Tso Morari coesite eclogite: pseudosection predictions v. the preserved record and implications for tectonometamorphic models

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Abstract: Ultrahigh-pressure eclogites of the Tso Morari area, NW Himalaya (Ladakh, India), have been intensively investigated petrographically and petrologically with surprisingly different results. Metamorphic subduction paths based on mineral isopleths in pressure–temperature pseudosections in some studies claim concave (to the temperature axis) pressure–temperature paths predicting significant Ca–Mg–Fe garnet growth in the lawsonite and glaucophane fields: a prediction at odds with abundant epidote/clinozoisite and sodic-calcic amphibole inclusions in garnet interiors more probable along a convex path. One study deduced strong heating still at high pressures and proposed a felsic diapir rising through the mantle wedge: an explanation strongly at odds with well-documented glaucophane cores to barroisite replacing matrix omphacite requiring a cold exhumation most likely back up the subduction channel. In addition, matrix magnesite rimmed by dolomite suggests pressures well into the coesite (if not diamond) stability field: something neglected in most studies. Despite the application of modern analytical and thermodynamic modelling tools, the peak conditions attained by Tso Morari ultrahigh-pressure rocks are often poorly deduced and at odds with simple observations. Is this problem perhaps hindering the reliable identification of new ultrahigh-pressure terranes?

The most spectacular testing ground for geological models of subduction–collision, plateau growth, erosion- or tectonically-driven exhumation, or even relief–climate interaction is the Himalayan–Tibetan region. The closure of the Neo-Tethys ocean and collision of the Indian and Asian continental plates, and intervening arcs, starting in Eocene times and continuing to the present day, has produced a >2500 km long mountain belt with Earth’s highest peaks and also a large high-elevation area, the Tibetan Plateau (e.g. Gansser 1964; Yin & Harrison 2000; Searle 2015). Several characteristic tectonometamorphic units have been identified running the length of the orogenic belt which aid in the subdivision but also give strong pointers to the magmatic and metamorphic evolution. These are, from north to south, the Transhimalayan Batholith, Indus–Tsangpo Suture Zone, the Higher Himalaya (comprising the Tibetan Tethys Zone and Central Crystallines), Lesser Himalaya and Subhimalaya (incorporating part of the Molasse and the Siwaliks) (Fig. 1). During Upper Cretaceous to Eocene northward subduction of the oceanic crust of Neo-Tethys, an Andean-type calcalkaline magmatic arc developed on the Asian margin the eroded core of which defines the Transhimalayan Batholith (Honegger et al. 1982; Wen et al. 2008). From east to west this is usually subdivided into the Gangdese, Ladakh and Kohistan Batholiths, respectively. The course of two major rivers, the Indus and Yarlung Tsangpo, follow the Indus-Tsangpo-Suture Zone (ITSZ) which contains thrust sheets of oceanic mafic, ultramafic and volcanioclastic rocks along with continental shelf and slope rocks all overlain by younger terrestrial conglomerates and sandstones of the Indus Molasse (e.g. Henderson et al. 2010). Blueschists (glaucophane-, lawsonite- and carpholite-bearing rocks) of Cretaceous age from the former accretionary prism occur as tectonic units within the IITSZ (Honegger et al. 1989; Oberhänsli 2013; Groppo et al. 2016). The rocks exposed south of the suture comprise Proterozoic, Paleozoic and Mesozoic rocks forming the Indian Plate that were initially deeply subducted then imbricated and stacked during metamorphism (e.g. Gansser 1964; Searle 2015). Indian Plate rocks incorporated early in the orogeny have undergone high-grade (kyanite- and anatectic sillimanite ± cordierite-bearing amphibole to granulite facies) metamorphism, mark the present-day high topography, and form the Higher (or Greater) Himalaya Crystalline (HHC) nappes. Crust incorporated later into the thrust stack, now tectonically below the HHC and separated by the Main Central Thrust, shows dominantly a much lower (greenschist facies or lower) grade of metamorphism (even in tectonic windows 100 km or more from orogenic front) and is
Fig. 1. Simplified geology of the Himalaya showing the distribution of the main units and locations of eclogites and other high-pressure rocks. The boundaries between the major units are tectonic.
generally known as the Lesser Himalaya (LH). In most of the orogen, the Phanerozoic cover of the plate became detached from the basement and, although folded and imbricated, underwent no high-grade metamorphism such that stratigraphy and fossils are preserved (e.g., Gaetani & Garzanti 1991). This Tibetan–Tethys Zone crops out in much of southern Tibet south of the ITSZ and displays a large difference in metamorphic grade to underlying HHC rocks emerging along a normal fault, the South Tibetan Detachment, which bound the two units. Tectonically below the LH, separated by the Main Boundary Thrust, Cenozoic marine and continental sedimentary rocks (Siwalik Group molasse) represent erosion products of the Himalaya, form the Subhimalaya (Burbank et al. 1996). Widening of the orogeny by southward propagation of the active main Himalayan thrust plane has led to these rocks being deformed and thrust southward along the Main Frontal Thrust onto the Indian Craton over the youngest foreland basin sediments.

Despite the general agreement over the overall structure of the present-day Himalaya – a structure outlined by Gansser (1964) and quantified by large-scale geophysical investigations (e.g. Zhao & Nelson 1993; Nelson et al. 1996; Jain et al. 2012; Caldwell et al. 2013) – the detailed evolution of the orogeny from subduction to collision and the interpretation of the rock record are still disputed. The discovery of coesite in Kaghan Valley (Pakistan) eclogite (O’Brien et al. 1999, 2001) and shortly after also in equivalent Tso Morari (NW India) eclogites (Mukherjee & Sachan 2001), required a significant modification to existing tectonometamorphic models for the Indian–Asia collision which were until then essentially based on crustal thickening. Understanding this magnitude, extent and duration of the additional component, namely deep crustal subduction, is critical to our interpretation of the transition between subduction (steep-angle) and collision/stacking (shallow-angle) processes and may allow development of models explaining the general mechanism that may help to interpret older, long equilibrated orogenic belts. Recent attempts to explain the pressure–temperature ($P$–$T$) evolution of the Tso Morari eclogites using advanced equilibrium thermodynamic techniques (St-Onge et al. 2013; Palin et al. 2014, 2017; Chatterjee & Jagoutz 2015) contrast with existing interpretations of these well-studied rocks. All three of these recent studies suggest a $P$–$T$ trajectory along a colder path than previously reported. Also, Chatterjee & Jagoutz (2015) take mineral compositions to indicate significant heating at high pressure interpreted as reflecting a diapiric rise of the eclogite-bearing units through the mantle wedge. These new findings and interpretations will be described and discussed in detail with the aim to test their validity and plausibility with respect to existing and newly presented microstructural and mineral chemical data.

**Metamorphic evolution of the Himalaya**

The metamorphic evolution of the Himalaya can be closely tied to the evolving stages of a classic subduction–collision orogeny that involves a transition from oceanic subduction to continental collision via a short-lived deep continental subduction (O’Brien 2001; Warren et al. 2008; Kohn 2014; Searle 2015). This is readily visible from the ages, relative locations and associated rocks of units containing the metamorphic indicator minerals carpholite, coesite, kyanite and cordierite (O’Brien 2018). Lawsonite-bearing blueschists (Groppo et al. 2016) and carpholite-bearing metasediments (Oberhansli 2013) in the former accretionary prism incorporated into the ITSZ of NW India and Pakistan record low-temperature subduction of Tethys oceanic crust below the Asian margin during Cretaceous times (Fig. 2, stage 1) i.e. at the same time as magmatism in the Transhimalayan Batholith (Honegger et al. 1982, 1989). Obducted ophiolites of this age exist in several parts of the ITSZ (Bédard et al. 2009). When Tethys was finally consumed, continental crust of the outermost margin of the Indian Plate was deeply subducted (Fig. 2, stage 2) as recorded in the coesite-bearing eclogites of the Kaghan/Neelum Valley in Pakistan and the Tso Morari areas of the NW Himalaya (de Sigoyer et al. 1997; Lombardo et al. 2000; O’Brien et al. 2001; Mukherjee & Sachan 2001; Wilke et al. 2010a). The eclogites represent metamorphosed equivalents of intrusive and extrusive Permian Panjal Trap basalts which are enclosed in Proterozoic basement and Paleozoic metasediments and orthogneissess (Papritz & Rey 1989; Greco & Spencer 1993; Girard & Bussy 1999; Rehman et al. 2016). Geochronological data from the eclogites point to deep subduction and substantial, extremely rapid exhumation all in Eocene (40–50 Ma) times (Tonarini et al. 1993; de Sigoyer et al. 2000; O’Brien & Sachan 2000; Kaneko et al. 2003; Schlup et al. 2003; Treloar et al. 2003; Parrish et al. 2006; Wilke et al. 2010b, 2012; Donaldson et al. 2013; Rehman et al. 2013, 2016). These rocks are located today directly adjacent to the ITSZ (Fig. 1) thus supporting exhumation along the subduction channel. Thermo-mechanical modelling has successfully reproduced the formation conditions, exhumation rate and tectonic location of such ultrahigh-pressure (UHP) eclogites assuming suitable input parameters (e.g. Warren et al. 2008; Beaumont et al. 2009). Subduction of continental India slowed plate convergence significantly at around 50 Ma (Patriat & Achache 1984; Klootwijk et al. 1992; van Hinsbergen et al. 2011) and the initial steep continental subduction stage
convert to a shallow underthrusting, crustal stacking and thickening stage as crustal buoyancy forces overcame the downward pull of the residual ocean slab that most likely began to tear off at this stage (Chemenda et al. 2000; O’Brien 2001; Massonne & O’Brien 2003; Replumaz et al. 2010). The thickened crust (Fig. 2, stage 3) in this now widened orogeny extending both north and south of the suture zone, would have been at kyanite-field conditions but thermal relaxation of the stacked slivers of upper crustal material, with their respective upper crustal radioactive heat production, would have led to significant heating, partial melting and the production of sillimanite ± cordierite-grade gneisses overprinting earlier kyanite-bearing assemblages. With further convergence the weak, anatectic crust is liable to flow – a process extensively modelled for Himalayan-type orogens (e.g. Beaumont et al. 2001, 2004, 2006; Jamieson et al. 2004) – and could extrude to shallower crustal levels – a process

Fig. 2. Summary of the evolution of the Himalaya as a subduction–collision orogeny (not to scale) indicating the timing of blueschist, coesite eclogite, kyanite and cordierite–sillimanite formation (loosely based on Jain et al. 2012). At stage 4 the arrows mark flow of anatectic crust with granulitized eclogites.
invoked to explain the anatectic, sillimanite-grade HHC below a detaching roof (Searle & Rex 1989; Searle et al. 2006: Fig. 2 stage 4) and also the characteristic inverted metamorphism induced in the footwall of the hot extruded bodies (LeFort 1975). Even today the base of the thickened crust is at eclogite facies conditions so it is not surprising that several locations of anatectic HHC gneisses, in some cases c. 200 km south of the ITSZ, contain eclogites overprinted by granulite-facies assemblages at Oligocene to Miocene times (Groppo et al. 2007; Cottle et al. 2009; Grujic et al. 2011; Warren et al. 2011a, b; Lombardo et al. 2016; Wang et al. 2015, 2017; Corrie et al. 2016).

Geology of the Tso Morari area

High-grade crystalline rocks crop out in the arid, high plateau of the Tso Morari area (Fig. 3) as a large dome dipping below the ITSZ in the NE (Zildat detachment fault) and the Tethyan Himalaya of the Zanskar area (Mata and Tetraogal Nappes) in the SW (Steck et al. 1998; Epard & Steck 2008). This c. 100 km long, 50 km wide, NW–SE-oriented body comprises mostly variably-deformed augen-gneisses (the Puga Gneiss) which formed from a porphyritic S-type granite of Cambro–Ordovician (479 ± 2 Ma: Girard & Bussy 1999) age. In some places, for example at Polokongka La, the rock appears as a hardly-modified granite but it has been demonstrated that both deformed and undeformed variants are identical in age (Steck et al. 1998; Girard & Bussy 1999). Hosting the granites are clastic and carbonate sediments of late Proterozoic–Cambrian age. Numerous folded and stretched layers and boudins of metabasalt – some bodies with relatively fresh eclogite facies minerals, other with eclogite relicts – represent intrusions into the granite and sedimentary sequence of sub-alkaline, within-plate basaltic rocks equivalent to the Permian Panjal Trap basalts documented from the weakly-metamorphosed Tethyan (Rao & Rai 2006; Jonnalagadda et al. 2017b). There are two nappes tectonically overlying the Puga Gneiss: the Tetraogal (Permian and younger shelf deposits with bimodal volcanic rocks) and Mata-Nyimaling-Tsarap (Late Proterozoic to Mesozoic clastic and carbonate metasediments intruded by the Ordovician I-type Rupshu granite) Nappes, which are mostly of greenschist facies grade and contain no relicts of the eclogite facies despite containing several metabasite bodies (Steck et al. 1998; Epard & Steck 2008). Minor serpentinites, metagabbros, chromitites and metabasalts of the Karzok ophiolites form mappable slivers at the Tetraogal–Mata Nappe boundary (Fig. 3).

Tso Morari: general geology and geochronology

It was in the Tso Morari area that Berthelsen (1953) made the first reliably-documented discovery

Fig. 3. Simplified geological map of the Tso Morari area, Ladakh, NW India (simplified after Epard & Steck 2008), showing the distribution of eclogites in the Puga orthogneiss.
of eclogites in the Himalaya and also convincingly showed, in his petrographic descriptions and sketches, their origin as metamorphosed Panjal Trap lavas and dykes (i.e. corona dolerite microstructure). Haydon (1904), mapping an area including the southern end of Tso Morari (lake), had already recognized garnet in basic dykes. A modern follow-up study of these rocks had to wait until the 1990s after reports of eclogite in the Kaghan Valley of Pakistan (Pognante & Spencer 1991) spurred a re-investigation of the minor mafic and pelitic bodies associated with the Puga Gneiss (Guillot et al. 1997; de Sigoyer et al. 1997) and led to the discovery of high-pressure (HP) indicators in other rock types too. In metapelites, all containing phengite + rutile + quartz, the additional phases garnet + jadeite + chloritoid + paragonite + glaucoaphane + zoisite + chloritite ± biotite only occur in Fe-rich types, whereas Mg-rich types (whiteschists) contain additional kyanite + Mg-chlorite + talc + garnet and intermediate types staurolite + kyanite + biotite + chlorite: meta-sandstones contain phengite (up to 3.57 Si per formula unit (pfu)) + chlorite + garnet + paragonite + zoisite (Guillot et al. 1997; Epard & Steck 2008). The $P$–$T$ conditions deduced for these samples of 2.0 ± 0.2 GPa, 550 ± 50°C for the eclogite facies peak, an eclogite–blueschist overprint at 1.8–1.3 GPa at the same temperature, a conspicuous subsequent heating at normal crustal depths (i.e. amphibolite facies) and finally partial retrogression to greenschist facies (c. 0.5 GPa at 500°C) conditions were the first solid data to indicate the tectonometamorphic evolution of this subducted crustal unit (Guillot et al. 1997). The parallel study of the metabasites by de Sigoyer et al. (1997) recognized a garnet + omphacite ± phengite + rutile + quartz peak assemblage as well as strong zoning of garnet and growth of glaucoaphane replacing omphacite on the retrograde path. Initial geothermobarometric results (de Sigoyer et al. 1997) matched the evolution of the metapelites with >2.0 ± 3 GPa, 550 ± 50°C for the peak followed by isothermal exhumation (to yield the glaucoaphane) and a high-temperature pulse at 1.0 GPa to yield pargasite humation (to yield the glaucophane) and a high-temperature (HT) peak of about 550°C for the peak followed by isothermal exhumation of the metapelites with >2.0 ± 3 GPa, 550 ± 50°C for the eclogite-facies stage whereas an amphibolite-facies overprint – 48 ± 2 Ma ($^{40}$Ar/$^{39}$Ar) and 45 ± 4 Ma (Rb–Sr) – was associated with garnet zoning, but detailed in-situ zircon geochronology (Donaldson et al. 2013; St-Onge et al. 2013), incorporating measurements from zircon grains in garnet overgrowth rims, strongly suggests an age below 50 Ma age for the peak of the eclogite facies metamorphism. A clear indication of the end of the metamorphism comes from zircon and fission track ages, indicating exhumation to about 10–15 km depth, by 35–40 Ma (Schlup et al. 2003).

**Tso Morari: extensive and intensive studies of the eclogite**

With NW India now joining the select few locations with UHP, coesite-bearing eclogites, there followed a proliferation of petrological studies of these rocks; in many cases aided by an excursion to this remote and travel-restricted region of Ladakh as part of the Himalaya-Karakoram-Tibet (HKT) Workshop in August 2008. Despite the existence of numerous mapped bodies of eclogite (Steck et al. 1998; Epard & Steck 2008: Fig. 3) an extraordinary number of studies have concentrated on just one outcrop (e.g. Sachan et al. 1999; O’Brien & Sachan 2000; Mukherjee & Sachan 2001; Mukherjee et al. 2003; Sachan et al. 2004; Konrad-Schmolke et al. 2008; Singh et al. 2013; St-Onge et al. 2013; Palin et al. 2014, 2017; Chatterjee & Jagoutz 2015; Wilke et al. 2015; Jonnalagadda et al. 2017a); the same outcrop depicted in Berghersen (1953, p. 367, Fig. 13). A summary P–$T$ diagram

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**Geochronology**

From initial multi-method geochronological investigation of both metabasites and metapelites (de Sigoyer et al. 2000) it was clear, despite analytical errors, that the HP metamorphism was of Eocene age. Ages of 55 ± 12 Ma (Lu–Hf, garnet–omphacite–whole rock (WR) metabasite), 55 ± 7 Ma (Sm–Nd garnet–glaucophane–WR metapelite) and 55 ± 17 Ma (U–Pb allanite) delimit the eclogite-facies stage whereas an amphibolite-facies overprint – 48 ± 2 Ma ($^{40}$Ar/$^{39}$Ar) and 45 ± 4 Ma (Rb–Sr) – was associated with garnet zoning, but detailed in-situ zircon geochronology (Donaldson et al. 2013; St-Onge et al. 2013), incorporating measurements from zircon grains in garnet overgrowth rims, strongly suggests an age below 50 Ma age for the peak of the eclogite facies metamorphism. A clear indication of the end of the metamorphism comes from zircon and fission track ages, indicating exhumation to about 10–15 km depth, by 35–40 Ma (Schlup et al. 2003).
showing the contrasting interpretations of peak conditions and trajectory of the prograde path (Fig. 5) outlines the main differences of opinion. Most critical are the cold $P$–$T$ paths (St-Onge et al. 2013; Palin et al. 2014, 2017; Chatterjee & Jagoutz 2015) and the diamond-field peak pressures (Mukherjee et al. 2003; Wilke et al. 2015). The cold subduction paths appear to be influenced by the modelled paths for slab surfaces worldwide (Syracuse et al. 2010: shaded region of Fig. 5) but the Tso Morari rocks are continental, not oceanic crust and were overlain by several kilometres of sediment (now existing as the Tetraogal and Mata Nappes). More relevant are thus the paths resulting from the modelling of a Tso Morari-type situation (Warren et al. 2008; Beaumont et al. 2009) which are also shown in Figure 5. Thermodynamic modelling of the Tso Morari eclogite bulk composition to produce equilibrium $P$–$T$ diagrams (i.e. pseudosections) with (Konrad-Schmolke et al. 2008), or without (except for specific steps: St-Onge et al. 2013; Palin et al. 2014, 2017; Chatterjee & Jagoutz 2015) fractionation along the followed $P$–$T$ path, allow prediction of phase assemblages and compositions. How reliable are these pseudosections?

**Pressure–temperature paths**

The metamorphic evolution favoured by Palin et al. (2017) follows a low-temperature concave (to the temperature axis) $P$–$T$ path (Fig. 5). The expectation from the model (Fig. 6a) is that garnet starts to grow as epidote is consumed and thus that lawsonite is as modally important as garnet for a significant part of the path leading up to the quartz = coesite univariant curve. In addition, although calcic (actinolite) and sodic-calcic (barroisite-winchite) amphiboles are expected, these are in small proportions and glaucophane is by far the modally dominant amphibole. The proposed peak pressure assemblage, at around 2.8 GPa is garnet + omphacite + phengite + rutile + coesite + lawsonite along with minor talc. By extending this modelled path from around 2.8 to 4.8 GPa, to investigate the pressure range speculated by Mukherjee et al. (2003) and Wilke et al. (2015) based on carbonate assemblages, the peak assemblage doesn’t change much: lawsonite slowly disappears to produce more garnet and omphacite. This is an important observation as it means that the record of this big pressure step will only be visible from the very outermost growth zone in garnet, exactly the
part of garnet most susceptible to later resorption or diffusive resetting. In contrast, the study of Konrad-Schmolke et al. (2008), testing equilibrium and fractionation (forward-modelling) approaches, implemented a P–T path staying outside the lawsonite stability field, essentially convex to the temperature axis, and predicting sequential consumption of chlorite, epidote and sodic-calcic amphibole during garnet growth (Figs 5 & 6c). Importantly, it was observed that equilibrium models predicted garnet growth along the whole of the path whereas fractionation models predict segments along the path with no garnet growth followed by renewed growth, close to the quartz = coesite reaction, with significantly different Ca–Mg–Fe such that a compositional step occurred. The P–T path favoured by Chatterjee & Jagoutz (2015) also stays in the cold, lawsonite field to a peak at 2.2–2.3 GPa, 400–425°C (below the quartz = coesite univariant curve) based on interpretation of garnet core and omphacite compositions, but subsequently shows heating to 670–720°C with only a slight sinking of pressures to 1.8–1.9 GPa. This contrasts with most other studies which propose isothermal decompression or initial cooling from peak conditions. In order to ascertain if these new studies have advanced our understanding of the dynamics of Indian Plate subduction it is first necessary to present critical details of the petrography by summarizing observations from this plethora of publications as well as from my own material.

Petrography

In hand specimen (Fig. 4c) the eclogite is recognized due to abundant 1–2 mm diameter, idiomorphic or elongate rather than spherical, pink to red garnet porphyroblasts, set in a finer matrix of partly dark- to emerald-green omphacite and bluish-grey
amphiboles. The proportion of amphibole increases from core to rim of boudins. Also, easily discernible in hand specimen are numerous sub-millimetric orange carbonate (mostly dolomite or Mg-calcite) porphyroblasts and individual flakes or small clusters of white mica. In thin section (Figs 7 & 8) individual grains or clusters of garnet (Fig. 7a) are commonly wrapped by a foliation (Fig. 7b, d) defined by oriented omphacite, phengite, poikilitic but in places idiomorphic carbonate porphyroblasts, and clusters of rutile occur; some samples also contain kyanite (Mukherjee et al. 2003; Wilke et al. 2015). This HP-matrix assemblage is irregularly overgrown by mostly sodic-calcic and calcic amphiboles, clinzoisite and paragonite. Locally, sodic-calcic amphibole is cored by glaucophane (Fig. 8d) (de Sigoyer et al. 1997; Wilke et al. 2015) but not all samples have preserved this phase. Matrix omphacite is commonly replaced by symplectites of Na-rich plagioclase and jadeite-poor clinopyroxene, rutile is rimmed by titanite, and carbonate porphyroblasts display magnesite cores and cracks or rims formed of dolomite (Fig. 8a, b). Later reaction around the carbonates has led to calcite, ferroan dolomite and siderite growth: the latter two phases giving the macroscopic orange colour (Figs 7a & 8a). Garnets contain abundant inclusions, often in S-shaped trails, in a darker red interior, whereas lighter-coloured margins show considerably less inclusions or are inclusion-free (Fig. 7). Inclusions in grain interiors are mostly sodic-calcic amphibole (barrosite), epidote, rutile and aegirine-rich omphacite along with subordinate paragonite, dolomite, calcite, ilmenite, quartz, apatite and zircon. In the paler, outer part of garnet the inclusions are jadeite-rich omphacite, usually with aegirine-rich relics in their cores (see also Fig. 8c), coesite/quartz (Fig. 7c), rutile and minor zircon. Compositional zoning in garnet in combination with changes in inclusion suites is an important indicator of metamorphic evolution in a rock but at the same time it is an obvious indicator of disequilibrium and thus must be adequately accounted for in interpretations utilizing P–T pseudosections. Microprobe compositional maps and line profiles presented by St-Onge et al. (2013) and Wilke et al. (2015) demonstrate important common features of less retrogressed garnets namely: a homogeneous interior zone with flat profiles (c. Alm0.625 Grs0.275 Prp0.08 Sp0.02) surrounded by an overgrowth showing a marked drop in Mn and Ca countered by increases in Mg and Fe (c. Alm0.65 Grs0.17 Prp0.17 Sp0.01) which then is zoned towards maximum pyrope content at the rim (c. Alm0.45 Grs0.195 Prp0.35 Sp0.003). Garnet compositional maps, back-scattered electron images and profiles presented by Chatterjee & Jagoutz (2015) show some of these features although the garnets they analysed show more evidence of partial resetting.
Eclogite garnet

From compositional maps (Fig. 9) and line profiles (Fig. 10) from a typical well-preserved eclogite garnet a number of important features can be recognized. Firstly, in this particular sample, high Mn patches (Sp > 0.20) indicate that the garnet coalesced from two separate cores. The zoning from core to rim (Fig. 10a) shows two critical knick points: at 550 microns where Mg starts to rise as Ca falls and then at 760 microns where a 20 mol% increase in pyrope is compensated by falls in both Fe and Ca. From the compositional maps for Ca and Mg (Fig. 9) it is visible that the Mg-rich overgrowth zone is irregularly developed around the garnet and is in many cases razor sharp. This is emphasized in the short profile across the boundary (Fig. 10b) showing 20 mol% changes in pyrope and grossular over a distance of less than 5 microns. Further growth of this outer zone to minimum Fe and maximum Mg (thus also maximum X_Mg) demonstrates the phenomenal variation in garnet composition in the space of a quarter of a millimetre (Fig. 10c). The distribution of inclusions in garnet as well as key matrix phases are readily visible in the compositional maps (Fig. 9). The Fe map, with its baseline raised to emphasize Fe in garnet, shows black holes where inclusions occur; isolated red spots denote ilmenite. Maps of Ca and Al indicate the location of epidote phases whereas Na and Si distribution show well the amphibole and omphacite inclusions. The highest Na + Al marks secondary albite in inclusions and at the edge of matrix symplectite. As the maps incorporate part of the matrix they also act as a useful guide to the distribution of matrix phases. Thus, the preserved omphacite-rich matrix domains are readily distinguishable in the NE and SE corners with Ca (green–yellow), Na (yellow) and Si (yellow) concentrations whereas retrograde amphibole domains mostly surrounding garnet, but also penetrating cracks, have lower Ca (blue–green), Na (blue) and Si (green). Likewise, phengite (K = red; Al = green, Mg = green) can be separated from paragonite (K = blue, Al = yellow, Mg = violet) and poikilitic
Mg-calcite in the SW corner (Ca = red, Mg = violet; Si = Na = Al = black) with its inclusions of omphacite and phengite is immediately visible.

There is little difference between this description of the garnet and its inclusions when compared to the reports of Chatterjee & Jagoutz (2015), St-Onge et al. (2013) and Palin et al. (2014, 2017). However, all of these assume garnet growth in the lawsonite field citing rare inclusions containing both paragonite and zoisite/clinozoisite as evidence for former lawsonite (e.g. St-Onge et al. 2013, Fig. 5f): not even a single lawsonite grain has ever been identified in these rocks. As both individual paragonite and epidote-phase inclusions are common in the garnet interiors it is not really surprising that now and again both occur together but considering the phenomenal number of epidote-phase inclusions, mostly monophase, of all different sizes, it seems inconceivable that all could result from an anisochemical breakdown of lawsonite without leaving any evidence of material transportation pathways. Likewise, the epidote-phase inclusions are predominantly elongate whereas oblong or rhombus shapes, as expected for a pseudomorph replacing lawsonite (e.g. St-Onge et al. 2013, Fig. 5f), occur extremely rarely. The most sensible conclusion is thus that the garnet grew in the epidote rather than the lawsonite stability field. Likewise thermobarometry, utilizing presumed Si-rich phengite (Fig. 4d of Chatterjee & Jagoutz 2015) inclusions with garnet core compositions, misinterprets the microtexture which instead represents later replacement of metastable garnet cores with hydrous phases (amphibole, white mica, clinozoisite) as is common in atoll garnets of some Tso Morari eclogite bodies (Jonnalagadda et al. 2017a).

**Eclogite clinopyroxene**

Further confusion arises from the interpretation of the jadeite content of omphacite. Careful analysis within a single thin section shows a wide range of compositions (Fig. 11). Although there is a very wide range of Ca (i.e. diopside + hedenbergite...
Fig. 9. Compositional maps for a typical garnet porphyroblast and its surrounding matrix. Width of imaged zone is 2.56 mm.
components: Ca-Tschermaks component is absent), Mg and Na contents (Fig. 11a), jadeite and aegirine components appear to be bimodally distributed (Fig. 11b). It is common to see ragged relicts of an older, more aegirine-rich clinopyroxene (Jd 30–40 Aeg 15–20) in the cores of more jadeite-rich grains (Jd 42–52 Aeg 5) both in the matrix and in the outer parts of garnet (Figs 8c & 7c, respectively). This is consistent with the jadeite-rich, incidentally also most Mg-rich pyroxene representing equilibrium with the Mg-rich garnet exterior. The jadeite-rich omphacite is therefore not the pyroxene in equilibrium with the pyrope-poor garnet core and thus the ≥2 GPa 400–425°C for the initial peak pressure stage deduced by Chatterjee & Jagoutz (2015) is erroneous. The spread of pyroxene compositions within a single thin section demonstrates well the problems encountered in the choice of key indicator phases for geothermobarometry in non-equilibrated rocks.

Eclogite amphiboles

Amphiboles are another important phase in the Tso Morari eclogites but appear in a number of sodic-calcic, calcic and even sodic varieties. Several studies have documented the range of compositions of inclusions in garnet as well as the differences between cores and rims of matrix amphiboles (e.g. St-Onge et al. 2013; Palin et al. 2014; Chatterjee & Jagoutz 2015). Due to the complexity of the amphibole nomenclature, combined with deduced
compositional trends (especially in calcium, sodium, silicon and aluminium), the rocks yield amphibole compositions plotting in the actinolite, actinolitic hornblende, magnesio-hornblende, ferroan pargasite, pargasite, winchite, barroisite, magnesio-taramite, magnesio-katophorite and glaucophane fields. The sodic amphibole glaucophane, first noted by de Sigoyer et al. (1997), is a core to some sodic-calcic amphibole grains in the matrix of better preserved eclogites (in some examples included in carbonates: Wilke et al. 2015) and minor inclusions in the outer core (not the overgrowth rim) of garnet (Mukherjee et al. 2003). Apart from glaucophane, compositional trends of all of the other amphiboles can be well visualized in a plot of tetrahedral Al (or Si) v. B-site Na (Fig. 12). Large matrix amphiboles are typically barroisite, as are most of the inclusions in garnet, and exhibit roughly equal Aliv and Alvi values of c. 0.7–0.9 (per 23 oxygens formula unit). The pleochroic, deep-blue amphiboles directly rimming garnet have significantly higher Aliv, mostly above 2.0 pfu and straddle the ferroan pargasite–magnesio-taramite boundary depending on their B-site Na content. Important to recognize is that many analyses of the small amphibole grains included in garnet plot between these two ranges in Aliv and most likely represent the effects of retrogressive modification of barroisite to Al-rich compositions rather than reflecting a large compositional range of amphiboles during garnet growth along the prograde path. The necessity for a non-isochemical evolution in these (as well as many other) eclogites, that is with pulses of fluid influx during exhumation driving amphibole (and paragonite and zoisite/clinozoisite) growth, is outlined by Palin et al. (2014). Important to recognize is that the modally abundant sodic amphibole predicted (Fig. 6a) from the concave P–T paths (St-Onge et al. 2013; Palin et al. 2014, 2017; Chatterjee & Jagoutz 2015) is not recorded, despite an abundance of amphibole inclusions in growth-zoned garnet. Instead, only minor glaucophane is reported in the outer core zone of garnet in some samples. This strongly suggests that the concave path is erroneous. The convex path (Konrad-Schmolke et al. 2008; Wilke et al. 2015) predicts the sodic-calcic amphibole– epidote rather than lawsonite–glaucophane evolution trend for the main growth phase of garnet: a trend consistent with changes in rare-earth-element patterns in the growing garnet based on modal changes of matrix phases due to prograde growth reactions. However, a number of authors report analyses of sodic amphibole (mostly glaucophane) cores to some barroisite porphyroblasts in the matrix (e.g. de Sigoyer et al. 1997; Mukherjee et al. 2003; Wilke et al. 2015) which, together with other microstructural observations, indicate that the amphibole-free UHP matrix of omphacite + phengite + rutile + carbonate was overgrown initially at the relatively low temperatures of the glaucophane field (see Fig. 8d). The same situation is reported in the Himalayan eclogites in Kaghan Valley, Pakistan (Wilke et al. 2010a; Lombardo et al. 2016). The important implication is that the exhumation followed the subduction channel thermal trend, as depicted in Figure 5 from the modelling of Warren et al. (2008) and Beaumont et al. (2009), and not the high temperature–high pressure trend proposed by Chatterjee & Jagoutz (2015). The proposed passage of the Tso Morari complex through the mantle wedge, accompanied by heating at high pressure, is thus shown to be completely inconsistent with the abundant mineralogical evidence.

Eclogite carbonates

A further aspect of the Tso Morari eclogites that has been neglected by some authors is the presence of carbonates. Abundant, macroscopic, mostly <1 mm, orange porphyroblasts of dolomite are evident with the naked eye in many hand specimens (Fig. 4c).
In thin section (Fig. 8a, b), back-scattered electron imagery combined with microprobe and micro-Raman analysis reveal that the matrix carbonate porphyroblasts comprise Mg-calcite, magnesite with dolomite-filled cracks or dolomite with relics of magnesite inside (Mukherjee et al. 2003; Chatterjee & Jagoutz 2015; Wilke et al. 2015). The rims as well as fractures in the carbonates contain further calcite (Mg-poor) and siderite related to several stages of retrograde reaction. The poikiloblastic, in places idiomorphic and twinned, matrix carbonates enclose omphacite and phengite. Garnet also contains inclusions of carbonates with dolomite and dolomite + calcite occurring in cores and outer cores, respectively (Wilke et al. 2015) but magnesite is reported in some overgrowth rims (Mukherjee et al. 2003). The presence of carbonates, especially magnesite, has important implications. Magnesite ± dolomite ± calcite is reported from both marble and metabasite in HP and UHP metamorphic terranes such as the Dabie Shan (Wang & Liou 1993), Kokchetav Massif (Zhu & Ogasawara 2002), western Tianshan (Zhang et al. 2003), German Erzgebirge (Massonne 2001) and SW Norway (Smit et al. 2008). In some of these terranes microdiamond has been reported so the fact that experimental and thermodynamic studies indicate the breakdown of dolomite to yield aragonite and magnesite to occur at pressures above 5 GPa for temperatures above 500°C (e.g. Schertl & Okay 1994; Ogasawara et al. 1995; Shat-sky et al. 1995; Luth 2001; Buob et al. 2006), well into the diamond field (e.g. Kennedy & Kennedy 1976; Bundy 1980), would appear to be complementary evidence for UHP metamorphism. Wilke et al. (2015) deduced c. 4.7 GPa at 610 ± 50°C for a magnesite-bearing Tso Morari eclogite based on the dolomite = aragonite + magnesite reaction and activities for the actual Fe-bearing magnesite and dolomite in the measured sample: conditions similar to those proposed by Mukherjee et al. (2003). However, in other magnesite-bearing eclogites (Smit et al. 2008), carbonate microstructural have been interpreted as a product of disequilibrium dissolution–precipitation reactions formed at lower (albeit still at coesite field) pressures. Another feature linked to carbonates is a decarbonation reaction of magnesite + quartz to produce a corona of talc as recognized in some samples (Mukherjee et al. 2003; Wilke et al. 2015); perhaps the true origin of talc described and interpreted by other studies (e.g. St-Onge et al. 2013; Palin et al. 2014, 2017) as supporting evidence for the cold, concave P–T path.

Summary and conclusions

It is clear from all descriptions of the Tso Morari eclogite that the rock is far from being perfectly equilibrated. Compositional zoning, as well as multiple generations of key phases, indicate that information from several different stages along the P–T evolution path are present in a single thin section. It is for this reason that petrological investigation of the rocks can only be successful if extreme caution is applied to all attempts using equilibrium thermodynamics to explain a non-equilibrium situation. The proposed evolution paths from recent papers using the pseudosection approach (St-Onge et al. 2013; Palin et al. 2014, 2017; Chatterjee & Jagoutz 2015) follow low-temperature subduction trends, as expected for the surface of subducted oceanic crust (Syracuse et al. 2010), and are concave in form. These paths predict a considerable segment of the prograde history within the glaucophane and lawsonite stability fields, so it is remarkable, considering the abundance of inclusions in growth-zoned garnet, that not a single lawsonite grain has been identified as an inclusion in garnet. Glaucophane is present as inclusions in the outer part of garnet but by far the dominant amphibole as inclusions in garnet is sodic-calcic and not sodic in nature. The validity of the subduction thermal paths of Syracuse et al. (2010) has been questioned in the light of the observation that generally hotter temperatures are recorded by subduction zone rocks (Penniston-Dorland et al. 2015). A hotter path was deduced from thermal–mechanical modelling of the Tso Morari subduction history (Warren et al. 2008; Beaumont et al. 2009) the general trend of which (see Fig. 5) corresponds with the two-step convex path implemented in the thermodynamic forward modelling study of Konrad-Schmolke et al. (2008). This latter study, although passing through the same near peak P–T point (2.8 GPa, 650–700°C) as that of Palin et al. (2017), predicts major garnet growth in the stability field of epidote and sodic-calcic amphibole: exactly as seen in the natural samples. The path modelled by Konrad-Schmolke et al. (2008) also predicts a break in growth of garnet that helps explain the incorporation of coesite as inclusions as renewed major growth of garnet occurred at UHP conditions. This is an important point: coesite only appears as inclusions after the break in growth and also in the inclusion-poor margin of garnet. In cases where the thin section has cut the marginal zone of a garnet it may appear that coesite is in the core (a problem also relevant for omphacite) but analysis of the adjacent garnet quickly shows Mg-rich, Ca-poor compositions, as typical for the overgrowth zone in all studies. The proposed path of Chatterjee & Jagoutz (2015) invokes even colder subduction than Palin et al. (2017) but there are serious issues relating to the correct choice of omphacite and phengite compositions which drive conventional equilibria to such low (2.2 GPa, 425°C) conditions. More problematic in the Chatterjee & Jagoutz (2015) study is...
the proposed heating, still at high pressures (to roughly 670–720°C at 1.8–1.9 GPa) which forms the basis for their model of exhumation through the mantle wedge. The fact that this hot exhumation bypasses the coesite field entirely and yields no change for the post peak-pressure growth of glaucophane (e.g. de Sigoyer et al. 1997; Wilke et al. 2015) gives this model little credibility. The true peak pressure reached by the eclogites is still uncertain. The common abundance of dolomite surrounding magnesite in the best preserved eclogites (e.g. Mukherjee et al. 2003; Wilke et al. 2015) strongly suggests very high pressures in the diamond stability field (>4.5 GPa) even when compensating for the c. 10% Fe component deviation from end member compositions. Thermodynamic models (e.g. Palin et al. 2017) expanded to c. 5 GPa do not predict much change in the eclogite apart from a minor growth of garnet at the expense of omphacite: a feature that would be very difficult to demonstrate in the actual Tso Morari samples considering the multiple retrogressive steps – reflected in amphibole, zoisite/clinozoisite and paragonite growth – linked to influx of fluids.

The very large P–T window (equivalent to the pressure range experienced by the whole continental crust) with essentially the same stable mineral assemblage (albeit with slightly different modal proportions and compositions) is a weakness in the pseudosection approach to geothermobarometry in HP and UHP metamorphic rocks: a problem already pointed out for HP granulites (O’Brien 2008). In order to obtain more precise formation conditions of UHP rocks it is necessary to determine intersects of mineral composition isopleths within the defined assemblage P–T window. This approach then suffers from exactly the same problems as for (often derived) conventional geothermobarometry with deduction of supposed equilibrium-phase compositions in a rock commonly with multiple reaction stages. In addition, the scarcity of experimental data to validate the, often unquestioned, end-member thermodynamic properties as well as mixing models (especially at the extreme compositions of UHP rocks) for rock-forming minerals should make us all more critical of how we interpret our samples. In summary, the commonly utilized tool in modern petrology, the fixed-composition equilibrium-phase diagram (pseudosection), is just that: a tool enabling us to speculate as to what could have occurred in a rock. Without time spent on petrography, detailed microanalysis and investigation of possible local dominant equilibrium, no sensible result should be expected: short cuts are rarely productive! Also, often forgotten is that rocks with long histories, and that definitely includes UHP rocks, will often exhibit different degrees of ‘memory’ even when they derive from the exact same outcrop. This is probably partly a factor leading to the wide range of interpretations of the Tso Morari eclogite.

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