Clark & Royden 2000; Beaumont *et al.* 2001*a*, *b*; 2004, 2006; Clark *et al.* 2005) while others oppose it (e.g. Toussaint & Burov 2004; Toussaint *et al.* 2004). Similarly, geophysical data can be interpreted both in favour of an active lower- or mid-crustal channel (Nelson *et al.* 1996; Klemperer 2006) or against it (e.g. Flesch *et al.* 2005; Hilley *et al.* 2005). In the models the details of channel flow are sensitive to the flow laws assumed to operate in the crust, the mechanical properties of bounding crustal layers and density distribution in the crust. A number of analytical solutions to Stoke's equation have been derived and experiments were developed for fundamentally different boundary conditions: conclusions will thus be model-dependent.

The Himalaya–Tibet orogen is the most studied putative example of both active- and palaeo-channel flow. The Greater Himalayan Sequence (GHS), also referred to as Higher Himalayan Crystallines, is a sequence of amphibolite- to granulite-facies ortho- and paragneisses, migmatites and syntectonic leucogranites. The GHS is both underlain and overlain by greenschist and lower grade metasediments and forms the metamorphic core of the Himalaya (e.g. Hodges 2000, 2006; Jamieson *et al.* 2004, 2006). The GHS is bounded by the Main Central thrust (MCT) at the base and by the South Tibetan detachment (STD) at the top. The two shear zones are subparallel, have opposite senses of shear and have operated coevally over an extended period of time (Hodges 2000; Godin *et al.* 2006). The change from burial to exhumation of the GHS is reflected in the change in shear sense along the upper bounding shear zone. The earlier phase of deformation is dominated by thrust-sense shear zones while the return channel flow and ductile extrusion are characterized by normal-sense shear along the roof of the exhumed channel. The presence of melt during both south- and north-directed shearing at the base and top of the GHS respectively (Fig. 1; e.g. Grujic *et al.* 1996, 2002; Davidson *et al.* 1997; Daniel *et al.* 2003; Harris *et al.* 2004) shows that high-temperature metamorphism and anatexis were an integral part of the exhumation process of the GHS (Godin *et al.* 2006; Hollister & Grujic 2006).

Two closely related terms, ductile extrusion and channel flow, are used in the literature in regard to tectonics of a weak crustal layer. Kinematic models for both are based on the concept of a pair of coeval subparallel dip-slip shear zones with opposite senses of shear bounding a crustal layer with significantly lower strength or viscosity than the bounding layers. The concept of channel flow is based on the physical laws of fluid dynamics with a particular set of boundary conditions. Ductile extrusion is more difficult to define

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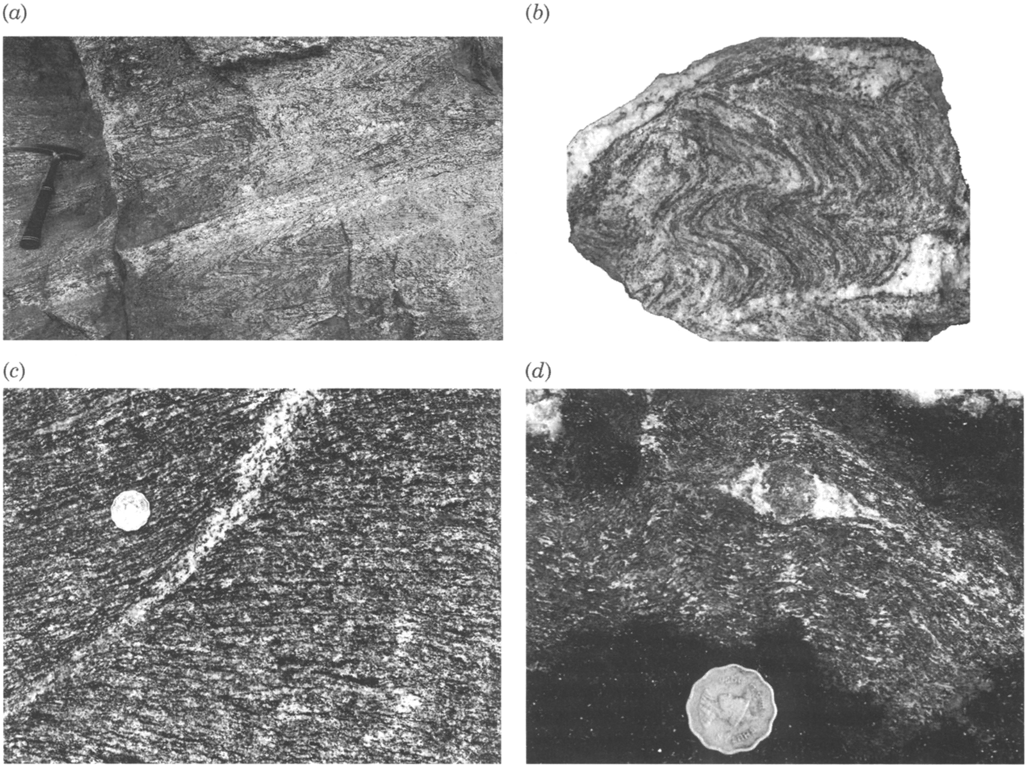

Fig. 1. Outcrop photographs of deformed migmatites from the metamorphic core of the Himalaya in central Bhutan (i.e. the Greater Himalayan Sequence; e.g. Grujic *et al.* 1996), suggesting an extended period of coeval partial melting and deformation. (**a**) A finely foliated migmatite has been folded into chevron-type folds. The axial planes are associated with a new generation of little-deformed, yet sheared leucosome (for discussion see Davidson *et al.* 1997). The shear planes are parallel to the dominant foliation associated with the top-to-the south thrusting. View to the east. (**b**) Folded stromatic migmatite with leucosome intruding along axial surfaces. Sample is *c.* 6 cm long. (**c**) Normal-sense, top-to-the north shear band with leucosome in the core of the shear band. (**d**) Leucosome in the pressure shadows around a garnet.

precisely and uniquely and will probably be of different forms in different orogens.

Ductile extrusion

Two types of extrusion along dip-slip fault or shear zones have been proposed in the literature. First is the extrusion of a rigid crustal sliver or wedge between discrete faults (e.g. Chemenda *et al.* 1995; Northrup 1996). This model is probably equivalent to the 'expulsion' model of Hodges (1998, his fig. 3). The second type is ductile extrusion of a deformable crustal layer between ductile shear zones (e.g. ductile extrusion of the metamorphic core of the Himalaya; Grujic *et al.* 1996; Grasemann & Vannay 1999; Grasemann *et al.* 1999). A similar process has been described in a number of other orogens, e.g. the Norwegian Caledonides (Northrup 1996), the Monashe Complex of the Canadian Rockies (Gibson *et al.* 1999; Williams & Jiang 2005) and the Hellenides (Xypolias & Koukouvelas 2001; Xypolias & Kokkalas 2006). In the former type strain is concentrated along the boundaries and there is little or no deformation in the core. In the latter, the deformation is pervasively distributed throughout the extruding wedge. In the case of pervasive deformation the strain may be simple shear with strain gradient toward the boundaries, or different degree of general shear (combination of simple and pure shear; e.g. Law *et al.* 2004; Jessup *et al.* 2006). Even deformation of the wedge by pure shear only would require simple shear deformation along the boundaries (see also Williams *et al.* (2006) for an argument against wedge extrusion). When the extruding wedge is composed of materials with yield strength, the deformation of the wedge is probably best represented by the Prandtl cell model (Price 1972).

Lateral extrusion of rigid crustal blocks between steep strike-slip faults as in SE Asia (e.g. Tapponnier *et al.* 1982, 2001) or the Eastern Alps (Ratschbacher *et al.* 1991) will not be discussed here. The driving forces of extrusion may be buoyancy forces (caused by different crustal densities), tectonic overpressure (caused by non-parallel boundaries of a weak crustal layer), pressure differences caused by varying crustal thickness (i.e. topography) or the combination of any of these. Both ductile extrusion and extrusion of a rigid wedge are coupled with surface erosion (e.g. Chemenda *et al.* 1995; Grujic *et al.* 1996; Hodges 1998).

Extrusion may thus be regarded as an exhumation process. It is evident that the volume of the extruding wedge limits the extrusion process. If ductile extrusion is linked to a crustal channel, the viscous part of the channel that cools below a critical temperature is converted rheologically into an extruding wedge: extended flow of a crustal channel provides replenishment of the extruding wedge. In this context, extrusion may mean the pumping or forcing out of material at the end of the viscous channel. The extrusion zone is, then, the part of the channel near the open end through which channel material is exhumed toward the surface (Beaumont *et al.* 2004, 2006). This means that in active orogens the exhumed metamorphic core may represent a fossil channel while the active channel is in the interior of the orogen at mid- to lower crustal level and beneath the related continental plateau. An important point is that the margins of the cell/wedge now exposed at the topographic surface may be more related to the processes of late stage extrusion/exhumation, rather than original channel flow within the more distal parts of the orogenic system. The classic example here may be the Greater Himalayan Sequence of the Himalaya (Hodges 2000; Grujic *et al.* 2002; Searle *et al.* 2003, 2006; Searle & Szulc 2005; Godin *et al.* 2006) and the putative active channel beneath the south Tibetan Plateau (Nelson *et al.* 1996; Hodges *et al.* 2001; Klemperer 2006).

Channel flow

Channel flow is a process in which viscous fluid flows through a channel lying between two rigid sheets that deform the viscous fluid between them through induced shear stress and pressure gradients within the fluid channel (e.g. Batchelor 2000; Turcotte & Schubert 2002). In geologically relevant analyses, boundary conditions that lead to channel flow solutions are diverse. In most cases the crustal plates bounding a weak crustal layer are rigid (i.e. the horizontal component of velocity is zero) but deformable (i.e. both the upper and lower surface of the layer can independently move vertically). The bounding plates may be also ductile and allowed to fail. The horizontal velocity profile in the channel depends on the geometry of the channel, but the simplest qualitative characteristic of these models in the case of rigid and undeformable bounding plates is that the one-dimensional (1D) velocity field is a hybrid between two end-member components: (1) Couette flow where the induced shear across the channel produces a uniform vorticity across the channel (Fig. 2a); and (2) Poiseuille flow (also known as the 'pipe-flow' effect) in which the induced pressure gradient produces highest velocities in the centre and opposite vorticity for the top and bottom of the channel (Fig. 2b). In the first case the bounding plates move parallel to each other; in the second case they are stationary. In order to avoid misunderstandings it is recommended that these end-members be referred to by their names and the term 'channel flow' be reserved for hybrid flow with specified boundary conditions. For the general channel flow the velocity profile will be a hybrid of the two end-members (Fig. 2c). The channel flow takes place in an infinitely long and wide channel (therefore a layer) whose lower boundary moves with velocity u_0 relative to its stationary upper boundary. The velocity profile of the Couette flow is a function of the displacement rate between the two bounding plates. For the simple Couette flow we assume that the velocity of the lower plate is $u_0 \neq 0$ and the velocity of the upper plate is $u = 0$. Non-slip boundary conditions apply; a viscous fluid in contact with a solid boundary must have the same velocity as the boundary. If the applied pressure gradient is zero, $p_1 = p_2$ or $\mathrm{d}p/\mathrm{d}x = 0$, the solution is a linear velocity profile:

$$u = u_0\left(1 - \frac{z}{h_{ch}}\right) \qquad (1)$$

If there is no relative displacement between the bounding plates, i.e. $u_0 = 0$, the velocity of Poiseuille flow is governed by the pressure gradient $\mathrm{d}p/\mathrm{d}x$, viscosity μ_{ch}, depth z, and channel thickness h_{ch}:

$$u = \frac{1}{2\mu_{ch}}\frac{\mathrm{d}p}{\mathrm{d}x}(z^2 - h_{ch}z) \qquad (2)$$

(Turcotte & Schubert 2002, equation 6–14). Pressure in the crust may be specified in three ways with increasing level of approximation: (a) real pressure (estimations of it are not obvious; however, some analyses of crustal flow use this approach, e.g. McKenzie *et al.* 2000; McKenzie & Jackson 2002); (b) lithostatic pressure in crust with density

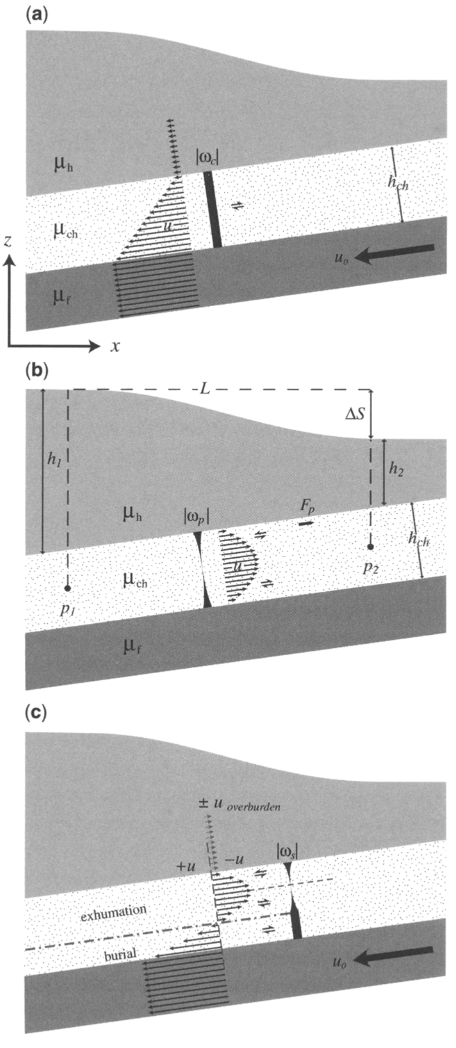

Fig. 2. Idealized picture of channel flow. Here the bounding plates of a weak viscous channel are depicted as rigid and undeformable; in nature the boundaries of a channel show smooth gradients in mechanical properties and both the weak channel and bounding crustal levels above and below are deforming. (**a**) Couette flow component. (**b**) Poiseuille flow component. (**c**) Return channel flow for a particular set of parameters. Symbols are discussed in the text. The vorticity ω values (rotational component of the flow profile) are schematically indicated by the width of the black bar, with highest vorticities (simple shear dominant) indicated by a wide bar segment (see also Jessup *et al.* 2006). Only the absolute value of the vorticity ω is indicated regardless of whether it is positive (sinistral simple shear) or negative (dextral simple shear) (after England & Holland 1979; Mancktelow 1995; Grujic *et al.* 2002). ω_c: vorticity in pure Couette flow. ω_p: vorticity in pure Poiseuille flow. ω_s: vorticity in a hybrid channel flow.

variations (e.g. Bertotti *et al.* 2000; Lehner 2000); (c) lithostatic pressure in crust with uniform density. This third level approximation is the simplest case and is assumed in the following discussions about channel flow in the crust.

The flux U of material in the channel can be expressed by integrating the velocity u of the material over the channel thickness:

$$U = \int_0^{h_{ch}} u(z)\,\mathrm{d}z \tag{3}$$

Applied to Equation 2 this leads to the flux U given by:

$$U = \frac{h_{ch}^3}{12\mu_{ch}}\frac{\mathrm{d}p}{\mathrm{d}x} \tag{4}$$

(Turcotte & Schubert 2002). For a linear viscous fluid Turcotte & Schubert (2002, equation 6–17) give the following equation for the mean velocity of flow (equal to flux/thickness) in a parallel-sided channel:

$$\bar{u} = -\frac{h_{ch}^2}{12\mu_{ch}}\frac{\mathrm{d}p}{\mathrm{d}x} + \frac{u_0}{2} \tag{5}$$

Equation 5 accounts for both Couette flow and Poiseuille flow (for the latter if $u_0 = 0$). In a crust with uniform densities the principal factors that influence channel flow are therefore the relative velocity of the bounding plates, the thickness of the channel, the viscosity of the channel material and the pressure gradient along the channel. Which of the end-members will be dominant in the hybrid velocity profile is a function of threshold value for a particular parameter (e.g. England & Holland 1979; Mancktelow 1995). For a constant plate convergence rate, pressure gradient and channel thickness, lowering of channel viscosity below a threshold value will shift the uniform Couette flow to heterogeneous flow where one part of the channel will be dominated by Poiseuille flow (Fig. 2c). A similar effect may be achieved with constant viscosity and channel thickness by slowing the plate convergence rate, or increasing the pressure difference (e.g. surface relief).

The pressure gradient or pressure difference along the channel may be provided by non-parallel channel walls as a tectonic overpressure (e.g. Mancktelow 1995) or by a different lithostatic load (which in most active orogens is related to topography). It is important to note here that in the case of a non-horizontal channel, when pressures are close to lithostatic and crustal densities are uniform (e.g. in models by Beaumont *et al.* 2004; Jamieson *et al.* 2004), the pressure gradient is approximately proportional to variations in the burial depth of the channel (e.g. Gemmer *et al.* 2004); this does not correspond to topography (i.e. surface relief) Δh because the channel is not horizontal (for comparison see Hodges *et al.* 2001). However, in an inclined channel beneath flat topography and without other driving forces (i.e. tectonic overpressure or buoyancy in the crust with density variations) there will be no flow (S. Medvedev, pers. comm. 2005). The pressure gradient can be approximated as the difference in pressure beneath thick and thin crust. Considering Airy compensation the lateral pressure gradient can be written as:

$$\frac{\mathrm{d}p}{\mathrm{d}x} \approx \Phi\frac{\rho_c g \Delta h}{L/2} \qquad (6)$$

(after Kruse *et al.* 1991), where Φ is the isostatic amplification factor $(\rho_m - \rho_c)/\rho_m$ (Turcotte & Schubert 2002), ρ_m is the density of the mantle, ρ_c the uniform density of crust, g is the acceleration due to gravity, Δh are the variations in crustal thickness, and L is the horizontal length scale of transport (characteristic horizontal scale of topographic perturbations). This equation shows that the rates of flow depend on the characteristic horizontal length of pressure distribution. The velocity in the viscous layer subject to variations of lithostatic pressure is given by Gemmer *et al.* (2004) as:

$$u_p = -\frac{\rho_c g}{2\mu_c}\frac{\partial h(x)}{\partial x} z(h_{ch} - z) \qquad (7)$$

The pressure gradient across a mountain range can increase due to build-up of surface topography, i.e. evolution of the continental plateaus (e.g. Medvedev & Beaumont 2006). Assuming lithostatic pressure, constant viscosity of the channel and uniform density of the crust, the material flux (Equation 4) becomes:

$$U = -\frac{\rho_c g h_{ch}^3}{12\mu_{ch}}\frac{\partial S}{\partial x} \qquad (8)$$

(Medvedev & Beaumont 2006) where S is the topographic surface elevation. Using Airy isostatic equilibrium the topographic elevation S over time, as a result of flux of crustal material in the lower crust, can be written as:

$$\frac{\mathrm{d}S}{\mathrm{d}t} = \Phi\left[\frac{1}{12}\rho_c g h_{ch}^3 \frac{\mathrm{d}}{\mathrm{d}x}\left(\frac{1}{\mu_{ch}}\frac{\mathrm{d}S}{\mathrm{d}x}\right)\right] \qquad (9)$$

(after Clark & Royden 2000). In other words a weak crustal channel will influence the topographic wavelength and in the same time a topographic gradient will influence the rate of flow in the channel. Using this approach and suggesting a viscosity of 10^{18} Pa s for the lower crust, Clark & Royden (2000) calculated that topography could be built by a constant flux of material into a crustal channel from beneath the thick part of the plateau. This may explain the eastward propagation of the Tibetan Plateau and the slope distribution along the eastern Plateau margin as a function of channel thickness and/or foreland crustal structure. Medvedev & Beaumont (2006) solved Equation 9 analytically for a constant boundary flux, $U =$ cons, for the evolution of upper surface with time:

$$S(x,t) = U\left[\sqrt{\frac{\alpha t}{\pi}}\exp\left(-\frac{x^2}{\alpha t}\right) - x\,\mathrm{erfc}\left(\frac{x}{\sqrt{\alpha t}}\right)\right] \qquad (10)$$

where $S(x, t = 0) = 0$ is the initial position of the surface, $\alpha = (\Phi\rho_c g h_{ch}^3)/3\mu_{ch}$, erfc is the complementary error function (Turcotte & Schubert 2002), and t is time. Testing for range of model parameters values, Medvedev & Beaumont (2006) propose three end-members for growth of continental plateaus in the presence of a weak crustal layer. However, the existence of a large-amplitude, short-wavelength topographic relief along the eastern margin of Tibet led McKenzie & Jackson (2002) to argue (assuming different boundary conditions from Clark & Royden 2000) that it is unlikely that extensive lower crustal flow has occurred there because of a short response time at short topographic wavelengths. Alternatively, Zhong (1997) used Maxwell rheology and assumed a minimum viscosity for the lower crust of 10^{21} Pa s to show that short-wavelength topography could remain uncompensated for extensive periods of time. Most recent landscape analyses of the Tibetan Plateau have indicated the existence of areas of anomalously high topography (Burchfiel 2004; Clark *et al.* 2005). At least along the eastern plateau margin, around the Sichuan basin, such high topography could be interpreted as dynamic (Clark *et al.* 2005). The authors apply the geometry of a thin viscous channel with a rigid cylindrical obstacle (simulating the crust of the Sichuan basin) perpendicular to the bounding plates that causes dynamic pressure to the base of the overlying elastic crust. Relative to the background channel flow, such a dynamic loading can cause surface uplift upstream, and subsidence downstream of the rigid obstacle. If this concept is correct, then dynamic topography may explain the short-wavelength topography observed along the eastern Tibetan margin, despite predictions that the short response time would prevent such landscape development in the case of an active crustal channel (McKenzie & Jackson 2002).

The shape of the velocity profile for Poiseuille and Couette flow is a function of the rheology of the channel material and of the temperature distribution across the channel, respectively. The velocity profile of channel flow of viscous fluids features a velocity gradient that decreases towards the centre of the channel. Hence the shear stress σ_s transmitted by a fluid layer also decreases toward the channel centre. Turcotte & Schubert (2002, equation 6–8) show the shear stress gradient in the channel.

$$\frac{d\sigma_s}{dz} = \frac{dp}{dx} \tag{11}$$

For a linear viscous fluid the Poiseuille flow velocity profile is a parabola that is symmetric about the centreline of the channel (Fig. 3a). In a power-law fluid, however, for increasing values of n (the exponent of power-law creep) the gradients in the velocity become large near the walls where the shear stress is a maximum (Fig. 3a). As a consequence, a nearly rigid core flow develops where the shear stress is low. A similar velocity profile is present in Bingham fluids (viscous–plastic materials that are characterized by a yield stress). As Bingham fluids become solid when the applied shear stress falls below the yield stress, the deforming material may become solid in the centre of the channel. There a solid 'plug' will be moving within the flow. The effect of the power-law exponent is especially clear for the change between $n = 1$ and $n = 2$, (Fig. 3a). However, the question remains whether it is actually possible to distinguish between linear and power-law flows from the finite strain data in nature (C. Beaumont, pers. comm. 2004).

The effective viscosity for channel flow of a power-law fluid is given by Turcotte & Schubert (2002, equation 7–127):

$$\mu_{eff} = \frac{\sigma_s}{du/dz} = \left(\frac{p_1 - p_2}{L}\right)\frac{h_{ch}^2}{4(n+2)\bar{u}}\left(\frac{2z}{h_{ch}}\right)^{1-n} \tag{12}$$

By assuming a crustal structure, flow laws for crustal material, and the geometry of the putative viscous channel, the viscosity of the channel material can be estimated. Studies aimed at constraining physical parameters of the crust have yielded estimates of effective viscosity of the lower crust of 10^{17}–10^{20} Pa s for assumed channel thicknesses of 10–25 km (e.g. Kruse *et al.* 1991; Wdowinski & Axen 1992; Kaufman & Royden 1994; Clark & Royden 2000; Medvedev & Beaumont 2006). However, the values of involved parameters (Equations 10 and 12) cannot be estimated precisely. The viscosity of the channel

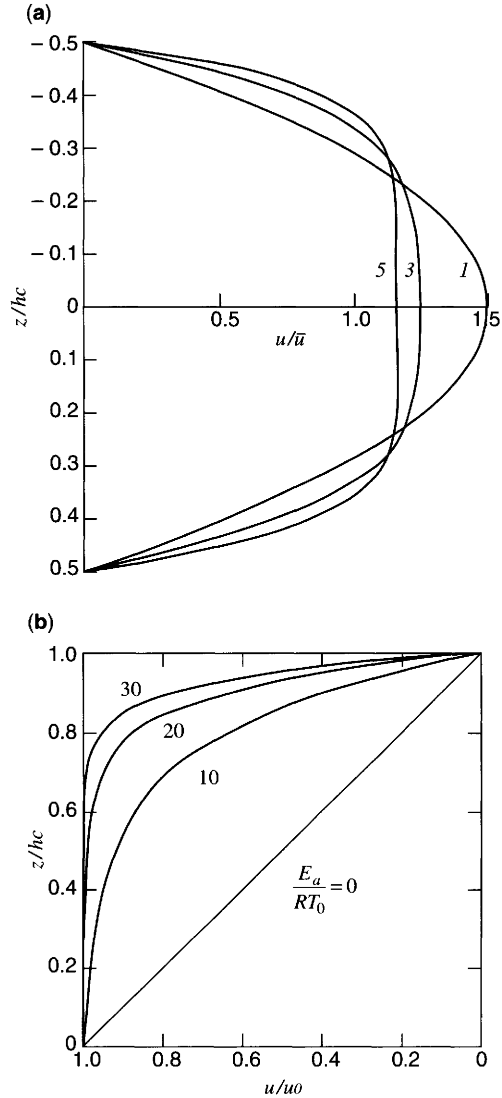

Fig. 3. Velocity profiles u/u_1 versus z/h in a channel flow with different rheologies. (**a**) Velocity profiles for Poiseuille flow showing the effect of power-law exponent: $n = 1$ (linear viscous), $n = 3$ and 5. Adapted from Turcotte & Schubert (2002, their fig. 7–13). The plug-flow appearance of the velocity profile is a consequence of the stress dependence of the effective viscosity. (**b**) Velocity profiles for Couette flow for different temperature dependences of viscosity (i.e. different values of the dimensionless activation energy parameter). Shear stress σ_s is a constant in the absence of a horizontal pressure gradient. The upper wall is maintained at temperature T_0 while the lower wall is kept at temperature T_1 ($T_1 > T_0$), assuming that the temperature difference $T_1 - T_0$ is small compared with T_0,. Adapted from Turcotte & Schubert (2002, their fig. 7–14).

can be estimated with an accuracy not better than two orders of magnitude, while h_{ch}^2 cannot be estimated with an accuracy better than one order of magnitude (S. Medvedev, pers. comm. 2005). The same effect can be produced by a thick channel with a high viscosity as by a thin channel with very low viscosity. Because the parameter α (Equation 10) scales as h_{ch}^3/μ, an uncertainty in h_{ch} of a factor slightly more than 2 causes an order of magnitude uncertainty in μ even when the flux U is known (Medvedev & Beaumont 2006). The results are also critically dependent on whether the channel thickness, flux and viscosity are constant or not (Clark & Royden 2000; Medvedev & Beaumont 2006). Moreover, if the channel absorbed all the deformation, the viscosity of the rest of the crust cannot be estimated.

The rheological structure, with Coulomb-plastic and power-law viscous regions, leads to a more complex flow structure in general. An important point about channel flow is that the Poiseuille flow produces shear traction on the base of the overburden resulting in the horizontal force F_p (Fig. 3b), which, in special cases, does not depend on viscosity:

$$F_p = \frac{h_{ch}}{2}\rho_c g(h_1 - h_2) \qquad (13)$$

(Fig. 2; Gemmer *et al.* 2004). This force, depending on the ratio between overburden thickness and channel thickness, and on overburden strength, may lead to failure of the overburden, i.e. failure of the upper crust (Beaumont *et al.* 2004; Gemmer *et al.* 2004). If the stresses are above the yield strength in Coulomb materials, near-surface extensional flow may occur giving rise to various structures produced by interaction between the viscous channel and failure of the overburden. For example 'plateau collapse' may be the effect of the channel flow, rather than of gravitational instability of a thickened crust. In numerical models (e.g. Beaumont *et al.* 2004, 2006) the flux and the channel thickness change and may lead to inflation or deflation of the channel where the flow rate changes along the channel. For example, the formation of North Himalayan Gneiss domes (Hodges 2000, 2006), or out-of-sequence thrusts within the GHS (Grujic *et al.* 2002, 2004) can be interpreted in terms of channel flow in an inflating channel (Beaumont *et al.* 2004).

The Couette flow velocity field is also affected by the temperature distribution in the channel which influences the viscosity with an exponential dependence on the inverse absolute temperature (Equation 14). When account is taken of heating by viscous dissipation in the shear flow, the temperature dependence of the viscosity couples the temperature $T(z)$ and velocity $u(z)$ profiles in the channel. Both quantities $T(z)$ and $u(z)$ must be determined simultaneously, since one depends on the other. The velocity u depends on T through the dependence of μ on T, and T depends on u because frictional heating depends on the amount of shear in the velocity profile. Turcotte & Schubert (2002, equation 7–248) give the equation for the effective viscosity μ_{eff} of a power-law material:

$$\mu_{eff} = \frac{1}{2C\sigma^{n-1}}\exp\left[\frac{E_a}{RT}\right] \qquad (14)$$

where C is the pre-exponential term of power-law creep, σ is the flow stress, E_a is the activation energy, R is the Boltzmann constant and T is the absolute temperature (K). When fluid viscosity is independent of temperature (the dimensionless activation energy parameter, $E_a/RT_0 = 0$), the velocity profile is linear (Fig. 3b). As the viscosity becomes increasingly temperature-dependent (larger values of E_a/RT_0), the shear is confined to progressively narrower regions where the fluid is hottest and the viscosity is smallest (Fig. 3b). Thus, frictional heating can have consequences on the shear flow of a fluid with a strongly temperature-dependent viscosity.

The transition from Couette flow- to Poiseuille flow-dominated channel flow is highly sensitive to viscosity and lateral pressure gradients. A critical change of viscosity within the channel will trigger changes in the flow pattern. For high viscosity material, Couette flow dominates and channel material will be underplated; for low viscosity material, Poiseuille flow will be dominant (e.g. Mancktelow 1995; Fig. 2c) and will cause flow opposite to the movement of the bounding plates, i.e. return flow of previously buried rock mass. Return flow in the channel is, therefore, a transient phenomenon that is recorded in the changing kinematic character of the structures within it (Fig. 4). The viscosity of the rock in the channel is likely to change significantly over short intervals of geological time (Beaumont *et al.* 2001*a*, *b*). A change in viscosity could be due to a metamorphic reaction leading, for example, to melting (e.g. Hollister 1993). Although certain details are disputed, it is evident that partial melting decreases the strength and the effective viscosity of rocks over several orders of magnitude (e.g. Cruden 1990). The volume of partial melt, not the temperature, controls the strength of rocks, and various values of critical melt percentage needed to dramatically reduce strength have been listed in the literature (Rosenberg & Handy 2005). Recent analyses suggest that in the range between solidus and a

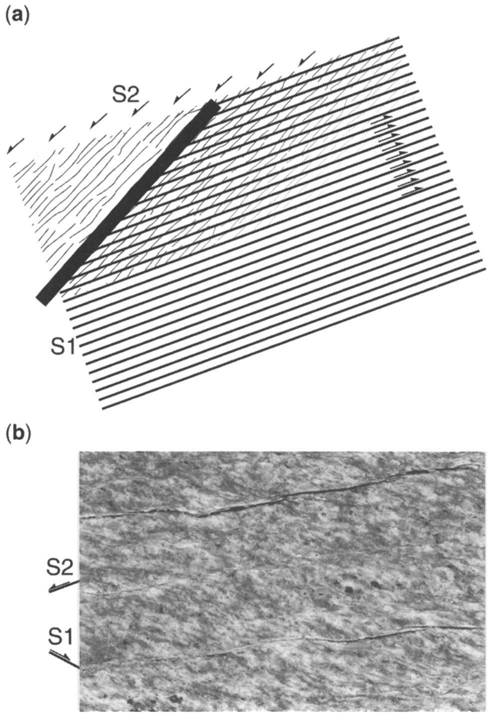

Fig. 4. (**a**) Schematic diagram of the deformation sequence along the upper boundary of a channel. S1, thrust shear zone formed by Couette flow (prograde metamorphism); S2, ductile normal shearing formed during return flow in the channel or during ductile extrusion (or both) at thermal peak of metamorphism and/or isothermal decompression. (**b**) Sample of deformed granite from the 2 km wide zone of the top-to-the north shear at the top of the GHS, NW Bhutan. Sample is *c.* 10 cm long.

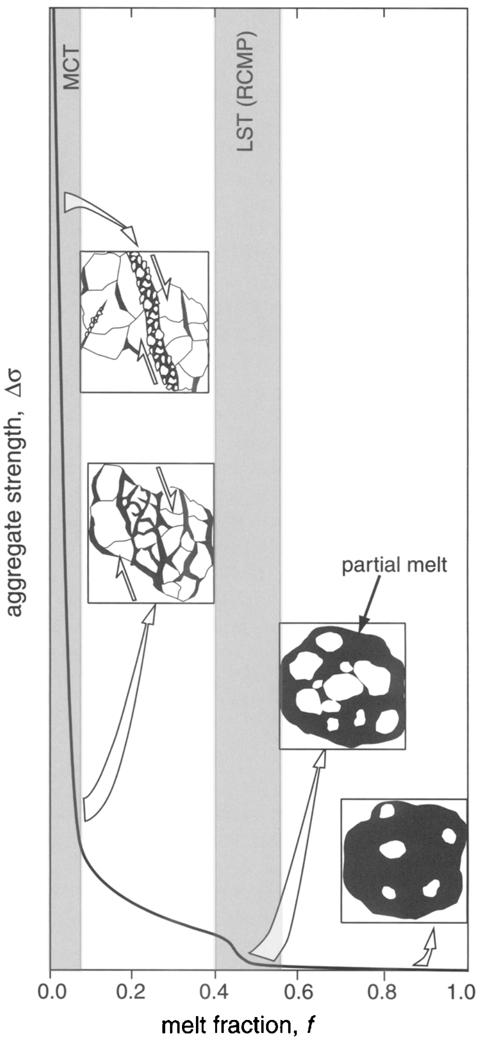

Fig. 5. Schematic plot of aggregate strength $\Delta\sigma$ versus melt fraction f for partially melted granite between the liquidus and solidus. The vertical scale of the lower part of the ordinate is exaggerated in order to make the LST (solid-to-liquid transition) visible. There are two steep segments of the strength curve corresponding to the MCT (melt connectivity transition) and to the LST, or RCMP (rheological critical melt percentage). Adapted from Rosenberg & Handy (2005).

7% melt fraction, the volumes of partial melt are sufficient to lower the strength by almost one order of magnitude (Fig. 5; Rosenberg & Handy 2005). Experiments on the mechanical behaviour of partially molten synthetic granite (Rutter *et al.* 2005) provide the most recent estimates of the rheological behaviour of partially molten rocks. The flow law for low strain data is of the form:

$$\dot{\varepsilon} = A \exp(Bf^m) \exp\left[\frac{-H}{RT}\right] \sigma^n \qquad (15)$$

where A and B are constants, m and n are exponents, H is the activation enthalpy, $\dot{\varepsilon}$ is the strain rate (s^{-1}), f is the melt fraction, σ is flow stress, and T is the temperature (Rutter *et al.* 2005). The extrapolation to geological strain rates using the above flow law suggests that in nature migmatites containing granitic melt will be extremely weak – much weaker than silicate rocks deforming by intra-crystalline plasticity (Rutter *et al.* 2005). The weakening of rocks with increasing melt fraction is neither linear ($m = 2$, Equation 15) nor gradual. As suggested by Rosenberg & Handy (2005), the mechanical response of a rock containing *c.* 8% melt may be very different from rock that contains

7% or less melt, but is not significantly different from rock that contains 50% melt. In rocks with a high melt percentage the stresses will be low and thus deformation-driven melt extraction processes hindered; weak rocks stay weak. This implies complex feedback mechanisms between rheological weakening by partial melting, melt extraction processes by deformation, and deformation localization. Moreover, the effective viscosity of the power-law fluid is proportional to σ_s^{1-n} (Turcotte & Schubert 2002). For large n the viscosity is high where shear stress is small and low where shear stress is large. The principle of channel flow is based on uniform distribution of stresses in the crust resulting in high strains in weak parts. However, weak parts in the centre of the channel have low stress, and consequently high viscosity if they deform by power-law rheology. Based on this principle there would be no channel flow. Rutter *et al.* (2005) propose that the stress exponent n for migmatites is 1.8 – smaller than for most geological materials (Twiss & Moores 2000; Turcotte & Schubert 2002; Pollard & Fletcher 2005). Accordingly, the effective viscosity of partially molten rock depends less on strain rate and stress but is more a function of temperature and pressure. In addition, near the walls where shear stress is high, the effective viscosity is low in a power-law fluid, and the velocity gradient is large (Fig. 3a). It follows that strain weakening caused by strain partitioning into shear zones is less important than hardening of the slower deforming material surrounded by shear zones (N. Mancktelow, pers. comm. 2005). Consequently, *in situ* melting is the most likely strength-lowering mechanism needed to initiate and sustain channel flow, although other weakening processes have also been suggested (Beaumont *et al.* 2004).

Discussion and conclusions

In summary, crustal extrusion can be defined as the forcing out of a crustal sliver or wedge above and below brittle or ductile shear zones, maintained by concurrent crustal driving forces and focused surface erosion. Channel flow is a lateral flow of a weak crustal layer between stronger bounding layers. In active orogens the viscous flow is driven by shear stresses induced by relative motion between moving bounding plates and by a pressure difference along the channel. In collisional orogens we may consider the upper plate stationary and the lower one underplating or subducting. However, the upper plate may fail and its extensional movement be influenced by the viscous channel. The movement in the upper plate may in turn influence the flow in the viscous channel. The pressure gradient in a crust with uniform density is influenced by the variations in crustal thickness and therefore surface relief. In addition, non-parallel channel walls may cause tectonic overpressuring, and the channel material may have a lower density than the surrounding crust and thus buoyancy forces may drive the flow as well.

The boundaries of a channel are more likely to be transitional than sharp. At depth the rheology of a channel changes as a function of temperature, pressure, lithology and melt content. Based on different numerical models of channels, and on field observations, at least three types of bounding conditions for a channel can be proposed: (a) rheological boundary – such a boundary may be due to lithologic contacts (e.g. salt–sediment boundary; Gemmer *et al.* 2004) or rheologic stratification within the crust (Lobkovsky & Kerchman 1991); (b) phase transitions – the most likely process is partial melting as briefly described here and assumed in some numerical models; (c) thermal boundary, e.g. exponential thermal weakening of crustal material (Bird 1991). In nature the first type of channel boundary may be marked by the sharpest gradient in mechanical properties, while the third type would be characterized by the smoothest transition. Yet the idealized thermal boundary condition is widely used in modelling (e.g. Clark & Royden 2000). The first two types of channel boundary may be mapped in the field as protolith boundaries (e.g. Davidson *et al.* 1997) while the thermal boundary may be mapped as a metamorphic isograd (e.g. Jamieson *et al.* 2004). If all these boundaries are present in an orogen they probably have parallel strikes but are differently inclined, and very likely located at different crustal levels. In the Himalaya the presence of very close, nearly parallel shear zones at both MCT and STD level (e.g. Searle & Godin 2003; Searle *et al.* 2003, 2006; Godin *et al.* 2006; Hollister & Grujic 2006) may indeed reflect multiple closely spaced boundary conditions.

Changing boundary conditions allow for changes in the channel flow regime, along the channel and with time, and do not restrict the definition of channel flow to idealized 1D flows of the type discussed by Turcotte & Schubert (2002). The velocity profile of channel flow is controlled by the balance between the shear stress gradient across the channel and the pressure gradient along the channel, while the exhumation history of the channel is controlled by the pattern of surface erosion (e.g. Jamieson *et al.* 2002; Beaumont *et al.* 2004). In an active orogen both channel viscosity and surface relief or crustal thickness (i.e. pressure gradient) are likely to change with time. In addition, numerical modelling allowing deformable and ductile bounding plates (e.g. Beaumont *et al.* 2004) suggests that the channel thickness is not constant, leading to much more complex tectonics. Further complication may arise due to the coupling between tectonics and surface processes. The results of thermal–mechanical models by Beaumont and co--workers (Beaumont *et al.* 2001*b*, 2004) suggest that surface denudation and strength of the upper crust may have a strong influence on flow in low-viscosity regions of the crust. Strong coupling between a crustal channel and topography indicates that concepts of coupling between surface processes (i.e. denudation) and tectonics that only take isostatic compensation into account may provide misleading results if there is an active channel.

Potential diagnostic techniques for determining the existence of active channelized crustal flow include geophysical and landscape analyses. Recent geophysical data indicate a weak crustal layer and the presence of fluids, perhaps melt, underneath southern Tibet (e.g. Nelson *et al.* 1996) and suggest channel flow may be active there (e.g. Klemperer 2006). However, some interpretations of GPS (e.g. Flesch *et al.* 2005) and seismic data (assuming that the mid- to lower crust deforms as a Maxwell visco-elastic rheology, e.g. Hilley *et al.* 2005) exclude the possibility of decoupling between mantle and crust in part of the Tibetan crust, and thus argue against a weak crustal layer there. Landscape analyses suggest that the topography of the Tibetan Plateau, its eastward propagation and steep southern boundary (i.e. the Himalaya) may be explained by the existence of a lower- or mid-crustal channel flow (e.g. Fielding *et al.* 1994; Clark & Royden 2000; Hodges *et al.* 2001; Shen *et al.* 2001; McKenzie & Jackson 2002; Clark *et al.* 2005). However, if different boundary conditions are assumed, the same landscape features may argue against a weak crustal channel (e.g. Zhong 1997).

In the case where a channel has been exhumed, geological criteria apply. (a) High grade rocks are bounded by subparallel, coeval ductile shear zones with opposite sense of shear. (b) The shear zone at the roof shows kinematic inversion. (c) In the estimated initial orientation, the dominant deformational/metamorphic foliation was subhorizontal (e.g. Williams & Jiang 2005). (d) Palaeo-channels should have unique structural and metamorphic histories; structures alone are not sufficient, structural histories are important. (e) The metamorphic histories of rocks should progressively change across the channel and show jumps across the boundaries of a channel (e.g. Jamieson *et al.* 2004, 2006). (f) The channel is represented by high-grade rocks, with syntectonic partial melt suggesting that during deformation these rocks were weaker relative to both under- and overlying crustal layers. (g) The high-grade rocks were extruded between the two bounding shear zones from beneath a coeval plateau. (h) There is strong orographic, focused erosion along the metamorphic belt interpreted as the palaeo-channel.

High-grade quartzo-feldspathic rocks and migmatites may thus be the most likely lithology in an exhumed crustal channel, and melt weakening is the most likely viscosity-lowering mechanism needed to promote channel flow. An outstanding question regarding channel dynamics therefore relates to the melting of rocks due to the pressure distribution in a channel (e.g. Harris *et al.* 2004), and encompasses processes such as partial melt segregation, magma coalescence, migration and emplacement. Melt weakening, however, is difficult to implement in numerical models (e.g. Beaumont *et al.* 2004) and at the moment there are no analytical models able to predict the style of deformation involving melt weakening and the development of channel flow.

In nature the boundaries of a channel show smooth gradients in mechanical properties and both the weak channel and bounding crustal layers above and below are deforming; the deformation is pervasively distributed throughout the channel but is concentrated in the weakest parts of the deforming crust causing channelized flows. Both channel flow and extrusion are influenced by surface denudation, which is in turn influenced by these two tectonics processes because of their control on surface uplift and surface slope. Therefore, the most important questions are what is a crustal channel in nature, and which boundary conditions produce intra-crustal channel flow?

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