Ductile extrusion in continental collision zones: ambiguities in the definition of channel flow and its identification in ancient orogens

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Abstract: Field characteristics of crustal extrusion zones include: high-grade metamorphism flanked by lower-grade rocks; broadly coeval flanking shear zones with opposing senses of shear; early ductile fabrics successively overprinted by semi-brittle and brittle structures; and localization of strain to give a more extensive deformation history within the extrusion zone relative to the flanking regions. Crustal extrusion, involving a combination of pure and simple shear, is a regular consequence of bulk orogenic thickening and contraction during continental collision. Extrusion can occur in response to different tectonic settings, and need not necessarily imply a driving force linked to mid-crustal channel flow. In most situations, field criteria alone are unlikely to be sufficient to determine the driving causes of extrusion. This is illustrated with examples from the Nanga Parbat–Haramosh Massif in the Pakistan Himalaya, and the Wing Pond Shear Zone in Newfoundland.

In this paper we review channel flow and crustal extrusion in relation to field geology. Our aim is to provide a critical evaluation of whether the predictions of channel flow models are testable using field observations, and also to consider the relevance of channel flow in relation to existing models of crustal extrusion. We illustrate the inherent difficulty and ambiguity of applying the channel flow concept with field examples from both active and ancient orogens (the Nanga Parbat–Haramosh Massif, Pakistan Himalaya, and the Wing Pond Shear Zone in the Appalachian orogen, Newfoundland). General aspects of orogenic thickening and crustal extrusion are presented first, followed by more specific discussion of channel flow and the nature and limitations of possible diagnostic characteristics.

Orogenic thickening and crustal extrusion

Field studies in orogenic belts have long shown that deformation arising from continental collision generally involves a combination of large-scale overthrusting and bulk crustal thickening. This combination is sometimes considered in terms of the relative importance of simple shear and pure shear to the mountain-building process, although these are end-member strain components and most parts of an orogen will experience more general shear (e.g. Sanderson 1982; Platt & Behrmann 1986; Passchier 1986, 1987; subsimple shear of Simpson & De Paor 1993). The variety and complexity of structures typically observed in orogenic belts illustrate that the distribution and localization of strain is highly heterogeneous (e.g. Jones et al. 2005). Overthrusting is associated with zones of high strain concentration and high vorticity that give the orogen marked polarity (i.e. asymmetry), recognizable in outcrop as dominantly foreland-verging structures, including shear zones with top-to-foreland shear sense. Regions of bulk crustal thickening represent a more widely distributed response to horizontal shortening across the orogenic belt, and show large variations in width and in localization of strain. The boundaries to these
regions can be well-defined shear zones or diffuse transitions with gradually varying style and intensity of strain. Some areas of large-scale crustal thickening can involve broad zones characterized by intense folding and widespread formation of pervasive tectonic fabrics. Crustal shortening can also be localized into narrow regions of more intense deformation (often marked by well-defined high strain boundaries), in which shortening is matched by thickening occurring as concentrated zones of crustal extrusion. Irrespective of whether strain is widely distributed across a broad area, or localized into a more concentrated zone of extrusion, diagnostic characteristics of crustal extrusion include the uplift of high-metamorphic-grade rocks in the core of the zone (e.g. the Kaoko Belt in Namibia; Goscombe et al. 2003a, b, 2005), and the reversal in the sense of vorticity across the region of deformation (e.g. southwestern Hellenides; Xypolias & Doutsos 2000; Xypolias & Koukouvelas 2001; Xypolias & Kokkalas 2006).

Numerical modelling of crustal extrusion

Regions of crustal extrusion bounded by coeval, opposing shear zones are a characteristic feature of numerical models of deformation that involve combinations of coaxial and non-coaxial strain (e.g. Sanderson 1982; Sanderson & Marchini 1984). Although strain matrix modelling is naïvely simplistic in approach, and explores the kinematics of deformation without any thermal–mechanical considerations, it is computationally efficient, and has the advantage that it can be used to model first-order relationships between bulk finite strain and boundary conditions. In addition, it also helps to demonstrate the innate complexity of progressive deformation in four dimensions (4D) (i.e. one temporal and three spatial dimensions). Strain matrix modelling has been applied extensively to improve understanding of various non-coaxial, non-plane strains that typify zones of transpressional and transtensional deformation (e.g. Fig. 1; Tikoff & Fossen 1993; Jones et al. 1997, 2004; Fossen & Tikoff 1998; De Paola et al. 2005). The significance of transpression is that deformation resulting from oblique plate collision cannot be accommodated by simple shear alone, and must involve a significant component of crustal extrusion, so numerical modelling of transpression can provide insight into the interplay between orogenic shortening, overthrusting and crustal thickening.

Application of transpressional strain modelling to areas of oblique convergence and divergence has shown close correlation between model predictions and observed aspects of crustal extrusion (e.g. Robin & Cruden 1994; Teyssier et al. 1995; Jones & Tanner 1995; Holdsworth et al. 1998, and references therein). In the case of transpression (Fig. 1c, d), upward extrusion of crustal material is caused by shortening across the deformation zone, and is thus equally applicable to plane-strain coaxial shortening during orthogonal collision (Fig. 1a, b). The thermal implications of crustal thickening and extrusion have been explored for both orthogonal and oblique shortening by Thompson et al. (1997a, b). Thermal–mechanical models (Ellis et al. 1998), representing an extra level of geological realism compared with simplistic strain modelling, reproduce similar combinations of large-scale overthrusting and broad zones of bulk crustal extrusion resulting from continental collision.

In general, strain modelling usually assumes that crustal extrusion occurs by upward migration, deformation and erosion of the Earth’s surface, although it is also possible for crustal material to be extruded laterally (i.e. axially, parallel to the orogenic trend; Dias & Ribeiro 1994; Jones et al. 1997; Dewey et al. 1998). Note that extrusion in this sense refers to the lateral escape of crustal blocks (e.g. Eastern Anatolia; Dewey et al. 1986). Lateral extrusion is also used in a slightly different sense to describe outward ductile flow at lower crustal levels, driven by lateral pressure gradients (e.g. Bird 1991; Clark & Royden 2000).

![Fig. 1. First-order kinematics of tectonically driven crustal extrusion caused by boundary displacement: (a) vertical extrusion, plane strain; (b) inclined extrusion, plane strain; (c) vertical extrusion, non-coaxial non-plane strain (transpression); (d) inclined extrusion, non-coaxial non-plane strain (inclined transpression).](image-url)
Evidence for crustal extrusion in central and eastern Himalayas

Recognition of north-dipping normal faulting over large distances along-strike in Tibet (Burg et al. 1984), led Burchfiel & Royden (1985) to suggest that the Greater Himalayan Sequence (GHS) has been extruded relative to Tibet and the Lesser Himalaya. Subsequent work has recently been extensively discussed and summarized by Searle et al. (2003), so only an overview of field-based evidence for crustal extrusion is given here. The possible relationship between crustal extrusion of the GHS and mid-crustal channel flow (together with evidence from regional geophysics) is discussed later.

The GHS is basally bounded by the Main Central Thrust (MCT), and separated from the overlying Tibetan crust by the South Tibetan Detachment (STD) system (Gansser 1964; Burg et al. 1984). Structural descriptions of normal displacement on the STD are given by Burchfiel et al. (1992), Edwards et al. (1996), Gruijic et al. (1996, 2002) and Law et al. (2004). Evidence for contemporaneity of movement on the MCT and STD between c. 22 and 17 Ma is documented by Hodges et al. (1992, 1993) and Walker et al. (1999). Detailed kinematic analyses from the MCT (Grasemann et al. 1999) and STD (Law et al. 2004; see also Carosi et al. 2006; Jessup et al. 2006) suggest that finite strain in the GHS is a subsimple shear (i.e. a combination of pure shear flattening and simple shear). Higher metamorphic grades are recorded from central parts of the GHS, with inverted pressure–temperature profiles across the MCT (e.g. Hodges et al. 1988, 1993; Searle & Rex 1989; Vannay & Grasemann 2001). Crustal-melt Miocene leucogranites were emplaced between the MCT and STD (e.g. Searle et al. 1993; Searle 1999; Edwards et al. 1996; Wu et al. 1998). Notwithstanding ongoing debate regarding some of the field evidence (e.g. Hubbard & Harrison 1989; Harrison et al. 1999), the combined stratigraphical, structural, metamorphic and geochronological data provide a strong indication that the Higher Himalayan slab has undergone southward-directed extrusion. The existence of this body of field-based evidence is independent of geodynamic interpretations that are based on regional geophysics, or that link crustal extrusion to mid-crustal channel flow beneath Tibet.

**Summary: crustal extrusion**

In this paper we use crustal extrusion as a general term to signify the commonly recognized thickening/extrusion of crust in response to bulk contraction. In this sense, the term implies no specific geodynamic process or driving force. The bulk geometry and kinematics of crustal extrusion can be modelled in...
4D using strain matrix modelling. Diagnostic field characteristics of crustal extrusion include: metamorphic inversion at the lower boundary of a high-grade zone orphaned by lower-grade rocks; high-grade regional metamorphism, possibly with partial melting in the core of the zone; broadly coeval shear zone margins with opposing senses of shear; early ductile fabrics progressively overprinted by later brittle structures; and localization of strain, with some phases of deformation only seen in the zone and not in the flanking rocks (Fig. 2a).

Channel flow

Ductile flow in the lower crust

Laboratory studies of rock strength (e.g. Goetze & Evans 1979; Brace & Kohlstedt 1980), as well as deep seismic data (e.g. Chen & Molnar 1983; Kuznir & Matthews 1988), have underpinned suggestions that layers of low viscosity in the lower crust can undergo ductile flow (e.g. Kuznir & Matthews 1988; Bird 1989, 1991; Block & Royden 1990; Molnar 1992; Bott 1999; see also Maggi et al. 2000, and Jackson 2002). These ductile zones may be weak enough to allow a degree of mechanical decoupling between upper and lower crust (e.g. Oldow et al. 1990; Royden 1996; Ellis et al. 1998; Teyssier et al. 2002; Tikoff et al. 2002; Grocott et al. 2004; Klepeis et al. 2004). Possible representatives of lower-crustal flow in the rock record include large tracts of granulite and migmatite terranes, and regions of large-scale ductile sheath folding.

Numerical modelling of ductile flow in the lower crust has shown that the flow may be driven by gravitational potential in regions of elevated topography, and could provide an effective mechanism for the compensation of local gravity anomalies (Bott 1999), and for large-scale variations in Moho topography to be gradually smoothed over geological time (Kuznir & Matthews 1988; Bird 1991). This approach to modelling has utilized standard fluid dynamics equations (e.g. Turcotte & Schubert 1982, 2002), generally assuming laminar flow within a channel geometry. Similar flow conditions have also been used to model exhumation of material along subduction channels by reverse flow (England & Holland 1979; Mancktelow 1995). Common to all the above models is that the location of channel flow is coincident with low-strength regions of the lower crust above the assumed rheology transition at the crust–mantle interface.

Channel flow in the mid- and upper crust

Grujic et al. (1996) subsequently suggested that comparable fluid dynamics principles might be applicable to mid- and upper-crustal levels, and used channel flow terminology in a conceptual model to explain field observations from Bhutan. This extended ideas developed by Nelson et al. (1995, 1996; see also Hauck et al. 1998, and Alsdorf et al. 1998), who presented geophysical data from the INDEPTH project, and postulated that mid-crustal flow is occurring under the Tibetan Plateau, with the southerly continuation of this mid-crustal low-viscosity layer being extruded as the GHS. An assumption of this model is that partial melting related to crustal thickening can radically alter the standard strength profile of the crust, allowing weaker zones of reduced viscosity to develop at mid-crustal levels that would generally be too strong for extrusive flow to occur. This concept has influenced thermal–mechanical models of the Himalaya, which have incorporated low-strength mid-crustal rheologies in order to induce channel flow to develop.

Thermal–mechanical models and channel flow

Recent advances in thermal–mechanical finite element modelling (e.g. Ellis et al. 1998; Jamieson et al. 2002) represent a new level of sophistication in the analysis of progressive evolution of orogenic belts. Thermal–mechanical models allow testable predictions to be made in relation to geodynamics, tectonics, geomorphology, and gross structural and metamorphic patterns. Application of this approach to lithospheric-scale modelling has been applied to the Himalaya–Tibet orogen (e.g. Beaumont et al. 2001, 2004, 2006; Jamieson et al. 2004, 2006). In these cases, the boundary conditions and input parameters are carefully chosen to be geologically realistic and to match actual Himalayan values as closely as possible. Conformity between the results of modelling and many first-order geological and geophysical characteristics of the Himalaya–Tibet belt suggests that the models are well calibrated to this specific orogen. The importance of this is that proper calibration is essential prior to using the modelling to test the sensitivity of different input parameters in relation to resultant tectonism. In other words, it should not be considered a weakness of the modelling method that a model is tuned to match a specific orogenic setting. On the contrary, this is the basis for a better understanding of which factors are likely to have most influence on dynamic orogenic processes such as orogenic thickening, crustal extrusion and channel flow.

Implications from thermal–mechanical models

Iterations of the thermal–mechanical modelling that test the sensitivity of various input parameters
emphasize that channel flow is unlikely to have developed unless certain boundary conditions were favourable. The aim of the following characterization of channel flow is to separate key attributes of the models into two main types, broadly corresponding to ‘cause and effect’: firstly, transient features relating to prerequisite tectonic and geodynamic boundary conditions shown to be important in inducing channel flow (‘cause’); and secondly, attributes relating to the resultant channel (‘effect’), evidence for which may remain in the rock record over significant periods of geological time, long after orogenesis has abated (Fig. 2c).

**Modelling: requisite conditions for channel flow to occur**

Large, hot, long-lived orogen. Modelling suggests that in order for a mid-crustal low-viscosity channel to develop, the crust must be suitably radiogenic, and large-scale crustal thickening must last long enough for partial melting to occur. An ancient orogen that was small, short-lived or had radiogenically depleted crust would be unlikely to have been able to support significant gravity-driven channel flow.

Extensive elevated plateau region. One of the main conclusions of modelling is that mid-crustal channel flow is driven by a large horizontal gradient in lithostatic pressure relating to a wide upland plateau region (corresponding to Tibet in the application of the model to the Himalayas), flanked by an erosional/topographic front. Indeed, the thermal–mechanical models for the Himalaya–Tibet orogen show that there appears to be such a strong dependency on gravitational loading that channel flow is unable to develop unless a large plateau is present. In an ancient orogen, lack of evidence of a sufficiently large source of gravitational potential (such as a plateau) will therefore make it significantly more difficult to demonstrate conclusively that gravity-driven channel flow has occurred. In many old orogens it may no longer be possible to demonstrate the presence of an ancient plateau region, particularly in orogenic belts that have been heavily reworked (e.g. Fenno-Scandian, Grenvillian, Grampian, Acadian, Variscan and many others), or those in which large lateral motions along strike-slip faults have dismembered much of the original collisional architecture (e.g. Western Cordillera, Scandian, Taconic, etc.).

Evidence of high denudation rates. Modelling has shown that a high rate of denudation at the margins of the elevated topographic front can have a major effect in causing the channel to propagate (‘tunnel’) upwards from mid-crustal levels, towards the Earth’s surface. Thus, high denudation rates are not a prerequisite for the channel to develop at depth (e.g. fig. 6a of Beaumont et al. 2004), but may be instrumental in promoting the exhumation of an active channel. In most ancient orogens, direct evidence of elevated rates of denudation will have disappeared, although indirect evidence may be present (e.g. from pressure–temperature–time (P-T-t) data, or from the sedimentary record in adjacent deposits; e.g. Galy et al. 1996).

**Field characteristics of resultant channel flow inferred from modelling**

Pressure, temperature and relative displacement paths in thermal–mechanical models allow inferences to be made about the resultant metamorphic and structural characteristics that should be visible in outcrop if channel flow had developed. Here we evaluate the degree to which specific characteristics will be useful as diagnostic attributes in ancient orogens. We consider features of the mid-crustal channel, as well as upper-crustal extrusion zones where the channel reaches the Earth’s surface.

Subhorizontal mid-crustal channel. Post-orogenic denudation and exhumation of mid-crustal rocks may expose evidence of the fossilized channel. This should contain regions of partial melting, migmatization and elevated metamorphic grade, flanked above and below by rocks that have experienced lower temperature metamorphism. Therefore, for positive recognition of possible channel flow, a full section showing both upper and lower channel margins should ideally be exposed. Modelling implies that structural evidence should be useful in recognizing whether channel flow may have occurred. One of the most important diagnostic characteristics is to demonstrate that shear zones were active simultaneously at both top and base of the channel, and for the zones to have demonstrably opposing shear sense (‘top-towards-orogenic-front’ for the basal shear zone, ‘top-towards-orogenic-core’ for the upper shear zone). Furthermore, the age of shearing should be broadly similar to the timing of migmatization in the channel. Difficulties may arise if the channel margins have been reworked or reactivated by subsequent tectonic episodes, as this may remove much of the evidence for earlier shear during the original channel flow.

Model predictions suggest that recognition of high strains might provide useful diagnostic information. In an idealized channel, intensity of shear strain should be greatest at the margins of the channel, and lower in the channel centre, though in reality strain is likely to be more heterogeneously
distributed. Consideration of particle path trajectories in thermal–mechanical models (e.g. figs 4 & 6 of Jamieson et al. 2004) show that for well-developed channel flow, shear strains (γ) of 10^2 or more may be common along the channel margins, with γ in excess of 10^3 also possible. Fault rocks that have experienced such high levels of strain are generally recognizable by their ultra-mylonitic fabric, even if the exact magnitude of shear strain is difficult to quantify precisely. Within the channel there should be evidence of subsimple shear strains (i.e. a non-coaxial shear with shortening across the deformation zone; examples are given by Grasemann et al. 1999; Xypolias & Doutsos 2000; Vannay & Grasemann 2001; Xypolias & Koukouvelas 2001; Law et al. 2004). In such cases, the strain magnitude in the channel margins should be lower towards the core of the orogen where the channel starts, with an increase in strain outwards towards the orogenic front (cf. ‘cream-cake tectonics’ of Ramsay & Huber 1987, pp. 610–613). Clearly this diagnostic attribute is only testable if a large extent of the mid-crustal channel is exposed across the orogen.

Strongly localized deformation due to high shear strain within the channel is likely to give rise to narrow terranes (domains) characterized by complex structural relationships, in which there is repeated overprinting of successive deformational fabrics. This may give the appearance of several phases of deformation (fabric crenulation, refolded folds, sheath fold development, fold and fabric transposition), though these should span a relatively short time period. Because these deformational ‘phases’ are caused by spatial and temporal heterogeneity along the channel, and relate to localized perturbations within the flow, the rocks outside the channel should be markedly less deformed, and some (if not most) of the ‘phases’ should be restricted to the channel, and will be largely absent beyond the channel margins.

**Zone of crustal extrusion in the upper crust.** Modelling implies that the exposure of a fossilized channel that has ‘tunnelled’ from mid-crustal to upper-crustal levels at the orogenic front, will share some diagnostic attributes with a corresponding channel at mid-crustal levels. As at greater crustal depths, the migmatitic channel should be flanked above and below by rocks of lower metamorphic grade, to give characteristic inverted and right-way-up metamorphic isograds at the base and top of the channel respectively (see section 10.1 of Jamieson et al. (2004) for discussion). This metamorphic inversion is a particularly important diagnostic feature in the upper crust, as there is likely to be greater contrast in metamorphic grade between the channel and its lower-grade flanks, with the high P-T migmatite-grade channel enveloped by markedly lower pressure and temperature greenschist- or amphibolite-grade rocks typical of upper-crustal orogenic levels. P-T-t conditions should indicate that rapid exhumation of the channel occurred synchronously with the timing of shear along the channel margins.

Coeval shear zones at the top and base of the channel should once again show opposing shear sense, with thrust displacement along the lower channel margin and normal-sense displacement at the upper margin. Modelling emphasizes that in order for the channel to have propagated towards the surface, both margins must have experienced very large strain magnitudes (material in the core of the channel must be transported large distances during displacement from mid- to upper crust). For flow in the channel to remain ductile at high crustal levels, high strain rates are needed, so that there is insufficient time for the channel to cool to temperatures at which increase in viscosity would impede flow. Thus in ancient orogens, typical characteristics of upper-crustal regions of the channel should include the presence of high strain mylonites or ultramylonites that are progressively overprinted by successive down-temperature and pressure phases of semi-ductile, semi-brittle and then brittle deformation, as material in the channel is exhumed and extruded towards the Earth’s surface. Younger brittle structures should generally show the same sense of shear as the older ductile fabrics that they overprint.

In summary, thermal–mechanical modelling allows specific predictions to be made about the field characteristics that can be expected in zones of channel flow. However, many of these characteristics are non-unique to channel flow: they are common to general zones of bulk crustal extrusion, and are widely recognized in areas in which channel flow is not believed to have occurred. The significance of this in relation to active and ancient orogens is discussed in the following sections.

**Discussion**

Application of the channel flow concept to explain Himalayan–Tibetan orogenesis has clearly been a positive catalyst in promoting vigorous scientific discussion. Much of the current debate focuses on the reliability and validity of geophysical data, and its relevance in relation to field observations of Himalayan geology. In the following discussion we focus on other inherent difficulties of the channel flow concept, particularly regarding recognition of channel flow in ancient orogens, and whether or not the concept is useful to field-based analysis of continental collision zones.
Ambiguity of channel flow semantics

Earlier sections of this paper illustrate that common usage of the term ‘channel flow’ can convey a range of slightly different meanings, each of which can carry subtle implications depending upon context. For example, with regard to fluid dynamics, channel flow is simply a generic term that encompasses a number of more specific types of quantitative flow equation. In contrast, when used with reference to outcrop observations (e.g. Grujic et al. 1996), channel flow is a conceptual term that is consistent with crustal extrusion (implying inverted metamorphism, high shear strains, and contemporaneous thrusting and normal faulting). In this context, for some workers the term ‘channel flow’ may also tacitly imply that continuation of the channel extends from the surface down to mid-crustal levels. Finally, within the context of the thermal–mechanical models of Beaumont et al. (2001, 2004), channel flow is a dynamic, gravity-driven process that requires high heat flow and large gravitational potential for significant amounts of flow to occur, as well as high rates of localized exhumation for the channel to reach the surface.

In relation to Himalayan geology, these semantic discrepancies are generally unimportant, since most proponents of channel flow generally include discussion that encompasses all the different contexts outlined above. However, in older orogenic belts, the distinction may be much more important, since in general there will be a very different balance of available data upon which interpretation can be based.

Nature of evidence in ancient orogens

The study of modern mountain belts is essential in order to improve our understanding of processes that produced ancient orogens. Conversely, ancient orogens can also increase our understanding of modern-day orogenic processes, because exhumation of ancient mountains allows us to study large tracts of deep levels of the orogen that are not widely exposed in active mountain regions today. In modern orogens such as the Himalayan–Tibetan system, many different types of data are available to help constrain tectonic interpretation and to guide the choice of suitable boundary conditions and geodynamic input parameters to use in modelling. Some of the most useful types of data are transient in nature (e.g. earthquake focal mechanisms, heat flow, bright spots on deep seismic reflection data, geodetic measurements showing relative displacement recorded by arrays of GPS stations). Other criteria can be longer-lived (e.g. topography, and sea-floor magnetic anomalies that record relative plate motion). However, in the majority of older orogens, these types of transient short-lived geodynamic data provide little or no constraint. Most ancient terranes do not have well-preserved boundary conditions, and it can be extremely difficult to infer the nature of the far-field driving forces that controlled deformation (e.g. Dewey et al. 1986). Furthermore, factors suggested by thermal–mechanical modelling to be prerequisite for channel flow to occur – such as high gravitational potential, heat flow and exhumation rates – are also transient in nature, and will generally be of much less use in constraining geodynamic interpretations of ancient orogens.

Thus, many of the types of data that are currently driving debate about channel flow in the Himalayas will not be as useful when testing the applicability of the channel flow concept to ancient orogens. In such cases, interpretation will inevitably place greater emphasis on field studies and analysis of samples collected at outcrop, and must rely heavily on a combination of critical types of metamorphic, structural and geochronological evidence. In our opinion, in order to test different geodynamic models as objectively as possible, it is important to maintain a descriptive framework for making field observations that avoids recording data in terms of a genetic description linked to a specific interpretation of causative processes.

Non-uniqueness of field characteristics inferred from modelling

Predicted field characteristics of channel flow inferred from thermal–mechanical modelling show close similarity with actual field observations from the Himalayan–Tibetan system (see above). However, while field data do provide strong evidence that crustal extrusion has taken place in the High Himalaya, it is much more contentious to demonstrate incontrovertibly that extrusion at the surface is linked to zones of channel flow beneath Tibet far to the north. Even in modern orogens where transient geodetic and geophysical data can be acquired, a fundamental challenge facing geoscientists is to test the range of possible geodynamic interpretations that are consistent with the available data. At the current level of understanding, the characteristics predicted by thermal–mechanical modelling, as well as actual field observations and analyses from the Himalayas, are non-unique in terms of causative geodynamic driving forces. In particular, regions of crustal extrusion can result from horizontal forces relating to convergent plate motion, and need not imply processes driven predominantly by gravitational potential. This is illustrated below with an example from the Nanga
Crustal extrusion in the Nanga Parbat–Haramosh Massif

The Nanga Parbat–Haramosh Massif (NPHM) lies at the NW end of the Himalayan arc, and is bordered to the west by the Hindu Kush, to the north by the Karakorum range, and to the SE by the High Himalayas (Fig. 3a). As part of the long-lived, hot Himalayan system, with the vast Tibetan plateau to the north, the massif shares many of the geodynamic attributes that characterize regions further to the east. The NPHM is marked by extreme topography, with over 7000 m of relief between the summit of Nanga Parbat (8125 m) and the base of the Indus valley. Rapid denudation is facilitated by the enormous sediment-bearing capacity of the Indus, which may have allowed an effective positive feedback mechanism to develop, where increased uplift rates are driven by rapid erosion (Zeitler et al. 2001; Koons et al. 2002).

The massif is a large asymmetric antiformal dome (Fig. 3b), forming a crustal-scale syntaxis (Fig. 3c) that marks a major change in the overall trend of the orogenic belt (Tahirkheli & Jan 1979; Coward 1983, 1985). Formation of the syntaxis has reworked the earlier structure of the Main Mantle Thrust (Butler & Prior 1988a, b), which represents the tectonic plate boundary between India and Asia. The age of the syntaxis post-dates peak metamorphism, and probably initiated within the last 10 million years (Treloar et al. 2000; Zeitler et al. 2001; Butler et al. 2002). The core of the massif consists of migmatitic crystalline basement rocks of Indian plate affinity that have been extruded through structurally overlying units of the Kohistan–Ladakh Arc, which had accreted to the SE margin of the Asian plate prior to collision with India from c. 55 Ma onwards (Treloar et al. 1989; Treloar & Coward 1991). Active crustal extrusion is ongoing, and is associated with extremely rapid uplift (Zeitler 1985) and high denudation rates (Bishop & Shroder 2000), suggesting exhumation rates of up to 7 mm a⁻¹ (Zeitler 1985; Zeitler et al. 2001). Seismic data suggest that the base of the upper-crustal seismogenic zone is significantly elevated beneath the syntaxis (Meltzer & Christensen 2001; Meltzer et al. 2001). The NPHM is one of the most intensely studied regions of Himalayan geology, and there

**Fig. 3.** Nanga Parbat–Haramosh Massif (NPHM), northern Pakistan: (a) satellite image of Himalaya, Tibet and the Indian subcontinent (courtesy of NASA/Visible Earth); (b) cross-section showing asymmetric antiformal nature of the massif and amount of material removed by erosion (after Butler et al. 2002); (c) map of the massif (collated from Butler et al. 1989, 2000; Searle & Khan 1996; Fêcher & Le Fort 1999; Argles 2000; Edwards et al. 2000).
is a wealth of published data that integrate a wide range of stratigraphic, metamorphic, structural, isotopic, geochemical and geophysical evidence. Detailed descriptions of the geology of the massif are given by Khan et al. (2000) and references therein, as well as Zeitler et al. (2001) and Butler et al. (2002).

Geochronological, metamorphic and structural data provide strong evidence that ductile extrusion of mid-crustal material is taking place in the core of the massif. Fission track data and muscovite, biotite and hornblende cooling ages show that the central parts of the NPHM have experienced very rapid exhumation relative to surrounding regions: \(40^{\text{Ar}}/39^{\text{Ar}}\) mica cooling ages suggest rates of between 3–4 mm a\(^{-1}\) (Whittington 1996) and 7 mm a\(^{-1}\) (Zeitler 1985; Zeitler et al. 2001). A range of P/T isotope data suggest that up to 25 km of crust may have been eroded from the central parts of the massif within the last 10 million years (Butler et al. 1997; Whittington et al. 1999). Uplift and exhumation are associated with partial melting and emplacement of small granitoid bodies. These yield zircon ages as recent as 1 Ma (Zeitler et al. 1993), and generally young towards the core of the massif (summarized in Zeitler et al. 2001).

Structural studies around the margins of the massif have revealed a protracted kinematic history in which earlier ductile fabrics are generally overprinted by later semi-ductile to brittle structures. The age of deformation is locally well constrained, either by direct dating of structural fabrics, or by dating of cordierite seams and anatectic leucogranites that are intimately associated with shearing (e.g. Zeitler & Chamberlain 1991; Zeitler et al. 1993; Butler et al. 1997; Schneider et al. 1999; Whittington et al. 1999). Currently active thrusting along the western margin of the NPHM is emplacing Nanga Parbat gneisses northwestwards over unconsolidated river gravels and rocks of the Kohistan arc (Butler & Prior 1988a; Butler 2000). In their hanging wall, these brittle structures carry semi-brittle and ductile fabrics that also have the same top-to-NW kinematics, including 1–2 km of mylonitic gneiss with well-developed S-C fabrics (Butler 2000). Detailed fabric analysis comparing variation in the ellipticity and preferred orientation of augen, shows that the finite strain in the gneiss is a subsimple shear, with a strong component of up-dip elongation (Butler et al. 2002). Boudinaged calc-silicate layers suggest elongations of 30%. There is also evidence of a component of right-lateral strike-slip along the western margin (Butler et al. 1989; Butler 2000), showing that the bulk deformation of the region is transpressional. The sequence of ductile to brittle structures now visible at the surface is likely to be very representative of comparable structures that are currently forming at different depths in the massif, and which will be progressively exhumed as crustal extrusion continues.

The eastern margin of the NPHM is not as widely studied or well understood as the western margin. There are local areas of top-to-NW thrusting (Argles 2000), while south of Nanga Parbat, top-to-south thrusting has been documented (Butler et al. 2000). However, most parts of the eastern margin appear to be marked by ductile fabrics of the Main Mantle Thrust (MMT), subsequently steepened during formation of the syntaxis (Butler et al. 1992; Wheeler et al. 1995), without significant overprinting of recent brittle or semi-brittle deformation (Argles 2000). Although further fieldwork is clearly needed to provide additional constraints on the kinematics of the eastern margin, available evidence suggests that later (post-MMT) strain is more distributed than in the west, and that top-to-SE deformation was accommodated largely by upwarping of the eastern antiformal limb during formation of the syntaxis.

Summary

There is a wealth of field data showing key characteristics of crustal extrusion in the NPHM, including: high-grade units in the core of the massif, with inverted metamorphism across the lower margin; evidence of partial melting, marked by younger ages in the centre of the zone; active deformation synchronous with crystallization of anatectic melts; ductile fabrics progressively overprinted by semi-ductile and brittle structures; and subsimple shear strains, with a large component of up-dip stretching. Recognizing that the prerequisite geodynamic conditions for channel flow are present (large, hot orogen, elevated plateau, extreme denudation), Beaumont et al. (2004) proposed that the NPHM is another potential example of active channel flow. However, it remains extremely difficult to establish the main causative driving force(s) of extrusion in the massif, and several alternative models for its development have been proposed. These include the NPHM as a fault-related antiform at the lateral tip to major Himalayan thrusts (Coward et al. 1988); a zone of NW shortening in response to orogen-parallel extension around the Himalayan arc (Seeber & Pécher 1998); upwelling of crustal gneiss domes (Pécher & Le Fort 1999; Koons et al. 2002); crustal-scale buckling (Burg & Podladchikov 2000); and north–south constriction/transpression above a thrust tip (Butler et al. 2000). Although the far-field boundary displacements between India and Asia are well established and reasonably straightforward (Treloar & Coward 1991), the
Crustal extrusion in the Wing Pond Shear Zone, Newfoundland

The Gander Zone in the Appalachian orogenic belt of northeastern Newfoundland (Fig. 4a) represents part of a Gondwanan-derived continental fragment thought to have been accreted to Laurentia during the closure of the Lower Palaeozoic Iapetus Ocean (e.g., Williams et al. 1988; Soper et al. 1992). The Iapetus suture in Newfoundland lies within the Dunnage Zone, the eastern part of which was obducted onto the Gander Zone during the Arenig with the boundary marked by an allochthonous unit of melanges (Gander River Complex, Fig. 4a) emplaced initially on eastward-directed thrusts. The present-day eastern margin of the NE Gander Zone is the Dover Fault Zone, a major reactivated brittle–ductile structure (Blackwood & Kennedy 1975; Holdsworth 1994). The Avalon Zone to the east comprises a characteristic Gondwanan Neoproterozoic tectonostratigraphy and Palaeozoic cover (O’Brien et al. 1983).

The metasedimentary rocks forming most of the Gander Zone in NE Newfoundland (the ‘Gander Group’) are a thick sequence of variably deformed and metamorphosed clastic metasediments and gneisses (psammites, pelites) intruded by numerous granites (Fig. 4a; Blackwood 1977; Hanmer 1981; O’Neill 1991; Holdsworth 1994). The deformational and metamorphic character of the Gander Zone is highly heterogeneous. Two regional ‘flat belts’ are recognized, characterized mainly by greenschist-facies (chlorite–white mica) assemblages which are deformed by generally eastward-verging recumbent folds locally termed ‘F2’ (e.g. Fig. 5a; Kennedy & McGonigal 1972) thought to have formed during overthrusting of the Dunnage Zone in the Arenig (c. 480 Ma). These are post-dated by two 10–25 km wide, high-strain transpressional steep belts (Fig. 4a) of similar age, both of which are characterized by significantly elevated P-T conditions (D’Lemos et al. 1997; King 1997). The easternmost of these, here termed the Hare Bay Gneiss Shear Zone (HBGSZ), is characterized by migmatization and syntectonic granite emplacement during sinistral shear thought to be related to initial docking of the Gander and Avalon terranes during the Silurian (e.g. Holdsworth 1994). Peak metamorphism, derived from mineral reactions, cordierite geobarometry and hornblende–plagioclase geothermometry in the HBGSZ was at high temperature–low pressure conditions (c. 740°C at 4.5 kbar; D’Lemos et al. 1997). Further to the west, the Wing Pond Shear Zone (WPSZ) (O’Neill 1991) is a structurally more complex region of focused metamorphism and high strain up to 10 km across (Fig. 4b); its age relative to the HBGSZ is uncertain, although existing geological constraints suggest that they are broadly contemporaneous.

The WPSZ is well exposed in roadside exposures along the Trans Canada Highway north and east of the later Gander Lake granite (Fig. 4b). Moving eastwards from Benton, early recumbent folds and associated greenschist-facies fabrics of the western Gander Zone are progressively overprinted and transposed by upright ‘F3’ folds and an associated steeply dipping, NE–SW trending ‘S3’ fabric (Fig. 5b; O’Neill 1991). S3 is a prograde foliation defined by increasing proportions of biotite and is associated with low-P prograde metamorphism resulting in formation of cordierite- and andalusite-bearing assemblages as the intensity of S3 increases. In places andalusite and cordierite are overprinted by S3 (Fig. 6a, b), whereas in other instances porphyroblast growth occurred late- or post-S3 (Fig. 6c). For the general range of bulk compositions in the Gander Group, the growth of cordierite- and/or andalusite-bearing assemblages implies prograde metamorphism at ≤ 3 kbar (Fig. 6f).

Further to the east, local ‘F4’ folds appear, tighten and are themselves transposed within planar, steeply dipping high-strain zones (Fig. 5c), with the preceding structural complexity only being apparent in local low strain windows associated with meso-scale F4 fold hinge zones (e.g. Fig. 5d). These planar high-strain panels define the structural geometry of the WPSZ and record sinistral transport parallel to a low-angle southward-plunging lineation. The high-strain shear fabric is associated with development of kyanite–staurolite ± garnet-bearing assemblages that formed at the expense of syn-S3 andalusite (Fig. 6d). The kyanite-bearing assemblages formed at around 5 kbar, 600°C (Fig. 6f) and are overprinted by sillimanite-bearing fabric (Fig. 6e) in the core of the WPSZ.
Along the eastern margin of the shear zone, local F5 folds appear and the high-strain fabric becomes mylonitic. It is characterized by greenschist-facies assemblages, and appears to correspond to the eastern margin of the WPSZ which is exposed along-strike, SW of the Deadman’s Bay Granite (Fig. 4a). The entire WPSZ domain is characterized by mainly steeply plunging mineral-stretching lineations, although there are local high-strain zones with shallow-plunging lineations. The F3–5 folds are geometrically identical sheath folds and appear to reflect progressive deformation and shear localization within the WPSZ. The eastern margin of the WPSZ is flanked by a system of older shallow-dipping fabrics that resemble the flat-lying system west of the shear zone. Viewed in this context the WPSZ has a simple geometry with a steeply dipping core flanked by domains of progressively lower strain.

Summary

The geometry of the WPSZ and its flanking structure is mirrored by both the kinematic and metamorphic character of the structure. The western margin of the shear zone records sinistral kinematics (Fig. 5e) with a component of east-up displacement, while the eastern margin of the shear zone records west-up displacement with dextral component of movement (Fig. 5f). The gradual increase in strain towards the shear zone core suggests that this is a zone of diffuse crustal
extrusion (cf. Fig. 2b), rather than a zone of low strain flanked by high strain margins. There is also an apparent progressive increase in metamorphic pressure towards the central region of the shear zone that is compatible with the kinematics, suggesting relative upward displacement of the shear zone core. On the flanks, andalusite-bearing assemblages formed at pressure of c. 3 kbar during progressive ‘D3’ deformation, while in the central region kyanite–staurolite-bearing

Fig. 5. Field evidence for channelized extrusion from the WPSZ. (a) Flat-lying low-grade psammites and pelites with tight ESE verging and facing fold pair (local F2), typical of the flat belts outside the WPSZ. Northern shore of Gander Lake. Notebook is 190 cm high. (b) Steeply dipping S3–S4 fabric typical of the WPSZ from the shores of Wing Pond. Note the abundant quartz segregations indicative of the higher metamorphic grades in these rocks. (c) Local intensification of S4–S5 fabric (in area to the left of the person) in the core region of the WPSZ; Square Pond quarry (east), just north of the Trans Canada Highway. (d) Typical lower strain augen in F4 fold-hinge zone from Mint Brook revealing underlying structural complexity in WPSZ. Close F4 folds refold tight F3 folds which themselves refold a spaced S2 solution fabric. Lens cap 55 mm diameter. (e) Plan view of sinistral asymmetric boudins of quartz segregations from a high strain zone close to the NW margin of the WPSZ, next to the Trans Canada Highway. (f) Plan view of dextral asymmetric boudins of quartz segregations from the SE margin of the WPSZ near Gambo.
Fig. 6. Petrological evidence for channelized extrusion from the WPSZ (long dimension of all photomicrographs = 7 mm). (a) Cordierite (now retrogressed) overprinted by S3 crenulations. (b) Partially retrogressed andalusite wrapped by S3 biotite–muscovite-bearing foliation. (c) S3 crenulations overgrown by andalusite. (d) Andalusite replaced by kyanite–staurolite-bearing ‘D4’ assemblage within the centre of the WPSZ. (e) Sillimanite-bearing ‘D4’ assemblage overprinting syn-D3 andalusite partially replaced by kyanite. (f) Metamorphic evolution of rocks in the WPSZ system depicted on a generalized KFMASH P-T pseudosection (+ muscovite-qtz) adapted from Reche et al. (1998) applicable to the aluminous metapelitic compositions that occur within the Gander Zone metasediments. On the flanks of the shear zone andalusite-bearing assemblages formed during progressive ‘D3’ deformation. Cordierite growth is generally later than andalusite, but still locally shows synkinematic relationships with ‘S3’. In the core of the shear zone, increasing pressure is implied by the replacement of andalusite by kyanite and staurolite. Subsequent growth of sillimanite is interpreted to reflect decompression associated with exhumation that differentially juxtaposed the now-exposed relatively high-P central region of the shear zone against the lower pressure flanking domains.
Conclusions

Intense study of active orogenic belts such as the Himalayan–Tibetan system emphasizes the inherent difficulty in deriving unique geodynamic interpretations based on available geodetic, geophysical and field-based evidence. In studies of ancient orogenic systems, transient geodetic and geophysical data are generally not available or are less useful, placing greater reliance on field-based metamorphic, structural and geochronological evidence.

Thermal–mechanical modelling can be extremely useful in testing the potential influence of various geodynamic boundary conditions, although many of the field characteristics predicted from modelling are common to general zones of crustal extrusion arising in other tectonic regimes, including transtension zones. Field examples from Nanga Parbat in northern Pakistan and the Wing Pond Shear Zone in Newfoundland display strong field evidence of significant crustal extrusion; however, there is a lack of evidence to show that extrusion is gravitationally driven or linked to subhorizontal mid-crustal channel flow. The main driving force of crustal extrusion in these areas is difficult to determine with certainty, but is at least as likely to be tectonically rather than gravitationally driven.

Because of the inherent difficulties in identifying the driving mechanisms of lithospheric-scale geodynamic processes, we believe that a clear distinction should be made between primary observations that describe field characteristics of crustal extrusion (Fig. 2a, b), and geodynamic interpretation that tries to relate the data to specific large-scale processes or driving mechanisms (Fig. 2c).

In this respect, existing usage of the term ‘channel flow’ is already ambiguous, since it is used variously to describe any or all of the following: (i) a general type of flow condition (in relation to fluid dynamics); (ii) gravitationally driven mid-crustal flow, which if aided by high denudation might propagate upwards to cause crustal extrusion at the surface (in the context of recent thermal–mechanical models); and (iii) generalized bulk crustal extrusion, which may extend downwards to connect to subhorizontal mid-crustal zones (in the context of a conceptual explanation for Himalayan field observations such as inverted metamorphism, high shear strains, and contemporaneous thrusting and normal faulting). Consequently, our recommendation is that the terminology used for field description and geometric/kinematic interpretation in orogenic belts should include terms such as ‘crustal extrusion’, ‘ductile extrusion’ or ‘channelized extrusion’. Such terms should be considered as broadly synonymous, and refer to zones that show inverted metamorphism, coeval marginal shear with opposing sense etc., without any implication of the far-field driving mechanism. Particularly in older orogenic belts, care should be taken to explain clearly the exact meaning of the term ‘channel flow’, to discuss any geodynamic implications in relation to the context in which the term is used, and to separate primary field observations from subsequent geodynamic interpretation.

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