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PRESSURE-TEMPERATURE-TIME-DEFORMATION PATH OF KYANITE-BEARING MIGMATITIC PARAGNEISS IN THE KALI GANDAKI VALLEY (CENTRAL NEPAL): INVESTIGATION OF LATE EOCENE-EARLY OLIGOCENE MELTING PROCESS

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1. Introduction

The Himalaya-Tibet orogenic system is often regarded as the classic "*type*" of a continental collisional belt. This belt is characterized by a continuity of both main tectonic units and discontinuities for almost 2400 km along strike. Among these units, the Greater Himalayan

Sequence (GHS) represents the now-exhumed metamorphic core of the Himalayan orogenic pile (Hodges, 2000), where medium- to high-grade metamorphic rocks and migmatites occur (*e.g.* Hodges, 2000). The pressure (P) – temperature (T) – time (t) paths of migmatites can carry valuable information on the thermal history and rheological evolution of collisional belts (*e.g.* Searle, 2013; Yakymchuk and Brown, 2013; Hallett and Spear, 2014; Weinberg and Hasalová, 2015). Melting has profound effects on the rheological properties of rocks such as a dramatic reduction of their viscosity even if only small volumes of melt (> 5%) are present (Rosenberg and Handy, 2005; Jamieson et al., 2011). This "melt-weakening effect", which should have affected a large part of Himalayan high-grade rocks, together with focused denudation/erosion, is the main prerequisite to apply channel flow models (Beaumont el al., 2001; Jamieson et al., 2004; Jamieson et al., 2011). These models can explain the evolution and exhumation of high-grade metamorphic rocks, although the mechanisms, the productivity and the timing of melting are far from being well understood in the Himalaya despite numerous experimental and empirical investigations made by many authors (*e.g.* Patiño-Douce and Harris, 1998; Harris et al., 2004; Guilmette et al., 2011; King et al., 2011; Searle 2013; Palin et al., 2014; Imayama et al., 2012; Groppo et al., 2010, 2012 and references therein). Several melting stages have been proposed for this mountain belt occurring at different times. For example, in the Sikkim region (Fig. 1) Rubatto et al. (2013), using monazite and zircon petrochronometers, identified two portions of the migmatitic GHS, which experienced partial melting and cooling at different times: structurally lower migmatites experienced melting and peak metamorphic conditions at c. 31-27 Ma, earlier than the structurally higher ones (26-23 Ma). Also Kohn et al. (2005) and Imayama et al. (2012) identified a clear diachronic evolution of GHS melting in the Langtang section, (Central Nepal) and in the Tamor-Ghunsa transect (Eastern Nepal) respectively (Fig. 1). In both cases the structurally higher samples experienced melting earlier than the lower ones (20 Ma vs. 15 Ma in Kohn et al., 2005, and 33-28 Ma vs. 21-18 Ma in Imayama et al., 2012).

Here, we focus on kyanite-bearing migmatitic paragneiss in the Kali Gandaki valley (Central Himalaya, Fig. 1) located in a key structural position (Vannay and Hodges, 1996) due to its closeness to the Main Central Thrust (MCT). Recently, Carosi et al. (2014a) identified within garnets coming from the same outcrop, the presence of crystallized melt inclusions, referred as "nanogranites" (Carosi et al., 2014a and references therein) showing a peculiar chemical composition of high-Ca melts (tonalites). According to Patiño-Douce and Harris (1998) and Prince et al. (2001) this melt types is interpreted as result of high-pressure (HP) melting, possibly in the presence of free water and through K-feldspar-absent reactions (see also King et al., 2011; Palin et al., 2014; Weinberg and Hasalová, 2015).

We present pressure (P) - temperature (T) - deformation (D) path for the kyanite-bearing migmatitic paragneiss, which is based on the pseudosection approach (*e.g.* Vance and Mahar, 1998) and trace element thermometry (*e.g.* Spear and Pyle, 2002; Hallett and Spear, 2014) coupled with careful meso- and micro-structural observations. Texturally and chemically controlled *in situ* U-Th-Pb monazite ages add important time constraints to this path (*i.e.* P-T-t-D paths). It has already turned out that in situ U-Th-Pb monazite dating, with the advantage that textural and chemical "grain-fingerprints" are preserved, coupled with the pseudosection approach is a powerful tool to temporally bracketing the P-T-D history of metamorphic rocks (see, *e.g.,* Williams and Jercinovic, 2002, 2012). At last, possible tectonic implications will be discussed.

2. Geological overview of the Himalayan Belt

The Himalayan mountain belt (Fig. 1a) is the result of the collision between the Asian and Indian continental plates around 55-50 Ma (Hodges, 2000) after the break-up of Gondwana and a long last-standing Andean-type active margin, which resulted from subduction of Neo-Tethys oceanic crust below the Lhasa Block, the intrusion of large (mainly I type) granitoid bodies and accretion of arc terranes. The precise age of the India-Asia collision as well as a possible diachronic collision is still under debate in the geological literature (*e.g.* Najman et al., 2010 and references therein).

Since the collision occurred, crustal rocks representing the northern front of the Indian plate have experienced a complex deformative and metamorphic history, building up part of the Himalaya as we observe this mountain range nowadays (Gansser, 1964; Heim and Gansser, 1939; Hodges, 2000; Yin, 2006).

The structural architecture of the Himalayan chain is made by different tectono-metamorphic units (Fig. 1a) separated by important tectonic structures (regional scale reverse and normal shear zones). In an "ideal" profile from south to the north, the following tectonic units are present according to Hodges (2000):

1) The Siwalik Unit (SU) is made of recent sediments tectonically sandwiched between the undeformed molasse of the Ganga plain and the upper unit, the Lesser Himalayan Sequence (LHS). The lower tectonic contact is a top-to-the south thrust system referred as Main Frontal Thrust (MFT), whereas the Main Boundary Thrust (MBT) separates the SU from the LHS.

2) The LHS is made of Lower Proterozoic to early Palaeozoic low to medium grade metasediments and meta-igneous rocks. This unit is tectonically overlain by medium- to high-grade metamorphic rocks of the GHS *via* a regional top to the S-SW (thrust sense) shear zone, called Main Central Thrust (MCT, Figure 1). Since the MCT is not a single thrust surface, but a thick ductile to brittle shear zone with a variable thickness (100 m up to several km, Searle et al., 2008), we prefer the term Main Central Thrust Zone (MCTZ) to identify its sheared rocks. According to Stephenson et al. (2001) both GHS and LHS rocks are ductily sheared by the MCT

activity, with the latter rocks ductily incorporated in the MCTZ during the shear zone activity (widening of the shear zone towards the S).

3) The GHS consists of Late Proterozoic to Cambrian metamorphic rocks. According to Searle and Godin (2003) and Searle (2010) three main units (referred as three *"formations"* in Le Fort, 1975) can be generally identified: *(i)* Unit 1 is made of kyanite-bearing metasediment with subordinate quartzite, calcsilicate and marble. Migmatites are present in the upper part of this unit; *(ii)* Unit 2 is mainly composed of medium to high-grade calcsilicate and minor marble, *(iii)* Unit 3 consists mainly of orthogneiss and minor kyanite/sillimanite migmatite. Structurally upwards (mainly within Unit 3) the GHS is intruded by Miocene leucogranite, referred as Higher Himalayan Leucogranites (HHL, Le Fort, 1975; Visonà et al., 2012), forming foliation-concordant sills and cross-cutting dykes up to large (kilometre sized) plutons. Melting is mainly the result of muscovite and biotite dehydration melting (Patiño -Douce and Harris, 1998; Visonà et al., 2012; Searle, 2013 and references therein). Moreover, Prince et al. (2001) recognized also the presence of an older "water-fluxed melting episode" at higher pressure (see also King et al., 2011).

The GHS evolution is often subdivided into two main metamorphic stages, classically (*e.g.* Vannay and Hodges, 1996) referred as Eohimalayan HP-MT event (M1) of Eocene-Oligocene age, in which the GHS reached the highest pressure (in the kyanite stability field) and the Neohimalayan MP-HT Miocene event (M2). For instance, the GHS rocks studied by Liu et al. (2007) experienced P-T conditions of 1.4 GPa and 750-800°C at 33 ± 2 Ma (zircon U-Pb SHRIMP) and then the original kyanite-bearing paragenesis in metapelites (e.g., Borghi et al., 2003) was overprinted by sillimanite- and cordierite-bearing paragenesis, associated with widespread melting, at 23 ± 2 Ma (Liu et al., 2007).

Deformation within the GHS is dominantly characterized by general shear (*e.g.* Larson and Godin, 2009 and references therein). The GHS is tectonically overlaid by the Tethyan

Sedimentary Sequence (TSS) *via* a large scale ductile to brittle system of normal faults, named South Tibetan Detachment System (Caby et al., 1983; Burg et al., 1984; Burchfiel et al., 1992; Carosi et al., 1998*,* 2002; Searle, 2010) active in the same time span of the activity of the MCT.

4) The TSS consists of early Palaeozoic to late Mesozoic multi-phase folded unmetamorphosed to low grade metamorphic sediments (Godin et al., 1999a,b; Antolín et al., 2011; Dunkl et al., 2011) originally deposited on the northern passive margin of the Indian Plate. Towards the N, the TSS is bounded by flysches and ophiolites (often with a blueschist metamorphic imprint, Hodges, 2000) of the Indus-Tsangpo suture zone (Fig. 1).

3. The GHS in the Kali Gandaki valley

The N-S trending Kali Gandaki valley (Fig. 1b) cross-cuts the whole Himalayan units and structures (Fig. 1a, b), offering a clear exposure of metamorphic rocks and their structural relationships. For these reasons the valley is probably one of the most classic study area of Himalayan geologists (*e.g.* Le Fort et al., 1986; Colchen et al., 1986; Vannay and Hodges, 1996; Godin et al., 1999a,b; Godin et al., 2001; Godin, 2003; Larson and Godin, 2009; Searle, 2010; Carosi et al., 2014a,b). Nevertheless up to now metamorphic P-T paths obtained with pseudosection modelling are lacking and very few *in situ* mineral ages are available along this transect.

In the Kali Gandaki valley the GHS appears as a homogeneous homoclinal slab with isoclinal folds (Brown and Nazarchuk, 1993; Vannay and Hodges, 1996) reaching a structural thickness of 10-15 km (*e.g.* Godin 2003). The main foliation (S_2) related to a second deformation phase (D_2), strikes NW-SE and moderately dips to the NE, whereas the main mineral lineation (L_2) strikes mostly E-W or SE-NE and plunges to the E (*e.g.* Vannay and Hodges 1996, Carosi et al., 2014b). Relicts of an older deformative event D_1, (Eohimalayan event of Vannay and Hodges, 1996) have been reported, for instance, by Vannay and Hodges, 1996. This eventi is testified by sporadic evidences (such as high-angle mica and kyanite) of an older foliation (S_1) and as an internal foliation in garnet porphyroblasts.

According to Vannay and Hodges (1996) the GHS in the Kali Gandaki could be subdivided into four “formations” (Formation 1, 2, 3, corresponding to Unit 1, 2, 3 of Searle and Godin, 2003) plus the upper Larjung Formation.

The lowest one (unit 1 of Fig. 1b) contains kyanite-bearing paragneiss and micaschist. The second one (unit 2 of Fig. 1b), representing the thick core of the GHS in this transect, consists of calcsilicate with the paragenesis of clinopyroxene, garnet, amphibole (often with titanite) and minor metapelite.

The third unit (Unit 3, Fig. 1), made of orthogneiss and minor metapelite, is overlaid by *c.* 200 m of amphibole-bearing calcsilicate-gneiss (Larjung formation). According to Godin (2003) the Larjung formation, interpreted by previous authors as the base of the TSS (*e.g.* Colchen et al., 1986), is deformed together with the upper part of the GHS by the Annapurna Detachment (AD), a 1500 m thick high-strain zone, representing a local segment of the STDS. Godin et al. (2001) suggested an age of *c.* 22 Ma for the cessation of the ductile shearing along the AD.

Within Unit 3, a ductile shear zone with a top-to-the southwest sense of shear, named Kalopani Shear Zone (KSZ in Fig. 1b), has been identified by Vannay and Hodges (1996) and is interpreted as an out of sequence thrust (*e.g.* Vannay and Hodges, 1996; Godin et al., 1999). Based on Ar-Ar white-mica geochronology an age older than 13-15 Ma has been suggested for the shearing (Vannay and Hodges, 1996). New U-Th-Pb monazite ages indicate an Eocene age (c. 30-40 Ma) for the shearing activity along the KSZ (Carosi et al., 2014b).

In the rocks belonging to the MCTZ–Lower GHS, Vannay and Hodges (1996) reported two groups of samples based on their P-T record. Eohimalayan Group 1 with P-T conditions of 610±40° C – 0.94±0.09 GPa, is interpreted to have mainly equilibrated close to the metamorphic peak (at the beginning of thrusting along the MCT), whereas the Neohimalayan Group 2 (540±30° C 0.65±0.03 GPa), which is more intensively sheared and retrogressed, records the exhumation stage during or after the MCT activity. Based on Ar-Ar geochronology, Vannay and Hodges (1996) suggested that the whole GHS rapidly cooled below the white-mica closure temperature (300-430°C in Vannay and Hodges, 1996) during the Early–Middle Miocene (13-15 Ma). Carosi et al. (2014a) identified "nanogranites" as inclusions in garnets from kyanite-bearing rocks (bottom of the upper GHS) testifying that these rocks have experienced melting starting at *c.* 41–36 Ma, based on U-Pb *in situ* monazite ages, further supporting the idea that these rocks should be better classified as migmatitic paragneisses (see also Searle, 2010).

4. Outcrop description and sample petrography

The studied gneiss comes from outcrops close to Titar Village (North of Dana village, Fig. 1b), > 1 km north of the MCT as mapped by Colchen et al. (1986) and Vannay and Hodges (1996) because top-to the south kinematic indicators are present. Recently, Parsons et al. (2014) assigned this thrust structure to the Chomrong Thrust (CT) shifting the MCT further to the south.

In the outcrops (Fig. 2a-d) kyanite-garnet-biotite-white mica-bearing migmatitic metapelite occur with thin intercalations (dm-thick) of garnet-amphibole-bearing gneiss.

Leucocratic layers (*in situ* leucosome) parallel to the main foliation commonly occur, which form tight to isoclinal folds with axial planes parallel to the main foliation (Sp). These leucosomes, mainly made of plagioclase and quartz, are stretched and the folds asymmetry points a top-to-the S sense of shear. Centimetric garnet and kyanite are abundant (Fig. 2b,c), reaching the largest grain size within the leucocratic layers. Mafic selvedges (*e.g.* Sawyer, 2008) of biotite were also observed (Fig. 2a, b). Moreover, late leucocratic pods are present

(Fig. 2d). Several hand specimens, covering the range of structures observed at the mesoscale were sampled (K28a to K28g).
Microscopically the main foliation is classified as a spaced anastomosing foliation (Passchier and Trouw, 2005) and is defined by lepidoblastic levels of biotite and white mica, where kyanite also occurs (Fig. 3a). Mineral lineation (Lp) is mainly defined by aligned kyanite and stretched quartz and feldspar. The microscale kinematic indicators, such as shear bands (Fig. 3b) and sigma type porphyroclasts, support a top-to-the S sense of shear.

Garnet is subhedral and full of tiny inclusions, which are in some cases moderately iso-oriented. Inclusions, often polymineralic, are made of plagioclase, rutile/ilmenite, white mica, biotite and minor chlorite. Garnet rims are frequently replaced by biotite and plagioclase. Late fractures, partially filled with chlorite/green biotite, are oriented at high angle with respect to the main foliation. They often cause retrogression of inclusions, with ilmenite replacing rutile and white mica being transformed to biotite/chlorite. The main foliation wraps around the garnet porphyroblasts, supporting their pretectonic occurrence (with respect to the top-to-the south shearing).

Kyanite forms large porphyroblasts (early synkinematic?) often aligned with the main foliation. This mineral contains inclusions of paragonite, quartz, potassic white mica and rutile. Plagioclase around quartz inclusions in kyanite was sporadically observed (*e.g.* Carosi et al., 2014a). This feature suggests that also kyanite was present during melting (Carosi et al., 2014a). Kyanite in the mesosome and less frequently in the leucosome is sometimes deformed (kinked) and locally partially replaced by fine-grained white mica. Very late sillimanite needles (Fig. 3c) rarely occur on kyanite, near the garnet and at the plagioclase boundaries.

Quartz and plagioclase show microstructural evidence of dynamic recrystallisation. Quartz has irregular, lobed grain boundaries interpreted as due to Grain Boundary Migration recrystallisation (Passchier and Trouw, 2005). Undulose extinction, often squarish (chessboard extinction, Fig. 3d), testifies a high temperature (≥ 650°C) deformation regime (Passchier and Trouw, 2005). Plagioclase shows lobed grain boundaries and in some cases deformation twinning is present. Strain-free grains of both minerals were also observed.

Within the sheared leucosomes, euhedral/subhedral plagioclase crystals, with well-developed crystal faces have been observed in some instances (Fig. 3f), testifying the heterogeneous nature of the deformation.

The stable Ti-phase in the matrix is ilmenite. Rutile is present only as relict cores in the former mineral and enclosed in kyanite and garnet. Other accessory phases are apatite (as the major phosphate), monazite, pyrite, zircon and tourmaline. These minerals occur in both garnet and matrix, whereas tiny xenotime grains were observed only in the matrix.

In spite of the strong deformation, several microstructural observations (see criteria review of Holness et al., 2011 and references therein) suggest that melt was present in the samples: (*i*) quartz grains with evidence of corrosion and rimmed by feldspar in quartzofeldspathic domains and within kyanite (Carosi et al., 2014a); (*ii*) tiny films of feldspars with cuspate low dihedral angles; (*iii*) “string of beads” microstructures (Fig. 3e); (*iv*) euhedral feldspar grains; and (*v*) “nanogranite” inclusions within peritectic garnet (Carosi et al., 2014a)**.**

From a petrological point of view the inferred “peak mineral assemblage” is interpreted to have been garnet-kyanite-biotite-plagioclase-quartz-white mica-rutile (+ melt), whereas ilmenite and the rare sillimanite needles are considered as post-peak minerals which grew during decompression and/or cooling. Chlorite locally on biotite and garnet suggests a very late fluid infiltration along some foliation planes (*e.g.* Vannay and Hodges, 1996). In a single case (sample K28c) hematite overgrowth on pyrite was detected within these alteration zones. It is very important to stress the lack of K-feldspar in the studied rocks (neither observed here nor reported by Vannay and Hodges, 1996 and Carosi et al., 2014a), while it was observed, together with prismatic sillimanite, in migmatites in similar structural position in the Sikkim Himalaya (Harris et al., 2004). The stromatic metatexite sample K28a was selected for a detailed petrological and geochronological investigation.

5. Methods

5.1 Mineral chemistry and compositional maps

After a careful optical inspection, K28a thin sections were studied with a CAMECA SX100 electron microprobe (EMP) hosted at Institut für Mineralogie und Kristallchemie (Universität Stuttgart) equipped with five wavelength dispersive spectrometers (WDS). The energy dispersive spectrometer (EDS) of the EMP was used for qualitative identification of minerals.

Chemical compositional maps (X-Ray mapping) were acquired on selected minerals/areas (micas and garnets) with a stepwise movement (100 ms per step) using an electron beam with a beam current of 60 nA, 150 nA, 30 nA for garnet, monazite and micas, respectively, and subsequent computer-aided evaluation. Garnets were mapped for Y, Ca, Mn, Fe, Mg, monazites for Y, Th, U, Ce, Si, and micas for Ba, Na, Mg, Fe, Ti.

Quantitative chemical analyses of points and transects were acquired on minerals present in all textural positions (matrix or included in porphyroblasts) using an acceleration voltage of 15 kV and a beam current of 15 nA. Monazite grains were analyzed following the procedure described in Massonne et al. (2007). Synthetic and natural standards were used for EMP calibration. The analytical uncertainties for the EMP measurements applied here are reported by Massonne (2012). For analyzing Zr in rutile (see below) a beam current of 100 nA and an acceleration voltage of 15 kV were selected. Structural formulae from mineral analyses were calculated with the software CALCMIN (Brandelik, 2009).

5.2 P-T estimates

In order to derive a P-T path, phase equilibria constraints, using pseudosection modelling, were derived for the selected sample (K28a). Previous authors have shown how powerful this tool is for bracketing the P-T evolution of Himalayan migmatites (*e.g.* Harris et al., 2004; Groppo et al., 2010, 2012; Guilmette et al., 2011).

P-T pseudosections were constructed for the P-T range of 0.3-1.3 GPa and 600-850°C, for a fixed bulk composition with the software PERPLE_X (*e.g.* Connolly 2005, version from August 2011, downloaded from the web site http://www.perplex.ethz.ch/). For this purpose we used the internally consistent thermodynamic database for minerals and water (CORK model, Holland and Powell 1991) given by Holland and Powell (1998, and updates). The following solid-solution (a-X) models were used: GlTsTsPg for amphibole, T for talc, Ctd(HP) for chloritoid, TiBio(HP) for biotite, Chl(HP) for chlorite, hCrd for cordierite, Gt(HP) for garnet, Opx(HP) for orthopyroxene, Omph(HP) for clinopyroxene, IlGkPy for ilmenite, Pheng(HP) for potassic white mica (with a maximum paragonite content of 50% mol), and St(HP) for staurolite (details on http://www.perplex.ethz.ch/perplex_solution_model_glossary.html). The models used for feldspars (plagioclase and K-feldspar) and paragonitic mica were reported by Massonne (2012 and references therein). Moreover, in order to calculate melting relationships, the model melt(HP) for haplogranitic melt (White et al., 2001) was used. The bulk composition was obtained with the X-ray fluorescence (XRF) spectrometer at the Earth Science Department of Pisa University, using the procedure of Tamponi et al. (2002-2003).

Calculations were performed in the MnO-Na_2O-CaO-K_2O-FeO-MgO-Al_2O_3-SiO_2-H_2O-TiO_2 system (MnNCKFMASHT). Titanium was included in order to determine the P-T stability of Ti-rich phases, while the O_2 (needed to consider ferric iron) was neglected because: *i)* magnetite is absent; *ii)* the amount of ferric iron in minerals is low and *iii)* rutile+ilmenite (± pyrite) should indicate low oxidation conditions (Diener and Powell, 2010; see also Groppo et al., 2010). The XRF composition was somewhat simplified for fitting the ten-component model chemical system as follows: *i)* CaO was reduced applying a correction for (ideally composed) apatite; *ii)* various amounts of H_2O were considered in the pseudosections calculations. The pseudosection results for 2.0 wt% of H_2O are presented here. Maybe such H_2O values are too high, but, for instance, Braga and Massonne (2012) have demonstrated for the Ulten Zone in the Eastern Alps that still higher water amounts can be present in HP anatectic metapelites. Calculations with different water (*e.g.* 2.5 wt%) or oxygen contents (despite the aforementioned negligance of O_2) have been explored resulting in minor changes of the pseudosection topology (*e.g.* magnetite occurrence, small T shift of the solidus, see also Massonne, 2014), with no significant changes in the considered phase-in boundaries (*e.g.* cordierite-in; K-feldspar-in) unless too low water amounts are assumed, stabilizing K-feldspar (not observed) in subsolidus assemblages (*e.g.* Massonne, 2014).

The pseudosections were contoured by molar amounts of phase components, like pyrope component in garnet (isopleths thermobarometry, *e.g.* Gaidies et al., 2006) and by modal amounts of phases (*e.g.* melt volume). Moreover, the "geothermobarometric potential of anatectic melts" as described by Massonne et al. (2013) and applied by Cruciani et al. (2014) has been explored comparing pseudosection predictions with results on "nanogranites" remelting experiments by Carosi et al. (2014a). The latter authors fully re-homogenized the crystallized melt inclusions during remelting experiments at a temperature of 820°C, a pressure of 1.2 GPa, and 24h experimental run, obtaining a melt composition with Si/Al ratio of 3.93 (± 0.25) and Na/K ratio of 2.7 (± 1.0) (Carosi et al., 2014a, their table 1).

The calculated raw P-T graphs were smoothed as shown by Connolly (2005). According to Massonne (2013), uncertainties of 10% on the P- and 5% on the T-estimates were considered for our P-T data resulting from the pseudosection modelling. For this reason, we also applied an independent method related to the Zr-in-rutile thermometry (Zack et al., 2004) using the pressure sensitive calibration of Tomkins et al. (2007). Recently, Ewing et al. (2013) demonstrated that the Zr-in rutile thermometer has good chances to record *"peak"* temperatures in high-grade rocks, especially for pristine (not-retrocessed/recrystallized) rutile, while Hallett and Spear (2014) testified that this thermometer provides important constraints for revealing the history of anatectic metapelites. Also Zr-in-rutile values, obtained for sample K28c by Carosi et al. (2014a), are reported for comparison.

5.3 Monazite in situ U-Th-Pb geochronology

Monazite, (LREE, Th)PO_4, can be a reliable geochronometer (*e.g.* Parrish, 1990). In recent years much attention has been paid in order to quantify its behaviour during deformation events (*e.g.* Williams and Jercinovic, 2002, 2012; Dumond *et al.* 2008) and metamorphic reactions (*e.g.* Foster and Parrish, 2003; Foster et al., 2000; Gibson et al., 2004; Spear and Pyle, 2002, 2010 and references therein). These studies have shown how monazite can record the timing of metamorphic processes for a wide spectrum of metamorphic conditions of the greenschist facies (*e.g.* Gasser et. al., 2012,) up to the granulite facies (*e.g.* Rubatto et al., 2001, 2013; Pyle and Spear 1999, 2003; Martins et al., 2009; Gasser et. al., 2012; Palin et al., 2014; Alcock et al., 2013; Massonne, 2014). Moreover, many efforts have been undertaken in order to link the chemistry of monazite to environmental parameters such as the temperature (see Spear and Pyle, 2002, for a review) using for example the monazite-xenotime thermometer (*e.g.* Spear and Pyle, 2002). *In situ* geochronological techniques offer the possibility of linking U-Th-Pb isotopic ages to particular chemical and/or textural domains related to metamorphic reactions and/or deformation events that could still be present also in unshielded matrix grains (*e.g.* Langone et al., 2011). For these reasons, in order to put time constraints in the evolution of the studied rock, monazite grains were carefully characterized.

Prior to isotopic dating of monazite grains, their textural position and internal features were imaged with a scanning electron microscope hosted at Earth Department of Pisa University, while the chemical characterization was achieved with the EMP as described above (section 5.1).

Monazite crystals were analysed *in situ* by laser-ablation, inductively coupled plasma mass spectrometry (LA–ICPMS) directly on 30 μm thick thin sections at the CNR–Istituto di Geoscienze e Georisorse U.O. Pavia (Italy) using an Ar–F 193-nm excimer laser (GeolLas 102 from Micro-Las) coupled with a magnetic sector ICP-MS (Element I from Thermo-Finnigan). The full description of the analytical procedure is reported in Paquette and Tiepolo (2007) and Tiepolo et al. (2003). Multiple laser spots are available for almost each grain using X-Ray maps as guide for spot position. Moreover, where it was possible, the laser spot was located at an area close to the EMP analytical spot.

Single analyses were performed by a one-minute acquisition of the background signal followed by recording, for at least 30 seconds, the ablation signal of the masses related to the isotopes ^{202}Hg, ^{204}(Hg+Pb), ^{206}Pb, ^{207}Pb, ^{208}Pb, ^{232}Th, and ^{238}U. The presence of common Pb was evaluated in each analysis on the basis of the net signal of ^{204}Pb (i.e. subtracted for the interference of ^{204}Hg and background). None of the sample revealed ^{204}Pb counts above the background level. However, the relatively high Hg signal in the gas blank does not exclude the effective presence of common Pb in the analysed monazite. Analytical conditions were 10 μm diameter of spot size, 12 J cm^{-2} of energy density, and 3 Hz of repetition rate. Time-resolved signals were carefully inspected to verify the presence of perturbations related to inclusions, fractures or mixing of different age domains. Laser-induced elemental fractionation and mass bias were corrected using matrix-matched external monazite standard (Moacir monazite: Seydoux- Guillaume et al., 2002a,b) considering the values, re-calibrated for isotopic disequilibrium, reported by Gasquet et al. (2010). In the analytical run eight to nine spots of the external standard were analysed. Only those close to the reference values (at least 4 in each run) were considered in order to reduce errors related to the standard reproducibility (Table A.1). External standards and unknowns were integrated over the same time intervals to ensure the efficient correction of fractionation effects. Data reduction was carried out with the GLITTER® software (van Achterbergh et al., 2001). In order to better estimate the uncertainty affecting the $^{206}Pb/^{238}U$, $^{207}Pb/^{235}U$ and $^{208}Pb/^{232}Th$ isotope ratios, the external reproducibility of the standard was propagated relative to individual uncertainties for the isotope ratios. After this error propagation each analysis is accurate within the quoted errors. The Isoplot 3.0 software by Ludwig (2003) has been used for age calculation and graphic representation.

6. Results

6.1 Mineral compositions (except monazite)

White mica has a composition showing a slight chemical variability, which deviates from ideal muscovite (Fig. 4a,c-Fig.5a). The contents of Si are between 3.10 and 3.16 a.p.f.u. (Fig. 5a) with the highest values observed mainly in the inner part of large grains (Fig. 4a,c) where typically relatively high Ti contents (up to 0.06 a.p.f.u.) were observed (Fig.4c). Mg/(Mg+Fe) (hereafter Mg#) is between 0.57-0.64. Moreover, paragonite (Fig. 5a) with XNa (*i.e.* Na/Na+K) values between 0.87 and 0.93 was detected within kyanite grains.

Biotite shows Mg# between 0.51 and 0.56 (Fig. 5b) with the highest value found in biotite included in garnet. Ti (a.p.f.u.) values range between 0.11 and 0.17 (Fig. 5b) with the lowest values occurring in biotite enclosed in garnet.

Garnet composition is slightly variable as shown by X-ray maps (Fig. 4c,d and Fig. 5c). This mineral is rich in almandine component (XAlm> 0.72). The garnet core is characterized by 23 mol% of pyrope component, 3 mol% of grossular (+ andradite) component and 2 mol% of spessartine component. Towards the rim, contents of Mn slightly increase (3% mol of spessartine), pyrope contents decrease down to 17 mol% and Ca contents are almost constant.

The composition of *plagioclase* varies with XAb between 0.86 and 0.92. The orthoclase component is always below 1% (Fig. 5d). A clear core to rim zoning is observed in matrix plagioclase were rims are somewhat richer in anorthite component.

Ilmenite contains some pyrophanite component (1.5-4 mol%). *Rutile* shows Cr (205-718 ppm) and Nb (750-15050 ppm) contents typical of metapelitic rocks (*e.g.* Meinhold 2010 and references therein). However, some analyses are also very rich in Nb (> 4000 ppm). Zr values between 270 – 650 ppm were determined. No clear correlation has been found between chemistry and rutile position within garnet, whereas lower Zr contents are present both in "pristine" rutile and more typically in the ilmenite-rimmed rutile. These chemical observations are valid for both K28a and K28c samples.

6.2 P-T constraints and P-T path

6.2.1 P-T Pseudosection

The P-T pseudosection for K28a (Fig. 6) shows a dominance of four-variance fields with quartz and plagioclase being ubiquitous. Garnet is absent at low P (fields n° 7, 8, and 9 in Fig. 6). Cordierite appears towards the high T side at low P, whereas biotite is completely consumed above *c.* 825°C. Melt is predicted to appear around 650°C. Rutile is stable above 0.70-0.80 GPa. Subsolidus kyanite occurs in a restricted P-T range (600-650°C, *c.* 0.7-0.8 GPa) whereas it is largely present at suprasolidus conditions. Interestingly, paragonite is present in the HP-LT corner of the pseudosection.

The assumed peak assemblage (garnet-kyanite-biotite-plagioclase-quartz-white mica-rutile + melt) is represented by a quatrovariant field (labeled as LWmPlGrtBtKyQzRt in Fig. 6) in the P-T range of *c.* 0.8-1.3 GPa and *c.* 650-800 °C, delimited by the disappearance of kyanite towards higher P and lower T (field n° 1) and by the appearance of K-feldspar (and ilmenite) towards higher high T and lower P. Melting along a nearly isobaric path, in this field is accompanied by the consumption of biotite-quartz-plagioclase and the formation of garnet, kyanite and white mica (for reactions of this type see Palin et al., 2014 and also King et al., 2011). Ilmenite and rare sillimanite, interpreted as post peak phases, formed later at relatively low P conditions (Fig. 6).

The relevant isopleths are displayed in Figs. 7a-d. XCa isopleths in garnet show a moderately positive dP/dT slope (Fig. 7a) and increasing Ca contents in garnet with rising pressure. Typically, the pyrope (XMg) content in garnet increases with rising temperature and pressure (Fig. 7a). This trend is opposed to the core to rim zoning of the studied garnets (Fig. 4d). For this reason, it is likely that the use of garnet compositions for deriving a P-T path, results only in minimum P-T conditions along a retrograde segment of the path.

Isopleths for Si contents in potassic white mica (Fig. 7b) show relatively flat dP/dT slopes at subsolidus and suprasolidus conditions. The albite content in plagioclase is predicted to increase with rising pressure (Fig 7b). The molar Si/Al ratio of the melt is mainly controlled by pressure (Fig 7.c), whereas the isopleths for Na/K ratio of the melt show moderate dP/dT slopes in agreement with results by Massonne et al. (2013) and Cruciani et al. (2014). The calculated melt volume (Fig. 7c) is below 10% in volume before crossing the white mica-out curve. After crossing a significant increase of the melt is observed ("*effective solidus*" of White et al., 2001).

6.2.2 Zr-in-Rutile

Consistent temperatures in the range of *c.* 650-720°C were obtained (Fig. 8), despite a large variation of Nb in rutile, a possible effect of which on the Zr-in-rutile thermometer was not experimentally constrained so far. Calculated temperatures close to 720°C can be easily explained to represent the “metamorphic T peak”, consistent with the pseudosection results (see below, Fig. 9). Lower temperatures (*c.* 650°C) were obtained for both rutile enclosed in “pristine” garnet and occurring in the matrix. This is interpreted as “prograde” T recorded in pristine rutile, whereas Zr remobilization (*e.g.* Luvinzotto and Zack, 2009) during rutile retrogression could be invoked for post “peak” temperatures obtained from rutile of the matrix.

6.2.3 P-T path

On the basis of the aforementioned petrography and the chemical compositions of relevant phases a P-T path was reconstructed (Fig. 9). The P-T conditions for garnet core formation are close to 700 °C and 1.0 GPa. Slightly higher pressures or lower temperatures are indicated by the highest Si contents in white mica, whereas somewhat higher temperature and pressure conditions are indicated by the Si/Al and Na/K ratios of the melt (despite the larger scatter, see above), the anorthite content in the plagioclase cores and the Zr-in-rutile thermometer. In this way a conservative P-T estimate for peak conditions reached by the studied sample K28a, is *c.* 710-720°C and 1.0-1.1 GPa (Fig. 9), where nearly 7 vol% melt is predicted to occur. Similar P-T results of 1.14 GPa -722°C (±0.21 GPa, ±34°C, sigfit 1.6) have been also obtained with AvePT thermobarometry (Powell and Holland, 1994) using the highest XMg, lowest XMn garnet composition with the average of matrix biotite, plagioclase and white mica. The compositions of the outermost garnet rim and white mica with the lowest Si contents on the main foliation provide a constraint that the retrograde path passed through P-T conditions of 650-670°C and 0.7-0.8 GPa, near the sillimanite-kyanite transition curve (Fig. 9). During this segment of the path characterized by decompression and cooling, garnet, white mica and the melt were consumed, whereas a minor increase in melt volume (≤ 1 % in vol) is predicted for the first stage (turning point) of the decompression. Plagioclase zoning, "reverse" zoning in garnet, and chemical compositions of potassic white mica and biotite are compatible with this retrograde segment of the P-T path. The proposed P-T path, lacking of large isothermal decompression, is also in accordance with the lack of K-feldspar in the sample (see also experimental results of Patiño-Douce and Harris, 1998).

Although speculative, the prograde P-T path characterized by both increasing P and T (dashed bold red curve in Fig. 9) is reconstructable based on the inclusion of rutile and paragonite in kyanite, the presence of "nanogranite" inclusions in peritectic garnet, and temperatures recorded by "pristine" rutiles (see paragraph 6.2.2).

6.3 Monazite dating

6.3.1 Monazite textural position and chemistry

Monazite was found in different microstructural positions (Fig. 10). This mineral is very common in both phyllosilicate and quartz-feldspar rich domains of the matrix forming grains up to 200 μm in diameter, which often have equilibrium grain faces with adjacent grains. Monazite shows inclusions of biotite, quartz and white mica. Small inclusions of monazite within porphyroblasts are much less common. In few cases (Mnz 19 in Fig. c) a grain completely armored in kyanite was detected. In one case a monazite grain was found to be located at the kyanite-quartz interface (Mnz 6). Monazite inclusions in garnet are also present. However, the complete shielding from the matrix was not always certain due to fractures in garnet. In one case (Mnz 1, Fig. 10a) monazite was found as inclusion in rutile within a garnet porphyroblast.

Backscattered electron (BSE) images show clear zoning in most grains with domains of different graytones (Fig. 10). X-ray compositional maps (Fig. 10a-f) and EMP analyses (Table

2, Fig. 11) reveal that such zoning is controlled by different distribution of HREE+Y *vs* LREE (Fig. 11) and Th. Matrix monazite often shows resorbed cores (*e.g.* Fig. 10c, e) with intermediate (Fig. 11) values of HREE+Y (1.5-1.9 wt% Y_2O_3) and variable Th contents. A brighter intermediate zone (mantle) is characterized by very low values of HREE+Y (Fig. 10e, Fig. 11) (0.2-0.7, wt% Y_2O_3) and generally higher Th contents. A darker outer zone, often forming a discontinuous rim on the intermediate zone, shows relatively high values of HREE+Y (2.5-3.3 wt% Y_2O_3). Monazite included in garnet is chemically similar to the cores of matrix monazite and monazite enclosed in kyanite. Mnz 19, fully included in kyanite, shows both the resorbed intermediate HREE+Y core and the mantle zone of the matrix monazite, whereas the high HREE rim is lacking. Such a rim domain has been observed in grain Mnz 6, which is only partially included in kyanite and in a few monazite grains within fracture zones in garnet.

In summary, based on the previously described textural and chemical arguments monazite shows three main growth domains/generations (Fig. 11) *i.e.*: *(i)* Mnz I with intermediate Y_2O_3 contents, is included in garnet and kyanite and occurs as resorbed cores in matrix monazite; *(ii)* Mnz II with very low HREE+Y contents is present in the matrix as mantle on resorbed cores and in monazite hosted in kyanite; *(iii)* Mnz III, forming discontinuous high HREE+Y rims, occurs only in the matrix. All three domains/generations are not always present in a single matrix grain. For instance, the Mnz I domain is absent in some grains. Nevertheless, monazite with all domains occurs in several matrix “Rosetta grains” (*sensu* Dahl et al., 2005) such as grain “Mnz 21” (Fig. 10e). Since the monazite HREE+Y budget in metapelites is largely controlled by garnet crystallization (*e.g.* Forster and Parrish, 2003; Spear and Pyle, 2010), the observed HREE+Y variations could be linked to several steps of garnet growth/resorption.

6.3.2 In-situ U-Th-Pb geochronology and age interpretations

The measured isotopic data are reported in Table 3, whereas results for the Moacir standard are reported in Table A.1. A total of 22 monazite grains covering the whole textural-chemical variability, were selected for *in situ* dating obtaining data from a total of 45 spots. According to Foster et al. (2000) the obtained results are plotted in a ^{238}U-^{206}Pb *vs* ^{232}Th-^{238}Pb concordia diagram (Fig. 12a). Also a probability density plot of Th-Pb ages is given (Fig. 12b). These ages are preferred for the discussion due to the ^{230}Th effect on U-Pb ages (*e.g.* Schärer 1984), even if this should be rather minor (*e.g.* Kellett et al., 2010). As it can be seen from Fig. 12, a large spread of ages ranging from 43 Ma to 18 Ma was determined. Ages in the range of 43-36 Ma were obtained from resorbed cores of matrix monazite and monazite included in garnet and kyanite (Mnz I). Ages around 29 Ma are recorded by the mantle of matrix monazite as well as monazite grains included within kyanite (Mnz II). Younger ages in the range of 25 Ma - 18 Ma are related to the high HREE+Y rims (Mnz III) of matrix grains. Large age variations are very common in medium-high grade metamorphic monazite (*e.g.* Foster et al., 2000, 2002; Martins et al., 2009; Rubatto et al., 2013; Massonne, 2014; Palin et al., 2014). Different mechanism could be invoked to explain such spread (see discussion in Foster et al., 2002): *(i)* presence of different chemical-age domains, as the result of monazite continuous or discontinuous growth; *(ii)* mixing of different age domains during laser ablation; *(iii)* Pb loss due to diffusive processes in a grain. Several studies (*e.g.* Spear and Pyle, 2002; Seydoux-Guillaume et al., 2002a; Cherniak et al., 2004; Gardés et al., 2007 and references therein) have shown that Pb diffusion in monazite is very slow and comparable with zircon. Thus, process (*iii*) can be excluded to explain the observed age spread. Mixing of different domains during ablation is possible despite careful spot position location and signal inspection, considering the relatively large ablation volume. However, common observations of systematic correlation of intra-crystalline zoning (for example in Y-HREE) with ages (*e.g.* Foster et al., 2000, 2002; Gibson et al., 2004; Williams and Jercinovic, 2002, 2012;) make the mechanism (*ii*) largely unlikely for being the main reason of the age spread. The observations in the present case, that the three different texturally and chemically (with different HREE+Y and Th contents) recognized populations (Mnz I-Mnz II-Mnz III see above) correspond to "age" populations, strongly support the idea that monazite records several growth stages along the experienced P-T path.

With these considerations, we interpret the obtained monazite ages as follow: Mnz I, due to its HREE+Y contents, grew during the prograde pre-melting (upper greenschist?-middle amphibolite facies?) part of the P-T path and experienced partial resorption during anatexis leading to the production of peraluminous melt (*e.g.* Pyle and Spear, 2003; Spear and Pyle, 2010). Mnz II, with its low HREE+Y contents, could be linked to the growth from a melt where garnet was part of the peritectic assemblage (*e.g.* Martins et al., 2009; Gasser et al., 2012) and partially shielded from the growth of the last generation (Mnz III) due to the entrapment in porphyroblasts. If this is true, crustal melting occurred in the time span between 36 and 28 Ma. Although Mnz III gave occasionally old ages as high as 30 Ma (mixing of domains ?), it is systematically younger than monazite of the other two groups. Because of high HREE+ Y contents this monazite (Mnz III) could be linked to the garnet breakdown during the exhumation stage accompanied by release of Y_2O_3 (including melt consuming back-reactions and/or a likely fluid infiltration) promoting monazite (Mnz III) re-growth. Assuming equilibrium conditions (see criteria reported by Spear and Pyle, 2002) between matrix xenotime and Mnz III with a X_{HREE+Y} of 0.096-0.087 (Table 2, Fig. 11, mole fractions calculated according to Pyle et al. 2001), the temperature span for crystallisation of Mnz III could be estimated in the range of *c.* 700-600°C (Table 2). This range is somewhat lower than the derived "peak" temperature condition and can be referred to cooling.

7. Discussions

7.1 P-T-D-t path of kyanite-bearing migmatite in the Kali Gandaki

Sheared migmatitic paragneiss of Unit 1 in the Kali Gandaki, according to petrological observations and thermodynamic calculations on sample K28a (see paragraph 6.2), experienced partial melting at "near peak" conditions around 710-720°C and 1.0-1.1 GPa (Figs. 9, 10) where kyanite-bearing tonalitic leucosomes developed. In addition to the presence of kyanite, the occurrence of rutile (within the supra-solidus porphyroblasts) is also diagnostic for melting at high pressure conditions, since the rutile-ilmenite transition is a useful pressure monitor in Barrovian pelitic rocks under reducing conditions (Weller et al., 2013; Massonne, 2014). According to Weller et al. (2013) rutile is typical of HP Barrovian metamorphism that is necessarily linked to a rock sequence metamorphosed below an overthrusting continental plate (see discussion in Weller et al., 2013).

The here estimated "near peak P-T" conditions (710-720°C, 1.0-1.1 GPa) are somewhat higher (610°C – 0.9 GPa, see paragraph 3) than the ones previously proposed by Vannay and Hodges (1996) using classical thermobarometry. In addition, the estimated temperatures are also *c.* 100°C lower than those resulting from remelting experiments of "nanogranites" reported by Carosi et al. (2014a). Although these authors considered the obtained melt composition as the true product of Himalayan anatexis, they regarded the obtained re-melting temperature unlikely for kinetic reasons linked to the experimental procedure (*e.g.* experimental time duration). This consideration is confirmed here, since temperatures of 800°C and more are not in agreement with the presence of white mica and the absence of K-feldspar in the rock and the result of the Zr-in-rutile thermometry.

The studied migmatitic paragneiss, after melting at near peak conditions, experienced decompression and cooling, associated with pervasive heterogeneous shearing, to reach P-T conditions of 650-670°C and 0.7-0.8 GPa. These conditions are consistent with the chessboard extinction observed in quartz. The retrograde part of the path is also compatible with an ilmenite-rutile transition at 0.7-0.8 GPa above 650°C (Massonne, 2014). Moreover, according to the melt isomodes (Fig. 7d) little melt (≤ 1.0 vol%) can be expected to have formed during the early stages of the here-inferred decompressional segment of the P-T path (see also Groppo et al., 2010, 2012 for higher T samples). It is worthy of note that the inferred P-T path has a similar shape as the P-T path proposed by Guillot (1999) (see his figure 5).

The timing of the P-T history has been constrained by *in situ* U-Th-Pb monazite dating. According to the obtained monazite ages the near peak-pressure melting of the sheared migmatitic paragneiss occurred in depths of 35-40 km during the ongoing prograde metamorphism (with a minimum starting age of 43 Ma, see also Carosi et al., 2010) in the time span between 36 and 28 Ma. This geochronological estimate, despite the different approaches (in situ dating *vs* mineral separation) and dating systematics, is compatible with the ages of c. 35-32 Ma reported by Godin et al. (2001) and of c. 41–36 Ma by Carosi et al. (2014a) for the GHS melting at HP conditions in the Kali Gandaki. Moreover, the present data potentially indicate that anatectic conditions could have been substained for c. 8-10 Myr (see also Palin et al., 2014). Starting from nearly 25 Ma up to 18 Ma the investigated rock experienced a substantial change of the dP/dT slope of their path and started to be exhumed (decompression and cooling segment of the P-T path) accompanied by melt crystallization, garnet breakdown and Mnz III formation (Fig. 10).

7.2 Formation of kyanite-bearing migmatites in the Himalaya

Along the Himalayan belt several kyanite (+ late sillimanite) bearing migmatites, occurring at different structural positions within the GHS, have been studied so far with respect to their metamorphic and temporal evolution. They are given here according to their regional position from east to west and compared with our results:

Guilmette et al. (2011) studied kyanite-bearing anatectic paragneiss from the Eastern Himalayan Syntaxis. These HP granulitic rocks experienced "peak conditions" of 820°C at pressures higher than 1.4 GPa (possibly 1.5 – 1.6 GPa), followed by decompression, cooling and melt solidification at 810°C and 0.9 GPa. Geochronological information on HP melting in the Eastern Himalayan Syntaxis was reported by Palin et al. (2014). These authors identified two high-grade metamorphic (and melting) events in their studied migmatites. The first one (71-50 Ma), within the sillimanite stability field, is related to the pre-collisional tectonic history of the Lhasa block. The second event, producing migmatites in the kyanite stability field and tonalitic leucosomes, occurred at minimum P-T conditions of nearly 700°C and 1.04 GPa (based on THERMOCALC Average P-T method) in the time span of 44–33 Ma (based on U-Th-Pb monazite geochronology). This event, that lasted nearly 10 Ma, was related by Palin et al. (2014) to the India-Asia collision.

Davidson et al. (1997) and Daniel et al. (2003) reported similarly deformed anatectic metasediments from the Bhutan Himalaya just few hundreds of metres above the MCT and, thus, in a structural position comparable with the studied samples. Davidson et al. (1997) suggested minimum P-T conditions of 0.8 GPa and >700°C, while Daniel et al. (2003), based on conventional geothermobarometry (see their Fig. 9), proposed P-T conditions of *c.* 750-800°C and 1.2 - 1.3 GPa, at which melting (*i.e.* in the kyanite stability field) occurred. The latter authors reported also U-Pb ages on separated monazite and xenotime grains suggesting the formation of kyanite-bearing migmatites around 18-16 Ma.

In the kyanite-sillimanite migmatitic gneisses of the Sikkim Himalaya (Harris et al., 2004) pre-decompressional garnet growth is dated at 23 ± 3 Ma (Sm-Nd systematics), while garnet growth at near-peak temperatures (750 °C - 0.8 GPa) occurred at 16 ± 2 Ma during the melting stage. According to Harris et al. (2004) the major melting stage is interpreted as the result of the decompression to the sillimanite stability field. Despite their similar approach for

P-T estimates (*i.e.* pseudosection), in the Kali Gandaki migmatites studied here, petrographic and petrological data testify that melting already occurred within the kyanite stability field (as also suggested by Daniel et al., 2003; see also Groppo et al., 2010), well before the Miocene STDS-related decompression and the sillimanite melting stage (overprinting ?).

In Far Eastern Nepal along the Arun-Makalu transect Groppo et al. (2010) described the P-T-t evolution of a kyanite-bearing migmatitic sample, coming from the structurally highest portion of the lower GHS, where also K-feldspar was part of the main peak assemblage. Their studied sample reached P-T conditions of 820°C, 1.3 GPa around 31 Ma (Early Oligocene) and then followed decompression, cooling and melt back-reactions down to c. 800 °C, 1.0 GPa in the time span of 27-29 Ma (see their figure 9). Also, Imayama et al. (2012) studied kyanite – sillimanite migmatites from Far-Eastern Nepal. These authors, combining pseudosections and trace-element constrained U-Pb zircon ages, demonstrated that their kyanite-sillimanite migmatites experienced melting at c. 21-18 Ma and isothermal decompression from P-T peak of 0.8–1.4 GPa and 720–770°C.

Sillimanite (after kyanite)-bearing migmatites from the Kharta valley in Tibet near Mt. Everest, studied by Liu et al. (2007), had experienced somewhat higher pressures and temperatures (1.4 GPa and 750-800°C) than our migmatites. However, the temporal evolution is well comparable with our results as the HP event for the Kharta valley rocks were dated at 33 ± 2 Ma (Liu et al., 2007). Also the retrograde evolution of these rocks at 23 ± 2 Ma (Liu et al., 2007) coincides with the time range of 18-25 Ma determined for our rocks.

The aforementioned different P-T-t conditions estimated for the formation of kyanite-bearing GHS migmatites, could be partially explained by different geothermobarometric and dating methods applied by the various authors. In part these differences can be due to samples taken from different structural positions. Despite these differences, the present data indicate that (1) melting at HP conditions produced tonalitic or leucogranitic melts (as function of the P-T conditions) already during the early stages of the India-Asia collision forming the Himalaya and, (2) other portions of the belt probably reached HP condition (and melting) at different times (up to the Miocene).

7.3 Tectonic implications

Several discontinuities identified within the GHS (*e.g.* Carosi et al., 2007; 2010; Corrie and Kohn, 2011; Imayama et al., 2012; Larson et al., 2013; Montomoli et al., 2013; see Montomoli et al., 2014 for a review) have shown that the GHS has a much more complex crustal architecture compared to simple models where a single coherent tectonic unit is bounded by only two tectonic discontinuities with opposite sense of shear. These recent findings are also compatible with diachronic melting within the GHS (Kohn et al., 2005; Corrie and Kohn, 2011; Imayama et al., 2012; Rubatto et al., 2013). As anticipated above our samples are localized ~ 1 km northern than the MCT location according to Colchen et al. (1986) and Vannay and Hodges (1996), recently mapped as CT by Parsons et al. (2014) whereas the MCT has been shifted dozen km to the South according to Searle (2010). The different localization of the MCT could arise some ambiguities on the tectonic meaning of our present results. Anyway, Montomoli et al. (2014) discussed how the location of MCT *vs* P-T-D-t discontinuities could be problematic since several processes (*e.g.* shear zone widening; ductile thinning due to pure shear component of deformation) could complicate the structural pattern in the GHS. They suggested that to characterize tectonic discontinuities field observations are not unique and a multidisciplinary approach (joining structural, metamorphic and chronological information) could help to solve ambiguity. Indeed, to better unravel the structural (and melting) evolution of the GHS, monazite geochronology could help in assessing the ages of the activity of the different tectonic discontinuities (*e.g.* Kohn et al., 2005; Corrie and Kohn, 2011). In particular, in the present case study, the rim monazite ages (*c.* 25-18 Ma, Mnz III), interpreted as retrograde, are useful for this assessment. The here presented Mnz III ages are older than the quoted ages for the MCT activity in the Kali Gandaki (*c.* 21 –16 Ma in Gibson et al. 2014; *c.* 22 Ma in Godin et al., 2006) and share much more similarities with intra GHS in sequence contractional shear zones, like the High Himalayan Discontinuity (HHD) of Montomoli et al. (2013, 2014). The kinematics, the P-T-t path and the age of the studied sample testify, for the first time, the occurrence of the HHD in the Kali Gandaki valley (Fig. 13). This occurrence at 25-18 Ma, in a structural higher position and older with respect to the MCT, dated at 22-16 Ma along the same section, proves a southward shifting of the shearing and exhumation withing the GHS (Fig. 13).

It is also important to stress that in this contribution the monazite ages related to the starting of the kyanite growth are significantly older compared to other kyanite-bearing gneiss ages within the MCT zone reported along other Himalayan transects (*e.g. c.* 22-15 Ma in Larson et al., 2013; 17-13Ma in Montomoli et al., 2013). This means that kyanite in the MCT zone grew at different times.

It also appears that the GHS melting history is not so simple as proposed in the current geodynamic models (*e.g.* extrusion and channel flow). Improved future models should fully account for the different melting scenarios and melting timing affecting the whole GHS such as: high pressure – decompressional – and low pressure (with peritectic andalusite) melting (see Visonà et al., 2012 and references therein). With respect to this point, we note that melting was produced at different times within the different GHS slices. The associated decrease of rock viscosity (*e.g.* Jamieson et al., 2011 and references therein) could be responsible for strain softening processes with the consequence of localization of deformation resulting in shear-zone nucleation and decoupling of the diverse GHS slices.

Conclusions

A comprehensive P-T-t-D path for kyanite-bearing migmatitic paragneiss of the GHS in the Kali Gandaki has been reconstructed integrating meso- and micro-structural and petrographic data, P-T estimates and *in situ* monazite geochronology (Fig. 13). The gneiss underwent prograde metamorphism from (at least) 43 to 28 Ma and experienced partial melting at P-T conditions of 710-720°C/1.0-1.1 GPa in the time span of 36-28 Ma, producing kyanite-bearing, K-feldspar-poor leucosomes (Fig. 13). This rock was subject to decompression and cooling associated with pervasive shearing during 25-18 Ma prior and at an upper structural position with respect to the MCT activity (Fig. 13). This study testifies for the first time the occurrence of a structural and tectonic discontinuity within the GHS in the Kali Gandaki valley and confirms its regional extent (HHD; Montomoli et al., 2013, 2014) and the occurrence of diachronous exhumation of the two portions of the GHS divided by the HHD. Exhumation in the GHS was not triggered only by the MCT but started, before it was formed, in the upper part of the GHS.

It is also suggested that the migmatitic paragneiss experienced a clockwise P-T loop as shown in Figure 9 represents a potential source of "high-Ca" melts recognized in some parts of the Himalaya (*e.g.* King et al., 2011), but the extent and the chronology of this HP melting event along the Himalaya strike deserves further attention in the future.

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Bibliography

Antolín, B., Appel, A., Montomoli, C., Dunkl, I., Ding, L., Gloaguen, R., El Bay, R., 2011. Kinematic evolution of the eastern Tethyan Himalaya: constraints from magnetic fabric and structural properties of the Triassic flysch in SE Tibet. In: Poblet, J., Lisle, R. (Eds.), Kinematic Evolution and Structural Styles of Fold-and-Thrust Belts: Geological Society of London Special Publications 349, pp. 99–121. http://dx.doi.org/10.1144/SP349.6.

Alcock, J., Muller, P.D., Jercinovic, M.J., 2013. Monazite ages and pressure–temperature–time paths from anatectites in the southern Ruby Range, Montana, USA: evidence for delamination, ultramafic magmatism, and rapid uplift at ca. 1780 Ma. Canadian Journal of Earth Sciences 50, 1069-1084, 10.1139/cjes-2013-0035

Beaumont, C., Jamieson, R.A., Nguyen, M.H., Lee, B., 2001. Himalayan tectonics explained by extrusion of a low-viscosity crustal channel coupled to focused surface denudation. Nature 414, 738–742.

Borghi, A., Castelli, D., Lombardo, B., Visona, D., 2003. Thermal and baric evolution of garnet granulites from the Kharta region of S Tibet, E Himalaya. European Journal of Mineralogy 15, 401–418.

Braga, R., Massonne, H.-J., 2012. H_2O content of deep-seated orogenic continental crust: the Ulten Zone, Italian Alps. International Geology Review 54, 633-641.

Brandelik, A., 2009. CALCMIN–an EXCEL™ Visual Basic application for calculating mineral structural formulae from electron microprobe analyses source. Computers and Geosciences 35, 1540-1551.

Brown, R.L., Nazarchuk, J.H., 1993. Annapurna detachment fault in the Greater Himalaya of Central Nepal. Geological Society of London 74, 41-473.

Burchfiel, B.C., Zhiliang, C., Hodges, K.V., Yuping, L., Royden, L.H., Changron, D., Jiene, X., 1992. The South Tibetan Detachment System, Himalayan Orogen. Geological Society of America Special Paper 269, 41 pp.

Burg, J.P., Brunel, M., Gapais, D., Chen, G.M., Liu, G.H., 1984. Deformation of leucogranites of the crystalline Main Central Sheet in southern Tibet (China). Journal of Structural Geology 6, 535–542.

Caby, R., Pêcher, A., Le Fort, P., 1983. Le grand chevauchement central himalayen: Nouvelles données sur le métamorphisme inverse à la base de la Dalle du Tibet. Revue de Géologie Dynamique et de Géographie Physique 24, 89–100.

Carosi, R., Lombardo, B., Molli, G., Musumeci, G., Pertusati, P.C., 1998. The south Tibetan detachment system in the Rongbuk valley, Everest Region. Deformation features and geological implication. Journal of Asian Earth Sciences 16, 299–311.

Carosi, R., Montomoli, C., Visonà, D., 2002. Is there any detachment in the Lower Dolpo (western Nepal)? Comptes Rendus Geoscience 334, 933-940.

Carosi, R., Montomoli, C., Visonà, D., 2007. A structural transect in the Lower Dolpo: Insights on the tectonic evolution of Western Nepal. Journal of Asian Earth Sciences 29, 407–423, doi:10.1016/j.jseaes.2006.05.001.

Carosi, R., Montomoli, C., Rubatto, D., Visonà D., 2010. Late Oligocene high-temperature shear zones in the core of the Higher Himalayan Crystallines (Lower Dolpo, western Nepal). Tectonics 29, DOI: 10.1029/2008TC002400.

Carosi, R., Montomoli, C., Langone, A., Turina, A., Cesare, B., Iaccarino, S., Fascioli, L., Visonà, D., Ronchi, A., Rai, S.M., 2014a. Eocene partial melting recorded in peritectic garnets from kyanite-gneiss, Greater Himalayan Sequence, central Nepal. Geological Society, London, Special Publications, 412, doi:10.1144/SP412.1

Carosi, R., Gemignani, L., Godin, L., Iaccarino, S., Larson, K., Montomoli, C., Rai, S.M., 2014b. A geological journey in the deepest canyon of the earth: the Kali Gandaki section. In: (Eds.) Chiara Montomoli, Rodolfo Carosi, Rick Law, Sandeep Singh, and Santa Man Rai, Geological field trips in the Himalaya, Karakoram and Tibet, Journal of the Virtual Explorer, Electronic Edition, ISSN 1441-8142, volume 47, paper 9.

Cherniak, D.J., Watson, E.B., Grove, M., 2004. Pb diffusion in monazite: a combined RBS/SIMS study. Geochimica et Cosmochimica Acta 68, 829–840. http://dx.doi.org/10.1016/j.gca.2003.07.012.

Colchen, M., Le Fort, P., Pêcher, A., 1986. Notice explicative de la carte géologique Annapurna–Manaslu–Ganesh (Himalaya du Népal) au 1:200 000e (bilingual: French–English), CNRS, Paris, 1986.

Connolly, J.A.D., 2005. Computation of phase equilibria by linear programming: A tool for geodynamic modeling and its application to subduction zone decarbonation. Earth and Planetary Science Letters 236, 524–541.

Corrie, S.L., Kohn, M.J., 2011. Metamorphic history of the central Himalaya, Annapurna region, Nepal and implications for tectonic models. Geological Society of America Bulletin 123, 18563-1879.

Cruciani, G., Fancello, D., Franceschelli, M., Scodina M., Spano, M.E., 2014. Geothermobarometry of Al-silicate-bearing migmatites from the Variscan chain of NE Sardinia, Italy: a P-T pseudosection approach. Periodico di Mineralogia 83, 19-40.

Dahl, S.P., Terry, M.P., Jercinovic, M.J., Williams, M.L., Hamilton, M.A., Foland, K.A., Clement, S.-M., Friberg, L.M., 2005. Electron probe (Ultrachron) microchronometry of metamorphic monazite: Unraveling the timing of polyphase thermotectonism in the easternmost Wyoming Craton (Black Hills, South Dakota). American Mineralogist 90, 1712–1728.

Daniel, C.G., Hollister, L.S., Parrish, R.R., Grujic, D., 2003. Exhumation of the Main Central Thrust from lower crustal depths, eastern Bhutan Himalaya. Journal of Metamorphic Geology 21, 317–334.

Davidson, C., Grujic, D.E., Hollister, L.S., Schmid, S.M., 1997. Metamorphic reactions related to decompression and synkinematic intrusion of leucogranite, High Himalayan Crystallines, Bhutan. Journal of Metamorphic Geology 15, 593–612.

Diener, J.F.A., Powell, R., 2010. The influence of ferric iron on the stability of mineral assemblages. Journal of Metamorphic Geology 28, 599–613.

Dumond, G., McLean, N., Williams, M. L., Jercinovic, M.J., Bowring, S.A., 2008. High resolution dating of granite petrogenesis and deformation in a lower crustal shear zone: Athabasca granulite terrane, western Canadian Shield. Chemical Geology 254, 175–196.

Dunkl, I., Antolín, B., Wemmer, K., Rantitsch, G., Kienast, M., Montomoli, C., Ding, L., Carosi, R., Appel, E., El Bay, R., Xu, Q., von Eynatten, H., 2011. Metamorphic evolution of the Tethyan Himalayan flysch in SE Tibet. In: Gloaguen, R., Ratschbacher, L. (Eds.), Growth and Collapse of the Tibetan Plateau: Geological Society of London Special Publications 353, 45–69. http://dx.doi.org/10.1144/SP353.4.

Ewing, T., Hermann, J., Rubatto, D., 2013. The robustness of the Zr-in-rutile and Ti-in-zircon thermometers during high-temperature metamorphism (Ivrea-Verbano Zone, northern Italy). Contributions to Mineralogy and Petrology 165, 757-779.

Foster, G., Parrish, R.R., 2003. Metamorphic monazite and generation of P–T–t paths. In: Vance, D., Müller, W. & Villa, I. M. (eds) Geochronology: Linking the Isotopic Record with Petrology and Textures. Geological Society, London, Special Publications 220, 25–47.

Foster, G., Kinny, P., Vance, D., Prince, C., Harris, N., 2000. The significance of monazite U–Th–Pb age data in metamorphic assemblages; a combined study of monazite and garnet chronometry. Earth and Planetary Science Letters 181, 327–340.

Foster, G.L., Gibson, H.G., Parrish, R.R., Horstwood, M.S.A., Fraser, J., Tindle, A., 2002. Textural, chemical and isotopic insights into the nature and behavior of metamorphic monazite. Chemical Geology 191, 183–207. http://dx.doi.org/10.1016/S0009-2541(02)00156-0.

Gaidies, F., Abart, R., De Capitani, C., Schuster, R., Connolly, J.A.D., Reusser, E., 2006. Characterization of polymetamorphism in the Austroalpine basement east of the Tauern Window using garnet isopleth thermobarometry. Journal of Metamorphic Geology 24, 451-475.

Gardés, E., Montel, J.M., Seydoux-Guillaume, A.M., Wirth, R., 2007. Pb diffusion in monazite: New constraints from the experimental study of $Pb^{2+}\leftrightarrow Ca^{2+}$ interdiffusion. Geochimica et Cosmochimica Acta 71, 4036–4043. DOI: 10.1016/j.gca.2007.06.036

Gasser, D., Bruand, E., Rubatto, D., Stüwe, K., 2012. The behaviour of monazite from greenschist facies phyllites to anatectic gneisses: An example from the Chugach Metamorphic Complex, southern Alaska. Lithos 134-135, 108-122.

Gansser, A., 1964. Geology of Himalayas. Wiley Interscience, London, 289.

Gasquet, D., Bertrand, J.-M., Paquette, J.-L., Lehmann, J., Ratzov, G., De Ascençâo Guedes, R., Tiepolo, M., Boullier, A.-M., Scaillet, S., Nomade, S., 2010. Miocene to Messinian deformation and hydrothermal activity in a pre-Alpine basement massif of the French western Alps: new

U–Th–Pb and argon ages from Lauzière massif. Bulletin de la Société Géologique de France 181, 227–241.

Gibson, H.D., Carr, D.S., Brown, L.R., Hamilton, A.M., 2004. Correlations between chemical and age domains in monazite, and metamorphic reactions involving major pelitic phases: an integration of ID‒TIMS and SHRIMP geochronology with Y–Th–U X‒ray mapping. Chemical Geology 211, 237–260.

Gibson, R., Godin, L., Cottle, M.J., Kellett, D.A., 2014. Along-strike strain variation and timing of deformation in the lower Himalayan metamorphic core, west-central Nepal. In Montomoli C., et al., eds., proceedings for the 29th Himalaya-Karakoram-Tibet Workshop, Lucca, Italy. Journal of Himalayan Earth Sciences (Special Volume), 45-47.

Godin, L., Brown, R.L., Hanmer, S., Parrish, R., 1999a. Back folds in the core of the Himalayan orogen: An alternative interpretation. Geology 27, 151-154.

Godin, L., Brown, R.L., Hanmer, S., 1999b. High strain zone in the hanging wall of the Annapurna detachment, central Nepal Himalaya, in Macfarlane, A., Sorkhabi, R. B., and Quade, J., eds., Himalaya and Tibet: Mountain roots to mountain tops: Geological Society of America (Special Paper) 328, 199–210.

Godin, L., Parish, R.R., Brown, L., Hodges, K.V., 2001. Crustal thickening leading to exhumation of the Himalayan Metamorphic core of central Nepal: insight from U-Pb geochronology and 40Ar/39Ar thermochronology. Tectonics 20, 729-747.

Godin, L., 2003. Structural evolution of the Tethyan sedimentary sequence in the Annapurna area, central Nepal Himalaya. Journal of Asian Earth Sciences 22, 307–328.

Gratz, R., Heinrich, W., 1997. Monazite–xenotime thermometry: experimental calibration of the miscibility gap in the system $CePO_4$–YPO_4. American Mineralogist 82, 772–780.

Groppo, C., Rubatto, D., Rolfo, F., Lombardo, B., 2010. Early Oligocene partial melting in Main Central Thrust Zone (Arun valley, eastern Nepal Himalaya). Lithos 118, 287–301.

Groppo, C., Rolfo, F., Indares, A., 2012. Partial melting in the Higher Himalayan Crystallines of Eastern Nepal: the effect of decompression and implications for the 'Channel Flow' model. Journal of Petrology 53, 1057–1088.

Guillot, S., 1999. An overview of the metamorphic evolution in Central Nepal. Journal of Asian Earth Sciences 17, 713-725.

Guilmette, C., Indares, A., Hébert, R., 2011. High-pressure anatectic paragneisses from the Namche Barwa, Eastern Himalayan Syntaxis: Textural evidence for partial melting, phase equilibria modeling and tectonic implications. Lithos 124, 66–81.

Hallett, B.W., Spear, F.S., 2014. The P-T history of anatectic pelites of the Northern East Humboldt Range, Nevada: Evidence for tectonic loading, decompression, and anatexis. Journal of Petrology 55, 3-36.

Harris, N.B.W., Caddick, M., Kosler, J., Goswami, S., Vance, D., Tindle, A.G., 2004. The pressure-temperature-time path of migmatites from the Sikkim Himalaya. Journal of Metamorphic Geology 22, 249-264.

Heim, A., Gansser, A., 1939. Central Himalaya: geological observations of the Swiss expedition 1936. Memoir Society Helvetica Science Nature 73, 1–245.

Herron, M.M., 1988. Geochemical classification of terrigenous sands and shales from core or log data. Journal of Sedimentary Research 58, 820-829, doi:10.1306/212F8E77-2B24-11D7-8648000102C1865D.

Hodges, K.V., 2000. Tectonics of Himalaya and southern Tibet from two perspectives. Geological Society of America Bulletin 112, 324–350.

Holness, M., Cesare, B., Sawyer, E.W., 2011. Melted rocks under the microscope: microstructures and their interpretation. Elements 7, 247–252.

Holland, T.J.B., Powell, R., 1991. A compensated-Redlich-Kwong (CORK) equation for volumes and fugacities of CO_2 and H_2O in the range 1 bar to 50 kbar and 100-1600°C. Contributions to Mineralogy and Petrology 109, 265-273.

Holland, T.J.B., Powell, R., 1998. An internally consistent thermodynamic data set for phases of petrological interest. Journal of Metamorphic Geology 16, 309-343.

Jamieson, R.A., Beaumont, C., Medvedev, S., Nguyen, M.H., 2004. Crustal channel flows: 2. Numerical models with implications for metamorphism in the Himalayan-Tibetan orogen. Journal of Geophysical Research 109, B06407, doi:10.1029/2003JB002811.

Jamieson, R.A., Unsworth, M.J., Harris, N.B.W., Rosenberg, C.L., Schulmann, K., 2011. Crustal melting and the flow of mountains. Elements 7, 253-260.

Kellett, D.A., Grujic, D., Warren, C., Cottle, J., Jamieson, R., Tenzin, T., 2010. Metamorphic history of a syn-convergent orogen-parallel detachment: the South Tibetan detachment system, Bhutan Himalaya. Journal of Metamorphic Geology 28, 785–808. http://dx.doi.org/10.1111/j.1525-1314.2010.00893.x.

King, J., Harris, N., Argles, T., Parrish, R., Zhang, H. 2011. Contribution of crustal anatexis to the tectonic evolution of Indian crust beneath southern Tibet. Geological Society of America Bulletin 123, 218–239.

Kohn, M.J., Wieland, M., Parkinson, C.D., Upreti, B.N., 2005. Five generation of monazite in Langtang gneisses: implication for chronology of the Himalayan metamorphic core. Journal of Metamorphic Geology 23, 399–406.

Imayama, T., Takeshite, T., Yi, K., Cho, D.-L., Kitajima, K., Tsutsumi, Y., Kayama, M., Nishido, H., Okumura, T., Yagi, K., Itaya, T., Sano, Y., 2012. Two-stage partial melting and contrasting cooling history within the Higher Himalayan Crystalline Sequence in the far-eastern Nepal Himalaya. Lithos 134–135, 1–22.

Langone, A., Braga, R., Massonne, H.-J., Tiepolo, M., 2011. Preservation of old (prograde metamorphic) U–Th–Pb ages in unshielded monazite from high-pressure paragneiss of the Variscan Ulten Zone. Lithos 127, 68–85.

Larson, K.P., Godin, L., 2009. Kinematics of the Greater Himalayan sequence, Dhaulagiri Himal: implications for the structural framework of central Nepal. Journal of the Geological Society 166, 25-43.

Larson, K.P., Gervais, F., Kellett, D.A., 2013. A P–T–t–D discontinuity in east-central Nepal: implications for the evolution of the Himalayan mid-crust. Lithos 179, 275–292.

Le Fort, P., 1975. Himalaya: the collided range. American Journal of Science 275, 1–44.

Le Fort, P., Pêcher, A., Upreti, B.N., 1986. A section through the Tibetan Slab in cental Nepal (Kali Gandaki Valley): mineral chemistry and thermobarometry of the Main Central Thrust Zone. Sciences de la Terre, Nancy, 47, 211-228.

Liu, Y., Siebel, W., Massonne, H.-J., Xiao, X., 2007. Geochronological and petrological constraints for the tectonic evolution of the central Greater Himalayan Sequence in the Kharta area, southern Tibet. Journal of Geology 115, 215-230.

Ludwig, K.R., 2003. Isoplot/Ex version 3.0. A geochronological toolkit for Microsoft Excel. Berkeley Geochronology Center, Special Publications, 4.

Luvinzotto, G.L., Zack, T., 2009. Nb and Zr behavior in rutile during high-grade metamorphism and retrogression: An example from the Ivrea–Verbano Zone. Chemical Geology 261, 303-317.

Martins, L., Vlach, S.R.F., Janasi, V.A. 2009. Reaction microtextures of monazite: correlation between chemical and age domains in the Nazaré Paulista migmatite, SE Brazil. Chemical Geology 261, 271–285.

Massonne, H.-J., 2012. Formation of amphibole and clinozoisite-epidote in eclogite owing to fluid infiltration during exhumation in a subduction channel. Journal of Petrology 53, 1969-1998, doi: 10.1093/petrology/egs040.

Massonne, H.-J., 2013. Constructing the pressure-temperature path of ultrahigh-pressure rocks. Elements 9, 267–279.

Massonne, H.-J., 2014. Wealth of P–T–t information in medium-high grade metapelites: Example from the Jubrique Unit of the Betic Cordillera, S Spain. Lithos 208–209, 137–157.

Massonne, H.-J., Cruciani, G., Franceschelli, M., 2013. Geothermobarometry on anatectic melts – a high pressure Variscan migmatite from northeast Sardinia. International Geology Review 55, 1490-1505.

Massonne, H.-J., Kennedy, A., Nasdala, L., Theye, T., 2007. Dating of zircon and monazite from diamondiferous quartzofeldspathic rocks of the Saxonian Erzgebirge — hints at burial and exhumation velocities. Mineralogical Magazine 71, 407–425.

Meinhold, G., 2010. Rutile and its applications in earth sciences. Earth Science Reviews 102, 1-28.

Montomoli, C., Iaccarino, S., Carosi, R., Langone, A., Visonà, D., 2013. Tectonometamorphic discontinuities within the Greater Himalayan Sequence in Western Nepal (Central Himalaya): insights on the exhumation of crystalline rocks. Tectonophysics 608, 1349–1370.

Montomoli, C., Carosi, R., Iaccarino, S., 2014. Tectonometamorphic discontinuities in the Greater Himalayan Sequence: a local or a regional feature? Geological Society, London, Special Publications, 412, http://dx.doi.org/10.1144/SP412.3

Najman, Y., Apple, E., Boudagher-Fadel, M., Bown, P., Carter, A., Garzanti, E., Godin, L., Han, J., Liebke, U., Oliver, G., Parrish, R., Vezzoli, G., 2010. Timing of India-Asia collision: Geological, biostratigraphic, and palaeomagnetic constraints. Journal of Geophysical Research: Solid Earth 115, DOI: 10.1029/2010JB007673

Palin, R.M., Searle, M.P., St-Onge, M.R., Waters, D.J., Roberts, N.M.W., Horstwood, M.S.A., Parrish, R.R., Weller, O.M., Chen, S., Yang, J., 2014. Monazite geochronology and petrology of kyanite- and sillimanite-grade migmatites from the northwestern flank of the eastern Himalayan syntaxis. Gondwana Research 26, 323–347.

Paquette, J.L., Tiepolo, M., 2007. High resolution (5 μm) U–Th–Pb isotope dating of monazite with excimer laser ablation (ELA)–ICPMS. Chemical Geology 240, 222–237.

Parrish, R.R., 1990. U–Pb dating of monazite and its applications to geological problems. Canadian Journal of Earth Sciences 27, 1431–1450.

Parsons, A.J., Law, R.D., Searle, M.P., Phillips, R.J., Lloyd, G.E., 2014. Geology of the Dhaulagiri-Annapurna-Manaslu Himalaya, Western Region, Nepal, 1:200,000. Journal of Maps, DOI: 10.1080/17445647.2014.984784.

Passchier, C.W., Trouw, R.A.J., 2005. Microtectonics. In: Springer (Ed.), p. 366.

Patiño-Douce, A., Harris, N., 1998. Experimental constraints on Himalayan anatexis. Journal of Petrology 39, 689–710.

Powell, R., Holland, T.J.B., 1994. Optimal geothermometry and geobarometry. American Mineralogist 79, 120-133.

Prince, C.P., Harris, N., Vance, D., 2001. Fluid enhanced melting during prograde metamorphism. Journal of the Geological Society London 158, 233–241.

Pyle, J.M., Spear, F.S., 1999. Yttrium zoning in garnet: coupling of major and accessory phase during metamorphic reactions. Geologica Material Research 1, 1–49.

Pyle, J.M., Spear, F.S., Rudnick, R.L., McDonough, F.W., 2001. Monazite-xenotime-garnet equilibrium in metapelites and a new monazite-garnet thermometer. Journal of Petrology 42, 2083-2107.

Pyle, J.M., Spear, F.S., 2003. Four generations of accessory-phase growth in lowpressure migmatites from SW New Hampshire. American Mineralogist 88, 338–351.

Rosenberg, C.L., Handy, M.R., 2005. Experimental deformation of partially melted granite revisited: Implications for the continental crust. Journal of Metamorphic Geology, 23, 19-28.

Rubatto, D., Williams, I.S., Buick, I.S., 2001. Zircon and monazite response to prograde metamorphism in the Reynolds Range, central Australia. Contributions to Mineralogy and Petrology 140, 458–468.

Rubatto, D., Chakraborty, S., Dasgupta, S., 2013. Timescale of crustal melting in the Higher Himalayan Crystallines (Sikkim, Eastern Himalaya) inferred from trace element-constrained monazite and zircon chronology. Contributions to Mineralogy and Petrology 165, 349–372. http://dx.doi.org/10.1007/s00410-012-0812-y.

Sawyer, E.W., 2008. Atlas of migmatites. The Canadian Mineralogist. In: Special Publication, vol. 9. NRC Research Press, Ottawa, Ontario, p. 371.

Schärer, U., 1984. The effect of initial ^{230}Th disequilibrium on young U–Pb ages: the Makalu case, Himalaya. Earth and Planetary Science Letters 67, 191–204.

Searle, M.P., 2010. Low-angle normal faults in the compressional Himalayan orogen; Evidence from the Annapurna–Dhaulagiri Himalaya, Nepal. Geosphere 6, 296–315.

Searle, M.P., 2013. Crustal melting, ductile flow, and deformation in mountain belts: Cause and effect relationships. Lithosphere 5, 547-554, doi:10.1130/RF.L006.1

Searle, M.P., Godin, L., 2003. The South Tibetan Detachment System and the Manaslu leucogranite: a structural reinterpretation and restoration of the Annapurna–Manaslu Himalaya, Nepal. Journal of Geology 111, 505–523.

Searle, M.P., Law, R.D., Godin, L., Larson, K.P., Streule, M.J., Cottle, J.M., Jessup, M.J., 2008. Defining the Himalayan Main Central Thrust in Nepal. Journal of the Geological Society, London 165, 523–534.

Seydoux-Guillaume, A.M., Paquette, J.L., Wiedenbeck, M., Montel, J.M., Heinrich, W., 2002a. Experimental resetting of the U–Th–Pb system in monazite. Chemical Geology 191, 165–181.

Seydoux-Guillaume, A.M., Wirth, R., Nasdala, L., Gottschalk, M., Montel, J.M., Heinrich, W., 2002b. XRD, TEM and Raman study of experimental annealing of natural monazite. Physics and Chemistry of Minerals 29, 240–253.

Seydoux-Guillaume, A.M., Wirth, R., Heinrich, W., Montel, J.M., 2002c. Experimental determination of thorium partitioning between monazite and xenotime using analytical electron microscopy and X-ray diffraction Rietveld analysis. European Journal of Mineralogy 14, 869– 878.

Spear, F.S., Pyle, J.M., 2002. Apatite, monazite, and xenotime in metamorphic rocks. Reviews in Mineralogy and Geochemistry 48, 293–335.

Spear, F.S., Pyle, J.M., 2010. Theoretical modeling of monazite growth in a low-Ca metapelite. Chemical Geology 273, 111–119.

Stephenson, B.J., Searle, M. P., Waters, D.J., Rex, D.C., 2001. Structure of Main Central Thrust zone and extrusion of the High Himalayan deep crustal wedge, Kishwar-Zanskar Himalaya. Journal of Geological Society, London 158, 637-652.

Tamponi, M., Bertoli, M., Innocenti, F., Leoni, L., 2002-2003. X-Ray fluorescence analysis of major elements in silicate rocks using fused glass discs. Atti Società Toscana di Scienze Naturali, Memorie Serie A 108, 73-79.

Tiepolo, M., Bottazzi, P., Palenzona, M., Vannucci, R., 2003. A laser probe coupled with ICP – Double–Focusing Sector-Field mass spectrometer from in situ analysis of geological samples and U–Pb dating of zircon. The Canadian Mineralogist 41, 259–272.

Tomkins, H.S., Powell, R., Ellis, J., 2007. The pressure dependence of the zirconium-in-rutile thermometer. Journal of Metamorphic Geology 25, 703-713.

van Achterbergh, E., Ryan, C.G., Jackson, S.E., Griffin, W., 2001. Data reduction software for LA-ICPMS. In: Sylvester, P. (ed.) Laser ablation ICPMS in the Earth sciences: Principles and Applications. Mineralogical Association of Canada Short Course Series 29, 239–243.

Vance, D., Mahar, E., 1998. Pressure.temperature paths form P-T pseudosections and zoned garnets: potential, pitfalls and examples form Zanskar Himalaya, NW India. Contributions to Mineralogy and Petrology 132, 225-245.

Vannay, J.C., Hodges, K.V., 1996. Tectonometamorphic evolution of the Himalayan metamorphic core between the Annapurna and Dhaulagiri, Central Nepal. Journal of Metamorphic Geology 14, 635–656.

Visonà, D., Carosi, R., Montomoli, C., Peruzzo, L., Tiepolo, M., 2012. Miocene andalusite leucogranite in central-east Himalaya (Everest–Masang Kang area): low-pressure melting during heating. Lithos 144, 194–208.

Weinberg, R., Hasalová, P., 2015. Water-fluxed melting of the continental crust: A review. Lithos 212–215, 158–188.

Weller, O.M., St-Onge, M.R., Waters, D.J., Rayner, N., Searle, M.P., Chung, S.-L., Palin, R.M., Lee, Y.-H., Xu, X., 2013. Quantifying Barrovian metamorphism in the Danba Structural Culmination of eastern Tibet. Journal of Metamorphic Geology 31, 909–935.

White, R.W., Powell, R., Holland, T.J.B., 2001. Calculation of partial melting equilibria in the system Na_2O-CaO-K_2O-FeO-MgO-Al_2O_3-SiO_2-H_2O (NCKFMASH). Journal of Metamorphic Geology 19, 139–153.

Williams, M.L., Jercinovic, M.J., 2002. Microprobe monazite geochronology: putting absolute time into microstructural analysis. Journal of Structural Geology 24, 1013–1028.

Williams, M.L., Jercinovic, M.J., 2012. Tectonic interpretation of metamorphic tectonites: integrating compositional mapping, microstructural analysis and in situ monazite dating. Journal of Metamorphic Geology 30, 739–752.

Yakymchuk, C., Brown, M., 2013. Consequences of open system melting in tectonics. Journal of the Geological Society of London 171, 21–40.

Yin, A., 2006. Cenozoic tectonic evolution of the Himalayan orogen as constrained by along-strike variation of structural geometry, exhumation history, and foreland sedimentation. Earth-Science Reviews 76, 1-131.

Zack, T., Moraes, R., Kronz, A., 2004. Temperature dependence of Zr in rutile: empirical calibration of a rutile thermometer. Contributions to Mineralogy and Petrology 148, 471–488.

Figures and tables captions

List of tables

Table 1: Representative silicate analyses (in wt%) for sample K28a. Mineral structural formulae were recalculated as follows: garnet = 24 O; micas = 11 O; plagioclase = 8 O; ilmenite = 3 O; n.a. = not analyzed.

Table 2: Representative analyses of monazite, recalculated on the basis of 4 O. Monazite textural position is indicated. For the monazite rims (interpreted in equilibrium with matrix xenotime) temperatures (in °C) are obtained with three different calibrations of monazite-xenotime thermometer, as indicate: $T_{(P01)}$ = Pyle et al., 2001; $T_{(G\&H97)}$ = Gratz and Heinrich 1997; $T_{(SG02)}$ = Seydoux-Guillaume et al., 2002c.

Table 3: LA-ICP-MS isotopic results and monazite ages. Textural position of the grain is indicated (abbreviations as in figure 3).

List of figures

Figure 1: a) Himalaya geological map with the location of the study area indicated (after Searle and Godin, 2003). Abbreviations (not explained in the figure) as follows: P = Peshawar basin; K = Kasmir Neogene basin; S = Sultej basin; MFT = Main Frontal Thrust; MBT = Main Boundary Thrust; MCT = Main Central Thrust; GCT = Great Counter Thrust; STDS = South Tibetan Detachment System; ZSZ = Zanskar Shear Zone; MMT = Main Mantle Thrust; MKT = Main Karakoram Thrust; SSZ = Shyok Suture Zone; b) Sketch geological map of the Kali Gandaki valley (modified after Vannay and Hodges, 1996; Carosi et al., 2014a). Numbers in the legend: 1 = Tethyan Sedimentary Sequence (TSS); 2 = orthogneiss (unit 3, GHS); 3 = metapelite and calcsilicate (unit 3, GHS); 4 = calc-silicate and marble (unit 2, GHS); 5 = kyanite-garnet gneiss (unit 1, GHS); 6 = quartzite of LHS in Vannay and Hodges, (1996) or lower GHS in Parsons et al. (2014); 7 = alluvial debris; 8 = STDS; 9 = minor normal fault; 10 = Kalopani shear zone (KSZ); 11 = MCT in Vannay and Hodges, (1996) or CT in Parsons et al. (2014); 12 = main foliation; 13 = object lineation; 14 = location of study samples.

Figure 2: Characteristics of the studied rocks. a) Outcrop view of the leucosome aligned along the main foliation (Sp). Black box is the location of the fig. 2b; b) Details of leucosome concentration around cm-sized (peritectic) garnet and mafic (biotite) selvedges; c) Kyanite and garnet bearing leucosome; d) Melt accumulation in dilational structures (intrafoliation boudin). Note the smooth boundaries between the melt pocket and the main foliation.

Figure 3: a) Panoramic view of garnet-kyanite migmatitic paragneiss (sample K28a, plane-polarized light); b) White mica shear band pointing to a top-to-the SW sense of shear (sample K28a, crossed polars); c) Tiny sillimanite needels growing at plagioglase boundaries and partially replacing kyanite (sample K28g, plane-polarized light); d) Quartz chessboard extinction. Note also the late white mica partially replacing a kyanite grain (sample K28g, crossed polars); e) Quartz string of pearls at a feldspar-feldspar boundary (sample K28a, crossed polars); f) Euhedral plagioclase faces within the leucosome (sample K28g, crossed polars with 530 nm λ plate). Abbreviations: Bt = biotite, Grt= garnet, Ky = kyanite, Pl = plagioclase, Qz = quartz, Sil = sillimanite, Wm = white mica.

Figure 4: X ray compositional maps for white mica (a for Ti, c for Mg) and garnet (b for Mn; d for Mg). Black to red colours indicate increasing element concentrations.

Figure 5: Main silicate chemistry: a) white mica and paragonite; b) biotite; c) garnet; d) plagioclase.

Figure 6: P-T pseudosection for sample K28a. The bulk composition (in wt%) used for the modelling is reported in the upper part of the P-T graph. According to Herron (1988) the sample plots in the shale field. Abbreviations as in figure 3 and Crd = cordierite, Kfs = K-feldspar, Ilm = ilmenite, L = melt, Opx = orthopyroxene, Pg = paragonite, Rt = rutile, St = staurolite.

Figure 7: Compositional and modal isopleths in the pseudosection of Fig. 6. a) XMg (red line) and XCa (green line) in garnet; b) Si atoms per formula units (a.p.f.u.) in potassic white mica (black line) and XAn in plagioclase (dashed black line); c) Si/Al (purple line) and Na/K (blue line) ratios of the silicate melt; d) modal amounts of melt (dashed blue line) and garnet (dashed red line).

Figure 8: Zr-in-rutile temperature (T in °C, using the Tomkins et al., 2007 calibration at 11 kbar reference pressure) *versus* Nb (ppm) content. An uncertainty of ± 30°C has been assigned to the calculated temperature (see discussion in Tomkins et al., 2007).

Fiugre 9: P-T path based on isopleths of Fig. 7 and Zr-in rutile thermometry (see paragraph 6.2). Melt-in curve refers to the first appearence of melt according to the Perple_X calculations.

Figure 10: Textural position, BSE images and Th, Y chemical maps of selected monazite. a) Mnz 1 in rutile included in garnet; b) Mnz 3 and Mnz 4 in garnet; c) Mnz 19 completely shielded in kyanite; note the absence of “high Y” rims; d) Mnz 7 in biotite; note the inclusion of biotite in monazite; e) Mnz 21 in the matrix; note the discontinuous high-Y rim; f) matrix grain Mnz 16. Th-Pb ages are reported and the quoted errors refer to a 2σ confidence level.

Figure 11: Monazite chemical variation of heavy rare earth elements (HREE) *versus* light rare earth elements (LREE). Core-mantle-rim “appellative” refers only to a geometric spot location of EMP analyses.

Figure 12: a) ^{206}Pb-^{238}U/^{208}Pb-^{232}Th concordia diagram; b) Probability density plot for Th-Pb ages. In the box inside, monazite populations (Mnz I, Mnz II, Mnz III) are indicated.

Figure 13: P-T-t-D path of the GHS kyanite-bearing migmatites (block diagrams modified after Vannay and Hodges, 1996) based on pseudosection modelling, Zr-in-rutile temperatures and monazite geochronological data (see text).

Appendix

Table A.1: Isotopic LA-ICP-MS results for the Moacir standard.