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Tethyan and Indian subduction viewed from the Himalayan high- to ultrahigh-pressure metamorphic rocks

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Abstract

The Himalayan range is one of the best documented continent-collisional belts and provides a natural laboratory for studying subduction processes. High-pressure and ultrahigh-pressure rocks with origins in a variety of protoliths occur in various settings: accretionary wedge, oceanic subduction zone, subducted continental margin and continental collisional zone. Ages and locations of these high-pressure and ultrahigh-pressure rocks along the Himalayan belt allow us to evaluate the evolution of this major convergent zone.

(1) Cretaceous (80–100 Ma) blueschists and possibly amphibolites in the Indus Tsangpo Suture zone represent an accretionary wedge developed during the northward subduction of the Tethys Ocean beneath the Asian margin. Their exhumation occurred during the subduction of the Tethys prior to the collision between the Indian and Asian continents.

(2) Eclogitic rocks with unknown age are reported at one location in the Indus Tsangpo Suture zone, east of the Nanga Parbat syntaxis. They may represent subducted Tethyan oceanic lithosphere.

(3) Ultrahigh-pressure rocks on both sides of the western syntaxis (Kaghan and Tso Morari massifs) formed during the early stage of subduction/exhumation of the Indian northern margin at the time of the Paleocene–Eocene boundary.

(4) Granulitized eclogites in the Lesser Himalaya Sequence in southern Tibet formed during the Paleogene underthrusting of the Indian margin beneath southern Tibet, and were exhumed in the Miocene.

These metamorphic rocks provide important constraints on the geometry and evolution of the India–Asia convergent zone during the closure of the Tethys Ocean. The timing of the ultrahigh-pressure metamorphism in the Tso Morari massif indicates that the initial contact between the Indian and Asian continents likely occurred in the western syntaxis at 57±1 Ma. West of the western syntaxis, the Higher Himalayan Crystallines were thinned. Rocks equivalent to the Lesser Himalayan Sequence are present north of the Main Central Thrust. Moreover, the pressure metamorphism in the Kaghan massif in the western part of the syntaxis took place later, 7 m.y. after the metamorphism in the eastern part, suggesting that the geometry of the initial contact between the Indian and Asian continents was not linear. The northern edge of the Indian continent in the western part was 300 to 350 km farther south than the area east of the Nanga Parbat syntaxis. Such “en baionnette” geometry is probably produced by north-trending transform faults that initially formed during the Late Paleozoic to Cretaceous Gondwana rifting. Farther east in the southern Tibet, the collision occurred before 50.6±0.2 Ma. Finally, high-pressure to ultrahigh-pressure rocks in the western Himalaya formed and exhumed in steep subduction compared to what is now shown in tomographic images and seismologic data.

Keywords: Himalaya; HP-UHP metamorphic rocks; Subduction; India–Asia collision

1. Introduction

Since the first petrological description of eclogites by Haüy (1822), eclogites have been reported from many locations with

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ages ranging from Proterozoic to Phanerozic (e.g. Godard, 2001 for review). Such high-pressure (HP) low-temperature (LT) metamorphic rocks (blueschist and eclogite facies rocks) in orogenic belts provide valuable information related to subduction processes of oceanic lithosphere (Coleman, 1971; Ernst, 1973). The occurrences of pelitic rocks and continental rocks metamorphosed under eclogite facies conditions suggest that they are subducted and later exhumed (Compagnoni, 1977; Carswell, 1990). The discovery of coesite in pyrope-bearing quartzites in the Alps (Chopin, 1984) introduced the term ultrahigh-pressure (UHP) metamorphism and demonstrated that continental rocks can be subducted at a depth greater than 100–120 km. Consequently, characterization of HP to UHP rocks is important for better understanding of paleogeodynamic evolution of convergent zones. This paper focuses on Himalayan eclogites and blueschists derived from the Tethyan oceanic crust and the margin of the Indian continent metamorphosed under eclogite facies conditions, described as group B and C, respectively, by Coleman et al. (1965). Their studies, in terms of origin, metamorphic evolution, and timing of exhumation, give us important constraints on the dynamics of the India–Asia convergence.

Argand (1924) and later Gansser (1964) described the Himalayan range as a continent–continent-collision belt. By the 1970's, researchers demonstrated that the Himalayan collision resulted from the closure of the Tethyan Ocean through the subduction of the oceanic lithosphere (Desio, 1977; Sengör, 1979). Although, HP rocks have been reported in the Alpine belt as early as 1822 by Hauÿ, subduction related metamorphic rocks in the Himalaya only started to be recognized in the early 20th Century. Hayden (1904) and Berthelsen (1953) described garnet-bearing mafic rocks in the Tso Morari gneiss in eastern Ladakh, India. Because the plate tectonic theory had not been widely accepted, the significance of their work was not recognized until the 1980's. The first modern descriptions of blueschists in the Himalaya were carried out by Shams (1972) and Desio (1977) in Pakistan along the Main Mantle Thrust (MMT) and by Franck et al. (1977) and Virdi et al. (1977) in Ladakh India along the Indus Tsangpo Suture zone (ITSZ). Eclogites were first reported by Chaudhry and Ghazanfar (1987) in the Kaghan valley, and UHP coesite-bearing eclogites were first described in 1998 in the same area (O'Brien et al., 1999). All these rocks have been observed in or near the India–Asia suture zone, Main Mantle Thrust (MMT) in Pakistan or Indus Tsangpo Suture zone (ITSZ) in India and China. In addition, Lombardo and Rolfo (2000) reported the occurrence of retrogressed eclogites in southern Tibet, about 200 km south of the ITSZ. The discovery raised the question of the distribution and preservation of HP and UHP rocks. Relatively rare findings of HP to UHP rocks in the Himalaya compared to the Alpine belt are not only related to the difficulty in access but also the intense activity of collision and erosion that probably obliterated such rocks.

This paper reviews the occurrences of HP to UHP along the Himalaya belt, and discusses, in the light of their metamorphic evolution, the initial history and geometry of the India–Asia convergent zone prior to and during the collision.

2. High-pressure and ultrahigh-pressure rocks in the Himalaya

The tectonostratigraphy of the internal part of the Himalaya is briefly reviewed here in order to understand the significance of eclogitic rocks formed from the Indian plate. From the Nanga Parbat spur in the west and Namche Barwa spur in the east, the 2400 km long Himalayan belt is classically divided into juxtaposed zones (Fig. 1). The westward extension of the Main Central Thrust (MCT) and the Higher (or Greater) Himalayan Crystallines (HHC) in Pakistan remain controversial (e.g., Di Pietro and Pogue, 2004), thus a major boundary east of the Nanga Parbat syntaxis separates the central-east Himalaya from the western Himalaya (Fig. 1).

South of the ITSZ, the central Himalaya is divided into five main zones: the Sub-Himalaya, the Lesser Himalayan Sequence (LHS) between the Main Boundary Thrust (MBT) and the MCT, the HHC between the MCT and the South Tibetan Detachment (STD), the Tethys Himalaya and the North Himalayan massifs (Yin et al., 1999; Hodges, 2000; DeCelles et al., 2000).

The LHS is mostly a thick (>10 km) succession of Early to middle Proterozoic low grade metasedimentary rocks locally crosscut by mafic and felsic intrusion. Augen gneisses in the metasedimentary rocks show U/Pb zircon ages of ca. 1850 Ma (e.g. Le Fort, 1989). Detrital zircon ages in the LHS range between 2.6 and 1.8 Ga, suggesting that they derived from Late Archean to Early Proterozoic rocks in the Indian continent (DeCelles et al., 2000; Richards et al., 2005; Robinson et al., 2006). Permian and Paleocene sedimentary rocks of the Gondwana sequence and foreland basin sequence overlie the LHS (Fig. 3a) (Sakai, 1989). The HHC thrust southward above the underlying LHS along the MCT. The HHC consists of metasedimentary rocks dominantly younger than 800 Ma. Detrital zircons from these metasedimentary rocks yield ages of 800 to 1700 Ma and the augen gneisses generally located at the top of the HHC give ages ranging between 500 and 480 Ma (Le Fort et al., 1986; Parrish and Hodges, 1996; Robinson et al., 2001; 2006; Richards et al., 2005). The HHC experienced Barrovian metamorphism and igneous intrusion during the Neoproterozoic to Cambrian Pan-African event, along the tectonically active northern margin of Gondwana; the MCT could correspond to an Early Paleozoic suture zone (Manickavasagam et al., 1999; DeCelles et al., 2000; Marquer et al., 2000).

North of the HHC, the Tethys Himalaya is separated from the HHC by the STD but the two were originally in stratigraphical continuity (e.g. Colchen et al., 1986; Garzanti et al., 1986). The marine sedimentary rocks of the Tethys Himalaya deposited on the Indian continental margin from the Ordovician to the Paleocene (Garzanti et al., 1986). The Tethys Himalaya sedimentary rocks are also characterized by abundant mafic lenses intercalated in the sedimentary sequence, corresponding to transposed dykes and sills. These mafic rocks are related to Carboniferous and Permian rifting (Garzanti, 1999) and are lateral equivalents of the Panjal Trap, which is abundant in Pakistan and Kashmir (e.g. Papritz and Rey, 1989).

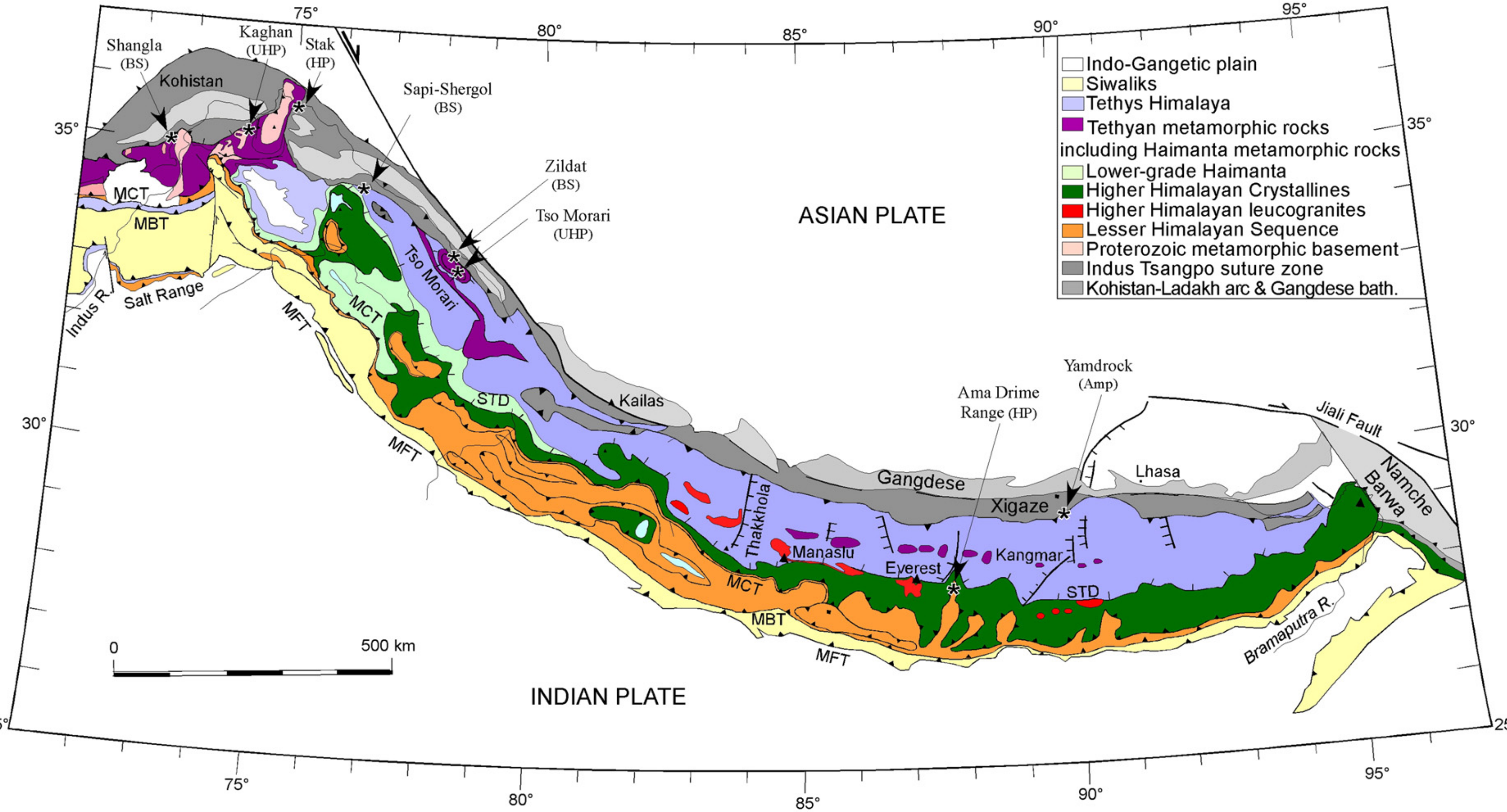

Fig. 1. Geological map of Himalaya (after Ding et al., 2001; Guillot et al., 2003; Di Pietro and Pogue, 2004). The western Himalaya is clearly distinct to the central-east Himalaya. The stars show the locations of the metamorphic rocks related to subduction processes discussed in the text. BS: Blueschist unit, Amp: amphibolitic unit; HP: high-pressure unit, UHP: ultrahigh-pressure unit.

The Tethys Himalaya is underlain by the North Himalayan domal massifs (Fig. 1). These domes consist of slightly metamorphosed Neoproterozoic to Cambrian sedimentary rocks (the Haimanta Formation) and/or 500 Ma old augen gneisses (Kangmar, Gurla Mandhata) and were intruded by Miocene-age two-mica granites in Southern Tibet (Mabja, Lhagoi Kangri, Kudai), or consist of UHP continental rocks in the western Himalaya (Tso Morari), cored by 500 Ma old augen gneisses (Fig. 1) (Burg et al., 1984; Brookfield, 1993; Steck et al., 1993; Guillot et al., 1997; Steck et al., 1998; de Sigoyer et al., 2004). The Permo-Carboniferous sedimentary rocks were deposited during the main rifting event onto the Haimanta Formation or the 500 Ma old augen gneisses (Brookfield, 1993; Steck et al., 1993; Di Pietro and Isachsen, 2001). The northern Himalayan massifs have a distinct stratigraphic sequence and metamorphic evolution, which is different from the HHC and likely, represents the distal part of the Indian continental margin (Guillot et al., 1997; Yin et al., 1999; Hodges, 2000).

In this general framework, HP to UHP rocks outcrop in the western Himalaya, within or just south of the ITSZ at the top of the nappe pile (Fig. 2). The exception is the Ama Drime Range eclogites in Arun valley, eastern Nepal, that crop out at the base of the HHC in Southern Tibet, just below the MCT (Fig. 2c) (Lombardo and Rolfo, 2000). This occurrence is described later.

3. High-pressure rocks from the Indus Tsangpo Suture zone

Blueschists facies rocks in the Himalaya have been reported only in the NW Himalaya, along the MMT and the ITSZ (Shams, 1972; Franck et al., 1977). In Pakistan, the Shangla blueschists outcrop in the Lower Swat district within the MMT between the Cretaceous Kohistan arc and the northern margin of the Indian plate (Fig. 1). Blueschists which originated from tholeiitic basalts occur as tectonic slices together with ophiolitic and metasedimentary rocks (Fig. 3a). Metabasites underwent metamorphism under blueschist facies conditions (0.7 ± 0.05 GPa and 400 ± 20 °C) followed by a minor retrogression under greenschist facies conditions ($P < 0.4$ GPa, $T < 400$ °C) (Fig. 4) (Guiraud, 1982; Jan, 1985). Rb–Sr, K–Ar and Ar–Ar dating of glaucophane and phengite suggests the blueschist facies metamorphism at ca. 80 Ma (Shams, 1980; Maluski and Matte, 1984; Anczkiewicz et al., 2000). These HP-LT metamorphic rocks originated from oceanic crust and associated sedimentary rocks. The absence of ages younger than 80 Ma in these rocks suggests that they were not affected by metamorphic imprint during the Himalayan collision and likely exhumed during oceanic subduction, within an accretionary wedge (Anczkiewicz et al., 2000).

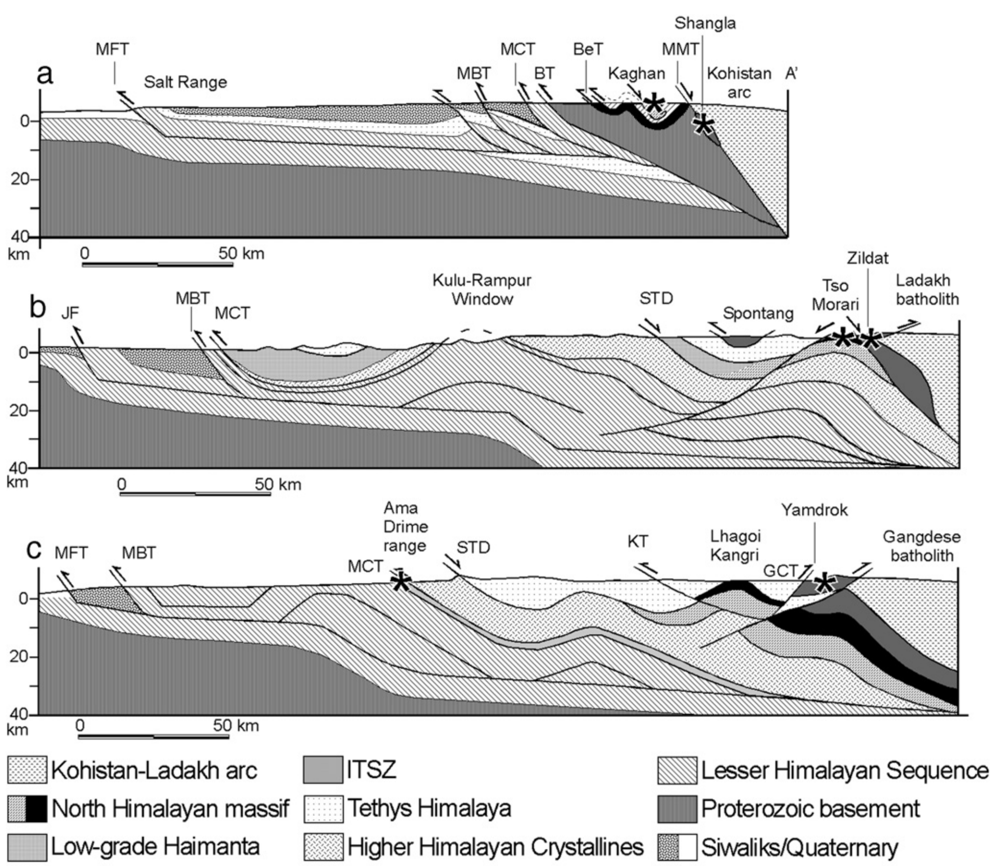

Fig. 2. Cross-sections throughout the Himalayan belt. (Modified after Di Pietro and Pogue, 2004). a) In Pakistan Himalaya showing the location of the Kaghan UHP unit at the top of the nappe pile along the ITSZ. b) In India Himalaya showing the location of the Tso Morari UHP unit at the top of the nappe pile along the ITSZ. c) In southern Tibet showing the location of the Ama Drime Range HP unit in the LHS, just below the MCT.

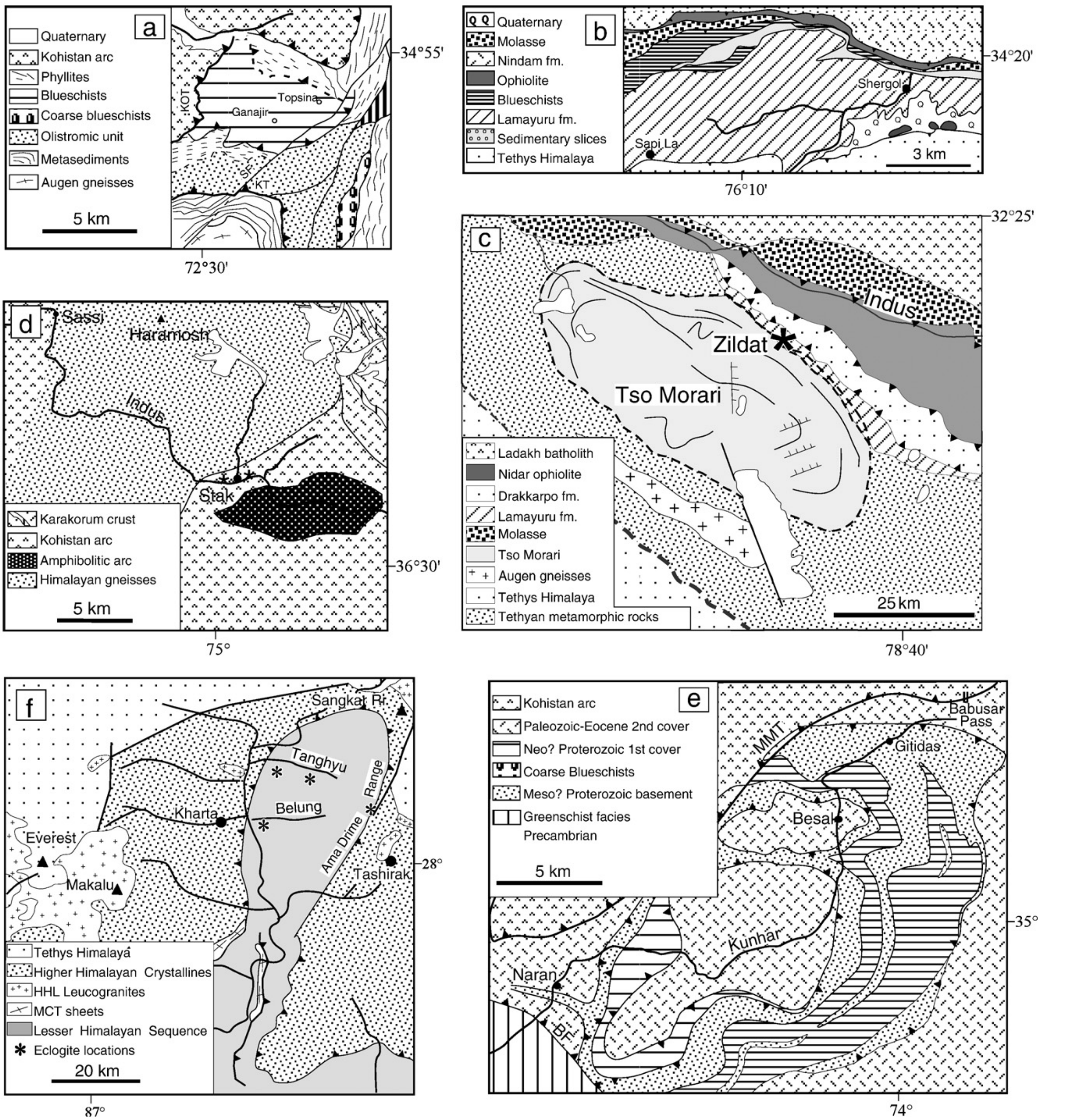

Fig. 3. Local geological maps of the of blueschists and eclogites' occurrences along the Himalayan belt. a) Geological map of the Shangla area in the Lower Swat district of Pakistan (modified after Anczkiewicz et al., 2000). The blueschist unit made of metavolcanites and metaschists formed a mélange zone of 5 * 5 km. KT: Kishora Thrust; KOT: Kohitsan Thrust; SF: Shinkad Fault. b) Geological map of the Sapi-Shergol zone mélange zone (modified after Honegger et al., 1989). The blueschist unit formed an elongated vertical zone made off lenses of serpentinites and metabasites interlayered with schists. c) Geological map of the Tso Morari area (modified after Guillot et al., 1997). The Zildat blueschists (*) formed a small decametric lense squeezed between the UHP unit and the ITSZ. The UHP Tso Morari unit formed a large area (100 * 50 km), south of the ITSZ, throughout the metamorphic Tethys Himalaya. d) Geological map of the Stak area (modified after Le Fort et al., 1997). The Stak eclogites outcrop along the Indus valley at the eastern rim of the Nanga Parbat syntaxis. In this area, the Ladakh arc is mainly composed of metabasites with lenses of serpentinised peridotites and boudins of retrogressed eclogites. e) Geological map of the Upper Kaghan valley (modified after Chaudhry and Ghazanfar, 1987). Coesite-bearing eclogites are common in the upper Kaghan valley (O'Brien et al., 2001; Kaneko et al., 2003; Treloar et al., 2003) but they are mainly found in the lowest unit, the Proterozoic basement as basic lenses within the gneisses. BF: Batal Thrust. MMT: Main Mantle Thrust (e.g. ITSZ). f) Geological map of the Arun valley (modified after Groppo et al., 2007). The retrogressed eclogites have been reported in the Ama Drime Range, below the MCT and belonging to the Lesser Himalayan Sequence.

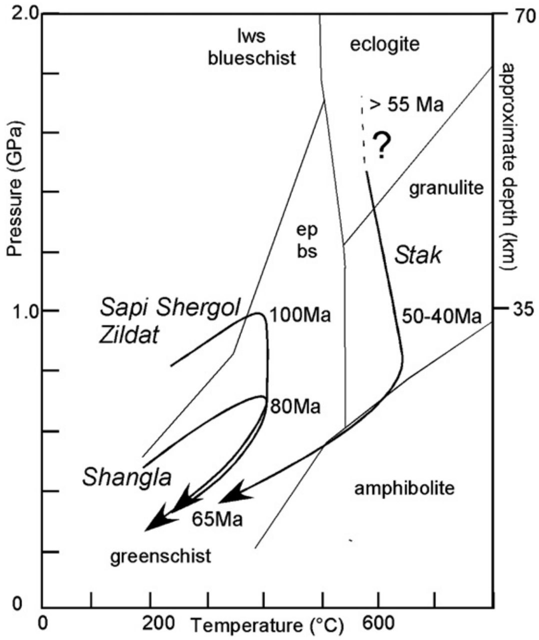

Fig. 4. Pressure–temperature–time paths of the blueschists units (Sangla, Sapi-Shergol and Zildat) and of the Stak eclogites (see text for references). Notice that the blueschist units recorded a cooled retrogressed path typical of syn-subduction exhumation while the Stak eclogites retrogressed under granulite facies metamorphic conditions, typical of syn-collision exhumation. Lws: lawsonite, ep: epidote, bs: blueschist.

Another occurrence of blueschists is in Ladakh, India, in the Sapi-Shergol mélange zone, south of Kargil (Fig. 1). They show peak metamorphism of 0.9–1.0 GPa and 350–420 °C (Fig. 4) (e.g. Mahéo et al., 2006). As in Shangla area, hectometric slices of metabasites and metasedimentary rocks are intercalated with ophiolitic rocks (Fig. 3b), defining the ITSZ (Honegger et al., 1982; 1989). K–Ar ages of whole-rocks and glaucophane provide a HP metamorphism at ~100 Ma (Honegger et al., 1989). Sapi-Shergol mélange is generally considered to consist of blocks of igneous rocks derived from an oceanic island and an intra-oceanic arc. This mélange also contain blocks of the Dras continental arc (Honegger et al., 1989; Mahéo et al., 2006). Mahéo et al. (2006) proposed that the Sapi-Shergol mélange incorporated part of the Ladakh arc during its early accretion to the Asian margin at the Cretaceous–Paleocene boundary (Fig. 5).

Blueschists were reported in eastern Ladakh in the Zildat valley by Virdi et al. (1977) and confirmed by de Sigoyer et al. (2004). They form a decametric lense squeezed between the UHP Tso Morari rocks to the south and the Nidar ophiolite to the north (Fig. 3c). These rocks show the same mineral assemblage as the Sapi-Shergol blueschists "and the two are probably lateral equivalents" (Jan, 1985). Finally, blueschists were also reported in southern Tibet, 60 km west of Xigaze (Xiao and Gao, 1984), but this occurrence still needs to be verified as no glaucophane schists are observed (Burg et al., 1987). Although the subduction of the Tethyan Ocean beneath the Asian continent was continuous from 120 Ma to 55 Ma, the metamorphic ages of blueschists are limited to a period ranging from 100 to 80 Ma. This suggests that exhumation of HP rocks took place only during a short period before the India–Asia collision (Fig. 5). Agard et al. (2006) in their review of HP oceanic rocks of the world pointed out that the exhumation of HP rocks is not continuous as subduction. They proposed that limited time periods for exhumation of HP rocks are explained by modifications of plate convergence, such as the start of subduction of buoyant material, a change in the convergence velocity, and the cessation of oceanic subduction. The eclogites reported in the ITSZ occur only in the Stak valley between the eastern contact of the Nanga Parbat syntaxis and the Ladakh island arc (Figs. 1 and 3d) (Le Fort et al., 1997). Eclogite facies metamorphism is recognized as the occurrence of relict minerals in garnet-bearing amphibolites that form metric to decametric lenses in serpentinites. The metabasites show the mineralogy common in retrogressed eclogites including garnet, symplectitic association of clinopyroxene and plagioclase, rutile and quartz overprinted by an amphibolite facies mineral assemblage of edenitic hornblende, secondary plagioclase and ilmenite. Le Fort et al. (1997) estimated the eclogite facies conditions, $>1.3\pm0.1$ GPa and $>610\pm30$ °C, followed by a retrogression under amphibolite facies conditions at 0.8 ± 0.1 GPa and 630 ± 30 °C (Fig. 4). The protolith and the metamorphic age remain uncertain, but its distribution in the ITSZ, and the association of metabasites and serpentinites suggest that the unit corresponds to oceanic or island arc rocks deeply subducted below the Ladakh arc (Fig. 5). A similar setting is well documented in the Piedmont zone in the western Alps, where successive tectono-metamorphic slices are exhumed in the paleo-oceanic

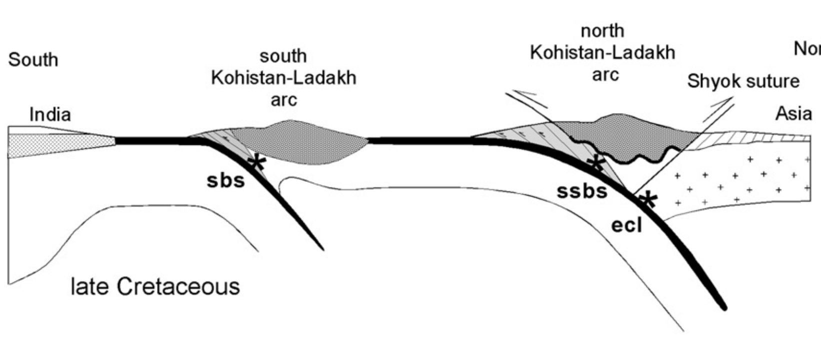

Fig. 5. Geodynamical model for the closure of the Tethys in western Himalaya (modified after Anczkiewicz et al., 2000) The closure of the Tethys Ocean was responsible for the development of double subduction zones with two accretionary wedges and the subduction of the oceanic lithosphere leading to the formation and exhumation of blueschists and eclogites. sbs: Shangla blueschists, ssbs: Sapi-Shergol blueschists, ecl.: Stak eclogites.

subduction zone (Schwartz et al., 2001; Agard et al., 2002; Guillot et al., 2004).

4. UHP metamorphism in the western Himalaya

4.1. The Kaghan massif

The eclogite–facies metamorphic rocks in the Kaghan and Neelum valleys in the Naran area, northern Pakistan (Fig. 1) (Pognante and Spencer, 1991; Fontan et al., 2000; O'Brien et al., 2001, Treloar et al., 2003) lie in the most northern margin of the Indian Plate on the immediate footwall of the Main Mantle Thrust (MMT) (Figs. 2a and 3e). In the Upper Kaghan valley, eclogite–facies rocks occur in two metasedimentary sequences and the underlying gneissic unit (Chaudhry and Ghafanzar, 1987; Greco et al., 1989; Greco and Spencer, 1993; Spencer et al., 1995). The north-facing detachment fault between the gneissic unit and the overlying sedimentary rocks may be an extension of the Besal Thrust. This fault is considered to have developed during the deformation under amphibolite facies conditions (Treloar et al., 2003), but it may have been started during UHP metamorphism. Reactivation of the MMT as a normal fault during Eocene through Oligocene time has been proposed by many researchers (Hubbard et al., 1995; Burg et al., 1996; Vince and Treloar, 1996; Argles and Edwards, 2002; Burg et al., 2006). In the southern part of the Kaghan massif, the Batal fault (Fig. 3e) is a major thrust, possibly associated with the exhumation of eclogites. The western extension of this fault is not known and it could merge with the ITSZ (Di Pietro and Pogue, 2004).

The protolith of the Kaghan massif is estimated from associated rocks. The lowest exposed gneissic unit consists of intensely deformed biotite-bearing gneisses and granulite–facies rocks that correlate with the Nanga Parbat gneisses (Treloar et al., 2003; Fig. 1). It may also correspond to LHS because similar granitic gneisses developed in the LHS during the 1.8 Ga event (Di Pietro and Pogue, 2004). The overlying gneissic unit is dominantly composed of biotite-bearing metagreywacke intercalated with garnet and/or kyanite bearing metapelites and strongly sheared granite sheets. They are considered to be Neoproterozoic (Treloar et al., 2003) or Lower Paleozoic (Spencer et al., 1995). The second sequence is metasedimentary rocks containing amphibolites, marbles and garnet±kyanite±staurolite-bearing calcareous metapelites of probable Upper Paleozoic to Early Mesozoic age and the amphibolites are metamorphosed Panjal Trap volcanic rocks (Greco et al., 1989; Papritz and Rey, 1989; Spencer et al., 1995) based on Permian U–Pb zircon ages (Spencer and Gebauer, 1996).

Coesite-bearing eclogites are common in the upper Kaghan valley (O'Brien et al., 2001; Kaneko et al., 2003; Treloar et al., 2003). The peak metamorphic condition for coesite-bearing unit is estimated at 3.0±0.2 GPa and 770±50 °C (O'Brien et al., 2001) and coesite-free unit at 610±30 °C at 2.4±0.2 GPa (Lombardo et al., 2000). The eclogites are retrogressed to form amphiboles, magnesio-hornblende, barroisite or glaucophane under blueschist and amphibolite facies conditions (Lombardo and Rolfo, 2000; O'Brien et al., 2001) and later overprinted by local greenschist facies conditions (Fig. 6).

The metamorphic evolution of the Kaghan eclogites is well dated. Tonarini et al. (1993) and Spencer and Gebauer (1996) first estimated the metamorphic ages ranging from 50 to 40 Ma using Sm–Nd and U–Pb isotope systems. Prograde metamorphism for the quartz-bearing eclogite–facies is dated at 50±1 Ma in the upper Kaghan valley by Kaneko et al. (2003). UHP conditions (3.0±0.2 GPa) for the same rocks are dated at 46.2±0.7 Ma and 46.4±0.1 using U–Pb zircon techniques by Kaneko

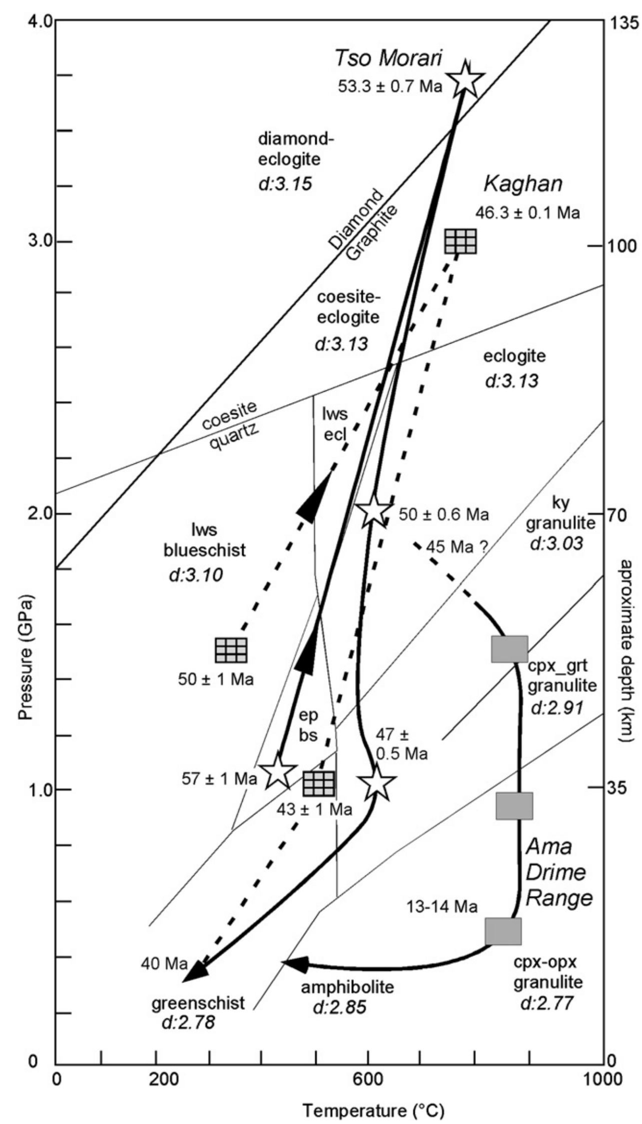

Fig. 6. Pressure–temperature–time paths of the UHP Tso Morari and Kaghan units and HP Ama Drime Range. See text for references. cpx: clinopyroxene, ecl: eclogite; grt: garnet; ky: kyanite; lws: lawsonite; opx: orthopyroxene; d: density after Bousquet et al. (1997). The Kaghan unit records roughly similar *P*–*T* conditions than the Tso Morari unit but it is 7 Ma younger during the prograde and retrograde paths. The Ama Drime Range HP unit records the lowest peak metamorphic conditions suggesting that the dip of the Indian plate was shallow in the southern Tibet during Eocene, as present-day observed. The age of peak metamorphic conditions is unknown in the Ama Drime Range, a 45 Ma old age is inferred from datings obtained in the Dudh Kosi valley (Catlos et al., 2002).

et al. (2003) and Parrish et al. (2006). Zircon and rutile from a coesite-free eclogite yielded ~44 Ma as the age of HP retrograde metamorphism (Spencer and Gebauer, 1996; Treloar et al., 2003; Parrish et al., 2006), implying a fast exhumation rate of ca. 3 to 8 cm/yr. The Kaghan unit cooled below 500–400 °C between 43 Ma and 40 Ma based on a Rb–Sr age of phengite (Tonarini et al., 1993), suggesting a slowed exhumation, ca. 0.3 cm/yr. This is supported by $^{40}Ar/^{39}Ar$ hornblende ages from surrounding amphibolitic rocks; 42.6±1.6 Ma (Chamberlain et al., 1991), 41±2 Ma (Smith et al., 1994) and 39.8±1.6 Ma (Hubbard et al., 1995) (Fig. 6).

4.2. The Tso Morari massif

The Tso Morari massif contains the only UHP rocks in North Himalayan massifs and outcrops south of the ITSZ in eastern Ladakh (Thakur, 1983; Fig. 1). It has an elongated shape of 100×50 km striking northwest and dipping 10° to the NW with a thickness of less than 7 km (Figs. 2b and 3c) (de Sigoyer et al., 2004). It consists of 500 Ma augen gneisses and overlying metasedimentary rocks, both are cut by metabasic dykes partly transposed by successive deformation events (de Sigoyer et al., 2004). The earliest phase of deformation (D1) under eclogitic conditions is characterized by steep, tight to isoclinal folds (F1) with a sub-vertical axial plane cleavage (S1). The second deformation event (D2) developed under blueschist then amphibolite facies conditions during the exhumation of UHP metamorphic rocks. The associated S2 foliation is horizontal in the central part of the dome and becomes steeper on its borders. Although D3 deformation geometrically superimposed the previous deformations, this normal shearing (Fig. 2b) probably developed at the beginning of the exhumation in order to accommodate the extrusion of the UHP metamorphic rocks relative to the surrounding rocks within the subduction channel (de Sigoyer et al., 2004). The existence of a subduction channel is supported by the occurrence of hectometric lenses of serpentinites on the footwall of the Tso Morari unit, the Zildat normal shear zone (Guillot et al., 2001). The southern limb of the Tso Morari massif is affected by a wide south dipping normal shear zone, the Karzog shear zone. This zone separates the Tso Morari massif from the less metamorphosed Mata-Karzog unit to the south.

Minor metabasic rocks intruded the Cambro-Ordovician orthogneiss and are associated with an Ordovician magmatic event (Girard and Bussy, 1999). Towards the rim of the dome, this orthogneiss is overlain by Upper Carboniferous to Permian metasedimentary rocks of the Tethys Himalaya (Stutz and Steck, 1986; Colchen et al., 1994; Fuchs and Linner, 1996). Metric to decametric lenses of metabasic rocks are correlated with Permian dolomitic limestones and show a continental tholeiitic geochemical signature similar to that of the Permian Panjal traps (de Sigoyer et al., 2004). Moreover, Ahmad et al. (2006) suggested that the coesite-bearing eclogites originated from mafic dykes of the Tethyan oceanic plate with transitional affinities, traversing through the northern margin of the Indian plate. These data suggest that the Tso Morari massif represents a remnant of the distal margin of the Indian continent.

The petrological and thermobarometrical studies of the Tso Morari rocks appear in several papers (Guillot et al., 1995; Guillot et al., 1997; de Sigoyer et al., 1997; O'Brien and Sachan, 2000; O'Brien et al., 2001; Sachan et al., 2001; Mukheerjee et al., 2003; Guillot et al., 2003; de Sigoyer et al., 2004). UHP paragenese is observed both in mafic and metapelitic rocks. This is further supported by the occurrence of carbonate-bearing coesite eclogite as lenses in kyanite/sillimanite-grade rocks in the northern part of the dome suggesting pressure of >3.9 GPa and temperature of >750 °C (Mukheerjee et al., 2003) (Fig. 6). Retrogression is characterized by the development of chlorite and chloritoid, which surround the garnet at a minimum pressure of 0.5 GPa and temperature of <500 °C (ibid). Mukheerjee et al. (2005) presented the first occurrence of microdiamond inclusions in zircon from coesite-bearing eclogites, which suggests a pressure greater than 3.5 GPa at 750 °C (Bundy, 1980). The data suggest the subduction down to a depth greater than 120 km. During their exhumation up to the depth of 40–30 km, the Tso Morari rocks underwent cooling under blueschist facies conditions (1.1±0.3 GPa; 580±50 °C) (Fig. 6). In the southwestern part of the unit, calcic amphibole (and not glaucophane) crystallized between omphacite and garnet in the metabasic rocks. In the metagreywacke, staurolite phengite and chlorite from the S2 foliation are destabilized to form kyanite and biotite, implying heating of rocks to 630±50 °C at the depth of 30 km (0.9±0.3 GPa) during the late exhumation (de Sigoyer et al., 1997; Guillot et al., 1997). The crystallization of chlorite and white mica are common in C3 shear bands in the entire Tso Morari massif. This marks the final deformation related to exhumation under the greenschist facies conditions.

Peak metamorphism of the Tso Morari massif was dated at 55±6 Ma using Lu–Hf, Sm–Nd and U–Pb isotope methods (de Sigoyer et al., 2000). Leech et al. (2005) obtained U–Pb SHRIMP zircon ages of 53.3±0.7 Ma and 50.0±0.6 Ma on quartzo-feldspathic gneiss from the Puga Formation and interpreted the former age as the age of UHP metamorphism, and the latter as the retrograde eclogite–facies metamorphism (Fig. 5). The UHP metamorphic age is very similar to a $^{40}Ar/^{39}Ar$ phengite age of 53.8±0.2 Ma obtained by Schlup et al. (2003). Close ages of peak metamorphism and cooling suggest a very fast exhumation. Indeed, the exhumation was very fast, ca. 1.7 cm/yr, compatible with modeling of garnet diffusion profile by O'Brien and Sachan (2000). The timing of amphibolite facies conditions (1.1±0.2 GPa and 630±50 °C; ibid.) is dated at 47±0.5 Ma on $^{40}Ar/^{39}Ar$ ages of phengite, Sm–Nd and Rb–Sr ages of amphiboles and U–Pb zircon ages (de Sigoyer et al., 2000; Leech et al., 2005). The final retrograde metamorphism under greenschist facies conditions is dated between 34±2 Ma and 45±2 Ma on 7 fission track analyses of zircon (Schlup et al., 2003). It corresponds to a decrease of the exhumation rate at ca. 0.3 cm/yr (Fig. 6).

5. HP metamorphism in southern Tibet

Granulitized eclogites were found in the Ama Drime Range (Arun valley) of southern Tibet (Figs. 1 and 3f), east of the Everest–Makalu massif (Lombardo and Rolfo, 2000; Groppo et al., 2007). The eclogites occur below the MCT in the western limb of the Arun north–south antiform (Fig. 2c). In this area, as observed elsewhere in southern Tibet (Pêcher, 1989; Guillot et al., 1999), the LHS consists of granitic orthogneisses and metasedimentary rocks containing garnet ± sillimanite ± cordierite. Farther north, slivers of quartzites, mylonitic marble, biotite–sillimanite schists and orthogneisses hosted lenses of relict eclogite, deriving from discordant dykes in the granitic orthogneiss. The protolith of eclogites is considered to range from 110 to 88 Ma based on SHRIMP zircon ages (Rolfo et al., 2005). Such an age rules out the possibility that the age of eclogitization is Paleozoic or Precambrian, and supports a Tertiary age for eclogitization (Lombardo and Rolfo, 2000). The ages correspond to the Cretaceous basalt volcanism in the LHS and the Tethys Himalaya (Sakai, 1983; Garzanti, 1993). Earlier eclogitization is indicated by garnet composition, occurrence of ilmenite replacing rutile (Groppo et al., 2007), and supported by the occurrence of albitic plagioclase + diopside ± orthopyroxene, which may be retrogressed omphacitic clinopyroxene (e.g., Joanny et al., 1991). Groppo et al. (2007) estimated that the symplectite after omphacite began to form at pressure <1.8 GPa. In addition, recalculated peak metamorphism pyroxene contains high jadeite components between 35 and 45 mol%, which are common in other HP rocks (Lombardo and Rolfo, 2000). The peak metamorphic condition is estimated to be pressures exceeding 1.5 GPa and probably reaching 2.0 GPa for a minimum temperature of 580 °C (Groppo et al., 2007) (Fig. 6). Subsequent granulite facies metamorphism took place at about 1.0 GPa and 750 °C and the retrogression under amphibolite facies conditions, at about 750 °C and 0.7 to 0.5 GPa (Groppo et al., 2007).

Lombardo and Rolfo (2000) considered that the eclogitic metamorphism was before 25 Ma, the Eohimalayan metamorphic event, based on $^{40}Ar/^{39}Ar$ ages of amphibole, 40 km east of the Ama Drime Range (Hodges et al., 1994). Rolfo et al. (2005) obtained SHRIMP U–Pb ages of 13–14 Ma from zircon grains and interpreted as the ages of post-granulite fluid circulation during thrusting along the MCT.

Another example of HP rocks was described in the HHC in the Namche Barwa spur (Fig. 1). Liu and Zhong (1997) described felsic granulite containing an assemblage of garnet + kyanite + feldspar + rutile + quartz and Zhong and Ding (1996) described mafic granulite with an assemblage of garnet + clinopyroxene + rutile + quartz + orthophyroxene + plagioclase. Zhong and Ding (1996) inferred P–T conditions of 1.7–1.8 GPa at 890 °C, but the occurrence of orthopyroxene in the mafic assemblage indicates that the pressure was lower than 1.0 GPa. Burg et al. (1998) re-evaluated the P–T conditions to be 0.8–0.9 GPa and 720–760 °C which correspond to typical granulite facies conditions. The peak metamorphic age of ca. 40 Ma on zircons (Ding et al., 2001) suggest that the P–T–t path is similar to the early Himalayan collisional event recorded along the HHC (Pêcher, 1989; Guillot et al., 1999).

6. Discussion

Despite the subduction of more than 5000 km of Tethys Ocean and Greater India, less than 10 occurrences of subduction related metamorphic rocks have been found along the 2400 km long Himalayan belt so far. Such paucity of HP to UHP rocks can be either explained by (1) the inability to produce such rocks, (2) inability to exhume such rocks and/or (3) the post-subduction processes which remove, alter or conceal HP to UHP rocks. These possibilities are discussed in the following sections: (1) HP rocks in the ITSZ related to the subduction of the Tethyan domain; and (2) eclogites enclosed in gneiss related to the subduction of the Indian plate. Therefore, two kinds of HP to UHP rocks are discussed separately.

6.1. Origin of the HP rocks in the ITSZ

Blueschists and retrogressed eclogites are only observed in the ITSZ, on both sides of the western syntaxis. The Cretaceous ages of the blueschists and the absence of young ages after 80 Ma clearly suggest that these rocks are produced and exhumed discontinuously in an accretionary wedge south of the Ladakh arc during the subduction of the Tethyan oceanic lithosphere (Fig. 5). The tectonic setting is similar to that of the Mariana forearc where fragments of blueschist were found in serpentinite mélange (Fryer et al., 1999). They represent an oceanic subduction similar to the coast of California, the western Alps, the Antilles, and Zagros (Amato et al., 1999; Schwartz et al., 2001; Guillot et al., 2004; Agard et al., 2006). Further work is necessary to determine the protolith, and the ages of their metamorphism and exhumation.

Blueschists are yet to be found in the central part of the ITSZ. However, south of the Xigaze ophiolitic sequence, Aitchison et al. (2000) and Dupuis et al. (2005) described a Cretaceous accretionary complex, the Yamdrock mélange, south of an intra-oceanic island arc (Fig. 1). The blocks of amphibolites may correspond to the subducted oceanic rocks and the lack of blueschists may be explained by a high geothermal gradient in the subduction zone, as has been explained for amphibolites in central Cuba (Garcia-Casco et al., 2002; Auzende et al., 2002) and the Franciscan Complex, California (Oh and Liou, 1990). High geothermal gradients are common in subduction zones where young oceanic crusts are subducted, such as SW Japan (Aoya et al., 2003). The evidence suggests that the absence of blueschists in the central part of the ITSZ does not preclude the existence of subduction related metamorphism.

6.2. Timing of the India–Asia collision

The timing of the India–Asia collision has been in debate (e.g. Rowley, 1996; Guillot et al., 2003; Leech et al., 2005; Najman et al., 2006). In Southern Tibet, the age of collision was dated between 40 and 46 Ma (Lutetian) by Rowley (1996) and 50.6 ± 0.2 Ma (Ypresian) by Zhu et al. (2005), based on detrital zircon ages and stratigraphic data assuming that the India–Asia collision immediately produced topographic highs to erode newly formed zircons. These ages contradict with paleomagnetic data showing a sharp decrease in India–Asia convergence rate between 60 and 55 Ma (e.g. Klootwijk et al., 1992; Lee and Lawver, 1995; Acton, 1999). In the western Himalaya, the first topographic highs likely appeared in the suture zone a few million years after initial contact below the sea level (Garzanti et al., 1987; Guillot et al., 2003). Therefore, detrital age of sedimentary rocks gives a minimum age, implying that the initial India–Asia contact in Southern Tibet occurred before 50.6 Ma.

In the western part of the Himalaya, along the Tso Morari–Zanskar transect, the initial India–Asia contact was before 53.3±0.7 Ma, the age of the UHP metamorphism, and after 58±1 Ma which corresponds to the onset of decreasing convergence rate between India and Asia, according to paleomagnetic data. de Sigoyer et al. (2000) and Guillot et al. (2003) proposed an age of 53–57 Ma for the initial India–Asia convergence, which was revised by Leech et al. (2005) to 56–58 Ma following a new date obtained from the Tso Morari unit. The combination of these data suggests initial India–Asia contact at 56±3 Ma in the Tso Morari area.

This age is substantially older than the initial India–Asia contact west of the Nanga Parbat in the Kaghan area. The UHP metamorphism is precisely dated at 46.2±0.7 and 46.4±0.1 Ma (Kaneko et al., 2003; Parrish et al., 2006) suggesting a difference of 7 Ma with the UHP Tso Morari massif. Some workers suggest that this difference is related to geochronological uncertainties (e.g. Treloar et al., 2003). However, even the retrogressed age of the Tso Morari at 47±0.5 Ma is older than the age of the peak metamorphism in the Kaghan massif, suggesting that the UHP rocks, east of the Nanga Parbat spur, had been already exhumed when the Kaghan rocks were being subducted. As the Kaghan unit belongs to the distal part of the northern Indian margin, the difference of 7 Ma is important and discussed in terms of the initial geometry of the Indian plate.

6.3. Initial geometry of north India continental margin

The main difference between the western Himalaya and southern Tibet is the absence of the HHC above the MCT in the western Himalaya (Figs. 1 and 2a). The Nanga Parbat massif is characterized by abundant granulites and other high-temperature metamorphic rocks as observed in the HHC. However Nd model ages of the basement of the Nanga Parbat range between 2.3 and 2.8 Ga, and are similar to the LHS instead of the HHC. In the same way, the metamorphic cover of the Nanga Parbat massif shows Nd model ages varying between 1.8 and 1.3 Ga, typical of the Haimanta Formation (Whittington et al., 1999; Di

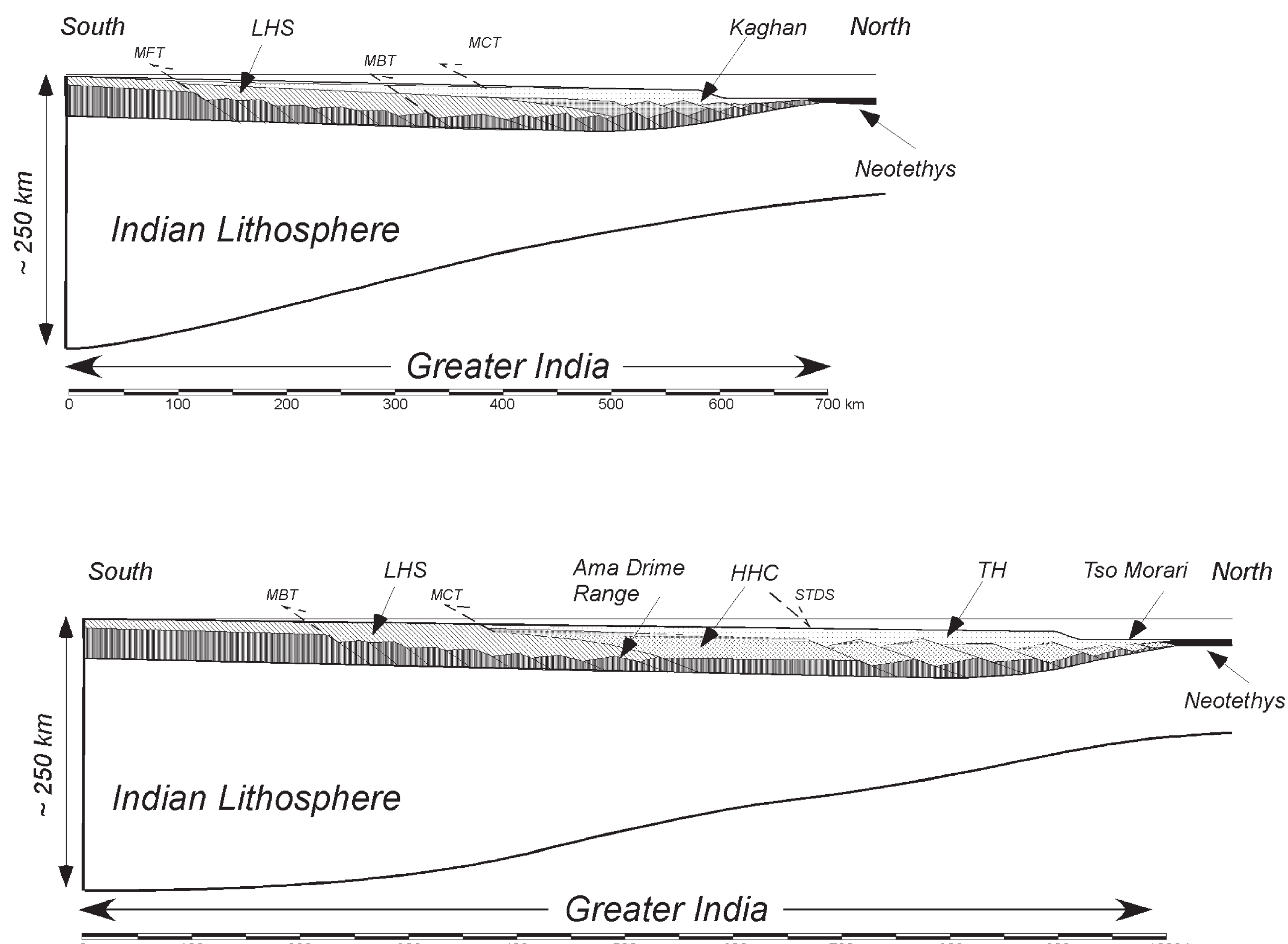

Fig. 7. a) Restored cross-section of the northern margin of the Greater India west of the Nanga Parbat syntaxis. Notice that the future internal zone is relatively narrow and the HHC is absent. b) Cross-sectional restoration of the northern margin of the Greater Indian, east of the Nanga Parbat syntaxis, valid for the Tso Morari and Ama Drime range sections. Notice that the internal zone is wider and the HHC is well developed (modified after DeCelles et al., 2002). MFT: Main Frontal Thrust; LH, Lesser Himalaya; MBT: Main Boundary Thrust; MCT: Main Central Thrust; HHC: Higher Himalayan Crystallines; STDS: South Tibetan Detachment System; TH: Tethys Himalaya.

Pietro and Pogue, 2004). Below the Panjal thrust, (the lateral equivalent of the MCT), the Lesser Himalayan Sequence is similar to what is observed in southern Tibet (Fig. 7b), but the main difference is that exposed Himalayan shortening is mainly concentrated south of the thrust, in the unmetamorphosed foreland (Fig. 2a). Directly above the Panjal thrust, the HHC is absent and the Haimanta formation, the product of the Pan-African orogeny (Garzanti et al., 1986), directly overlies the Archean to Paleoproteroic metasediments of the Indian shield (Fig. 7a). In the western Himalaya the Lower Paleozoic and upper Mesozoic series are mostly absent (e.g. Di Pietro and Pogue, 2004), the Tethyan succession is reduced to the Carboniferous and Triassic sediments with abundant Upper Permian volcanic rocks, the so-called Panjal volcanic rocks. The absence of pre-Permian sedimentary rocks in the western Himalaya is explained by the Permo-Triassic rifting (Pogue et al., 1992; DiPietro et al., 1993) or an early obduction (Dipietro et al., 2000) that took place at about 65 Ma (Searle et al., 1987; Beck et al., 1995). Another possible explanation involves the removal of post-rifting sediments (post-Triassic) during the initial subduction of the distal part of the Indian margin (Guillot et al., 2000). These possibilities can explain the absence of the sedimentary cover but not the absence of the HHC basement in the western Himalaya (Fig. 7a).

In addition, shortening in the western Himalaya during the India–Asia collision is less than 600–700 km (Coward and Butler, 1985; Di Pietro and Pogue, 2004), whereas it is greater than 1000 km in southern Tibet (Le Pichon et al., 1992; DeCelles et al., 2002; Guillot et al., 2003). The difference suggests that the north south length of Greater India that was subducted beneath Asia was shorter west of the present-day Nanga Parbat spur. Assuming a similar velocity of convergence along strike of the Himalayan belt, the two different ages of the UHP metamorphic rocks on both sides of the Nanga Parbat spur suggest a later contact between the Indian and Asian continents in the western part. Based on the ages of the initial India–Asia contact and the velocity of the Indian plate, Guillot et al. (2007) estimated that the north–south extension of Greater India in the western part is shorter by 300 to 400 km than its extension in the eastern part (Fig. 6). The value is similar to 350 km estimated by Ali and Aitchinson (2005) based on paleogeographic data and 325 km estimated from the lengths of unfolded Indian margin in the western part and in the central part after removing the HHC (Fig. 7). We propose that the abrupt decrease in thickness of the HHC and a shorter extension of Greater India in the western Himalaya are due to pre-existing north-trending faults (Fig. 8). Considering the lack of the Lower Paleozoic and Mesozoic sediments in the western part, transform faults probably formed during the rifting from Devononian to Albian times (e.g. Di Pietro and Pogue, 2004). The present-day "en baionnette" geometry of the far western part of the Himalaya towards the Chaman fault may also be explained by the reactivation of these earlier faults.

6.4. Subductability of the Indian continental plate

Buoyant continental lithosphere is not easily subducted during the collision of two continents, yet the Indian continent

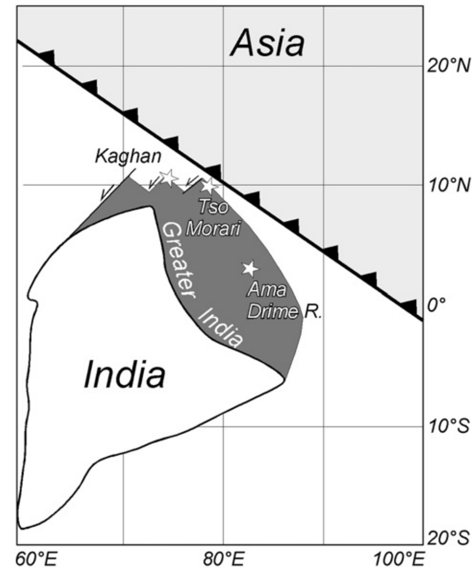

Fig. 8. Proposed Greater India geometry before India–Asia collision. The eastern geometry of Greater Indian fit with the Wallaby–Zenith Fracture zone (Ali and Aitchinson, 2005) whereas its western part fit with the initial location of the UHP Kaghan and Tso Morari units. The position of the Indian craton and the Asian margin at ~55 Ma using pole of Acton (1999). Greater India (dark grey) corresponds to the Indian continent involved in the subduction–collision zone.

subducted beneath the Asian continent. There are several factors that likely contributed to the subduction of the Indian continent during the collision with the Asian continent. South of the collision zone, tomography images show that the Indian lithosphere has the thickness of about 250–300 km (Van der Voo et al., 1999), which is similar to that of the Canadian Shield (Jaupart and Mareschal, 1999). However, the northern margin of the Indian continent becomes thin towards the Himalayan collision zone due to successive tectono-thermal events, such as the Pan-African orogeny, and Permo-Triassic and Albian riftings (Fig. 7). The thinned continental lithosphere can be subducted (Ranalli et al., 2000). The earlier rifting also resulted in denser mass of Greater India by incorporating flood basalts and underplating mafic magmas. A significant volume of flood basalts occurs in the Kashmir basin where the basaltic flows reach a thickness of about 2.5 km. Panjal Traps related to the Permian rifting event are also described in Ladakh (Honegger et al., 1982), Zanskar (Vannay and Spring, 1993) and in the western syntaxis of NE Pakistan (Papritz and Rey, 1989; Spencer et al., 1995). In the Kaghan valley, the Panjal volcanic rocks are metamorphosed to eclogite and amphibolite facies and represent a total thickness of up to 2 km (Spencer et al., 1995). Metamorphosed upper crust containing 10% basalts is calculated to have an increased density by 100 kg m^{-3} (modified after Bousquet et al., 1997) (Fig. 6), which allows the continental crust to subduct (e.g. Ranalli et al., 2000).

6.5. Formation and exhumation processes of UHP rocks

UHP rocks are only recognized in the western Himalaya. In the eastern part, rocks with similar protoliths are only affected by greenschist to amphibolite facies metamorphism (Burg et al., 1984; Chen et al., 1990; Guillot et al., 1998). The scarcity of UHP rocks in the Himalaya could be either due to insufficient exploration in the internal zone, no exposure of such rocks, or no exhumation of such rocks. The oblique convergence of the western Indian margin with the Asian plate may explain the difference between the western Himalaya and southern Tibet (de Sigoyer et al., 2004). The strain partitioning associated with this obliquity has been noted in the western syntaxis (Seeber and Pêcher, 1998). Early dextral motion in the ITSZ has probably assisted the exhumation of the UHP massifs (de Sigoyer et al., 2004) as strain partitioning plays a major role in the vertical motion of HP and UHP rocks (e.g., Fossen and Tickoff, 1998). Rapid exhumation rates of the UHP rocks are compatible with such vertical motion (Leech et al., 2005; Parrish et al., 2006; Guillot et al., 2007). Guillot et al. (2007) estimated an initial subduction angle of ca. 40° and greater than 20° west and east of the present-day Nanga Parbat massif, respectively (Fig. 9). Such a steep subduction is now shown to the depth of 200–300 km beneath the Hindu Kush by tomographic images and seismic studies (Roecker, 1982; Van der Voo et al., 1999; Negredo et al., 2007), which suggests the on-going formation of UHP metamorphic rocks at present (Searle et al., 2001). In contrast, tomographic images and seismic data show that the Indian plate dips 9° beneath southern Tibet with a maximum crustal depth of 80 km. This gently dipping continental subduction occurred in a context of frontal collision. (Nelson et al., 1996; Van der Voo et al., 1999; Zhou and Murphy, 2005). Sapin and Hirn (1997) and Schulte-Pelkum et al. (2005) suggested that the high-velocity of the Indian crust beneath the Tethys Himalaya and the ITSZ is related to a partially eclogitized Indian lower crust. But the low angle of underthrusting can only produce HP rocks in the lower crust (maximum pressure of 25 kbar). Moreover, as only the upper crust is exhumed in the Himalaya, the maximum pressure recorded by exhumed rocks in this shallow subduction context, should be of a maximum of 20 kbar. Thus, we propose that the absence of early steep subduction of the Indian plate in southern Tibet did not allow formation of UHP rocks.

6.6. HP rocks in the LHS: subduction or collision processes?

The occurrences of eclogites in the western Himalaya and their apparent absence in southern Tibet are considered to be a major difference between the two regions (Guillot et al., 1999, 2003). Lombardo and Rolfo (2000) and Groppo et al. (2007) showed that eclogitic metamorphism in southern Tibet had been later obliterated by the later thermal events related to MCT. They suggested that the LHS in southern Tibet also underwent subduction processes and that the Ama Drime Range eclogites are presumably synchronous with those in the NW Himalaya. However, there are two major differences between the western Himalaya and the southern Tibet HP rocks; the location of the eclogites and their thermal evolution. As previously discussed, the western Himalayan UHP rocks are related to the early subduction of the distal part of the Indian margin (Fig. 9a) and their rapid exhumation to upper crustal level, before 40 Ma (e.g. Treloar and Rex, 1990; de Sigoyer et al., 2000; Treloar et al., 2003; Schlup et al., 2003; de Sigoyer et al., 2004) occurred along the ITSZ. The Ama Drime Range eclogites in southern Tibet are different from eclogites in the western Himalaya; they are exhumed within the MCT zone, farther South of the ITSZ, during Miocene time. The work of Catlos et al. (2002) in the Dudh Kosi valley, 20 km east of the Arun valley, shows a monazite Th–Pb age at 45.2±2.1 Ma in HHC rocks, suggesting

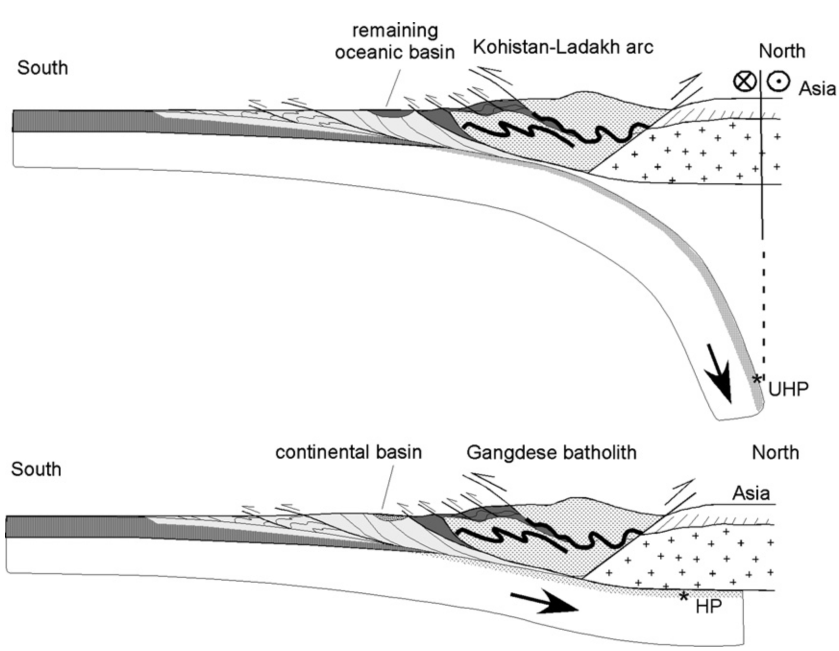

Fig. 9. A geodynamical model for the initial subduction of the Indian continental margin during the Paleocene–Eocene. In western Himalaya (a), the steep subduction of the Indian continent resulted in the development of wrench tectonics on the Asian side due to strain partitioning and the formation of UHP metamorphism on the Indian side. In southern Tibet (b), frontal subduction leads to the underthrusting of Indian slab beneath southern Tibet producing only HP rocks.

an earlier underthrusting of this zone during Eocene time (Burg, 2006) along a flat plane (Fig. 9b). As previously discussed, a shallow underthrusting of the Indian plate beneath southern Tibet would give pressures necessary for eclogite facies metamorphism, but not UHP metamorphism. The *P*–*T* conditions are slightly higher than those reported by Pognante et al. (1990) in the HHC of Zanskar (1.0±0.5 GPa and 650–720 °C) at the transition from HP granulite to eclogites during the so-called Eohimalayan metamorphic event (before 30 Ma). From 45 to 20 Ma, the long-lived slip along the MCT (Hodges et al., 1996; Guillot and Allemand, 2002) allowed the metamorphic rocks to be heated (transformation from low-pressure eclogite to granulite) before its final exhumation along a ramp structure at about 14 Ma, as recorded in the Ama Drime Range retrogressed eclogites but also in the MCT in the adjacent Dudh Kosi valley (Catlos et al., 2002). Thus, we propose that the Ama Drime Range eclogites formed during underthrusting of the Indian plate beneath southern Tibet but underthrusted rocks probably never reached the crust–mantle boundary. The Late Miocene exhumation of these rocks most likely took place during collisional processes.

7. Conclusion

HP and UHP rocks which originated from a variety of protoliths occur in different settings in the Himalayan belt: accretionary wedge, oceanic subduction, continental subduction and continental collision. Although the subduction of the Tethys Ocean and Greater India was continuous since 120 Ma, exhumation of HP and UHP rocks took place for a short period between 100 and 80 Ma. Exhumation processes are controlled by plate velocity change, occurrence of oceanic aspirities and finally introduction of the buoyant Indian margin beneath the southern Asian margin. Oblique convergence of the Indian continent in the western Himalaya also played a significant role in the exhumation of UHP rocks by producing steep subduction. In contrast, frontal convergence in southern Tibet induced a shallower buoyant subduction, only producing HP rocks. The timing of UHP metamorphism in the western Himalaya constrains the initial geometry of the Indian continental margin. The difference of 7 Ma in the peak metamorphic ages between the Kaghan and Tso Morari units is explained by "en baionnette" geometry of the northern Indian margin. The geometry is mostly produced by north-trending faults inherited from successive rifting since the Late Paleozoic. Our proposed interpretation about HP and UHP locations along the Himalaya predicts that other HP to UHP rocks should be present in the ITSZ, in the eastern Himalayan syntaxis. The dating of these hypothetical HP to UHP rocks would further constrain the geometry of the Indian plate during the final closure of the Tethys Ocean.

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