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#### Petrology of blueschist from the Western Himalaya (Ladakh, NW India): Exploring the complex behavior of a lawsonite-bearing system in a paleo-accretionary setting

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1	PETROLOGY OF BLUESCHIST FROM THE WESTERN HIMALAYA (LADAKH, NW INDIA):			
2	EXPLORING THE COMPLEX BEHAVIOUR OF A LAWSONITE-BEARING SYSTEM IN A			
3	PALAEO-ACCRETIONARY SETTING			
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#### 24 **1. Introduction**

25 Lawsonite-bearing blueschists and eclogites are witnesses of cold subduction processes occurred along ancient convergent margins. Metamorphic processes involved in the generation and preservation of 26 lawsonite are crucial in many research areas, ranging from petrology to geochemistry, geodynamics and 27 28 geophysics (e.g. Hacker et al., 2003; Bebout, 2007; Hacker, 2008; Davis, 2011; Martin et al., 2011; Vitale 29 Brovarone et al., 2011; Chantel et al., 2012; Abers et al., 2013; Cao et al., 2013; Kim et al., 2013; Spandler & 30 Pirard, 2013). Therefore, lawsonite-bearing eclogites and, to a lesser extent, lawsonite-bearing blueschists 31 have been the focus of several studies, especially in recent years (Tsujimori and Ernst, 2014 and references 32 therein). Compared to the rare occurrences of lawsonite eclogites worldwide (see the review paper by 33 Tsujimori et al., 2006), lawsonite blueschist units are reported from several orogenic belts (e.g. Agard et al., 34 2009; Tsujimori and Ernst, 2014 and references therein); however, in many cases, the lawsonite blueschist-35 facies assemblages formed at peak metamorphic conditions are widely overprinted by epidote blueschist-36 and/or greenschist-facies retrograde assemblages during exhumation (e.g. Ernst, 1988; Agard et al., 2001, 37 2006; Jolivet et al., 2003; Schumacher et al., 2008; Plunder et al., 2012). Lawsonite preservation requires 38 exhumation along cold geothermal gradients, comparable to those required for its formation during 39 subduction. Such geothermal regimes are typical of ancient Pacific-type plate convergent margins (see 40 Tsujimori and Ernst, 2014 for a review); the occurrence of well-preserved high-pressure lawsonite 41 blueschists and eclogites in an orogenic belt is therefore an appealing clue of a peculiar tectonic setting.

42 Although the Himalaya is the archetype of collisional orogens, formed as a consequence of the closure of 43 the Tethyan ocean separating India from Asia followed by continental collision between the two plates, 44 high-pressure metamorphic rocks are rare along the orogen (e.g. Lombardo and Rolfo, 2002; Guillot et al., 45 2008). Moreover, most of the eclogites reported so far from the Himalaya correspond to the 46 metamorphosed continental Indian crust dragged below Asia (NW Himalaya: Kaghan, Tso Morari and Stak 47 massifs; Pognante and Spencer, 1991; Guillot et al., 1997, 1999, 2007, 2008; de Sigoyer et al., 2000; O'Brien 48 et al., 2001; Sachan et al., 2004; Lanari et al., 2013), or underthrusted beneath southern Tibet (E Himalaya: 49 Kharta and Bhutan; Lombardo and Rolfo, 2002; Groppo et al., 2007; Chakungal et al., 2010; Grujic et al., 50 2011; Warren et al., 2011). Evidence of the ancient Tethyan oceanic crust subducted below Asia are also 51 rare and locally occur within the Indus-Tsangpo Suture (ITS) zone, which separates the northern margin of 52 the Indian plate to the south (i.e. the Himalaya s.s.) from the southern margin of the Asian plate to the 53 north (represented, from west to east, by the Kohistan Arc, the Ladakh block and the Lhasa block). These 54 evidences are: (i) few lawsonite blueschists from the western part of the ITS zone in Pakistan (Shangla: 55 Shams, 1972; Frank et al., 1977) and Ladakh (NW India) (Sapi-Shergol: Honegger et al., 1989; Zildat: Virdi et 56 al., 1977; de Sigoyer et al., 2004), interpreted as related to paleo-accretionary prisms formed in response to 57 the subduction of the Neo-Tethyan ocean below the Asian plate (e.g. Robertson, 2000; Mahéo et al., 2006; 58 Guillot et al., 2008); (ii) few eclogite, lawsonite- and epidote blueschist -facies rocks reported from the 59 Indo-Burmese Ranges (Nagaland Ophiolite Complex: Ghose and Singh, 1980; Acharyya, 1986; Chatterjee 60 and Ghose, 2010; Ao and Bhowmik, 2014; Bhowmik and Ao, 2015; Chin Hill Ophiolite: Socquet et al., 2002), 61 interpreted as the eastern extension of the ITS zone. These rare high-pressure/low-temperature (HP-LT) 62 rocks are therefore crucial for constraining the evolution of the India-Asia convergence zone during the 63 closure of the Neo-Tethyan ocean (Guillot et al., 2008); in this framework, the detailed reconstruction of 64 their P-T paths is a fundamental step toward a reliable geodynamic interpretation.

The P-T evolution of the eclogites and blueschists from the Indo-Burmese Ranges has been recently constrained by means of modern petrological methods (e.g. pseudosections); variable peak P-T conditions have been reported from different portions of the suture zone, ranging from ~340 °C, ~11.5 kbar (lawsonite blueschists: Ao and Bhowmik, 2014) to 540 ± 35 °C, 14.4 ± 2 kbar (epidote blueschists: Bhowmik and Ao,

69 2015) to 580-610°C and 17-20 kbar (eclogites: Chatterjee and Ghose, 2010). On the opposite, modern 70 petrologic studies aimed at constraining the P-T evolution of the blueschist-facies rocks from the western 71 sector of the ITS zone are lacking. Some 25 years ago, Honegger et al. (1989) reported peak metamorphic 72 conditions of 350-420 °C, 9-11 kbar for the Sapi-Shergol lawsonite blueschists using conventional 73 thermobarometry. P-T estimates for the Shangla blueschists were published even earlier (Guiraud, 1982; Jan, 1985) and suggest peak P-T conditions of ca. 400 °C, 5 kbar. Although detailed, these petrological 74 75 studies are based on conventional methods and need to be updated using more recent and powerful 76 petrological approaches (e.g. isochemical phase diagrams).

- 77 In this paper, the lawsonite blueschists from Sapi-Shergol have been petrologically re-investigated with the
- aims of: (i) constraining their P-T evolution; (ii) evaluating the influence of  $Fe_2O_3$  and of  $H_2O$  on the stability
- of the high pressure mineral assemblages; (iii) understanding the processes controlling lawsonite formation
  and preservation, and (iv) interpreting the P-T evolution of the Sapi-Shergol blueschists in the framework of
  India-Asia collision.
- 81 I 82

### 83 **2. Geological setting**

In the India–Asia convergence system, the ITS zone records the closure of the Neo-Tethyan ocean from Late Cretaceous to Tertiary time (Frank et al., 1977; Honegger et al., 1989; Cannat and Mascle, 1990). Among the few occurrences of high-pressure rocks along the ITS, those of Ladakh (NW India) are the best in terms of rock freshness, areal extent and metamorphic assemblages. Blueschists in the Ladakh area occur along the ITS in few localities: from SE to NW these are Puga, Urtsi, Hinju and Sapi-Shergol (Honegger et al., 1989). The largest outcrop is that of Sapi-Shergol (35 km south of Kargil), where the blueschists form a 12 km x 1 km E-W trending narrow zone.

91 Tectonically, the Sapi-Shergol blueschists belong to a narrow belt called "Ophiolitic Mélange Unit" 92 (Honegger et al., 1989) (Fig. 1), which outcrops over a distance of 250 km along the ITS suture. This belt 93 consists of several thrust slices sandwiched between the Nindam-Naktul-Dras nappes to the north, and the 94 Lamayuru-Karamba nappes to the south. The Ophiolitic Mélange Unit is interpreted as a relic of a paleo-95 accretionary prism formed in response to the northward subduction of the Neo-Tethyan ocean, originally 96 separating the Ladakh arc to the south from the southern Asian active margin to the north (Mahéo et al., 97 2006). This paleo-accretionary prism consists of sedimentary units including blocks of (mainly) basic 98 lithologies that have been metamorphosed under variable P-T conditions, ranging from low-grade 99 metamorphism to lawsonite blueschist -facies metamorphism (Frank et al., 1977; Honegger et al., 1989; 100 Jan, 1987; Reuber et al., 1987; Sutre, 1990; Ahmad et al., 1996; Robertson, 2000; Mahéo et al., 2006).

101 The Sapi-Shergol Ophiolitic Mélange (SSOM) is a complex unit which includes slices of the paleo-102 accretionary prism, intercalated with numerous slices of other units including the Nindam and Lamayuru 103 turbidites and low grade meta-ophiolitic slices consisting of serpentinized peridotites intruded by basic 104 dikes ("sheared serpentinites" of Robertson, 2000). The narrow blueschist zone cropping out close to the 105 village of Shergol (Fig. 1, 2a) is overlain discordantly by the Shergol conglomerate of post-Eocene (Oligo-106 Miocene?) age (Honegger et al., 1989). Blueschist lithologies are dominated by volcanoclastic sequences of 107 basic material (Fig. 2b,c) with subordinate interbedding of cherts and minor carbonatic lithologies. Mahéo 108 et al. (2006) suggested that the blueschists derive from calc-alkaline igneous rocks formed in an intra-109 oceanic arc environment. K-Ar ages of whole-rocks and glaucophane suggest an age of ca. 100 Ma for the 110 high-pressure metamorphism (Honegger et al., 1989).

111

#### 112 **2.1 Main blueschist lithologies of the SSOM**

113 Metabasic and metavolcanoclastic rocks are the dominant lithologies in the SSOM, and they are associated 114 to subordinate interbedded metasediments. These lithologies have been described in detail by Honegger et 115 al. (1989); the most relevant petrographic features are therefore only summarized here.

116

### 117 2.1.1 Metabasic and metavolcanoclastic rocks

118 Metabasic rocks are mainly represented by fine-grained glaucophane-bearing schists (Fig. 2c,e) with 119 variable amounts of lawsonite and minor clinopyroxene and phengite. Lawsonite can be either fine-grained 120 or porphyroblastic and it generally overgrows the main foliation defined by the alignment of glaucophane ± 121 phengite (Fig. 2e); where present, phengite often shows a slightly greenish pleochroism. Clinopyroxene 122 (omphacite/aegirine-augite) generally occurs as fine-grained dusty and fibrous aggregates, probably 123 replacing former magmatic clinopyroxene fenocrysts. Fine-grained titanite aggregates are often aligned to 124 the main foliation (Fig. 2e); opaque minerals can be locally abundant and surrounded by pressure fringes of 125 albite. Locally, remnants of a strongly vesciculated structure are evidenced by the alignment of fine-grained 126 titanite.

127 Metavolcanoclastic rocks are characterized by a clastic structure and consist of irregular fragments of 128 metabasic rocks set in a very fine-grained matrix (Fig. 2b). Clasts of metabasic rocks are either rounded or 129 sharp and vary in size from few millimeters to several centimeters (Fig. 2b, d). The clasts generally consist 130 of blue amphibole + lawsonite ± minor clinopyroxene in different modal abundances and with different 131 grain-size (Fig. 2d). The matrix is generally very fine-grained and mainly consists of blue amphibole, green 132 clinopyroxene (aegirine/omphacite) forming fine-grained dusty aggregates, porphyroblastic lawsonite and 133 minor phengite and chlorite. Fine-grained aggregates of titanite (leucoxene) replace former ilmenite.

- Both metabasic rocks and metavolcanoclastic rocks can be crosscut by glaucophane veins and/or late albite
   ± calcite, and albite + chlorite ± quartz veins.
- 136

#### 137 2.1.2 Metasediments

138 Both silicic and impure carbonatic metasediments occur as intercalations in the metabasic and 139 metavolcanoclastic rocks. Among the silicic metasediments, glaucophane + lawsonite + phengite  $\pm$  garnet 140 schists, lawsonite + glaucophane + phengite + garnet quartzitic-micaschists and glaucophane + garnet + 141 phengite quartzites (Fig. 2f) are the most common types. Lawsonite and garnet can be either fine-grained 142 or pophyroblastic. Lawsonite and garnet porphyroblasts can reach few centimeters and few millimeters in 143 size, respectively, and generally overgrow the main foliation; lawsonite porphyroblasts are locally dusty due 144 to the presence of abundant fluid inclusions. Glaucophane and phengite are always fine-grained and define 145 the main foliation, which is often intensely crenulated. Titanite is ubiquitous as accessory mineral. The 146 lawsonite blueschists investigated in detail in this paper belong to this group of metasediments.

The impure carbonatic metasediments are very fine-grained and mainly consist of lawsonite, calcite, glaucophane and minor phengite ± prehnite (Fig. 2g). Calcite often occurs as large poikiloblasts including idioblastic lawsonite. Prehnite is rare and occurs as reniform globular aggregates of fine-grained brownish fibrous crystals.

Late quartz, albite  $\pm$  quartz and calcite  $\pm$  albite veins crosscut the main schistosity in most metasediments.

151

# 152153 **3. Methods**

# 154 **3.1 Micro-X-ray fluorescence (μ-XRF) maps**

155 The micro-XRF maps of the whole thin sections (Fig. 3 and Fig. SM1, SM2) were acquired using a  $\mu$ -XRF 156 Eagle III-XPL spectrometer equipped with an EDS Si(Li) detector and with an Edax Vision32 microanalytical 157 system (Department of Earth Sciences, University of Torino, Italy). The operating conditions were as follows: 100 ms counting time, 40 kV accelerating voltage and a probe current of 900  $\mu$ A. A spatial resolution of about 65  $\mu$ m in both x and y directions was used. Quantitative modal percentages of each mineral were obtained by processing the  $\mu$ -XRF maps with the software program "Petromod" (Cossio et al. 2002).

162

# 163 **3.2 Mineral chemistry**

164 Minerals were analysed with a Cambridge Stereoscan 360 SEM equipped with an EDS Energy 200 and a 165 Pentafet detector (Oxford Instruments) at the Department of Earth Sciences, University of Torino. The 166 operating conditions were as follows: 50 s counting time and 15 kV accelerating voltage. SEM–EDS 167 quantitative data (spot size = 2  $\mu$ m) were acquired and processed using the Microanalysis Suite Issue 12, 168 INCA Suite version 4.01; natural mineral standards were used to calibrate the raw data; the  $\rho\phi$ Z correction 169 (Pouchou and Pichoir, 1988) was applied. Absolute error is 1  $\sigma$  for all calculated oxides.

170 Mineral chemical data of representative minerals are reported in Tables SM1, SM2. Structural formulae 171 have been calculated on the basis of 12 oxygens for garnet, 6 oxygens for omphacite, 8 oxygens for 172 lawsonite, 11 oxygens for phengite and 23 oxygens for amphibole. Fe<sup>+3</sup> has been calculated by 173 stoichiometry except for amphibole (average Fe<sup>+3</sup> values).

174

#### 175 **3.3 Phase diagrams computation**

Isochemical phase diagrams were calculated in the MnNKCFMASH(O) system using Perple\_X (version 6.7.1,
Connolly 1990, 2009) and the thermodynamic dataset and equation of state for H<sub>2</sub>O–CO<sub>2</sub> fluid of Holland
and Powell (1998, revised 2004). The following solid solution models were used: garnet (Holland and
Powell, 1998), amphibole (Diener et al., 2007, 2012), omphacite (Green et al., 2007; Diener et al., 2012),
chlorite (Holland et al., 1998), phengite (Holland and Powell, 1998), plagioclase (Newton et al., 1980) and
epidote (Holland and Powell, 1998). Quartz, lawsonite, and zoisite were considered as pure end-members.

182 The bulk rock compositions of the studied samples have been calculated by combining the mineral 183 proportions obtained from the modal estimate of micro-XRF maps (Fig. 3, Table 1) with mineral chemistry 184 acquired at SEM-EDS, and are reported in Table 1: these whole rock compositions have been used to 185 model: (i) the whole prograde P-T evolution in sample 14-4B; (ii) the growth of garnet core + mantle in 186 sample 14-6F. For this last sample, the possible effects of chemical fractionation of the bulk composition 187 due to the growth of the strongly zoned garnet porphyroblasts have been also considered. The bulk 188 composition effectively in equilibrium during the growth of garnet rim has been therefore calculated by 189 subtracting the garnet core and mantle compositions (i.e. the modal amount of garnet core + mantle was 190 estimated from the micro-XRF maps as 3.5 vol%) to the whole rock composition (Table 1).

191

# 192 **4. Petrography and mineral chemistry**

Among the various lithologies observed in the study area, two metasediments (samples 14-4B and 14-6F/G) have been petrologically investigated in detail; they are both characterized by a relatively simple and very well preserved mineral assemblage, but differ for the grain size and the modal abundance of each phase (Table 1). Samples 14-6F/G derive from the same hand specimen (two different thin sections cut parallel and perpendicular to the main lineation); petrography and mineral chemistry refer to both thin sections, whereas the micro-XRF map and the thermodynamic modeling refer to sample 14-6F only.

### 200 **4.1 Sample 14-4B**

201 Sample 14-4B is a fine-grained lawsonite + glaucophane + garnet -bearing quartzitic-micaschist 202 characterized by mm-thick quartz-rich layers alternating with mm-thick lawsonite + phengite-rich layers

- (Fig. 4a). The main foliation, defined by the preferred orientation of phengite and glaucophane in both
   domains, is crosscut by late quartz-bearing and calcite ± albite -bearing veins (Fig. 3, and Fig. SM1).
- 205 Microstructural relationships between quartz (42 vol%), lawsonite (21 vol%), phengite (22 vol%), 206 glaucophane (12 vol%) and garnet (3 vol%) suggest that these minerals all belong to the equilibrium 207 assemblage (Fig. 7). Abundant titanite and minor pyrite occur as accessory minerals.
- Lawsonite occurs as fine-grained idioblasts (Fig. 4) with quartz ± titanite inclusions; in the quartz-rich layers, lawsonite is often crowded of quartz inclusions, locally assuming a skeletal habit. It is almost pure in composition, with a very low Fe content (0.00-0.30 a.p.f.u. on the basis of 8 oxygens).
- The fine-grained phengite (Fig. 4b,c) shows a relatively large compositional spread in Si,  $AI_{tot}$  and (Mg + Fe<sub>tot</sub>). Its Si content ranges between 3.53 and 3.81 a.p.f.u. (on the basis of 11 oxygens), with the most frequent values in the range 3.53-3.64 a.p.f.u (Fig. 6d). Most of the phengite compositions broadly lie along the celadonite-muscovite compositional joint, reflecting the dominant role of Tschermak's substitution; phengite with the lowest Si contents, however, plot slightly away from the celadonite- muscovite joint in both Si vs.  $AI_{tot}$  and (Mg + Fe<sub>tot</sub>) vs. Si diagrams, thus suggesting the existence of very low Fe<sup>+3</sup> contents (Vidal and Parra, 2000).
- Blue amphibole occurs as fine-grained idioblasts associated to phengite and lawsonite (Fig. 4a,b), and it is slightly zoned, with a lighter blue core and a darker blue rim. Both cores and rims are ferroglaucophane according to the classification of Leake et al. (1997), but are characterized by slightly different Si (on the basis of 23 oxygens), XNa (XNa=Na/Na+Ca) and XFe<sup>+3</sup> (XFe<sup>+3</sup>=Fe<sup>+3</sup>/Fe<sub>tot</sub>) contents (core: Si = 7.62-7.74 a.p.f.u., XNa=0.92-0.95, XFe<sup>+3</sup>=0.23-0.27; rim: Si = 7.91-7.97 a.p.f.u., XNa=0.98-1.00, XFe<sup>+3</sup>=0.10-0.21) (Fig. 6e,f).
- Garnet occurs as small slightly zoned idioblasts (up to 0.3 mm in diameter) (Fig. 4a), particularly enriched in
   Mn (Fig. 6a). XSps decreases and XAIm and XPrp increase from core to rim, whereas XGrs is almost
   homogeneous (core: Sps<sub>55-60</sub>AIm<sub>20-25</sub>Grs<sub>15-22</sub>Prp<sub>0-0.6</sub>; mantle: Sps<sub>50-54</sub>AIm<sub>25-28</sub>Grs<sub>16-23</sub>Prp<sub>0.4-0.9</sub>; rim: Sps<sub>44-47</sub>AIm<sub>30-34</sub>Grs<sub>18-23</sub>Prp<sub>0.6-1.3</sub>) (Fig. 6a).
- 228

# 229 4.2 Sample 14-6F/G

- Sample 14-6F/G is a lawsonite + glaucophane + phengite + garnet schist, dominated by glaucophane (44 vol%) + lawsonite (22 vol%) + phengite (9 vol%) + garnet (4 vol%) layers alternating with discontinuous quartz (21 vol%) -rich domains. The main foliation, defined by the preferred orientation of glaucophane and minor phengite, is overgrown by large lawsonite and garnet porphyroblasts and it is intensely crenulated (Fig. 5a,b). Lawsonite and garnet porphyroblasts crystallization occurred prior to the crenulation event (Fig. 7). Titanite occurs as accessory mineral aligned to the main foliation. Late quartz ± albite ± chlorite veins crosscut the main foliation (Fig. 5d, e).
- The fine-grained blue amphibole nematoblasts in the matrix (Fig. 5a-c) are quite homogeneous in composition; they are glaucophane according to the classification of Leake et al. (1997) and have Si = 7.71-7.99 a.p.f.u., XNa=0.85-1.00 and XFe<sup>+3</sup>=0.14-0.24, with Si and XNa decreasing and XFe<sup>+3</sup> increasing toward the rim (Fig. 6e,f).
- Lawsonite occurs as large porphyroblasts, up to few centimeter in size, overgrowing the main foliation (Fig.
  3, 5a-d and Fig. SM2). Lawsonite porphyroblasts are often boudinated; the boudinage still occurred in the
- 243 lawsonite stability field because lawsonite + quartz + glaucophane are also found in the pressure shadows
- (Fig. 5c). The Fe content in lawsonite is very low (Fe = 0.03-0.05 a.p.f.u. on the basis of 8 oxygens).
- Garnet porphyroblasts, up to 2-3 mm in diameter, overgrow the main foliation and are also included in lawsonite (Fig. 5). They are strongly zoned (Fig. 5e, 6a, 6b), with spessartine decreasing and almandine and pyrope increasing from core to rim (core: Sps<sub>50-55</sub>Alm<sub>16-21</sub>Grs<sub>26-28</sub>Prp<sub>1.1-1.4</sub>; mantle: Sps<sub>40-48</sub>Alm<sub>22-27</sub>Grs<sub>29-31</sub>Prp<sub>1.5-1.8</sub>; rim: Sps<sub>24-30</sub>Alm<sub>37-42</sub>Grs<sub>27-32</sub>Prp<sub>2.5-3.4</sub>) (Fig. 6a). Garnet porphyroblasts include glaucophane,

actinolite, quartz and chlorite in the core and mantle domains, and few omphacite (Jd<sub>16-33</sub>Acm<sub>9-17</sub>) (Fig. 6c),
 phengite (Si = 3.80 a.p.f.u.) and quartz in the mantle and rim domains (Fig. 5e,5f, 6d).

- Phengite occurs as small flakes in equilibrium with glaucophane (Fig. 5, 7); it is locally zoned, with the highest Si content in the rim (core: Si=3.36-3.57 a.p.f.u.; rim: Si=3.61-3.84 a.p.f.u. on the basis of 11 oxygens). The  $Fe^{+3}$  content in phengite is low since most of the phengite compositions lie along the celadonite-muscovite compositional joint (Fig. 6d).
- 255

# 256 **5. Phase equilibria and P-T evolution**

## 257 5.1 P-T pseudosection in the MnNKCFMASH system

258 The thermodynamic modeling approach was used to constrain the P-T evolution of the two blueschist 259 samples. P-T pseudosections have been first calculated in the MnNKCFMASH model system (MnO-Na<sub>2</sub>O-K<sub>2</sub>O-CaO-FeO-MgO-Al<sub>2</sub>O<sub>3</sub>-SiO<sub>2</sub>-H<sub>2</sub>O), and two assumptions were made: (1) H<sub>2</sub>O was considered in excess; (2) 260 261 Fe<sup>+3</sup> was not included in the calculation. The influence of these two important components on the stability of mineral assemblages will be discussed later. Concerning sample 14-6F, the fractionation effects on its 262 263 bulk composition due to the growth of large garnet porphyroblasts have been considered, and two 264 different pseudosections have been calculated: (i) a first pseudosection, calculated using the whole rock 265 composition, has been used to model the growth of garnet core + mantle; (ii) a second pseudosection, 266 calculated using the effective bulk composition derived by subtracting garnet cores and mantles to the 267 whole rock composition (Table 1), has been used to model the growth of garnet rim. Fractionation effects 268 on the bulk composition are negligible for sample 14-4B, because garnet is very small.

269

### 270 5.1.1 P-T evolution constrained for sample 14-4B

The topology of the pseudosection calculated for sample 14-4B is very simple and dominated by three- and four-variant fields (Fig. 8a). The observed peak assemblage Grt + Gln + Lws + Phe is modelled by a relatively narrow five-variant field at P > 19 kbar, which separates a chlorite-bearing field (at lower T) from an omphacite-bearing field (at higher T). At P < 19 kbar, both chlorite and omphacite coexist in the threevariant Grt + Gln + Lws + Phe + Chl + Omp field.

276 The modeled garnet compositional isopleths (core: X<sub>Sos</sub>=0.60, X<sub>Alm</sub>=0.24, X<sub>Grs</sub>=0.15, X<sub>Pro</sub>=0.006; mantle: X<sub>Sos</sub>= 277 0.52, X<sub>Alm</sub>=0.28, X<sub>Grs</sub>=0.18, X<sub>Prp</sub>=0.009; rim: X<sub>Sps</sub>=0.44, X<sub>Alm</sub>=0.30, X<sub>Grs</sub>=0.23, X<sub>Prp</sub>=0.013) constrain the growth 278 of garnet core, mantle and rim at about 365 °C, 19.5 kbar (in the Grt + Gln + Lws + Phe + Chl field), 390°C, 279 20.5 kbar (in the Grt + Gln + Lws + Phe field) and 420°C, 22 kbar (in the Grt + Gln + Lws + Phe + Omp field) 280 (Fig. 8a and Fig. SM3). The modeled modal amounts of chlorite and omphacite in equilibrium with garnet 281 core and rim, respectively, are lower than 0.5 vol%. The modeled phengite compositional isopleths (Si = 282 3.80-3.82 a.p.f.u.) constrain the growth of phengite at P-T conditions compatible with the growth of garnet 283 core and mantle.

284 The resulting prograde P-T evolution of sample 14-4B is therefore characterized by an increase in both P 285 and T, up to peak conditions of about 420°C, 22 kbar (Fig. 8a). The modeled isomodes of the main mineral 286 phases are consistent with the prograde growth (i.e. increase in its modal amount) of garnet along this P-T 287 path, but predict the (slight) consumption of lawsonite (Fig. 8b; the P-T path crosses the Lws-isomodes 288 downward), opposite to microstructural observations which suggest that garnet and lawsonite grew almost 289 simultaneously (Fig. 7). This apparent discrepancy between the results of the thermodynamic modeling and 290 the observed microstructure will be discussed in the following. The modeled H<sub>2</sub>O isomodes show that 291 during the inferred prograde evolution, a moderate de-hydration occurred, thus implying that mineral 292 assemblages were H<sub>2</sub>O saturated (Guiraud et al., 2001).

293

#### 294 5.1.2 P-T evolution constrained for sample 14-6F

- The topologies of the two pseudosections calculated for sample 14-6F using the whole rock composition and the fractionated bulk composition are simple and dominated by three- and four-variant fields. Because the two pseudosections are quite similar (the main difference is the shift of the Grt-bearing fields toward higher temperatures in the fractionated pseudosection), they have been condensed in the same figure (Fig. 9a). Two large three-variant fields, separated by a narrow di-variant field, dominate the two pseudosections: at higher P (and lower T) is stable the Chl + Grt + Act + Gln + Lws + Phe assemblage, whereas at lower P (and higher T) is stable the Chl + Grt + Gln + Lws + Omp + Phe assemblage.
- 302 The modeled garnet compositional isopleths from the unfractionated pseudosection (core: X<sub>sps</sub>=0.55, X<sub>Alm</sub>=0.18, X<sub>Grs</sub>=0.27, X<sub>Prp</sub>=0.011; mantle: X<sub>Sps</sub>= 0.40, X<sub>Alm</sub>=0.27, X<sub>Grs</sub>=0.30, X<sub>Prp</sub>=0.018) constrain the growth of 303 304 garnet core and mantle at about 395 °C, 18.5 kbar (in the Chl + Grt + Act + Gln + Lws + Phe field) and 435°C, 305 19.5 kbar (in the Chl + Grt + Gln + Lws + Omp + Phe field) (Fig. 9a and Fig. SM4). The transition from the Act-306 bearing (Omp-absent) field to the Omp-bearing (Act-absent) field is consistent with the occurrence of 307 actinolite inclusions within garnet core, and omphacite inclusions within garnet mantle. The modeled 308 phengite compositional isopleths (Si = 3.81-3.83 a.p.f.u.) constrain the growth of phengite at P-T conditions 309 slightly lower than the growth of garnet core.
- The modeled garnet compositional isopleths from the fractionated pseudosection (rim:  $X_{sps}$ =0.24,  $X_{Alm}$ =0.42,  $X_{Grs}$ =0.31,  $X_{Prp}$ =0.034) constrain the growth of garnet rim at about 470 °C, 20 kbar (in the Chl + Grt + Gln + Omp + Lws + Phe field) (Fig. 9a). The modeled modal amount of chlorite in equilibrium with garnet rim is lower than 1 vol%.
- Peak P-T conditions for sample 14-6F are therefore constrained at about 470 °C, 20 kbar. Overall, the prograde P-T evolution of sample 14-6F is similar in shape to that predicted for sample 14-4B but at lower P and slightly higher T (i.e.  $\Delta T = +50$  °C,  $\Delta P = -2$  kbar). Similarly to sample 14-4B, the modeled isomodes do not predict the growth (i.e. increase in modal amount) of lawsonite along this P-T path (Fig. 9b), opposite to microstructural observation which clearly show that lawsonite grew simultaneously (or even later) to garnet (Fig. 7).
- 320

#### 321 **5.2** The influence of Fe<sup>+3</sup>

- Although low, the Fe<sup>+3</sup> content in glaucophane from both the samples is not negligible, suggesting that the metasediment bulk compositions were slightly oxidized. In order to test the influence of Fe<sup>+3</sup> on the stability of the equilibrium assemblages and on the peak P-T conditions, two P-XFe<sub>2</sub>O<sub>3</sub> and T-XFe<sub>2</sub>O<sub>3</sub> pseudosections were calculated at 420°C, 22 kbar (sample 14-4B; Fig. 10a,b), and 470°C, 20 kbar (sample 14-6F; Fig. 10c,d), respectively, i.e. at the peak P-T conditions estimated for the two samples in the Fe<sup>+3</sup>-free MnNKCFMASH system. A XFe<sub>2</sub>O<sub>3</sub> range of 0-0.5 was considered, with XFe<sub>2</sub>O<sub>3</sub> = Fe<sub>2</sub>O<sub>3</sub>/FeO<sub>tot</sub> (i.e. XFe<sub>2</sub>O<sub>3</sub> = 0 means that all Fe is bivalent; XFe<sub>2</sub>O<sub>3</sub> = 0.5 means that FeO and Fe<sub>2</sub>O<sub>3</sub> are present in equal amounts).
- The  $P-XFe_2O_3$  and  $T-XFe_2O_3$  pseudosections modeled for sample 14-4B and contoured for the garnet rim compositional isopleths, show that peak-P conditions decrease of about 2-3 kbar with increasing  $XFe_2O_3$ , whereas peak-T conditions do not significantly change at variable  $XFe_2O_3$  values. The  $XFe_2O_3$  is constrained
- to a maximum of 0.20, above which the modeled peak assemblage (Grt + Gln + Lws + Phe + minor Omp) is
  no longer stable (Fig. 10a,b).
- The same effects are also observed for sample 14-6F, but in this case the decrease of peak-P conditions is less pronounced (ca. 1 kbar). The stability field of the peak assemblage (Grt + Gln + Omp + Lws + Phe + minor Chl) constrains the maximum  $XFe_2O_3$  value to 0.4, but for  $XFe_2O_3 > 0.15$  the modeled garnet
- 337 compositional isopleths diverge, therefore constraining XFe<sub>2</sub>O<sub>3</sub> to values in the range 0-0.15 (Fig. 10c,d).

The P-T paths of the two studied samples calculated for  $XFe_2O_3=0.10$  mostly overlap, thus suggesting that the prograde P-T evolution of the SSOM blueschists was characterized by an increase in P and T from ca. 370 °C, 17 kbar to peak conditions of ca. 470°C, 19 kbar.

341

### 342 6. Discussion

# 343 **6.1 H<sub>2</sub>O-saturated vs. H<sub>2</sub>O under-saturated conditions**

344 The results obtained so far are based on the assumption that H<sub>2</sub>O was in excess during the whole 345 metamorphic evolution: this is a common assumption in the modeling of lawsonite-bearing blueschist and 346 eclogites (e.g. Davis and Whitney, 2006, 2008; Clarke et al., 2006; Groppo and Castelli, 2010; Endo et al., 347 2012; Wei and Clarke, 2011; Vitale Brovarone et al., 2011; Ao and Bhowmik, 2014; Tian and Wei, 2014; 348 Bhowmik and Ao, 2015). In many cases H<sub>2</sub>O is considered in excess because lawsonite-bearing assemblages 349 demand that high water amounts are available in the system. Opposite to this common assumption, it has 350 also been demonstrated that lawsonite can grow during subduction (at increasing P and T) at H<sub>2</sub>O-351 undersaturated conditions (e.g. Ballevre et al., 2003; Lopez-Carmona et al., 2013). H<sub>2</sub>O-undersaturated 352 conditions would significantly influence phase equilibria and hence P-T estimates; therefore, the possibility 353 that prograde metamorphism could have occurred under H<sub>2</sub>O-undersaturated conditions should be 354 carefully evaluated.

355 This issue was explored by calculating two  $P/T-X(H_2O)$  pseudosections for sample 14-4B (Fig, 11); similar 356 results are obtained for sample 14-6F (see Fig. 12a). These pseudosections report the H<sub>2</sub>O content (in wt%) on the horizontal axis and a P/T gradient on the vertical axis. Two P/T gradients have been considered: the 357 358 first one (gradient A: Fig. 11a) is coincident with the P-T path constrained using the P-T pseudosection 359 calculated with H<sub>2</sub>O in excess, whereas the second one (gradient B: Fig. 11b) is steeper and similar to the 360 early prograde P-T evolution of Eastern Himalayan blueschists reported in the literature (Ao and Bhowmik., 361 2014). The two pseudosections are contoured for garnet core and rim compositions. The intersection 362 between garnet compositional isopleths should provide information about: (i) whether the growth of 363 garnet with the measured composition could have occurred along the previously discussed P/T gradient A 364 but at H<sub>2</sub>O-undersaturated conditions, and (ii) whether the alternative (steeper) P/T gradient B would be 365 compatible with the growth of garnet with the measured composition under H<sub>2</sub>O-undersaturated 366 conditions.

367 The white dotted lines in the calculated P/T-X(H<sub>2</sub>O) pseudosections represent the H<sub>2</sub>O-saturation surface 368 and divide the pseudosections in a  $H_2O$ -saturated part on the right and in a  $H_2O$ -undersaturated part on the 369 left. A H<sub>2</sub>O amount of 3-4 wt% (depending on T and P) is required to reach H<sub>2</sub>O-saturated conditions in 370 sample 14-4B. Garnet compositional isopleths show that: (i) the steeper P/T gradient B (Fig. 11b) is not 371 compatible with the observed garnet compositions because the modeled compositional isopleths of garnet 372 core do not overlap; (ii) concerning the P/T gradient A, the intersection of garnet compositional isopleths 373 on the H<sub>2</sub>O-saturation surface confirms that the growth of garnet with the measured composition occurred 374 at H<sub>2</sub>O-saturated conditions (Fig. 11a, 12a), thus suggesting that the assumption of H<sub>2</sub>O in excess for the 375 modeling of garnet growth was correct.

376

#### 377 **6.2** When and how did lawsonite grow

The P/T-X(H<sub>2</sub>O) pseudosections calculated at H<sub>2</sub>O saturated conditions for both samples 14-4B (Fig. 8) and 14-6F (Fig. 9) fail in modeling the contemporaneous growth of lawsonite and garnet; in fact, the inferred prograde P-T path crosses the garnet isomodes upward (Fig. 8c, 9c), but the lawsonite isomodes are crossed downward (Fig. 8b, 9b), thus suggesting that lawsonite was (slightly) consumed when garnet was growing (i.e. lawsonite modal amount was slightly decreasing while garnet modal amount was increasing). Two different hypothesis can be proposed to explain the discrepancy between the observed microstructures and the prediction of thermodynamic modeling: (i) the first hypothesis is still based on an equilibrium model for prograde metamorphism, which is the classical paradigm that is the basis of isochemical phase diagrams; (ii) the second hypothesis explores the possibility that the prograde appearance of lawsonite was controlled by nonequilibrium processes rather than by equilibrium ones (i.e. kinetics factors prevailing over equilibrium thermodynamics).

389

#### 390 *6.2.1 The equilibrium approach*

391 Following an approach based on the principles of equilibrium thermodynamics, the  $P/T-X(H_2O)$ 392 pseudosection calculated for sample 14-6F (Fig. 12a), contoured for lawsonite and garnet modal amounts 393 (Fig. 12b, 12c) is useful to explain the inconsistency between the observed and predicted sequence of 394 porphyroblasts growth (i.e. Lws contemporaneous with Grt vs. Lws earlier than Grt) (see also Fig. SM5 for 395 sample 14-4B). Fig. 12b shows that H<sub>2</sub>O addition is required to form lawsonite (i.e. to increase its modal 396 amount). The observed microstructures suggest that lawsonite growth was contemporaneous to garnet 397 growth (Fig. 7), thus implying that H<sub>2</sub>O was introduced in the system at the relatively high pressure of ca. 398 17-18 kbar (large white arrow in Fig. 12b). Once reached H<sub>2</sub>O-saturated conditions, garnet (with the 399 measured composition of Grt core) started to form; the simultaneous growth of high modal amounts of lawsonite, however, subtracted H<sub>2</sub>O to the system ("-H<sub>2</sub>O" arrows in Fig. 12b), that eventually became 400 401 again  $H_2O$ -undersaturated. A protracted  $H_2O$  influx at high pressure ("+ $H_2O$ " arrows in Fig. 12b) is therefore 402 required in order to allow the contemporaneous growth of garnet (which requires H<sub>2</sub>O-saturated 403 conditions) and lawsonite (whose growth subtracts  $H_2O$  to the system).

- 404 Our model thus suggests that the system might have been H<sub>2</sub>O-undersaturated during the early prograde 405 subduction (i.e. prior to the appearance of garnet). According to the modeling, at  $H_2O$ -undersaturated 406 conditions, the Ca-rich precursor of lawsonite should have been epidote: the small epidote inclusions 407 observed in garnet porphyroblasts (sample 14-6F: Fig. 5f) would support this assumption. This hypothesis 408 confirms what has been already predicted by previous studies, i.e. the H<sub>2</sub>O-rich character of lawsonite-409 bearing assemblages requires the addition of  $H_2O$  at elevated pressure to allow them to form (Clarke et al., 410 2006; Tsujimori and Ernst, 2014). Significant fluid release is predicted at these P-T conditions (e.g. Ulmer 411 and Trommsdorff, 1995; Scambelluri et al., 2004; Poli and Schmidt, 1995; Poli et al., 2009) through 412 metamorphic devolatilization reactions occurring in the subducting slab (Bebout, 1991, 1995; Jarrad, 2003). 413 Our results suggest that fluids released at P > 17-18 kbar by the de-hydrating subducting slab can be largely 414 re-incorporated in lawsonite, and confirm that the pervasive growth of lawsonite represents an efficient 415 mechanism for fixing water in the high pressure accretionary prism, thereby delaying its ascent toward the 416 surface (Ballèvre et al., 2003; Vitale Brovarone and Beyssac, 2014).
- 417

#### 418 6.2.2 The nonequilibrium approach

419 Alternatively to what discussed in the previous point, the inconsistency between the observed 420 microstructures and the equilibrium phase relations predicted by the pseudosections could suggest that 421 nonequilibrium processes controlled the prograde appearance of lawsonite and garnet. Transient 422 nonequilibrium states can be common during prograde metamorphism (e.g. Ague & Carlson, 2013), 423 especially at low temperatures such those inferred for the early prograde evolution of the studied 424 blueschists. Previous works addressed the question of the interplay between the approach to equilibrium 425 on one hand, and reaction kinetics on the other hand (see Ague & Carlson, 2013 for a review). Crucial to the 426 discussion is the concept of reaction affinity, which is an energetic expression of the easiness of a reaction 427 to overstep the kinetic barriers to nucleation and growth (e.g. Waters & Lovegrove, 2002; Pattison et al., 428 2011; Ketcham & Carlson, 2012). It has been demonstrated that mineral reactions which release large quantities of H<sub>2</sub>O have higher reaction affinity per unit of temperature/pressure overstep than those which
 release little or no H<sub>2</sub>O. The former are expected to be overstepped in temperature and/or pressure less
 than the latter (see Pattison et al., 2011 for further details). Reactions with lower reaction affinity may be
 strongly influenced by kinetic factors, or may not occur at all.

433 Some authors considered nucleation as the main rate-limiting process in metamorphic reactions (e.g. 434 Waters & Lovegrove, 2002; Gaidies et al., 2011; Pattison et al., 2011). They demonstrated that low reaction 435 affinity (and consequently high overstepping) of a prograde metamorphic reaction may cause the delayed 436 nucleation (and growth) of porphyroblastic phases. Microstructurally, this becomes evident when the 437 observed sequence of porphyroblasts growth does not coincide with the sequence predicted by 438 thermodynamic modelling (e.g. Waters & Lovegrove, 2002). Other authors argued that intergranular 439 diffusion is the main kinetic component controlling the nucleation and growth of porphyroblastic phases 440 (e.g. Carlson, 1989, 2002; Hirsch et al., 2000; Ketcham & Carlson, 2012). In this case, delayed porpyroblasts 441 growth would be related to the sluggishness of intergranular diffusion. More in detail, growing 442 porphyroblasts extract nutrients from the immediate surroundings, suppressing the nucleation of new 443 crystals in diffusionally depleted zones surrounding pre-existing crystals.

444 A quantitative treatment of these concepts is well beyond the aim of this paper; nevertheless, it is worth 445 nothing that the modelled pseudosection for sample 14-6F predicts that lawsonite is mainly produced at low P-T conditions (i.e. at P < 5 kbar, and T < 300°C; Fig. SM6) through the epidote or prehnite (depending 446 447 on T) breakdown, much earlier than the onset of garnet growth. Both the epidote- and prehnite-448 consuming (lawsonite-producing) reactions are hydration reactions, i.e. they consumes H<sub>2</sub>O. Qualitatively, 449 it is therefore to be expected that reaction affinity of these reactions is very low and that they might be 450 significantly overstepped in temperature and pressure. The discrepancy between the observed and 451 predicted sequence of porphyroblasts growth can be therefore explained by a delayed growth of lawsonite 452 porphyroblasts, possibly due to: (i) low reaction affinity of the Lws-producing reaction (either Ep- or Prh-453 consuming), and/or (ii) difficulty of nucleation of lawsonite.

454 Both the equilibrium- and nonequilibrium- hypothesis are compatible with microstructural observations 455 (e.g. the rare occurrence of small epidote inclusions within garnet) and they are complementary rather 456 than mutually exclusive.

457

#### 458 **6.3** Interpretation of the P-T evolution and geodynamic implications

Prior to this study, P-T estimates based on conventional thermobarometry suggested peak P-T conditions of 459 460 350-420°C, 9-11 kbar for the SSOM blueschists (Honegger et al., 1989). The results of our petrological 461 modeling point to peak P-T conditions significantly higher than those previously estimated, i.e. ca. 470°C, 462 19 kbar (Fig. 13), thus suggesting that the careful re-examination (by means of modern petrological 463 approaches) of previous P-T estimates obtained using conventional thermobarometry can provide new 464 insights on the subduction history of the Neo-Tethyan ocean. The obtained results suggest that the SSOM 465 blueschists experienced a cold subduction history along a very low to low thermal gradient ("early" prograde: ca. 5-6°C/km; "late" prograde: ca. 7-8°C/km; Fig. 13a). Furthermore, in order to preserve 466 467 lawsonite in the studied lithologies, exhumation must have been coupled with significant cooling (i.e. 468 without crossing the lawsonite-out boundary; Zack et al., 2004). The resulting P-T path is therefore 469 characterized by a clockwise hairpin loop along low thermal gradients (< 8-9 °C/km) (Fig. 13a).

This P-T evolution is consistent with a cold subduction zone system in an intra-oceanic subduction setting, as also suggested by Ao and Bhowmik (2014) for blueschists from the far eastern Himalaya. Moreover, the observed lithological associations (i.e. mainly volcanoclastic rocks and minor sediments), the estimated peak P-T conditions (very close to the eclogite stability field but still inside the lawsonite blueschist -facies) and the clockwise hairpin P-T trajectory, are all consistent with the interpretation that the SSOM represents a relic of an oceanic paleo-accretionary prism, related to the northward subduction of the northern Neo-

- Tethyan ocean beneath the Ladakh Asian margin (Robertson, 2000; Mahéo et al., 2006; Guillot et al., 2008).
- 477 Interestingly, the estimated peak P-T conditions of ca. 470°C, 19 kbar roughly coincide with the maximum
- P-T estimates predicted by thermo-mechanical models for the metasediments exhumed in the accretionary
  wedge (Yamato et al. 2007) (Fig. 13a), and with the maximum P-T conditions registered by natural
- occurrences of blueschist accretionary complexes worldwide (Fig. 13b) (e.g. the Schistes Lustres Complex of
  the Western Alps and Alpine Corsica, Turkey, Zagros, Oman, New Caledonia, Franciscan Complex: e.g.
  Banno et al., 2000; Agard et al., 2001a,b; Warren al., 2005; Agard et al., 2006; Page et al., 2006; Tsujimori et
  al., 2006; Warren and Waters, 2006; Ernst and McLaughlin, 2012; Plunder et al., 2012, 2015; Agard and
- 484 Vitale Brovarone, 2013; Ukar and Cloos, 2014; Vitale Brovarone et al., 2014).
- 485 Most of the studies focused on subduction-related HP-LT terranes from different localities point to a 486 continuous increase of peak-T and associated P in adjacent tectonometamorphic units (Fig. 13b). A 487 continuous metamorphic gradient is thus recorded in most of the blueschist-facies terranes worldwide, up to maximum P-T conditions of ca. 470°C, 18-19 kbar (e.g. Oman: Yamato et al., 2007; Corsica: Vitale 488 489 Brovarone et., 2014; Schistes Lustres of the Western Alps: Plunder et al., 2012; New Caledonia: Vitale 490 Brovarone & Agard, 2013; Turkey: Plunder et al., 2015). This metamorphic zonation might reflect the 491 repeated accretion of the ocean-floor sediments subducted at different depths and offscraped at the base 492 of the accretionary prism (e.g. Agard et al., 2009 and references therein).
- 493 A similar metamorphic zonation from greenschist to pumpellyite-diopside and up to lawsonite-blueschist -494 facies conditions has been recently reported by Ao & Bhowmik (2014) for the Nagaland Ophiolite Complex 495 of far-eastern Himalaya, whose geological setting is very similar to that of the SSOM (i.e. it is mainly 496 dominated by metavolcanoclastic rocks, with minor intercalations of metasediments). Although a detailed 497 discussion of the SSOM metamorphic units adjacent to the blueschist one is beyond the aim of this paper, it 498 is worth mentioning that preliminary data suggest that a similar metamorphic zonation might characterize 499 also the western portion of the ITS zone. Chlorite + epidote + green/blue-green amphibole -bearing 500 metavolcanoclastic rocks, and prehnite-pumpellyte -bearing metagabbros occur in the thin metamorphic 501 slices associated to the blueschist unit in the SSOM. Further petrological investigations could eventually 502 confirm the existence of a continuous metamorphic gradient in the SSOM.
- 503 504

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# 777 Figure captions

Fig. 1 – Geological sketch map of the studied area (modified after Honegger et al., 1989). (1) Dras-Naktul
volcanoclastics and flysch; (2) pillow lavas, sill and dyke series; (3) ultramafic lenses; (4) Shergol
conglomerate; (5) mèlange formation; (6) blueschist zone; (7) Karamba and Lamayuru unit; (8) Zanskar unit.
Star: samples location. Inset: simplified tectonic map of the Himalayan orogen showing the locations of the
blueschist facies rocks in the Indus Tsangpo suture zone (ITS). 1, Shangla; 2, Sapi-Shergol; 3, Zildat; 4, Sans
Sang; 5, Yamdrock; 6, Nagaland. Other abbreviations used: NP, Nanga Parbat; NB, Namche Barwa; MBT,
Main Boundary Thrust; MFT, Main Frontal Thrust.

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Fig. 2 – (a) Panoramic view of the Sapi-Shergol Ophiolite Melange (in blue). View looking north-westward.
Landscape width is about 20 km. (b,c) Outcrop appearance of the most abundant blueschist lithologies in
the SSOM: volcanoclastic rocks (b) and metabasic rocks (c). (d-g) Representative microstructures of
volcanoclastic (d) and metabasic (e) rocks and of silicic (f) and carbonatic (g) metasediments. The dotted
white line in (d) separates a pluri-mm clast (lower right) from the reddish matrix (upper left). The inset in
(g) shows a detail of a large poikiloblast of calcite including idioblastic lawsonite. Plane Polarized Light (PPL).

Fig. 3 - Processed major elements μ-XRF maps of the whole thin sections of samples 14-4B and 14-6F. The
 unprocessed μ-XRF maps for each element are reported in Fig. SM1 and SM2.

Fig. 4 – Representative microstructures of sample 14-4B. (a) Detail of a discontinuous quartz-rich layer
 alternated to thicker lawsonite + phengite + glaucophane layers. Note the small dark garnet on the right.
 (PPL). (b, c) Detail of a phengite + lawsonite + glaucophane layer: phengite and glaucophane define the
 main foliation. PPL (b) and Crossed Polarized Light (XPL) (c).

801 Fig. 5 – Representative microstructures of sample 14-6F/G. (a) The main foliation, defined by the preferred 802 orientation of glaucophane and minor phengite, is overgrown by large lawsonite and garnet porphyroblasts 803 and is intensely crenulated. PPL (a), XPL (b). (c) Detail of a boudinated lawsonite porphyroblast overgrowing 804 the fine-grained glaucophane + phengite matrix. Lawsonite and quartz occur in the pressure shadows. Note 805 the garnet porphyroblasts, overgrowing the main foliation and included in lawsonite. PPL. (d) Detail of a 806 lawsonite porphyroblast including several garnet crystals, crosscut by thin quartz veins. PPL. (e) Processed 807 X-ray map of garnet reported in (d), highlighting the inclusion distribution within garnet and its chemical 808 zoning. (f) Back-scattered (BSE) image of a garnet porphiroblast, showing the distribution of inclusions. 809 Note the occurrence of a small omphacite inclusion in garnet rim and of a small epidote inclusion in garnet 810 mantle.

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**Fig. 6** - Compositional diagrams for the main mineral phases analysed in samples 14-4B and 14-6F/G. (a) Garnet compositions plotted in the Grs-(Sps+Andr)-(Alm+Prp) diagram. (b) Fe, Mg and Mn X-ray maps of the same garnet reported in Fig. 3d, e. (c) Omphacite compositions (inclusions in garnet) plotted in the Jd-Quad-Aeg diagram. (d) Phengite compositions plotted in the Si vs. (Mg + Fe) (a.p.f.u.) diagram. The black line represents the ideal celadonitic substitution. (e,f) Na-Amphibole compositions plotted in the Si (a.p.f.u.) vs. Mg/(Mg+Fe<sup>+2</sup>) (e), Si (a.p.f.u.) vs. Na/(Na+Ca) and Si (a.p.f.u.) vs. Fe<sup>+3</sup>/Fe<sub>tot</sub> (f). (g) Ca-Amphibole compositions plotted in the Si (a.p.f.u.) vs. Mg/(Mg+Fe<sup>+2</sup>).

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**Fig. 7** – Metamorphic evolution inferred for samples 14-4B and 14-6F. Sm is the main foliation.

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822 Fig. 8 – (a) P-T pseudosection calculated for sample 14-4B in the MnNKCFMASH model system and at  $H_2O$ saturated conditions using the whole rock bulk composition. The variance of the fields varies from two (i.e. 823 824 8 phases, white fields) to five (i.e. 5 phases, darker grey fields). Garnet compositional isopleths are reported 825 for garnet core, mantle and rim in dark, medium and light red, respectively (Alm: dashed; Grs: continuous; 826 Prp: dotted; Sps: dashed-dotted lines); phengite compositional isopleths are reported in yellow. The 827 modeled peak assemblage is reported in bold. The black arrow is the prograde portion of the P-T path 828 inferred from the pseudosection. The entire set of garnet compositional isopleths is reported in Fig. SM3. 829 (b, c) Same pseudosection of (a), contoured for lawsonite (b) and garnet (c) modal amount (vol%). Note 830 that lawsonite is predicted to be slightly consumed along the inferred P-T path, whereas garnet is predicted 831 to increase in modal amount.

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833 Fig. 9 – (a) P-T pseudosections calculated for sample 14-6F in the MnNKCFMASH model system and at  $H_2O$ 834 saturated conditions using the whole-rock (unfractionated: lower left) and the fractionated (upper right) 835 bulk compositions, respectively, and used to model the growth of garnet core and mantle (unfractionated 836 bulk) and garnet rim (fractionated bulk). The variance of the fields varies from two (i.e. 8 phases, white 837 fields) to four (i.e. 6 phases, darker grey fields). Garnet compositional isopleths are reported for garnet 838 core, mantle and rim in dark, medium and light red, respectively (Alm: dashed; Grs: continuous; Prp: 839 dotted; Sps: dashed-dotted lines); phengite compositional isopleths are reported in yellow. The black 840 arrows are the prograde portions of the P-T path inferred for the growth of garnet core, mantle and rim. 841 The entire set of garnet compositional isopleths is reported in Fig. SM4. (b, c) Same pseudosections of (a), 842 contoured for lawsonite (b) and garnet (c) modal amount (vol%). Note that lawsonite is predicted to be 843 slightly consumed along the inferred P-T path, whereas garnet is predicted to increase in modal amount.

Fig. 10 – (a, b) P-X(Fe<sub>2</sub>O<sub>3</sub>) and T-X(Fe<sub>2</sub>O<sub>3</sub>) pseudosections calculated for sample 14-4B in the 845 MnNKCFMASHO model system at T = 420°C and P = 22 kbar, respectively. (c, d) P-X(Fe<sub>2</sub>O<sub>3</sub>) and T-X(Fe<sub>2</sub>O<sub>3</sub>) 846 847 pseudosections calculated for sample 14-6F (fractionated bulk composition) the MnNKCFMASHO model 848 system at T =  $470^{\circ}$ C and P = 20 kbar, respectively. In all the pseudosections the variance of the fields varies 849 from two (i.e. 8 phases, white fields) to six (i.e. 5 phases, darker grey fields). Garnet compositional isopleths 850 are reported for garnet rim in red (Alm: dashed; Grs: continuous; Prp: dotted; Sps: dashed-dotted lines). 851 The modeled equilibrium assemblages are reported in bold. For both the samples, peak-P conditions 852 decrease with increasing XFe<sub>2</sub>O<sub>3</sub> ( $\Delta P$  = 2-3 kbar for sample 14-4B and  $\Delta P$  = 1 kbar for sample 14-6F), 853 whereas peak-T conditions do not significantly change at variable XFe<sub>2</sub>O<sub>3</sub> values.

Fig.  $11 - P/T-X(H_2O)$  pseudosection calculated for sample 14-4B in the MnNKCFMASH model system along 855 856 two different P/T gradients: gradient A (a) coincides with the P-T path constrained using the P-T 857 pseudosection calculated with H<sub>2</sub>O in excess (black arrow in Fig. 8a); gradient B (b) is steeper (similar to the 858 early prograde P-T evolution of Eastern Himalayan blueschists reported in the literature; Ao and Bhowmik., 859 2014). The variance of the fields varies from two (i.e. 7 phases, white fields) to five (i.e. 5 phases, darker 860 grey fields). Garnet compositional isopleths are reported for garnet core and rim in dark and light red, 861 respectively (Alm: dashed; Grs: continuous; Prp: dotted; Sps: dashed-dotted lines). The observed 862 equilibrium assemblage is reported in bold. The white dotted lines in both pseudosections represent the 863 H<sub>2</sub>O-saturation surface and divide the pseudosections in a H<sub>2</sub>O-saturated part on the right and in a H<sub>2</sub>O-864 undersaturated part on the left. The intersection of garnet compositional isopleths on the H<sub>2</sub>O-saturation 865 surface in (a) confirms that garnet growth (with the measured composition) occurred at H<sub>2</sub>O-saturated 866 conditions; garnet core compositional isopleths do not intersect in (b), thus implying that gradient B is not 867 compatible with the observed mineral assemblage and compositions.

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Fig. 12 – (a,b) P/T-X(H<sub>2</sub>O) pseudosections calculated for sample 14-6F in the MnNKCFMASHO model system 869 870 along the same gradient A as in Fig. 11a and using the whole-rock (unfractionated: a) and the fractionated 871 (b) bulk compositions. The variance of the fields varies from two (i.e. 7 phases, white fields) to five (i.e. 5 872 phases, darker grey fields). Garnet compositional isopleths are reported for garnet core and rim in dark and light red, respectively (Alm: dashed; Grs: continuous; Prp: dotted; Sps: dashed-dotted lines). The white 873 874 dotted line represents the H<sub>2</sub>O-saturation surface and divides the pseudosections in a H<sub>2</sub>O-saturated part 875 on the right and in a H<sub>2</sub>O-undersaturated part on the left. The intersection of garnet compositional 876 isopleths on the  $H_2O$ -saturation surface confirms that garnet growth (with the measured composition) 877 occurred at H<sub>2</sub>O-saturated conditions.(c-f) Same P/T-X(H<sub>2</sub>O) pseudosection of (a, b) contoured for lawsonite 878 (c, d) and garnet (e, f) modal amounts (in vol%). The red ellipses indicate the  $P-T-X(H_2O)$  conditions inferred 879 for the growth of garnet core and rim from Fig. 12a, 12b. H<sub>2</sub>O addition is required to form lawsonite (c, d). 880 A protracted H<sub>2</sub>O influx at high pressure is required in order to allow the contemporaneous growth of garnet, which requires H<sub>2</sub>O-saturated conditions (e, f) and lawsonite, whose growth subtracts H<sub>2</sub>O to the 881 882 system (c, d) (see text for further details).

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884 Fig. 13 - (a) P-T path constrained for the Sapi-Shergol blueschist unit (red thick vs. dotted arrows are the P-T paths constrained in the MnNKCFMASHO vs. MnNKCFMASH system, respectively; the P-T path previously 885 constrained by Honegger et al., 1989 and Guillot et al., 2008 is reported in orange) compared with the P-T 886 887 paths of the other Himalayan blueschist rocks: Shangla (in yellow: Guillot et al., 2008) and Nagaland 888 Ophiolite Complex (in green: Ao and Bhowmik, 2014). The dashed grey arrow is the schematic P-T path 889 followed by the sedimentary particles in the accretionary wedge, as resulting from the thermomechanical 890 numerical study of Yamato et al. (2007). (b) P-T diagram comparing the maximum P-T conditions for well-891 documented examples of accretionary terranes in subduction zones (modified from Agard and Vitale 892 Brovarone, 2013 and Plunder et al., 2015, with references therein) with the P-T conditions experienced by 893 the Sapi-Shergol blueschist unit (this study; red squares) and the Nagaland Ophiolite Complex of far-eastern 894 Himalaya (Ao and Bhowmik, 2014; green square). Data are mainly derived from: Agard et al. (2001b), 895 Plunder et al. (2012): Western Alps; Ravna et al. (2010), Vitale Brovarone et al. (2011, 2013), Agard and 896 Vitale Brovarone (2013): Corsica; David and Whitney (2008), Plunder et al. (2015): Turkey; Warren et al. 897 (2005), Warren and Waters (2006), Agard and Vitale Brovarone, (2013): Oman; Agard et al. (2006): Zagros; 898 Fitzherbert et al. (2003, 2004, 2005), Agard and Vitale Brovarone (2013), Vitale Brovarone and Agard 899 (2013): New Caledonia; Banno et al. (2000), Page et al. (2006), Tsujimori et al. (2006), Ernst and McLaughlin 900 (2012), Ukar and Cloos (2014): Franciscan Complex (western USA).

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Fig. 2 Click here to download high resolution image







Fig. 5 Click here to download high resolution image





















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Sample	14-4B	14-6F	
Qtz	42	21	
Lws	21	22	
Phe	22	9	
Gln	12	44	
Grt	3	4	
Total	100	100	
Sample	14-4B	14-6F	
		unfractionated	fractionated
SiO <sub>2</sub>	70.26	61.69	62.91
$Al_2O_3$	14.40	15.76	15.48
FeO	4.20	7.14	7.01
MgO	1.80	4.75	4.96
MnO	1.05	0.93	0.06
CaO	4.66	5.32	4.95
Na <sub>2</sub> O	0.96	3.34	3.51
K <sub>2</sub> O	2.67	1.07	1.12
Total	100.00	100.00	100.00

# Table 1 - Modal (vol%) and bulk (wt%) compositions of samples 14-4B and 14-6F