Crustal structure, restoration and evolution of the Greater Himalaya in Nepal–South Tibet: implications for channel flow and ductile extrusion of the middle crust

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Abstract: Recent suggestions that the Greater Himalayan Sequence (GHS) represents a mid-crustal channel of low viscosity, partially molten Indian plate crust extruding southward between two major ductile shear zones, the Main Central thrust (MCT) below, and the South Tibetan detachment (STD) normal fault above, are examined, with particular reference to the Everest transect across Nepal–South Tibet. The catalyst for the early kyanite + sillimanite metamorphism (650–680°C, 7–8 kbar, 32–30 Ma) was crustal thickening and regional Barrovian metamorphism. Later sillimanite + cordierite metamorphism (600–680°C, 4–5 kbar, 23–17 Ma) is attributed to increased heat input and partial melting of the crust. Crustal melting occurred at relatively shallow depths (15–19 km, 4–5 kbar) in the crust. The presence of highly radiogenic Proterozoic black shales (Haimanta–Cheka Groups) at this unique stratigraphic horizon promoted melting due to the high concentration of heat-producing elements, particularly U-bearing minerals. It is suggested that crustal melting triggered channel flow and ductile extrusion of the GHS, and that when the leucogranites cooled rapidly at 17–16 Ma the flow ended, as deformation propagated southward into the Lesser Himalaya. Kinematic indicators record a dominant south-vergent simple shear component across the Greater Himalaya. An important component of pure shear is also recorded in flattening and boudinage fabrics within the STD zone, and compressed metamorphic isogradic limits within both the STD and MCT shear zones. These kinematic factors suggest that the ductile GHS channel was subjected to subvertical thinning during southward extrusion. However, dating of the shear zones along the top and base of the channel shows that the deformation propagated outward with time over the period 20–16 Ma, expanding the extruding channel. The last brittle faulting episode occurred along the southern (structurally lower) limits of the MCT shear zone and the northern (structurally higher) limits of the STD normal fault zone. Late-stage breakback thrusting occurred along the MCT and at the back of the orogenic wedge in the Tethyan zone. Our model shows that the Himalayan–South Tibetan crust is rheologically layered, and has several major low-angle detachments that separate layers of crust and upper mantle, each deforming in different ways, at different times.

The Tibetan plateau (Fig. 1), covering an area of >5 × 10⁶ km², is an arid plateau with low relief and low erosion rates, and forms the largest area of high elevation (average elevation of 5023 m) and thick crust (65–80 km) on the planet (Fielding et al. 1994). The plateau is bordered to the south by the Himalayan range, which has a similar average elevation, but much greater topographic relief, ranging from 2000 to 8850 m, intense fluvial and glacial erosion, and high erosion rates (Duncan et al. 2003). The horizontal gradients in lithostatic pressure between the Tibetan plateau and the Indian shield south of the Himalaya, with its normal crustal thickness (c. 35–40 km) and low elevation (<1.5 km), is the driving force behind deformation involving crustal flow away from the plateau. Two major types of crustal flow have been proposed for Tibet: firstly, lower crustal flow in eastern Tibet extruding east and SE around the eastern Himalayan syntaxis; and secondly, middle crustal flow in a layer or channel extruding southward from beneath the southern part of the plateau to the Greater Himalayan range (Fig. 1).

The concept of planar channel flow (combined ‘Couette flow’ (simple shear) and ‘Poiseuille flow’ (also known as the ‘pipe-flow effect’) of the weak lower crust, which removes crust from beneath mountains and levels out topography, was first proposed by Bird (1991). The crustal structure of the eastern part of the Tibetan plateau has been interpreted in terms of lateral flow of a ductile layer of lower crust by several authors (Royden et al. 1997; Clark & Royden 2000). This appears to be supported by GPS observations that show the ground surface of north Tibet rotating around the eastern syntaxis as far as Yunnan and Burma...
Velocity vectors show that the south Tibetan crust and Himalaya (south of 32°N) are moving along an azimuth of 020°NNE with respect to stable Asia. North of this, the velocity vectors swing around to ENE in northern Tibet, east along the eastern margin of the plateau and continue to the south in Yunnan and northern Burma (Wang et al. 2001). This eastward flow of material is restricted to northern Tibet (Qiangtang and Kunlun terranes), north of the Jiale fault and south of the Kunlun fault (Fig. 1). Northern Tibet has a thin, hot lithosphere, high Poissons’ ratio (ratio of latitudinal to longitudinal strain), a seismic anisotropy, possibly developed by lateral flow in the mantle, a probable fragmental lower crust and abundant Tertiary–Quaternary volcanic activity (Owens & Zandt 1997; Kosarev et al. 1999; Hacker et al. 2000). Southern Tibet has a thicker, cooler lithosphere consistent with underthrusting of crust and mantle of Indian plate affinity north as far as the latitude of the Bangong suture at c. 32°N (Nelson et al. 1996; Owens & Zandt 1997).

The lower crustal flow of the Qiangtang and Songpan Ganze terranes of northern Tibet contrasts with the channel flow model for the Greater Himalaya–south Tibet crust, in which a layer of hot, ductile deforming and partially molten middle crust was extruded south from beneath the southern part of Tibet (e.g. Grujic et al. 1996, 2002; Beaumont et al. 2001, 2004; Searle et al. 2003; Jamieson et al. 2004). Here, the mid-crustal layer of Indian plate rocks is represented by the Greater Himalayan Sequence (GHS) which is exposed in areas that form most of the highest topography, and the deepest exhumed rocks, along the Himalayan arc. The lower crust, comprising Archean Indian shield granulites, has been underthrust to the north, and is relatively rigid and melt-free. Since the crust beneath the Tibetan plateau is up to 80 km thick (c. 20–22 kbar pressure, assuming average crustal densities) it is possible that the lower crust has transformed into eclogite beneath the Lhasa block of southern Tibet. However, whereas granulites are dry, strong and capable of supporting thick crust like Tibet, eclogites may be hydrous, and have a density similar to, or possibly even greater than, that of the upper mantle. Ultra-high temperature and near
ultra-high pressure xenoliths in young alkali basalts in central Tibet include felsic granulites and both felsic and mafic eclogites, interpreted as tapping deep levels of the Tibetan crust (Hacker et al. 2000; Ducea et al. 2003).

This paper synthesizes data from the Everest transect, Tibet–Nepal in the context of the channel flow model. For the subsurface structure, we rely on the various seismic profiles from project INDEPTH (e.g. Nelson et al. 1996; Alsdorf et al. 1998; Hauck et al. 1998). For the surface geological mapping we use our data (e.g. Searle 1999a, b, 2003; Searle et al. 2002, 2003) as well as that of Lombardo et al. (1993), Carosi et al. (1998, 1999a, b), and for the microstructural interpretation we use the data from Law et al. (2004). We attempt a restoration of the entire Nepal–south Tibet GHS and use this restored section as a template to determine the protoliths of GHS rocks, the depth of melting of the Everest leucogranites, and the thrust and normal fault trajectories. Finally, we discuss the role of the South Tibetan Detachment (STD) normal faults as a passive roof fault during channel extrusion, and the role of the Main Central Thrust (MCT) in the extrusion and exhumation of the GHS.

**Greater Himalayan Sequence**

The GHS, as exposed along the High Himalaya, is composed of a mid-crustal layer of high-grade metamorphic rocks and migmatites with sheets of crustal melt leucogranites prominent along the higher structural levels. The upper contact of the crustal melt leucogranites prominent along the metamorphic rocks and migmatites with sheets of composed of a mid-crustal layer of high-grade

The GHS, as exposed along the High Himalaya, is composed of a mid-crustal layer of high-grade metamorphic rocks and migmatites with sheets of crustal melt leucogranites prominent along the higher structural levels. The upper contact of the GHS is a low-angle normal fault system, the STD, which generally dips north (5–20°), beneath the Tibetan plateau (Burg 1983; Burg et al. 1984; Burchfiel et al. 1992; Searle et al. 1997). The lower boundary of the GHS is a high strain ductile shear zone, the MCT, which is coincident with a zone of inverted metamorphic isograds from kyanite grade down to biotite–chlorite grade (Hubbard 1996; Stephenson et al. 2000, 2001). The MCT also dips at low angles to the north and places Proterozoic and Palaeozoic rocks, metamorphosed during the Himalayan orogeny, south over Lesser Himalayan thrust sheets, which were largely unaffected by Himalayan metamorphism (Gansser 1964, 1983; Hodges 2000; Searle & Godin 2003).

The large-scale structure and pressure–temperature (P–T) constraints across the GHS have led to numerous thermal–mechanical models (e.g. Searle & Rex 1989; Hodges et al. 1992; Harrison et al. 1998; Searle et al. 1999b). Several recent papers interpreted the structural and thermal data from the GHS in terms of a channel flow model, whereby the middle crust was extruded southwards between coeval thrust- and normal-sense shear zones (Grujic et al. 1996, 2002; Beaumont et al. 2001, 2004; Searle et al. 2002, 2003; Searle & Szulc 2005). Although these models differ in some aspects along the strike of the mountain range, several first-order field- and laboratory-based data sets form the basis for meaningful models (see next section). Figure 2a and b show a map of the Nepal–South Tibet Himalaya, and a cross-section across the STD in the Everest region. Figure 2c is a restoration of the profile in Figure 2b using pressure-depth constraints from thermobarometry of footwall gneisses and assuming a constant dip of the STD (Searle et al. 2002, 2003). These authors suggested horizontal transport or flow of at least 100 km (possibly as much as 200 km) of GHS rocks along the footwall of a passive roof fault, the STD, during ductile channel flow during the Miocene.

The thermal structure of the Greater Himalaya has been interpreted with the help of petrologic and thermobarometric data. P–T profiles across the GHS generally show an inversion of temperature and pressure across the narrow MCT zone (e.g. Hubbard 1996; Searle et al. 1999b, 2003; Stephenson et al. 2000, 2001). Structurally above the kyanite-grade rocks, the Everest section shows a 50 km horizontal width of sillimanite-grade gneisses (Hubbard 1989, 1996; Simpson 2002; Searle et al. 2003). Temperatures remain high for 50 km along a north–south profile from the footwall of the STD at Rongbuk south as far as Karikhola, south of Lukla. Pressures are highest (7–8 kbar) in the south above the MCT zone and drop off to the north to 4–5 kbar (Searle et al. 2003, fig. 8). Pressures of Everest greenschist rocks above the Lhotse detachment are poorly constrained, but they must be relatively low (<3 kbar) because of the low metamorphic grade (upper greenschist facies). Recent discoveries of staurolite-grade schists from the south face of Lhotse (Jessup et al. 2004) suggest that metamorphism decreases up-structural section along the top of the GHS slab in a structural style similar to that observed in Zanskar (Searle et al. 1992, 1999b).

Several authors have suggested that metamorphic isograds were affected by crustal-scale post-peak metamorphic folding and shearing (Searle & Rex 1989; Hubbard 1996; Walker et al. 1999; Stephenson et al. 2000, 2001). At least in the NW Himalaya, it can be demonstrated that isograds were inverted by south-vergent folding and associated simple shear along the MCT zone at the base of the extruding layer. Metamorphic isograds at the top of the GHS extruding layer are right-way-up and affected by post-metamorphic normal-sense shearing and flattening (Searle & Rex 1989; Dezès et al. 1999; Walker et al. 1999;
Fig. 2. (a) Geological map of the central Himalaya in Nepal, Sikkim, Bhutan and south Tibet. K, Kathmandu; MBT, Main Boundary Thrust; MCT, Main Central Thrust. (b) Geological section across the Everest–Rongbuk profile, marked a and b on (a) (after Searle 2003); vertical and horizontal scales are equal. (c) Restoration of the Everest profile along the line of section marked on (a) after Searle et al. (2002, 2003). The Qomolangma Detachment (QD) and Lhotse Detachment (LD) are part of the South Tibetan Detachment system of low-angle normal faults. Crosses mark the restored positions and depths of three samples from Kala Patar and Pumori, 2 km SW of Everest, which record pressures of 4.0, 4.1 and 4.9 kbar. Using a mean 10° dip of the STD passive roof fault, the relative southward displacement of footwall Greater Himalayan sequence (GHS) rocks is 90–108 km. The true dip of the STD on the north face of Everest is between 3 and 5° N; at this angle the relative displacement along the STD would be 180–216 km. The approximate depth and original position of Everest leucogranites prior to emplacement beneath the passive roof fault is also shown.
Jessup et al. 2004). In many parts of the Garhwal and Nepal Himalaya, late-stage brittle normal faulting has cut out the flattened and sheared right-way-up isograds at the top of the GHS layer, and the STD fault places unmetamorphosed Palaeozoic sediments directly on top of sillimanite-grade gneisses and/or leucogranites (e.g. Shisha Pangma (Searle et al. 1997), Garwal (Searle et al. 1999a), Rongbuk valley (Hodges et al. 1998; Murphy & Harrison 1999; Searle 1999a; Searle et al. 2003) and Annapurna (Godin et al. 2001)).

Uniform P–T conditions of the sillimanite-grade gneisses along the Everest transect between the MCT zone at the base and the Lhotse detachment at the top, together with the lack of major structural discontinuities (except the Khumbu thrust; Searle 1999a, b) suggest that the GHS acted as a relatively homogeneous high-temperature, partially molten slab during the early Miocene. This thermal structure supports the channel flow model (Fig. 3), in which the ductile deforming channel is flowing south, bounded by major crustal-scale, low-angle ductile shear zones both below (thrust-related) and above (normal sense of shear). The timing of motion along the MCT zone along the base, and the STD zone along the top of the channel, has been extensively reviewed (Hodges et al. 1992; Hodges 2000; Searle et al. 2003; Searle & Godin 2003; Godin et al. 2006a). Both ductile shear zones were active during the early Miocene (23–15 Ma) and hornblende and mica ages are consistently older than 16–14 Ma, providing a minimum time constraint on when both shear zones were exhumed from the ductile into the brittle regime. It is possible that both shear zones are still active (e.g. Hodges et al. 2001). Evidence for this is the major difference in geomorphology and relief across both the bounding shear zones, evidence for young and recent motion along the lower MCT thrusts coincident with knick-points in the rivers, and the fact that the highest topography occurs largely across the GHS.

Himalayan channel flow model

In the western Himalaya, several studies combining structural mapping with thermobarometry show that metamorphic isograds were folded after peak metamorphism, resulting in a right-way-up metamorphic sequence along the top of the GHS beneath the Zanskar shear zone (ZSZ), and an inverted metamorphic sequence along the MCT zone at the base of the slab (Searle & Rex 1989; Searle et al. 1992, 1999b; Walker et al. 1999; Dezes et al. 1999; Stephenson et al. 2000, 2001; Robyr et al. 2002; Vannay et al. 2004). Peak sillimanite + K-feldspar-grade metamorphism was concomitant with partial melting at c. 22–20 Ma (Noble & Searle 1995). The resulting 5–10 km thickness of migmatites, with abundant leucogranite sills, formed by dehydration melting of muscovite in the underlying metapelites at 4–7 kbar and 650–750 °C (Searle et al. 1992, 2003). The right-way-up isograds along the footwall of the ZSZ were mapped, linking with the inverted isograds along the MCT hanging wall via a large-scale, SW-vergent, NW-plunging recumbent fold in western Zanskar (Searle & Rex 1989; Searle et al. 1992, 1999b; Stephenson et al. 2001). These data are consistent with extrusion of the GHS by channel flow and imply at least 60–100 km of motion occurring along both the bounding shear zones, the MCT and the ZSZ.

In the eastern Himalaya, a similar model was developed for the Bhutan Himalaya, where Grujic et al. (1996, 2002) proposed that the GHS formed the core of a low-viscosity channel extending under the Tibetan plateau. Lateral lithostatic pressure gradients produced a ‘pipe-flow’ effect with the highest velocities in the centre of the channel. Beaumont et al. (2001, 2004) presented a quantitative model in which this crustal channel was coupled to surface denudation at the High Himalayan front. In contrast, Grasemann et al. (1999) and Vannay & Grasemann (2001) proposed a general shear model due to the requirements of strain compatibility, whereby the centre of the channel extruded by pure shear flow, rather than the heterogeneous simple shear model invoked by Grujic et al. (1996).

We propose the following points as fundamental boundary conditions for the development of an orogenic channel flow model in the Himalaya (Fig. 3).

1. Channel flow is driven by the topographic and crustal thickness variation between Tibet (c. 5 km elevation; 70–80 km thick crust) and the Indian foreland (<1 km elevation; c. 35 km thick crust).
2. Presence of a mid-crustal layer, between c. 15 and 40 km depth, composed of ductilely deforming rocks with in situ partial melting (Zhao et al. 1993; Nelson et al. 1996).
3. A zone of inverted, telescoped and flattened metamorphic isograds and isotherms is present along the basal contact of the channel (Main Central Thrust zone). A ductile shear zone within kyanite-grade rocks propagates down-section to a later brittle thrust with time (Stephenson et al. 2000, 2001).
4. A zone of right-way-up, telescoped and flattened isograds and isotherms is present along the top of the extruding channel (STD). A ductile, normal-sense shear zone propagates upwards to a later brittle normal fault with
Fig. 3. Channel flow model for the Greater Himalaya proposed in this paper. The deep structure beneath southern Tibet is based on an interpretation of the INDEPTH seismic profile (Nelson et al. 1996; Hauck et al. 1998) and shows the reflectors bounding the partial melt layer of middle crust and the "bright spots" interpreted here as pockets of leucogranite magma forming today. The Miocene Himalayan leucogranites are shown as small crosses along the top of the Greater Himalayan slab composed of metamorphic rocks, migmatites and in situ melting (shaded area). The lower crust is the relatively rigid, anhydrous granulite-facies metamorphism of the underthrust Indian shield. Also shown are the recumbent folded metamorphic isograds of the Greater Himalayan sequence (Searle & Rex 1989) linking the right-way-up metamorphism along the footwall of the STD at the top with the inverted isograds along the MCT ductile shear zone at the base of the extruding middle crust. The horizontal hatched layer in the suture zone is interpreted as a layer of ophiolitic rocks and the large crosses mark the Gangdese pre-collisional granite batholith.
5. The two shear zones bounding the extruding channel above (STD) and below (MCT) must be structurally linked and must move synchronously (Searle & Rex 1989; Hodges et al. 1992, 1993, 1996).
6. Presence of partial melts (migmatites) or leucogranite magmas lubricating the flow within the extruding layer, but not cutting across the bounding brittle fault zones (Searle et al. 1992, 1999a, 2003).
7. Coupling between channel flow and surface erosion. Erosion is highest where the channel reaches the margin of the plateau in the High Himalaya (Hodges et al. 2001).

Figure 3 shows our interpretation of the channel flow model, based on geological (Searle 1999a, b; Searle et al. 2003; Law et al. 2004; Jessup et al. 2006) and geophysical (Zhao et al. 1993; Nelson et al. 1996; Hauck et al. 1998) data from the Nepal and south Tibet Himalaya. The shaded part of the middle crust in Figure 3 represents the channel of high-grade gneisses, migmatites and leucogranites that, during the Miocene, acted as a ductile layer sandwiched between the brittle deforming upper crust and a rigid, anhydrous, granulite-facies lower crust.

Deep crustal structure from the INDEPTH profile

The subsurface structural interpretation beneath south Tibet is based entirely on deep seismic reflection profiling from project INDEPTH (Zhao et al. 1993; Nelson et al. 1996; Brown et al. 1996), combined with broadband earthquake and magnetotelluric data (Kind et al. 1996; Wei et al. 2001). The crustal structure of southern Tibet shows several reflectors, which have been successfully matched to the major faults of the Himalaya to the south, notably the STD and the Main Himalayan Thrust (MHT; Hauck et al. 1998; Alsdorf et al. 1998). The MHT is the basal detachment, which dips at about 10° north and progressively ramps up-section to the south, with both the MCT and Main Boundary Thrust (MBT) splaying off it at depth.

Magnetotelluric data (Wei et al. 2001) suggest that an electrically conductive layer at 15–20 km depth corresponding to zones of seismic attenuation, may reflect partial melts and/or aqueous fluids in the middle crust of southern Tibet (Nelson et al. 1996; Alsdorf & Nelson 1999). The ‘bright spots’ imaged on the INDEPTH profile beneath south Tibet have been interpreted by some as leucogranitic melts forming today, at a similar structural horizon and depth to the High Himalayan leucogranites which crystallized during the Miocene and were extruded southward to their present position along the High Himalaya (Searle et al. 1993, 1997, 2003; Searle 1999a, b; Gaillard et al. 2004). It has further been suggested that the ages of crustal melt leucogranites in the footwall of the STD decrease to the north (Wu et al. 1998) and that the Greater Himalayan crystalline rocks, bounded by the STD above and the MCT below, have been effectively extruded southwards from beneath the middle crust of south Tibet (Grüjic et al. 1996, 2002; Searle 1999a; Searle et al. 2003; Searle & Szulc 2005).

Several factors such as radiogenic heating, crustal shortening and concomitant thickening could have induced the high temperatures at these mid-crustal levels beneath southern Tibet. Francheteau et al. (1984) measured unusually high heat flow in the lake sediments beneath Yamdrock Tso SE of Lhasa (Fig. 1), which they attributed to the presence of a magma body at 10–20 km depth. Satellite data reveal a magnetic low over Tibet indicating hot crust, and the deeper levels of the high-conductivity zone have been interpreted as a zone of in situ partial melting (Alsdorf & Nelson 1999). Seismological data show that relatively fast (cool) upper mantle extends from the Himalaya northwards to roughly the centre of the Tibetan plateau around the latitude of the Bangong suture (Jin et al. 1994; Owens & Zandt 1997; Kosarev et al. 1999). This has been interpreted as the northern limits of cold Indian lithosphere underthrust beneath the plateau.

Crustal structure of the Everest transect

The surface structure of the GHS in the Everest transect is shown in Figure 4, together with U–Pb ages of metamorphic rocks and leucogranites (Simpson et al. 2000; Viskupic & Hodges 2001; Searle et al. 2003) and Th–Pb ages of Murphy & Harrison (1999) from Rongbuk, and Catlos et al. (2002) from the MCT zone. The structure of the profile has been presented previously (Searle 1999a, b, 2003; Searle et al. 2003; Law et al. 2004; see also Jessup et al. 2006, fig. 3) and only the main points pertinent to the discussion on channel flow need be repeated here. The upper part of the GHS slab in the Everest region is marked by two low-angle normal faults, which merge towards the north along the Rongbuk and Kharta valleys into one major shear zone (Searle et al. 2002, 2003). The structurally lower Lhotse detachment is a ductile shear zone that separates sillimanite-grade gneisses with abundant leucogranite sills beneath from lower amphibolite...
Fig. 4. (a, b) Cross-section through the Greater Himalayan sequence along the Everest transect from the South Tibetan detachment along the Rongbuk valley in Tibet to the Main Central Thrust zone in Nepal showing all U–Th–Pb age data from this profile. MCT, Main Central Thrust; LD, Lhotse detachment; QD, Qomolangma detachment.
Fig. 4. (Continued)
(Jessup et al. 2004)–greenschist-facies pelitic and calcareous rocks (Everest series) with few, if any, leucogranites above. The structurally higher normal fault, the Qomolangma detachment, separates Everest Series metapelite beneath from unmetamorphosed Ordovician sedimentary rocks above. Beneath the Lhotse detachment a coherent leucogranite sill, up to 3.5 km thick, composed of tourmaline + biotite + muscovite ± garnet leucogranite, extends at least from the Makalu massif westwards to Cho Oyu across the lower edifice of Mount Everest and the Nuptse–Lhotse massif (Searle 1999a, b, 2003). Numerous leucogranite sills extend from the Everest massif north for at least 57 km along the Kharta and Rongbuk valleys as parallel sheets within the gneisses. Early leucogranite sills are parallel to the ductile fabrics within the gneisses, whilst a few later dykes cross-cut the fabric, but are themselves truncated by the brittle Qomolangma detachment above (Murphy & Harrison 1999; Searle et al. 2002, 2003).

Most of the high peaks in the Khumbu Himalaya (including the base of Cho Oyu, Gyachung Kang, Nuptse, Everest, and most of the peaks Chomolonzo, Makalu, Baruntse, Ama Dablam, Kangteiga, Tramserskou) are composed of flat-lying leucogranite sills within the high-grade sillimanite gneisses (Searle, 1999a, b; Weinberg & Searle 1999; Visona & Lombardo 2002). Some sills are very narrow (<1 m), whilst others reach 1–2 km in thickness. A few interconnecting dykes link the foliation-parallel sills, but the rarity of discordant dykes implies that magma transport was dominantly along the foliation planes. In the Everest–Lhotse–Nuptse massif, the leucogranites lie above the north-dipping Khumbu Thrust (Searle 1999a, b), well exposed along the south face of Nuptse and Lhotse, which has transported these rocks south over sillimanite gneisses that contain very few leucogranites.

Along the Everest–Lukla transect in Nepal, sillimanite remains the stable aluminium silicate over the 45 km horizontal width (and >15–20 km structural depth) of the GHS, whereas the extruding slab of middle crust had a high-temperature core during the early Miocene. Although pelites are dominant throughout this section, there are numerous horizons of calc-silicates and K-feldspar augen gneisses throughout the whole sequence down as far as the MCT zone. We apply the original definition of the MCT, and map it as the main zone of high strain separating rocks metamorphosed during the Tertiary above (Phaplu augen gneiss and structurally higher gneisses), from underlying rocks largely unmetamorphosed during the Himalayan orogeny (Lesser Himalaya). This corresponds to the lower MCT-1 position of Arita (1983) and not to the structurally higher ‘MCT’ (north of Karikhola, south of Lukla; Catlos et al. 2002). Thus, rocks previously described as ‘upper Lesser Himalayan crystallines’ (e.g. Catlos et al. 2002; Vannay et al. 2004) at kyanite or sillimanite grade that have P–T conditions similar to rocks higher up the section throughout the GHS slab, are here considered part of the GHS, above the MCT. The entire base of the GHS is a large-scale shear zone, corresponding to the zone of inverted metamorphic isograds, but the precise distribution of strain has yet to be determined by a detailed kinematic investigation.

Restoration of the Everest Himalaya

Figure 5 is a schematic restoration of the Everest transect through the GHS, showing the trajectories of a number of the major faults. The restoration shows the position of the two major low-angle normal faults that cut through Everest itself: the upper Qomolangma detachment which flattens out along the top of the Ordovician ‘Yellow Band’; and the lower Lhotse detachment, which bounds most of the large leucogranite sills (Searle 1999a, b; Searle et al. 2003). The Everest granites must have been sourced within the Proterozoic black shales that formed the protolith of the sillimanite gneisses comprising Unit 1 in the GHS, also called the Barun gneiss (Pognante & Benna 1993; Lombardo et al. 1993). We discard the previous ‘stratigraphic’ nomenclature of the GHS (e.g. Le Fort 1981; Colchen et al. 1986) because we believe that it oversimplifies the complex metamorphic lithologies and structure of the GHS (Searle & Godin 2003). For example, there are numerous horizons of augen gneiss (previously called Formation III) throughout the GHS from the base of the slab (Ulleri and Phaplu augen gneisses) up to the highest levels beneath the leucogranites (e.g. Pumori augen gneiss; Searle 2003). Likewise, calc-silicate bands (previously Formation II) and less common amphibolites are present at several levels throughout the GHS sequence.

The Namche K-feldspar orthogneisses, together with similar orthogneisses higher up the slab near Pumori, were intruded into the sillimanite gneisses and the metamorphosed Cambrian–Ordovician limestones–calc-silicates. The orthogneisses contain biotite, sillimanite, cordierite, quartz, plagioclase and K-feldspar. Monazite–zirconite thermochronology suggests that the original protolith crystallized at, or before, 466 Ma; metamorphism began at, or before, 28.37 Ma, and anatectic melting occurred at 25.4–24.75 Ma (Viskupic & Hodges 2001). The Namche orthogneisses are therefore interpreted as part of the suite of late Cambrian–Ordovician graniteoids that intruded the Indian plate (Miller et al. 2001),
Fig. 5. Schematic restored section across the Greater Himalayan sequence (GHS) in the Everest Himalaya. Shaded area represents the protolith rocks of GHS mid-crustal channel. Unit I is dominantly pelitic and interpreted as metamorphosed Proterozoic shales. Minor layers of calc-silicate and amphibolite occur throughout this unit. Unit II is dominantly calc-silicate and marble, interpreted as metamorphosed Cambro-Ordovician carbonates. K-feldspar orthogneiss includes Proterozoic augen gneisses (Phaplu gneiss) and metamorphosed Cambro-Ordovician gneisses (Namche gneisses). The protolith source of the Everest leucogranites is interpreted as the Proterozoic black shale sequence beneath the Lhotse Detachment (LD) and above the Khumbu thrust (KT; Searle 1999a, b). The Qomolangma Detachment (QD) places unmetamorphosed Ordovician–Silurian sedimentary rocks on top of Ordovician greenschist-facies marbles (Yellow Band) and Everest Series pelites. The QD and LD merge to the north, off this section.
and have subsequently been metamorphosed during the Oligocene Himalayan event, resulting eventually in Miocene partial melting.

Farther south within the GHS, another prominent level of K-feldspar augen gneisses crops out near the village of Phaplu (Fig. 4). These gneisses contain K-feldspar, quartz, plagioclase, biotite and hornblende, with a metamorphic assemblage including muscovite, biotite, garnet and sillimanite. A Th–Pb monazite age of 1219 ± 9 Ma is interpreted as the age of the protolith (Catlos et al. 2002), significantly older than the protolith ages of K-feldspar augen gneisses higher in the section. Restoration of the section strongly suggests that the protolith of the sillimanite-grade metapelites lying structurally above the Phaplu augen gneiss was the Proterozoic Haimanta Group shales and greywackes (Fig. 5).

A final requirement from the restored section through the GHS is that the footwall and hanging wall rocks across the MCT must match up. The GHS has a sedimentary provenance, dominantly of late Proterozoic age, based on detrital zircon ages of 0.8–1.0 Ga, whereas Lesser Himalayan zircons have older 1.87–2.60 Ga ages (Parrish & Hodges 1996). However, there is no evidence of a Palaeozoic suture and no evidence of any pre-Himalayan displacement on the metamorphic schistosity and have a weak fabric. Set 2 leucogranite dykes cut the metamorphic fabric, and both sets 1 and 2 are cut by the later set 3 dykes. Set 3 dykes are undeformed garnet þ tourmaline þ muscovite þ biotite leucogranites that cross-cut the ductile fabrics. No leucogranites cut across the brittle STD fault anywhere along the Himalaya, despite earlier reports to the contrary (Manasl granite (LeFort 1981; Harrison et al. 1999), and the Rongbuk granite (Burchfiel et al. 1992; Hodges et al. 1998)). Along the East Rongbuk glacier the Lhotse detachment truncates foliation in the footwall gneisses (Fig. 8a). The Yellow Band calc-silicates thin towards the north, but form a prominent layer between the high-grade gneiss below and the unmetamorphosed sediments above (Fig. 8b and c). Rocks above the Qomolangma detachment are sedimentary and composed dominantly of calcite with minor amounts of dolomite, quartz and illite-muscovite (Fig. 8d). Rocks beneath the Lhotse detachment show ductile deformation with lower set 1 leucogranite sills folded in with the pelitic and calc-silicate gneisses (Fig. 8e). Structurally higher in the section, flattening fabrics are common with the pelitic and calc-silicate gneisses (Fig. 8f). Microstructures within the zone indicate top-to-north kinematics, compatible with extrusion of footwall gneisses to the south beneath a static

**STD as a passive roof fault to the extruding channel**

The upper boundary of the GHS channel is marked by the low-angle normal faults of the South Tibetan Detachment (STD). This detachment is an orogenic-scale fault present along 2000 km of the High Himalaya from the Zanskar region of NW India to Arunachal Pradesh in the east. It places the unmetamorphosed or anchizone sedimentary rocks of the Tethyan zone above the high-grade metamorphic rocks, migmatites and leucogranites of the Greater Himalayan Series. The STD zone includes a structurally lower ductile shear zone, the Lhotse detachment, and a higher-level brittle fault, the Qomolangma detachment. In the Everest transect massive leucogranite sills occur below the structurally lower Lhotse detachment (Searle 1999a, b, 2003). The two shear zones merge towards the north to form one very large high-strain zone with leucogranite sills and ductile shear decreasing structurally upwards towards the Tethyan sedimentary rocks (Fig. 6a). Along the Rongbuk and Kharta valleys, thin leucogranite sills are also present up into the lower-grade rocks below the Qomolangma detachment (Fig. 6b). This suggests that the STD system cuts progressively down-section to the north as shown on the restoration in Figures 2c and 5.

Leucogranites in the GHS are mostly sill complexes intruded parallel to the foliation, with less common late dykes that cross-cut the ductile fabrics, but are truncated by the brittle fault above. Searle et al. (2003) defined three major sets of leucogranite sills and dykes, which are broadly comparable across the entire region. These three sets are illustrated from outcrops in the Hermit’s Gorge, above the Rongbuk Base camp in Tibet (Fig. 7a,b). Set 1 leucogranites are parallel to the ductile foliation, folded with the metamorphic schistosity and have a weak fabric. Set 2 leucogranite dykes cut the metamorphic fabric, and both sets 1 and 2 are cut by the later set 3 dykes. Set 3 dykes are undeformed garnet þ tourmaline þ muscovite þ biotite leucogranites that cross-cut the ductile fabrics. No leucogranites cut across the brittle STD fault anywhere along the Himalaya, despite earlier reports to the contrary (Manasl granite (LeFort 1981; Harrison et al. 1999), and the Rongbuk granite (Burchfiel et al. 1992; Hodges et al. 1998)). Along the East Rongbuk glacier the Lhotse detachment truncates foliation in the footwall gneisses (Fig. 8a). The Yellow Band calc-silicates thin towards the north, but form a prominent layer between the high-grade gneiss below and the unmetamorphosed sediments above (Fig. 8b and c). Rocks above the Qomolangma detachment are sedimentary and composed dominantly of calcite with minor amounts of dolomite, quartz and illite-muscovite (Fig. 8d). Rocks beneath the Lhotse detachment show ductile deformation with lower set 1 leucogranite sills folded in with the pelitic and calc-silicate gneisses (Fig. 8e). Structurally higher in the section, flattening fabrics are common with thin leucogranite sills showing boudinage structures (Fig. 8f).
Fig. 6. (a) Northernmost outcrop of the South Tibetan (STD) along the Rongbuk valley in Tibet, approximately 38 km north of the summit of Everest. Massive cliffs of layered leucogranite (Lg) pass up into ductile deformed gneisses and calc-silicates with fewer leucogranite sills. Metamorphic grade decreases up-section to the Qomolangma Detachment, which places unmetamorphosed limestones and shales above the Greater Himalayan sequence (GHS) metamorphic rocks and leucogranites. (b) STD shear zone approximately 60 km north of Everest along the Kharta valley showing upward decrease in volume of leucogranite. The entire section is a large-scale shear zone with ductile strain decreasing upwards towards the base of the Tethyan sedimentary sequence. Cliff section is approximately 80 m high.
Fig. 7. (a, b) Hermit’s Gorge, 5430 m altitude, east of Everest Base camp, Rongbuk glacier. Early, set 1 leucogranite sills are parallel to, and folded in with, the metamorphic schistosity. Set 2 dykes cut the metamorphic schistosity and some of the folds. Both sets 1 and 2 sills and dykes are cut by later, vertical leucogranite dykes (set 3) that abruptly cut folds, metamorphic schistosity and ductile fabric in the gneisses. These late dykes have little or no fabric within, but are cut by the flat-lying brittle Qomolangma Detachment at higher structural levels. Cliff section in (b) is approximately 20 m high.
estimated at least 100 km of southward motion may have occurred in the footwall GHS rocks along the Rongbuk valley, north of Everest. This supports the channel flow model of ductile extrusion of the middle crust. Geochronology suggests that localized partial melting may have triggered channel flow, and that the rapid cooling of the entire GHS slab may have caused channel flow to end abruptly at around 16–14 Ma (Godin et al. 2006a). $^{40}$Ar/$^{39}$Ar muscovite ages, interpreted as a proxy for timing of the ductile to brittle transition, across the entire slab also lend support to this. It is possible, however, that the GHS channel may still be active to the north, at depth beneath south Tibet, and the rocks along the High Himalaya record the 20–16 Ma palaeo-channel.

**Timescales of metamorphism, melting and channel flow**

Figure 9 shows geochronological data for the Everest transect plotted on a temperature versus time diagram, with the known closure temperatures for individual minerals in each system on the right. U–Pb dating of monazites growing in equilibrium with garnet, kyanite and sillimanite–cordierite shows that peak metamorphism spanned at least 14 million years, from 32.2 ± 0.4 Ma to 17.9 ± 0.5 Ma (Simpson et al. 2000; Searle et al. 2003). Large-scale melting to form the Everest leucogranites occurred between 21.3 and 20.5 Ma (Simpson et al. 2000) corresponding to the timing of peak high-temperature sillimanite–cordierite-grade metamorphism along the top of the GHS slab. Rapid cooling, and therefore rapid exhumation and probably a period of high erosion rates, followed immediately after crustal melting.

If crustal thickening and regional metamorphism started soon after the India–Asia collision at c. 50 Ma then approximately 15–20 million years were required to thicken the crust and metamorphose it to kyanite–sillimanite-grade P–T conditions. Temperatures within the GHS slab must then have remained very high for at least 16 million years (from 32 to 16 Ma), with sporadic episodes of

![Figure 9. Temperature–time plot showing all geochronological data from the Everest region. Closure temperatures of individual minerals in various isotopic systems are shown on the right-hand side. Monazite ages are interpreted as timing the growth of metamorphic assemblages at or close to the peak, and in granites as timing the crystallization of the granite. Sources of data are Hubbard & Harrison (1989), Bergman et al. (1993), Hodges et al. (1998), Murphy & Harrison (1999), Hubbard & House (2000), Simpson et al. (2000) and Searle et al. (2003). The steep part of the cooling path at c. 16 Ma is interpreted as the timing of rapid motion of STD footwall rocks as they were exhumed towards the surface.](image-url)
crustal melting resulting in partial anatexis of the Namche migmatite at 25 Ma (Viskupic & Hodges 2001), and the main pulse of crustal melting forming the Everest—Nuptse—Makalu leucogranites at 21 Ma (Simpson et al. 2000). Temperatures within the slab remained high enough for leucogranite generation until 16 Ma when the youngest granite dykes were formed in the Rongbuk area (Murphy & Harrison 1999). All 40Ar/39Ar and K–Ar muscovite and biotite ages of GHS metamorphic rocks are older than 14 Ma, suggesting that by that time, the entire GHS slab had cooled to below 350–300°C as the channel cooled, ductile strain had turned to brittle faulting, and thrusting propagated down-structural-section into the Lesser Himalaya.

The timing of metamorphism and melting in the Everest transect closely matches other profiles along the Himalayan chain. In Zanskar, the GHS was metamorphosed to kyanite grade at 33–31 Ma as evidenced by Sm–Nd dating of garnet (Vance & Harris 1999) and U–Pb dating of metamorphic monazite (Walker et al. 1999). Temperatures remained high during active crustal shortening and thickening for at least a further 12 million years until 22–20 Ma, when peak sillimanite-grade metamorphism across the GHS slab was coincident with widespread migmatization and crustal melting producing tourmaline, garnet and muscovite-bearing leucogranites (Noble & Searle 1995; Dezes et al. 1999; Walker et al. 1999). Protophilths of the GHS metamorphic rocks are Proterozoic to early Mesozoic sediments and Permian volcanic rocks (Panjal Traps), partly lateral equivalents to the unmetamorphosed rocks in the Tethyan zone to the north (Searle 1986; Walker et al. 2001).

Th–Pb monazite ages within the sillimanite-grade gneisses of the GHS from Pheriche south to Lukla in the Everest transect (Fig. 4) are all within the range 25.3–20.7 Ma (Catlos et al. 2002). Only in the far south around the village of Karikhola do the Th–Pb monazite ages become younger, from 15.2 to 11.8 Ma. The monazites are inclusions within garnet that was apparently growing at 530 ± 50°C (Catlos et al. 2002), a temperature similar to that required for monazite growth in pelite (525°C; Smith & Barreiro 1990). Catlos et al. (2002), following Harrison et al. (1997), interpreted these ages as representing late Miocene reactivation of the MCT. Even younger Th–Pb monazite ages (c. 7–3 Ma) have been dated in rocks from a similar position along the MCT in central Nepal and Garwhal (Harrison et al. 1997; Catlos et al. 2002).

Young Th–Pb monazite ages from the MCT present a conundrum for interpreting the tectonic evolution of the MCT. 40Ar/39Ar hornblende ages of c. 21 Ma from the upper part of the MCT zone south of Lukla, record the timing of cooling through the 500°C isotherm (Hubbard & Harrison 1989). Along most of the Greater Himalaya, 40Ar/39Ar and K–Ar muscovite and biotite ages, which record timing of cooling through 300–350°C isotherms, are always older than 14 Ma across the entire slab (Godin et al. 2001). How is it possible, therefore, that 530°C metamorphic temperatures in the MCT samples with young monazite ages do not reset the 40Ar/39Ar and K–Ar systematics? Recent studies have shown that monazite and even zircon are not only mobile, but can crystallize at temperatures below 350°C in slates (Rasmussen et al. 2001; Dempster et al. 2004). Rubin et al. (1993) showed that zirconium can be a mobile phase during hydrothermal alteration. Young matrix monazites within the MCT zone could therefore have grown during post-metamorphic hydrothermal alteration, but it is harder to explain the monazite inclusions within armouring garnet, unless minute cracks in the garnet allowed access of fluids during hydrothermal activity. The MCT zone has been, and still is, a conduit for boiling fluids, with numerous hot springs located along the thrust (the location of many villages called Tatopani – Nepali for ‘hot water’). If the young monazite ages along the MCT zone do record a period of elevated temperatures, it is a very restricted metamorphism solely along the MCT and has no record higher up throughout most of the GHS. The Th–Pb monazite ages south of Kharikhola span 15.2–11.8 Ma (Catlos et al. 2002) and may indicate the youngest period of garnet growth in the Phaplu gneiss, immediately prior to cooling below the ductile–brittle transition.

Discussion

In the channel flow model (Fig. 3) for the Himalaya and Tibet, hot, weak, partially molten and thickened crust flows from the Tibetan plateau towards the plateau margins in response to lithostatic pressure gradients between the high elevation and thick crust of Tibet and the low elevation and normal thickness of the Indian plate crust. Channel flow processes have been incorporated into both linear viscous models and coupled thermal–mechanical models (Royden et al. 1997; Beaumont et al., 2001, 2004). Previously, the deformation has been thought of in terms of a ‘jelly sandwich’ model in which the upper crust and upper mantle are strong, and are separated by a weak and ductile lower crust (Jackson et al. 2004). However, in the Himalaya it is the middle crust that is hot, weak and flowing, whereas the
granulitic lower (Indian) crust is relatively strong. The lower crust of Tibet is not exposed so we have to rely on gravity data (Jin et al. 1994), seismic data (Nelson et al. 1996) and volcanic xenoliths (Hacker et al. 2000) to interpret the structure.

The INDEPTH seismic experiments in southern Tibet indicate the presence of a fluid-like layer of middle crust (Nelson et al. 1996; Brown et al. 1996) interpreted as containing a degree of partial melt. This mid-crust layer can be followed south to link with the GHS and the bounding shear zones of the MCT and STD. We suggest that the crust beneath the Himalaya and southern Tibet must be rheologically layered, with a major decoupling of the upper, middle and lower crust (Fig. 10).

The geotherm is not linear with depth but shows a nipping of the upper, middle and lower crust (Fig. 10). The geotherm is not linear with depth but shows a peak temperature at high levels in the middle crust, corresponding to the location of the 'bright spots' imaged by magnetotelluric data beneath the Lhasa block (Kola-Ojo & Meissner 2001). The lower crust is composed of the more rigid, dry granulites of the underthrusting Indian shield. Earthquakes in southern Tibet are abundant in the upper crust down to c. 18 km and again in the lowermost crust at 60–80 km, but not in the middle crust (Jackson et al. 2004), in agreement with the non-linear isotherm shown in Figure 10. There is however, some dispute as to whether the deep earthquakes beneath the Himalaya and south Tibet are actually in the lower crust or in the upper mantle (Chen & Yang 2004).

The obvious question now is why do melting temperatures exist at relatively shallow levels of the thick crust beneath the south Tibetan plateau? The restoration of the Everest transect (Fig. 5) shows that the leucogranites were derived from the Proterozoic shales along a stratigraphic horizon that must have formed the protolith of the migmatites and leucogranites. The heat source for melting cannot have been frictional heating along the MCT because the site of anatexis is along much higher crustal levels than the MCT, and in situ melting is not seen along the MCT. We suggest that the shallow level of high temperatures producing leucogranite melts must have been the consequence of a high concentration of radioactive heat-producing elements at that unique stratigraphic horizon in the crust. The concentration of U, Th and K could have been the result of sedimentary processes within the late Proterozoic basin that accumulated debris off the eroding Archaean mountain belt of peninsula India.

Another intriguing question remaining is: why did channel flow operate within a relatively narrow time span (24–17 Ma), and why did it apparently end at c. 16 Ma? High temperatures, close to leucogranite melting temperatures (680–720°C), are required to maintain ductile channel flow. Numerous 40Ar/39Ar and K–Ar cooling ages on micas within the GHS are almost entirely older than 15–14 Ma (see Searle & Godin (2003) and Godin et al. (2006a,b) for a summary). Temperatures in the GHS metamorphic rocks and leucogranites must therefore have been below 350°C by 15–14 Ma, and the whole channel must have cooled through the ductile–brittle transition. This fact makes the subsequent late Miocene–Pliocene monazite ages from the MCT zone (Harrison et al. 1997; Catlos et al. 2002) hard to interpret as indicators of a regional metamorphic event.

It is possible that the Himalayan mid-crustal channel could still be active today (Hodges et al. 2001), although there is no seismic evidence for recent activity on the MCT or the STD. The GHS along the High Himalaya corresponds to the zone of highest topography, although some of the highest peaks along the range (e.g. Dhaulagiri–Annapurna...
peaks) are composed of the lower Tethyan sedimentary rocks above the STD. Hurtado et al. (2001) suggested that one branch of the STD north of Annapurna may have been active during the Quaternary. The MCT is coincident with prominent knick-points in Himalayan river profiles (Seeber & Gornitz 1983), but the hanging wall uplift could be due either to MCT reactivation (Harrison et al. 1997, 1998) or to uplift above a ramp along the deeper and younger Main Himalayan thrust (MHT) fault (Cattin & Avouac 2000). If the high topography of the Himalaya is due to channel flow during the period 24–16 Ma, why does the range remain so high today? One solution would be to have the Miocene channel preserved along the Himalaya today, as recorded by the geological structure and pressure—temperature—time conditions, as an outer rind or crust surrounding the hot, active channel presently at mid-crustal depths beneath the southern part of the Tibetan plateau (Fig. 11). This could be envisaged as analogous to a cooling Hawaiian-type lava flow with a hot, molten core surrounded by a cooling brittle carapace.

Conclusions
Field mapping, macro- and micro-structural observations, and thermal modelling all support the conclusion that the channel flow model is appropriate for the GHS. A slab of Indian plate middle crust, between 5 and 20 km thick, was at fairly uniform high temperatures (c. 600–700 °C) for a period of about 12–16 million years during the late Oligocene and early Miocene (between 30 and 17 Ma), sandwiched between a brittle-deforming upper crust (Tethyan sedimentary zone) and a rigid granulitic lower crust (subducting Indian shield). The mid-crustal channel was metamorphosed at sillimanite-grade temperatures and contained a zone of in situ early Miocene migmatization, up to 5–8 km thick. The GHS channel was bounded by two major ductile shear zones with thrust sense of shear along the base (MCT), and a normal sense of shear above (STD).

Structural geometry, thermobarometry and geochronology suggest that there was a mid-crustal layer composed of high-temperature sillimanite-grade rocks that were close to partial melting temperatures during the period 32–17 Ma. Migmatites and crustal melt leucogranites are widespread at the upper structural levels of this layer. U–Pb dating of metamorphic monazites crystallizing in equilibrium with garnet and kyanite, indicates that early metamorphism in the Everest area peaked at 32.2 ± 0.4 Ma (Simpson et al. 2000). Similar ages were obtained from Sm–Nd dating of garnets in the Zanskar and Garhwal regions in the western Himalaya (Vance & Harris 1999;
Prince et al. 2001). It appears that, although high temperatures were attained by 32 Ma, melting did not occur along the top of the GHS until about 24–21 Ma, at least in the Everest–Makalu region (Schärer 1984; Simpson et al. 2000). These ages suggest that it could have been partial melting that triggered channel flow and ductile extrusion. The timing of granite melting also coincides with the timing of increased cooling rates, erosion rates and exhumation of the GHS, and increased sedimentation rates in the Siwalik foreland basin. As the granites cooled, flow ended and lateral transport was achieved by brittle faulting along the outer margins of the channel, both below (MCT) and above (STD).

Melting resulted in several large leucogranite bodies that reach a maximum thickness of c. 3 km in the Makalu–Chomolonzon leucogranite body. Himalayan leucogranites contain tourmaline, garnet, muscovite and biotite and are pure crustal melts, derived from a highly radiogenic Proterozoic sedimentary source, corresponding to the Haimanta Group shales (or Cheka Group in Bhutan). The shallow level of crustal melting within the Himalayan crust must have been the result of a high concentration of heat-producing elements within the stratigraphic horizon of the Proterozoic black shales. The Himalayan granites were intruded as sill complexes within the sillimanite gneisses for large distances (over 70–100 km north of Everest) laterally along the foliation planes.

Microstructures within the GHS channel indicate that there was a significant component of penetrative vertical pure shear, as well as the dominant south-directed simple shear (Law et al. 2004; Jessup et al. 2006). However, despite the evidence for a reduction in the thickness of the channel through flattening, there is also good evidence that the active channel expanded with time, during a change from ductile to brittle deformation, as MCT zone thrusts propagated down-structural-section with time, and STD normal faults propagated up-section with time. Thrust faults along the base of the channel and normal faults along the top must have been synchronous, as previously suggested (Searle & Rex 1989; Hodges et al. 1993, 1996; Searle et al. 1997, 2003).

STD normal faults were active in a wholly compressional environment, and acted as a passive roof fault during crustal shortening and thickening in the footwall. Although pure shear flattening did compress the sequence, new material was constantly being fed into the channel from crustal subduction beneath the active MCT during the late Oligocene and Miocene. Folding and thrusting in the Tethyan sedimentary rocks above the STD occurred earlier than shortening and thickening in the GHS beneath, although some breakback thrusting and north-directed backthrusting did occur late in the sequence (Searle 1986; Corfield & Searle 2000; Murphy & Yin 2003). The STD normal faults do not relate to wholesale crustal thinning, whole crustal extension or topographic collapse.

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References


