Dominant simple-shear deformation during peak metamorphism for the lower portion of the Greater Himalayan Sequence in West Nepal: new implications for hybrid channel flow-type mechanisms in the Dolpo region.

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ABSTRACT

I conducted new vorticity and deformation temperatures studies to test competing models of the exhumation of the mid-crustal rocks exposed in the Dolpo region (west Nepal). My results indicate that the Main Central Thrust is located ~5 km structurally below the previous mapped locations. Deformation temperature increasing up structural section from ~450°C to ~650°C and overlap with peak metamorphic temperature indicating that penetrative shearing was responsible for the exhumation of the GHS occurred at “close” to peak metamorphic conditions. I interpreted the telescoping and the inversion of the paleo-isotherms at the base of the GHS as produced mainly by a sub-simple shearing (Wm = 0.88-1) pervasively distributed through the lower portion of the GHS. My results are consistent with hybrid channel flow-type models where the boundary between lower and upper portions of the GHS, broadly corresponding to the tectonometamorphic discontinuity recently documented in west Nepal, represents the limit between buried material, affected by dominant simple shearing, and exhumed material affected by a general flow dominates by pure shearing. This interpretation is consistent with the recent models suggesting the simultaneous operation of channel flow- and critical wedge-type processes at different structural depth.

Key words:

Great Himalayan sequence, quartz petrofabrics, vorticity of flow, deformation temperatures, hybrid channel flow mechanisms, West Nepal.

1. Introduction

The Himalayan belt has developed in response to continental collision between the Asian and Indian plates that initiated at *c.* 55-50 Ma (Hodges, 2000). Three main tectono-metamorphic units are classically recognized along the >2500 km east-west length of the orogen. From the bottom to the top they are: (1) the Lesser Himalayan (LH), (2) the Greater Himalayan sequence (GHS) and (3) the Tibetan Sedimentary sequence (TSS) (Fig. 1). The GHS, the strongly metamorphosed and anatectic core of the orogen, may be interpreted as a low-viscosity tabular body of mid-crustal rocks (Godin et al., 2006 and references therein) extruded southward in Miocene times beneath the Tibetan plateau between two parallel and opposite-sense crustal-scale shear zones: the Main Central thrust (MCT) at the base, and the South Tibetan Detachment system (STDs) at the top (Burchfiel and Royden, 1985; Hodges et al., 1993). The pre-/syn-shearing mineral assemblage documented within these crustal-scale shear zones indicates that the metamorphic grade increases toward the core of the GHS producing an inverted and a normal thermal gradient respectively on the top and on the bottom of the slab (Hodges, 2000 and references therein). In addition, thermal profiles estimated using both petrology- and microstructures/fabrics-based thermometers indicate that the metamorphic isograds are condensed (e.g. Law et al., 2011).

To document the deformation pattern recorded within the GHS, several kinematic and vorticity investigations, coupled with deformation temperatures estimates, have occurred over last decade across the entire belt from NW India to Bhutan (e.g. Grujic et al., 1996; Grasemann et al., 1999; Law et al., 2004; 2011; 2013; Larson and Godin, 2009; Langille et al., 2010; Larson and Cottle, 2014). Although horizontal extension and vorticity estimates collected across the GHS could be strongly biased by the criteria used to define the map position of the MCT (Searle et al., 2008 and references therein), published vorticity data document general shear flow (1>Wk>0) within the slab with a pure-shear component of flow slightly predominant within the core of the GHS whereas the simple-shear component seems to dominate at the top of the slab (e.g. Law et al., 2014). The lower boundary of the GHS records a general shear flow with a comparable contribution of simple and pure shearing. The associated crustal extrusion is compatible with Couette - Poiseuille velocity flow profile as assumed in crustal-scale channel flow-type models (see overviews by Grujic, 2006; Godin et al., 2006; Jessup et al., 2006; Larson and Godin, 2009; Jamieson and Beaumont, 2013). The same tectonic model can explain the thermal profiles and the telescoping of the paleoisotherms documented at the top and at the bottom of the GHS. At shallower structural levels (i.e. in the Mohr-Coulomb regime), however, alternative models involving ductile wedge extrusion (Grujic et al., 1996; Grasemann et al., 1999), tectonic (Leger et al. 2013 with references therein) or frictional (the critical taper model: Kohn, 2008) wedging, hybrid channel flow-critical taper model (see review of Cottle et al, 2015), high-temperature contractional shear zone (Carosi et al., 2010;

Montomoli et al., 2013) or development of a foreland-propagating fold-thrust belt (De Celles et al., 2001; Robinson et al., 2006), have also been used to explain the Oligo-Miocene exhumation of the GHS rocks and its thermal profiles.

In this study, the quartz c-axis petrofabrics, vorticity and deformation-temperature studies are integrated with microstructures and metamorphic studies to individuate a new location of the MCT and to document the spatial distribution of ductile deformation patterns across the lower portion of the GHS exposed in the Chaudabise river valley in western Nepal (Figs. 1, 2, 3 and 4). Deformation temperatures and vorticity of flow profiles are then compared to those obtained in the adjacent valleys to test competing models responsible for the exhumation of the GHS and of the inversion and telescoping of the paleoisotherms documented at its base. My results re-locate the MCT and re-evaluate the operation of the channel-flow-type mechanisms in the Dolpo region (West Nepal).

2. Previous geological studies in the Jumla region

The first geological studies in the Jumla region were carried out during the 1960's and 1970's by Franks and Fuchs (1970), Fuchs (1977) and Arita et al. (1984). According to these authors, low-grade metasedimentary rocks (quartzites, limestone and phyllite), locally preserving sedimentary structures and rare meter-thick amphibolitic bodies crop out around the Jumla village and in the westernmost 10-12 km of the Chaudabise river valley (i.e. Chail Nappe of Franks and Fuchs (1970) and Fuchs (1977); Jumla Formation -Gulwa-Jumla zone, Midland meta-sedimentary rocks of Arita et al., 1984; Fig. 2). While these authors broadly agreed on the lithostratigraphy, their tectonic interpretations for the footwall of the MCT differ. (Fig. 2). Fuchs and co-workers, suggested the presence of thrusts, located a few kms west of Talphi (Fig. 3), that duplicated the low-grade metasedimentary sequence (Fig. 2; i.e. broadly correspond to the Ramgarh thrust of DeCelles et al., 2001 and Pearson and DeCelles, 2005), whereas Arita et al. (1984) documented in the same location, the presence of late WNW-ESE trending folds. Although Fuchs and co-workers focused on the lithostratigraphy, Arita et al. (1984) focused more on the spatial distribution of shear zones. They defined a MCT zone (MCTz in Fig. 2; Fig. 3) as a ~ 700-800 m-thick mylonitic shear zone made of sheared chlorite ± garnet -bearing schists, phyllites, quartzites and amphibolites sandwiched between highly-deformed garnet + kyanite ±sillimanite-bearing gneisses and migmatites of their Himalayan Gneiss (HGz in Fig. 2), and the weakly-deformed quartzite and phyllite of the Midland meta-sediments (GJz in Fig. 2). According to these authors, the MCT zone is bounded by two "thrusts" called MCTI and MCTII (Figs. 2, 3 and 4) and recorded a metamorphic grade that increased upward reaching the greenschist facies conditions. In the easternmost outcrops of the Midland meta-sediments, Arita et al. (1984) documented a metamorphic grade moving up structural section toward the MCTz, that they interpreted as produced by shear heating.

Broadly in the same location of the MCTz defined by Arita e al. (1984), Fuchs and co-workers defined the Lower Crystalline nappe (LCn in Fig. 2), as an assemblage of graphite schists, dark-grey garnet-bearing phyllites, schists and light quartzites that is topped by the augen gneiss and the garnet + muscovite + biotite-bearing paragneisses of the Upper Crystalline nappe (UCr in Fig. 2). According to these authors, the MCT is located at the base of the Lower Crystalline nappe (Figs. 2, 3 and 4). Both Arita et al. (1984) and Fuchs and co-workers described at ~ 10 km east of Jumla, NW-SE trending sub-vertical faults that lowered the western block. For this reason, they found augen gneisses (the Himalayan Gneisses of Fuchs and co-workers; the Upper Crystalline nappe of Arita et al., 1984) immediately eastward of Jumla village (i.e. the Jumla klippe of Arita et al. 1984; Figs. 3 and 4). More recently, Carosi et al. (2002, 2013) suggested that the low-grade metasedimentary sequence cropping out near Jumla belongs to the Lesser Himalaya (LH in Fig. 2). They also suggested that the MCT is located at the boundary between the migmatitic gneisses (Higher Himalayan Crystalline, HHC: Carosi et al., 2002; GHS: Carosi et al., 2013; Fig. 2) and a <5 km- thick sequence that they attributed either to the upper portion of the LH (the Lesser Himalayan Crystalline, LHC in in Fig. 2: Carosi et al., 2002, 2007) or to the Main Central Thrust zone (MCTz in Fig. 2: Carosi et al., 2013).

The easternmost accessible outcrops in the Chaudabise river valley document the emplacement of a leucogranite body that was firstly recognized by Arita et al. (1984). The granite emplaced at the boundary between GHS and the overlying deformed Tibetan sediments at ~25-23 Ma (BP in Fig. 3; Bura Buri pluton, Carosi et al., 2013). Personal mesoscopic observations carried out on the Bura Buri granite are in agreement with Carosi et al. (2013). The outermost portions of the pluton is composed of coarse-grain leucogranite with a well-developed magmatic foliation (Fig. 4) defined by alignment of undeformed cm-scale muscovite (1-10 cm in size), feldspar and biotite crystals. Moving farther to the east (i.e. at higher topographic levels in the inner exposed portion of the pluton), the granite is of fine to medium grain size with a homogeneous texture and contains abundant nodules of tourmaline (1-8 cm). Samples collected two valleys to the north of the Chaudabise river valley (Fig. 3) from north-easternmost glacial deposits suggest that these granites intrude both the high-grade gneisses of the GHS and the low-grade metasandstones and metalimestones of the lower part of the Tibetan Sequence.

Robinson et al. (2006) working 40-50 km to the northwest of the study area (Fig. 1), suggested that the metasedimentary sequence cropping out along the Chaudabise river valley represents the lower portion of the Lesser Himalayan sequence (LH; i.e. Lakharpata Group and the Syangia, Ranimata and Kushma formations) that was affected by thrust-related repetition and duplexing. In the Mugu Karnali river valley (~35 km N-NW of Jumla) these authors mapped the Ramgarh thrust (a first order thrust developed within the LH and carrying the Kushma Formation on to the Ranimata Formation) in the same position of the MCT as recently mapped by Montomoli et al. (2013) (compare Fig. 3 of Robinson et al., 2006 and Fig. 4 of Montomoli et al., 2013).

3. The study area: the Chaudabise river valley

The geology of the Chaudabise river valley, that extends eastward from Jumla village for about 35 km (Figs. 3 and 4), is slightly described by previous authors that studied this sector of the Himalayan belt. Fuchs (1977 and Arita et al. (1984), in fact, represent the geology of the Chaudabise river valley only in the easternmost part of their large-scale maps. Recently, Carosi et al. (2013) present a more detailed geological sketch map, but they focused mainly on the Bura Buri pluton (Fig. 3).

Mesoscopic descriptions on metamorphism and deformation made during my fieldwork contrast to previous documentation by Arita et al. (1984) and Fuchs (1977) and improve our knowledge of the tectono-metamorphic evolution of the GHS exposed in the valley (e.g. Fig. 2). I documented two major episodes of penetrative deformational (D1 and D2) overprinted by two weaker deformation events (D3 and D4) and by a metamorphism that increased up structural section from lower greenschist- to amphibolite-facies conditions. The S2 foliation represents the main structural element in the field area (Fig. 3).

Based on structures (compare the orientation of D2 structural elements in Fig. 3), stratigraphy and metamorphism, the Chaudabise river valley can be schematically divided in two sectors, representing different structural GHS domains (Fig. 3). The boundary between the two sectors is represented by a system of down-dip normal faults oriented ~NW-SE. The west sector of the valley, stretching ~12 km far to east of Jumla village, is characterized by numerous sub-vertical normal faults that cut both the metasedimentary sequence and the migmatitic gneisses (i.e. the Jumla klippe of Arita et al., 1984) (Fig. 3). In contrast to Arita et al. (1984), no sedimentary structures were documented in this sector of the valley. Close to Jumla, in fact, the S2 foliation is marked by quartz, muscovite and rare tourmaline with a complete absence of biotite. Where it is not marked by normal faults, the contact between the metasedimentary sequence and the structurally higher migmatites is ambiguous. Relicts of the D1 deformation phase (S1 foliation) are recorded mainly close to Jumla village. Elsewhere, they are completely transposed by the S2 foliation that locally shows a pervasive mylonitic fabric with S-C/S-C' foliation, foliation fish and σ-type porphyroclasts that point to a top-to-the S/SW sense of shear (Fig. 5a-b). The east sector of the valley is located east of the fault systems (Figs. 3 and 4). In this sector, the presence of extensive outcrops and the absence of faults led me to document a continuous rock sequence composed of decimetre- to meter-thick intercalated layers of white quartzite (Fig. 5c) and Grt-bearing micaschist, with occasional graphitic schist, metre-thick bodies of Grt-bearing amphibolite and rare marble (QP in Figs. 2, 3 and 4). In contrast with Arita et al., (1984) and Fuchs and co-workers no evidence of folds or/and tectonic repetitions was documented. Moving eastward, I documented a reduction of micaschist content, a concurrent increase of quartzite content and an increase in biotite and garnet grain size. This sequence is capped by ~500 m of Grt±Ky-bearing paragneisses (Pg in Figs. 2, 3 and 4). Gt+Bt+Ms+Ky±Sil-bearing migmatitic gneisses (Gn in Figs. 2, 3 and 4; Fig. 5d) have been documented above the Grt-bearing paragneisses for about 3-4 km toward the east. In this sector of the valley, S1 foliation is rarely preserved and usually completely overprinted by the D2 deformation phase that produced rare tight to isoclinal asymmetric F2 folds (Fig. 5e) and a pervasive mylonitic fabric. Within Grt-bearing micaschists north of Talphi, the mylonitic fabric contains S-C/S-C' mylonitic domains, whereas to the east it is marked by a pervasive and strongly developed S2 mylonitic foliation (Fig. 5c) and by well-developed mineral lineations (L2) indicated by streaking of muscovite and/or biotite aggregates, quartz ribbons and elonged quartz-feldspar aggregates. Kinematic indicators observed at both the meso-scale (S-C/S-C' foliations, bookshelf structures in cm-size feldspar crystals, garnet and feldspar grains with σ-type tails of recrystallized quartz) and at the microscopic scale (garnet and feldspar porphyroclasts with sigma-type tails of recrystallized quartz, muscovite and biotite, muscovite fish, S-C/S-C' foliations and quartz and feldspar grain shape preferred orientation; Fig. 6) indicate a top-to-the S/SW sense of shear. S2 foliation is marked by muscovite, biotite, feldspar and occasionally pre-/syn-kinematic garnet (Fig. 6a). In particular, within quartzite and quartz-rich micaschist and paragneiss, it is highlighted by elongate quartz and feldspar grains and thin layers of phyllosilicates with an increase in biotite grain size (from 200 to 700 μm) eastward (i.e. toward the base of the migmatitic gneisses) (Fig. 6b). Within the pelitic paragneisses and gneisses, the S2 mylonitic foliation is defined by lepidoblastic layers of muscovite and biotite crystals and by granoblastic layers of quartz, feldspar, garnet, kyanite and fibrolitic sillimanite (Fig. 6c-d).

As direct thermo-baric constraints are not available for the studied samples, I estimated metamorphism temperatures from the syn-kinematic D2 metamorphic mineral assemblages documented within quartzites and quartz-rich micaschist. The results indicate a broadly increase up structural section eastward, passing from the biotite zone (greenschist facies conditions: ~400-500 °C) up to the sillimanite+muscovite zone (amphibolite facies conditions: ~550-650 °C).

Finally, D1 and D2 structures are deformed by F3 folds with sub-vertical axial planes and variably trending fold hinges ranging from NNW-SSE to ENE-WSW in the Jumla and Talphi areas respectively, and by F4 folds with NW-SE trending axis and moderately dipping axial planes (Fig. 5f).

4. Ductile deformation patterns

To characterize the style of flow and the thermal profile recorded in the rocks exposed in the Chaudabise river valley, 48 samples have been collected at different structural positions (Fig. 3). To limit the biases introduced in the Qtz c-axis fabrics by phyllosilicates content and by differences in strain induced by differences in texture and bulk composition, 6 representative samples were selected for petrofabric and flow vorticity analyses: five impure quartzite (phyllosilicates content: 5-10 vol%; feldspar contents: 18-25 vol%) and one quartz-rich garnet micaschist (D38; phyllosilicates content: 30-35 vol%; feldspar content: 40-50 vol%). To also avoid biases in the data set being introduced by sampling in areas affected by tectonic repetition, these samples were exclusively selected from those collected along the west sector of the valley (Figs. 3 and 4). The number of samples selected and used in this work is comparable with the number of samples used to characterize the kinematic of flow along the Karnali river (Yakymchuk and Godin, 2012: 9 samples along a section of 55 km), 30-35 km east of the study area (Carosi et al., 2007: 4 samples) and the kinematic of flow, deformation temperatures and finite strain ratio along each of the 6 transects investigated in the Sutlej valley (Law et al., 2013: from 4 to 8 samples along each transect).

Microstructural, petrographic and quartz petrofabric data were collected from thin sections cut parallel to mineral lineation and orthogonal to the mylonitic foliation. The kinematic and petrographic results were analysed using structural coordinates defined with respect to the base of the migmatitic gneisses (Fig. 4).

4.1. Quartz petrofabrics

Analysis of quartz crystal-preferred orientation patterns in naturally deformed rocks is a powerful tool for understanding: (1) active slip mechanisms; (2) shapes of the finite strain ellipsoids; (3) strain magnitudes; (4) strain paths (e.g. increase in simple shear); (5) kinematics of flow (e.g. Fueten et al., 1991; Stünitz, 1991; Sullivan, 2008); (6) vorticity numbers associated with flow (e.g. Wallis et al., 1993; Law et al., 2004; Frassi et al., 2009) and, (7) deformation temperatures (e.g. Kruhl, 1998; Morgan and Law, 2004; Larson and Cottle, 2014). Observed relationships between quartz-crystal fabrics can, however, be strongly influenced by the presence of pre-existing crystal orientations or fabrics (Vissers, 1993), the growth of phyllosilicates along preferred crystallographic orientations (Hippertt, 1994) or by dynamic recrystallization due to grain-boundary migration (Heilbronner and Tullis, 2006). Parsing the key deformational information from samples with one or more of these issues involves great caution during the data interpretation and a careful choice during samples collection. Although these biases exist, quartz c-axis fabrics were commonly used to constrain the vorticity of the flow and deformation temperatures within the GHS along the entire Himalayan belt from NW India (e.g. Grasemann et al., 1999) to Buthan (Grujic et al., 1996). Most data were collected within the South Tibetan Detachment system in the Everest area (Law et al., 2011 with references therein) and across the base of the GHS in central (Larson and Godin, 2009; Larson et al., 2010; Larson and Cottle, 2014) and western Nepal (Yakymchuk and Godin, 2012), and in NW India (e.g. Law et al., 2013).

Quartz c-axis fabrics data for this study were measured manually using a universal stage mounted on an optical microscope and are presented in equal-area lower hemisphere projections (Fig. 7). They are mostly Type I (*sensu* Lister, 1977) asymmetric cross-girdle fabrics that indicate approximately plane-strain deformation conditions (Lister and Hobbs, 1980) (Fig. 7 and Supplementary Table 1). The fabric asymmetry, observed both in the internal fabric skeleton and in the contouring of the density distribution, increases up structural section toward the base of the migmatitic gneisses from sample D11 to D38. The observed fabric asymmetry confirms the top-to-the SW shear sense indicated by micro- and mesoscopic D2 kinematic indicators. The geometry of the quartz c-axis fabric skeletons is consistent with those obtained by numerical, experimental and natural studies conducted since the 1970's (see Law, 1990 with references therein). As a consequence, I interpret the measured L2 mineral lineation as being representative of the true stretching lineation oriented parallel to the principal extensional direction.

Opening angles of the c-axis fabrics (see Fig. 8a for definition) generally increase up structural section eastward. The poorly defined c-axis fabric in sample D38 is interpreted to reflect its greater phyllosilicates content (30-35 vol%; cf. Starkey and Cutforth, 1978).

4.2. Deformation temperatures

Deformation temperatures are used in the Himalaya to evaluate possible tectonic processes responsible for the extrusion / exhumation of the GHS (e.g. Grujic et al., 1996; Grasemann et al., 1999; Law et al., 2004, 2011, 2013 and references therein; Langille et al., 2010; Wagner et al., 2009; Yakymchuk and Godin, 2012; Larson and Cottle, 2014). I estimated, deformation temperatures using: (1) quartz and feldspar deformation microstructures (Hirth et al., 2001; Stipp et al., 2002a, b; Rosenberg and Stünitz, 2003), (2) opening angles of quartz c-axis fabrics (Kruhl, 1998; Law et al., 2004; Morgan and Law, 2004) and (3) distribution of quartz c-axis fabrics within the shear zone (Mainprice et al., 1986). These techniques all assume that dynamic recrystallization and dislocation creep occurred simultaneously at "peak" metamorphic temperatures while experiencing "average" geological strain rates. Consequently, they assume that microstructures and quartz petrofabrics are mainly controlled by temperature rather than strain rate, hydrolytic weakening or 3D strain-type geometry. For this reason, temperatures estimates should be used with caution. For a complete and exhaustive discussion about these caveats and their implications for inferring deformation temperatures see the review of Law (2014).

4.2.1. Quartz and feldspar deformation microstructures

Quartz grains in the study area generally have weak undulatory extinction, occasionally deformation bands, grain-shape alignments and lobate grain-boundaries microstructures (Fig. 9a-b). These features indicate intracrystalline deformation accompanied mainly by grain-boundary migration (GBM) recrystallization indicating a deformation temperature of at least ~500 °C (e.g. Stipp et al., 2002; Faleiros et al., 2010). In the westernmost sample (D11) in addition to GBM recrystallization microstructures locally straight grain boundaries and triple points between adjacent quartz grains indicative of recovery processes were also documented. Up structural section toward the east, the recrystallized quartz grain-size progressively increases. In the western part of the valley feldspar grains exhibit undulose extinction and weak grain-shape preferred orientation. In contrast, up structural section toward the base of the migmatitic gneisses, they show undulose extinction, deformation twins and lobate boundaries where adjacent to quartz grains (Fig. 9b), documenting the basis for an interpretation of a progressive rise in deformation temperatures from ~450 °C to ~600 °C from W to E (Rosemberg and Stüniz, 2003; Passchier and Trouw, 2006 and references therein).

Even if the different recrystallization mechanisms occurred simultaneously as a function of a variation strain rate versus position (Stipp et al., 2002b), and/or water content (Mancktelow and Pennacchioni, 2004), microstructures related to a later occurrence of Subgrain Rotation (SGR) and Bulging (BLG) recrystallization were found in sample D44 (see Fig. 3 for sample location). In contrast to the other samples, quartz grains show strong intracrystalline deformation (i.e. undulose extinction, deformation bands) and domains of small and elonged sub- and new grains (~ 60 to ~ 150 μm) oriented parallel to the mylonitic foliation (Fig. 9c) that locally produce core-and-mantle structures. In addition, sub- and new- grains documented along high-mobility grain boundaries that preserve dragging and pinning structures (Fig. 9d) indicate that GBM microstructures were overprinted by "slow" grain-boundary migration (BLG) and by elongate new grains produced by progressive misorientation of subgrains (SGR). As these microstructures are not found elsewhere along the studied section, they are interpreted as being produced during a localized shear event that occurred under deformation temperatures of <500 °C (Stipp et al., 2002). The transition from GBM to SGR/BLG, however, may be interpreted to result from a decrease in deformation temperatures, a decrease in water content, an increase of strain rate and/or a combination of these factors (see review of Law, 2014).

In same sample (D44), feldspar grains show deformation twins and rare undulose extinction that are consistent with deformation temperatures of 450–600 °C (Passchier and Trouw, 2006 and references therein). As microfractures are absent in the feldspar grains, it may be possible to assume that deformation temperatures were in the upper portion of this interval (i.e. close to 600°C). In addition, the presence of fish-shaped monomineralic feldspar grains, showing deformation twins, straight cusp and boundaries (Fig. 9d) oriented parallel to the main mylonitic foliation may indicate that deformation involved both dislocation and diffusional creep during high-grade metamorphic conditions (ten Grotenhuis et al., 2003). No evidence of a later shear event at lower temperatures has been documented in feldspar.

4.2.2. Fabric opening-angle thermometer

Assuming that deformation occurred during peak metamorphic temperature under plane strain conditions, Kruhl (1998) proposed a graphical approach to estimate the deformation temperature using the opening angle (O.A.) of quartz c-axis fabric (Fig. 8a). According to Kruhl (1998) the method introduces an error of ±50°C that reflects the range in metamorphic temperature obtained using petrology-based thermobarometers from the samples used for the original O.A. *vs.* temperature plot. Variation in strain rate and potential hydrolytic weakening are thought to be encompassed in the methodological error whereas the influence of 3D strain type and the recrystallization mechanisms on the opening-angle of the quartz c-axis population remain poorly understood and not encompassed by this fabric-based thermometer (see review by Law, 2014). Poorly defined quartz c-axis fabrics in the stereonet plots can introduce an additional source of error not considered in the development of in the Kruhl's thermometer**.** However, even though the contribution of these different biases in the error is not fully understood, this thermometer is commonly used worldwide: in the Himalayan region (e.g. Law et al., 2004, 2011, 2013; Larson and Godin, 2009; Langille et al., 2010; Larson and Cottle, 2014), in the Zagros belt (Faghih and Sarkarinejad, 2011; Faghih and Soleimani, 2015), in NW Scotland (Thigpen et al., 2010, 2013) and in the Sardinia island (Frassi et al., 2009).

In the specific case of the Chaudabise river valley, the microstructures and petrofabric data documented in the selected samples validate the assumption made by Kruhl (1998). Plane strain deformation is confirmed by the approximately Type I (*sensu* Lister, 1977) asymmetric cross-girdle fabrics documented in all samples (Fig. 7). The simultaneity of deformation shearing event (D2 shearing) and peak metamorphism is confirmed by the broad agreement existing between the obtained deformation temperatures and the metamorphic temperatures estimated using both S2 syn-kinematic mineral assemblage (see Section 3) and classical thermo-barometry on samples collected at analogous structural levels in the adjacent valleys (T: ~600 ±40 °C: Carosi et al., 2010; T: ~638 ±70 °C, Yakymchuk and Godin, 2012; T: ~682 ±42 °C, Montomoli et al., 2013).

To quantify the error introduced during the measurement of opening angle, O.A. were measured using two methods. The first method considers the intersections between the primitive circle and the higher class of density contour, and the second used the interpolation of c-axis projections close to the primitive circle (i.e. the trailing edge; see Supplementary Table 1 and Supplementary Fig. 1). Using this approach, each opening-angle estimate has an error ranging from ± 0.3° to ±1.3° (see Supplementary Table 1). Errors less than ±1° are negligible because they are encompassed in the size of the symbols used to represent the sample on the regression line in the Kruhl (1998) diagram. Whereas, errors greater than ±1° allow a range in deformation temperatures of at least ±12 °C. As a consequence, assuming that the error of ±50 °C included in the Kruhl thermometer is valid and that a measurement error exists, deformation temperatures estimated for samples D11, D45 and D38 may have a total error of ±62 °C (see Supplementary Table 2).

The error of ±50°C suggested by Kruhl (1998), however, is valid only for the samples used to compile the original O.A. *vs.* deformation temperature plot. As the role of hydrolytic weakening, strain rate and temperature within the operation of glide systems is not quantifiable (as well as the 3D strain type and recrystallization mechanisms), deformation temperatures obtained from sample collected in other localities may have errors that could be or not, equal to what suggested by Kruhl (1998; see Section 2.2 in Law, 2014). In addition, given the abundance of a-slip rather than c-slip in the Qtz-c axis stereographic projections (Fig. 7), the error can legitimately be smaller than ±50 °C. Considering the difficulties in quantitatively estimating the error introduced by each factor, I prefer not insert the error associated with deformation temperature estimates thoughout the text and figures.

Quartz c-axis opening angles plotted using the Kruhl's thermometer indicate that deformation temperature increased progressively eastward and up structural section over ~2545 m toward the base of the migmatitic gneisses, from ~511 °C to ~678 °C (Fig. 8b). This trend is interrupted at ~ 1375 m beneath the migmatitic gneisses, where a deformation temperature of ~490 °C is estimated for sample D44 (Fig. 8). For this sample, quartz and feldspar microstructures yield similar deformation estimated temperatures (quartz: <500 °C, feldspar: 450-500°C; see Section 4.2.1), which I infer to result from the occurrence of a late localized shear event developed for lower temperature and/or during an increase in strain rate and/or a decrease in water content. In sample D44 however, evidence of an early higher temperature shear event consistent with deformation temperature documented in other samples, is the rare occurrence of dragging structures in quartz (indicative of GMB recrystallization; Fig. 9d) and the fish-shaped monomineralic feldspar grains (indicative of dislocation and diffusion creep; Fig. 9c).

4.2.3. Quartz c-axis distribution-based thermometer

Using the same assumptions and caveats for the Kruhl's geothermometers, density variations in the quartz c-axis fabrics may be used to infer operative slip systems and at least qualitatively estimate deformation temperatures (see reviews by Toy et al., 2008; Law, 2014 and references therein). Inferred basal <a> slip coupled with a combination of rhomb <a> and prism <a> slip in the westernmost collected samples and the dominance of prism <a> slip and lack of basal <a> prism in the sample closest to the base of the migmatitic gneisses (Fig. 7) may indicate that deformation fabrics were recorded under temperatures that increased toward eastward structurally higher levels.

4.3. Thermal gradient

The samples plotted in the deformation temperature versus structural position diagram yield an apparent thermal gradient of ~68 °C/km with a coefficient of determination R^2 of 0.94 (Fig. 8b). It is worth noting that because of the presence of peculiar quartz microstructures indicative of distinctive deformation conditions (see Sections 4.2.1 and 4.2.2), samples D44 has been excluded during the interpolation process. The probability of fit increases approaching 1 ($R^2 = 0.98$) using a polynomial line along which the apparent thermal gradient increases toward the base of the migmatitic gneisses from ~42 °C/km to ~105 °C/km (Fig. 8b). It should be emphasize however, that the samples are heterogeneously distributed along the investigated crustal section (sample D11 is located ~1,600 m below the rest of the samples that are distributed within only ~900 m) and samples from both the base and the top of the GHS are lacking.

4.4. Flow Vorticities

Vorticity analyses are commonly used to quantifying the degree of non-coaxiality of the flow in deformed rocks (see review by Xypolias, 2010). For plane-straining conditions, the kinematic vorticity number (Wk) is a measure of the instantaneous contribution of pure and simple shearing components of deformation, and can be employed to estimate shear-zone perpendicular thinning and transport-parallel stretching associated with exhumation/extrusion of deformed rocks. For 2D (plane strain) deformation, Wk = 1 indicates that only simple shearing occurred, whereas Wk = 0 indicates that only pure shearing occurred. As demonstrated by Law et al. (2004), equal contributions of pure and simple shearing occur at Wk = 0.71. Vorticity of flow was estimated using the δ/β method (Wallis, 1995). The method uses the quartz oblique-grain-shape orientation and the obliquity of the quartz c-axis fabric to estimate the degree of non-coaxiality that occurred during ductile flow by using the equation (Wallis, 1995):

$$Wk = \sin 2(\beta + \delta) \tag{1}$$

where $(\beta + \delta)$ is equal to the angle ξ, the acute angle between flow apophysis A1 (lines of zero instantaneous rotation) and the principal instantaneous stretching axes (ISA1) (Passchier, 1987) (Fig. 10). As Wk defines the vorticity of flow during instantaneous deformation, to integrate its variations over space and time, the use of the mean vorticity kinematic number, Wm, is more appropriate (Passchier, 1988). Typically, deformation is assumed to be steady-state, so that Wm is equivalent to Wk, however, more complex strain histories can be considered for determining Wm if sufficient time-related constraints for the deformation path are known.

Wallis (1995) determined that β can be measured as the acute angle between the central girdle of the quartz c-axis fabric and mylonitic foliation (Sa) within the XZ plane (Fig. 10a), and that δ is the angle between oblique foliation (Sb) and mylonitic foliation (Sa) (Fig. 10a). However, as the oblique-grain-shape foliation Sb, is the result of complex and intimate relationships between nucleation, dynamic recrystallization and rotation, only quartz neoblasts nucleated/recrystallized, with the greatest measured acute angle δ between oblique foliation (Sb) and mylonitic foliation (Sa) (Fig. 10a-b) should be used (Wallis, 1995). For this study, β angles were measured from the quartz-c axis fabrics, whereas the angle δ was measured from high-resolution photomicrographic images with the software ImageJ v.1.47i (Abramoff et al., 2004). Long axes orientations for at least 400 quartz grains were obtained for each thin section (Supplementary Fig. 2 and Table 3).

Vorticity estimates from the selected quartzites and quartz-rich micaschist range from 0.87 to 1.00 (Fig. 10c), indicating that deformation during the late steady-state increments of D2 shearing was dominated by a non-coaxial shearing. Moving up structural section, the pure shearing component decreases from up to 33% to 0%, indicating that the rocks immediately below the migmatitic gneisses recorded a deformation that approaches 100% simple shearing.

4.5. Strain and magnitude of principal stretches

Finite strain ratios were estimated using the nomogram recently proposed by Xypolias (2009). Assuming steady-state flow and monoclinic geometry for deformation, the nomogram graphically represents theoretical relationships between β, δ, vorticity of flow (Wm) and finite strain rate (Rxz) (Fig. 11a; Xypolias, 2009). It provides Rxz values that are particularly sensitive to $\beta > 5°$ (Xypolias,

2009) and assumes that the microstructures and quartz crystal fabrics used to estimate the β and δ angles represent the entire strain history.

Rxz strain ratios obtained from the studied samples range from 4.5 to >50 (Fig. 11; Supplementary Table 4) and drastically increase when traced structurally down toward the MCT. The two lower samples, D11 and D46, in fact, have much larger strain ratios (~ 60±15 and ~ 43±15) than the four upper samples (D45, D44, D31 and D38) that have strain ratios between 8 and 15, except for sample D44 that has both atypical microfabrics and data for its deformation temperature (Fig. 11b). Assuming that the orientation of the quartz oblique foliation (i.e. δ) and the asymmetry of fabric skeleton (i.e. β) in the upper four samples were acquired during the later stages of incremental deformation and considering that most of the parameters that control the Rxz calculation are unknown (e.g. diffusive mass transfer deformation mechanisms, strain rate, post-mylonitic recrystallization recovery), the strain results may provide only an indication of strain developed during the late D2 stage of top-to-the SW penetrative shearing.

Even if samples D11 and D46 record significantly greater strain, they did not record much difference in vorticity estimates as compared to other samples (Fig. 10b). Methods used to estimate vorticity of flow and Rxz both use δ and β parameters. The former has quite homogeneous values for the six samples, whereas the latter has values for samples D11 and D46 that are 4 to 8 time less than measured in the other samples (Supplementary Table 4). The parameter that controls the difference in Rxz estimates and consequently in the magnitude of stretches (see below) is thus the angle β or, in other words, the asymmetry of Qtz c-axis petrofabrics. It is classically interpreted as reflecting an increase in the strain during simple shear and/or an increase of simple shear component of deformation (Schmidt and Casey, 1986). In the Chaudabise valley, the slight increase of the simple-shear component documented by the vorticity technique (Fig. 10c) up structural section is consistent with this interpretation, whereas the very large strain values (Fig. 11) recorded in the two lower samples (D11 and D46) contrast to this hypothesis and requiring alternative explanations. In addition, the very weak asymmetry of the crossed girdle fabrics obtained for the lower samples is not consistent with the large contribution of simple shearing documented by vorticity estimates. These discrepancies could be explained by assuming that relicts of a pre-existing quartz preferred orientation, not completely obliterated by the D2 top-to-the SW shearing, are present in samples D11 and D46. Consequently, the strain values obtained for these two samples overestimate the strain developed during the later increments of top-to-the SW penetrative shearing.

Assuming plane-strain deformation (consistent with quartz c-axis fabric skeletons (see section 4.1 and Fig. 7), Rxz finite strain ratios can be integrated with vorticity data to estimate shortening measured orthogonal to the flow plane using the following equation (Wallis et al., 1993):

$$S=\left\{\frac{1}{2}(1-W_m^2)^{1/2}\left[\left(R_{xz}+R_{xz}^{-1}+2\frac{(1+W_m^2)}{(1-W_m^2)}\right)^{1/2}+(R_{xz}+R_x^{-1}-2)^{1/2}\right]\right\}^{-1} \quad (2)$$

Additionally, assuming volume-constant deformation and parallelism between the S2 mylonitic foliation and flow plane developed during D2 shearing as suggested by the important contribution of non-coaxial shear deformation, it is possible to obtain maximum principal stretch (s) estimates parallel to the top-to SW transport direction (i.e. parallel to the flow plane) by simply assuming that $s = S^{-1}$.

Shortening orthogonal to the flow plane decreases up structural section from ~60-70% to ~0-17% and the extension parallel to the tectonic transport changes in the same direction from ~150-215 % to ~0-20% (Supplementary Table 4). Considering samples D45, D44, D31 and D38 that represent the strain intensity between ~1400 and ~600 meters from the base of the migmatitic gneisses, the shortening and stretching magnitudes perpendicular and parallel to the flow plane for each specimen, respectively can be calculated and compared using mean Rxz and Wm values. Assuming Wm = 0.96 and Rxz = 9.7, the four upper samples would have experienced a mean shortening perpendicular to the flow plane of ~20% and a principal stretch within the flow plane trending parallel to the top-to-the SW transport direction of ~67%. Taking a median Rxz ratio of ~51:1 and Wm value of 0.91 for samples D11 and D46, I estimated a ~68% shortening perpendicular to flow plane and a ~200% stretch parallel to the transport direction as representative of the lowest crustal levels exposed in the Chaudabise river valley.

However, if deformation occurred entirely under simple shear conditions (Wm = 1), the velocity field is parallel to the shear zone boundaries/flow plane. This kinematic behaviour would mean that in samples D31 and D38 in which the maximum vorticity number is 1, the shortening perpendicular to flow plane could be zero. As already discussed for the strain estimations, the shortening and stretching values obtained for the 4 upper samples (D45, D44, D31 and D38) provide only a rough guide of the magnitudes of stretches recorded during the later stages of D2 penetrative top-to-the SW shearing. The interpretation of the magnitude of stretches values obtained for the two lower are more problematic because of the probably presence of a pre-existing quartz preferred orientation.

5. Discussion

5.1. The lower boundary of the GHS: implications for mid-crustal extrusion processes

The location of the MCT is still debated within the Himalayan scientific community (Searle et al., 2008). Its position, in fact, is critical for the interpretation of the kinematics, thermal profile and strain distribution across the GHS as well as to discriminate the conceptual model responsible of the southward extrusion of the GHS. Until now, the limited thickness of the GHS documented 30-35 east of the study area (2-5 km: Carosi et al., 2002, 2010) has been used to exclude or/and strongly limit the operation of channel-flow-type mechanisms in the Dolpo region. According to Carosi et al. (2010) and Montomoli et al. (2013), in fact, the exhumation of the GHS in this region occurred by ductile in-sequence propagation-style mechanisms accommodated by km- to m-thick shear zone (High Himalayan Discontinuity, HHD: Montomoli et al., 2014; Fig. 12). In the Seti region (West Nepal), Robinson et al. (2006) suggested instead that the exhumation of GHS occurred by vertical thinning and horizontal shortening accompanied by brittle and ductile-brittle thrusts and duplexing. Recent work conducted in far-northwest Nepal by Yakymchuk and Godin (2012), instead, suggests that the Dolpo region represents a cooler and thin foreland section of the extruding paleo-channel deformed by propagating-style deformation and that the HHD (Montomoli et al., 2013), corresponding to their metamorphic discontinuity (MD, Fig. 12), may represent the foreland-hinterland transition zone of the belt.

Detailed geological mapping (Fig. 3), microstructural (Figs. 6 and 9) and quartz c-axis fabrics analyses (Fig. 7), deformation temperatures estimates and vorticity of flow results (Fig. 8) carried out on Qtz-rich mylonites allow me to interpret the crustal section exposed in the Chaudabise river valley as belonging entirely to the GHS. My results, in fact, are consistent with the MCT fault defined by Searle et al. (2008) as the boundary between penetrative ductile sheared rocks affected by greenschist to amphibolite facies conditions and unmetamorphosed or low-grade rocks preserving sedimentary structures. Deformation temperatures estimated using quartz fabric opening-angles and recrystallization regimes, in fact, indicate that quartzites located at ~3100 m below the previously mapped MCT (e.g. samples D11) are not affected by very low-grade metamorphism but instead record penetrative D2 shearing and deformation temperatures of ~520 °C. The ductile penetrative shear zone with a top-to-the SW sense of shear in the hanging wall of the MCT should be at least ~5 km further structurally downward (i.e. toward SW) and the location of the MCT previously mapped by Franks and Fuchs (1970), Arita et al. (1984) and Carosi et al. (2002) should be repositioned (see Figs. 2 and 4). More precisely, the metasedimentary sequence documented in the Chaudabise valley represents the lower portion of the GHS (GHS_L) whereas the migmatitic gneisses, cropping out in the eastern portion of the valley, are the base of the upper portion of the GHS (GHS_U, as defined by Larson et al., 2010). The presence of syn-kinematic biotite and garnet crystals within the micaschists cropping out in the western portion of the East sector of the Chaudabise valley (i.e. located west of sample D11; Figs. 3 and 4) indicate that these rocks are near the base of the GHS_L. Unfortunately, the presence of normal faults that repeated the metasedimentary sequences and offset the base of the GHS_U (i.e. the Jumla klippe of

Arita et al., 1984; Figs. 3 and 4) prevents exposure of the MCT fault on the geological map of Fig. 3. However, I speculate that it is located immediately beneath the outcrops located at the beginning of the East sector of the Chaudabise river valley (Fig. 4). Consequently, the thickness of GHS changes from 2-5 km, as estimated by Carosi and co-workers, to at least 7-10 km. In addition, the absence of stratigraphic repetitions and the presence of the high-temperature crystal–plastic syn-metamorphic pervasive shearing in the studied transect is more consistent with a syn- to late-metamorphic ductile extrusion rather that with a brittle and brittle-ductile propagation-style models as suggested by Robinson et al. (2006). These considerations imply that the mid-crustal extrusion processes active in this region should be re-visited and the role of the channel-flow-type mechanism re-considered.

5.2. Vorticity of flow and tectonic implications

Studies conducted across the GHS along the entire belt (e.g. Grasemann et al., 1999; Law et al., 2004; Jessup et al., 2006; Carosi et al., 2006, 2007; Larson and Godin, 2009) seems corroborate, even with some discrepancy, the velocity profiles predicted by channel flow-type mechanisms. They document a pervasive ductile deformation throughout the GHS with an increase of pure shear component (small Wm value) toward the base of the GHS_U and a general shearing with an equal contribution of pure and simple shear components within the GHS_L. The occurrence of pure shear at the base of the GHS_L may be attribute to lithostatic loading (e.g. Larson and Godin, 2009; Wagner et al., 2010) or explained by considering a decelerating strain-path (Grasemann et al., 1999). Recent detailed vorticity analyses conducted in the immediate hanging wall of the MCT in NW India indicate that the base of the GHS was affected by sub-simple shear. A more complex vorticity profile that suggests the operation of a hybrid flow regime with a combination at deepest structural levels of Couette and Poiseuille flow, has been recently proposed in the Northern Himalayan gneiss dome (Langille et al., 2010; Wagner et al., 2010). Unfortunately, no vorticity studies are available across the HHD.

In contrast to previous studies, Wm results obtained in the Chaudabise valley indicate that the entire GHS_L is pervasively affected by sub-simple shear where the small contribution of pure shear component (0-33%) may be exclusively produced by lithostatic loading. These results are inconsistent with frictional wedging implied in the foreland propagating-style models. In addition, as my results constraint the kinematic of flow during the later stages of D2 shearing, the operation of the decelerating strain-path model (predicting the increase of pure shear component with time; Grasemann et al., 1999) can be excluded.

The vorticity profile obtained integrating my data to those collected at the upper portion of the GHS (e.g. Grasemann et al., 1999; Law et al., 2004) is consistent with hybrid channel flow-type models where the combination of Couette and Poiseuille flow produces a velocity profile dominated by simple shearing in a wide portion of the extruded mid-crust (Fig. 2c in Grujic, 2006). The GHS_U-GHS_L boundary (broadly corresponding to the tectonometamorphic discontinuity recently documented in West Nepal by Yakymchuk and Godin, 2012, MD, and Montomoli et al., 2013, HHD) may so represent the limit between buried material, affected by dominant simple shearing (GHS_L), and exhumed material affected by a general flow dominates by pure shearing (the GHS_U; Fig. 13). Consequently, the HHD-MD discontinuity may be interpreted as a discrete structure developed at the boundary between pure- and simple-shearing-dominated domains, under deformation conditions that reflect the transition from hinterland- to foreland-style deformation.; see also Fig. 7d of Cottle et al., 2015). To test this reconstruction, however, my results should be integrated with detailed vorticity investigations conducted immediately above the MCT and across the HHD and by further geochronological studies on deformation and metamorphism in the GHS exposed in the Chaudabise river valley.

5.3. Inversion and telescoping of paleo-isotherms in the GHS_L

The metamorphic and tectonic interpretation of temperature estimates from rocks in the GHS is a function of the chosen tectonic reference frame. The placement of this frame is complicated by data gathered where a complete section of the GHS is absent and/or the placement of the MCT within a study region that may differ in the same region from author to author. Additionally, recent P-T-t-d studies carried out within the GHS in western and central Nepal (Carosi et al., 2010; Larson et al., 2013; Montomoli et al., 2013; Larson and Cottle, 2014; Figs. 1 and 12a) document a diachronous metamorphism related to the operation, under quite similar temperature conditions, of tectono-metamorphic discontinuities that internally thickened the GHS and likely involved a sequence of discontinuities that propagated forelandward with time (see review of Cottle et al., 2015).

For these reasons, the estimates of deformation temperatures from this study are only compared to similar data from the closest GHS crustal profiles (Srivastava and Mitra, 1996; Yakymchuk and Godin, 2012; Antolin et al., 2013; Fig. 12). On the northern boundary of the Almora klippe (NW India), located ~300/350 km west of the Chaudabise river valley, Srivastava and Mitra (1996) documented an inverted thermal gradient above the North Almora thrust (i.e. the MCT). Using microstructures and deformation mechanisms, they documented structurally upward increasing deformation temperatures ranging from 350-400 °C close to the thrust to 500-600 °C at ~1.3-2.2 km above the thrust, indicating an apparent thermal field gradient of ~75-115 °C/km. In the Dadeldhura klippe, Antolin et al. (2013) using dynamic recrystallization microstructures in quartz grains documented a slighty inverted deformation temperature profile in the lower part of GHS and a right way up profile in the upper part of the GHS. In the Karnali valley, Yakymchuk and Godin (2012) used both quartz and feldspar microstructures and the fabric opening-angle thermometer to infer a syn-deformational geothermal gradient ~60-70 °C/km.

Even though it is located ~500 km along strike to the west of the Chaudabise river valley, the study by Law et al. (2013) in the Sutlej valley of NW India (Figs. 1 and 12a) is the first detailed analysis of the syn-deformational geothermal conditions across the hanging wall of the MCT. Using the quartz opening-angle thermometer and a nonlinear regression lines to interpolate the deformation temperature versus structural position data, they documented in the valleys located in the SW sector of their study area (i.e. in the foreland position) an inverted thermal field gradients ranging from ~300-170 °C/km, at less than 200 m above the MCT, to a quite “normal” ~30-40 °C/km at 600-1100 m above the MCT. In contrast, in the valleys located in the NE sector of their study area (i.e. in the hinterland position), they documented a quite “normal”, even if inverted, thermal gradient at of >200 m above the MCT.

In summary, the syn-deformational geothermal gradient of ~ 68 °C/km inferred from the GHS_L exposed in the Chaudabise valley (see Section 4.2 and Fig. 8b) is broadly comparable to thermal gradients obtained in the Karnali valley (Yakymchuk and Godin, 2012) and in the northern margin of the Almora klippe (Srivastava and Mitra, 1996). Yet, it differs from the thermal profiles obtained by Law et al. (2013).

The broadly agreement existing between temperatures estimated in the Chaudabise river valley using D2 metamorphic mineral assemblage and microstructures- and opening angle-based thermometers and temperatures estimated using petrology-based thermometers on samples collected at analogous structural positions in the adjacent valleys (Carosi et al., 2010; Montomoli et al., 2013; Figs. 1 and 12a) indicated that the inversion of the isotherms documented in the study area may be occurred “at close” to peak metamorphism conditions during the ductile top-to-the SW lateral extrusion/exhumation of the GHS. As a consequence, assuming originally sub-horizontal or shallowly N-dipping isograds and considering that the ductile top-to-the SW penetrative shearing occurred under simple shear dominated-regime, I would suggest that the inversion of paleo-isotherms was probably produced by the heterogeneous simple-shearing-dominated flow pervasively distributed across the GHS_L and responsible for the late D2 stages of foreland-directed extrusion/exhumation of the GHS.

The telescoping of paleo-isotherms documented both at the top and the base of the GHS could be produced during (Law et al., 2011, 2013): (1) a vertical shortening associated with a dominant pure-shear component of deformation; (2) a dominant simple-shear component of deformation associated with lateral extrusion parallel to the GHS boundaries; or (3) a general shear flow.

Vorticity estimates collected in the Chaudabise river valley indicate that the telescoping of the paleo-isotherms during the last stages of exhumation/extrusion of the GHS cannot be produced by vertical shortening due to pure shearing across the flow plane. Wm, ranging from 0.88 to 1, in fact, indicate that a general shear flow with a dominant component of simple shear (68-100%) would be mostly responsible for the telescoping.

Given a change of ~167°C over ~2500 meters in the Chaudabise river valley, to achieve the same change in temperature for a "normal" geothermal gradient of 30 to 40°C, the sequence would have to be ~4200-5500 meters thick. If the sequence thickness decreased from ~4200-5500 to ~2500 meters normal to the flow plane and the paleo-isotherms were parallel to the flow plane, the shortening orthogonal to the flow plane should be ~ 40-54% that would correspond to stretches parallel to flow plane of 66-120%. It is worth noting, however, that these predicted values provide information for the finite deformation history whereas my estimates constrain the late stages of shearing. The predicted values of shortening and extension in fact, are greater than the mean values obtained for the four upper samples (20% shortening, 67% extension), and at the same time are less than the mean values calculated for the lower two samples (32% shortening, 200% extension). They are comparable to the mean values of shortening orthogonal to flow plane of 36% and stretching parallel to flow plane of 88% obtained using the 6 samples Consequently, a heterogeneous general shear flow dominated by the simple shear component may have contributed significantly, at least in the later stages of D2 exhumation, to the shortening orthogonal to the isotherms and the consequent stretching parallel to the southward lateral exhumation of the GHS.

These results coupled with the presence of a syn-to late metamorphic pervasive ductile shearing and crystal plastic deformation and the complete absence of stratigraphic repetitions induced by thrusts indicate that the telescoping and the inversion of the paleoisotherms in the Chaudabise river valley was mainly produced, at least during the late Oligocene-early Miocene, by the heterogeneous simple shear homogeneously distributed throughout the GHS_L (Fig. 13).

6. Conclusions

Qtz c-axis petrofabrics analyses, vorticity of flow measurements and deformation-temperature estimates integrated with microstructures and metamorphic studies in the Dolpo region indicate that the rocks exposed in the Chaudabise river valley experienced a pervasive top-to south ductile shearing. They belong entirely to the GHS_L and the location of MCT should be moved at least ~5 km structurally downward from the previously mapped location in the Chaudabise river valley. Consequently, the thickness of the GHS in the Dolpo region should be increased to 7-10 km. The integration of microstructure- and petrofabric-based thermometers indicates that deformation temperature during the pervasive shearing increase up structural section toward the GHS_L-GHS_U boundary from ~450-500 °C to ~650 °C suggesting an apparent inverted thermal field gradient of ~68 °C/km. These temperatures are consistent with the inverted metamorphism documented using metamorphic mineral assemblages and with peak metamorphic temperatures estimated in adjacent valleys obtained using petrology-based thermometer. This overlap indicates that the pervasive ductile shearing and the inversion of the isotherms may have occurred during syn-to late peak metamorphism conditions. Vorticity studies document a sub-simple shear pervasive distributed throughout the GHS_L (Wm=0.88-1.00, 100-67% simple shear). The slight increase of pure shearing (0-33%) toward the base of the GHS_L is probably induced by lithostatic loading. The general shear flow dominated by simple shear component produced the inversion and the telescoping of isotherms and the stretch parallel to the tectonic transport documented in the Chaudabise valley. The integration of my vorticity results with those collected within the GHS_U suggest a vorticity profile throughout the GHS that contrasts to critical taper-type models and differs from the typical ductile extrusion models. It is more consistent with a hybrid channel flow model where the combination of Couette and Poiseuille flow produces a velocity profile dominated by Couette flow (i.e. simple shear) in the lower portion of the mid-crustal channel (Fig. 13). This hypothesis is also consistent with recent models suggesting the simultaneous operation, at different structural depth and time, of channel flow- and critical wedge-type processes (Larson et al., 2010, 2013; Yakymchuk and Godin, 2012; Larson and Cottle, 2014, Cottle et al., 2015 and references therein). The GHS_L, exposed in the Chaudabise valley, is so interpreted as a slice of GHS bounded at the top by the tectonometamorphic discontinuities recently documented in West and Central Nepal (i.e. MD and HHD) and at the bottom by the MCT, buried and then accreted to the foreland-propagating wedge since late Oligocene- early Miocene (Fig. 13). The MD and HHD may be so interpreted as a discrete structure developed at the boundary between pure- and simple-shearing-dominated domains in response to a change in the flow conditions that occurred at the transition between extending (i.e. channel flow model) and compression flow (i.e. frictional fault-propagating-type models).

Acknowledgements

I am very grateful to Rick Law (Virginia Tech, VA) for the logistic support and for the English revision. I would like to thank P. Mosca for constructive discussions on various aspects of Himalayan geology and for his informal review of an earlier version of the manuscript and G. Musumeci. I thank R. Carosi and C. Montomoli for providing financial support for fieldwork and partly for the laboratory work associated with this study (MiUR-PRIN and CNR-EV-K2 projects). I very appreciated the detailed and constructive revisions of K. Larson, J. Singleton, P. Xypolias, J. Langille and B. Dunne that significantly improved the manuscript.

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Figure captions

Figure 1. Simplified tectono-stratigraphic map of the Himalayan belt highlighting the first-order geological framework and location studies discussed in the text. The study area is also indicated.

Figure 2. A comparison of the location of the Main central Thrust (MCT) used in previous studies in the Jumla area with that used in this study. This study: BB: Bura Buri pluton; Gn: migmatitic gneisses; Pn: pelitic paragneisses; QP: quartzites and phyllites; GHS: Greater Himalayan Sequence. Fuchs and co-worker: GG: granite gneiss; UCr: Upper Crystalline nappe; LCr: Upper Crystalline nappe; C1: Krol and Shali Fms; C2: Blaini Fm. Arita at al. (1984): TGr: Tourmaline granite; HGz: Himalayan Gneiss zone; MCTz: Main Central Thrust zone; GJz: Galwa-Jumla zone. Carosi et al. (2002; 2013): BB: Bura Buri pluton; HHC: Higher Himalayan Crystallines; LHC: Lesser Himalayan Crystallines; MCTz: Main Central Thrust zone; LH: Lesser Himalaya. The works Carosi et al. (2007, 2010) use the same tectonometamorphic subdivision proposed by Carosi et al. (2002).

Figure 3. New geologic map of the Chaudabise river valley (Dolpo region) and lower-hemisphere stereographic projections of the main D2 structural elements. S2 foliation measurements in adjacent valleys are from Fuck (1977), Arita et al (1984) and Carosi et al. (2002). The location of photographs shown in Figs. 5 and 6 and the previous locations of the MCT are indicated.

Figure 4. Geological cross-section along the Chaudabise river valley (see Fig. 2 for location). Locations of sample used for CPO measurements and vorticity analyses and the locations of the MCT used in previous studies in the Jumla area are also shown.

Figure 5. Examples of key-rock types and structures exposed in the Chaudabise river valley (see Fig. 3 for location). (a) Foliation fish with top-to-the S shear sense in phyllites and micaschists close to the base of the Jumla klippe (Sm: mylonitic foliation; C': shear plane). (b) Bt + Grt-bearing micaschists cropping out north of Talphi, cut by S-C/S-C' structures indicating a top-to-the S/SW shear sense (S: S2 mylonitic foliation; C and C': shear planes; Grt: garnet). (c) White quartzites with bedding transposed in to the S2 mylonitic foliation (Sm). (d) Cm-size kyanite crystal (Ky) within Ky + Sil- bearing gneisses cropping out in the easternmost part of the valley. (e) F2

isoclinal folds within the Bt + Ms + Grt + Ky ± St ±Sil- bearing paragneisses (Sm: mylonitic foliation; A.P.2: F2 fold axial plane). (f) Later F4 folds with sub-horizontal axis and axial plane (A.P.4) within micaschists west of Talphi (Sm: S2 mylonitic foliation).

Figure 6. Microstructures and metamorphic mineral assemblages (see Fig. 3 for the location of samples used to obtain these thin sections). (a) Pre-/ syn-D2 Grt-bearing quartz-rich micaschist with top-to-the S/SW shear sense indicators (Grt: garnet; Bt: biotite; Sm: mylonitic foliation). A simplified sketch showing the relation between internal and external foliation in garnets is also showed. (b) Muscovite-rich quartzite (sample D46) with incipient S-C/S-C' structures indicating a top-to-the SW shear sense (Bt: biotite; Sm and S: S2 mylonitic foliation; C and C': shear planes). (c) Quartz-rich paragneiss with S-C/S-C' structures indicating a top-to-the S/SW shear sense (Sm and S: mylonitic foliation; Bt: biotite; Grt: garnet; C': shear plane). (d) Kyanite-bearing paragneiss collected in the upper portion of the GHS_L. Large flakes of biotite (Bt) define the mylonitic foliation (Sm and S) and shear planes (C') (Ky: kyanite).

Figure 7. Optically measured quartz c-axis fabrics from the six selected quartz rich-mylonites (see Fig. 4 for sample locations). The fabric data are shown on lower hemisphere equal-area projection using version 2.1 OSXStereonet developed by N. Cardozo and R. D. Allmendinger.

Figure 8. Kruhl (1998) opening-angle (O.A.) thermometer, adapted from Morgan and Law (2004) and Law et al. (2004). (a) Relationship between O.A. and deformation temperature with sample data plotted. (b) Deformation temperature estimates plotted against distance beneath the base of the migmatitic gneisses. Linear regression line through data points indicates an apparent thermal field gradient of 68°C/km with a coefficient of determination R^2 of 0.94. See text for further discussions on errors related to deformation temperatures and to the estimations of thermal gradient using an interpolation curve. The thickness of circles used to represent the samples encompasses an error in sample positioning of ± 50m.

Figure 9. Quartz microstructures. (a) Sample D46. Quartz (Qtz) showing heterogeneous grain size and irregular grain boundaries indicating GBM recrystallization. Bt: biotite; Feld: feldspar. (b) Sample D38. Lobate boundaries between quartz (Qtz) and feldspar (Feld) indicate the occurrence of high temperature conditions during later D2 shearing. Bt: biotite. (c) Sample D44. Bulging structures in quartz (Qtz; indicative of SGR recrystallization) along high-mobility grain boundaries pinned by biotite crystal (Bt) (i.e. dragging structure: black arrow) characteristic of GBM recrystallization. (d) Sample D44. Domains of small and elonged sub- and new grains oriented

parallel to the mylonitic foliation develop around the quartz grains (ng: newgrain; sg: subgrain) inferred to mean that SGR was the dominant recrystallization mechanism. The fish-shaped monomineralic feldspar grains (Feld), showing deformation twins, straight cusp and boundaries, indicate that the rock experienced high-grade metamorphic conditions (ten Grotenhuis et al., 2003).

Figure 10. Vorticity analysis. (a) Angular relationship between oblique grain shape (Sb), quartz c-axis fabrics and flow plane (Wallis et al., 1993; Wallis, 1995; Law et al., 2004). (b) Example of preferred orientation histograms used to estimate δ (see Supplementary Fig. 2 for all histograms). (c) Bar chart showing vorticity results. Bt + Grt metamorphic isograd location is approximate (see Fig. 4 for sample locations) (q: quartzite; m: quartz-rich micaschist).

Figure 11. Strain ratio measurements. (a) Sa Λ Sb angle *vs.* ß angle normogram used for Rxz strain ration estimates (Xypolias, 2009). (b) Strain ratio estimates plotted against structural position beneath the top of the MCTz (i.e. the base of the GHS gneisses). The nomogram-based method is unreliable under high-strain condition (Rxz > 10-15) with $\beta < 5°$ (see Xypolias, 2009). For this reason Rxz values from samples D46 and D11 should be viewed with caution.

Figure 12. Studies conducted in NW India and in West and central Nepal discussed in the text are located both in a regional geological sketch map (a; see Fig. 1 for map location) and along a simplified schematic cross –section (b) across the GHS. It's worth to note that the location of the vorticity results from Carosi et al. (2007) have been located respect to the new position of the MCT defined in this work. The type of study (vorticity of flow, strain, deformation temperatures and P-T-t-d paths) is indicated with different colours. In (b), the location of the different studies along the MCT surface doesn't reflect their reciprocal position in a hypothetical foreland-hinterland transect.

Figure 13. (a) Schematic reconstruction of the Himalayan orogen during the development of the MD-HHD tectonometamorphic discontinuity in the late Oligocene-early Miocene (adapted from Cottle et al., 2015). The star indicates the hypothetically location of the crustal section now exposed in the Chaudabise river valley. See text for explanation. Hypothetical velocity profiles and kinematics for the edge (b) and the deep portion of the GHS (c) adapted from Grujic (2006). The absolute value of vorticity is proportional to the width of the black bar with wide bar that indicates portion dominated by simple shear. Bold line: active structure; grey line: formerly active structure; white colour: undifferentiated rock of Indian affinity.

Supplementary Material

Table 1. Opening angles of quartz c-axis fabrics measurements and associated errors.

Table 2. Opening angles and deformation temperatures obtained using the Kruhl thermometer (1998).

Table 3. Summary of vorticity data.

Table 4. Summary of strain and magnitude of principal stretches data.

Figure 1. Parameters used to measure the opening angles of quartz c-axis fabrics.

Figure 2. Preferred orientation histograms used to estimate Wk.