**Eclogitic metagabbro from the Lanzada Window, eastern Central Alps:
confirmation of subduction beneath the Malenco Unit**

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Running title: Eclogite from the Lanzada Window

Abstract

Epidote-rich eclogitic metagabbro forms a small body within the Lanzada Window, upper Val Malenco, where it is associated with serpentinites and supracrustal rocks of the Lanzada – Santa Anna Zone (LSZ), which lies structurally beneath the Malenco unit. Conventional garnet–clinopyroxene geothermometry and garnet–clinopyroxene–phengite geobarometry indicate that the peak-metamorphic mineral assemblage (garnet + omphacite + epidote + phengite + titanite + apatite + pyrite) equilibrated at *ca.* 2.0 GPa and 525 °C. The bulk composition is Ca-rich (wollastonite-normative), suggesting that the rock underwent Ca-metasomatism prior to high-*P* metamorphism. The presence of eclogite within the LSZ strengthens the correlation of the LSZ with the blueschist-bearing Avers Bündnerschiefer, and confirms the former existence of a southerly-dipping subduction zone beneath the Malenco unit.

Key words: epidote-eclogite; Lanzada – Santa Anna Zone; Ca-metasomatism; subduction; Piemont-Ligurian ocean.

1 Introduction

It has long been recognised that there is a close connection between rocks of high-pressure, low-temperature facies, such as blueschists and low-/medium-*T* eclogites, and subduction zones (e.g. Ernst 1973), to the extent that the existence of such rocks is generally taken as good evidence for the site of a former subduction zone (e.g. Evans and Brown 1986; Carswell 1990). The purpose of this paper is to report on a chance find of an eclogite in the Lanzada – Santa Anna Zone, a tectonic unit of the eastern Central Alps from which no high-*P*

rocks have previously been recorded. Because of the link between high-*P* rocks and subduction, the find is significant as it helps to constrain the tectonic evolution of the Alpine orogen.

1.1 Geotectonic background

According to the models of Froitzheim et al. (1996), Schmid et al. (2004, 2008) and Handy et al. (2010), the Alpine mountain belt is the product of two orogenic episodes: (i) the closure, in Cretaceous times, of the Meliata ocean between the northern and southern parts of the Apulian plate (Alcapia and Adria, respectively), and (ii) the sequential closure, in the Tertiary, of the Piemonte-Liguria and Valais branches of Alpine Tethys, between the Apulian and European plates. The two branches of Alpine Tethys were separated by the Briançonnais continental mass (Frisch 1979), thought to be the eastern extension of the Iberian continent. In the Central Alps, Tertiary continental convergence and collision was accommodated by south-easterly subduction and/or underthrusting of the Alpine-Tethyan oceanic lithosphere and of slices of the adjacent continental margins, including the Briançonnais. The resulting nappe stack (Fig. 1) includes basement nappes of European provenance (e.g. the Adula nappe), remnants of the Valais basin (e.g. the Chiavenna ophiolite and Rhenodanubian flysch), basement nappes from the Briançonnais domain (e.g. the Tambo and Suretta nappes), ophiolitic remnants of the Piemonte-Ligurian ocean (e.g. the Avers Bündnerschiefer, Platta nappe and Forno portion of the Malenco unit), and Austroalpine basement nappes from the north-western margin of the Apulian plate (e.g. the Err and Bernina nappes and higher Austroalpine nappes) (Schmid et al. 2004). Most of the continental basement nappes are structurally overlain by thin and highly deformed cover sequences of Permo-Mesozoic passive-margin or oceanic sedimentary rocks.

1 The palaeogeographical origin of the lowest Austroalpine continental basement
2 nappes, the Margna and Sella nappes ('Ma' and 'Se', respectively, in Fig. 1), is somewhat
3 controversial. Hermann et al. (1997) consider them to represent the distal part of the Apulian
4 plate, but Frisch (1979), Froitzheim et al. (1996) and Schmid et al. (2004) argue that they
5 represent a micro-continent between north-western (Avers–Forno) and south-eastern (Platta)
6 portions of the Piedmont-Ligurian ocean generated during mid-Jurassic rifting. The NW part of
7 the Malenco unit ('M-F' in Fig. 1), which preserves an original transition from continental
8 lower crust to mantle, directly underlay part of the Margna nappe prior to rifting
9 (Trommsdorff et al. 1993; Müntener and Hermann 1996; Hermann et al. 1997). The Malenco
10 unit and its incorporated lower crust show widespread evidence of exhumation to the sea floor
11 during rifting (Müntener et al. 2000).

12 The present disposition of the aforementioned units in the eastern Central Alps (Fig. 1)
13 is the product of a complex interplay of rifting (Mohn et al. 2010, 2011), imbrication,
14 exhumation and repeated plastic deformation, culminating in intrusion by granitoid magmas
15 of the Bergell complex at *ca.* 33-26 Ma (von Blanckenburg 1992; Hansmann 1996; Oberli et
16 al. 2004) and back-thrusting of the nappe stack along the Periadriatic Line between *ca.* 28-18
17 Ma (Schmid et al. 1996).

1.2 Alpine metamorphism in the Central Alps

18 In the Central Alps, relict Alpine high-*P* mineral assemblages appear to be confined
19 to units underlying the Austroalpine nappes (the 'Penninic' units), but not all Penninic units
20 preserve high-*P* relics. Well preserved and intensively studied eclogite-facies rocks occur in
21 the Adula and Cima Lunga nappes (e.g. O'Hara and Mercy 1966; Ernst 1977; Evans and
22 Trommsdorff 1978; Heinrich 1986; Meyre et al. 1997; Nimis and Trommsdorff 2001; Dale
23 and Holland 2003), but variably retrogressed eclogites also occur in Valais-derived ophiolitic

1 associations in the Antrona zone and the Mergoscia and Orselina zones of the Southern Steep
2 Belt (Colombi and Pfeifer 1986; Pfeifer et al 1991; Tóth et al. 2000; Brouwer et al. 2005).
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4 Trommsdorff (1990) and Brouwer et al. (2005) viewed this assemblage of high-*P* units as a
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6 lithosphere-scale tectonic mélange and interpreted it as the product of a tectonic accretion
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8 channel that developed during convergence and collision along the Apulia-Europe plate
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10 boundary. Well constrained ages for the eclogitic metamorphism are mostly in the range 40 –
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12 35 Ma (Becker 1993; Gebauer 1996; Brouwer et al. 2005).
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17 Lower-grade high-*P* assemblages are preserved in the Valais-derived Bündnerschiefer
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19 of the Vals area (Wiederkehr et al. 2009) and Engadine Window (Goffé and Oberhänsli 1992;
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21 Bousquet et al. 1998), respectively, north and north-east of the area shown in Fig. 1. Relict
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23 blueschist-facies assemblages in these units developed at peak-*P* conditions of 1.2 – 1.4 GPa
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25 (Bousquet et al. 2002) at *ca.* 42 – 40 Ma (Wiederkehr et al. 2009). These dates agree well
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27 with those from the eclogite-facies rocks further south and with them provide a tight
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29 constraint on the timing of the closure of the Valais basin.
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34 Previously published reports of high-*P* metamorphism in the Briançonnais- and
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36 Piemont-Liguria-derived Penninic units overlying the Adula nappe in the eastern Central Alps
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38 are confined to relict blueschist-facies assemblages in the Avers Bündnerschiefer (Oberhänsli
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40 1978; Ring 1992) and the northern parts of the Suretta and Tambo nappes (Ring 1992;
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42 Nussbaum et al. 1998). Estimated peak-*P* conditions for these units are as follows:- Avers:
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44 350 – 400 °C, 0.9 – >1.2 GPa (Ring 1992); Suretta: 380 – 420 °C, 0.8 – >1.0 GPa (Ring
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46 1992), 400 – 450 °C, 1.0 GPa (Nussbaum et al. 1998); Tambo: 400 – 480 °C, 0.8 – 1.0 GPa
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48 (Ring 1992). Subduction of these units to greater depths in Eocene times (50 – 40 Ma) is
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50 implicit in the balanced cross-sections of Schmid (1996). In the Western Alps, the peak of
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52 ultra-high-*P* metamorphism in the Piemont-Liguria-derived ophiolitic Lago di Cignana unit
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54 has been dated at 44 Ma (Rubatto et al. 1998), in agreement with this timing.
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1 No high-*P* mineral assemblages have been reported from the Malenco unit, or the
2 Margna, Sella or Platta nappes. Rb/Sr mineral isochron ages of *ca.* 50 – 47 Ma from
3 pervasively deformed rocks from the Platta nappe and overlying Austroalpine basement have
4 been interpreted by Bachmann et al. (2009) as dating the cessation of subduction along the
5 northern margin of Apulia.
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11 The dominant metamorphic imprint in the Central Alps west of the Bergell intrusive
12 complex is the Tertiary Barrovian greenschist- to amphibolite-facies metamorphism centred
13 on the Lepontine area (e.g. Frey and Ferreiro-Mählmann 1999). This metamorphism
14 overprinted the high-*P* assemblages to varying extents and post-dated collision-related nappe
15 refolding (Wiederkehr et al. 2008), as testified by the smooth, concentric pattern of isograds
16 transecting nappe boundaries (e.g. Thompson 1976; Trommsdorff 1980; Frey and Ferreiro-
17 Mählmann 1999). It developed as the result of thermal relaxation within the tectonically
18 thickened nappe stack (e.g. Todd and Engi 1997; Wiederkehr et al. 2008; Berger et al. 2011),
19 possibly augmented by additional mantle heat flow following slab break-off (von
20 Blanckenburg and Davies 1995). In the sillimanite zone near Bellinzona, the thermal peak of
21 metamorphism has been dated at *ca.* 33 Ma by Gebauer (1996; 1999) but the data of Rubatto
22 et al. (2009) indicate that temperatures high enough for anatexis persisted in these rocks until
23 *ca.* 22 Ma. East of the Bergell, however, the effects of the Lepontine overprint are largely
24 absent (Frey and Ferreiro-Mählmann 1999), and the regional metamorphism is at least partly
25 of Cretaceous age, as documented by ³⁹Ar–⁴⁰Ar amphibole ages of 67–73 Ma for the latest
26 detected crystallisation in the Malenco unit (Villa et al. 2000).
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51 1.3 The Lanzada – Santa Anna Zone

52 The Lanzada – Santa Anna Zone (LSZ) is an enigmatic assemblage of rocks of
53 uncertain palaeogeographical origin structurally underlying the Malenco unit (Trommsdorff et
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1 al. 2005). With an outcrop area totalling $<6 \text{ km}^2$ (Montrasio et al 2005), the LSZ is exposed in
2 two small tectonic windows in Val Malenco, one (the Lanzada window) in the centre of the
3 Malenco unit outcrop and the other (the Santa Anna window) near its southern boundary (Fig.
4 1). The outcrop pattern is controlled by the interference of two sets of upright folds related to
5 back-folding (Trommsdorff et al. 2005). The LSZ, as defined by Trommsdorff et al. (2005),
6 corresponds partly to the Lanzada–Scermendone Zone of Montrasio (1984), but the latter
7 author identified a third window (Alpe Airale) west of the other two which is not included in
8 the LSZ in the map of Montrasio et al. (2005).
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11 The rocks of the LSZ include Permo-Mesozoic supracrustal rocks of pre-rift, passive-
12 margin origin, such as quartzites, marbles and dolomites, as well as those of oceanic affinity,
13 such as greenschists, calc-schists, ophicarbonates and Mn-rich schists (Trommsdorff et al.
14 2005). Also present are small bodies of serpentinite and metagabbro, interpreted as either
15 olistoliths or tectonic slices (Montrasio 1984). The tectonic map of the Alps (Spicher 1980)
16 shows an outcrop of orthogneiss in the centre of the Lanzada window and, although this does
17 not appear on the map of Montrasio et al. (2005), we confirm the existence of an augen gneiss
18 exposure 2 km NE of the town of Lanzada at Swiss map coordinates 7894 1281 (latitude 46°
19 $16' 43''$ N; longitude $9^\circ 53' 52''$ E). Orthogneisses, interpreted as pre-Triassic crystalline
20 basement, occur within the Alpe Airale window and are included in the Lanzada–
21 Scermendone Zone by Montrasio (1984).
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46 According to Trommsdorff et al. (2005), the LSZ may have been derived from the
47 ocean-continent transition of the Briançonnais realm. On the basis of lithological similarities
48 and tectonic position, Montrasio (1984) correlated the metasediments and ophiolitic rocks of
49 the LSZ with the Avers Bündnerschiefer and the orthogneisses with the Suretta nappe. A
50 similar lithological assemblage along the eastern and south-eastern margins of the Bergell
51 complex (Fig. 1) has been confidently correlated with the Suretta nappe and its Avers cover
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(Gieré 1985; Trommsdorff et al. 2005). The possibility of the Suretta nappe outcropping within the Lanzada Window was also noted by Trümpy (1980).

The dominant metamorphism in the LSZ is of albite-epidote-amphibolite facies, with tremolite developed in marbles, rare garnet in pelites, and amphibole + epidote + chlorite + albite in metabasites (Montrasio 1984). The age of this metamorphism is unknown. The eclogitic metagabbro described in this paper is the first evidence of high-*P* metamorphism in the LSZ and in the Malenco area as a whole.

2 Field occurrence and petrography

The eclogitic metagabbro outcrop is located on the southern flank of Monte Motta *ca.* 0.8 km NNE of Lanzada, at Swiss map coordinates 7882 1278 (latitude 46° 16' 33" N; longitude 9° 52' 53" E). It forms a small (*ca.* 2m-wide) exposure on the footpath that contours westwards from a W-closing hairpin bend in the Lanzada-Franscia road. It is closely associated with schistose serpentinites and greenschists that crop out along the SE side of a prominent band of marble which, according to the map of Montrasio et al. (2005), defines the NW margin of the Lanzada Window. Unfortunately, the field relations of the metagabbro body are unclear as its contact with the associated serpentinites and greenschists are not exposed. Nonetheless, the position of the metagabbro locality relative to the marble means that there is no doubt that it lies within the Lanzada Window, rather than within the Malenco Unit or along the boundary between the two.

The metagabbro is a medium-grained, weakly foliated L-S tectonite consisting mainly of omphacite and epidote with subordinate garnet. The omphacite and epidote each form irregular, nearly monomineralic aggregates, commonly measuring *ca.* 20 × 10 × 5 mm but locally up to 40 × 20 × 10 mm. Within the aggregates, the omphacite and epidote prisms are

1 imperfectly aligned parallel to the long axes of the aggregates. Assuming that the epidote and
2 omphacite aggregates represent former plagioclase- and pyroxene-rich domains, the size of
3 the aggregates implies that the gabbroic protolith was very coarse grained.
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5 The full mineral assemblage of the metagabbro is garnet + omphacite + epidote +
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7 phengite + titanite + hornblende + albite + biotite + pyrite + apatite. Modal proportions based
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9 on a point-count of a thin-section (1000 points) are: Grt 6%, Omp 52%, Ep 35%, Phe <<1%,
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11 Ttn 3%, Hbl 1%, Ab 1%, Bt <<1%, Pyt <<1%, Ap <<1% (abbreviations from Kretz, 1983).
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13 The small total mode of retrograde minerals (hornblende, albite and biotite) reflects the fact
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15 that, compared to many Alpine eclogites, the rock shows little post-eclogitic alteration.
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18 The garnet forms irregular porphyroblasts up to 8mm in diameter with abundant
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20 inclusions of titanite and epidote. Garnet rims are lobate and commonly intergrown with
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22 omphacite and epidote (Fig. 2a). Omphacite forms green prisms measuring up to 4×1.5 mm
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24 but are more typically 1–2 mm long. The cores of many of the larger omphacites contain
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26 numerous dark, fluid-filled inclusions aligned parallel to the c-axes of their host crystals.
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28 Epidote prisms tend to be smaller than those of omphacite, measuring up to 2×0.4 mm, and
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30 better aligned. Most prisms are optically zoned with slightly higher-birefringence rims and
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32 contain rare fluid inclusions. Small amounts of yellow, highly birefringent epidote coexist
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34 with albite, hornblende and biotite in retrogressed domains, particularly close to garnet.
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36 Phengite is rare, forming either isolated crystals enclosed by omphacite, or irregular flakes, up
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38 to 0.8mm across, surrounded by retrogressive biotite, albite and yellow epidote. Titanite is the
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40 only Ti-phase present and forms sub-/euhedral crystals up to 0.4 mm long. Scattered,
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42 randomly oriented euhedral hornblende prisms up to 6×1 mm cut across the foliation defined
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44 by omphacite and epidote and locally contain relict inclusions of omphacite (Fig. 2b). Albite
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46 forms localised, irregular, interstitial or poikiloblastic grains replacing garnet and omphacite,
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48 and is usually associated with hornblende.
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On the basis of the textural relations, the peak-metamorphic mineral assemblage of the metagabbro is Grt + Omp + Ep + Phe + Ttn + Pyt + Ap.

The occurrence of titanite in the peak assemblage, rather than rutile, the more common Ti-phase of eclogites (Carswell 1990), is intriguing and may be due to the unusually Ca-rich bulk composition (see below).

3 Mineral chemistry

Minerals were analysed by wavelength-dispersive spectrometry using the Cameca SX-100 electron-microprobe at the University of Manchester. Standards used were wollastonite (Si, Ca), rutile (Ti), corundum (Al), eskolaite (Cr), fayalite (Fe), tephroite (Mn), periclase (Mg), adularia (K), and jadeite (Na). Analyses were obtained at an accelerating voltage of 15kV, beam diameter of *ca.* 1 μ m, and count times of 20s and 10s on peaks and backgrounds, respectively. X-ray counts were processed using PAP software (Pouchou and Pichoir 1991). Representative analyses are listed in Table 1.

3.1 Garnet

Garnet formulae were calculated assuming $\Sigma(\text{Al, Cr, Fe}^{3+}, \text{Fe}^{2+}, \text{Mn, Mg, Ca}) = 5.00$ per formula unit (pfu). The analyses show small Si deficiencies (mean $X_{\text{Si}}^{\text{IV}} = 0.98$), suggesting a small but finite amount of Si=4H substitution due to hydrogarnet component. Calculated Fe^{3+} contents scatter from zero to 0.035 pfu. Garnet porphyroblasts show relatively simple zoning profiles, with Mn and Fe decreasing, and Mg and Ca increasing from core to rim (Fig. 3). The lobate porphyroblast rim regions are compositionally fairly uniform compared to garnet interiors. Extreme compositions are $\text{alm}_{61.0} \text{py}_{4.6} \text{grs}_{28.2} \text{sps}_{4.5} \text{adr}_{0.2} \text{hy-gr}_{1.4}$ (core) and $\text{alm}_{54.4} \text{py}_{6.6} \text{grs}_{33.3} \text{sps}_{3.3} \text{adr}_{0.2} \text{hy-gr}_{2.2}$ (rim).

3.2 Omphacite

Omphacite formulae were calculated assuming 4.00 cations per 6(O) (Droop 1987). Compositions are shown on the jadeite–acmite–augite diagram of Essene and Fyfe (1967) in Fig. 4. Almost all compositions plot within the omphacite field in this diagram. Individual crystals show no consistent zoning patterns, but there are distinct differences between the compositions of omphacites close to garnet and epidote and those of omphacites from the interiors of monomineralic aggregates (Fig. 4). The former have a mean composition of $\text{jd}_{38}\text{acm}_{13}\text{aug}_{49}$ and range from 28%–45% jd and 9%–16% acm, closely resembling omphacites from many low- and medium-T eclogites (e.g. those of the Zermatt-Saas zone – Fig. 4). The mean $\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$ value is 0.80. In contrast, the crystals from the interiors of omphacite aggregates are consistently richer in ‘augite’ components and poorer in acmite and generally also in jadeite, having a mean composition of $\text{jd}_{27}\text{acm}_{06}\text{aug}_{67}$ and mean $\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$ value of 0.83.

3.3 Epidote

Epidote formulae were calculated assuming all Fe to be Fe^{3+} . Away from retrogressed areas, epidote cores show $\text{Fe}^{3+}/(\text{Al}+\text{Fe}^{3+})$ values ranging from 0.11 to 0.15 whilst rims show values ranging from 0.16 to 0.19. The yellow retrogressive epidotes have $\text{Fe}^{3+}/(\text{Al}+\text{Fe}^{3+})$ values in the range 0.26–0.30.

3.4 Phengite

Phengite formulae were calculated assuming all Fe to be Fe^{2+} . Owing to the scarcity of phengite in the section, only two crystals were analysed, and we are forced to assume that their compositions are representative. The mean composition (based on four analyses) is

approximately $K_{0.84}Na_{0.07}Mg_{0.33}Fe_{0.17}Ti_{0.02}Al_{2.28}Si_{3.29}O_{10}(OH)_2$, with $Mg/(Mg+Fe^{2+})$ and $Na/(Na+K)$ values of 0.65 and 0.08, respectively.

3.5 Titanite

Titanite analyses show Ca and Si contents close to the ideal value of 1.00 per 5(O). Al contents range from 0.04 to 0.11 pfu and are inversely correlated with Ti (0.91–0.96 pfu).

3.6 Hornblende

Hornblende formulae were calculated assuming $\Sigma(Si, Ti, Al, Cr, Fe, Mg, Mn) = 13.00$ per 23(O) anhydrous (Droop 1987). With Ca contents of 1.47 to 1.55 pfu, these amphiboles are marginally subcalcic and plot close to the boundary between the magnesiohornblende and barroisite fields in the classification scheme of Leake et al. (1997). Si and Al contents are quite variable, ranging from 7.06 to 7.36 and 1.07 to 1.47 pfu, respectively, as are $Mg/(Mg+Fe^{2+})$ values (0.69–0.76). Most of the variation is inter-crystalline, but within individual crystals there is slight zoning with rimwards increases in Ca and Al and decreases in Si, Na and $Mg/(Mg+Fe^{2+})$. On the $100Na/(Na+Ca)$ vs. $100Al/(Al+Si)$ plot of Laird and Albee (1981), the hornblendes plot in the high-P field (Fig. 5) and, in this respect, resemble the retrograde Ca- and NaCa-amphiboles of the Zermatt-Saas zone more closely than the Ca amphiboles of the overlying Malenco unit and Margna nappe.

3.7 Albite

Retrograde albite compositions are close to end-member (>98% ab).

4 Whole-rock geochemistry

1 The bulk composition of the Lanzada metagabbro was obtained by X-ray fluorescence
2 spectrometry (XRF), using the PANalytical™ Axios XRF spectrometer at the University of
3 Manchester. Duplicate sample pellets were prepared from 100g of powdered rock by mixing
4 12g rock powder with 3g wax binder. X-ray counts were processed using the IQ+ program.
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6 The two analyses were the same, within error; mean major and trace element concentrations
7 are given in Table 2. A crude estimate of the bulk $\text{Fe}^{2+}/\text{Fe}_{\text{tot}}$ of 0.46 (atomic) was obtained
8 from mineral modes and compositions and used to compute an amended major-element
9 analysis and CIPW weight norm (Table 2).
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12 Compared that of Le Maitre's (1976) average gabbro (Table 2), the major-element
13 analysis of the Lanzada metagabbro is characterised by low concentrations of SiO_2 , MgO and
14 K_2O , very high CaO , high Na_2O and MnO , and similar TiO_2 , Al_2O_3 , P_2O_5 and total Fe oxides.
15 Although the overall Fe oxide total is unremarkable, the high proportion of Fe_2O_3 is unusual;
16 oxide-rich ferrogabbros commonly show this trait (as well as low SiO_2) but are usually
17 characterised by high total Fe oxides and correspondingly high TiO_2 , which is not the case
18 here. The most striking feature of the metagabbro analysis is the very high CaO (~18%)
19 which is *ca.* 3σ above that of Le Maitre's (1976) average gabbro. Coupled with modest Al_2O_3
20 and alkalis and low SiO_2 , this produces appreciable wollastonite and nepheline in the norm
21 as well as much diopside (Table 2).
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24 Without appealing to calcite or melilite in the protolith, the appearance of significant
25 normative wollastonite cannot be explained in terms of compositional or modal variations
26 among the 'usual' minerals of gabbros (olivine, pyroxenes, plagioclase, Fe,Ti-oxides). Instead,
27 it implies that the protolith was subjected to Ca-metasomatism prior to metamorphism. Ca-
28 metasomatism of basic igneous rocks in a variety of geological settings is well documented
29 and includes (i) rodingite formation accompanying serpentinisation of peridotite (e.g.
30 Coleman 1967; Honnorez and Kirst 1975; Wenner 1979; Austrheim and Prestvik 2008), (ii)
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epidosite formation during high-*T* hydrothermal alteration of oceanic crust at ocean ridges (e.g. Bickle and Teagle 1992; Nehlig et al. 1994; Jowitt et al. 2012, Wang et al. 2012) and fore-arcs (e.g. Banerjee et al. 2000), and (iii) hydrothermal alteration of high-level dolerite and gabbro intrusions in a continental setting (e.g. Dickin and Jones 1983; Rose and Bird 1994). However, in the absence of the protolith composition, it is impossible to establish with certainty which, if any, of these types of alteration affected the Lanzada metagabbro, particularly as (i) they are likely to generate similar geochemical trends (e.g. decreasing Na₂O and LILE), and (ii) the degree of metasomatic alteration was probably modest (as testified by e.g. the unexceptional Na₂O content).

Ca enrichment is one of several types of metasomatism for which there is evidence in Alpine ophiolitic rocks (Widmer et al. 2000). As well as rodingitisation (Vuagnat 1967; Li et al. 2008; Zanoni et al. 2011) and epidotisation (Bowtell et al. 2007), metabasalts of the Zermatt-Saas zone collectively show evidence of spilitisation (Beauregard and Stern 1971) and hydrothermal chloritisation (Martin and Tartarotti 1989; Widmer et al. 2000; Martin et al. 2008). On the MgO-CaO-Na₂O plot of Widmer et al. (2000), which summarises the geochemical trends associated with these types of alteration, the Lanzada metagabbro plots in the part of the diagram characterised by rodingitisation and epidotisation (Fig. 6). However, it plots well to the high-Na side of the rodingitisation trend defined by the metarodinite data of Evans et al. (1981) and arguably matches the poorly defined epidotisation trend better (Fig. 6).

If the likelihood of metasomatism is accepted, there is little scope for using the geochemistry of the metagabbro to elucidate the tectonic setting of its protolith. One exception may be the use of the relatively immobile trace elements Ti, Nb and Y. On the TiO₂ versus Y/Nb plot of Floyd and Winchester (1975), the Lanzada metagabbro plots within the field of tholeiitic basalts but in the area of overlap between continental and oceanic tholeiites (Fig. 7). However, acceptance of this result would require us to assume that the trace element

concentrations were not modified by cumulus processes during crystallisation of the gabbro protolith, which is hard to justify.

5 Thermobarometry

The high variance of the Lanzada metagabbro mineral assemblage means that few equilibria are available for constraining its peak-metamorphic P-T conditions. Three well-calibrated equilibria may be written amongst end-member components of the eclogitic assemblage Grt + Omp + Phe:

garnet omphacite garnet omphacite

garnet garnet phengite omphacite phengite

garnet phengite garnet phengite.

Equilibrium A is temperature-sensitive and has been calibrated as a geothermometer by many workers, including Råheim and Green (1974), Ellis and Green (1979), Krogh (1988), Pattison and Newton (1989), Ai (1994) and Krogh-Ravna (2000). Equilibrium C can also be used as a thermometer and has been calibrated for metabasic compositions by Krogh and Råheim (1978). Equilibrium B is pressure-sensitive and has been calibrated as a geobarometer by Krogh-Ravna and Terry (2004).

The application of equilibria A, B and C relies on the identification of equilibrated garnet, omphacite and phengite compositions. The variability of garnet and omphacite chemistry indicates that the metagabbro did not achieve equilibrium on a thin-section scale. Garnet cores and omphacites from the interiors of omphacite aggregates clearly failed to equilibrate at the peak of metamorphism. Because Fe-Mg partitioning between garnet and

clinopyroxene (controlled by equilibrium A) results in $(\text{Mg/Fe})_{\text{Cpx}} > (\text{Mg/Fe})_{\text{Grt}}$ and decreases with increasing temperature, zoning profiles in eclogitic garnets that show smoothly increasing $(\text{Mg/Fe})_{\text{Grt}}$, such as observed in our data, may be assumed to record prograde garnet growth. Mn is also strongly partitioned into garnet. Thus, the lack of Fe,Mn-enriched garnet outer rims argues that garnet and omphacite underwent little or no retrogressive Fe-Mg exchange, and thus that the garnet outer Mg-rich rims represent the thermal peak of metamorphism. For present purposes, therefore, the compositions chosen for thermobarometry were the garnet rim with the highest $\text{Mg}/(\text{Fe}+\text{Mg})$ value (0.107), the adjacent omphacite, and the mean phengite.

Application of the equation of Krogh-Ravna and Terry (2004) for equilibrium B requires input of activity values for the component end-members. Activity-composition relations of minerals used with these calculations were: *omphacite*: ordered $P2/n$ pyroxene model of Holland (1990); *white mica*: ideal ionic mixing; *garnet*: regular solution model with $W_{\text{py-alm}} = 2.5 \text{ kJmol}^{-1}$ and $W_{\text{py-gr}} = 33 \text{ kJmol}^{-1}$, as used in the program AX of Holland & Powell (<ftp://www.esc.cam.ac.uk/pub/minp/AX/>), but multiplied by $(X_{\text{Si}}^{\text{IV}})^3$ to take account of the minor Si=4H substitution on the tetrahedral sites.

The independent calibrations of Krogh-Ravna (2000) and Krogh-Ravna and Terry (2004) for equilibria A and B, respectively, yield an intersection at $1.99 \pm 0.32 \text{ GPa}$ and $525 \pm 80 \text{ }^\circ\text{C}$ (Fig. 8). The calibration of Krogh and Rheim (1978) for equilibrium C suggests a slightly higher temperature and, if given equal weight to A and B, yields a ‘best-fit’ intersection close to 1.9 GPa and $530 \text{ }^\circ\text{C}$ (Fig. 8).

Apart from the errors inherent in the calibrations, the main potential sources of error are (i) uncertainties in the Fe^{3+} contents of the minerals, and (ii) the susceptibility of phengite to retrograde diffusional Fe-Mg exchange. With regard to the first point, the temperatures returned by Fe,Mg-exchange thermometers are particularly sensitive to errors in the Fe^{3+}

1 contents of clinopyroxene and phengite because these phases have low Fe/Mg ratios (as
2 discussed for Grt-Cpx thermometry by Krogh Ravna 2000). In principle, Fe³⁺ contents could
3 be obtained on omphacite separates by Mössbauer spectroscopy, but given the spatial
4 variation in omphacite chemistry, this method could not produce trustworthy Fe³⁺ contents for
5 the omphacites in equilibrium with garnet rims. Allowing some of the Fe in phengite to be
6 Fe³⁺ has no effect on the temperature returned by the thermometer of Krogh and Råheim
7 (1978) because this calibration rests on the assumption that all Fe in phengite is Fe²⁺.
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9 Fortunately, the fact that diopside and celadonite activities are relatively insensitive to the
10 Fe³⁺ contents of their host minerals means that the pressures generated by equilibrium B are
11 robust.
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13 With regard to the issue of retrograde Fe-Mg exchange between mica and other phases,
14 Spear (1993) has shown that the thermometric error will be greatest where the mica mode is
15 small compared to that of other Fe,Mg-silicates, as in this case. Even though no retrograde
16 Fe,Mg-exchange is detectable in garnet rims, the possibility remains that the compositions of
17 the rare phengites were modified during cooling, which casts doubt on the robustness of
18 garnet-phengite Fe-Mg exchange thermometry. In view of this, we prefer to take the
19 intersection between equilibria A and B alone (1.99 GPa, 525°C; Fig. 8) as the best estimate
20 of peak-metamorphic conditions.
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22 The geobarometric results clearly indicate that the Lanzada metagabbro underwent
23 metamorphism at modest temperature and great depth (*ca.* 70 km). The implied ‘average’
24 geothermal gradient (obtained by dividing peak-metamorphic temperature by peak-
25 metamorphic depth) is *ca.* 7 ± 2 °C/km.
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27 **6 Discussion and conclusions**

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1 The *P-T* conditions recorded by the Lanzada metagabbro are comparable to those
2 recorded by high-pressure rocks in other ophiolitic units of the Alps, and they map onto the *P-*
3 *T* array defined by those data (Fig. 8). As with those other ophiolitic units, the low ‘average’
4 geothermal gradient obtained for the Lanzada metagabbro is entirely consistent with
5 metamorphism in a subduction zone.
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11 This is the first record of high-pressure metamorphism in the Lanzada window. The
12 implication of this finding is that at least some parts of the cover sequence of the LSZ
13 underwent subduction during Alpine ocean closure. Whether or not the whole of the LSZ
14 cover was subducted to depths of ≥ 70 km is debateable. If it did, the other rocks within the
15 window must have had eclogite-facies mineral assemblages totally overprinted by
16 retrogressive assemblages. Given the small degree of retrogression in the metagabbro (which
17 is consistent with the low grade of Lepontine regional metamorphism in the Penninic units
18 east of the Bergell intrusion – Frey and Ferreiro-Mählmann 1999), this seems unlikely. An
19 alternative scenario is one in which the LSZ comprises a tectonic mélange within which small
20 fault-bouded lenses of subducted rock have become tectonically interleaved with lower-grade
21 rocks during exhumation, as documented in some other subduction complexes (e.g. the Sistan
22 Suture Zone, Iran – Rad et al. 2005). Establishing which of these two scenarios is correct will
23 require further field and petrographic work.
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43 Given the existence of blueschist-facies mineral assemblages in the Suretta nappe and
44 Avers Bündnerschiefer, and the absence of high-*P* assemblages in the Malenco unit, Platta,
45 Margna and Sella nappes, our findings support Montrasio’s (1984) correlation of the
46 ophiolitic part of the LSZ with the Avers Bündnerschiefer and the orthogneisses with the
47 Suretta nappe. If true, this would mean that the Avers unit preserves a southwards increase in
48 peak pressure of *ca.* 0.7 GPa over a map distance of *ca.* 30 km, providing further evidence in
49 support of a southerly-dipping subduction zone. The only other unit for which a peak-*P*
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1 gradient is documented for high-*P* metamorphism is the Adula nappe (Heinrich 1986), in
2 which peak pressure increases to the south by *ca.* 1.0 GPa over a map distance of *ca.* 40 km
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4 (Dale and Holland 2003). The correspondence between these two gradients is remarkable,
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6 especially when one recalls (a) that the LSZ and Adula nappe were subducted at different
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8 times and locations, and (b) that the present-day map-lengths of these units probably bear
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10 only an approximate relation to their true down-dip lengths within their respective subduction
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12 zones, as both units have undergone several phases of post-high-*P* plastic deformation (Ring
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14 1992; Zulbati 2008).
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19 If the correlation of the ophiolitic component of the LSZ with the Avers
20 Bündnerschiefer is correct, the Lanzada metagabbro is the most south-easterly-derived rock
21 discovered so far in the eastern Central Alps that provides direct evidence of subduction of
22 the Piemont-Ligurian oceanic crust. The peak pressure recorded by the Lanzada metagabbro
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24 is much higher than that recorded by the overlying Malenco unit and Margna nappe (Benning
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26 and Sidler 1992; Bissig and Hermann 1999; Müntener et al, 2000), indicating that the base of
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28 the Malenco unit represents an important metamorphic and tectonic discontinuity. That the
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30 retrograde P-T path of the Lanzada metagabbro was also different from that of the Malenco
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32 unit is qualitatively indicated by marked differences in the retrograde amphibole compositions
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34 (Fig. 5).
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43 Bachmann et al. (2009) assumed that the deformation that they dated within the
44 structurally higher (and therefore more south-easterly-derived) Platta nappe and lowermost
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46 Austroalpine basement reflected subduction, but as no high-*P* metamorphic relics have been
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48 found in those units, the validity of the assumption is questionable. Equally possible, in our
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50 view, is that the dated deformation reflects a phase of crustal (re-)thickening by nappe
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52 stacking at the northern margin of Apulia prior to the onset of ‘deep’ subduction north-west of
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54 the original position of the Malenco unit. If, as argued by Froitzheim and Manatschal (1996)
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1 and Froitzheim et al. (1996), the protoliths of the Platta nappe became part of the Piemont-
2 Ligurian ocean floor by denudation of the mantle along extensional normal faults rather than
3 by spreading at an ocean ridge, then the deformation dated by Bachmann et al. (2009) within
4 the Platta and lowermost Austroalpine nappes might merely reflect compressive (re-
5)thickening of the crust and restoration of the original superposition of Apulian continental
6 crust upon Platta mantle. Such motion would still record an early phase of orogenic
7 compression but would not require subduction.
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10 Our geochemical data do not allow us to determine the original tectonic setting of the
11 Lanzada metagabbro protolith. There is clear evidence that some Alpine (meta-)gabbros are of
12 Permian age and of sub-continental origin (e.g. the Braccia gabbro of the Malenco unit –
13 Müntener and Hermann 1996; Hermann et al. 1997; Trommsdorff et al. 2005 –, the Fedoz
14 gabbro of the Margna nappe – Ulrich and Borsien 1996 – and the Mont Collon gabbro of the
15 Dent Blanche klippe – Monjoie et al. 2005) while others are of Jurassic age and of oceanic
16 origin (e.g. the Mellichen and Allalin gabbros of the Zermatt-Saas zone – Rubatto et al. 1998).
17 If the Lanzada metagabbro formed originally at an ocean ridge, its inferred Ca-metasomatism
18 could have been due to epidotisation caused by ridge-related hydrothermal fluid circulation
19 (e.g. Bickle and Teagle 1992). If, on the other hand, the Lanzada metagabbro was originally
20 sub-continental, its inferred Ca-metasomatism is unlikely to have occurred while it was
21 situated close to the crust-mantle boundary as there is little scope in that environment for the
22 circulation of the necessary hydrous fluids. However, it is noteworthy that the LSZ contains
23 ophicarbonates; similar rocks in the Malenco unit have been interpreted by Hermann and
24 Müntener (1996) as having formed when sub-continental mantle peridotite was exposed at the
25 ocean floor by rifting-related extensional normal faulting. Serpentinisation of the peridotites
26 during this process could have resulted in partial rodingitisation of associated gabbro bodies.
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1 The age of the high-*P* metamorphism of the Lanzada metagabbro is unknown. If the
2 correlation of the LSZ with the Avers Bündnerschiefer is correct, an age within the range 50 –
3
4 40 Ma would be expected. Dating the eclogitic assemblage may be achievable through U-Pb
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6 dating of titanite or Sm-Nd dating of garnet-omphacite pairs, though the latter would be
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8 complicated by the documented thin-section-scale disequilibrium.
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11 **7 Conclusions**

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19 1. Epidote-rich eclogitic metagabbro forms a small body within the Lanzada Window,
20 where it is associated with serpentinites and supracrustal rocks of the Lanzada – Santa Anna
21 Zone (LSZ). This is the first record of high-pressure metamorphism in the Lanzada Window
22 and in the Malenco region.
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- 29 2. The peak-metamorphic mineral assemblage (garnet + omphacite + epidote + phengite
30 + titanite + apatite + pyrite) crystallised at *ca.* 2.0 GPa and 525 °C and underwent only limited
31 retrogressive alteration to an albite-epidote-amphibolite-facies assemblage. The deduced P-T
32 conditions imply that at least parts of the LSZ were subducted to depths of *ca.* 70 km.
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- 39 3. The presence of eclogite within the LSZ strengthens the correlation of this unit with
40 the blueschist-bearing Avers Bündnerschiefer, and confirms the former existence of a
41 southerly-dipping subduction zone beneath the Malenco unit. If the correlation is true, the
42 Lanzada metagabbro is the most south-easterly-derived rock discovered so far in the eastern
43 Central Alps that provides direct evidence of subduction of Piemont-Ligurian oceanic crust.
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- 50 4. The Lanzada metagabbro shows trace-element geochemical affinities with tholeiitic
51 basalts. The bulk composition is Ca-rich (wollastonite-normative), suggesting that the rock
52 underwent Ca-metasomatism prior to high-*P* metamorphism. The metasomatism presumably
53 occurred either by partial epidotisation during ocean-ridge-related hydrothermal circulation or
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1 by partial rodingitisation during extensional denudation and serpentinitisation of mantle during
2 slow spreading of the Piemonte-Ligurian ocean in the Jurassic.
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Figure captions

Fig. 1. Tectonic map of the eastern Central Alps (simplified from Montrasio et al., 2005, with boundaries extrapolated across unexposed areas). Star: eclogite locality. A-G: Adula Nappe including Gruf mass; Av: Avers Bündnerschiefer; B-D: Bellinzona-Dascio Zone; Ca: Campo and Languard nappes; Ch: Chiavenna Ophiolite; E-B: Err-Bernina nappe complex; EL: Engadine Line; Gr: Grosina Nappe; IL: Insubric Line; M-F: Malenco-Forno Unit; Ma: Margna Nappe; Pl: Platta Nappe; Se: Sella Nappe; Su: Suretta Nappe; Ta: Tambo Nappe; To: Tonale Series.

Fig. 2. (a) Photomicrograph showing the irregular lobate rim region of a garnet porphyroblast (Grt) and associated omphacite (Cpx), epidote (Ep), titanite (Ttn) and retrograde albite (Ab) and hornblende (Hbl). (b) Photomicrograph showing retrograde hornblende (Hbl) cutting across a foliation defined by aligned omphacite (Cpx) (locally rich in dark fluid inclusions) and epidote (Ep) and containing relict inclusions of those minerals.

Fig. 3. Compositional zoning profile across a garnet porphyroblast.

Fig. 4. Clinopyroxene compositions shown on the diagram of Essene and Fyfe (1967). 'Z-S Cpx': mean of 11 omphacite analyses from the Zermatt-Saas zone (Ernst & Dal Piaz 1978).

Fig. 5. Amphibole compositions shown on the $100\text{Na}/(\text{Na}+\text{Ca})$ vs. $100\text{Al}/(\text{Al}+\text{Si})$ plot of Laird and Albee (1981), in comparison with retrograde hornblendes of the Braccia Gabbro in the Malenco unit (Müntener et al. 2000), hornblendes of the Margna nappe (Benning & Sidler 1992), and hornblendes and barroisites of the Zermatt-Saas zone (Ernst & Dal Piaz 1978).

Fig. 6. Bulk composition of the Lanzada metagabbroic eclogite compared with compositions of variably metasomatised metabasalts from the Zermatt-Saas Zone, Western Alps, on

the CaO–MgO–Na₂O (wt.%) plot of Widmer et al. (2000). Field of mid-ocean ridge basalts (MOR) and all Zermatt-Saas data from Widmer et al. (2000). Compositions of Cima Lunga basalt-metarodingite suite from Evans et al. (1981).

Fig. 7. Bulk composition of the Lanzada metagabbroic eclogite on the TiO₂ versus Y/Nb plot of Floyd and Winchester (1975).

Fig. 8. Garnet-clinopyroxene-phengite thermobarometry of the Lanzada metagabbroic eclogite. See text for explanation. Mineral component abbreviations from Kretz (1983). Circle: approximate ‘best-fit’ intersection generated by independently calibrated thermobarometers. Shaded area: error box for equilibria A and B. Facies boundaries from Okamoto and Maruyama (1999). *P-T* conditions estimated for high-*P* metamorphism in other Penninic ophiolitic units in the Alps: AL: Zermatt-Saas Zone - Allalin Gabbro (Bucher and Grapes, 2009); AV: Avers Bundnerschiefer (Ring, 1992); LC: Lago di Cignana Unit (Reinecke, 1991); LZ: Lanzo peridotite (Pelletier and Muntener, 2006); MV: Monviso Complex (Messiga et al., 1999); SM: Zermatt-Saas Zone - St Marcel (Martin et al., 2008); TE: Tauern Eclogite Zone (Hoschek, 2007); TG: Tauern Glockner Nappe (Dachs and Proyer, 2001); ZS: Zermatt-Saas Zone (Angiboust et al., 2009).

Table headings

Table 1. Representative electron-microprobe analyses of peak-metamorphic and retrograde minerals. Garnet analyses 1, 2 and 4 are from the same porphyroblast; epidote analyses 1 and 2 are from the same crystal; all other analyses are from different crystals. Abbreviations: ‘cont.’: contacting; ‘dist.’: distant from; ‘Cpx agg.’: in interior of Cpx aggregate; ‘poik.’: poikiloblast.

Table 2. Bulk chemical composition of the Lanzada metagabbroic eclogite.

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Figure 1

Figure 2a
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Figure 2b
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Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Figure 8

Table 1

	Garnet 1	Garnet 2	Garnet 3	Garnet 4	Omphacite 1	Omphacite 2	Omphacite 3	Omphacite 4	Epidote 1	Epidote 2	Epidote 3	Phengite 1	Phengite 2	Titanite 1	Homblende 1	Homblende 2	Albite 1
	core	intermed.	near rim	rim cont. Cpx	core Cpx agg.	core near Grt	core cont. Ep	rim cont. Grt	core dist. Cpx	rim dist. Cpx	rim cont. Grt	core	core	small cont. Cpx	core	rim cont. Cpx	poik
<i>Oxides</i>																	
SiO2	37.20	37.50	37.66	37.63	54.71	55.65	55.06	55.55	38.66	38.29	36.94	49.50	49.49	30.34	49.18	48.93	67.87
TiO2	0.16	0.11	0.08	0.07	0.24	0.06	0.07	0.04	0.13	0.13	0.11	0.35	0.41	38.60	0.20	0.20	0.00
Al2O3	21.14	21.05	21.35	21.63	6.86	10.95	10.31	10.80	29.35	27.00	22.19	29.19	28.93	1.30	7.72	8.13	20.11
Cr2O3	0.03	0.02	0.02	0.01	0.00	0.02	0.04	0.02	0.03	0.02	0.03	0.01	0.00	0.02	0.01	0.03	0.00
Fe2O3	-	-	-	-	2.26	3.59	4.60	4.82	5.87	8.59	15.08	0.00	0.00	-	4.25	4.55	0.10
FeO	27.77	27.64	26.18	25.18	4.23	4.18	3.92	3.16	-	-	-	3.12	3.18	0.32	8.80	9.03	-
MnO	2.01	1.89	1.06	1.51	0.25	0.31	0.32	0.25	0.05	0.20	0.37	0.05	0.02	0.07	0.45	0.46	0.00
MgO	1.16	1.27	1.70	1.69	10.10	6.34	6.25	6.46	0.05	0.04	0.00	3.26	3.23	0.01	13.44	13.11	0.00
CaO	10.53	10.82	12.18	12.84	17.01	11.31	11.42	10.98	23.88	23.62	23.09	0.04	0.03	28.37	9.52	9.66	0.29
Na2O	0.00	0.00	0.00	0.00	4.61	7.83	7.74	8.08	0.00	0.00	0.00	0.51	0.57	0.01	2.96	2.79	11.46
K2O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	9.97	9.89	0.00	0.36	0.43	0.04
Total	100.00	100.30	100.23	100.57	100.26	100.24	99.73	100.17	98.02	97.89	97.82	95.97	95.74	99.04	96.88	97.30	99.86
<i>Formula</i>	12(O)	12(O)	12(O)	12(O)	6(O)	6(O)	6(O)	6(O)	12.5(O)	12.5(O)	12.5(O)	11(O)	11(O)	5(O)	23(O)	23(O)	8(O)
Si	2.976	2.988	2.982	2.966	1.978	1.991	1.987	1.986	2.987	2.996	2.971	3.289	3.297	0.999	7.151	7.100	2.969
Ti	0.009	0.006	0.005	0.004	0.006	0.002	0.002	0.001	0.008	0.008	0.007	0.017	0.020	0.956	0.022	0.022	0.000
Al	1.994	1.977	1.992	2.009	0.292	0.462	0.439	0.455	2.673	2.490	2.104	2.286	2.271	0.050	1.323	1.391	1.037
Cr	0.002	0.001	0.001	0.001	0.000	0.001	0.001	0.001	0.002	0.001	0.002	0.001	0.000	0.000	0.001	0.003	0.000
Fe3+	-	-	-	-	0.061	0.097	0.125	0.130	0.341	0.506	0.913	0.000	0.000	-	0.465	0.497	0.003
Fe2+	1.858	1.842	1.733	1.660	0.128	0.125	0.118	0.095	-	-	-	0.173	0.177	0.009	1.070	1.096	-
Mn	0.136	0.128	0.071	0.101	0.008	0.009	0.010	0.008	0.003	0.013	0.025	0.003	0.001	0.002	0.055	0.056	0.000
Mg	0.139	0.151	0.201	0.199	0.544	0.338	0.336	0.344	0.006	0.005	0.000	0.322	0.321	0.000	2.913	2.835	0.000
Ca	0.902	0.924	1.033	1.084	0.659	0.433	0.441	0.421	1.977	1.980	1.990	0.003	0.002	1.001	1.482	1.501	0.014
Na	0.00	0.00	0.00	0.00	0.323	0.543	0.541	0.560	0.000	0.000	0.000	0.065	0.073	0.001	0.833	0.784	0.972
K	0.00	0.00	0.00	0.00	0.000	0.000	0.000	0.000	0.000	0.000	0.001	0.845	0.841	0.000	0.067	0.080	0.002
Total	8.017	8.017	8.017	8.025	4.000	4.000	4.000	4.000	7.997	7.999	8.013	7.005	7.004	3.019	15.382	15.365	4.998
Mg/(Mg+Fe2)	0.069	0.076	0.104	0.107	0.810	0.730	0.740	0.784	-	-	-	0.650	0.644	-	0.731	0.721	-

Table 2

<i>Major oxides</i>	Weight % oxides	Weight % oxides	Weight % oxides
	<i>Lanzada metagabbro (XRF)</i>	<i>Lanzada metagabbro (anhydrous normalised)</i>	<i>Average of 1317 gabbros (anhydrous normalised) & std. deviations (LeMaitre, 1976)</i>
SiO ₂	45.24	46.24	51.06 ±4.56
TiO ₂	1.03	1.05	1.17 ±1.00
Al ₂ O ₃	14.57	14.90	15.91 ±3.66
Fe ₂ O ₃	11.13	6.13	3.10 ±2.15
FeO	-	4.72	7.76 ±3.06
MnO	0.48	0.49	-
MgO	4.57	4.67	7.68 ±3.48
CaO	17.74	18.13	9.88 ±2.78
Na ₂ O	3.00	3.07	2.48 ±1.17
K ₂ O	0.29	0.30	0.96 ±0.85
P ₂ O ₅	0.28	0.28	-
SO ₃	0.02	0.02	-
H ₂ O	0.68	-	-
Total	99.04	100.00	100.00

Trace elements (ppm): Lanzada metagabbro

Sc	7.8	Sn	8.6
V	153.0	Sb	bd
Cr	274.2	Te	27.0
Co	66.7	I	bd
Ni	185.5	Cs	4.6
Cu	60.6	Ba	54.9
Zn	77.5	La	55.0
Ga	22.6	Ce	103.1
Ge	2.9	Nd	41.1
As	5.7	Sm	8.8
Se	bd	Yb	bd
Br	bd	Hf	6.9
Rb	7.7	Ta	bd
Sr	489.3	Hg	bd
Y	45.6	Tl	bd
Zr	139.9	Pb	25.4
Nb	13.0	Bi	bd
Mo	0.9	Th	14.5
Ag	bd	U	5.3
Cd	bd		

CIPW Weight Norm: Lanzada metagabbro

Plagioclase	39.77	Diopside	30.32
<i>An</i>	25.53	<i>Di</i>	25.65
<i>Ab</i>	14.24	<i>Hd</i>	4.67
Orthoclase	1.77	Wollastonite	10.00
Nepheline	6.57	Ilmenite	1.99
Na ₂ SO ₄	0.04	Magnetite	8.89
Apatite	0.65		

Supplementary Material

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