Young eclogite from the Greater Himalayan Sequence, Arun Valley, eastern Nepal: P–T–t path and tectonic implications

S.L. Corrie, M.J. Kohn, J.D. Vervoort

Abstract

Garnet geochronology was used to provide the first direct measurement of the timing of eclogitization in the central Himalaya. Lu–Hf dates from garnet separates in one relict eclogite from the Arun River Valley in eastern Nepal indicate an age of 20.7 ± 0.4 Ma, significantly younger than ultra-high pressure eclogites from the western Himalaya, reflecting either different origins or substantial time lags in tectonics along strike. Four proximal garnet amphibolites from structurally lower horizons are 14–15 Ma, similar to post-eclogitization ages published for rocks along strike in southern Tibet. P–T calculations indicate three metamorphic episodes for the eclogite: i) eclogite-facies metamorphism at ~670 °C and ≥15 kbar at 23–16 Ma; ii) a peak-T granulite event at ~780 °C and 12 kbar; and iii) late-stage amphibolite-facies metamorphism at ~675 °C and 6 kbar at ~14 Ma. The garnet amphibolites were metamorphosed at ~660 °C. Three models are considered to explain the observed P–T–t evolution. The first assumes that the Main Himalayan Thrust (basal thrust of the Himalayan thrust system) cuts deeper at Arun than elsewhere. While conceptually the simplest, this model has difficulty explaining both the granulite-facies overprint and the pulse of exhumation between 25 and 14 Ma. A second model assumes that (aborted) subduction, slab breakoff, and ascent of India's leading edge occurred diachronously: ~50 Ma in the western Himalaya, ~25 Ma in the central Himalaya of Nepal, and presumably later in the eastern Himalaya. This model explains the P–T–t path, particularly heating during initial exhumation, but implies significant along-strike diachronity, which is generally lacking in other features of the Himalaya. A third model assumes repeated loss of mantle lithosphere, first by slab breakoff at ~50 Ma, and again by delamination at ~25 Ma; this model explains the P–T–t path, but requires geographically restricted tectonic behavior at Arun. The P–T–t history of the Arun eclogites may imply a change in the physical state of the Himalayan metamorphic wedge at 16–25 Ma, ultimately giving rise to the Main Central Thrust by 15–16 Ma.

1. Introduction

Although eclogites have been found in only a few localities in the Himalaya, they are key in understanding the orogen's metamorphic and tectonic evolution (Fig. 1). In the northwest Himalaya, ultra-high pressure eclogites (UHP) have been documented in the Kaghan Valley of Pakistan (Pognante and Spencer, 1991; Tonarini et al., 1993) and the Tso Morari dome in the Ladakh region of India (de Sigoyer et al., 1997, 2000). In the central Himalaya, eclogites have been documented in the Ama Drime range in the Kharta region of Tibet (Lombardo and Rolfo, 2000; Groppo et al., 2007; Cottle et al., 2009), and the Makalu- Everest region of the Arun River Valley in eastern Nepal (Parkinson and Kohn, 2002). Both NW Himalayan eclogite localities occur within the Greater Himalayan Sequence immediately south of the Indus Suture Zone (O'Brien et al., 2001), and are considered to indicate subduction of the leading edge of the Indian continental crust during the initial stages of Indo–Asian collision. These eclogites reached conditions of 725 ± 25 °C and 28–30 kbar (Mukherjee and Sachan, 2001; O'Brien et al., 2001). Accessory zircon (Kaneko et al., 2003; Parrish et al., 2006) and garnet (Tonarini et al., 1993), whose chemistry or texture link them to UHP metamorphism, have been directly dated at ~46 Ma in the Kaghan Valley, and a garnet isochron yields ~54 Ma for Tso Morari (de Sigoyer et al., 2000).

In contrast, eclogites in the central Himalaya do not appear to have attained UHP conditions (Lombardo and Rolfo, 2000; Groppo et al., 2007) and direct correlation with eclogites in the NW Himalaya is unclear. The Ama Drime eclogites occur north of the Arun eclogite, up-drainage along the Arun/Phung Chu River within the Ama Drime Massif. Both eclogites have been strongly overprinted by subsequent metamorphism, leaving sparse evidence for eclogitization. Groppo et al. (2007) proposed four phases of metamorphism for the Ama Drime eclogites on the basis of microstructural observations, pseudosection analysis, and conventional thermobarometry: an initial (M1) eclogitic metamorphism at >15 kbar and >580 °C; a peak-T
and 4 kbar. Attempts to date the Ama Drime eclogites have yielded U–SHRIMP ages of 13 Ma (Kohn et al., 2010). Few data distinguish rocks in one sequence from the other. In this study, we have followed closely the lithostratigraphic classification and mapping of LHS versus GHS in eastern Nepal and adjacent Tibet (e.g., Goscombe et al., 2006; Groppo et al., 2007; Cottle et al., 2009), and other than detrital zircon ages and c. 500 Ma felsic intrusions (DeCelles et al., 2001, 2002; Gehrels et al., 2003). The LHS consists of greenschist- to amphibolite-facies pelitic schists, quartzites, marbles, and calc-silicates, with interbedded orthogneisses, felsic metavolcanic rocks, and metabasalts in the lower stratigraphic levels. Older and younger LHS rocks are separated by a major unconformity contemporary with Pan-African tectonics (Valdiya, 1995). The lower, older formations are interpreted as proximal to a Paleoproterozoic (c. 1830 Ma) continental arc that bordered the northern margin of the Indian plate, the leading edge of which is now subducted beneath the Himalaya and Tibet (Kohn et al., 2010). Above the unconformity are Permian to Paleocene sedimentary rocks of Gondwanan affinity (Sakai, 1983).

Initial movement on the MCT has been generally viewed as having occurred at 20–22 Ma based on hornblende ⁴⁰Ar/³⁹Ar and monazite U–Pb ages (Hubbard and Harrison, 1989; Hodges et al., 1996; Johnson et al., 2001). However, more recent work dating monazite from leucogranites and migmatites in the MCT thrust sheet has suggested initial movement on the MCT at 16±1 Ma and melt crystallization by 20–22 Ma (DeCelles et al., 2001, 2002; Robinson et al., 2003, 2006; Kohn et al., 2004, 2005). The displacement on the thrust is estimated to be 125 km (DeCelles et al., 2001, 2002; Robinson et al., 2003, 2006; Kohn et al., 2004), with a displacement rate of about 2 cm/year between 16 and 10.5 Ma (Kohn et al., 2004).

Samples from both the LHS and GHS were collected in eastern Nepal along the west side of the Arun River (Fig. 2). The metabasites occur as decimeter- to meter-scale boudinaged sills and small bodies within the surrounding felsic schists and gneisses, much as documented by Cottle et al. (2009) along strike in southern Tibet. Amphibolite samples were collected from LHS (meta)basaltic sills that are stratigraphically continuous with the surrounding schists and gneisses. The granulitized eclogites from the GHS Barun Gneiss were collected from the core of a 3 meter diameter boudinaged sill.

Fig. 2. STDS = South Tibetan Detachment System; MCT = Main Central Thrust; MBT = Main Boundary Thrust.

2. Geologic setting

In general, laterally extensive, crystalline thrust sheets of the Nepalese Himalaya can be grouped into two main lithotectonic units: the Greater Himalayan Sequence (GHS) and the Lesser Himalayan Sequence (LHS), bounded above by the South Tibetan Detachment System (STDS), and below by the Main Boundary Thrust (MBT). There is little agreement in the timing of eclogitization in the central Himalaya, as garnet growth would have been initiated during the prograde transition to eclogite facies. The timing of this event is then interpreted in the context of models of Indo–Asian collision and Himalayan tectonics.


Sample descriptions, mineral compositions, and supplemental X-ray maps are located in the Data Repository.

3. Analytical methods

Elemental compositions and X-ray maps were collected using the Cameca SX-100 electron microprobe housed in the Department of Earth and Environmental Sciences at Rensselaer Polytechnic Institute, Troy, New York. Natural and synthetic silicates and oxides were used for calibrations, and quantitative measurements were made using an accelerating voltage of 15 kV and a current of 20 nA (major silicates) or 100 nA (rutile and titanite). A minimum beam size was used on most minerals, except plagioclase and micas (10 μm), and rutile and titanite (2 μm). Peak count times were 10 s (Na, Ca, Fe, Mn, Si, Al), 20 s (Mg, Ti, K), 45 s (Zr in rutile), and 75 s (Zr in titanite). Operating conditions for the X-ray maps consisted of an accelerating voltage of 15 kV, current of 200 nA, pixel time of 30 ms, and step size of 2–5 μm/pixel.

Understanding the distribution of Lu in garnet is important for interpreting Lu–Hf isochrons within the context of progressive metamorphism. Therefore, trace element profiles were collected across eclogite garnets using a UP193 laser and Element XA magnetic sector ICP-MS housed at the University of Lausanne, Switzerland. Prior to data collection, line traverses were pre-ablated with a spot size of 25 μm, a stage movement rate of 50 μm/s, and a power intensity of 6.2 J/cm², as metered directly at the surface. Traverse data were then collected with a spot size of 15 μm, a stage movement rate of 10 μm/s and a power intensity of 6.2 J/cm². Intensities were measured in low resolution mode for a large suite of major and trace elements ranging from Mg to Hf, the most important issue being the distribution of Lu. Raw counts of 175Lu were normalized to 27Al, and standardized against NIST612 glass. For sample AR01-43c, concentrations of Lu were calculated from averages measured across the core vs. rim. Whole-rock samples were crushed to <750 μm. Approximately 0.25 to 1 g of garnet was handpicked under a binocular microscope, excluding grains with visible inclusions like rutile and zircon. Sample dissolution, chemical separations, and isotopic analysis protocols are described by Vervoort et al. (2004) and Cheng et al. (2008). For each geochronological point, ∼0.25 g of garnet or whole-rock powder was digested. Two aliquots of whole-rock powders were digested per sample. The first followed the same tabletop dissolution procedure as that of garnet to determine the zircon-free whole-rock composition. The second used high pressure Teflon bombs to ensure total dissolution of refractory phases including zircon. Isotopic data were collected on a ThermoFinnigan Neptune™ multi-collector inductively coupled plasma mass spectrometer (MC-ICP-MS) in the GeoAnalytical Laboratory at Washington State University. Protocols for Lu analysis and data reduction are described by Vervoort et al. (2004). Hafnium ratios were corrected for mass fractionation using 179Hf/177Hf = 0.7325 and the exponential law for mass bias correction. Isotopic compositions were then normalized using an external Hf standard, JMC475 (176Hf/177Hf = 0.282160). The average [176Hf/177Hf] for JMC475 taken...
over the course of this study was $0.282142 \pm 0.000025$ (2σ, n = 54). External uncertainties applied to measured data are 0.5% for $^{176}$Lu/$^{177}$Hf and the quadratically combined 2-sigma in-run error and a blanket 0.005% uncertainty for $^{187}$Hf/$^{177}$Hf. Lutetium–Hafnium ages were calculated using the $^{176}$Lu decay constant of $1.867 \times 10^{-11}$ (Scherer et al., 2001; Söderlund et al., 2004) and isochrons were produced using the program Isoplot/Ex (Ludwig, 2003). Errors are reported at 95% confidence.

4. Petrology

Samples AR01-43c and AR01-43e from this study are similar to the granulitized eclogites described previously from the Ama Drime massif (Lombardo and Rolfo, 2000; Groppo et al., 2007; Cottle et al., 2009). These undeformed, medium-grained rocks contain garnet, clinopyroxene, plagioclase, amphibole, quartz, trace orthopyroxene (in AR01-43c only), biotite, scapolite, calcite, ilmenite, titanite, zircon, and apatite. The high-calcium bulk composition of these rocks stabilizes titanite as the primary titanium-bearing phase rather than rutile, which is typically assumed to be the higher-P phase. Direct evidence for eclogitization was all but eliminated during subsequent metamorphic events. Calcium-rich (~30% grossular) cores of large garnet are the only direct record of the prograde eclogitic assemblage, and were overgrown by thin rims during the later thermal peak of metamorphism (Fig. 3). True omphacite is not preserved in the matrix or as inclusions in garnet, although sparse inclusions of relatively sodic clinopyroxene (up to 15% jadeite) are preserved in garnet (Fig. 4). However, these pyroxene inclusions may be prograde, rather than peak eclogitic pyroxene (Groppo et al., 2007). In the matrix, presumed omphacite has been entirely replaced by a plagioclase + Na-poor clinopyroxene symplectite that has been documented in other retrogressed eclogites as the result of the transition from eclogite to granulite facies (Lombardo and Rolfo, 2000; Groppo et al., 2007; Liu et al., 2007; Cottle et al., 2009). Similar to Cottle et al. (2009), no pseudomorphs after phengite, i.e., symplectites of high-Fe biotite and plagioclase as described by Groppo et al. (2007), were observed in any samples.

The thermal peak metamorphic assemblage is characterized by relatively coarse-grained diopsidic clinopyroxene (Fig. 5), plagioclase, garnet, quartz, and trace orthopyroxene. A final thermal event is recorded in the retrograded eclogite by the almost total replacement of orthopyroxene by brown amphibole, which also grew in the clinopyroxene-plagioclase symplectite.

The LHS mafic amphibolites (AR01-2b, -11b, -35b, and -41) generally have a mineralogy consisting of quartz + amphibole + garnet + biotite + ilmenite ± clinopyroxene ± plagioclase ± rutile or titanite. Regionally, different amphibolites exhibit a range of mineral assemblages (Goscombe and Hand, 2000), probably reflecting basaltic to hydrothermally altered volcanic protoliths. There is no petrographic or textural evidence that these rocks ever experienced eclogite-facies conditions. The samples contain no evidence for omphacite (no pseudomorphs, symplectites, or inclusions in garnet), and garnets exhibit simple growth zoning of major elements (see Data Repository Fig. 1).

5. Thermobarometry

While no direct barometric measurements can be made for eclogite-facies conditions, an estimate for the minimum pressures experienced by the rocks can be obtained by reintegrating sodic plagioclase into pyroxene (e.g., Will and Schmädicke, 2001). This integration assumes: 1) at eclogite facies, all sodium was hosted in clinopyroxene; and 2) the albite and anorthite components of plagioclase texturally associated with matrix pyroxene (either in the symplectites or surrounding large diopсидic clinopyroxenes; Fig. 5) were hosted in omphacite as the jadeite and Ca-Tschermaks

Fig. 3. Ca X-ray map of garnet from sample AR01-43c with sketch showing core–rim relationship. Garnets exhibit broad cores with a relatively high Ca content ($X_{\text{Grs}} \geq 0.28$) and lower-Ca rims ($X_{\text{Grs}} \leq 0.23$). Lutetium concentrations are in ppm and show a slight enrichment of Lu in the overgrowth rims relative to the cores. Also note that matrix plagioclase is zoned from more sodic cores to more calcic rims. Scale bar is 500 μm. Grt = garnet, Hbl = hornblende, Pl = plagioclase, Scp = scapolite.

Fig. 4. (Wollastonite + enstatite + ferrosilite) − jadeite − aegirine pyroxene compositional diagram of Morimoto et al. (1988). Matrix clinopyroxene is Na-poor (diopсидic). Clinopyroxene inclusions in garnet contain up to 15 mol% jadeite and plot close to the omphacite field. Re-integrated pyroxenes plot well within the omphacite field. $n = 57$. 

Fig. 5. (Wollastonite + enstatite + ferrosilite) − jadeite − aegirine pyroxene compositional diagram of Morimoto et al. (1988). Matrix clinopyroxene is Na-poor (diopсидic). Clinopyroxene inclusions in garnet contain up to 15 mol% jadeite and plot close to the omphacite field. Re-integrated pyroxenes plot well within the omphacite field. $n = 57$. 

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components, respectively. This does not account for sodium in matrix hornblende or calcium originally hosted by garnet, and provides only a minimum estimate of the jadeite content. Assuming an average composition for plagioclase and diopside as measured by electron microprobe, the number of moles of each species is calculated via their respective modal abundances and molar volume. In sample AR01-43e, plagioclase has an average composition of An$_{1.2}$Ab$_{39}$ and modal abundances are 23 and 16% for plagioclase and clinopyroxene (estimated via ~ 300 point counts per sample). This reintegrates to an omphacitic pyroxene with a jadeite content of 39.1 mol%, which corresponds to a minimum pressure of 15 kbar at 670 °C (Fig. 4; Holland, 1980), with temperature estimated via Zr-in-titanite thermometry (see below). The same calculation for AR01-43c (An$_{20}$Ab$_{59}$; 21% plagioclase, 12.5% clinopyroxene) returns a jadeite component of 30.9 mol%, corresponding to 14.1 kbar at 670 °C. Calculations based on small regions of symplectite yield greater scatter, but comparable results.

Pressure–temperature conditions for the post-eclogite metamorphic events were calculated using the garnet–clinopyroxene (Pattison and Newton, 1989) and garnet–hornblende (Graham and Powell, 1984) thermometers and the garnet–plagioclase–clinopyroxene–quartz (Eckert et al., 1991) and garnet–plagioclase–hornblende–quartz (Kohn and Spear, 1990) barometers (Fig. 6A). Compositions for each calculation were selected based on spatial and textural relationships among the individual components; i.e., symplectitic and coronal plagioclase was chosen for garnet–plagioclase–clinopyroxene–quartz barometry and matrix plagioclase associated with hornblende for garnet–plagioclase–hornblende–quartz barometry. Intersection of the garnet–clinopyroxene and the garnet–plagioclase–clinopyroxene–quartz equilibria indicate pressure–temperature conditions of about 780 °C and 12 kbar for the granulite-facies event. A later amphibolite-facies overprint is recorded at ~765 °C, ~6 kbar by garnet–hornblende (Graham and Powell, 1984) and garnet–plagioclase–hornblende–quartz (Kohn and Spear, 1990) equilibria. (B) Pressure–temperature plot showing Zr-in-titanite equilibrium lines plotted with the barometers Jd–Ab–Qtz (Holland, 1980, calculated from the re-integration of plagioclase into clinopyroxene), Grt–Pl–Cpx–Qtz and Grt–Pl–Hbl–Qtz from A. Small black arrows indicate minimum pressure estimate for Jd–Ab–Qtz equilibria; gray and black boxes represent average P–T conditions from A; small white boxes denote points along the P–T path determined by combining Zr thermometry with conventional barometry; and large gray arrow indicates general P–T trajectory.

Zirconium-in-titanite thermometry (Hayden et al., 2008) was used to calculate temperatures for retrograded eclogites AR01-43c and -43e. Back-scattered electron images of titanite show no intracrystalline chemical zoning, although different grains have different compositions. Zircon is evident in all samples, either as observable matrix grains, or in mineral separates, or by relatively high Hf contents of bombed whole-rock powders. Rutile is not present in samples where Zr-in-titanite thermometry was applied, and the activity of TiO$_2$ was estimated to be 0.75 based on equilibrium among garnet, plagioclase, titanite, and quartz. Variations in $\alpha$(TiO$_2$) of ± 0.1 change apparent temperatures by 7 °C. Temperatures were calculated for matrix grains at pressures previously obtained via conventional thermobarometry for both the granulite and amphibolite metamorphic events, about 12 and 6 kbar, respectively (Table 1). At 12 kbar, core and rim temperatures of matrix titanite closely bracketed the 780 °C calculated via garnet–clinopyroxene thermometry (Fig. 6B). Yet, at 6 kbar, temperatures were ~710 °C, 35 °C higher than garnet–hornblende temperatures. It is possible titanite growth ceased prior to final growth and equilibration of hornblende.
Zirconium concentrations were measured from titanite inclusions in large matrix pyroxenes from sample AR01-43e (Table 1; Fig. 5B). The resultant Zr-in-titanite temperatures (Hayden et al., 2008) were 35–130 °C lower than matrix temperatures calculated for the same pressure (e.g., 650–745 °C vs. ∼780 °C at 12 kbar), although apparent temperatures generally increase towards the pyroxene rim. A minimum pressure estimate for the lowest temperature titanite inclusion from a pyroxene core (670 °C), which presumably pre-dates the granulite event, is 15 kbar (Fig. 6B; Holland, 1980), well within the eclogite facies. This is consistent with initial formation of titanite in the eclogite facies, with continued growth or recrystallization as the system began to exhume and heat from the eclogite facies toward the peak-T granulite event.

The Zr-in-rutile thermometer (Watson et al., 2006) was also used on amphibolite samples AR01-35b and -41 and the temperatures calculated were approximately 660 °C, similar to temperatures calculated for the amphibolite-facies overprint in the eclogites. The Zr-in-rutile thermometer is only weakly pressure dependent (Tomkins et al., 2007), and as these two samples do not contain plagioclase, an independent evaluation of pressure was not possible.

### Table 1
Zr-in-rutile and Zr-in-titanite results.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Zr in Rt</th>
<th>Zr in Ttna*</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>35b matrix</td>
<td>41 matrix</td>
</tr>
<tr>
<td># of analyses</td>
<td>5</td>
<td>8</td>
</tr>
<tr>
<td>ppm Zr (±1σ)</td>
<td>367±25</td>
<td>355±20</td>
</tr>
<tr>
<td>T °C (±1σ)</td>
<td>659±6</td>
<td>656±5</td>
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</table>

*Activity of TiO₂ assumed to be 0.75.

### Table 2
Lu-Hf isotopic results for garnet and whole rock separates from eclogite and amphibolites.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Fraction</th>
<th>Lu (ppm)</th>
<th>Hf (ppm)</th>
<th>176Lu/177Hf</th>
<th>176Hf/177Hf</th>
<th>Initial 176Hf/177Hf</th>
<th>Isochron age (Ma)</th>
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<tbody>
<tr>
<td>AR01-2b</td>
<td>grt1</td>
<td>1.60</td>
<td>0.162</td>
<td>1.400</td>
<td>0.283291±13</td>
<td>0.282890±15</td>
<td>15.1±0.7</td>
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<tr>
<td></td>
<td>grt2</td>
<td>1.60</td>
<td>0.142</td>
<td>1.593</td>
<td>0.283327±13</td>
<td>0.282894±16</td>
<td>14.1±0.7</td>
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<tr>
<td></td>
<td>WR</td>
<td>0.160</td>
<td>0.799</td>
<td>0.0291</td>
<td>0.282869±06</td>
<td>0.282854±06</td>
<td>6.1±0.7</td>
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<td>WR'</td>
<td>0.158</td>
<td>1.97</td>
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<td>0.282854±06</td>
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<td>grt1</td>
<td>1.87</td>
<td>0.115</td>
<td>2.312</td>
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<td>0.282480±11</td>
<td>14.1±0.4</td>
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<td>grt2</td>
<td>1.95</td>
<td>0.114</td>
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<td>0.269</td>
<td>0.591</td>
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<td>0.282869±06</td>
<td>14.1±0.4</td>
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<td>WR'</td>
<td>0.284</td>
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<td>0.282869±06</td>
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<td>AR01-35b</td>
<td>grt1</td>
<td>1.54</td>
<td>0.155</td>
<td>1.412</td>
<td>0.282635±14</td>
<td>0.282480±48</td>
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<td>grt3</td>
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<td>0.182</td>
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<td>0.282252±06</td>
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<td>WR'</td>
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<td>3.38</td>
<td>0.0107</td>
<td>0.282263±10</td>
<td>0.282252±06</td>
<td>13.9±2.5</td>
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<td>grt1</td>
<td>2.33</td>
<td>0.147</td>
<td>2.244</td>
<td>0.282865±17</td>
<td>0.282270±86</td>
<td>14.5±2.7</td>
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<td>grt2</td>
<td>2.23</td>
<td>0.204</td>
<td>1.550</td>
<td>0.282716±11</td>
<td>0.282270±86</td>
<td>14.5±2.7</td>
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WR, whole rock by savillex digestion; WR’, whole rock by bomb digestion. Reported errors are 2σ.

**Data not used in regression.**

*Calculated from isochron.

### 6. Garnet ages

Garnet–whole-rock isochrons yield ages ranging from 13.9 to 15.1 Ma for amphibolite samples, and 20.7±0.4 Ma for a retrograded eclogite (Table 2, Fig. 7). Analysis of bomb-digested whole rock revealed significantly lower Lu–Hf ratios and higher Hf concentrations compared to tabletop digestions due to dissolution in the former of zircon. Such data were omitted from calculated isochrons if the 176Hf/177Hf of the bomb-digested whole rock differed from the tabletop-digested whole rock by more than 1.5×10⁻⁵, presuming the zircon in these samples contained an inherited component not in equilibrium with garnet growth. If used in the regression, these data cause the calculated age to be spuriously old (e.g., Scherer et al., 2000). Models of Lu uptake in garnet (Lapen et al., 2003; Kohn, 2009) imply that garnet Lu–Hf ages should reflect early growth of garnet. Garnet X-ray maps from AR01-43c indicate that garnet cores represent ∼75% of total garnet volume, but rims are enriched in Lu by a factor of ∼1.4 compared to cores (Fig. 3). Thus, 60–70% of the bulk garnet Lu–Hf age reflects eclogite-facies garnet growth, while 30–40% reflects growth in granulate facies. Given the...
chemical simplicity of the amphibolite-facies LHS garnets in the underlying MCT footwall, their Lu–Hf ages are interpreted simply to reflect later amphibolite-facies metamorphism.

7. Discussion

Three main metamorphic stages are recognized in the retrograded eclogite samples from eastern Nepal. Pressure–temperature conditions of primary eclogite-facies metamorphism are loosely constrained at 670 °C and > 15 kbar (≥ 50 km depth) by albite–jadeite–quartz barometry and Zr-in-titanite thermometry. These are similar to conditions inferred by Lombardo and Rolfo (2000) and Groppo et al. (2007) for rocks along strike in Tibet. The eclogite was then strongly overprinted by a peak-temperature, granulite-facies metamorphism at ∼ 780 °C and 12 kbar, recorded by garnet–plagioclase–clinopyroxene equilibria and Zr-in-titanite thermometry. This was followed by a later, lower pressure and temperature event at ∼ 675 °C and 6 kbar, represented by the widespread growth of amphibole in the matrix.

Fig. 7. Lu–Hf isochron plots for eclogite and amphibolites. An age of ∼ 21 Ma is calculated for the GHS Arun eclogite, and LHS amphibolite ages range from 14 to 15 Ma. All errors are at 95% confidence.

Fig. 8. Schematic illustrations of potential tectonic models for the P–T–t evolution of Arun eclogites showing overall thrust geometries, implied thermal structure, and migmatite zone (T ≥ 700 °C, gray shading). MCT = Main Central Thrust. Gray ellipse = GHS rocks from Arun; Gray squares = GHS rocks elsewhere. Model 1: steady-state, critical taper model for the Himalaya (based on Henry et al., 1997 and Bollinger et al., 2006). At Arun, the MCT may have transported material further (model 1a) or cut deeper (model 1b) than elsewhere in the orogen. This model does not explain the granulite-facies overprint at Arun. Model 2: model of slab breakoff (model 2a; after Kohn and Parkinson, 2002) or delamination (model 2b; after Chemenda et al., 2000). Detachment and removal of the down-going plate or mantle lithosphere led to asthenospheric upwelling and buoyant extrusion of a sliver of Indian crust, causing rapid exhumation of the GHS and the eclogites. Both models explain granulite-facies overprint. Dark gray shaded area = Indian continental crust. Model 3: channel flow model (based on model LHO-2 from Beaumont et al., 2006) in which weak areas in the lower plate detach, transfer to the upper plate, and exhume. This model may explain granulite-facies overprint. Stippled area = strong lower plate; W = weak portion of lower plate.
and Zr-in-rutile thermometry. These conditions produce a clockwise P–T path characterized by heating during initial exhumation (Fig. 8B). The general shape of the P–T path agrees well with Groppo et al. (2007), but specific P–T results differ in that no evidence was found for the low pressures (e.g., 4 kbar) they reported. Our estimated final pressures correspond well with conventional thermobarometric results of Goscombe and Hand (2000) in the region for metapelites and amphibolites.

The 20.7 ± 0.4 Ma age for the GHS eclogite is the first attempt to measure directly the age of eclogitization in the central Himalaya, although several attempts have been made to date the granulite metamorphism in the region. Cottle et al. (2009) ascribed ≤13 Ma monazite dates to peak granulite metamorphism and prograde anatexis. However, partial melting should consume, not produce, monazite (Pyle and Spear, 2003; Kohn et al., 2004; Kelsey et al., 2008), and these ages more likely represent post-peak cooling and crystallization, presumably related to exhumation, thrust transport, or both. Groppo et al. (2007) also interpreted zircon rim ages of 13–14 Ma to represent late-stage granulite-facies metamorphism, but provide little direct evidence for this interpretation and concede that the zircon ages could instead be linked to the amphibolite-facies event. Overall, we interpret the 13–14 Ma monazite and zircon ages to reflect post-granulite-facies overprinting associated with cooling during MCT transport. This interpretation agrees with the four 14–15 Ma prograde ages for garnet amphibolites from the LHS in the MCT footwall, and is consistent with relatively late initiation of the MCT along strike in Bhutan and central Nepal (Daniel et al. 2003; Kohn et al., 2004). Considering the weighted proportions of Lu in garnet cores (eclogite facies: 60%–70%) vs. rims (granulite facies: 30%–40%), and assuming that these components are mixed in these proportions in the garnet fractions, it implies an age of eclogitization of 23–26 Ma, and a latest possible granulite-facies age of 13–15 Ma.

The P–T–t evolution of the HP Arun and Ama Drime eclogites contrasts with the UHP eclogites from the northwest Himalaya in three fundamental ways: the Arun rocks were metamorphosed at much lower pressures; they show significant heating during initial exhumation; and they were metamorphosed 20–30 Myr later (Groppo et al., 2007; Cottle et al., 2009; this study). These differences potentially signal different tectonic processes. With respect to P–T paths, the nearly complete granulite-facies overprinting of the Arun eclogite may require either an extra heat source or a combination of slow erosion plus strongly diminished thrust transport in the mid-crust after loading. Rapid erosion alone should exhume rocks isothermally, whereas rapid thrusting on lower level thrusts should cool rocks isobarically. In principle, heating might also occur upon expulsion of relatively cooler lower crustal rocks into an inverted geotherm. The occurrence of rapid exhumation (~30 km in 7 Myr) with heating points to additional heat sources or an inverted geotherm, and we evaluate possible competing models within that context.

It is important to note that all models of Himalayan orogenesis predict eclogites in the deep-crust. Thus, their key question is not the genesis of ~25 Ma eclogites, but rather why they were exhumed and exposed at Arun and apparently not elsewhere. One possible model is that their appearance simply reflects deepening or longer transport of the MCT in the Arun area (Fig. 8, models 1a and 1b). This explanation is supported by relatively low pressures of eclogitization — as little as 50 km depth, not ≥90 km as evident in the northwest Himalaya. Peak metamorphic pressures near the base of the MCT in central Nepal are already documented at 12 kbar (Kohn, 2008), and slight changes to thrust geometry and dynamics could allow sampling of eclogite facies, rather than amphibolite-facies GHS rocks (Henry et al., 1997). The disparity in eclogite ages then reflects different exhumation processes — slab breakoff at ~50 Ma and “normal” Himalayan thrust transport at ~20 Ma. This model, however, does not adequately explain the granulite-facies overprint at Arun. Within the context of fold-thrust belts, exhumation between 25 and 14 Ma implies the eclogites were transferred to the hanging wall, but thrust transport in the hanging wall should cause cooling, not heating.

A second possible model is that (aborted) subduction, slab breakoff, and buoyant ascent of Indian crust occurred diachronously: ~50 Ma in the western Himalaya, ~25 Ma in the central Himalaya, and presumably even younger in the eastern Himalaya. By this model (Fig. 8, model 2a), mechanical coupling between the oceanic and Indian continental lithospheres subducted India beneath Tibet to eclogite-facies conditions. Decoupling and removal of the down-going oceanic plate would cause asthenospheric upwelling and buoyant extrusion of a sliver of Indian crust out of the mantle lithosphere, causing rapid exhumation of the GHS and its eclogites (Kohn and Parkinson, 2002). Increased heat flow due to mantle upwelling could have increased temperatures to the granulite facies during exhumation. Differences along strike in exhumation rate or thermal structure might have allowed eclogites in the northwest Himalaya to exhum isothermally. Subsequent thrusting and transport would have cooled the Arun eclogites as they were incorporated in the Himalayan fold-thrust belt. While this model does explain our data, there is little supporting evidence elsewhere in the Himalaya for such profound along-strike diachronity in fundamental tectonic processes.

A third, “hybrid” model recognizes the possibility of repeated loss of mantle lithosphere, possibly by different mechanisms (Chemenda et al., 2000). For example, in the northwest Himalaya, UHP eclogites may have been exhumed early by breakoff of the oceanic slab (O’Brien, 2001; Mahéo et al., 2002; Treloar et al., 2003; Parrish et al., 2006), whereas the Arun HP eclogites may have been exhumed later by delamination of Indian lithosphere (Fig. 8, model 2b). Like slab breakoff, delamination is expected to produce mantle upwelling, leading to a granulite-facies overprint. Again, we know of no other supporting evidence elsewhere along strike for this process, but the possibly unique occurrence of such young eclogites at Arun may require a geographically restricted process.

Regarding more specific thermal–mechanical models of thrust-belt development in the Himalaya, both channel flow (e.g., Beaumont et al., 2001, 2004) and critical taper (Robinson et al., 2006) have been proposed to explain the metamorphism of the GHS (Jamieson et al., 2004; Kohn, 2008), and could in principle accommodate the presence of eclogites. However, neither model explains some key features of the Arun rocks. Channel flow models predict a focused erosional front for the last c. 25 Myr (Beaumont et al., 2001; Jamieson et al., 2004). Yet the Arun rocks were exhumed by ~30 km between 14 and 25 Ma, apparently far from any erosional front. This admittedly local observation casts doubt on how channel flow models account for erosion distributions and hence their predictive power for the Arun rocks. Moreover, the upper Arun River Valley is already differentiated from other areas along strike in the Himalaya by local E–W extensional structures that formed ≤13 Ma (Jessup et al., 2008; Cottle et al., 2009). As described previously, critical taper has trouble explaining P–T paths that combine exhumation with heating. Major changes must be invoked for the mechanical behavior of the wedge (to accommodate such profound thinning) and thrust rates (to accommodate heating) between 25 and 14 Ma.

One way of reconciling both channel flow and critical taper models with the Arun eclogite P–T–t path postulates that as parts of the lower crust heat and weaken, they may be expelled to the mid-crust, e.g. as modeled by Beaumont et al. (2006, model LHO-2; Fig. 8, model 3). An inverted geotherm is required to produce heating during expulsion, but this is predicted both in steady-state and time-dependent models (e.g., Royden, 1993; Henry et al., 1997; Beaumont et al., 2001, 2004). Less clear are the effects of expulsion on thermal state and later structures, particularly given the young age of exhumation (14–25 Ma). In this regard, the P–T–t history of the Arun and Ama Drime eclogites may signal a change in the physical state of the Himalayan metamorphic wedge at 16–25 Ma, ultimately giving rise to the MCT by 15–16 Ma.
References


