Initiation of crustal-scale thrusts triggered by metamorphic reactions at depth: Insights from a comparison between the Himalayas and Scandinavian Caledonides

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[1] Active eclogitization has recently been inferred at depth beneath the Himalaya from geophysical observations, and the mechanical consequences of eclogitization can be observed in the field in the eroded and extended nappe stack of the Scandinavian Caledonides. There, Proterozoic metastable granulites and igneous protoliths underwent partial eclogitization during the collision of Baltica with Laurentia. The reaction began in pseudotachylites and veins and eventually formed a connected network of eclogite-facies shear zones that localized deformation and weakened the lower crust of Baltica during the collision with Laurentia. All these features can be compared with the seismic activity of the Indian Lower Crust, its strength loss beneath the Himalayan ranges, and its delayed density increase regarding its penetration in the eclogite facies. The Caledonian Bergen Arc eclogites and the Himalayan Ama Drime eclogites are both derived from continental crust. In both cases, these eclogites were formed contemporaneously with the activation of the main thrusts responsible for the construction of the orogenic wedges, the Main Central Thrust in the Himalayas, and the main thrust below the Jotun Nappe Complex in the Caledonides. The similarities in these two orogens, which compare both in size and structure, highlight the importance of eclogitization at depth as a mechanism for weakening of the lower crust and for decoupling of the crust and lithospheric mantle in collision zones. **Citation:** Labrousse, L., G. Hetényi, H. Raimbourg, L. Jolivet, and T. B. Andersen (2010), Initiation of crustal-scale thrusts triggered by metamorphic reactions at depth: Insights from a comparison between the Himalayas and Scandinavian Caledonides, *Tectonics*, *29*, TC5002, doi:10.1029/2009TC002602.

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1. Introduction

[2] Large overthrusts involving the continental basement are the main structures responsible for crustal shortening and thickening in Phanerozoic orogens. The Main Central Thrust (MCT) in the Himalayas or the Austro-Alpine thrust in the Alps are well-known examples. Models explaining the exhumation of high-pressure (HP) and low-temperature (LT) metamorphic nappes also involve large thrusts. The kinematic and mechanical relations between deep-seated processes in subduction zones and the formation of thrusts at the surface are not entirely clear. The decoupling of large units of continental crust from the subducting lithosphere requires an efficient mechanism to weaken the crust at depth and near the surface. At shallow depth, thrusts form along lithological interfaces and décollement levels propagate along weak stratigraphic levels. More homogeneous crystalline basement units require strain-localizing processes involving fluids and metamorphic reactions to produce weak phases [*Wibberley*, 2005]. At greater depth, in the stability field of eclogite (deeper than 50 km), synkinematic eclogitization assisted by fluids also leads to softening and localization of shear zones [*Austrheim*, 1991]. Such deep-seated phenomena can be observed in the exhumed roots of mountain belts but can only be inferred from active orogens.

[3] The Scandinavian-NE Greenland Caledonides and the Himalayan-Tibetan ranges are comparable in size, and the Caledonides are often considered to have been of a height similar to the Himalaya [*Andersen*, 1993; *Andersen et al.*, 2002]. The internal architectures of both orogens are similar, and the orogenic phases in both orogens have similar timescales. The exhumed deep levels of the Caledonian Orogen may be equivalent to the deep root of the Himalayas [*Jackson et al.*, 2004; *Ladenberger et al.*, 2009], although the latter cannot be observed directly and must be inferred from geophysical observations. It is therefore possible to use the field studies from the Caledonides as a guide to interpreting geophysical images of the Himalaya.

[4] On the basis of this comparison, we propose here that eclogitization of the lower crust is the decoupling process necessary for the initiation of large-scale overthrusts characterizing collisional wedges and the syncollisional exhumation of HP metamorphic rocks in their hanging wall.

2. Himalayas Versus Caledonides: A Comparison

[5] The Appalachian-Caledonian Orogen, from Svalbard through Scandinavia and East Greenland to the British Isles

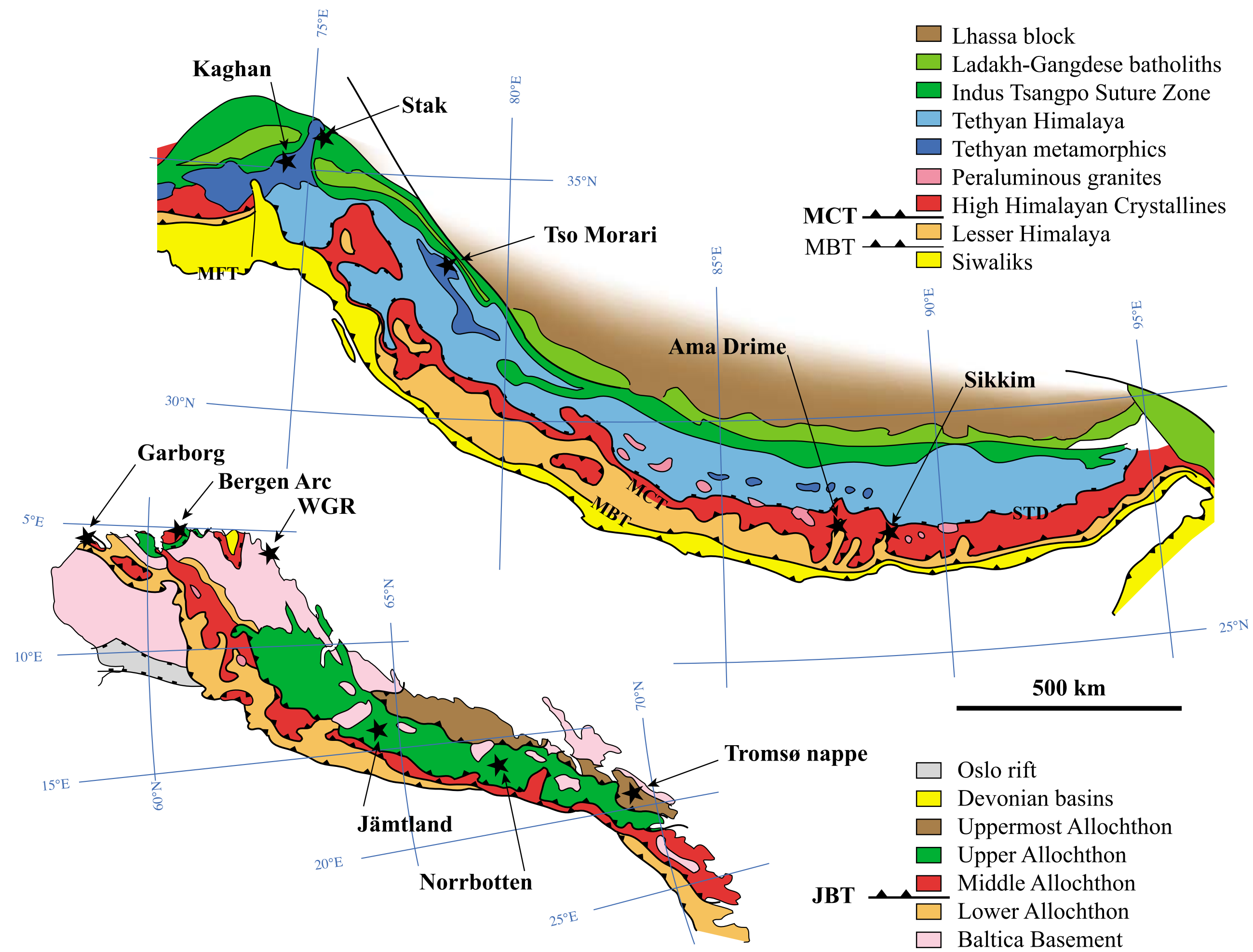

Figure 1. Compared tectonostratigraphies for (top) Himalayas and (bottom) Scandinavian Caledonides modified from *Guillot et al.* [2008] and *Gee and Sturt* [1985]. Projections are different but the scale bar is the same for both maps. MFT, Main Frontal Thrust; MBT, Main Boundary Thrust; MCT, Main Central Thrust; STD, Southern Tibet Detachment; JBT, Jotunheimen Basal Thrust; Stars, eclogite occurrences.

and into the Atlantic North America, is the major Palaeozoic structure that can be considered to represent the characteristic behavior of the continental lithosphere 470–400 Myr. It has previously been compared with the Hercynian orogeny [*Rey et al.*, 1997]. The attempts to compare it with the Alpine-Himalayan Orogeny [*Zhou*, 1985; *Atherton and Ghani*, 2002] mainly use the recent example as a model to understand the ancient and less constrained one. Rare attempts compared field evidence in the Caledonides with geophysical data in the recent Alpine orogen [*Austrheim*, 1991; *Jackson et al.*, 2004]. The Himalaya-Karakoram-Tibetan Orogen was also used as a reference model compared with the Trans-Hudson Orogen in order to compare ancient and recent tectonic processes [*St-Onge et al.*, 2006], showing that the uniformitarian approach is possible over the past 1.7 Ga. Geological processes now active in the Himalayas were thus also active 430–400 Myr. The Caledonides have a size and an internal structural architecture that compare well with the Himalayas in terms of tectonostratigraphy. A comparison of the time sequence of Himalayan and Caledonian tectonic events shows striking similarities, supporting the idea that a generic orogenic process is common to both belts. Both mountain ranges developed from the stacking of passive margins that were first subducted to great depths and then emplaced onto their respective continents.

[6] In the Himalaya (Figure 1a), five main units can be distinguished from bottom to top [*Gansser*, 1964; *Le Fort*, 1975]. (1) The Subhimalayan Ranges, above the Main Frontal Thrust (MFT), represent the deformed part of the Siwalik foreland basin. (2) The Lesser Himalayan Sequence (LHS), bounded by the MBT and MCT, is derived from the Precambrian Indian continent. (3) The Higher Himalayan Crystallines (HHC), with younger rocks affected by PanAfrican orogeny, represent the Gondwana margin of Greater India, that has been thrust above the LHS. This thick crustal nappe stack is limited at its top by the South Tibetan Detachment (STD). (4) The Tethyan Himalaya comprise a Palaeozoic to Cenozoic marine sequence and mafic intrusives from the distal part of the Indian margin. The metamorphic rocks cropping out within the North Himalayan

Figure 2. Pressure-temperature estimates for peak conditions of the different eclogite occurrences in Himalayas (light gray) and Scandinavian Caledonides (dark gray) compared with the pressure-temperature distribution in the Indian Lower Crust from India toward Tibet deduced from thermokinematic modeling [*Hetényi et al.*, 2007]. TM, Tso Morari; K, Kaghan; S, Stak; AD, Ama Drime; NS, North Sikkim; WGR, Western Gneiss Region; TN, Tromsø nappe; G, Garborg; N, Norrbotten; BA, Bergen Arc; J, Jåmtland. References are as in Table 1.

domes are considered to be remnants of the subducted basement of the Tethyan sequences. (5) Farther north, ophiolites and flysch units as well as arc-related magmatic bodies of Kohistan-Ladakh and Gangdese represent the Indus-Tsangpo Suture Zone (ITSZ) between the Indian terranes and the Lhasa block.

[7] The Scandinavian Caledonides (Figure 1b) consist of four main allochthons thrust southeastward onto Baltica [*Roberts and Gee*, 1985]. (1) The Lower Allochthon (LA) comprises shelf and continental series of Baltic origin. Basement windows bring Baltica or correlated basement to the surface through the nappe stack. The Western Gneiss Region (WGR) is the largest of these windows. (2) The Middle Allochthon (MA), comprising the Jotun Nappe Complex and the Lindås Nappe, among others, involves more distal cover and basement units. The MA is considered here to be the main unit of the nappe wedge thrust above the basal imbricate décollement zone below the Jotun Nappe Complex (hereafter called Jotunheimen Basal Thrust, JBT; *Fossen* [1992]). (3) The Upper Allochthon (UA) has ophiolites and magmatic arcs representing vestiges of the Iapetus Ocean domain [*Stephens et al.*, 1985]. (4) The Uppermost Allochthon (UMA), on top of the nappe pile, has Laurentian affinities [*Roberts et al.*, 2002].

[8] The HHC and the MA are comparable in position and origin. Both represent a thick sequence of high-grade metamorphic rocks derived from the cratonal basement and its cover, which were thrust over an autochthon through thick reverse ductile shear zones, the MCT and JBT, respectively.

3. Himalayan and Caledonian Eclogites

3.1. Occurrences

[9] Eclogites (Figure 1, Table 1) are known from several structural levels in both orogens. In Himalaya, eclogites are found close to the ITSZ in the North Himalayan massifs (Tso Morari, Stak) and the HHC (North Sikkim). The occurrence of metamorphic Tethyan sediments in the Kaghan unit suggest that these eclogites belong to the HHC [*Guillot et al.*, 2008]. The Ama Drime eclogites are found in orthogneisses probably derived from the upper part of a lower crust correlated to the LHS [*Groppo et al.*, 2007; *Cottle et al.*, 2009]. On the basis of lithological associations, pressure-temperature paths and ages of metamorphism, the crystallization of these eclogites can be assigned to different burial processes. The Stak eclogites occur as lenses within continental gneisses, and reached 2.3 GPa (Figure 2) at 51 Myr [*Riel et al.*, 2008]. The Kaghan metasedimentary sequence was metamorphosed at ultrahigh pressure (UHP) at 46 Ma [*Parrish et al.*, 2006] and the Tso Morari metabasites, included in orthogneiss, reached UHP conditions between 53 and 55 Ma [*Guillot et al.*, 2003; *de Sigoyer et al.*, 2004; *Leech et al.*, 2005]. These three massifs are markers of the subduction of Tethys and the Indian margin. The Ama Drime and North Sikkim eclogites, at lower tectonostratigraphic levels, recorded lower pressures (Figure 2) and a warmer *P–T* evolution with a posteclogite granulite overprint [*Groppo et al.*, 2007]. Lu-Hf dating of garnets from the Ama Drime eclogite relicts yield a 21 Ma age for their maximum burial [*Corrie et al.*, 2007]. The North Sikkim eclogites are undated but are likely equivalent to the Ama Drime eclogites, representing syncollisional burial of crustal slices derived from the Indian shield [*Rolfo et al.*, 2008].

[10] In the Caledonides, eclogites are present in the UMA (Tromsø nappe), UA (Norrbotten and Jämtland), MA, LA, and WGR basement. Some eclogites have pillow-lava remnants, and their ages (452 Ma for the Tromsø eclogites, 454 Ma for the Jämtland occurrence, and 500 Ma for Norrbotten [*Brueckner and van Roermund*, 2004; *Corfu et al.*, 2003; *Essex et al.*, 1997]) suggest that they are markers of early Iapetus Ocean subduction or isolated Laurentian or Baltica crustal rafts. The Bergen Arc eclogites, in the Lindås Nappe, which is part of the MA, reached the eclogite facies at 425–430 Myr [*Glodny et al.*, 2008] during the initial stages of the main Scandian Caledonian continental collision.

Table 1. Main Characteristics of the Eclogite Occurrences in the Himalayas and Scandinavian Caledonides[a]

Occurrence	Lithologies	Tectonostratigraphy	Peak PT Conditions	Retrogression Conditions	Age of HP (Ma)	Geodynamic Context	Ref.[b]
			Himalayan Eclogites (From W to E)				
Kaghan	Metasedimentary sequences	Nanga Parbat gneisses in the HHC	3.0 ± 0.2 Gpa 770°C ± 50°C	A or BS then GS	46	Subduction of Indian margin	1
Stak	Metabasites in orthogneiss	North Himalayan massif (HHC)	2.3 GPa, 750°C	A	51	Subduction of Indian margin	2, 3
Tso Morari	Metabasites in orthogneiss	North Himalayan massif (HHC)	>3.9 GPa, >750°C	A or BS then GS	55–53	Subduction of Indian margin	1
Ama Drime	Metabasites in orthogneisses	LHS	1.5–2.0 GPa > 580°C	G then A	21	Collision	4–6
North Sikkim	Metabasites in orthogneisses	HHC	>1.5 Gpa > 600°C	G then A	?	Collision	7
			Scandinavian Eclogites (From N to S)				
Tromsø nappe	Mestabasites in calcsilicates	Uppermost Allochthon	2.8 Gpa 725°C	A	452	Laurentian m. subduction	8
Norrbotten	Pillow-lavas	Seve nappe complex (Upper Allochthon)	1.5–2.7 Gpa 610–750°C	A	500475	Continental subd. in the north	9
Jämtland	Metabasites in paragneisses	Seve nappe complex (Upper Allochthon)	1.4–1.8 Gpa 550–780°C	A	454 ± 4.5	Iapetus subduction	10
Western Gneiss	Basic lenses in gneisses pseudotachylites internal eclogites in mantle peridotites	Western Gneiss basement window	up to 3.5 GPa up to 800°C	A G locally	425–400	Late continental subduction	11
Bergen Arc	Shear zones in granulites pseudotachylites	Middle Allochthon	1.8–2.1 GPa 700°C	A	425–430	Collision	12
Garborg	Lenses within gneisses	Middle Allochthon	2.3–2.8 GPa 585–655°C	A	455–471	Subduction ?	13, 14

[a]G, Granulites facies; A, Amphibolite facies; BS, Blueschists facies; GS, Greenschists facies.
[b]References (recent syntheses have been privileged when available) 1, *Guillot et al.* [2003]; 2, *Le Fort et al.* [1997]; 3, *Riel et al.* [2008]; 4, *Lombardo and Rolfo* [2000]; 5, *Corrie et al.* [2007]; 6, *Cottle et al.* [2009]; 7, *Rolfo et al.* [2008]; 8, *Corfu et al.* [2003]; 9, *Essex et al.* [1997]; 10, *Brueckner and van Roermund* [2004]; 11, *Hacker* [2007]; 12, *Glodny et al.* [2008]; 13, *Smit et al.* [2008]; 14, *Smit et al.* [2009] . Gray level, Crystallization ages of North Sikkim eclogites is not known.

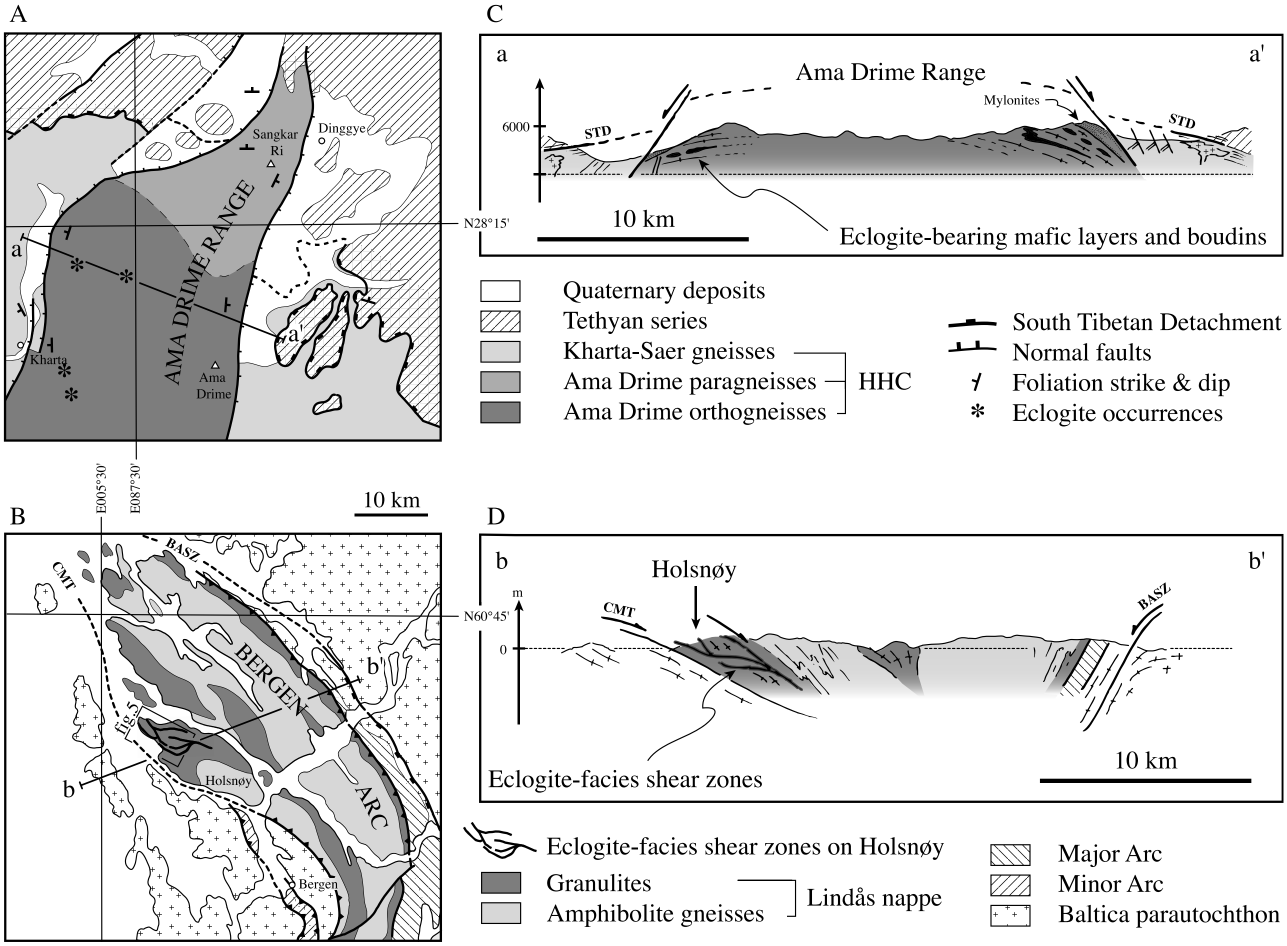

Figure 3. Geological context and structural positions of the Ama Drime and Holsnøy eclogites. (a) Structural sketch of the Ama Drime Range [after *Kali et al.*, 2010]. Eclogite occurrences after *Groppo et al.* [2007]. (b) Structural sketch of the Bergen Arc [after *Ragnhildsveit and Helliksen*, 1997]. The eclogite-facies shear zones on Holsnøy are simplified from *Austrheim et al.* [1996b]. CMT, Caledonian Main Thrust; BASZ, Bergen Arc Shear Zone. (c) Cross section through the Ama Drime Range along a-a′ [after *Kali et al.*, 2010]. (d) Cross section through the Bergen Arc along b-b′.

Eclogites, present as lenses within the Jaeren nappe, also part of the MA give ages between 471 and 455 Myr for peak metamorphism [*Smit et al.*, 2009]. Large numbers of eclogites in WGR gneisses reached HP or UHP between 425 and 400 Myr [*Kylander-Clark et al.*, 2007] during a late stage of continental subduction. The retrogression of Caledonian eclogites occurred mainly at amphibolite- to greenschist facies [*Cuthbert and Carswell*, 1990], although local post-eclogite granulites have been described in the NW [*Straume and Austrheim*, 1999].

[11] The peak PT estimates for Caledonidan eclogites (Figure 2) range from 1.4 to 4.0 GPa and from 550°C to 800°C. Except for the UHP domains in the WGR, the eclogites derived from continental protoliths equilibrated at lower pressure than eclogites from oceanic rocks. In both orogens the eclogites young downward in the orogenic wedge similar to the ages of major thrusts. In addition to their structural position, another distinctive feature, examined hereafter, is the timing of eclogitization with respect to the general evolution of the subduction and collision in both mountain ranges.

3.2. Ama Drime and Holsnøy Eclogites Geological Setting

[12] The Ama Drime Range (Figures 3a and 3c) is a horst bounded by N-S trending normal faults crosscutting the South Tibetan Detachment Zone [*Kali et al.*, 2010]. The orthogneisses, where eclogites occur in metabasite layers or boudins are correlated to the lower Himalayan Cristalline, the surrounding Kharta gneisses are part of the upper Himalayan Crystalline [*Groppo et al.*, 2007]. The orthogneisses show a foliation parallel to the flanking faults with mylonitization on a kilometer-thick zone [*Kali et al.*, 2010]. The eclogites, intensely retrogressed in the granulite-facies conditions and scattered in the gneisses, do not exhibit any clear mark of their deformation in the eclogite-facies conditions. The Holsnøy partly eclogitized granulites

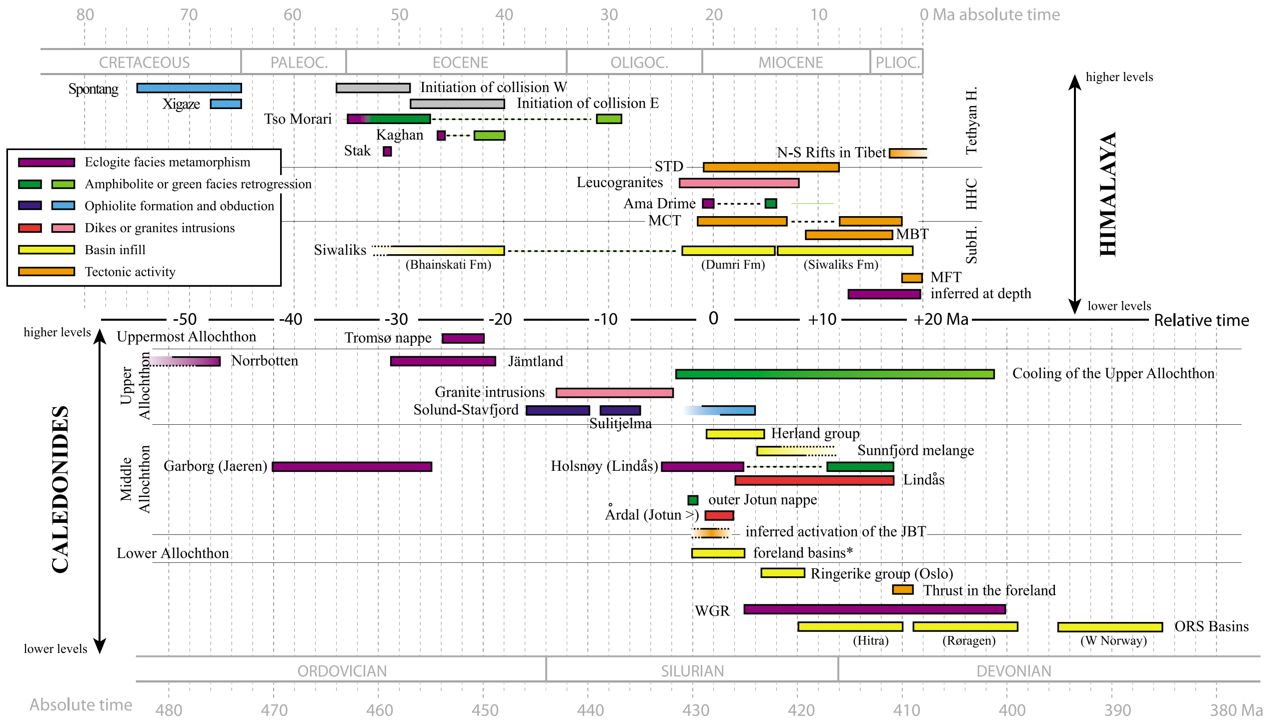

Figure 4. Time tables for the Himalayan and the Caledonian Orogenies. Epoch and periods limits after ICS 2004. The ages of the Ama Drime eclogites and Holsnøy eclogites, Bergen Arcs are used as references for the relative time line (−50 to +20 Ma). Ages for eclogites are detailed in Table 1. In the Lower Allochthon, the Ekeberg greywacke and Bruflat sandstones and Röde formations [*Hossack et al.*, 1985] have been grouped as "foreland basins." Other references are as in the text.

(Figures 3b and 3d) are part of the Lindås nappe, in the Bergen Arc, which is bounded to the East by the Bergen Arc Shear Zone, that accommodated strike slip and extension during the late stages of the Scandian phase. The Holsnøy granulites were preserved from the amphibolite and greenschist-facies retrogression associated with their latest stages of exhumation. The deformation they show happened in the eclogite-facies conditions, where shear was accommodated by a connected network of shear bands [*Austrheim et al.*, 1996b].

4. Comparison of Tectonic Histories

4.1. Himalayan Orogeny

[13] In the Himalaya, the ages of eclogites can be correlated with two main tectonic phases (Figure 4). The inner eclogites (Tso Morari, Kaghan, and Stak occurrences; Table 1) postdate obduction events in the western (Spontang and Nidar ophiolites) or eastern (Xigaze ophiolite) parts of the mountain range [*Corfield et al.*, 2001; *Ding et al.*, 2005]. They are synchronous with the initiation of collision, i.e., the beginning of steady-state convergence, at 57–50 Ma in the west and 49–40 Ma in the east [*Rowley*, 1996; *Guillot et al.*, 2003; *Leech et al.*, 2005]. The Ama Drime eclogites age of 21 Ma (Table 1) is contemporaneous with the oldest leucogranites (e.g., 22–23 Ma [*Hodges*, 2000]), and with the activation of the MCT at 22 Ma and the STD at 20 Ma in central Himalaya [*Hodges et al.*, 1996]. Movement along these major shear zones stopped before the development of N-S trending rifts at ~2–3 Ma, which accommodate E-W extension of the overthickened crust in South Tibet [*Hodges*, 2000; *Yin and Harrison*, 2000; *Mahéo et al.*, 2007]. The main subsidence phase in the foreland basin was recorded by the discordant fluvial deposits of the Dumri formation or lateral equivalents at 23 Ma (early Aquitanian [*DeCelles et al.*, 1998]). The slicing and stacking of the continental slab, which led to the activation of the MCT and the development of the foreland basin on the Indian continent, was therefore synchronous with the crystallization of eclogites in the subducted Indian Lower Crust. This southward progression of the deformation continued at the front of the orogen when the MBT initiated at 11 Ma [*Burbank et al.*, 1997] and the MFT at 2 Ma [*Van der Beek et al.*, 2006].

4.2. Caledonian Orogeny

[14] In the Scandinavian Caledonides, a similar tectonic history is recorded across the 90 Ma time span from 480 to 390 Ma (Figure 4). The earliest eclogites, dated between 459 and 449 Ma (*Brueckner and van Roermund* [2004] and *Essex et al.* [1997] in Table 1) crystallized before, and largely outboard of Baltica [*Andersen and Andresen*, 1994; *Corfu et al.*, 2006], and therefore have limited relevance here. Subsidence has been recorded from lower Ordovician

to lower Devonian by several detrital deposits (Figure 4) related to island arc and back-arc basins during the closure of the Iapetus Ocean.

[15] The main Scandian phase is marked by a number of events. (1) Ophiolites in the UA, such as the Solund-Stavfjord or Sulitjelma ophiolites [*Dunning and Pedersen*, 1988; *Pedersen et al.*, 1991], were obducted after the intrusion of granitic plutons at 443–432 Ma [*Hacker et al.*, 2003]. (2) The Herland Group and overlying Sunnfjord Mélange record obduction of the Solund-Stavfjord ophiolite onto the MA before 425 Ma [*Andersen et al.*, 1990]. (3) Eclogites from Holsnøy in the Bergen Arcs record subduction and exhumation at ~430 Ma [*Glodny et al.*, 2008] contemporaneous with the intrusion of synkinematic dike swarms in the Lindås nappe and in western parts of the Jotun nappe between 427 and 418 Ma [*Kuhn et al.*, 2001; *Lundmark and Corfu*, 2007]; these granitic intrusions are considered to be products of anatexis of deeper levels of the Caledonian nappe stack [*Lundmark and Corfu*, 2007]. (4) $^{39}Ar/^{40}Ar$ ages between 432 and 402 Ma indicate cooling of the UA through the amphibolite and greenschist facies [*Eide et al.*, 2005; *Hacker and Gans*, 2005]. (5) In the Lower Allochthon and the foreland of Baltica, subsidence and detrital influx is initially recorded by the transition from marine (Ekeberg greywacke and Bruflat sandstones) to continental formations of the Ringerike Group [*Hossack et al.*, 1985]. (6) The diachronous continental infill in the foreland ranges from Ludlow to Pridoli in the Ringerike Group (423–418 Ma) in the Oslo graben [*Davies et al.*, 2005].

[16] The final stages of contractional deformation recorded in the foreland [*Andersen*, 1998] were synchronous with the late stages of (U)HP eclogites recrystallization in the WGR [*Kylander-Clark et al.*, 2007, 2009; *Root et al.*, 2005]. At that time, synorogenic collapse of the inner part of the mountain chain was triggered by body forces within the orogenic wedge. Intramontane collapse basins began to form at ~420 Ma in the Hitra basin [*Eide et al.*, 2005].

[17] From these chronological data, it appears that the formation of the Bergen Arc and WGR eclogites was accompanied by (1) construction of the orogenic wedge, (2) subsidence in the foreland, (3) cooling of high-level units, (4) obduction/emplacement of marginal basins, and (5) intrusion of anatectic melts. Using the Himalayan sequence of events as a guide, it is possible to interpret these events to mark the actual activation of the main thrust system of the Scandinavian Caledonides, i.e., the Jotunheimen Basal Thrust [*Fossen and Dunlap*, 1998], at 430 Ma. The most abundant eclogites in the WGR are younger (418–400 Ma) and correspond to the continued underthrusting and eventual subduction of the more external parts of Baltica below the MA, UA, and UMA.

4.3. Comparison of Caledonian and Himalayan Event Sequences

[18] The Caledonian and Himalayan tectonic event sequences show striking similarities if the ages of the Ama Drime and the Bergen Arc eclogites are taken as a common reference (Figure 4). When juxtaposing the Caledonian and Himalayan histories to the same reference time, at 428 and 20 Myr, respectively, the relative age of subsidence episodes in the forelands compare (Figure 4). Eventually, the MFT and the youngest deformation at the front of the Scandinavian Caledonides share a +18 Ma relative age. The succession of eclogite formation stages is the same in both orogenies, with older eclogite formation and exhumation related to subduction dynamics in the upper tectonostratigraphic units being 30 Ma older than collision-related eclogites (i.e., Ama Drime and Holsnøy eclogites). The late WGR eclogites would have the same relative age as the eclogites imaged now at depth beneath the Himalayas and Tibet. The main difference between these two time lines is the interval between the onset of collision and the development of intracrustal thrusts. The initiation of the MCT indeed happened 30 Ma after the onset of the Himalayan collision, whereas the JBT initiation inferred here at ca. 430 Ma seems synchronous with the age considered for the onset of the main Caledonian collision, known as the Scandian phase [*Roberts*, 2003].

[19] The Himalayan and Caledonian orogens show close similarities in structural geometry and size as well as in the succession and duration of tectonic events. The Himalayan sequence of tectonic events shows a close temporal association between foreland subsidence, anatexis, and activation of the MCT. If a similar temporal association applies to the Caledonides, the activation of the main thrust system, i.e., the Jotunheimen Basal Thrust, can be inferred. The compilation of ages in the Caledonides show that near-surface events can be correlated with the formation of large tracts of continental eclogites at depth; a similar correlation may apply to the Himalaya. The Caledonides, now deeply eroded and extended, must present field evidence of a possible coupling between HP recrystallization and crustal-scale thrust activation, whereas the same deep processes must be inferred for the Himalaya from geophysical data.

5. Discussion

5.1. Geophysical Signature of Eclogitization Processes

[20] Receiver-function images beneath the Nepalese Himalaya and the Tibetan Plateau (Figure 5a) delineate features within the orogenic wedge [*Hetényi*, 2007; *Nábělek et al.*, 2009, *Hetényi et al.*, 2010]. The sub-Himalayan Moho apparently flattens at 70–75 km depth 250 km to the north of the MFT. A double interface at ~60 and ~75 km is interpreted as the top and bottom of the underplated Indian Lower Crust (ILC). Thermokinematic and petrological modeling of gravity data shows that the Bouguer anomalies beneath Tibet are best explained by a progressive densification of the ILC from 3100 to 3300 kg·m^{-3} (Table 2) between 250 and 350 km to the north of the MFT [*Hetényi et al.*, 2007] (Figure 4b). The same study concluded that the most probable cause of this density change is delayed eclogitization of the ILC. The distance between the limit of the eclogite facies in the crustal wedge and the actual densification of the ILC is ~150–200 km, which implies a time delay of between 7 and 10 Ma for a convergence rate of 21 mm yr^{-1} [*Bollinger et al.*, 2006]. Local earthquake tomography beneath the Nepalese Himalaya [*Monsalve et al.*,

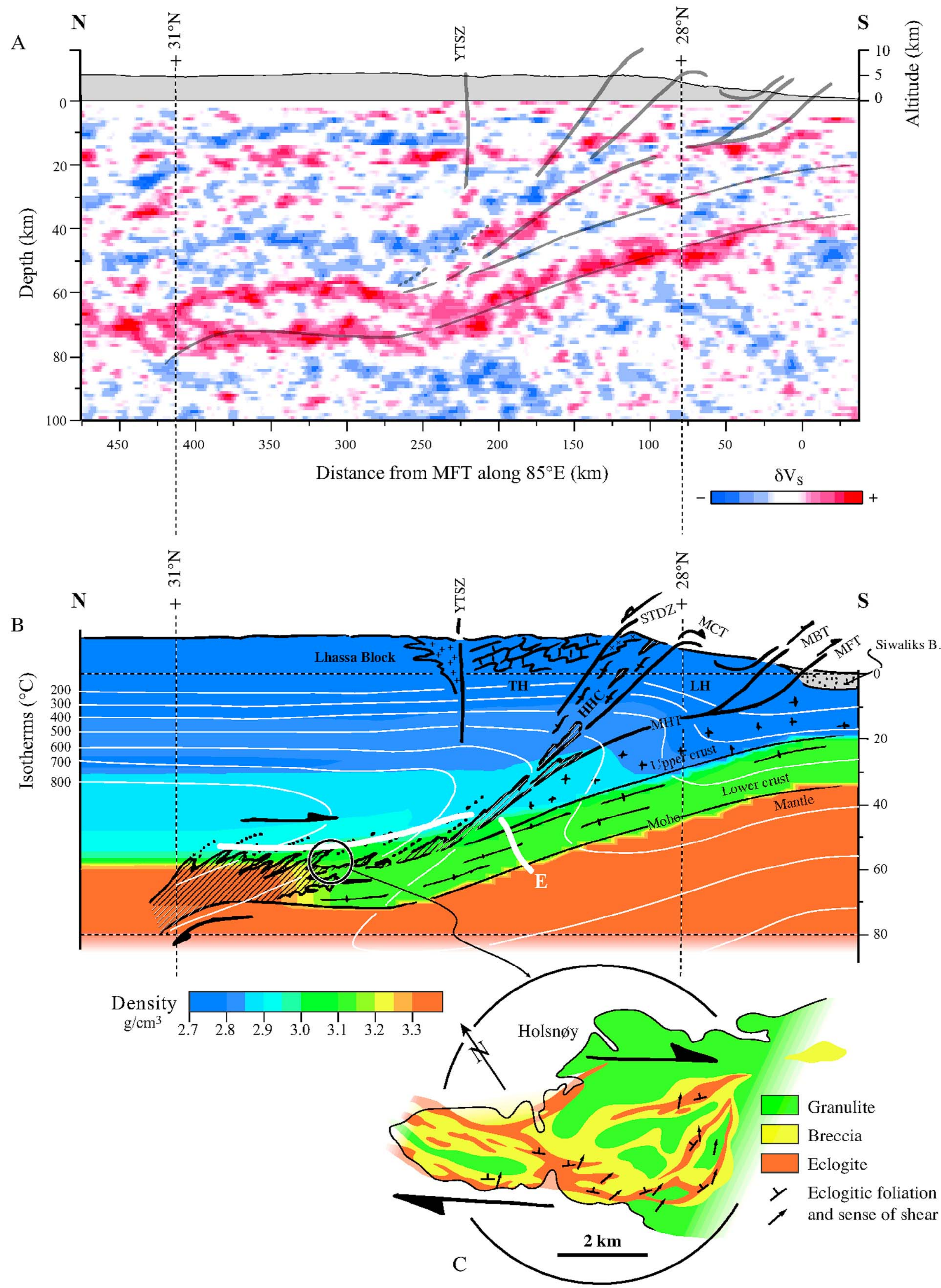
A
N
S
31°N
28°N
YTSZ
Altitude (km)
Depth (km)
Distance from MFT along 85°E (km)
δVs
B
Lhassa Block
TH
LH
HHC
STDZ
MCT
MBT
MFT
Siwaliks B.
MHT
Upper crust
Lower crust
Mantle
Moho
E
Isotherms (°C)
Density
g/cm3
Holsnøy
Granulite
Breccia
Eclogite
Eclogitic foliation
and sense of shear
2 km
C

Figure 5

Table 2. Compared Petrological and Mechanical Characteristics of the Indian Lower Crust and the Holsnøy Granulites and Eclogites

	ILC	Eclogitized ILC	Holsnøy Granulites	Holsnøy Eclogites
Chemistry	Gabbroic diorite[a]	Gabbroic diorite[a]	Gabbroic anorthosite[b]	Gabbro anorthosite[c]
Mineralogy	19%	14%	66%	0 7%
pl	19%	14%	ε	4 6%
q	17%	4%	15%	1 1%
ky	25%	6%	ε	0 27%
bi	20%	28%	18%	39 51%
wm		32%	scap, sp	40 10%
amph				chl, rt
ep				
cpx				
opx				
grt				
access.				
ρ (k gm^{-3})	3100[d]	3300[d]	2750–3100[e]	3060–3480[e]
	3100[f]	3370[g]	3050[g]	3370–3530[g]
V_P (km s^{-1})	7.0[f]	7.5 ± 0.1[g]	7.14[g]	7.83–8.14[g]
		≥8.4[h]	7.16[i]	8.01–8.29[i]
			7.5 at 600 MPa[j]	8.3–8.5 at 600 MPa[j]
V_P/V_S	1.73[f]	1.75[g]	1.80 [g]	1.81 1.78[g]
	1.73–1.75[k]	1.76[h]		
H_2O (wt %)	1[a]	1[a]	0.1–1.1 [l]	0.5–2[l]

[a]As considered for best fit of density changes beneath Himalayas, mineralogies computed via PerPle_X [*Connolly*, 2005] for a gabbroic diorite protolith with 1% water [*Hetényi et al.*, 2007].
[b]Average mode and mineral analysis compiled from *Austrheim and Mørk* [1988], *Fountain et al.* [1994], and *Raimbourg et al.* [2007].
[c]Modes and analysis from *Boundy et al.* [1992].
[d]Deduced from gravity anomalies modeling [*Hetényi et al.*, 2007].
[e]Measured in *Austrheim and Mørk* [1988].
[f]Calculated according to *Hacker and Abers* [2004] at 0.8 GPa and 450°C.
[g]Calculated according to *Hacker and Abers* [2004] at 2 GPa and 700°C.
[h]Deduced from tomography at 70 km BSL [*Monsalve et al.*, 2008].
[i]Calculated according to *Hacker and Abers* [2004] at 600 MPa and 25°C.
[j]Measured in *Fountain et al.* [1994].
[k]Deduced from receiver function analysis beneath the Indian shield [*Kumar et al.*, 2001].
[l]Compiled from *Austrheim and Mørk* [1988], *Boundy et al.* [1992], and *Raimbourg et al.* [2007].

2008] reveals high P wave velocities (above 8.4 km s^{-1}, Table 2) and regular V_p/V_s ratios (1.76, Table 2) below the seismological Moho beneath the Tethyan Himalaya. The presence of eclogites is mentioned as one of the possible causes for this velocity structure. These data show first-order similarities with petrophysical data available for the Holsnøy granulites and eclogites (Table 2). The eclogitization of the Holsnøy granulites, anorthositic to gabbroic in composition [*Austrheim and Mørk*, 1988] is responsible for their densification and the rise of their seismic velocities. Density measurements give a maximum value of 3100 kg·m^{-3} for granulites and between 3060 and 3480 kg·m^{-3} for eclogites [*Austrheim and Mørk*, 1988]. The calculation of densities of Holsnøy eclogites at their peak conditions (700°C and 2 GPa) from their modal composition [*Hacker and Abers*, 2004] gives values between 3370 and 3530 kg·m^{-3}, i.e., the same order as values inferred for the ILC at depth. Measured seismic velocities of Holsnøy rocks change from 7.5 km·s^{-1} to 8.4 km·s^{-1} when eclogitized [*Fountain et al.*, 1994] or from 7.14 km·s^{-1} for granulites to 7.83–8.14 km·s^{-1} for eclogites, when calculated from modal compositions at 700°C and 2 GPa [*Hacker and Abers*, 2004]. The structure of the Holsnøy granulite massif partially eclogitized along kilometer-scale eclogitic shear bands with contrasting seismic velocities would make it a potential highly reflective structure in seismic or receiver function analysis profiles [*Fountain et al.*, 1994].

5.2. Eclogitization as a Weakening Process

[21] Estimates of the effective elastic thickness (EET) of India constrained by gravity anomalies, receiver-function analysis, and thermomechanical modeling [*Hetényi et al.*, 2006] indicate a sharp drop in lithospheric strength, which is interpreted to correspond to decoupling of the crust from the lithospheric mantle [*Hetényi et al.*, 2006]. A dense seismic swarm located close to the Moho in an adjacent zone by independent studies [*Monsalve et al.*, 2006] could be related to this decoupling process. Geophysical data

Figure 5. (a) Receiver function migration profile across Nepalese Himalaya and southern Tibet [*Hetényi*, 2007; *Nábělek et al.*, 2009; *Hetényi et al.*, 2010]. Deduced lithological units labeled in Figure 5b. (b) Density and thermal structures compatible with gravity anomalies [*Hetényi et al.*, 2007]. Geometry deduced from Figure 5a. Hatched, eclogitized Indian Lower Crust. Eclogite facies upper limit deduced from PT field of *Hetényi et al.* [2007]. (c) Holsnøy Island map, Bergen Arcs. Pristine Granulites (cpx + opx + gt + pl), partially eclogitized ("Breccia"), and completely transformed zones (omph + gt + phe ± ky) after *Austrheim et al.* [1996b]. Main foliation and shear senses after *Raimbourg et al.* [2005].

Table 3. Rheological Parameters and Effective Viscosity for Dislocation Creep of Synthetic Mineral Aggregates From the Literature[a]

Material	log A ($MPa^{-n} \cdot s^{-1}$)	n	Q ($kJ \cdot mol^{-1}$)	Experimental conditions T (°C)	P (GPa)	$d\varepsilon/dt$ (s^{-1})	log η at T = 700°C (Pa s) $d\varepsilon/dt = 10^{-14}$ s^{-1}	$d\varepsilon/dt = 10^{-12}$ s^{-1}	References
Omphacitite (jd42)	−2.0	3.5	310	1300–1500	3	10^{-5}–10^{-4}	21.3	19.9	*Zhang et al.* [2006]
Jadeitite	−3.3	3.7	326	?	2.5	?	21.8	20.4	*Orzol et al.* [2002]
Diopsidite (1 wt% H_2O)	5.2	3.3	490	900–1271	0.3	10^{-5}	22.2	20.8	*Bolland and Tullis* [1986]
Eclogite (50% gt)	3.3	3.4	480	1177–1327	3	$10^{-4.4}$–$10^{-3.2}$	22.5	21.1	*Jin et al.* [2001]
Anorthitite (dry)	12.7	3	648	864–1104	0.3	10^{-6}–10^{-4}	22.7	21.4	*Rybacki and Dresen* [2000]

[a] A is preexponential constant, n is stress exponent, and Q is the activation energy in the power law $d\varepsilon/dt = \sigma^n e^{(-Q/RT)}$.

therefore suggest that the ILC is decoupled from the lithospheric mantle at least 200 km north of the MFT and then progressively densified between 250 and 350 km. This early loss of strength could also be explained by eclogitization.

[22] In the Bergen Arcs, a 10 km thick portion of lower crustal granulites on Holsnøy was partially eclogitized in veins and shear bands (Figures 3 and 5c). Models of garnet zoning indicate that these granulites resided at eclogites facies conditions at least 2 Ma before fluid influx triggered reaction (1–4 Ma according to *Erambert and Austrheim* [1993], 0.7–1.8 Ma according to *Raimbourg et al.* [2007a]). The first stages of eclogitization, in Holsnøy and also in the WGR, occurred in discrete structures such as pseudotachylites and cracks accompanied by fluids that are markers of a brittle deformation of the granulite and probably seismic activity [*Austrheim*, 1987; *Austrheim and Boundy*, 1994; *Austrheim et al.*, 1996a; *Jackson et al.*, 2004; *John et al.*, 2009; *Lund and Austrheim*, 2003]. Strain was subsequently localized in the eclogitized parts of the massif, and eclogite-facies shear bands developed, forming a breccia-like structure with meter-scale blocks of granulite in a foliated eclogite-facies matrix. The granulite-eclogite transition remained incomplete because of the lack of fluids [*Austrheim*, 1987] or deformation gradients [*Bjørnerud et al.*, 2002]. Similar observations have also been made in the WGR, where large portions of Proterozoic igneous protoliths or granulites remained unreacted [*Austrheim*, 1987; *Krabbendam et al.*, 2000]. The overall finite strain is asymmetric and suggests a component of top-to-the-NE shear [*Raimbourg et al.*, 2005]. The eclogite-facies shear zones were clearly a weaker phase than the granulites [*Austrheim*, 1991; *Jolivet et al.*, 2005]. This field observation is confirmed by experimental studies: effective viscosities calculated from rheological parameters for dislocation creep of pure synthetic omphacitite, jadeitite, or wet diopsidite at 700°C, and for strain rates of 10^{-14} s^{-1} range between $10^{21.3}$ and $10^{22.2}$ Pa.s (Table 3 and references therein). Experimental parameters for an eclogite with 50% garnet, 40% omphacite, and 10% quartz yield an effective viscosity of $10^{22.5}$ Pa.s [*Jin et al.*, 2001]. This value can be considered as a maximum estimate for the Holsnøy eclogite viscosity in their peak condition, since garnet, considered as responsible for strengthening of the eclogites in the dislocation creep regime [*Jin et al.*, 2001] represents only 10%–40% of these rocks (Table 2). White micas, representing 4%–6% of the Holsnøy eclogites must also lower their effective viscosity. The effective viscosity of dry anorthitite in the same conditions is $10^{22.7}$ Pa.s and can be considered as a minimum value for Holsnøy granulites, containing ~20% of garnet. The viscosity contrast between granulites and eclogites can be amplified by 1 order of magnitude when strain is localized in the eclogite shear bands, where strain rate estimates range between 10^{-13} and 10^{-12} s^{-1} [*Terry and Heidelbach*, 2006]. Mechanical study of eclogitization [*Raimbourg et al.*, 2007b] shows that the drop in effective strength for crust undergoing eclogitization is early and abrupt: when the eclogite-facies shear bands connect into a network isolating boudins of preserved granulite. The effect on density is, however, proportional to the bulk-eclogitized fraction of the protolith and therefore progressive from 2750 to 3100 $kg \cdot m^{-3}$ for gabbroic anorthosites equilibrated in the granulite facies up to 3060–3480 $kg \cdot m^{-3}$ in the eclogite facies (Table 2). The partial eclogitization of the Holsnøy granulite must have lowered its resistance, whereas the bulk density of the crust, close to 3100 $kg \cdot m^{-3}$ for a gabbroic anorthosite with 50% granulitic minerals and 50% eclogitic minerals, remained lower than mantle values. Eclogitization of the lower crust is therefore a possible mechanism for decoupling the crust and the mantle when incomplete and responsible for densification of the lower crust when completed. The partially eclogitized granulites of Holsnøy, exhumed from the eclogite facies without penetrative retrogression, could therefore be preserved crustal slivers equivalent to the ILC now underplated at the base of the Himalayan wedge. P-T conditions, calculated for the Holsnøy eclogites (650°C –750°C at 1.5–2.1 GPa after *Glodny et al.* [2008, Supplementary Table]), are comparable to the modeled thermal state of the base of the Tibetan crust [*Hetényi et al.*, 2007]. The ~2 Ma time lapse for fluid-granulite reaction in the eclogite facies deduced from petrological studies in Holsnøy is also compatible with the offset between the entry of the ILC into the eclogite facies stability field and its actual densification (Figure 5b).

5.3. Delamination of Crust During Orogeny

[23] Models of development of crustal-scale thrust zones responsible for slicing of subducting lithosphere and exhuming of large portions of the crust, such as the HHC or the CMA, require a weak rheological level in the lower crust [*Brun and Faccenna*, 2008; *Chemenda et al.*, 1995]. The examination of the thermal structure of India and Baltica prior to their involvement in the collision can tell us whether they presented such a decoupling level or not. The latest contractional episode recorded in Baltica was the

Sveconorwegian orogeny [*Gorbatschev*, 1985], ~400 Ma before the Scandian collision. In India, a late Pan-African imprint [*Gaetani and Garzanti*, 1991] was also 400 Ma prior to the Himalayan collision. The formation of their passive margins was the last thermal event experienced by the two shields. It is 660–600 Ma in Baltica [*Torsvik et al.*, 1996] and Permian in India [*Gaetani and Garzanti*, 1991]; in both cases the time delay between rifting and collision is ~200 Ma. In both orogens, therefore, the lower plates had similar thermal histories and presumably strengths at the time of collision. Both collisions were initiated at high convergence rates, i.e., in the order of 5–10 cm/yr [*Torsvik et al.*, 1996; *Klootwijk et al.*, 1992]. The present-day rheological stratification of India, estimated far south from the Himalayan front, may be a good model of the strength of the shields prior to collision. The dynamic compensation of the Himalayan topography implies a strong mantle fully coupled to the Indian shield crust [*Cattin et al.*, 2001; *Hetényi et al.*, 2006]. The decrease in EET, due to intralithospheric decoupling, only happens related to the collision wedge, beneath the Tethyan Himalaya [*Hetényi et al.*, 2006]. The Baltica and Indian lithospheres were thus initially strong without a preexisting decoupling level. A weakening process must therefore have happened within the continental wedge in order to allow the decoupling of the upper and lower parts of the lithosphere. (1) Progressive heating of buried material can lower viscosity and trigger exhumation of buoyant portions of the crust [*Warren et al.*, 2008; *Beaumont et al.*, 2009]. (2) Partial melting of the lower crust can also promote the exhumation of dense and strong lenses of eclogites [*Labrousse et al.*, 2002]. (3) Eclogitization, as summarized above, can be a weakening mechanism, able to promote the exhumation of rocks when incomplete [*Austrheim*, 1991; *Raimbourg et al.*, 2007b]. Of these three weakening mechanisms, both partial melting and eclogitization can localize deformation in the lower crust and exhume coherent upper crust panels. Anatectic granites are present in the Caledonides and Himalaya and formed during partial melting contemporaneous with the development of the main crustal-scale thrusts and with amphibolite- or granulite-facies retrogression of the HP rocks. Even if partial melting might be one of the weakening mechanisms promoting the exhumation of HP rocks, the first stage of decoupling must happen at depths where eclogite is stable in order to explain their own exhumation. The observed synchronicity between the development of eclogites at depth and the activation of the crustal-scale thrusts responsible for exhuming the eclogite-bearing lower crustal slices is a key observation supporting the idea that eclogitization is the first weakening phenomenon responsible for the strength loss in the lower crust and the decoupling of the crust and mantle [*Jolivet et al.*, 2005]. Eclogitization in the Bergen Arc, representing the most distal and buried parts of the CMA allowed the initiation of the JBT at 430 Ma. The Ama Drime eclogites crystallized in the gneisses of the Ama Drime unit 21 Myr at the same depth where the ILC is now underplated and initiated the development of the MCT. Their exhumation within the HHC gneisses and their granulite-facies overprint obliterated the record of their deformation in the eclogite facies [*Kali et al.*, 2010].

6. Conclusion

[24] Comparison of the early Palaeozoic Scandinavian Caledonides and the Cenozoic Himalayan orogens reveals a close relationship between HP metamorphism and the development of crustal-scale thrust zones responsible for thickening the crustal wedge. Both orogens have eclogites derived from continental material in the hanging wall of their main crustal thrusts, i.e., the MCT in the Himalaya and the JBT in the Scandinavian Caledonides. The crystallization ages of those eclogites are equivalent to the activation ages of the main large-scale thrusts.

[25] Field evidence, offered by the collapsed, extended, and eroded Caledonian orogen in the Bergen Arc and the WGR, shows that eclogitization can account for the two main changes in rock properties seen at depth in the Himalaya by geophysics. Receiver-function analysis, gravity anomalies, and EET estimates indeed suggest crust-mantle decoupling and a subsequent ILC density increase 200–350 km north of the MFT [*Cattin et al.*, 2001; *Hetényi et al.*, 2006, 2007]. Field study of Baltica lower crust now exposed on Holsnøy, in the Bergen Arc, shows that eclogites are weak compared with granulites or peridotites. Weakening of the crust occurs before complete eclogitization and coeval densification, when the eclogites form an interconnected network of veins and shear bands, so that decoupling of the crust and mantle is possible and exhumation of the still buoyant crust is transiently favored by eclogitization. The first step of eclogitization at Holsnøy may have been accompanied by seismicity. This same process may be at work beneath the Himalaya where seismicity occurs in the zone of inferred eclogitization. The progressive decoupling of the continental crust and the lithospheric mantle occurs during eclogitization reactions promoted by fluids and deformation. Retrograde metamorphism and partial melting can further weaken the crustal material during its exhumation. Thermomechanical modeling of collision zones with density and viscosity changes of the rocks governed by metamorphic reactions has to be performed to further constrain the role of eclogitization in the localization of deformation at the crustal scale.

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