



## Research Paper

## Geodynamic evolution of the Tethyan lithosphere as recorded in the Spontang Ophiolite, South Ladakh ophiolites (NW Himalaya, India)

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## ABSTRACT

The Spontang Ophiolite complex represents the most complete ophiolite sequence amongst the South Ladakh ophiolites and comprises mantle rocks (depleted harzburgites, dunites and minor lherzolites) as well as crustal rocks (basalt, isotropic gabbros, layered gabbros etc.). In the present study, detailed geochemistry (whole rock as well as mineral chemistry) and Sr-Nd isotopic analyses of thirty-six ultramafic-mafic samples have been attempted to constraint the evolution and petrogenetic history of the Tethyan oceanic crust. Major, trace-element and REE patterns of the peridotites and their minerals indicate that the lherzolites experienced lower degrees of partial melting resembling abyssal peridotites (at higher temperatures,  $T_{\text{REE}} = \sim 1216^\circ\text{C}$ ) than the harzburgites (6%–8% versus 15%–17%). Elevated  $\varepsilon_{\text{Nd}}(t)$  and variable  $^{87}\text{Sr}/^{86}\text{Sr}(t)$  ratios along with REE patterns suggest that the Spontang mafic rocks display N-MORB affinity with negligible participation of oceanic sediments in their genesis are originated from a depleted upper mantle with little contribution from subduction-related fluids. MORB-type Neotethyan oceanic crust is associated with the earliest phase of subduction (of older Jurassic age) through which a younger intra-oceanic island arc (Spong arc) subsequently developed. Harzburgites REE display typical U-shaped patterns, suggesting that these rocks have been metasomatized by LREE-enriched fluids. On the other side, mafic rocks are characterized by heterogeneous  $(\text{Nb/La})_{\text{PM}}$  and  $(\text{Hf/Sm})_{\text{PM}}$  and relatively homogeneous  $\varepsilon_{\text{Nd}}(t)$ , indicating interaction of subduction-related melts with the upper mantle during the initiation of subduction, in Early Cretaceous times.

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## 1. Introduction

Ophiolites are the remnants of fossil oceanic lithosphere that have been tectonically emplaced on continental margins and are considered key to reconstruct the magmatic and tectonic history of the oceanic lithosphere (Dilek and Furnes, 2011). Ophiolites depict fragments of upper mantle and oceanic crust (Dewey and Bird, 1971; Coleman, 1977; Nicolas, 1989) integrated into continental margins during continent–continent and arc–continent collisions (Dick and Flower, 2003), ridge–trench interactions (Cloos, 1993; Lagabrielle et al., 2000), and/or subduction-accretion

occurrences (Cawood et al., 2009). Ophiolites although dismembered, compared to mantle xenoliths which are volumetrically smaller in number and in size, offer a unique opportunity to study the relationships within and between the mantle and crustal units. Systematic studies using sophisticated instrumentation have made determination of geochronological, geochemical and isotopic characteristics of both crustal and mantle units of ophiolites possible. Specifically, geochemical and isotopic studies of mantle peridotites and crustal rocks from oceanic floor and ophiolitic type settings are important to characterize the chemical and isotopic heterogeneity of the mantle (Zindler and Hart, 1986). Detailed geochemical studies of mantle peridotites from ophiolitic suites provide important information on melting processes, melt extraction episodes and interaction of oceanic crust, mantle melts and fluids in the paleo-oceanic lithosphere–asthenosphere system under active tectonic

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conditions (Kelemen et al., 1992; Barth et al., 2003; Dilek et al., 2007; Ishikawa et al., 2007). Geochemical studies of these ancient oceanic lithospheric terranes are therefore important in order to constrain the magmatic processes involved in oceanic crust genesis, i.e. partial melting, fractional crystallization and melt/fluid-rock reactions (Shervais, 2001; Pearce, 2003; Aldanmaz et al., 2008).

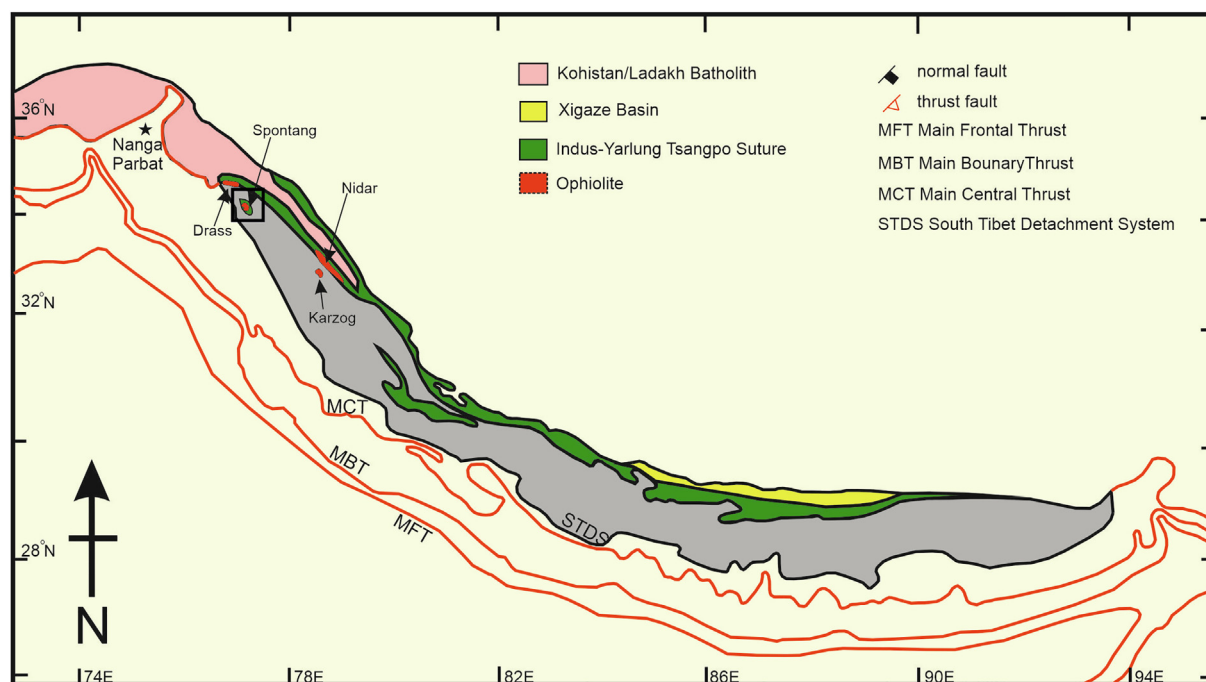
The Tethyan Ophiolitic Belt formed as a consequence of the closure of the Tethyan Ocean and collision of the Indian and Eurasian tectonic plates (Pal et al., 2003; Ghosh et al., 2017) and are marked by several events such as the subduction of the Indian plate beneath Eurasia, ophiolite obduction onto the continental margin, high pressure – low temperature metamorphism and collision-accretion activities (Dilek et al., 1999; Robertson 2000; Aldanmaz et al., 2008). The Tethyan ophiolites stretching from western Europe to India and China represent multiple suture zones within the Alpine, Mediterranean and Himalayan-Tibetan orogenic belts. Several ophiolitic slivers are exposed in several locations from Spain through the Alps, Turkey, Iran, Oman, Pakistan, India, Tibet, Burma and Indonesia (Saha, 2010). Departing from the original 1972 Penrose definition (Anonymous, 1972), numerous studies on the Tethyan ophiolites have significantly improve our understanding of both ophiolites and oceanic crust, in general. The Tethyan orogenic belt preserve ophiolitic slivers in two tectonic settings viz. within the suture zone (e.g. Xigaze ophiolite, south Tibet; Nicolas et al., 1981; Göpel et al., 1984) and those obducted onto the passive continental margin (e.g. Oman ophiolite; Lippard, 1986). Considering the expanse of the Himalayas and the distance the Indus suture runs, few ophiolite complexes are observed within the Himalayan belt (Fig. 1), mainly because they form the highest structural thrust sheet and more often than not, they are subjected to intense erosion. The Spontang ophiolite complex located south of the Indus Suture Zone (Fuchs, 1979; Reuber, 1986; Searle, 1986) forms the highest level of the allochthon above the marine sediments of the Lamayuru accretionary complex. In this study, we present petrology and whole rock (major- and trace-element and Sr/Nd isotopic) geochemical compositions of

both mafic and ultramafic rocks along with mineral chemistry (major and REE) of the ultramafic minerals with a view to evaluate their petrogenesis and geodynamic implications.

## 2. Geological setting

The Indus Tsangpo Suture Zone (ITSZ) is a significant crustal lineation representing the collisional boundary between the continental plates of Southern India and Northern Eurasia (Le Fort, 1975; Molnar and Tapponnier, 1975; Gansser, 1977). It is regarded to depict the remnants of Neo-Tethys ocean sandwiched between the Karakoram zone in the north and the Zaskar zone of the Tethys Himalayas in the south (Gansser, 1980; Srikantia and Razdan, 1981). Remnants of this Neo-Tethyan ocean are preserved along the suture as ophiolitic slivers in the Ladakh area (Gansser, 1977; Srikantia and Razdan, 1981; Honegger et al., 1982; Robertson, 2000; Maheo et al., 2004; Ahmad et al., 2008) with correlates found in the east in Tibet (Nicolas et al., 1981; Girardeau et al., 1985a, b; Zhou et al., 1998; Aitchison et al., 2000; McDermid et al., 2002; Miller et al., 2003; Xia et al., 2003; Dubois-Côté et al., 2005; Dupuis et al., 2005; Aitchison et al., 2007; Bédard et al., 2009) and the west in Pakistan (Jan et al., 1985; Sarwar, 1992; Zaigham and Mallick, 2000; Kakar et al., 2013a, b).

The Ladakh ophiolites are arranged in an east–west direction and divided into the North and South Ladakh group of ophiolites based on their geographic position (Maheo et al., 2004). The South Ladakh ophiolites consists of the Spontang massif which is the most complete fragments of the Neo Tethys ocean to be preserved along the ITSZ (Fuchs, 1982; Reuber, 1986; Searle, 1986). The Nidar ophiolite (~140 Ma) located about 120 km west is the closest nearly complete ophiolite series to the Spontang Massif (Maheo et al., 2004; Ahmad et al., 2008; Zyabrev et al., 2008) whereas the Karzog ophiolitic slice is considered to be detached klippe of Nidar ophiolite (de Sigoyer et al., 1997). The northern Ladakh alpine ophiolite group is preserved as extremely deformed and dismembered discrete slices best exposed in the Dras and Suru valleys (Gansser, 1980; Srikantia and Razdan, 1981; Honegger et al., 1982;



**Fig. 1.** Regional geological map of the Himalayas (after Hébart et al., 2003) showing location of the South Ladakh (Spontang, Nidar and Karzog) and North Ladakh ophiolite (Drass) sequences of Maheo et al. (2004). The study area has been demarcated by a box.

Radhakrishna et al., 1984; 1987; Reuber et al., 1989; Sinha and Mishra, 1994; Robertson, 2000).

The Spontang massif located 30 kms south of the Indus Suture Zone (Fuchs, 1979; Reuber, 1986; Searle, 1986) is exposed near the village of Photoksar with elevations ranging between 4000 and 6000 m (Fig. 2). Forming part of the South Ladakh group of ophiolites (after Maheo et al., 2004), the Spontang ophiolite displays most elements of a complete ophiolite sequence as defined by the Penrose-style stratigraphic sequence (Anonymous, 1972). The area has been a major attraction to many geologists who are credited with provided detailed geological maps as a result of several expeditions (Fuchs, 1979; Honegger et al., 1982; Reibel and Reuber, 1982; Bassoulet et al., 1983; Keleman and Sonnenfeld, 1983; Colchen and Reuber, 1986; Colchen et al., 1987; Reuber, 1986; Reuber et al., 1989; Reuber et al., 1992; Corfield et al., 2001). The Spontang Ophiolite can be subdivided into a western mantle unit (Reuber et al., 1986; Corfield et al., 2001) and an eastern crustal unit (Corfield et al., 1999; 2001; Pedersen et al., 2001) with much of the sequence being tectonically disturbed by faulting and the maximum exposed vertical thickness of the deformed ophiolite of 1 km (Corfield et al., 1999). The complete ophiolite sequence is significantly dismembered by deformation. The Spontang ophiolite consists of an upper mantle depleted harzburgites and dunites (Reuber et al., 1989) and some lower crustal cumulates (Reibel and Reuber, 1982; Reuber, 1986) along with gabbros, sheeted dykes and pillow lavas (Corfield and Searle, 2000) (Fig. 2a). Under the Spontang ophiolite, the Photang Thrust sheet (Corfield et al., 1999) consists of tectonic mixtures and volcanic units connected with marine sediments. Detailed field relationships between mantle and crustal units have been presented in Jonnalagadda et al. (2019) and are summarized below for ready reference.

Approximately 5 km west of Photoksar village along the Phu River, an extensive sequence of crustal rocks composed of high level gabbros and pillow lavas is exposed along the eastern flank of the Photang Valley at base of the massive Photang Kangri. A gradational contact of the sheeted dykes (Fig. 2b) with the overlying pillow lavas is observed downstream in the eastern side of the Photang valley (Fig. 2c). Some exposures reveal characteristic basalt – basalt and gabbro – basalt intrusive relationships. Isotropic gabbros were identified at the base of the dyke sequence with absence of massive and layered varieties.

The N-S ridge separating the Photang and Spong valleys, reaches over 5500 m and exposes an excellent structural section through the Photang Thrust Sheet. The mantle sequences exposed in the Spong valley consisting of depleted harzburgites (Fig. 2d) overlain by an upper lherzolitic unit are well exposed in the high cliffs of the upper Spong valley and near the upper peaks of Sisir La. Dunitic bodies are commonly observed either as large boulders (Fig. 2e, f) or intruding the peridotite units. Several gabbroic and trondhjemitic dykes are seen to crosscut the mantle units (Fig. 2g), some several meters in width consistent with earlier reports by Corfield et al. (2001).

Lower crustal sequences viz. weathered cumulates of wehrlites and pyroxenites (Fig. 2h) with associated gabbros are exposed in the Marling Chu valley and the Bumiktse valleys. The highly tectonized gabbros in the Bumiktse valley are exposed below the snout of the glacier and are thrust directly onto the volcano-sedimentary sequence of the lower Bumiktse Valley. Crustal sections exposed along both valleys and west of Bumiktse are observed to continue across the inaccessible vertical cliffs and glaciers of the Photang Kangri. South of the Photoksar village, a gradational transition of ophiolitic units to a volcano-sedimentary sequence of pillow lavas and interbedded volcanoclastic deposits on the northern edge of the Photang Kangri and along the Bumiktse and Marling valleys is observed (Fig. 2i). These volcanic

and volcanoclastic sequences of island-arc affinity (Spong arc) which tectonically overly the Spontang ophiolite are believed to be associated with an immature array of the island forming above the oceanic substratum of the Spontang ophiolite (Reuber et al., 1992; Corfield et al., 2001). There is no clearly defined contact between the units; however, depending on the start of volcanoclastic deposits in the field, a transition can be noted. The sequence of several 100 m thick units, consisting of angular and poorly sorted clasts, is widely displayed in the Bumiktse valley with occasional clasts of chert.

### 3. Samples and petrography

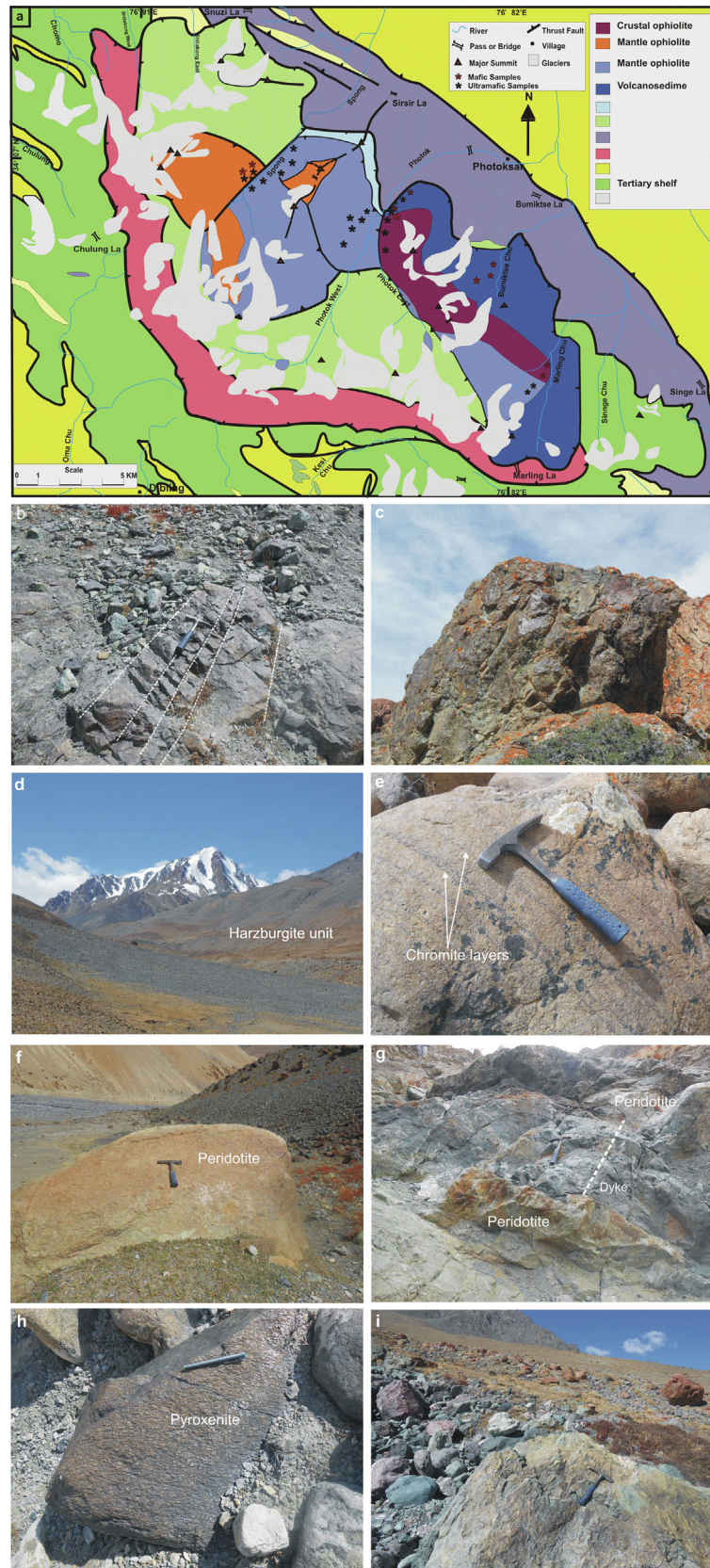
A total of thirty-six samples consisting of ultramafic and mafic rocks were analyzed in the present study (Table 1). The mafic samples are represented by basaltic dyke intrusions into the peridotites and gabbros whereas the ultramafic rocks comprise depleted harzburgites, minor lherzolites and dunites. Detailed petrography of the peridotite samples is presented in Jonnalagadda et al. (2019) and summarized below for ready reference.

Harzburgites are coarse-grained and display porphyroclastic to protogranular textures. Their primary modal mineralogy includes olivine (60 to 75 vol.%), orthopyroxene (15 to 30 vol.%), minor clinopyroxene relics (<1 vol.%) with accessory spinel (less than 3 vol%). Olivines are coarse, exhibit undulose extinction surrounded by smaller crystals of olivine and display curvilinear boundaries meeting at triple junctions. Zoning, kink bands, and recrystallization are commonly observed (Fig. 3a). Orthopyroxenes are colorless, generally occurring as big prismatic crystals, 0.2 to 3.5 mm in size with curvilinear boundaries and often showing exsolution lamellae, kink bands and gliding (Fig. 3b, c). Minor clinopyroxenes are interstitial, 0.1 to 0.3 mm in length and show pale green color in polarized air light with sutured boundaries.

Lherzolite specimens demonstrate textural shift from porphyroclastic to equigranular mosaic texture. Feeble alteration to serpentine is observed at places. Subhedral olivine occurs as porphyroclasts (0.3 to 1.5 mm) whereas anhedral grains are observed as mosaic crystals with curved boundaries. Both generations show curvilinear, serrated boundaries, recrystallization, and the presence of kink bands (Fig. 3d). Clinopyroxene may be present in the olivine matrix as porphyroclasts with exsolution lamellae of orthopyroxenes (Fig. 3e). Spinel exsolution lamellae are noted in the orthopyroxenes at places.

Dunites are fresh, unaltered, with granular texture. Coarse olivine grains are large, porphyroclastic, elongated and display undulose extinction and intense zoning (Fig. 3f). They range in size from 0.9 to 4.00 mm and are surrounded by smaller subhedral grains. Matrix olivines are subhedral to anhedral 0.5 to 2 mm in size and display polygonization and serrated boundaries. Euhedral spinels are commonly observed along olivine boundaries or as inclusions within olivines.

Gabbros are medium- to fine-grained foliated rocks displaying inequigranular to protomylonitic texture and predominantly contain hornblende, relic pyroxenes and plagioclase phenocrysts, Fe-Ti oxides and secondary minerals such as epidote and quartz. Plagioclase crystals are elongate and cloudy with anorthositic core and albitized rims (Fig. 3g). They rarely show polysynthetic twinning. Pyroxenes are augitic in composition ( $\text{En}_{0.26-0.36} \text{Fs}_{0.38-0.55} \text{Wo}_{0.08-0.28}$ ) and occur as relict anhedral grains with two distinct set of cleavages at right angles to each other and display simple twinning. In most samples, the pyroxenes are completely altered to amphiboles with greenish magnesio-hornblende core and brownish Ti-rich magnesio-hornblende rims. The gabbros are affected to low degrees of sea-floor metamorphism as observed



**Fig. 2.** (a) Detailed geological map of the Spontang ophiolite showing the major tectonic units (after [Corfield et al., 2001](#)). Red and black stars indicate sample locations of the mafic and ultramafic samples respectively. (b) Contact of the sheeted dykes with the overlying pillow lavas as observed downstream in the eastern side of the Photang valley. (c) Outcrops of pillow lavas exposed in the western side of the Photang Valley. (d) Harzburgite unit exposed in the upper Spong valley near the upper peaks of Sisir La. (e) Chromite layering visible in boulders of Dunite. (f) Boulders of peridotite (dunite) exposed in the upper Spong valley. (g) Mafic dykes are observed intruding the mantle units with sharp boundaries. (h) Boulders of pyroxenites exposed in the Marling Chu valleys. (i) Volcano-sedimentary sequence of pillow lavas and interbedded volcanoclastic deposits exposed in the Bumiktse valley.

**Table 1**

Summary of location as well as the coordinates of ultramafic and mafic samples of the Spontang ophiolite complex.

| Sr. No. | Location                                | Sample No.  | Abbreviation Used | Rock Type  | Latitude     | Longitude   | Altitude (m) |
|---------|-----------------------------------------|-------------|-------------------|------------|--------------|-------------|--------------|
| 1       | Photok valley                           | MAR/1/PB    | A1                | Basalt     | 34° 04.419'  | 76°47.991'  | 4277         |
| 2       |                                         | MAR/2/PB    | A2                | Basalt     | 34° 04.398'  | 76°47.940'  | 4285         |
| 3       |                                         | SH/1/P      | B1                | Harburgite | –            | –           | –            |
| 4       |                                         | SH/2/D      | B2                | Lherzolite | 34°06.756'   | 76°46.512'  | 4512         |
| 5       |                                         | SH/3/G      | B3                | Gabbro     | –            | –           | –            |
| 6       |                                         | SH/4/P      | B4                | Harburgite | –            | –           | –            |
| 7       |                                         | SH/5/G      | B5                | Gabbro     | –            | –           | –            |
| 8       |                                         | SH/6/H      | B6                | Harburgite | –            | –           | –            |
| 9       |                                         | SH/8/G      | B8                | Gabbro     | 34°06.670'   | 76°46.380'  | 4564         |
| 10      |                                         | SH/10/P     | B10               | Lherzolite | –            | –           | –            |
| 11      |                                         | SH/12/G     | B12               | Gabbro     | 34°07.412'   | 76°46.812'  | 4455         |
| 12      |                                         | ST/3/G      | B15               | Gabbro     | –            | –           | –            |
| 13      |                                         | St/4/D      | B16               | Harburgite | 34°07.412'   | 76°46.812'  | 4455         |
| 14      | Marling Valley                          | MAR2/8/P    | C8                | Harburgite | 34°04.013'   | 76°47.082'  | 4394         |
| 15      |                                         | MAR2/9/P    | C9                | Lherzolite | 34°04.013'   | 76°47.082'  | 4394         |
| 16      |                                         | MAR2/10/G   | C10               | Gabbro     | 34°03.943'   | 76°46.997'  | 4416         |
| 17      | Western Side of Photok Valley           | SH2/2/P     | D2                | Harburgite | 34°07.103'   | 76°46.575'  | 4519         |
| 18      |                                         | SH2/5/P     | D5                | Harburgite | 34°06.990'   | 76°46.500'  | 4534         |
| 19      |                                         | UR/5/P      | D11               | Lherzolite | 34°03.940'   | 76°46.828'  | 4463         |
| 20      |                                         | UR/8/G      | D14               | Gabbro     | 34°03.940'   | 76°46.828'  | 4463         |
| 21      | Spontang valley                         | SPO/1/P     | E1                | Harburgite | 34°06.591'   | 76°46.645'  | 4526         |
| 22      |                                         | SPO/2/G     | E2                | Gabbro     | 34°06.445'   | 76°46.314'  | 4587         |
| 23      |                                         | SPO/4/H     | E4                | Harburgite | 34°06.190'   | 76°46.044'  | 4636         |
| 24      |                                         | SPO/5/D     | E5                | Dunite     | 34°06.143'   | 76°45.994'  | 4640         |
| 25      |                                         | SPO/7/D     | E7                | Harburgite | –            | –           | –            |
| 26      | Western Marling Valley                  | PYX 1/16    | G1                | Dunite     | 34°00.546'   | 76°52.300'  | 4338         |
| 27      |                                         | PYX 2/16    | G2                | Harburgite | 34°00.530'   | 76°52.379'  | 4335         |
| 28      |                                         | G 3/16      | G7                | Harburgite | 34°00.348'   | 76°51.954'  | 4404         |
| 29      |                                         | PD 2/16     | G6                | Harburgite | 34°00.368'   | 76°51.500'  | 4465         |
| 30      | Spong Valley                            | PDT/SPO/16  | I1                | Harburgite | 34°06.470'   | 76°46.542'  | 4581         |
| 31      |                                         | PYX3/SPO/16 | I4                | Lherzolite | 34°06.158'   | 76°46.285'  | 4615         |
| 32      |                                         | G2/SPO/16   | I6                | Gabbro     | –            | –           | –            |
| 33      |                                         | HZ/SPO/16   | I7                | Harburgite | 34°06.210'   | 76°46.328'  | 4607         |
| 34      |                                         | D/SPO/16    | I8                | Harburgite | 34°05.946'   | 76°46.010'  | 4630         |
| 35      | Bhumiktse valley                        | B/BHU/16    | J3                | Basalt     | –            | –           | –            |
| 36      | From Photoksar to Wanla - Sisir La pass | SP3/15/14   | SP3               | Dunite     | 34° 06. 410' | 76° 46.766' | 4654         |

by albitization of feldspars, uralitization of pyroxenes and presence of secondary epidote and quartz (Fig. 3h).

## 4. Analytical methods

### 4.1. Whole-rock analyses

Whole rock major- and trace-element data are presented in [Supplementary Data, Table S1](#). Whole-rock major-element analyses were obtained by X-ray fluorescence (XRF) on fusion glass disks using a Bruker S4 Pioneer model wavelength dispersive X-ray spectrometer at the National Centre for Earth Science Studies (NCESS), Kerala. The sample preparation method was carried out using a Katanax X-300 X-Fluxer triple station fusion instrument using standard procedures in the sample preparation lab of NCESS. Reliable and international calibration standards were used during the analysis. The analytical uncertainty is estimated to be  $\pm 1\%$  for all major oxides.

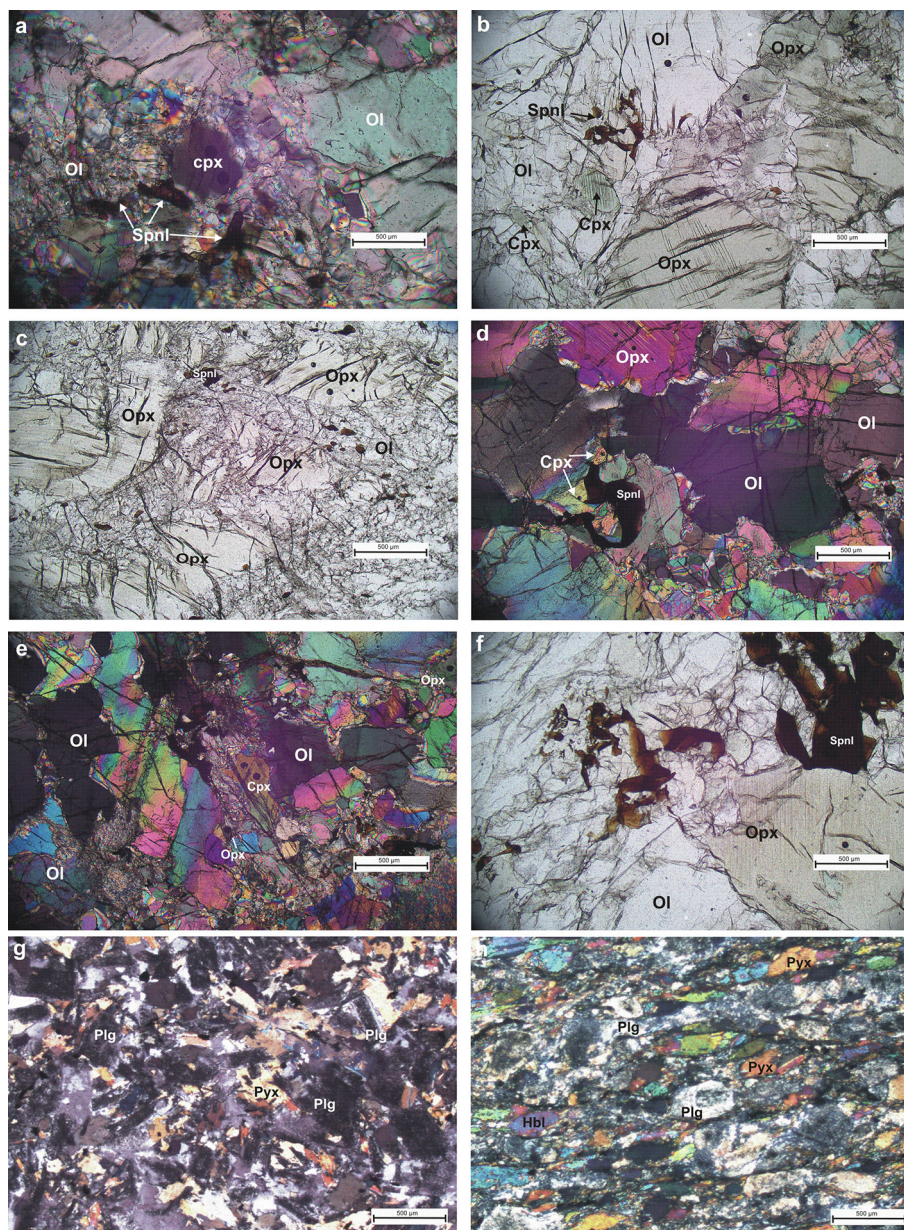
Trace elements, including rare earth elements (REE), were analyzed following detailed sample preparation and analytical procedures described by [Barrat et al. \(2007\)](#). For the mafic samples, about 100 mg of whole-rock powder was accurately weighed and transferred to 22 ml Savillex PFA beakers. An amount (between 1.5 and 15 ng) of Tm (in solution) was added to each sample, except to the blank and to the reference material BHVO-2. The sample was dissolved in a mixture 1:1 HF/HNO<sub>3</sub>. After total dissolution, samples were treated with 5 ml 2% HNO<sub>3</sub> and aliquots were taken, diluted and their concentrations determined with an

ELEMENT XR, Thermo-Scientific inductively coupled plasma mass spectrometer (ICP-MS) following [Barrat et al. \(2007\)](#) protocols. The <sup>87</sup>Rb/<sup>86</sup>Sr and <sup>147</sup>Sm/<sup>144</sup>Nd ratios were specifically monitored and international geostandards BEN, UBN and blanks were regularly run during the set. Analytical procedures for ultramafic samples have been elaborated in [Jonnalagadda et al. \(2019\)](#).

Whole-rock Sr- and Nd-isotope compositions for 12 mafic and solely Sr compositions for 14 peridotite samples were analyzed ([Table 2](#)) in the isotope laboratory at Geoscience Environment Toulouse laboratory-OMP, Toulouse France using a TRITON + Thermo-Scientific mass spectrometer (TIMS), following procedures described by Muñoz et al. (2018). Sr and Nd have been separated and purified by column chromatography from an aliquot of the solution taken after sample digestion for trace-element determination using a combination of Sr-Spec, Thru-spec and Ln-Spec resins. An equivalent of 250 ng Sr and 100 ng Nd were measured together with NBS987 and La Jolla isotopic standards. Standard analyses yielded  $0.510848 \pm 6$  (n = 50) for La Jolla and  $0.710255 \pm 20$  (n = 70) for NBS-987 which fall within the GeoRem recommended values (<http://georem.mpch-mainz.gwdg.de/>). Typical blanks are 20 pg for Nd and 150 pg for Sr. <sup>87</sup>Sr/<sup>86</sup>Sr and <sup>143</sup>Nd/<sup>144</sup>Nd ratios were normalized against <sup>86</sup>Sr/<sup>88</sup>Sr = 0.1194 and <sup>146</sup>Nd/<sup>144</sup>Nd = 0.7219.

### 4.2. Mineral chemistry

Representative mineral compositions of olivine, pyroxenes, and spinel from the investigated samples are given in [Supplementary](#)



**Fig. 3.** Microphotographs of the ultramafic and mafic rocks of the Spontang ophiolite complex. (a) Olivines displaying curvilinear boundaries displaying zoning, kink bands, and recrystallization in the harzburgites. (b) and (c) Colorless orthopyroxenes occurring as big prismatic crystals displaying exsolution lamellae, kink bands and gliding. (d) Porphyroclasts of olivine show curvilinear, serrated boundaries, with the presence of kink bands in the harzburgites (e) Interstitial clinopyroxene observed in the matrix displaying exsolution lamellae of orthopyroxenes. (f) Dunites exhibiting coarse porphyroclastic olivine grains that are elongated and display undulose extinction and intense zoning. Anhedral spinels are commonly observed along olivine boundaries. (g) Elongate plagioclase crystals with a cloudy appearance along with relic pyroxenes Fe-Ti oxides observed in the gabbros. (h) Alteration of feldspars, pyroxenes and presence of secondary epidote and quartz observed in the gabbros.

**Data, Table S2.** Major-element analyses of minerals were performed with an CAMECA SX-Five electron microprobe at the Centre de Micro-caractérisation Raimond Castaing UMS-CNRS 3623, Paul Sabatier University (Toulouse). The operating conditions were set at 15 kV with 20 or 10 nA beam currents. Natural minerals and synthetic pure oxides were used as reference materials. The electron microprobe used in this study had one energy dispersive and five-wavelength dispersive spectrometers capable of automated quantitative analysis.

Trace-element including REE concentrations of clinopyroxene, orthopyroxene and olivine were determined by laser ablation (LA)-ICP-MS via a NWR213 Laser Ablation system coupled to a Thermo ELEMENT XR ICP-MS system at Geoscience Environment

Toulouse laboratory-OMP, Toulouse, France ([Supplementary Data, Table S3](#)). The reference materials used as external standards were synthetic calcium-sodium aluminosilicate glass with a concentration of approximately 400 ppm (NIST 610) and 40 ppm (NIST 612) for a wide range of trace elements. Ca concentrations determined by electron microprobe were used as an internal standard for clinopyroxene analyses. The laser-spot diameter ranged from 30 to 50 µm, with a frequency of 10 Hz, and the fluence ranged from 5 to 10 J·cm<sup>-2</sup> (energy density). The Glitter 4.0 software ([Van Achterbergh et al., 2001](#)) was used for raw data processing. Detection limits for most elements are in the range of 10 to 60 ppb except for Sc and V (100 ppb) and Ti (2 ppm), Ni and Cr (0.7 ppm).

**Table 2**  
Rb-Sr and Sm- Nd isotopic compositions of the mafic and ultramafic rocks of the Spontang ophiolite complex.

| Sample No                                          | A2       | A1       | J3       | B8       | B3       | E2       | D14      | C10      | B15      | B5       | B12      | I6       |
|----------------------------------------------------|----------|----------|----------|----------|----------|----------|----------|----------|----------|----------|----------|----------|
| Rock Type                                          | Basalt   | Basalt   | Basalt   | Gabbro   | Gabbro   | Gabbro   | Gabbro   | Gabbro   | Gabbro   | Gabbro   | Gabbro   | Gabbro   |
| <sup>87</sup> Rb/ <sup>86</sup> Sr ICPMS           | 0.007    | 0.104    | 0.035    | 0.23     | 0.040    | 0.22     | 0.044    | 0.022    | 0.060    | 0.005    | 0.038    | 0.034    |
| <sup>87</sup> Sr/ <sup>86</sup> Sr measured        | 0.708943 | 0.704243 | 0.704863 | 0.705166 | 0.705121 | 0.704918 | 0.703795 | 0.703746 | 0.703792 | 0.705795 | 0.703703 | 0.70409  |
| error, 10–6                                        | 5        | 5        | 5        | 4        | 4        | 5        | 4        | 5        | 5        | 5        | 5        | 5        |
| <sup>147</sup> Sm/ <sup>144</sup> Nd ICPMS         | 0.216    | 0.198    | 0.201    | 0.208    | 0.217    | 0.209    | 0.206    | 0.217    | 0.213    | 0.223    | 0.224    |          |
| <sup>87</sup> Sr/ <sup>86</sup> Sr age corrected   | 0.708924 | 0.703982 | 0.704775 | 0.704599 | 0.705020 | 0.704366 | 0.703684 | 0.703690 | 0.703641 | 0.705784 | 0.703608 | 0.704004 |
| <sup>143</sup> Nd/ <sup>144</sup> Nd measured      | 0.513014 | 0.513040 | 0.513099 | 0.513087 | 0.513028 | 0.513042 | 0.513044 | 0.51304  | 0.513054 | 0.513085 | 0.513039 | 0.513040 |
| error, 10–6                                        | 3        | 3        | 5        | 20       | 11       | 3        | 4        | 5        | 5        | 4        | 5        | 9        |
| Ages (Ma)                                          | 177      | 177      | 177      | 177      | 177      | 177      | 177      | 177      | 177      | 177      | 177      | 177      |
| <sup>143</sup> Nd/ <sup>144</sup> Nd age corrected | 0.512764 | 0.512811 | 0.512866 | 0.512846 | 0.512777 | 0.512800 | 0.512806 | 0.512788 | 0.512808 | 0.512826 | 0.512779 | 0.512801 |
| eNd(t)                                             | 6.9      | 7.8      | 8.3      | 8.5      | 7.2      | 7.6      | 7.7      | 7.4      | 7.8      | 8.1      | 7.2      | 7.1      |
| Sample No                                          | C8       | B4       | E7       | B16      | D2       | B1       | D5       | E4       | B10      | B2       | D11      | E5       |
| <sup>87</sup> Rb/ <sup>86</sup> Sr ICPMS           | 0.22     | 0.21     | 0.094    | 0.12     | 0.099    | 0.080    | 0.23     | 0.053    | 0.14     | 0.14     |          | 0.062    |
| <sup>87</sup> Sr/ <sup>86</sup> Sr measured        | 0.708035 | 0.711189 | 0.709547 | 0.709507 | 0.709346 | 0.709504 | 0.708596 | 0.708915 | 0.706846 | 0.710276 | 0.711053 | 0.709599 |
| error, 10–6                                        | 6        | 11       | 12       | 23       | 12       | 7        | 12       | 16       | 11       | 13       | 30       | 5        |
| <sup>87</sup> Sr/ <sup>86</sup> Sr age corrected   | 0.707488 | 0.710653 | 0.709310 | 0.709197 | 0.709097 | 0.709303 | 0.708020 | 0.708781 | 0.706503 | 0.709931 | 0.709333 | 0.709442 |
| Ages (Ma)                                          | 177      | 177      | 177      | 177      | 177      | 177      | 177      | 177      | 177      | 177      | 177      | 177      |

5. Results

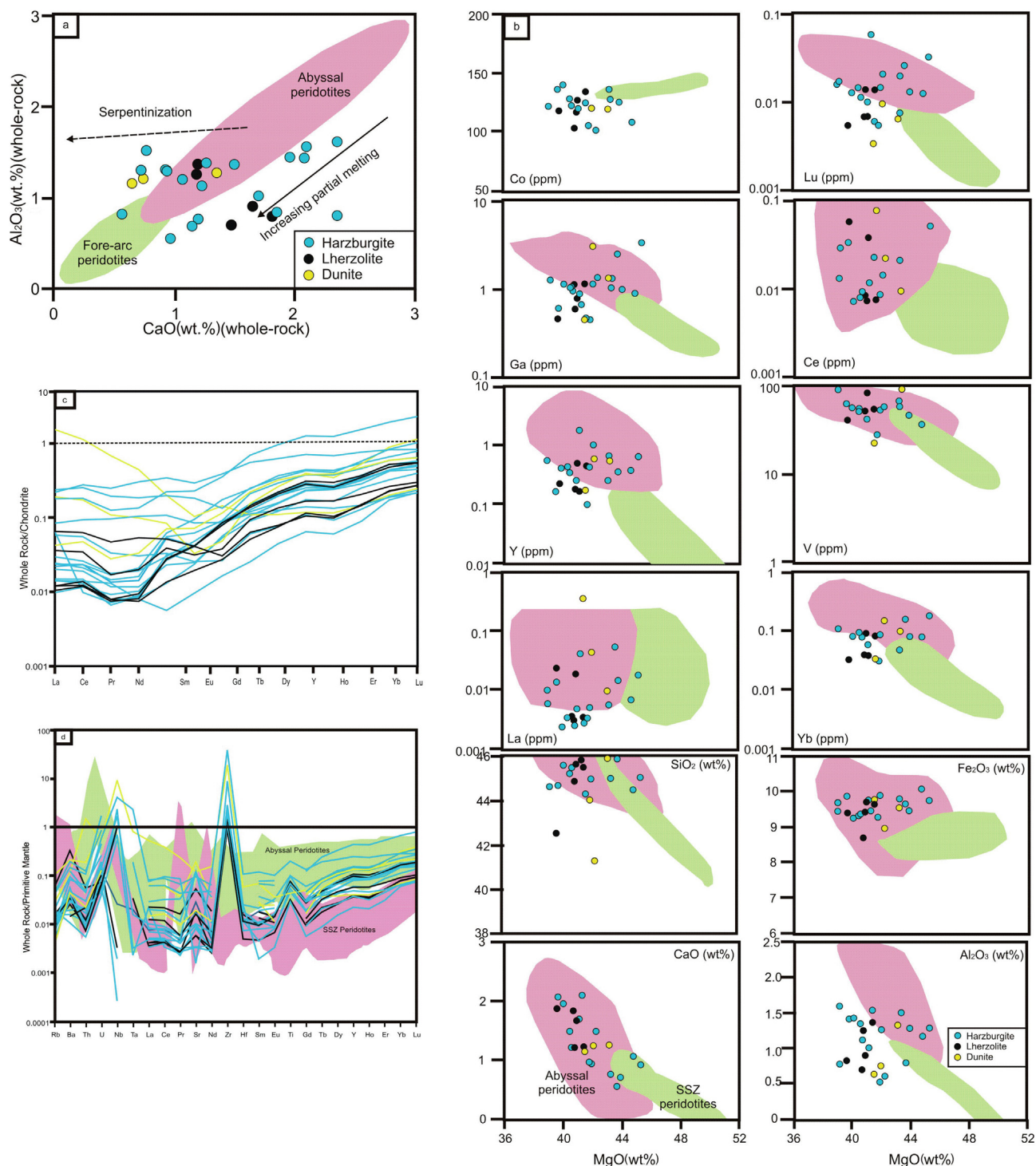
5.1. Whole-rock major- and trace-element compositions

5.1.1. Ultramafic samples

All samples are relatively fresh and show low degree of alteration as evidenced by relatively low loss-on-ignition (LOI) values (0.76 to 5.6 wt.%). The major oxides presented here are calculated on a volatile free basis to reduce the effect of element dilution related to serpentinization. The peridotites from Spontang ophiolite complex show low and restricted contents of SiO<sub>2</sub> (41.37–47.66 wt.%) with low abundances of Al<sub>2</sub>O<sub>3</sub> (0.54–1.58 wt.%), TiO<sub>2</sub> (0.02–0.07 wt.%), CaO (0.56–2.35 wt.%), moderate Fe<sub>2</sub>O<sub>3</sub> (8.7–10.1 wt.%) and high MgO (39.1–45.2 wt.%). The Mg# ([molar ratio of 100 × Mg<sup>2+</sup>/(Mg<sup>2+</sup> + Fe<sup>2+</sup>)] range from 90.4 to 91.9 and resemble residual oceanic peridotites (Bonatti and Michael, 1989; Bodinier and Godard, 2007). Lherzolites show Mg# ranging between 90.6 and 91.8 with Mg# in harzburgites between 90.4 and 91.9 with the dunites displaying Mg# between 91.0 and 92.0. Al<sub>2</sub>O<sub>3</sub> and CaO in the lherzolites vary between 0.68 to 1.36 wt% and 1.19 to 1.81 wt.%, respectively, which is higher than the harzburgites. Dunites contain the lowest Al<sub>2</sub>O<sub>3</sub> (0.66–1.36 wt.%) and CaO values (1.15–1.27 wt.%). On the basis of CaO and Al<sub>2</sub>O<sub>3</sub> concentrations (Fig. 4a), all samples plot towards the lower end of the abyssal peridotite field with a couple of samples plotting within the fore-arc peridotite field. To understand the degree of depletion, major oxide concentrations were plotted against MgO (Fig. 4b). SiO<sub>2</sub> and CaO display well defined inverse correlations whereas Fe<sub>2</sub>O<sub>3total</sub> and Al<sub>2</sub>O<sub>3</sub> show a strong positive correlation. The studied peridotites are enriched in transition elements viz. Cr (1226–12335 ppm), Ni (2076–9001 ppm) Co (102–459 ppm) and Cu (4–76 ppm), much higher than primitive mantle values (Ni > 400 ppm and Cr > 800 ppm), and show a positive correlation when plotted against MgO (e.g. Co). Chondrite-normalized patterns for the studied peridotites reflect low ΣREE concentrations of 0.05 to 0.81 ppm with depleted light rare earth elements (LREE) relative to middle and heavy rare earth elements i.e. (La/Sm)<sub>N</sub> = 0.36 to 2.4) and (La/Lu)<sub>N</sub> = 0.01 to 0.38) with an exception of the dunite sample (E5). Middle rare earth elements (MREE) are observed to be depleted relative to heavy REE (HREE; (Sm/Yb)<sub>N</sub> = 0.03 to 0.99). In the REE diagram (Fig. 4c), the peridotites show gradual decrease from Lu to Sm, however the MREE-HREE concentration in the lherzolites are lower than the harzburgites and dunites. The samples display U-shaped patterns with the lowest points at Nd and then slightly increasing from Pr to Ce and La being relatively flat. The dunites on the contrary, show increasing contents from Eu to La (except sample SP3). In a primitive mantle normalized multi-element plot, the Spontang peridotites display enriched concentrations of moderately to slightly incompatible elements (e.g. Ti and Zr, MREE – HREE). The samples are relatively enriched in large-ion lithophile elements (LILE) viz. Ba, Sr and fluid-mobile elements like U (Fig. 4d).

5.1.2. Mafic samples

Whole-rock geochemical analysis of twelve mafic samples including nine isotropic gabbro samples and three basaltic samples were performed. All samples exhibit low level of alteration (LOI = 0.53–1.01 wt.%) and are characterized by varying concentrations of SiO<sub>2</sub> (49.4–57.1 wt.%) and MgO (4.3–9.5 wt.%). Mg# for the basalts ranging from 63.8 to 70.9 and from 49 to 58 for the gabbroic rocks indicate that they derive from evolved rather than primitive magmas. TiO<sub>2</sub> concentrations vary from 0.66 wt.% to 1.74 wt.% which closely resemble mid-ocean ridge basalt (N-MORB) values (1.61 wt.% on average; Hofmann, 1988). The total alkali (Na<sub>2</sub>O + K<sub>2</sub>O) contents of the studied samples range between

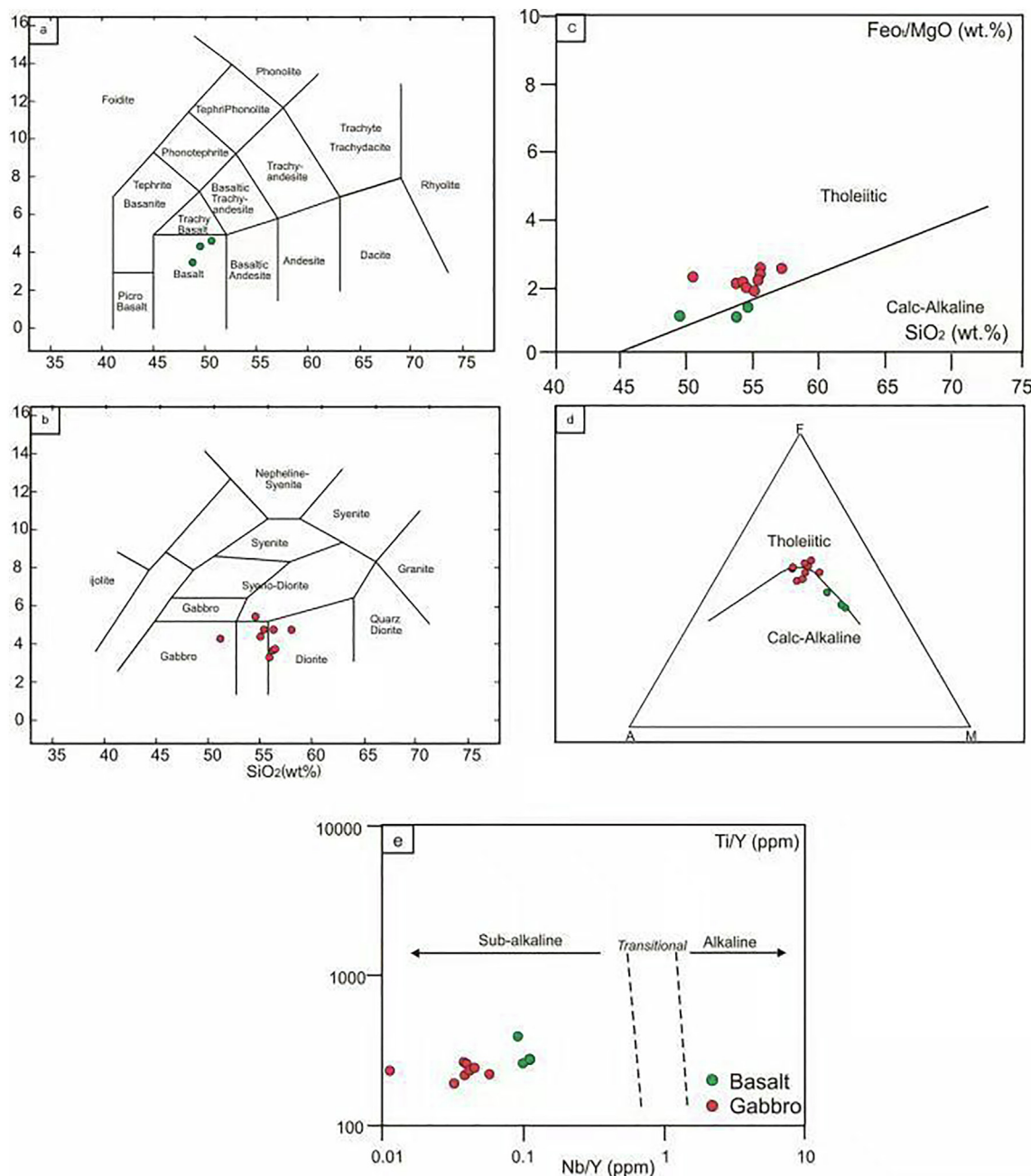


**Fig. 4.** (a) Whole-rock  $\text{Al}_2\text{O}_3$  (wt.%) versus  $\text{CaO}$  (wt.%) plot for the Spontang peridotites. Abyssal and forearc peridotite fields, as well as the partial melting trend, are taken from Pearce et al. (1992). (b) Variation diagrams of major oxides and trace elements vs.  $\text{MgO}$  (wt.%) in bulk rock Spontang peridotite samples. Abyssal (after Niu et al., 1997) and supra-subduction zone peridotite (after Parkinson and Pearce, 1998) fields are also shown for comparison. (c) Chondrite normalized REE patterns. (d) Primitive mantle normalized multi-element spider diagrams of the peridotites of the Spontang ophiolite complex. Normalizing values are after Sun and McDonough (1989). In Fig. 4d, Abyssal peridotite fields are from Hellebrand et al. (2002), Johnson and Dick (1992) and Niu et al. (1997). SSZ peridotite fields are from Parkinson and Pearce (1998).

3.7 and 6.1 wt.% and are characterized by low  $\text{K}_2\text{O}/\text{Na}_2\text{O}$  ratios of 0.01–1.8. In the Total Alkali versus Silica diagrams (TAS), the studied basaltic samples plot in the basalt fields whereas the gabbro plot in the gabbro, diorite and syeno-diorite fields (Fig. 5a, b). The gabbros when plotted in the AFM and  $\text{SiO}_2$  versus  $\text{FeO}_t$  diagrams (Fig. 5c, d) plot mainly in the tholeiitic fields with the basalts falling in the calc-alkaline fields. However, all studied mafic sam-

ples plot in the sub-alkaline area when plotted in the Nb/Y versus Ti/Y diagram (Fig. 5e) confirming their mildly alkaline nature.

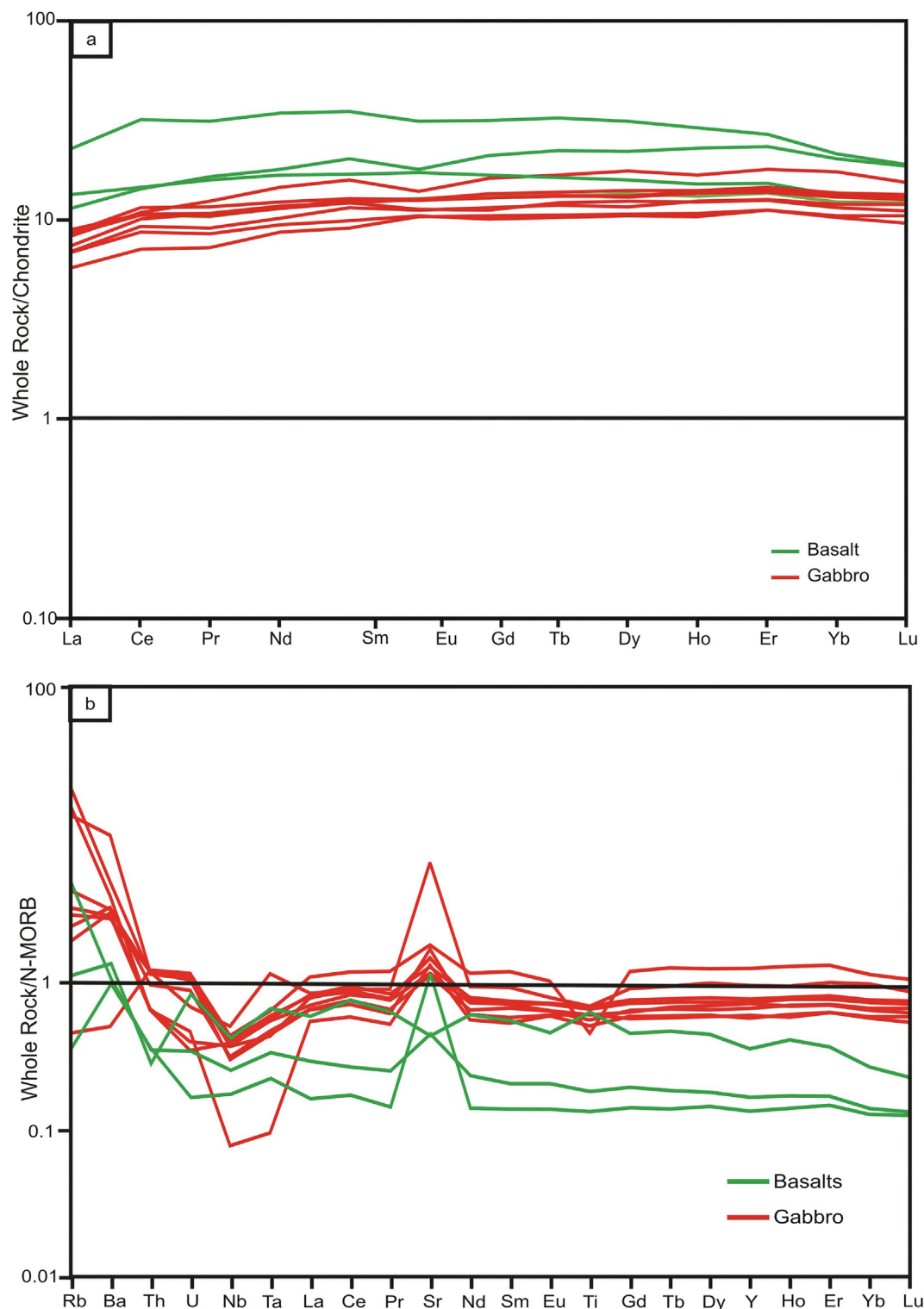
Major elements like CaO and  $\text{TiO}_2$  display positive trends when plotted against  $\text{Mg}\#$ , however  $\text{SiO}_2$  and  $\text{Fe}_2\text{O}_3$  displaying negative trends (See Supplementary Data, Fig. S1). Transition elements like Ni are relatively lower in the studied mafic dykes (79 ppm to 149 ppm) than in primary basalts (Ni = 300–400 ppm; Frey et al.,



**Fig. 5.** (a) Classification diagrams of the mafic rocks of the Spontang ophiolite complex. (b) Total alkali versus Silica diagrams (after Le Bas et al., 1986; Cox et al., 1979). (c) SiO<sub>2</sub> (wt.%) versus FeO/MgO. The boundary line between tholeiite and calc-alkaline rock types is from Miyashiro (1974). (d) AFM ternary diagram (Irvine and Baragar, 1971). (e) Nb/Y versus Ti/Y diagram.

1978) whereas Cr contents range between 261 and 347 ppm for the investigated basalts. Mg# when plotted against Zr shows a positive correlation for all samples. Cr and Ni concentrations of the gabbros range from 6 ppm to 58 ppm and 14 ppm to 31 ppm respectively and show a positive trend when plotted against Mg# (See Supplementary Data, Fig. S1) indicating clinopyroxene fractionation. Gabbros have variable total rare earth element ( $\Sigma\text{REE}$ ) concentrations ranging from 9.46 to 19.3 ppm when compared with the basalts (12.3 to 29.3 ppm). Chondrite-normalized patterns for the studied mafic rocks display smooth flat patterns showing enrichment up to 10 to 30 times that of chondrite values ( $\Sigma\text{REE}_N = 9$  to 29) (Fig. 6a). The gabbros are LREE-depleted ( $(\text{La}/\text{Yb})_N = 0.47$  to 0.76 and  $(\text{La}/\text{Sm})_N = 0.52$ –0.73) with slightly elevated patterns

depicted by the basalts ( $(\text{La}/\text{Yb})_N = 0.71$ –1.06 and  $(\text{La}/\text{Sm})_N = 0.64$ –0.79). REE abundances in gabbros are observed to increase from MREE to HREE ( $(\text{Dy}/\text{Yb})_N = 0.97$ –1.09) similar to normal (N)-MORB like patterns, with the basalt samples showing slightly elevated HREE patterns ( $(\text{Dy}/\text{Yb})_N = 1.09$ –1.46). All samples exhibit no Eu anomaly ( $\text{Eu}/\text{Eu}^* = 0.97$  on average) indicating absence of plagioclase fractionation. In the N-MORB-normalized incompatible trace-element diagram (Fig. 6b), the gabbro and basalt samples exhibit enriched LILE (Rb, Ba and Sr) typical of subduction related magmas (Pearce and Robinson, 2010). Gabbros exhibit negative Nb-Ta-Ti depletions along with Zr and Hf (not shown) whereas the basalt samples show negative Nb and positive Ta and Ti anomalies.



**Fig. 6.** (a) Chondrite normalized REE patterns. (b) Primitive mantle normalized multi-element spider diagrams of the mafic rocks of the Spontang ophiolite complex. Normalizing values are after [Sun and McDonough \(1989\)](#).

## 5.2. Whole-rock Sr-Nd isotopic results

A total of twenty-nine samples consisting of fifteen mafic and fourteen ultramafic samples were selected for Sr and Nd isotopic analyses. Three basalts, nine gabbros, one dunite, nine harzburgites and four lherzolites were processed. For the ultramafic series, only Sr isotopes have been determined, the REE content being too low to

allow the extraction of enough Nd for isotopic measurements. The  $^{87}\text{Sr}/^{86}\text{Sr}$  and  $^{143}\text{Nd}/^{144}\text{Nd}$  have been corrected for radioactive decay using an age  $t = 177$  Ma, which is the age proposed by [Pedersen et al. \(2001\)](#) for the Spontang ophiolitic formations. All mafic samples having low  $^{87}\text{Rb}/^{86}\text{Sr}$  (typically below 0.2), and similar  $^{147}\text{Sm}/^{144}\text{Nd}$ , the age corrections have either little or no effect on the corrected isotopic ratios. Except for one sample, (sample

A2, 0.7089), basalts and gabbros display relatively close  $^{87}\text{Sr}/^{86}\text{Sr}(t)$  ratios (0.7036 to 0.7048) and radiogenic Nd isotopic compositions ( $\epsilon_{\text{Nd}}(t)$  from +6.3 to +8.3), the sample A2 having the least radiogenic value. Ultramafic samples are characterized by elevated initial Sr isotopic ratios (0.7056 to 0.7106), which are partly correlated to  $1/(\text{Sr})$  (figure not shown). For those samples, part of their initial Sr isotopic compositions is higher than the  $^{87}\text{Sr}/^{86}\text{Sr}$  cretaceous seawater signature. These relatively radiogenic initial  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios can be related to an underestimation of the actual time-integrated  $^{87}\text{Rb}/^{86}\text{Sr}$  owing to a recent change in their Rb/Sr ratio, as Rb was likely to be preferentially remobilized during greenschist facies metamorphism. Crustal fluid contaminations during the obduction processes may also change the  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios of the samples, especially the depleted ones, towards more radiogenic values.

Mafic samples are plotted in the  $\epsilon_{\text{Nd}}(t)$  vs  $^{87}\text{Sr}/^{86}\text{Sr}(t)$  diagram in order to compare our data to those published, of same petrological affinities, along the Indus Suture Zone (Fig. 7). Both mafic (basalts, andesites and gabbros) and published ultramafic samples define different tendencies as observed in this figure, highlighted by previous authors (Bignold and Treloar, 2003; Bignold et al., 2006; Dhuime et al., 2007, 2009; Zhu et al., 2007, 2008, 2009; Lee et al., 2012; Kang et al., 2014). The positioning of our newly measured samples comparatively to these tendencies will be discussed below. However, it is worth noting that the Spontang samples fall within the most radiogenic reported samples for  $\epsilon_{\text{Nd}}(t)$  along the Indus Suture Zone and display typical shift towards radiogenic  $^{87}\text{Sr}/^{86}\text{Sr}(t)$  classically attributed to sea water contamination (Lanphere et al., 1981; McCulloch et al., 1981; Benoit et al., 1999). Finally, despite the fact that they display  $\epsilon_{\text{Nd}}(t)$  values close to the most radiogenic data obtained by Bignold et al. (2006) and

Dhuime et al. (2007, 2009) for the Kohistan island-arc formations, they also fall within the radiogenic part of Indian MORB domain (Mahoney et al., 1998, 2002), indicating juvenile characteristics of their source.

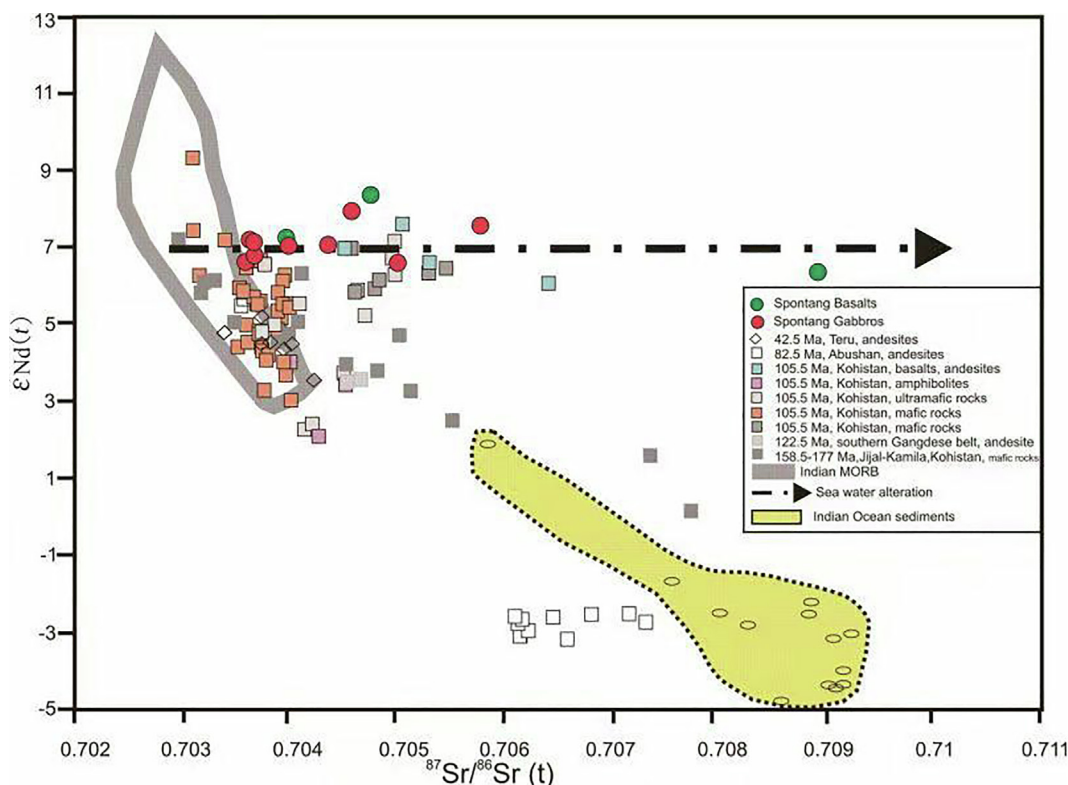
### 5.3. Mineral chemistry

#### 5.3.1. Olivine

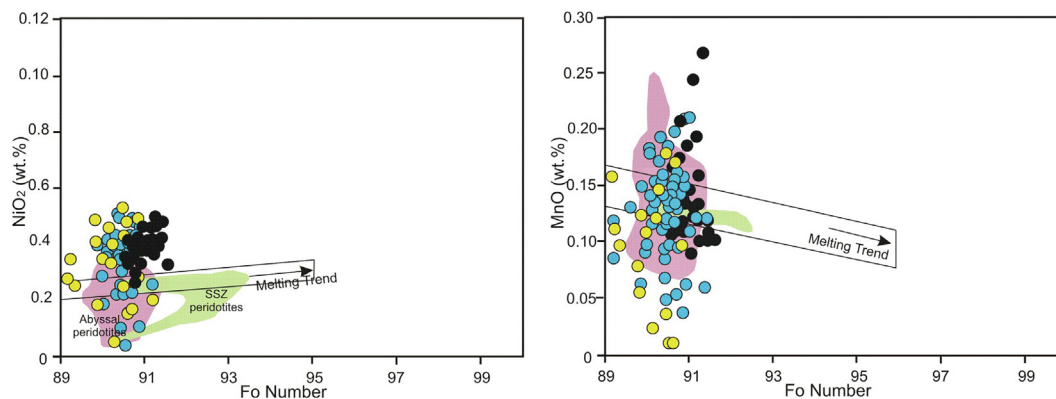
Olivine from the Spontang peridotites is forsteritic and show limited compositional range from  $\text{Fo}_{87.2-91.5}$ , avg.  $\text{Fo}_{90.3}$  close to those of olivine from abyssal peridotites (average Fo content = 91; Dick and Bullen, 1984). Olivines from both lherzolites and harzburgites show similar Mg# between 90 and 93 while dunites display Mg# contents of 88–92. NiO contents in olivine are variable and range between 0.34 and 0.51 wt.% and 0.25 and 0.49 wt.% in harzburgites and lherzolites, respectively while dunitic olivines display a wider range from 0.04 to 0.53 wt.%. Ca contents in all the studied samples are very low ( $\sim 0.07$  wt.%). MnO contents in lherzolitic olivine range between 0.09 and 0.27 wt.% versus 0.09 to 0.19 wt.% in harzburgites. MnO contents of dunitic olivine vary from 0.01 to 0.22 wt.%. The Fo contents display a positive correlation with NiO (Fig. 8a) and are negatively correlated with MnO (Fig. 8b).

#### 5.3.2. Pyroxenes

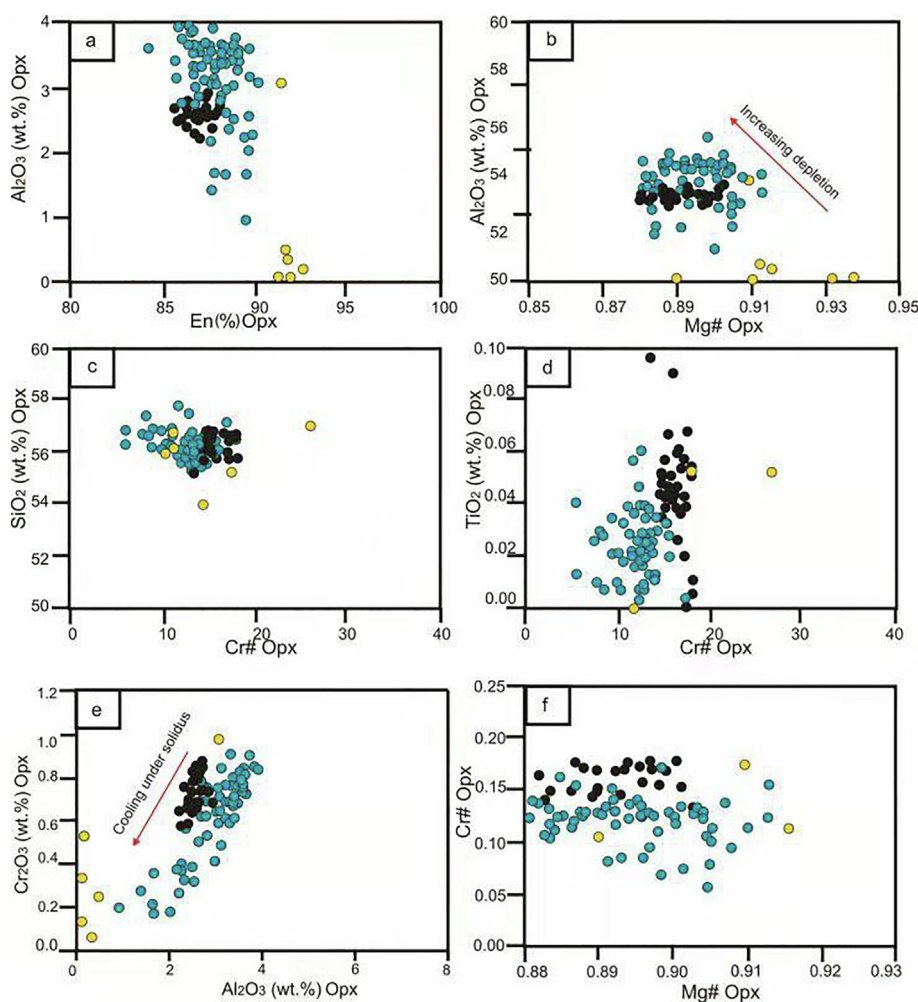
Pyroxenes from the studied peridotites are of enstatite and Cr-diopside varieties according to the nomenclature of Morimoto et al. (1988). Enstatite contents range between 84.1 and 90.2 and 85.7 to 89.9 in the harzburgites and lherzolites respectively, while dunitic orthopyroxenes display En contents ranging from 85.6 to 93.7. Similarly, Mg# of orthopyroxene in harzburgites (88.1 to



**Fig. 7.**  $\epsilon_{\text{Nd}}(t)$  vs  $^{87}\text{Sr}/^{86}\text{Sr}(t)$  diagram for Spontang basalts and gabbros, superimposed on available isotopic data for mafic and ultramafic rocks from the ITSZ (Khan et al., 1997, 2004; Bignold and Treloar, 2003; Bignold et al., 2006; Dhuime et al., 2007, 2009; Zhu et al., 2009; Kang et al., 2014; Li et al., 2015). All isotopic data have been age-corrected according to their measured or estimated emplacement ages (177 Ma, Pedersen et al., 2001 for Spontang samples). Actual, Indian MORB domain and actual sediments, extracted from Earthchem database (<http://www.earthchem.org/>) are also reported. The dotted lines illustrated the Cretaceous Sea water contamination trend (triple bold), the Oceanic Arc (bold dotted) and Continental (double dotted) mixing trends.



**Fig. 8.** Compositional variation plots of olivine (Fo content) versus NiO(a), and MnO(b). Partial melting trends are from Ozawa (1994), abyssal peridotite and fore arc peridotite are from Pagé et al. (2008).

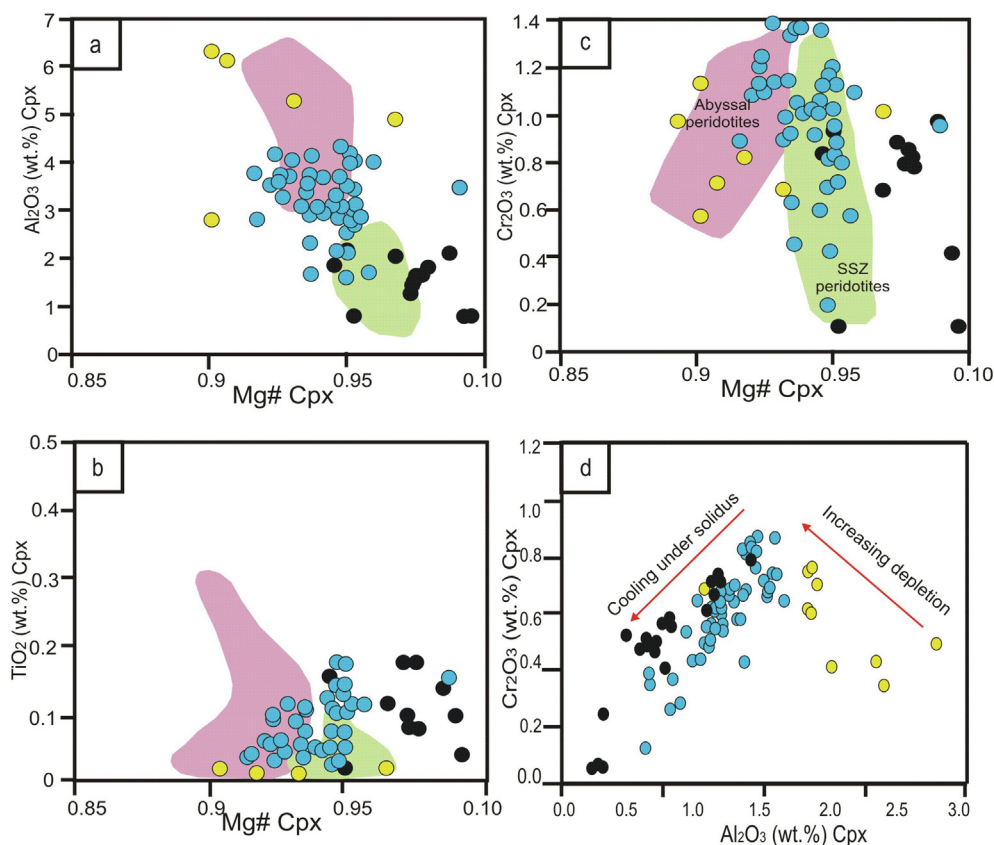


**Fig. 9.** Compositional variation plots of orthopyroxene (a) En% versus Al<sub>2</sub>O<sub>3</sub> (wt.%); (b) Mg# versus Al<sub>2</sub>O<sub>3</sub> (wt.%), Cr# versus; (c) SiO<sub>2</sub> (wt.%) and (d) TiO<sub>2</sub> (wt.%). (e) Al<sub>2</sub>O<sub>3</sub> (wt.%) versus Cr<sub>2</sub>O<sub>3</sub> (wt.%). (f) Mg# versus Cr#.

91.3) is relatively higher than that in lherzolites (88.1–90.5). Orthopyroxene Al<sub>2</sub>O<sub>3</sub> contents are variable and range between 1.63 and 3.89 wt.% in the harzburgites, with relatively close values in the lherzolites (2.21–2.93 wt.%) and are negatively correlated with En and Mg# content (Fig. 9a, b). Orthopyroxenes from lherzolites display similar Cr<sub>2</sub>O<sub>3</sub> and MnO contents (0.37–0.87 wt.%; 0.09–0.19 wt.%) comparable to the harzburgites (0.22–0.91 wt.%;

0.08–0.20 wt.%) and exhibit higher CaO contents (0.32–0.68 wt.%) when compared with the harzburgites (0.26–1.30 wt.%).

Cr# (Cr# = (Cr/(Cr + Al))) are high in lherzolites (0.13–0.18) compared to the harzburgites (0.074–0.17) and are negatively correlated with SiO<sub>2</sub> (Fig. 9c) and Mg# (Fig. 9f). TiO<sub>2</sub> ranges between 0.03 and 0.11 wt.% in both peridotite varieties and show a positive correlation when plotted against Cr# (Fig. 9d). In both peridotite types,



**Fig. 10.** Compositional variation plots of clinopyroxene (a) Mg# versus Al<sub>2</sub>O<sub>3</sub> (wt.%), (b) TiO<sub>2</sub> (wt.%), (c) Cr<sub>2</sub>O<sub>3</sub> (wt.%), (d) Al<sub>2</sub>O<sub>3</sub> (wt.%) versus Cr<sub>2</sub>O<sub>3</sub> (wt.%). Abyssal peridotites are after Johnson et al. (1990) and forearc peridotites are after Ishii et al. (1992).

the opx grains have lower Al<sub>2</sub>O<sub>3</sub> (Fig. 11a) and Cr<sub>2</sub>O<sub>3</sub> (Fig. 10d) contents and high Mg# values (Fig. 9e, f) possibly as a result of re-equilibration during subsolidus cooling (Xiong et al., 2017).

Clinopyroxenes from the studied peridotites are chemically homogeneous diopsides with average compositions of En<sub>45-74</sub>Wo<sub>27-51</sub>Fs<sub>6-15</sub>. Clinopyroxenes from lherzolites display high Al<sub>2</sub>O<sub>3</sub> (1.68–4.40 wt.%) and Cr<sub>2</sub>O<sub>3</sub> contents (0.41–1.30 wt.%) comparable to those of harzburgites (Al<sub>2</sub>O<sub>3</sub> = 1.64–4.25 wt.%; Cr<sub>2</sub>O<sub>3</sub> = 0.20–1.44 wt.%). The Mg# values of clinopyroxenes in harzburgites show a more varied range from 88.7 to 96.5 in comparison to the Mg# displayed by clinopyroxenes from the lherzolite samples (89.8–95.1). In both peridotite types, the cpx grains have lower Al<sub>2</sub>O<sub>3</sub> (Fig. 10a) and Cr<sub>2</sub>O<sub>3</sub> (Fig. 10d) contents and high Mg# values (Fig. 10a–c) possibly as a result of re-equilibration during subsolidus cooling (Xiong et al., 2017). Cr# in Cpx from harzburgites (0.78–0.21) are lower than those of the lherzolitic Cpx (0.74–0.31). TiO<sub>2</sub> content of clinopyroxenes is positively correlated with Mg# and is higher in the lherzolites (0.09 wt.% to 0.21 wt.%) than in the harzburgites (0.03 wt.% to 0.18 wt.%) (Fig. 10c). Similarly, Na<sub>2</sub>O content of clinopyroxene in lherzolites is higher (0.08–0.53 wt.%) than those in harzburgites (0.002–0.30 wt.%). Dunitic clinopyroxenes show negligible TiO<sub>2</sub> contents and low Na<sub>2</sub>O contents.

### 5.3.3. Spinel

Detailed presentation of chemical compositions of spinels from the studied peridotites is available in Jonnalagadda et al. (2019) and is summarized below. Spinel is classified as spinel – spinel-hercynite and display high Cr<sub>2</sub>O<sub>3</sub> and low TiO<sub>2</sub> contents confirming their depleted character. Al<sub>2</sub>O<sub>3</sub> contents vary between

37.60–43.9 wt.% in the spinel from dunites while those from the lherzolites and harzburgites display Al<sub>2</sub>O<sub>3</sub> values ranging from 23.8 to 35.6 wt.% and from 33.2 to 62.4 wt.%, respectively. Cr# in the spinel's from lherzolites are moderately low (0.10–0.34) especially in comparison to Cr# from dunitic spinels (Cr# = 0.79). Mg# decreases with increasing Cr# (Fig. 11a). Based on the Cr# versus Fo content of spinel – olivine pairs, all peridotite samples plot within the mantle array (Arai, 1994a, b) (Fig. 11b). Low TiO<sub>2</sub> contents observed in the spinels of the studied peridotites (0.1–0.3 wt.%) are consistent with their residual nature.

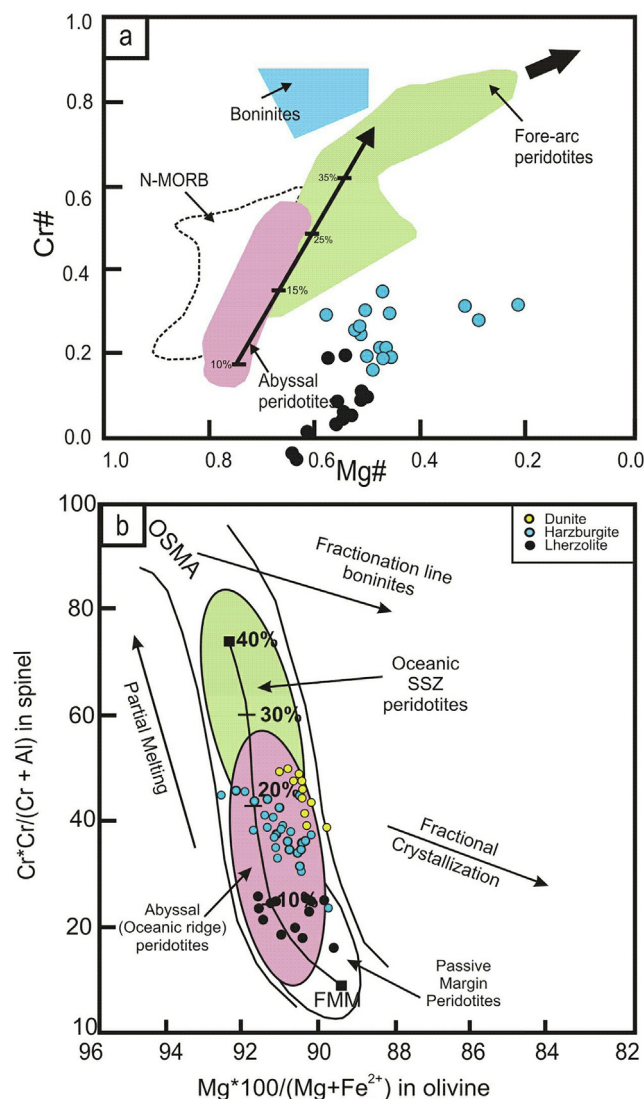
### 5.4. Mineral trace-element compositions

#### 5.4.1. Olivine

Studied olivines are characterized by low Ca (177 to 362 ppm) and Ti contents (3.82 to 9.26 ppm) and fall within the established range of mantle derived olivines (Ca > 700 ppm, Ti > 70 ppm; Foley et al., 2013). Trace-element compositions of olivine from the studied peridotites display very low concentrations close to the detection limits and hence the observations are not relevant and not discussed further.

#### 5.4.2. Clinopyroxenes

Trace-element including REE compositions of clinopyroxenes from the Spontang lherzolites and harzburgites differ from those of cpx from depleted MORB-source mantle (DMM). In the chondrite-normalized REE diagram (Fig. 12a), clinopyroxenes from the harzburgites and lherzolites display smooth patterns with depleted LREE and a gradual increase in MREE with enriched HREE above chondritic values. In the primitive-mantle-normalized dia-



**Fig. 11.** Spinel compositions of the Spontang peridotite samples plotted on (a)  $Mg\#$  [ $100 \times Mg/(Mg + Fe^{2+})$ ] vs.  $Cr\#$  [ $100 \times Cr/(Cr + Al)$ ] diagram. Data for spinel in modern abyssal peridotites is from Dick and Bullen (1984) and Juteau et al. (1990). Field for spinel in boninites is taken from Dick and Bullen, (1984). Data for spinel in fore-arc peridotites are from Ishii et al. (1992) and Ohara and Ishii (1998). Partial melting trend (black arrow) is from Arai (1994). (b)  $Cr\#$  in spinel versus  $Fo$  content of olivine in peridotites of Spontang ophiolite. Fields for spinels occurring in abyssal (and oceanic ridge), oceanic SSZ and passive margin peridotites are after Dick and Bullen (1984) and Pearce et al. (2000). OSMA means Olivine Spinel Mantle Array and fractionation line of boninites are after Arai (1994a, b).

gram (Fig. 12b), clinopyroxenes show overall depleted patterns with negative Ba, Th, Sr and Zr anomalies and slightly positive anomalies in U, Ta, Nb, Hf and Ti.

#### 5.4.3. Orthopyroxenes

Orthopyroxenes of the Spontang peridotites are more depleted than the DMM. HFSE of orthopyroxenes from both peridotite varieties pronounced depletion in Zr, Nb, Ta and positive enrichments in Ti and Hf. In the chondrite-normalized REE patterns (Fig. 12c), orthopyroxenes from the lherzolites and harzburgites are characterized by depleted LREE and a gradual increase from MREE to HREE. In the primitive-mantle-normalized diagram (Fig. 12d), all samples show obvious negative anomalies of Ba, La, Sr, Zr while U, Hf and Ti, show positive anomalies.

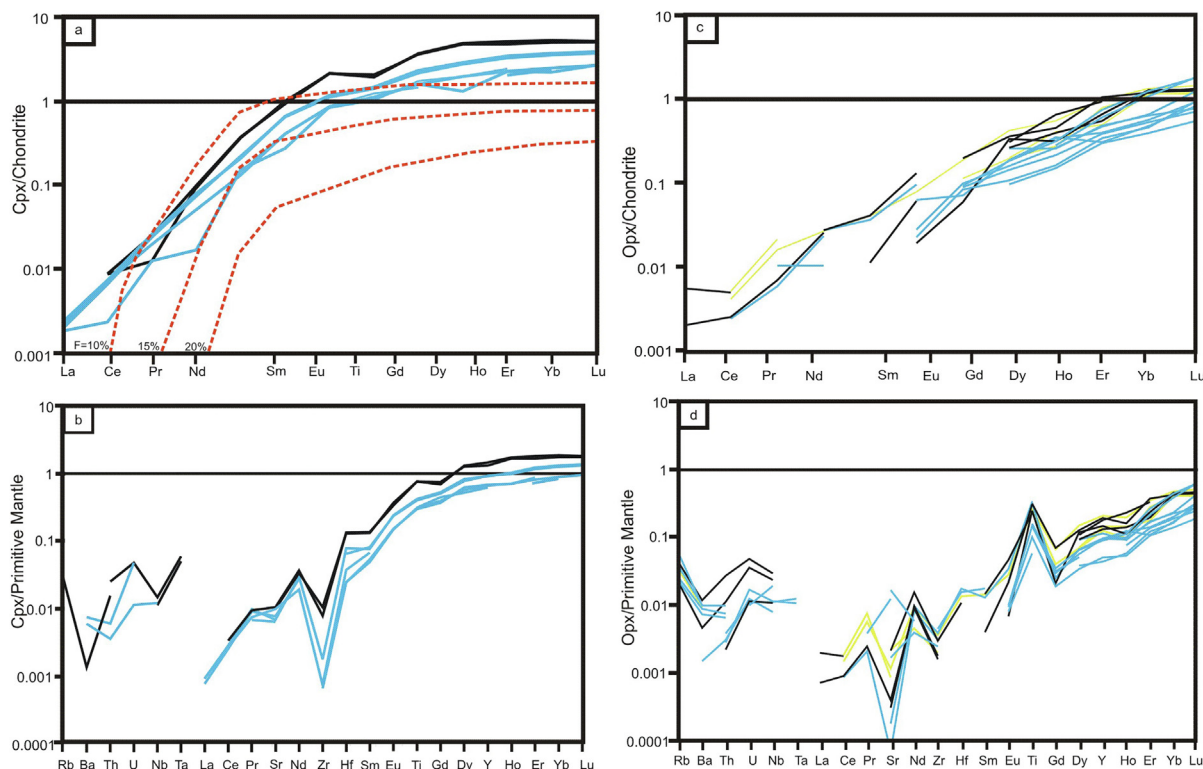
#### 5.5. Temperature estimates

Equilibration temperatures for the Spontang peridotites have been calculated using co-existing mantle phase compositions. The Spontang peridotites are plagioclase- and garnet-free spinel-facies peridotites for which no robust geobarometers exist. The following thermometers have thus been used assuming a pressure of 1.5 GPa typical of upper mantle peridotites within the spinel stability field (Gasparik, 1987). (1) Fe-Mg exchange between Opx and Cpx ( $T_{BK90}$ ; Brey and Köhler, 1990); (2) Ca in Opx ( $T_{BKCa90}$ ; Brey and Köhler, 1990) and (3) Fe-Mg exchange between olivine and spinel (Fabries, 1979). Using the thermometers based on the Ca and Fe-Mg exchange in pyroxenes, average temperature estimates for the lherzolites are  $T_{BKCa90} = 779$ – $1046$  °C and  $T_{BK90} = 762$ – $1054$  °C respectively. In contrast, the harzburgites record lower temperatures ( $T_{BKCa90} = 519$ – $996$  °C and  $T_{BK90} = 730$ – $1006$  °C). Using the partitioning of Mg and  $Fe^{2+}$  between olivine and spinel, the calculated temperatures range from 646 °C to 795 °C, significantly lower than the pyroxene-based thermometers. The Spontang lherzolites thus have systematically higher equilibration temperatures than coexisting harzburgites using pyroxene-based thermometers while thermometers using olivine and spinel give similar temperatures for both peridotite varieties. For the samples for which both cpx and opx trace-element compositions were available (two harzburgites and one lherzolite), the REE in two-pyroxene thermometer of Liang et al. (2013) was also applied. The temperature inversions were restricted to the HREE and Y while the LREE visibly deviate from a linear regression, most likely affected by secondary processes and/or poor analytical precision due to their low abundances in opx. The harzburgites yield  $T_{REE} = 1115$  and  $1277$  °C, but only the former value corresponds to a robust regression passing through the origin (see See Supplementary Data, Fig. S2), while the latter may indicate some extent of disequilibrium between cpx and opx. In comparison, the lherzolite yields  $T_{REE} = 1216$  °C, which is significantly higher than the reliable value obtained for the harzburgites. These relatively high temperatures potentially indicate that harzburgites and lherzolites were only little affected by sub-solidus re-equilibration. Limited cooling is also indicated by the overall agreement, at least for the lherzolites, between  $T_{REE}$  (1216 °C) and  $T_{BKCa}$  (779–1046 °C), considering that REE closure temperatures are commonly higher than major-element ones (e.g. Le Roux et al., 2016; Tilhac et al., 2016). However, this interpretation is not entirely consistent with the differences observed between  $T_{BKCa}$  and  $T_{BK}$ , although these may simply reflect pressure and compositional dependencies, such as the influence of Na on the Ca content in opx (Nimis and Grütter, 2010).

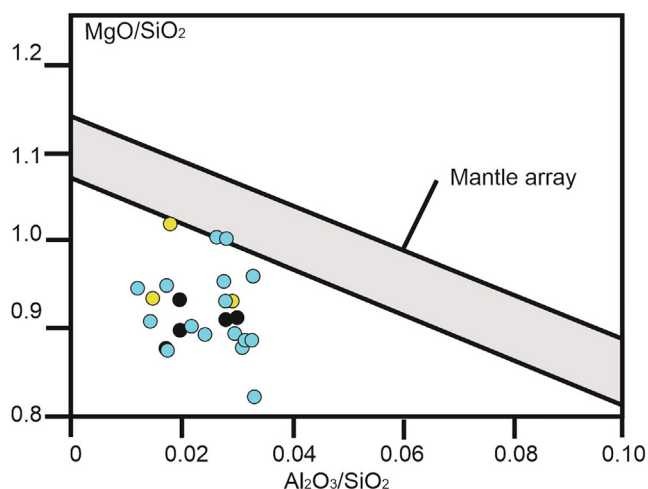
## 6. Discussion

#### 6.1. Effects of alteration

Based on petrographic and geochemical studies carried out, evidences of alteration in the studied Spontang samples cannot be ruled out leading us to assess their potential effect on the mobility of key elements. It is known that LILE concentrations such as Sr and Rb could be affected by seawater alteration and low-grade metamorphism of ocean floor rocks. In contrast, immobile elements such as high field strength elements (HFSE; Nb, Ta, Zr and Hf) and REE which commonly are less affected by low-grade hydrothermal alteration and metamorphism (Thompson, 1973; Sheraton, 1984; McCulloch and Gamble, 1991; Pearce et al., 2000) provide good insights on primary igneous processes (Pearce and Norry, 1979). In the Fig. 13, the studied mantle peridotites plot below the mantle array implying a possible loss of Mg during hydrothermal metamorphism (Niu, 2004). To test this



**Fig. 12.** (a) and (c) REE and (b) and (d) multi-element patterns of clinopyroxenes and orthopyroxenes in representative Spontang peridotites, normalized to values for chondrites and the primitive mantle Sun and McDonough (1989), respectively. Red dashed curves in Fig. 12a represent the modelled REE compositions of residual clinopyroxenes [melting degrees (F) of 10%, 15%, and 20%] after perfect nonmodal fractional melting within spinel-facies mantle, described by Hellebrand et al. (2002).



**Fig. 13.** Mantle array plot of MgO/SiO<sub>2</sub> versus Al<sub>2</sub>O<sub>3</sub>/SiO<sub>2</sub> diagram of Spontang peridotites. Mantle array is from Hart and Zindler (1986).

further, Al<sub>2</sub>O<sub>3</sub> is used as a depletion index to constrain the behavior of major elements (Snow and Dick, 1995). Major oxides like SiO<sub>2</sub> (Fig. 14) along with transition elements like Cr and Ni display a positive correlation when plotted against Al<sub>2</sub>O<sub>3</sub> whereas the negative trend observed between Al<sub>2</sub>O<sub>3</sub> and MgO is typical of residual mantle rocks (Niu, 2004). Binary plots of immobile HFSE and their neighboring REE, e.g. Hf against La, Ce and Yb in the studied peridotites indicate that LREE were immobile during alteration and primarily controlled by magmatic processes (Niu, 2004; Paulick et al., 2006) (Fig. 15). In the mafic rocks, behavior of Zr (Fig. 16) against major oxides like TiO<sub>2</sub>, Fe<sub>2</sub>O<sub>3</sub>, P<sub>2</sub>O<sub>5</sub> and REE like Y and Ce (not

shown) display good positive trends implying that these elements were immobile during alteration. Transition elements like Ni and V also display a positive correlation when plotted against Zr.

Sea water contamination is known to have an effect on Sr isotopic compositions in ophiolites, which is classically characterized by a horizontal drift in Sr-Nd space towards more radiogenic values (Lanphere et al., 1981; Benoit et al., 1999). Upper Cretaceous seawater has an isotopic Sr composition of about 0.709 (Spooner, 1976; Burke et al., 1982) and therefore, during seafloor high temperature (HT) water circulations or late HT/LP metamorphism, samples may lose their pristine magmatic isotopic signatures, according to the water/rock ratios and initial Sr concentrations. However, this low-grade metamorphism has no or little effect on the Nd isotopic ratios (Benoit et al., 1999; Nonnotte et al., 2005), which therefore can be used to discuss high-temperature mantle processes. It is worth noting that for both mafic and ultramafic series, the drift in Sr isotopic signatures is inversely proportional to their Sr concentrations, confirming that seawater contamination can be approximated by simple mixing processes. As a consequence, the ultramafic rocks (i.e. with the lowest primary Sr concentrations) are highly affected, despite their paleo-depth in the ophiolitic section.

## 6.2. Partial melting

Mineral chemistry along with whole-rock geochemical compositions of upper-mantle peridotites can be instrumental to decipher the extent of partial melting as well as melt/fluid phase enrichment and mantle-melt interaction processes subsequent to melt extraction (e.g. Johnson et al., 1990; Hellebrand et al., 2002; Seyler et al., 2007; Uysal et al., 2007; 2012; Choi et al., 2008; Dilek and Morishita 2009; Morishita et al., 2011). Based on the mineral major-element compositions, it is evident that the

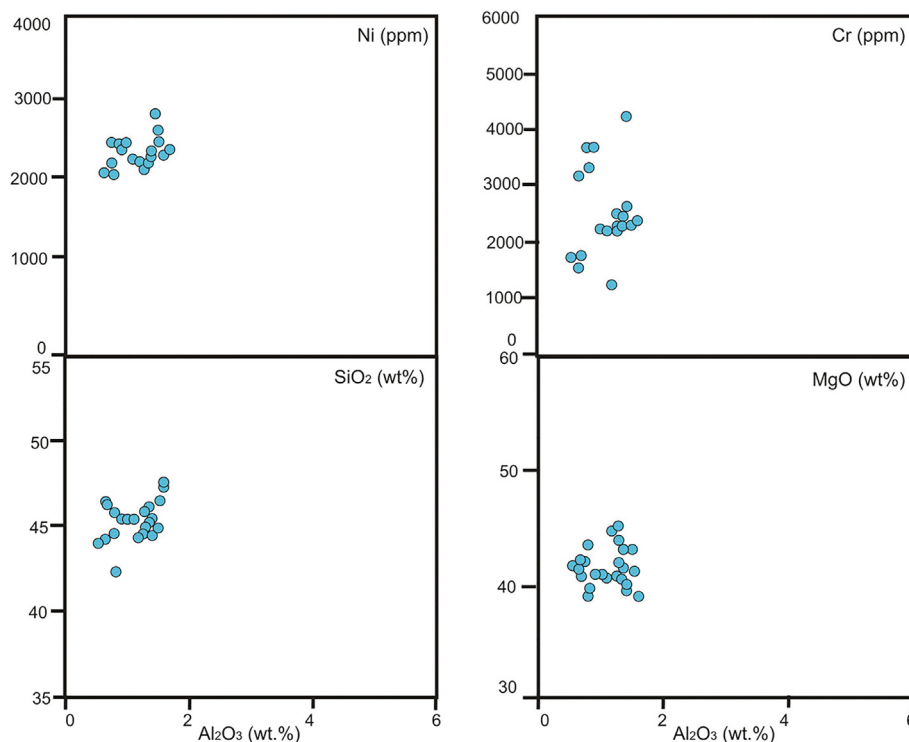


Fig. 14. Variation diagrams of major oxides, trace elements and REE versus  $\text{Al}_2\text{O}_3$  (wt.%) of Spontang peridotites.

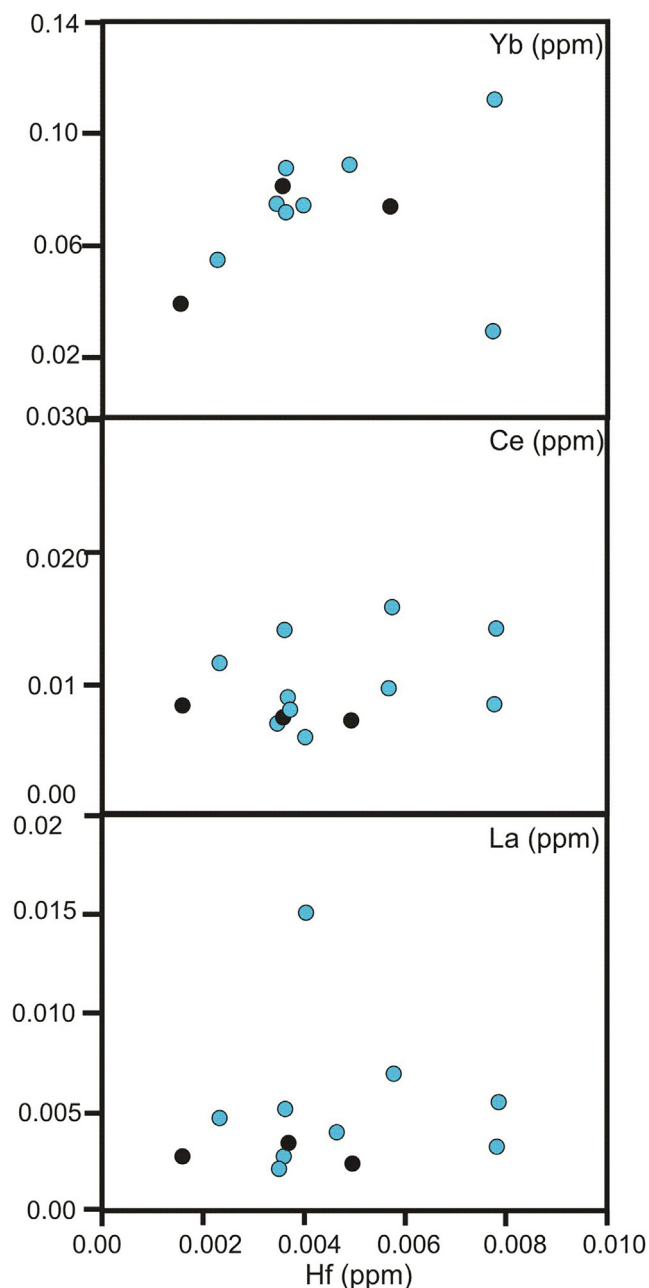
Spontang peridotites were subjected to differing degrees of partial melting (Jonnalagadda et al., 2019). Cr# contents in spinel as well as Al contents in clinopyroxene reflect the relative degree of partial melting (e.g., Hellebrand et al., 2001) and are observed to decrease with the degree of depletion (e.g. Dick and Natland, 1996; Zhou et al., 2005). Using the Hellebrand et al., (2001) equation ( $F = 10 \times \ln(\text{Cr\#}) + 24$ ), fractional melting of the studied peridotites was estimated at around 12–14% for the lherzolites and 17–21% for the harzburgites and about 23% for the dunites. These estimates of partial melting degree are similar to those obtained by using the Mg# versus Cr# plot and the Olivine-spinel mantle array plot (OSMA: Arai, 1994a, b).  $\text{Al}_2\text{O}_3$  contents of clinopyroxene within the Spontang peridotites show negative correlation when plotted against Cr# of spinel (Fig. 17a) whereas in the  $\text{Al}_2\text{O}_3$  versus  $\text{TiO}_2$  plot, the studied samples show variations consistent with an increase of the degree of partial melting (Fig. 17b).  $\text{Al}_2\text{O}_3$ ,  $\text{Cr}_2\text{O}_3$  and  $\text{TiO}_2$  contents of pyroxenes from the lherzolites are lower than those of the harzburgites with higher Mg# and Cr# values (Fig. 10). In addition, weakly to moderately incompatible elements (e.g. MREE-HREE, Y, Hf, Ti) concentrations in both pyroxenes show that the harzburgites are depleted when compared to the lherzolites suggesting primary control by partial melting process. The lherzolites resemble abyssal type peridotites which underwent depletion in a mid-ocean ridge system while the harzburgites and dunites are analogous to supra-subduction zone (SSZ) peridotites owing to the relatively high degrees of melting recorded by their whole-rock and mineral chemistry.

Trace-element contents in whole rock as well as in pyroxenes were used to perform several melting and enrichment models. We compare the studied lherzolites to closed system dynamic melting model of a PM source (after Palme and O'Neill, 2004). In the Ni/Yb versus Yb diagram (Fig. 18a), the lherzolites are residual of up to 6–8 percent partial melting of the PM whereas the harzburgites range from 8 up to 15 % partial melting. When an open-system dynamic melting models was applied using Model B (from Eiler et al., 2000, 2007), the clinopyroxenes from the

harzburgites show higher degrees of partial melting while when modelled on La/Yb versus Yb diagram (Fig. 18b) and point out around 15 to 17% partial melting. This is in-fact consistent with fractional melting results using whole rock as well as major oxide mineral chemistry. The clinopyroxenes from the lherzolites have in fact negligible La contents and hence were not shown in the above figure (Fig. 18b).

### 6.3. Melt-peridotite interaction and metasomatic signature

Mantle peridotites containing spinels with  $\text{Cr\#} > 0.70$  are believed to result from high degrees of partial melting resulting in pyroxene-free dunites (Kostopoulos, 1991). The Spontang dunites however, contain spinels with high Cr# of 0.79 and low  $\text{TiO}_2$  contents (0.05–0.1 wt.%) along with minor interstitial clinopyroxenes that may result from fertilization processes as proposed by Barth et al. (2003) and Godard et al. (2008). Relationship between spinel Cr# and trace-element compositions in the studied pyroxenes were investigated in order to understand if the peridotites were affected by metasomatic and/or partial melting processes. Incompatible trace elements like Yb, Ti, Hf and Zr contents in the clinopyroxenes from lherzolites and harzburgites display a positive correlation when plotted versus spinel Cr# indicating that the host rocks may have been affected by late-stage fluid/melt metasomatism, as un-metasomatized abyssal peridotites would ideally display gradually decreasing trends in the same plots (Bodinier and Godard, 2004; Warren, 2016) (Fig. 19). The Spontang peridotites exhibit REE patterns which are U-shaped, depleted in MREE and variably enriched in LREE, especially the harzburgites and dunites, which display enriched LREE and other highly incompatible elements (like Ba, Sr, Zr, Nb, Ti, Th, and U, the latter being a fluid-mobile element) when compared with the lherzolites. Such elevated concentrations of LREE and other incompatible elements clearly reflect that the harzburgites were affected by metasomatic enrichment; an observation a priori not consistent with their higher depleted nature, but in agreement with enrichment due to



**Fig 15.** Variation diagrams of Hf versus La (ppm), Ce (ppm) and Yb (ppm) of Spontang peridotites.

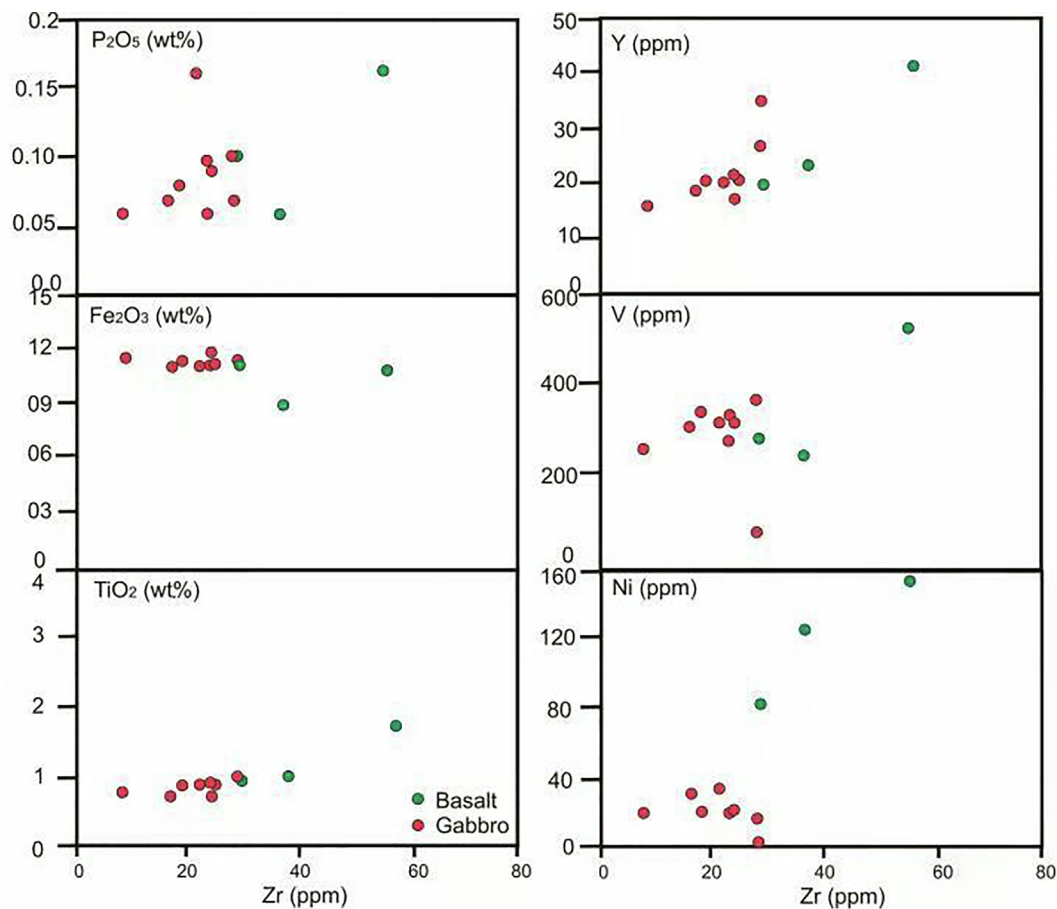
late metasomatic events. REE and other trace element compositions of both pyroxenes from the harzburgites show similar enrichments in incompatible and fluid mobile elements like Rb, Th, U, LREEs (except La), Nd, Hf, Ti with MREE-HREE's. In summary, both whole-rock trace-element and REE patterns from the Spontang peridotites indicate that the enrichment in incompatible trace elements as well as LREE reflect at least two processes i.e. mantle-melt interaction that may have resulted in U-shaped REE patterns as well as enrichment in LREE following initial partial melting (depicted by MREE-HREE patterns) by low-degree partial melts derived from a deeper mantle region and reacting with the overlying depleted mantle. Positive spikes in U, detectable Ba, depleted Zr and negligible Nb and Ta signatures observed in the pyroxenes indicate that the studied samples have been metasomatized.

#### 6.4. Nature of the magma sources

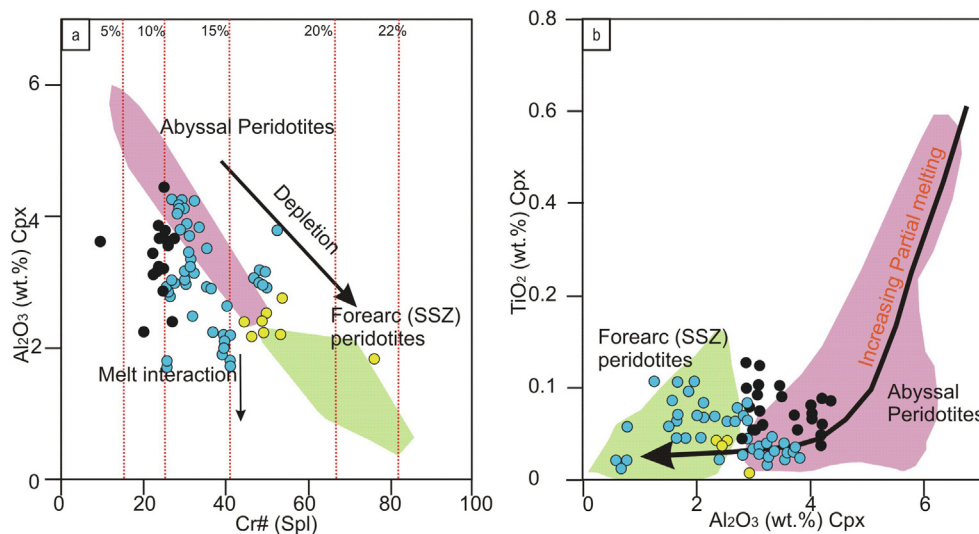
Considering their immobile nature during subduction, HFSE's such as Nb, Zr, Ta and Ti are useful to decipher the nature of the magma source (Pearce and Cann, 1973; Pearce and Peate, 1995; Xu et al., 2003; Pearce, 2005; Pearce and Stern, 2006). Geochemical signatures such as enriched LILE and depleted HFSE observed in the Spontang mafic rocks indicate that they could be derived from subduction related magmas. The studied samples exhibit minor REE fractionation with  $\text{La/Yb}_N$  ratios ranging between 0.47 and 1.06 and low  $(\text{Dy/Yb})_N$  (0.97–1.46),  $(\text{Sm/Yb})_N$  (0.88–1.63) and  $(\text{La/Sm})_N$  (0.52–0.79) ratios. In the Zr versus Nb plot, the samples are observed to plot in the depleted mantle field, indicating derivation from a depleted mantle source (Fig. 20a). When plotted in the La/Yb versus Dy/Yb diagram, the samples fall near the N-MORB domain within the spinel lherzolite fields (Fig. 20b) indicating derivation from a spinel-peridotite mantle source. This trend is consistent in the Nb/Yb versus Th/Yb,  $\text{TiO}_2/\text{Yb}$  and Zr/Yb diagrams (See Supplementary Data, Fig. S3) along with negative Na, Ta and positive LILE anomalies in gabbros indicating that some subducted material contaminated their mantle source (Pearce, 2008). The  $(\text{Ta/La})_{PM} = (0.16\text{--}1.20)$  and  $(\text{Hf/Sm})_{PM} = (0.56\text{--}0.87)$  ratios of the studied mafic rocks link them to the depleted upper mantle but indicate that the mantle source was metasomatized by subduction-related fluids (Fig. 21a). However, as shown in the Th/Zr versus Hf/Sm plot (Fig. 21b), the samples follow the melt addition trend as a result of low Th/Zr and high Hf/Sm values implying that the mantle source was enriched following metasomatic processes (Kepezhinskis et al., 1997). The Spontang mafic rocks display Sr/Nd ratios (19.06–85.16) higher than that of N-MORB (12.33; Sun and McDonough, 1989) as well as the upper crust (11.85; Xu, 2014) implying an origin from magmas originated from a source modified by slab-derived fluids. Therefore, contamination of a mantle source by addition of subducted oceanic sediment-derived melts cannot be ruled out as it can increase concentrations of relatively immobile elements like Th and LREE (Woodhead et al., 2001; Xu, 2014). Th/Yb ratios observed in the mafic samples are relatively higher than N-MORB values (0.04; Sun and McDonough, 1989) implying that the magma source was modified by sediment-derived melts.

#### 6.5. Isotopic constraints

Spontang mafic samples are characterized by systematically elevated  $\varepsilon_{Nd}(t)$  and variable  $^{87}\text{Sr}/^{86}\text{Sr}(t)$ . Compared to the available data on mafic lithologies of the Indus Suture Zone (Khan et al., 1997; 2004; Bignold and Treloar, 2003; Bignold et al., 2006; Dhuime et al., 2007, 2009; Zhu et al., 2007; 2008; Li et al., 2015), the studied samples plot along a single seawater contamination line, and do not present any clear isotopic evidence of oceanic sediments participating in their genesis (Fig. 8). It is worth noting that part of Kohistan mafic and ultramafic samples analysed by Dhuime et al. (2007, 2009) display radiogenic Nd isotopic compositions close to those from this study, typically referred to as Type-1 by these authors. In their scenario, Type-1 magmatic event corresponds to the initiation of an intra-oceanic subduction system. Therefore, a depleted mantle source slightly contaminated by sediment-related fluids cannot be ruled out for Spontang mafic samples. However, no isotopic evidence of source evolution through time, like proposed by Dhuime et al. (2009) or Bignold et al. (2006) for the Kohistan magmas, is found in the Spontang mafic samples. All analyzed basalt and gabbro samples in this present study have radiogenic Nd isotopic compositions. In that sense, the source of Spontang melts may be simply MORB-related. In order to test the hypothesis of an upper mantle which may have been contaminated by a plume, as it has been proposed for the



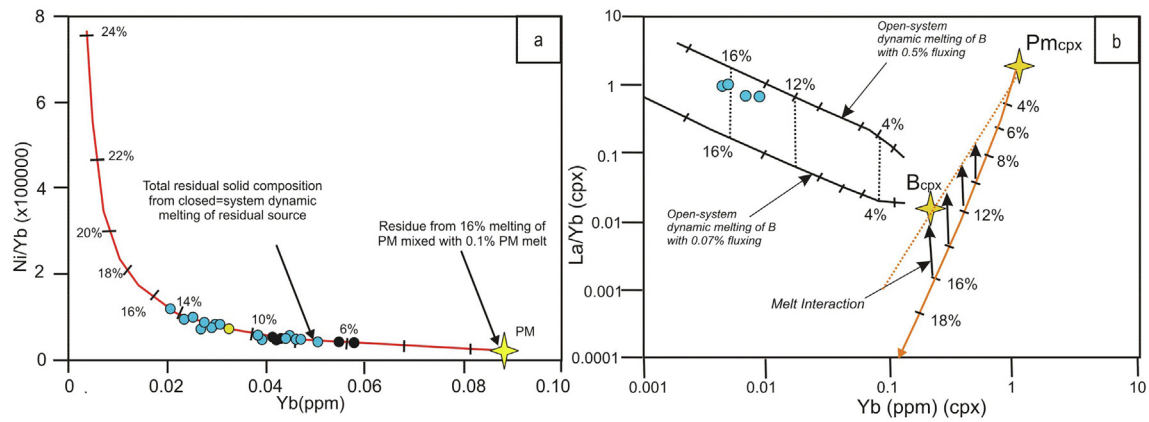
**Fig. 16.** Variation diagrams of Zr (ppm) versus major oxides and trace elements of mafic rocks of Spontang ophiolite complex.



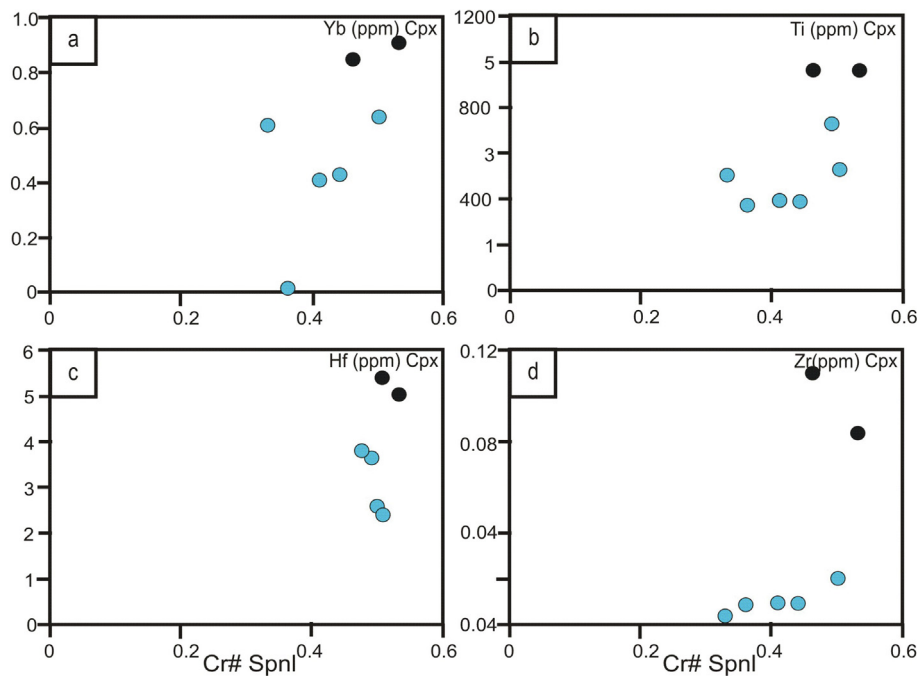
**Fig. 17.** Variation diagrams of a)  $\text{Al}_2\text{O}_3$  (wt.%) clinopyroxene versus  $\text{Cr}\#$  [ $100\text{Cr}/(\text{Cr} + \text{Al})$ ] of coexisting spinel. b)  $\text{TiO}_2$  (wt.%) versus  $\text{Al}_2\text{O}_3$  wt.% contents in clinopyroxene of the Spontang peridotites.

Cona massif by Zhu et al. (2008) for Late Jurassic or early Cretaceous mafic magmatism, we have reported our data in the  $\varepsilon_{\text{Nd}}(t)$  vs  $(\text{Nb}/\text{La})_{\text{PM}}$  plot (Fig. 22). In this diagram, the Spontang gabbros and basalts fall on the right-hand side of the MORB towards the arc trend and do not show any evidence of an upper mantle, plume-contaminated, source. Furthermore, they are characterized

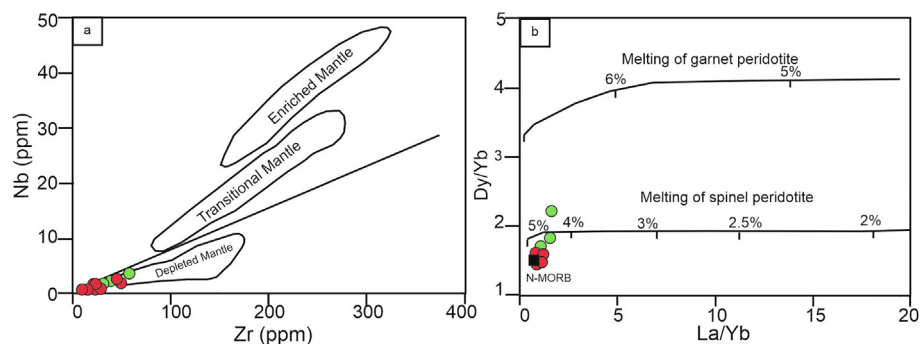
by a large range of  $(\text{Nb}/\text{La})_{\text{PM}}$  for almost similar radiogenic  $\varepsilon_{\text{Nd}}(t)$ . This range is observed in Dhuime et al. (2009) data but the increase in  $(\text{Nb}/\text{La})_{\text{PM}}$  is systematically associated with an increase in  $\varepsilon_{\text{Nd}}(t)$ . For Spontang, two samples escape this tendency and are characterized by high  $\varepsilon_{\text{Nd}}(t)$  and low  $(\text{Nb}/\text{La})_{\text{PM}}$ . This peculiar signature, observed to a lesser extent in all Spontang mafic rocks, may be



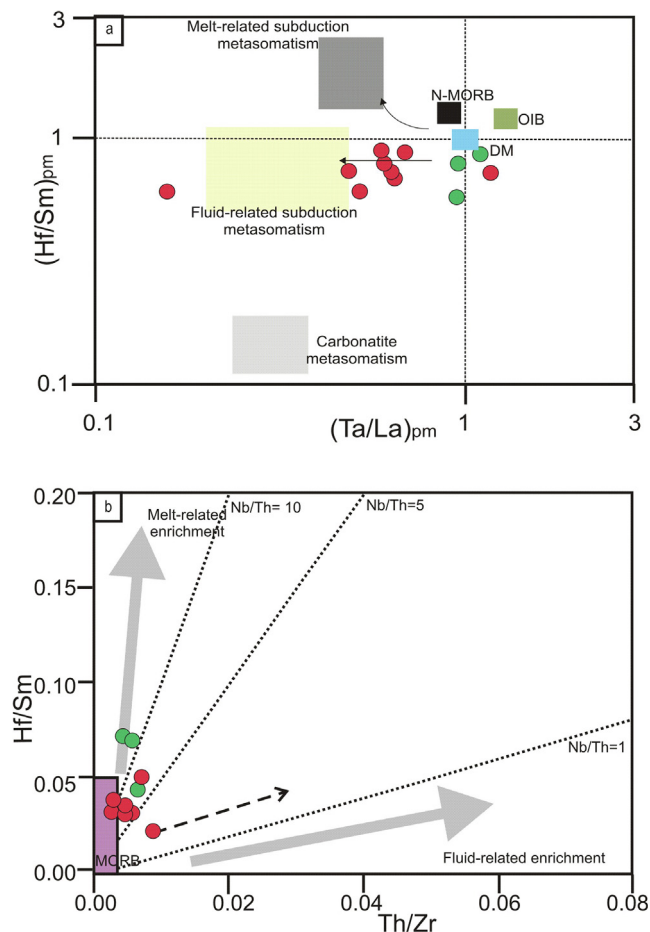
**Fig. 18.** Trace element melting models of the Spontang peridotites. a) Closed- and open-system dynamic melting of model source B are shown on (a) Ni/Yb versus Yb (ppm) and b) La/Yb versus Yb (ppm) bivariate diagram for clinopyroxene LA-ICPMS data.



**Fig. 19.** Variations of spinel Cr# versus (a and b) Yb<sub>N</sub> in Cpx and Opx (normalized by chondrite); (c and d) Ti in Cpx and Opx (ppm), (e and f) Hf in Cpx and Opx (ppm), (g and h) Zr in Cpx and Opx (ppm), (i and j) Nd in Cpx and Opx (ppm) in the Spontang peridotites.



**Fig. 20.** Zr versus Nb diagram (after Geng et al., 2011) and (b) La/Yb versus Dy/Yb diagram (after Jung et al., 2006) for the Spontang mafic rocks. The normalizing values for N-MORB are from Sun and McDonough (1989).

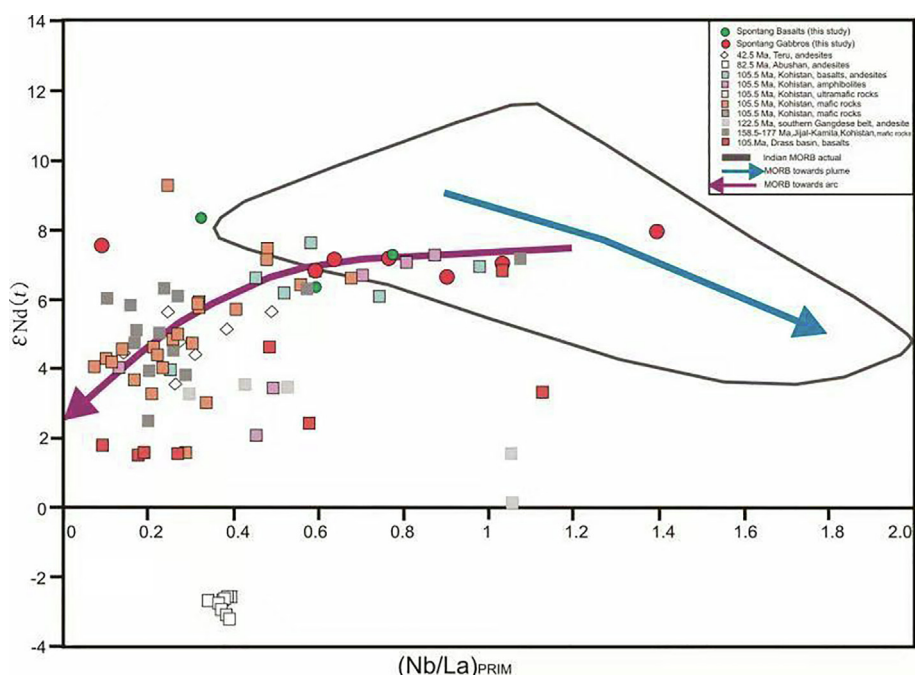


**Fig. 21.** a)  $(\text{Ta/La})_{\text{PM}}$  versus  $(\text{Hf/Sm})_{\text{PM}}$  (after Laflèche et al., 1998); b)  $\text{Th/Zr}$  versus  $\text{Nb/Zr}$  (Kepezhinskis et al., 1997) of the Spontang mafic samples.

related to the introduction of subduction-related fluids in the upper mantle source during the initiation of subduction. In this scenario, Nd isotopic composition may be buffered by the upper-mantle signature and Nb (and Ta) affected by metamorphic/magmatic processes happening during subduction.

#### 6.6. Tectonic setting of the Spontang ophiolite

Peridotites in forearc settings are known to have a variety of origins, ranging from supra-subduction zone peridotites that equilibrated with hydrous fluids and melts derived from the subducting slab to previously formed abyssal peridotites trapped within a forearc region (Parkinson et al. 1992; Parkinson and Pearce, 1998; Pearce et al., 2000). This record of multiple tectonic settings may be attributed to complex melt-rock interaction (e.g. Elthon, 1992; Michibayashi et al., 2009; Dygert et al., 2016) and/or re-melting and re-crystallization processes that constituent mineral compositions through the ascending melts (Niu, et al., 1997). Ophiolitic rocks especially from the Ladakh area of the Indian Himalayas play a key role in reconstructing the timing and processes involved during Indo-Asia collision and deciphering the nature of the Neo-Tethyan oceanic crust and upper mantle (Hebert et al., 2012; Hu et al., 2016). Formation of the Spontang ophiolites although debated, is generally interpreted to represent supra-subduction tectonic settings (e.g. Reuber, 1986; Reuber et al., 1989, 1992; Aitchison et al., 2000; Pedersen et al., 2001; Aitchison and Davis, 2004; Maheo et al., 2004; Baxter et al., 2010; Clift et al., 2014). However, based on distinct whole-rock, trace and mineral chemistry of the studied samples presented above, formation is believed to have occurred in several distinct stages. The petrographic features as well as the major and mineral chemistry of the Spontang lherzolite and harzburgite domains are distinguished from each other in terms of the extent of mantle depletion, and metasomatic history, thereby defining a multi-stage evolution of the oceanic lithosphere formed during two



**Fig. 22.**  $\epsilon\text{Nd}(t)$  vs  $(\text{Nb/La})_{\text{PRIM}}$  diagram for Spontang basalts and gabbros, superimposed on available isotopic data for mafic and ultramafic rocks from the ITSZ. Trace element data are normalized to the primitive mantle from Sun and McDonough (1989). All Nd data have end age corrected according to their measured or estimated emplacement ages. Present day Indian MORB domain is superimposed on the age-corrected data. The two arrows are illustrating the MORB toward Arc (double plain) and MORB toward Plume (dotted double) tendencies. References are same as in Fig. 8, except for Clift et al. (2000).

different episodes of lithosphere accretion from the convective mantle. The depleted lherzolites formed as products of 6%–8% partial melting of an anhydrous MOR-type melting beneath the mid-ocean ridge system and represents the first stage. This decompressional partial melting of a MOR mantle beneath a cold lithospheric lid at the spreading centre resulted in the lherzolites to be formed at high temperatures ( $T_{\text{REE}} = 1216^\circ\text{C}$ ) and probably reflect the original thermal state preserved just after their adiabatic decompression close to the estimated potential temperatures of the Phanerozoic mantle (approximately 1280–1350  $^\circ\text{C}$ ; Hezberg et al., 2010). As the subduction of the Neo Tethyan oceanic crust initiated in an intra-oceanic setting during the Middle Jurassic (177 Ma; Pedersen et al., 2001), the previously formed MOR- oceanic lithosphere was subsequently emplaced in the north-dipping subduction zone of the upper Neo-Tethyan plate. In this stage, the refractory harzburgites and dunites formed as residues after higher degrees of partial melting at lower temperatures ( $T_{\text{REE}} = 1115^\circ\text{C}$ ) of a depleted MORB source as a result of subduction initiation magmatism. Furthermore, metasomatism due to percolation of slab-derived fluids originating from the dehydrated serpentinites in the down going slab in the mantle wedge affected the host mantle rocks heterogeneously at all scales.

The Spontang massif exposes two groups of mafic rocks (viz. gabbros and basaltic units) and exhibit distinct geochemical differences i.e. MORB type of older age of 177 Ma (Pedersen et al., 2001) while the younger gabbros have primitive arc-related affinities of 136–133 Ma (Buckman et al., 2018) indicating the gabbroic magmatism with arc related affinity marks a second stage associated with the development of the Spong arc (Buckman et al., 2018). Field observations demonstrate a clear intrusive relationship between the mafic rocks and the harzburgites suggesting that accretion of the harzburgitic lithosphere must have occurred much earlier than the mafic magmatism. This inference is consistent with the Spontang harzburgite domain being located close to the crustal section within the Tethyan oceanic lithosphere as is observed in several other associated mantle – crustal units (e.g. Zedong ophiolite; Xiong et al., 2017). REE and isotopic abundances of the mafic samples analysed during the present study clearly exhibit elevated MREE–HREE abundances as well as elevated  $\varepsilon_{\text{Nd}}(t)$  and variable  $^{87}\text{Sr}/^{86}\text{Sr}(t)$  displaying distinct N-MORB affinities and correspond to the older Jurassic ( $\sim 177$  Ma; Pedersen et al., 2001) MORB-type Neotethyan oceanic crust associated with the earliest phase of subduction through which a younger intra-oceanic island arc (Spong arc) subsequently developed (Buckman et al., 2018). The mantle and crustal sections of the Spontang ophiolite were intruded by these younger gabbros with a distinct supra-subduction geochemical signature signifying an intra-oceanic subduction initiation and fore-arc spreading in  $\sim 177$  Ma oceanic crust, which led to the development of the Spong island-arc during the Cretaceous. Systematic enrichment in LILE's,  $(\text{Ta}/\text{La})_{\text{PM}}$  and  $(\text{Hf}/\text{Sm})_{\text{PM}}$  ratios for ultramafic rocks along with increase in  $(\text{Nb}/\text{La})_{\text{PM}}$  versus  $\varepsilon_{\text{Nd}}(t)$  in the gabbros confirm their upper mantle affinity and clearly indicate that the mantle source of the mafic rocks was eventually metasomatized by subduction-related fluids during the Early Cretaceous.

## 7. Conclusions

- (1) The Spontang ophiolites expose ultramafic rocks consisting of depleted lherzolites, refractory harzburgites and dunites associated with mafic rocks (viz. gabbros and basalts) exposed along the Indus Tsangpo Suture Zone (closure of the Tethyan Ocean and collision of the Indian and Eurasian tectonic plates).

- (2) Petrological and geochemical data highlight the pristine characteristics of the Tethyan ocean lithosphere, with the Spontang peridotites transitioning from mid-ocean ridge spreading center to a subduction zone setting before closure of the Neo-Tethyan basin.
- (3) Distinct major, trace element and REE patterns (whole rock and mineral patterns) along with differing thermal conditions is consistent with the presence of both abyssal and SSZ signatures.
- (4) Initiation of subduction allows slab-derived fluids to interact with the mantle wedge, resulting in contrasted, transitional, geochemical signatures for the Spontang ophiolite.
- (5) Both ultramafic and mafic rocks studied here display isotopic and trace element signatures typical from N-MORB Neo-Tethyan oceanic crust generated first in an intra-oceanic setting to further evolve into SSZ.

## Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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## Appendix A. Supplementary data

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.gsf.2021.101297>.

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