Peridotitic minerals of the Nidar Ophiolite in the NW Himalaya: sourced from the depth of the mantle transition zone and above

SOUVIK DAS¹, BARUN K. MUKHERJEE¹*, ASISH R. BASU² & KOUSHICK SEN, JR¹

¹Petrology and Geochemistry Group, Wadia Institute of Himalayan Geology, 33 GMS Road, Dehra Dun, 248001, India
²Department of Earth and Environmental Sciences, University of Texas at Arlington, Arlington, TX 76019, USA

*Corresponding author (e-mail: barun@wihg.res.in)

Abstract: The Nidar Ophiolite Complex (NOC) consists of a c. 10 km thick ophiolite suite in the NW Himalaya, India. The c. 7 km thick lower ultramafic part of the ophiolite body terminates against the Tso Morari Crystallines, which represent the leading edge of the Indian continental margin. Mineral inclusions from the peridotites in the lower ultramafic part of the NOC were studied, including C2/c clinopyroxenite, disordered coesite and high-pressure Mg2SiO4 (probably β-Mg2SiO4). These minerals, found in two harzburgite bodies from the ophiolite’s mantle section, were characterized by laser Raman spectroscopy and electron micro-probe analysis. Textural evidence supporting decompression from an ultra-high-pressure condition was also observed, such as Cr spinel exsolution needles in olivine crystals. The systematic mineral phase transitions of coesite → quartz, high-pressure clinopyroxenite → orthoenstatite and β-Mg2SiO4 → Cr spinel exsolution needles in olivine suggest that the mantle section of the Nidar Ophiolite evolved from the deep mantle beneath a palaeo-spreading centre. The phase stabilities of these high-pressure minerals require derivation from the depth of the mantle transition zone (410–660 km). A transport mechanism for these minerals is suggested via dunite channels along a mantle adiabat in the focused convective flow below the spreading centre. This mechanism brought these deep mantle phases into the ultramafic part of the NOC. These observations suggest that some part of the mantle section of the NOC in the NW Himalaya originated in a mid-ocean ridge setting.

Supplementary material: Representative mineral chemical compositions of the lherzolites (1M1 and 1NU27) and the host channelized dunites are available at http://www.geolsoc.org.uk/SUP18836.

The long-standing controversy of the early twentieth century concerning the presence of ultramafic rocks in mountain belts (Steinman 1905; Hess 1955) was finally resolved with the advent of plate tectonic theory, which showed that these rocks were examples of exposed lithospheric mantle at plate margins (Moores 1970). The Alpine–Himalayan ophiolites were reported to represent oceanic crust and upper mantle (Hess 1962; Moores & Vine 1971). Ophiolites are generally divided into a top magmatic group formed at the oceanic ridge (constituting the bulk of the oceanic basaltic crust) and a lower group consisting of peridotites with olivine, ortho-clinopyroxene and accessory minerals. This lower group shows plastic deformation and solid-state flow resulting from the high-pressure and high-temperature conditions (Nicolas & Jackson 1972).

Ophiolites are key elements in determining the complicated geological process by which orogenic belts are formed. However, the original tectonic setting of ophiolites is often not known exactly and therefore their geodynamic relationship with mountain-building processes is still speculative. The environment of formation of ophiolites has been a major controversy since the first publication of the plate tectonic concept and their presence along suture zones has led geoscientists to the famous ‘ophiolite conundrum’. Earlier geologists considered ophiolites to be representatives of ancient oceanic crust on the basis of the similarity between the internal structure of ophiolites and the modern oceanic lithosphere (Nicolas 1989). A preferred hypothesis is that ophiolites originated in spreading centres, which then collided with continental crust in subduction zones and finally obducted over the continental crust. A revolution in thought occurred when results from geochemical analyses began to suggest the derivation of ophiolites from depleted mantle at shallow depths resembling the subduction zone environment (Miyashiro 1973) rather than at a mid-oceanic ridge setting. This result progressively led geologists to propose a supra-subduction zone origin for ophiolites (Pearce et al. 1984). Ultra-high-pressure minerals were later
discovered in the podiform chromitites of the Luobusa Ophiolite (Tibet) (Robinson et al. 2004; Yang et al. 2007), which suggested that such chromitites evolved through upwelling from the deep mantle in a spreading centre setting. However, Arai (2013) rejected this deep mantle origin theory for podiform chromitites and suggested an alternative mechanism of the deep recycling of low-pressure chromitites via mantle convection. Apart from the podiform chromitites, it is widely accepted that the ultramafic rocks of ophiolite mantle sections are the products of shallow-level magmatic activity and that their depth of formation is restricted to within the stability field of spinel peridotite (<60 km depth; Coleman 1977).

We report here an extraordinary set of mineral assemblages identified by micro-laser Raman spectroscopy and electron probe micro-analysis (EPMA). The mineral assemblages are of ultra-deep origin from the lower peridotites of the Nidar Ophiolite Complex (NOC), Indus Suture Zone in the western Indian Himalaya. The current study contributes to previous work by the discovery of a unique set of ultra-high-pressure mineral assemblages in ophiolitic peridotites. This demonstrates contiguity in their pressure–temperature stability space, linking the mantle transition zone (410–660 km) with the regional scale as an eye-shaped body (Fig. 1a).

Geological setting

The Nidar Ophiolite in SE Ladakh is well exposed in the Indus Suture Zone that separates the Indian plate in the south from Asia to the north (Fig. 1a). The northern part of the Nidar Ophiolite is occupied by the calc–alkaline Ladakh Batholith and the Indus Molasse; to the south the Indian plate consists of continental margin sedimentary units and the Tso Morari Crystalline, which bears ultra-high-pressure mineral assemblages (Mukherjee et al. 2003).

The Nidar Ophiolite appears in plan view on the regional scale as an eye-shaped body (Fig. 1a). The NOC displays variations in thickness (such as the 3 km Mahe–Sumdo section and the 10 km Nidar Valley section) across its NW–SE trend (Fig. 1a, b). The ophiolite sequence is best exposed in the Nidar Valley, where the lower ultramafic part of the ophiolite suite terminates against the ultra-high-pressure eclogite-bearing Tso Morari core complex (Mukherjee et al. 2003) to the south and its upper basaltic part rests against the Indus Molasse.

The Nidar Valley displays the classical ophiolite sequence (Fig. 1c, d). The ophiolite consists of three units from south to north: (1) topmost volcanic rocks with a chert association; (2) gabbroic rocks in the middle; and (3) an ultramafic lower unit. The uppermost part, approximately 1.5 km thick, consists of volcanic rocks, vesicular pillow lavas, sheeted dykes and volcanogenic sediments with chert (Fig. 1c, d). The pillow lavas show a concordant relationship with the gabbroic parts of the sequence.

The rocks of the middle part of the ophiolite show distinct grain size variations from fine- to coarse-grained gabbroic rocks. Units 1 and 2 are up to 3 km thick (Fig. 1d). The gabbronoritic unit 2 show a zigzag contact with the serpentinites of the lowermost ultramafic unit (Fig. 1c, d). This contact is identified as the petrological Moho of the Nidar Ophiolite. This Moho is best developed in the Mahe–Sumdo section.

The ultramafic unit is about 7 km thick in the Nidar–Kyon Tso section (Fig. 1c, d). The mantle section consists of dunite and varieties of peridotite (harzburgite and lherzolite). The base of the mantle section is occupied by a spinel-bearing dunite body (about 3 km thick), concordant with the trend of the Indus Suture Zone (NW–SE) with moderate (about 10–15%) serpentinitization. A metre-thick chromite vein (Sachan & Mukherjee 2003; Sachan et al. 2007) consisting of thin parallel layers is present within the basal dunite. Some channelized networks (about 3 km long, 100–200 m wide and trending NE–SW) with the same lithology (olivine + disseminated chromite) emerge from the basal dunite body, cross-cutting the entire ultramafic body that consists mostly of harzburgites (Siddaiah 2001; Sachan & Mukherjee 2003) and lherzolites (Fig. 2). Some peridotite rock bodies are also observed as entrapped xenoliths within the channelized dunites (Fig. 2). On the basis of their modes of occurrence, the ultramafic part of the basal Nidar Ophiolite can be classified into three types: (1) peridotites cut by the discordant dunite; (2) the discordant dunite; and (3) the peridotites entrapped within the channelized dunites. All these peridotites occur as entrapped xenoliths, caught in the dunite channels. Mineralogically, these peridotites show variable assemblages of olivine (40–50 vol%), orthopyroxene (c. 25 vol%), clinopyroxene (c. 15 vol%), Cr spinel (2–5 vol%) and serpentine (5–18 vol%).

Methodology

Nine samples from the three ultramafic units of the Nidar Ophiolite were used in this study. Thin sections were prepared from the rock samples for petrographic study at the Wadia Institute of Himalayan Geology (WIHG), Dehradun, India. The thin sections were studied optically under transmitted and reflected light using a Nikon Eclipse LV 100...
Fig. 1. (a) Geological map of the Nidar Ophiolite Complex (NOC) of the Indus Suture Zone in the Indian Himalaya; inset shows the location of the NOC. (b) Digital elevation model of the NOC. A–B represents a traverse along the Nidar Valley for our geological mapping of the ophiolite complex. (c) Cross-section along A–B showing the detailed lithological units of the ophiolite suite. (d) Pseudo-stratigraphic reconstruction of the NOC based on geological mapping, showing the location of the peridotite body of samples 1M1 and 1NU27 and adjacent rocks.
POL microscope and image analyses were performed using NIS-Elements software. The mineral compositions were analysed with a CAMECA SX 100 microprobe equipped with five wavelength-dispersive spectrometers. The pyroxene and olivine grains were analysed at an accelerating voltage of 20–15 kV with a beam current of 20 nA and a fully focused c. 5 μm beam diameter. The raw data were corrected using the ZAF method. Detection limits were c. 0.1% and an accuracy better than 5% was obtained. Some of the minerals were analysed using a JEOL 8900 microprobe at the Cornell Center for Materials Research, Cornell University (Ithaca, NY, USA). These analyses were performed at an accelerating voltage of 15 kV and a 20 nA beam current. The beam diameter was c. 3–4 μm. Using both these probes the detection limits were c. 0.1% and an accuracy better than 5% was obtained.

Laser Raman spectroscopy was performed on thin rock wafers using a LabRAM HR-Horiba Jovin Yvon instrument at the WIHG. Spectra were generated at room temperature with the 514.5 nm line of a green 12 mW Ar laser and detected with a charge-coupled detector array through Olympus 10×, 50× and 100× objective lenses. The laser spot on the surface had a diameter of approximately 1 μm and a power of c. 15–33 mW. The light was dispersed by a 600 grooves/mm holographic grating. The real-time display exposure times for the minor mineral phases were selected between 1 and 40 s with an exposure time of 1–15 s and an accumulation number of 1–10. The Raman spectroscopy was calibrated with a synthetic Si standard (at 520.5 cm⁻¹). The shift was <1 cm⁻¹ with respect to the synthetic standard.

The Raman spectra of high-pressure Mg₂SiO₄ in peridotite 1NU27 were analysed using the

Fig. 2. (a) Field photograph of the lowermost ultramafic rocks of the Nidar Ophiolite. Some channelized discordant dunites originate from the base. (b) In some places the discordant dunites show variations in grain size. Coarse elliptical portions of the dunite channel show a NE trend along the channel. (c) The peridotite bodies are trapped as xenoliths in the dunite flow. (d) Dunite intruded as a secondary injection in an early formed peridotite body.
Ranishaw in Via Raman spectroscopy at the Cornell Center for Materials Research using a 488 nm laser. The exposure time was 10–30 s and the accumulation number was 1. The laser spot diameter was approximately 1–2 μm with a power of c. 4 mW. Light was dispersed by a holographic grating with 600 and 1800 grooves/mm. All the data were generated with a 50× objective lens. The instrument was calibrated with a synthetic Si standard at 520.5 cm⁻¹ and the shift was ≤1 cm⁻¹.

Mineralogy of the studied samples

The dunite channels cross-cut the entire ophiolite mantle section and several metre-scale bodies of ultramafic rocks are present as entrapped xenoliths in the dunite channel. Two of these xenoliths were the focus of this work and bear the signature of a deep mantle origin. These peridotites, along with some of the adjacent rocks, were characterized by petrography and their mineral chemistry.

Petrography

The dunite sample 1NU18 is composed of olivine (c. 80–85%) and Cr spinel (c. 5%). The olivine grains are cross-cut by serpentine veins to create a mesh texture. The channelized dunite bodies are moderately serpentinized (c. 10–15%), whereas the degree of serpitinization varies (10–100% serpentinites) from place to place in the basal dunite section.

The lherzolite xenolith 1M1 is also serpentinized (18–20%) and is composed of olivine (40%), orthopyroxene (25%), clinopyroxene (15%) and chromite (2%). The other lherzolite sample (1NU27) is much less serpentinized (c. 5%). This rock shows an assemblage of olivine (50%), orthopyroxene (25%), clinopyroxene (15%) and Cr spinel (c. 5%). The pyroxenes occur as porphyroclasts in an olivine-dominated matrix in both lherzolites.

Ultra-high-pressure minerals in the Nidar peridotites

The lherzolites show coarse-grained ortho- and clinopyroxene (Fig. 3) porphyroclasts in an olivine matrix. A thin section of 1M1 with two square-shaped translucent grains attached to opaque phases in an enstatite host is of special interest (Fig. 4a). Their boundaries with the host enstatite are diffuse and indistinct. In one grain a typical palisade texture is observed, which is a characteristic texture of the transition from coesite to low-pressure quartz. A colourless foam-like crystalline mineral (size 10 × 5 μm) with higher relief (Fig. 4b) is seen in the elliptical, elongated core of this palisade grain. The rest of the palisade grain is colourless and devoid of cleavage, but has a lower relief than the mineral in the core. Some radiating fractures are observed from the core to the rim of the palisade texture. This texture is typical of the solid-state phase transformation from coesite to quartz and the mineral in the core displays the typical optical character of coesite. Points 1–3 in Figure 4b indicate the spots analysed by laser Raman spectroscopy. Another square-shaped translucent silica grain (Fig. 4c) was also probed by laser Raman spectroscopy (spot 4).

Part of a thin section from another lherzolite (1NU27) is also shown (Fig. 4d, e). This displays two spinel grains with euhedral inclusions of a mineral that is opaque in transmitted light. The same lherzolite (1NU27) displays an orthopyroxene grain with unusual long (several hundred microns; up to 700 μm) and thin (1–2 μm) coherent lamellar inclusions (Fig. 5a). The lamella can be easily distinguished under the crossed nicols of an optical microscope. The lamellae show slightly higher relief under plane polarized light.

In this lherzolite (1NU27), a dense population of rectilinear exsolution needles 12–30 μm long and >1 μm wide with a strong preferred orientation are also seen in the host olivine (Fig. 6a, b). The needles are translucent and have a higher relief than the host olivine. From the polished surface of the thin sections, the area occupied by the needles was determined using Nikon software to be about 0.3% by volume of the host olivine and locally as high as 0.7%.

All these mineral phases were characterized by laser Raman spectroscopy and EPMA. These results are documented in the following sections along with their respective Raman spectra and major element compositions.

Mineral chemistry

All of the mineral chemistry data were generated using EPMA. Representative mineral chemical compositions of both the lherzolites (1M1 and 1NU27) and the host channelized dunites are given in the Supplementary material. Olivine and spinels are the most common minerals in both the lherzolites and the dunites. The lherzolite xenoliths and channelized dunite show notable variations in composition.

The forsterite content of the 1M1 lherzolite olivines is 89.9–90.5% (S2). The spinels show an Mg# (Mg/Mg + Fe) of c. 52.6–52.8% and a Cr# (Cr/Cr + Al) of c. 54.5–55.2% (S2). The orthopyroxenes and clinopyroxenes have an Mg# of 89.9–91.35% and 92.5–93.3%, respectively (S4, S2). The variation in Mg# (89.9–91.35%) across an orthopyroxene similar to the host of the high-
pressure phases in 1M1 (S3, S4) suggests a relatively Fe-rich core. The olivines of the 1NU27 lherzolite have a forsterite content of 90.6–91.2% (S5). The Mg# in the orthopyroxene and clinopyroxene were 91.1–91.3% and 95%, respectively (S5). The Mg# and Cr# in the spinels were 48.9–49.23% and 67.9–70.3%, respectively (S5). The same parameters in the host dunite (1NU18) were: forsterite in olivine c. 93.34–93.57%; Mg# in spinel 58.38–62.4%; and Cr# in spinel 61.9–72.3% (S1). The olivines and spinels in the lherzolites (1M1 and 1NU27) were also richer in Fe than the host dunite, whereas the Cr# was lower in the 1M1 spinel than in the dunites and the 1NU27 lherzolite.

The lamellar inclusions in the pyroxenes and the exsolution needles in lherzolite 1NU27 were also analysed by EPMA (Table 1). At the time of the analyses of the lamellar inclusions in orthopyroxene, the spot size of the EPMA was c. 5 µm and the width of the lamellae was c. 2 µm. The acquired data therefore represent data of the lamellae + host pyroxene. However, a comparison of these data with the host pyroxene showed that both the lamellar inclusions and the host were MgSiO₃ pyroxene with Mg# 90.9–91.4%. It was also noted that there was no difference in the Ca content between the lamellae and the host enstatite.

Similarly, the compositions of the 1 µm thick exsolution needles in the olivines of the 1NU27 lherzolite were also determined (Table 2). The chemistry data for the host olivine + exsolution needles were compared with the data for the olivine. It was found that the needles were comparatively enriched with Cr, Al and Fe.

The euhedral inclusion in the Cr spinel of lherzolite 1NU27 was also analysed (Table 3). It was found that the inclusions were composed of an Mg₂SiO₄ mineral. However, it was also observed that their composition showed some differences with respect to the matrix olivine. The Mg# of the high-pressure Mg₂SiO₄ was much higher (92.07–92.35%) than that of the olivine (90.79%) in the matrix. It was also observed that these inclusions were richer in Cr.
Identification of the ultra-high-pressure minerals

Representative Raman spectra of the spots in Figure 4 confirmed the presence of high-pressure mineral phases (Fig. 7). The core of the palisade texture (Fig. 4b) shows the Raman trend for coesite, with peaks at 118, 159, 176.5 and 524 cm$^{-1}$ (Fig. 7B1). The principal peak at c. 524 cm$^{-1}$ is broadened and has a lower intensity than usual (Fig. 7B1). This may be due to its exsolved origin (Zhang et al. 2005) or retrogression.
from tetragonal stishovite, as indicated by its square shape.

The EPMA data revealed that the host mineral was MgSiO₃ pyroxene (S4). The low intensity peaks at 244, 369, 432 and 573 cm⁻¹ from the core of this palisade texture with a coesite spectrum (Fig. 7B1) indicated the presence of clinoenstatite, high P Cen or low P Cen (Ulmer & Stalder 2001). However, the high intensity sharp peak at 673 cm⁻¹ was strongly indicative of high-pressure C2/c Cen (Chopelas 1999). The splitting of the c. 673 cm⁻¹ peak from the core (Fig. 4b, spot 1) to the rim (Fig. 4b, spots 2 and 3) and their corresponding spectra in Figure 7B1, B2 and B3 also suggest transformation from a C-centred MgSiO₃ to a P-centred MgSiO₃ pyroxene (Ross & Reynard 1999; Tribaudino et al. 2012). The other peaks in Figure 7B1, B2 and B3 are assigned to orthoenstatite. Where coesite is visible, the signatures of C2/c clinoenstatite are distinguishable. The other peaks indicate mixed clino–orthoenstatite phases in the three spots of the palisade-textured grain (Fig. 7B1, B2 and B3). The coesite and clinoenstatite peaks in Figure 7B2 are weaker than the orthoenstatite peaks. From the core to the rim of the palisade texture, the C2/c clinoenstatite progressively transforms into orthoenstatite.

The other translucent section of the grain in Figure 4c also shows an assemblage of C2/c Cen – coesite (Fig. 7C4), similar to the portion bearing the palisade texture.

The isotropic euhedral inclusions in the Cr spinels (Fig. 4d, e) display the Raman spectrum of forsteritic olivine: a doublet at 822–823, 854 with a 961 cm⁻¹ peak (Fig. 7D5, E6). The peaks near c. 542 and 918 cm⁻¹ have been reported previously from shocked olivine (Farrell-Turner et al. 2005) under very high pressure conditions. The 918 cm⁻¹ peak may be due to primary wadsleyite. To understand the behaviour of the peaks of the high-pressure Mg₂SiO₄, the Raman spectrum of high-pressure Mg₂SiO₄ was compared with the forsteritic olivine in the groundmass of lherzolite 1NU27. An interesting phenomenon was observed (Fig. 8) when the Raman trend of olivine and high-pressure Mg₂SiO₄ were compared. The principal olivine doublets in high-pressure Mg₂SiO₄ showed inversion with respect to forsteritic olivine. The doublets of the high-pressure Mg₂SiO₄ also showed a shift to a higher value. This inversion of the doublet along with the shift of the high-pressure Mg₂SiO₄ peaks with respect to the olivine in lherzolite 1NU27 (Fig. 8) can also be considered as evidence of decompression from the high-pressure Mg₂SiO₄ of the mantle transition zone to a forsteritic olivine (Durben et al. 1993). Mineral chemical analyses were carried out on the high-pressure Mg₂SiO₄ and its host spinel and were compared with the matrix olivine (Table 3). As expected from the Raman spectrum, the inclusion was Mg₂SiO₄ in composition. It was noticed that the high-pressure Mg₂SiO₄ was slightly enriched in Cr and that the Mg# of the high-pressure Mg₂SiO₄ was distinctly
higher (92.07–92.35%) than that of the Mg# of the olivines (90.79%) in the matrix. The unusual isotropic euhedral inclusions were characterized as high-pressure Mg$_2$SiO$_4$, which has the dominant characters of forsteritic olivine, but also bears the signature of a retrogressed high-pressure spinelloid (probably b-Mg$_2$SiO$_4$).

The lamellar inclusions (Fig. 5a) showed a clinopyroxene trend, whereas the host displayed a Raman trend of orthopyroxene (Fig. 5b, c). From the EMPA results it was clear that the lamellar inclusions were MgSiO$_3$ pyroxene (Table 1). It could be inferred that the host was orthoenstatite (Oen), whereas the lamellae indicated a mixed clinoenstatite (Cen) and orthoenstatite phase (Das et al. 2013) (Fig. 5b). The mixing of the Oen–Cen phase could be due to either the thin nature of the lamellae or retrogression from the high-pressure polymorph. The splitting of the 666.6 cm$^{-1}$ peak into a 659.2, 681.4 cm$^{-1}$ doublet from the lamellae to the host is indicative of a C-centred to P-centred pyroxene structure (Tribaudino et al. 2012) (Fig. 5b, c). There are some subtle differences between the Cen + Oen lamellae and the host Oen in the 324–389 cm$^{-1}$ region (Fig. 5b, c). The 252.8 and 356.6 peaks appeared in the lamellae, which may be assigned to clinoenstatite (Ulmer & Stalder 2001) (Fig. 5b).

Laser Raman spectroscopy and EPMA were used to characterize the thin needles in the olivine of lherzolite 1NU27. In the case of the needles, the spot size of the laser Raman was c. 2 μm, whereas the needles were c. 1 μm wide (Fig. 6b). As a result of this discrepancy, the spectrum shows the peaks of the host olivine in the background (Fig. 6d). To identify the exsolved needles, the Raman spectra of the needles and the olivines were compared (Fig. 6c and d). The Raman spectra of the needles showed the characteristic peaks of Cr spinel at 659.3 and 701.7 cm$^{-1}$ (Fig. 6d). A peak at 601.8 cm$^{-1}$ can be assigned to Cr spinel because of its high intensity with respect to the

Fig. 6. (a, b) Photomicrographs of the unusual exsolution micro-textures in the 1NU27 lherzolite showing a dense population of rectilinear 12–30 μm long and 1 μm wide exsolution needles with a strong preferred orientation in the host olivine. These translucent needles have a higher relief than the host olivine. From the polished surface of the thin sections using Nikon software, the area occupied by the needles was measured to be about 0.3% by volume of the host olivine and locally as high as 0.7%. (c, d) Strongly oriented needles in olivine were analysed by Raman spectroscopy. The analysed spots are marked in (a, b). (c) Raman trend of olivine used to deduct the peaks of the needles from (d). (d) The Raman trend suggests the presence of both Cr spinel and olivine.
607 cm$^{-1}$ olivine band (Fig. 6d). The chemistry of the needles was determined by EPMA (Table 2), which yielded mixed analyses with the host olivine. Comparison of the major element data with the host olivine indicated the presence of Cr spinel. The presence of similarly oriented rectilinear Cr spinel needles in olivines in the Alpe Arami peridotite massif in the Alps have been interpreted as a product of exsolution under ultra-high-pressure conditions (Dobrzhinetskaya et al. 1996).

### Discussion

The mineral phase transitions and micro-textures seen in this work have revealed that these lherzolites of the Nidar Ophiolite evolved from ultra-deep mantle depths. The observed textures and inclusions indicate a systematic decrease in the pressure–temperature conditions, which ultimately have implications for their emplacement below the ophiolitic crust in a mid-oceanic ridge environment.

In the orthopyroxene porphyroclasts of the 1M1 and 1NU27 lherzolites, the clinoenstatite (Figs 7B1, C4 and 5) could be any one of the two types of $C_{2}/c$ Cen–high T Cen and high P Cen. From the Raman spectra it is difficult to decipher the high P Cen. However, high T Cen has a small stability field at $P < 1$ GPa and $T > 1500$ °C (Paccolo & Gasparik 1990). The association of coesite, exsolved Cr spinel and high-pressure Mg$_2$SiO$_4$ rules out high T Cen and indicates the presence of high P Cen. The phase transition from high P Cen to Oen takes place at a minimum of c. 8 GPa (Gasparik 1990).
The Cr spinel exsolution needles in olivine suggest high-pressure primary conditions, as reported in the Alpe Arami peridotite, Swiss Alps (Dobzhinskaya et al. 1996), where it was proposed that these exsolution textures were the result of the decompression of a higher pressure polymorph of olivine, probably $\beta$-Mg$_2$SiO$_4$. However, this suggestion of primary $\beta$-Mg$_2$SiO$_4$ has been questioned (Hacker et al. 1997) as a result of a lack of other supporting evidence for mantle transition zone ($\approx 410–660$ km depth). In the present example, however, we found high-pressure Mg$_2$SiO$_4$ (or retrogressed $\beta$-Mg$_2$SiO$_4$) as inclusions in Cr spinel in the same lherzolite (1NU27) sample that contained the needles. The mineral chemistry indicated that the high-pressure Mg$_2$SiO$_4$ was rich in Cr with respect to the matrix olivine (Table 3). It was therefore considered that decompression from the depth of the mantle transition zone resulted in the possible breakdown of $\beta$-Mg$_2$SiO$_4$ to produce the observed Cr spinel exsolution needles in the host olivine.

The mineral phase transitions $\beta$-Mg$_2$SiO$_4$ $\rightarrow$ Cr spinel exsolution in olivine and high P Cen $\rightarrow$ Oen indicate the initiation of mantle flow from the mantle transition zone (Fig. 9). Petrologists and tectonophysicists generally consider that ophiolitic peridotites evolve from $\approx 2$ GPa (c. 60 km) depth (e.g. Coleman 1977). Petrological knowledge of the generation of mid-oceanic basalts also suggests that melting in an upwelling mantle is restricted to 70–85 km depth (e.g. O’Leary et al. 2010). However, geophysical studies have shown that the root of the mantle upwelling beneath a spreading centre is at much greater depths (Su et al. 1992; Zhao et al. 1997). Our observations may be able to fill the knowledge gap between these geophysical and petrological views. The mineralogical observations presented here indicate an ultra-deep origin for the mineral phases in the lower ultramafic rocks of the NOC and their pressure–temperature stability relations (Fig. 9) allow a tentative exhumation pathway for these mineral phases. For the system MgO–FeO–SiO$_2$, this pathway, the adiabatic route, terminates beneath the mid-ocean ridge basalt eruption centres. The in situ mineralogical data presented here suggest that this pathway originates from the depth of the mantle transition zone, passes through the stability fields of wadsleyite, continues successively through the high-pressure

| Element | Host spinel | Host spinel | High-pressure Mg$_2$SiO$_4$ | Olivine |
|---------|-------------|-------------|-----------------------------|---------|
| NaO     | 0.04        | 0.02        | 0                           | 0       |
| MgO     | 9.25        | 9.52        | 49.49                       | 49.15   |
| Al$_2$O | 15.13       | 14.69       | 0                           | 0       |
| SiO$_2$ | 0.03        | 0.06        | 41.05                       | 41.33   |
| P$_2$O$_5$ | 0            | 0.01        | 0                           | 0       |
| K$_2$O  | 0           | 0.01        | 0                           | 0       |
| CaO     | 0.02        | 0           | 0.02                        | 0.01    |
| TiO$_2$ | 0.05        | 0.04        | 0.02                        | 0.02    |
| Cr$_2$O$_3$ | 51.26     | 52.19       | 0.64                        | 0.44    |
| MnO     | 0.27        | 0.34        | 0.16                        | 0.13    |
| FeO     | 20.57       | 19.88       | 7.28                        | 7.53    |
| Total   | 96.62       | 96.76       | 98.67                       | 98.58   |

| Cation | Host spinel | Host spinel | High-pressure Mg$_2$SiO$_4$ | Olivine |
|--------|-------------|-------------|-----------------------------|---------|
| Na     | 0.02        | 0.011       | 0                           | 0       |
| Mg     | 3.708       | 3.809       | 1.811                       | 1.799   |
| Al     | 4.796       | 4.649       | 0                           | 0       |
| Si     | 0.009       | 0.015       | 1.008                       | 1.015   |
| P      | 0.001       | 0.002       | 0                           | 0       |
| K      | 0           | 0.005       | 0                           | 0       |
| Ca     | 0.007       | 0           | 0.001                       | 0.001   |
| Ti     | 0.014       | 0.008       | 0                           | 0       |
| Cr     | 10.902      | 11.078      | 0.012                       | 0.008   |
| Mn     | 0.062       | 0.078       | 0.003                       | 0.003   |
| Fe     | 4.627       | 4.464       | 0.15                        | 0.155   |
| Total  | 24.141 (O = 32) | 24.119 (O = 32) | 2.985 (O = 4) | 2.981 (O = 4) |
| Mg#    | 44.49       | 46.04       | 92.35                       | 92.07   |

The mineral chemistry of high-pressure Mg$_2$SiO$_4$ was compared with the matrix olivine. The high-pressure Mg$_2$SiO$_4$ was slightly enriched in Cr and the Mg# of high-pressure Mg$_2$SiO$_4$ was much higher (92.07–92.35) than the Mg# of the olivines in the matrix (90.79).
clinoenstatite field and the stability field of coesite and, finally, terminates in the orthoenstatite field (Fig. 9). We conclude that some of the mantle peridotites below the Nidar Ophiolite show evidence of high-temperature–high-pressure recrystallization, plastic deformation and solid-state flow that began during their journey at mantle transition zone depths of 410–660 km before their arrival beneath the ophiolitic volcanic crust at oceanic ridges. This conclusion has obvious implications for the nature of upper mantle flow and sources of oceanic ridge basalts in a spreading centre.

Several studies have reported ultra-high-pressure minerals, such as diamonds, coesite, high-pressure forms of TiO₂ and other minerals, in the chromitites of a Tibetan ophiolite (Robinson et al. 2004; Yang et al. 2007) and, more recently, the high-pressure polymorph of olivine, ringwoodite,
has been found as inclusions in the diamonds of a Brazilian kimberlite (Pearson et al. 2014). Another group of micro-mineralogical studies in subducted peridotites from continental collision zones have established the exhumation of the host peridotites from 300 to 400 km depths in the mantle (e.g. Dobrzhinetskaya et al. 1996). The current study, however, can be distinguished from previous work by the discovery of a unique set of mineral phase transitions from deep mantle to upper crust conditions that demonstrate contiguity in their pressure–temperature stability space, linking the mantle transition zone to the basaltic crust formed at the ocean ridge. The absence of prograde metamorphism in the associated units of the ophiolite ultramafic rocks does not support the deep subduction origin of our reported high-pressure minerals. The systematic phase transitions infer only the decrease in pressure–temperature conditions, which suggests the evolution of high-pressure lherzolites as a result of deep mantle flow in a spreading centre. Early work on the Nidar Ophiolite suggested different original settings based on studies of discrete and segmented parts. Geochemical studies of the gabbros inferred an origin in an intra-oceanic arc setting (Mahéo et al. 2004; Ahmad et al. 2008). Ravikant et al. (2004) proposed a supra-subduction setting origin by studying the layered and disseminated chromitites. However, the evolution of ultra-high-pressure signatures is difficult to explain using the previously described tectonic models, which refer to magmatic activity at shallower depths. We prefer to suggest that at least part of the mantle section of the Nidar Ophiolite originated in a mid-oceanic ridge setting.

Unlike previous studies, the newly reported high-pressure mineral phases of the Nidar Ophiolite show a continuous pathway from the mantle.

Fig. 8. (a, b). Raman spectra of high-pressure Mg$_2$SiO$_4$ and forsteritic olivine of the 1NU27 lherzolite matrix were compared with the aid of a LabRAM HR-Horiba Jovin Yvon instrument at WIGH. (c) In the spectrum of high-pressure Mg$_2$SiO$_4$, the doublets show inversion and a slight shift to the higher side with respect to the forsteritic olivine.

![Image of Raman spectra and mineral inclusions](image_url)
The pressure—temperature path of the journey of the relict grains from their place of origin to final emplacement in the peridotitic host beneath the ophiolitic volcanic crust. The inferred adiabatic route as a function of pressure ($P$) and temperature ($T$) for the system $\text{MgO-FeO-SiO}_2$ starts from the mantle transition zone and ends at the oceanic spreading centre. The representative phase diagrams were adapted from Pacalo & Gasparik (1990) and Gasparik (1990). The forsterite ($\alpha$)–wadsleyite ($\beta$)–ringwoodite ($\gamma$)–$\text{Mg}_2\text{SiO}_4$ phase boundaries were adapted from Akaogi et al. (1989).

**Conclusions**

1. The presence of $C2/c$ clinoenstatite, disordered coesite and high-pressure $\text{Mg}_2\text{SiO}_4$ (probably $\beta$-$\text{Mg}_2\text{SiO}_4$) in the herzolites of the NOC challenge the conventional idea of the 'origin of ophiolite peridotites in a spinel lherzolite stability field' (Coleman 1977). The presence of high-pressure clinoenstatite and $\beta$-$\text{Mg}_2\text{SiO}_4$ indicates the origin of host peridotite at $>240$ km, that is, the depth of the mantle transition zone (410–660 km).

2. This paper reports the first evidence in favour of a high-pressure origin for ophiolitic peridotites. Early reports of high-pressure phases from other ophiolites were from only the podiform chromitite parts of these ophiolites.

3. The high-pressure mineral phases do not advocate a shallow depth supra-subduction origin for some portions of the Nidar Ophiolite. The evolution of these systematic mineral phase transitions to the uppermost mantle region. These minerals began their journey upwards from the mantle transition zone by solid-state flow, arriving beneath the ophiolite’s volcanic crust, which was generated by partial melting as a result of this upwelling at the oceanic ridge (Fig. 9).
transitions can be best explained by a spreading centre setting.
(4) We also suggest that retrogression of $\beta$-$Mg$_2$SiO$_4$ may produce Cr spinel exsolution needles in olivine during decompression. With the aid of combined EPMA and laser Raman spectroscopy, a rare natural analogue of this highly debated phase transition (Dobrzyniecka et al. 1996; Hacker et al. 1997) was described in this paper.
(5) The presence of deep mantle phases in the ultramafic part of the NOC suggest that at least some part of the mantle section of the NOC in the western Himalaya originated at a mid-oceanic ridge setting.

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