# Variation of mineral composition, fabric and oxygen fugacity from massive to foliated eclogites during exhumation of subducted ocean crust in the North Qilian suture zone, NW China

Y. CAO,<sup>1</sup> S. G. SONG,<sup>1</sup> Y. L. NIU,<sup>2</sup> H. JUNG<sup>3</sup> AND Z. M. JIN<sup>4</sup>

<sup>1</sup>MOE Key Laboratory of Orogenic Belts and Crustal Evolution, School of Earth and Space Sciences, Peking University, Beijing 100871, China (sgsong@pku.edu.cn)

<sup>2</sup>Department of Earth Sciences, Durham University, Durham DH1 3LE, UK

<sup>3</sup>Tectonophysics Laboratory, School of Earth and Environmental Sciences, Seoul National University, Seoul 151-747, Korea <sup>4</sup>State Key Laboratory of Geological Processes and Mineral Resources, China University of Geosciences, 388 Lumo Road, Wuhan 430074, China

ABSTRACT Eclogites from the North Qilian suture zone are high-pressure low-temperature metamorphic rocks of ocean crust protolith, and occur in both massive and foliated varieties as individual blocks of tens to hundreds of metres in size. The massive type is weakly deformed and shows granoblastic texture characterized by a coarse-grained peak mineral assemblage of  $Grt^1 + Omp^1 + Ph + Rt \pm Lws$  (or retrograde Cz). In contrast, the foliated type is strongly deformed and shows a fine-grained retrograde mineral assemblage of  $\text{Grt}^2 + \text{Omp}^2 + \text{Cz} + \text{Gln} + \text{Ph.}$  Both total FeO and aegirine contents in omphacite, as well as  $X_{\text{Fe}}[=\text{Fe}^{3+}/(\text{Fe}^{3+} + \text{Al}^{\text{VI}})]$  in clinozoisite/epidote, increase significantly from massive to foliated eclogites. Lattice preferred orientation (LPO) of omphacite, determined by electron back-scatter diffraction analysis, is characterized by weak and strong SL-type fabrics for massive and foliated eclogites, respectively. Clinozoisite/epidote also developed SL-type fabric, but different from the LPOs of omphacite in <010> and <001> axes, owing to their opposite crystallographic long and short axis definitions. The transition of deformation mechanism from dislocation creep to diffusive mass transfer (DMT) creep in omphacite and the concomitant retrograde metamorphism both are efficiently facilitated when the original coarse-grained  $Omp^1 + Grt^1 + Lws$  assemblage is dynamically recrys-tallized and retrogressed into the fine-grained Fe<sup>3+</sup>-rich assemblage of  $Omp^2 + Grt^2 + Cz + Gln$ . The DMT process with concomitant anisotropic growth assisted by fluids is considered to be an important deformation mechanism for most minerals in the foliated eclogite. P-T estimates yielded 2.3–2.6 GPa and 485–510 °C for the massive eclogite and 1.8–2.2 GPa and 450–480 °C for the foliated eclogite. The significant increase in total Fe and  $Fe^{3+}$  contents in omphacite and clinozoisite/epidote from massive to foliated eclogite suggests changes in mineral compositions accompanied by an increase in oxygen fugacity during ductile deformation associated with exhumation. The LPO transition of omphacite, clinozoisite and rutile from weak SL-type in massive eclogites to strong SL-type in foliated eclogites is interpreted to represent the increment of shear strain during exhumation along the 'subduction channel'.

**Key words:** exhumation; lattice preferred orientation; low-temperature eclogite; North Qilian suture zone; oxygen fugacity; retrograde metamorphism and deformation.

## INTRODUCTION

Orogenic eclogites are HP–UHP metamorphic rocks of basaltic/gabbroic protoliths with primarily a bi-mineralogical assemblage of garnet and omphacite. They record a complete history of subduction-zone metamorphism and later processes associated with exhumation. Because of the crystal habit and abundant cleavages, the omphacite in the matrix is weak and thus bears much of the strain during deformation, with garnet mostly showing a passive response and rotation (Zhang *et al.*, 2006; Zhang & Green, 2007a,b). This is commonly the case for eclogite formed at high temperatures (>600 °C) with minor or no hydrous phases. However, low-temperature (<550–600 °C) eclogite may be more complex because of the presence of hydrous phases such as sodic amphibole, epidote– clinozoisite–zoisite, chlorite, phengite and lawsonite.

Previous studies of eclogite deformation have contributed a great deal to the understanding of the rheology of the subducting slab (e.g. Avé Lallemant, 1978; Buatier *et al.*, 1991; Ji & Martignole, 1994; Karato *et al.*, 1994, 1995; Godard & van Roermund, 1995; Cordier *et al.*, 1996; Voegelé *et al.*, 1998a,b; Prior et al., 2000; Mauler et al., 2000, 2001; Piepenbreier & Stöckhert, 2001; Brenker et al., 2002; Terry & Heidelbach, 2004; Storey & Prior, 2005; Zhang et al., 2006; Zhang & Green, 2007a,b), seismic properties in the deep crust (e.g. Bascou et al., 2001; Ji et al., 2003; Wang et al., 2005) and exhumation process (e.g. Kurz et al., 2004; Kurz, 2005). These studies have, however, mainly focused on simple eclogite dominated by omphacite and garnet. Deformational behaviour of hydrous phases in eclogite remains poorly understood, and whether there is any compositional change during deformation is essentially unknown.

It is generally accepted that garnet from 'dry' eclogite is too strong to be deformed plastically because of its extremely high Pierls stress to resist the dislocation glide, thus having rather low dislocation density even at high temperatures (Karato et al., 1994, 1995; Ingrin & Madon, 1995). However, garnet showing a pronounced shape preferred orientation (SPO) requires certain deformation mechanisms such as grain boundary processes facilitated by diffusion creep (den Brok & Kruhl, 1996; Terry & Heidelbach, 2004, 2006; Storey & Prior, 2005; Zhang et al., 2006; Zhang & Green, 2007a,b) or dislocation creep promoted by thermal or hydrolytic weakening (Ji & Martignole, 1994; Cordier et al., 1996; Voegelé et al., 1998a,b; Wang & Ji, 1999; Su et al., 2003; Ji et al., 2003). In contrast, plastically deformed features of omphacite are more widely observed in nature, and a number of experimental studies have been done on both single crystal and polycrystalline diopside to simulate deformation processes of omphacite (e.g. Avé Lallemant, 1978; Raterron & Jaoul, 1991; Mauler et al., 2000; Zhang & Green, 2007a). However, application of these experimental results to natural omphacite is not straightforward because they are inadequate to explain the much greater strains obvious in actual rocks. As a result, other factors including lower melting point of omphacite relative to that of diopside (Buatier et al., 1991; Godard & van Roermund, 1995), high deviatoric stress (Philippot & Van Roermund, 1992), and hydrolytic weakening (Lorimer et al., 1972; Buatier et al., 1991; Su et al., 2003) have been invoked to explain this apparent inconsistency. Moreover, many studies on various LPOs (lattice preferred orientations) of omphacite [L-type, S-type, transitional SL- (S > L) and LS- (L > S) types originally defined by Helmstaedt et al., 1972], combined with TEM studies, suggest that dislocation creep is the dominant deformation mechanism in omphacite with the Burgers vectors of (100) < 001 >,  $\{110\} < 001 >$ and  $\{110\}1/2 < 110 >$  (van Roermund & Boland, 1981; van Roermund, 1983; Ingrin et al., 1991; Raterron et al., 1994). The development of LPOs has also been theoretically simulated by applying both single and multiple slip systems (e.g. Bascou et al., 2002). However, the inferred slip system (010) < 001 > from the LPOs in terms of the single slip model has not been observed as a dominant slip system so far. Some researchers thus invoke diffusive mass transfer (DMT) at grain boundaries to reconcile this difference (Godard & van Roermund, 1995; Mauler *et al.*, 2001). In addition, most authors believe that the variation between L- and S-type fabric is caused by finite strain geometry change rather than the temperature-induced omphacite space group transformation (Helmstaedt *et al.*, 1972; Mauler *et al.*, 2001; Brenker *et al.*, 2002; Kurz *et al.*, 2004; Kurz, 2005).

Field and petrographic observations indicate that plastic deformation is mostly associated with fluids (e.g. veins and oriented hydrous phases). Hence, it is possible that  $P-T-X_{H2O}$  variation and chemical changes may in fact be important in facilitating the deformation. However, little research exists on how the chemistry of minerals may change in response to plastic deformation. In this article, investigations on the petrography, major-element composition of minerals, microstructure and LPOs of omphacite, clinozoisite/epidote, glaucophane, and rutile in low-Teclogites from the North Qilian suture zone are presented. These results allow us to characterize deformation-induced micro-fabrics defined by newly formed minerals and existing minerals under change accompanied by the changing  $P-T-f_{\Omega_2}$  conditions. We then discuss the deformation mechanisms and the relationship between fabrics and exhumation process.

# GEOLOGICAL SETTING

The North Oilian suture zone is an elongated NW-stretching belt between the Alashan block, the western part of North China Craton to the north, and the Qilian-Qaidam block to the south, and is transected by the left-lateral strike-slip Altyn-Tagn Fault system in the west (inset of Fig. 1a). It consists of Early Palaeozoic subduction complexes including ophiolitic mélanges, blueschists and eclogites, Silurian flysch formations, Devonian molasse and Carboniferous to Triassic sedimentary cover sequences (Fig. 1a; see Song et al., 2007, 2009 for details). Three high-pressure metamorphic slices A, B and C, which trend sub-parallel to each other, are composed of eclogite, blueschist, meta-pelite and meta-greywacke. Eclogites of N- and E-type MORB or ocean island basalt (OIB) protoliths commonly occur as tensto-hundreds-metre-sized blocks surrounded by blueschist, metapelite or metagreywacke (Song et al., 2009). Previous thermobarometric calculations and phase equilibrium analysis yield peak P-T conditions of 2.10-2.65 GPa and 460-590 °C for the eclogite (Song et al., 2007; Zhang et al., 2007; Wei et al., 2009). Comparable results have also been obtained from the eclogite facies metapelites, i.e. 2.15-2.6 GPa and 470-540 °C (Song et al., 2007; Wei & Song, 2008). SHRIMP U-Pb dating of eclogitic zircon yields ages of 489-464 Ma (Song et al., 2004, 2006; Zhang et al., 2007), and Ar-Ar ages of glaucophane and phengite from high-grade blueschists range from



Fig. 1. Geological maps of the study area showing (a) middle part of North Qilian Suture Zone (after Song *et al.*, 2007), (b) rock assemblage of Baijingsi cross-section and (c) field occurrence for massive and foliated eclogites.

462 to 448 Ma (Liou *et al.*, 1989; Wu *et al.*, 1993; Zhang *et al.*, 1997; Liu *et al.*, 2006). All these data, combined with the recognition of lawsonite (Zhang &

Meng, 2006; Song *et al.*, 2007; Zhang *et al.*, 2007) in eclogite and Mg-carpholite in metapelites (Song *et al.*, 2007), confirm that the North Qilian HP metamorphic

belt is the product of 'cold' oceanic-type subduction in the Early Palaeozoic (Song *et al.*, 2006).

# FIELD OCCURRENCE AND SAMPLE PETROGRAPHY

The eclogite occurs as blocks of tens to hundreds of square metres in size in the three HP slices A, B and C (Fig. 1a). Their country rocks are blueschist to eclogite facies metamorphosed greywacke and pelite with blocks of serpentinite, marble and Fe-Mn-rich chert, forming a typical mélange (Song et al., 2009). On the basis of mineral assemblages, the North Qilian eclogites can be divided into phengite-eclogite (Type I) with a peak-stage mineral assemblage of Grt + Omp +  $Ph + Rt \pm Cz$  and epidote-eclogite (Type II) with an assemblage of  $Grt + Omp + Cz + Gln + Rt \pm$ Otz plus minor phengite and paragonite. Geochemical analyses suggest that Type I eclogite has an E-type MORB or OIB protolith, whereas Type II has an N-type MORB or gabbroic protolith (Song et al., 2006, 2009). Mineral abbreviations are after Whitney & Evans (2010).

The samples were collected in the Baijingsi crosssection.  $\sim 30$  km to the east of Oilian Town (Fig. 1a). Detailed petrology is given in Song et al. (2007) and Zhang et al. (2007). In this cross-section (Fig. 1b), most eclogites outcrop as individual 50-200 m<sup>2</sup> sized blocks, and are intercalated within metapelitic schist or blueschist country rocks. Most sedimentary country rocks are also eclogite facies metamorphosed with mineral assemblages of Grt + Omp + Ph + Gln + Ep + Qtz + Rt, Grt + Cld + Ph + Gln + Qtz +Rt and  $Grt + Cld + Car + Ph + Qtz + Rt \pm Gln$ with peak conditions of 2.15-2.60 GPa and 495-540 °C (Song et al., 2007). These eclogites are mostly epidote-eclogite (Type II), and can be subdivided into massive and foliated ones on the basis of field appearances and microstructures (Fig. 1c).

## Massive eclogite

The massive eclogite blocks commonly exhibit weak foliation/lineation and varying extent of retrograde overprinting. They show a typical granoblastic texture with 20–30 vol.% idioblastic garnet (~2 mm in size) uniformly distributed in the matrix of omphacite (~45–60 vol.%) and minor Cz + Ph + Rt (Fig. 2a,b). Omphacite, lawsonite and its pseudomorphs (rectangle-shaped Cz + Pg aggregates) occur as inclusions in idioblastic garnet, suggesting a peak metamorphic assemblage of Grt + Omp + Lws + Rt + Ph (Song *et al.*, 2007; Zhang *et al.*, 2007). In addition, some garnet also contains orientated clinozoisite inclusions in the core domain, suggesting that the deformation occurred during progressive metamorphism in the subduction zone.

Omphacite of the first generation (denoted as Omp<sup>1</sup>) shows a granoblastic texture with average grain size of

~0.5–1 mm; some grains are deformed, elongated, and show undulatory extinctions, and subgrains (Fig. 2c,d). Clinozoisite in the matrix is minor in modal content and some of the grains are interpreted as forming from lawsonite (Song *et al.*, 2007; Zhang *et al.*, 2007; Wei *et al.*, 2009). Rutile crystals are on average ~100  $\mu$ m in average size, ~1–2% in modal content, and occur both in the matrix and in garnet. A second generation of fine-grained equigranular (0.1–0.2 mm in size) glaucophane is also observed in the matrix.

Some massive eclogite samples also contain orientated  $Cz + Omp^2 (+Gln)$  bands (Fig. 2e,f), probably developed during retrograde metamorphism. These bands cut and overprint the early Grt  $+ Omp^1$  assemblage.

## **Foliated eclogite**

In contrast, foliated eclogite samples (e.g. 2Q22, 2Q19, 08BSY-38, 08BSY-39) are strongly deformed with the foliation defined by a fine-grained (0.05-0.1 mm in width) mineral assemblage of  $Cz + Omp^2 + Gln$ . Some of the foliated eclogites retain weakly strained domains in which the first generation of coarse-grained  $Grt^1 + Omp^1 + Ph$  is preserved (Fig. 3a,b,h). Garnet crystals in the weakly strained domains are euhedral, rich in randomly orientated mineral inclusions, and are partly replaced by late glaucophane or chlorite/epidote along cracks. On the contrary, garnet (Grt<sup>2</sup>) in the strongly deformed domains is somewhat elliptical and contains orientated Cz + Omp inclusions that are sub-parallel to the external foliation in the matrix (Fig. 3g,h), suggesting syntectonic growth and in equilibrium with matrix minerals of  $Omp^2 + Cz +$ Gln. The modal content of garnet in the foliated eclogite is  $\sim 10-15\%$ , significantly less than that in the massive eclogite.

Omphacite in the foliated eclogite shows two generations. Omp<sup>1</sup> of the first generation is coarse grained and is present within the weakly deformed domains.  $Omp^2$  of the second generation shows stretched columnar or lenticular shape with an assemblage of Cz + Gln (+Ph); these omphacite grains define the lineation, micro-folds and the overall foliations in the matrix (Fig. 3b-f,i). Both omphacite and clinozoisite display straight foliation-parallel grain boundaries and visible chemical zonation patterns, but lack dislocation-related intracrystalline plasticity features. Fine-grained glaucophane is mainly distributed in the matrix parallel to the foliation, whereas coarse-grained glaucophane is oriented slightly oblique to the foliation. Rutile (on average  $\sim 100 \ \mu m$  in size, 1–2 mod.%) is evenly spread in the matrix with the long axis subparallel to the foliation.

## Whole-rock compositions

All studied eclogite samples were analysed for bulk-rock major and trace elements to compare



**Fig. 2.** Photomicrographs of massive eclogites from Baishiya area (sample: Q05-1). All images are in cross-polarized light. (a, b) The garnet-rich eclogite showing an overall granoblastic texture with a weak foliation defined by coarse-grained  $Omp^1$ . (c) Coarse  $Omp^1$  grain exhibiting subgrained microstructure. (d) Undulatory extinction and subgrain microstructures of omphacite, which has the same composition as coarse-grained  $Omp^1$ . (e) Elongated  $Omp^2$  grains in the garnet-poor domain, fine-grained retrogressive clinozoisite occurs within and between omphacite grains. (f) Shape preferred orientation of a clinozoisite band intercalated between two garnet-rich domains, which indicates the fluid-assisted DMT creep.

compositional variation between massive and foliated eclogites (see Song *et al.*, 2010 for detailed analytical details). In the SiO<sub>2</sub> v. total alkali (TAS) diagram, two foliated eclogite samples (2Q22 & 08BSY38) plot in the

trachybasalt (alkaline) field and all others in the subalkaline basalt field (Fig. 4a). Table 1 shows that the massive eclogite has higher  $Al_2O_3$  (17.19–18.01 wt%) and MgO (7.47–8.49 wt%), but lower  $Fe_2O_3t$ 



**Fig. 3.** Microstructures of the foliated eclogites from Baijingsi area except for (b) which came from Xiangzigou area in slice B west of Qilian Town. (a) Development of a foliation after shear deformation is shown where randomly oriented coarse-grained  $Omp^1 + Grt$  (first generation) are severely altered into the fine-grained mineral assemblage of  $Omp^2 + Cz + Gln$  at relatively lower *P*–*T* conditions. Note randomly oriented inclusions in porphyroblastic garnet. (b) The relict coarse-grained  $Omp^1 + Ph$  in the foliated eclogite (m2Q22). (c–f) Four foliated eclogites in XZ section show a clear foliation and an increase of grain size, aspect ratio and SPO of Omp and Cz from (c) to (f). (c: 2Q22, d: 08BSY-38, e: 08BSY-39 and f: 2Q19). (g, h) Oriented Cz + Omp<sup>2</sup> inclusions in the syn-tectonic garnet (Grt<sup>2</sup>) (foliated eclogite 2Q19). (i) BSE image showing elongated grains, long straight grain boundaries between omphacite and epidote and chemical zonations along the foliation (2Q19). Optical photomicrographs are shown in cross-polarized light except for (g) and (i) in backscattered electron image.

(8.50–9.38 wt%) and CaO (7.38–8.77 wt%) than the foliated eclogite. In the primitive mantle normalized (Sun & McDonough, 1989) multi-element diagram (Fig. 4b), the two alkaline basaltic samples (2Q22 & 08BSY38) show patterns consistent with present-day OIB and others are similar to N-type MORB. Therefore, the geochemistry suggests that protoliths of massive and foliated eclogites are fragments of ocean crust that derived from ocean ridge and near ridge seamount or OIB (Song *et al.*, 2009).

## LATTICE PREFERRED ORIENTATION

Foliation and lineation of samples were determined by the SPO of minerals. Each sample was cut parallel to the XZ plane (X: parallel to the lineation, Z: normal to foliation) for measuring the LPO of minerals. Thinsections were made and coated with a 2–5 nm thick carbon layer to prevent charging under SEM. The subsequent electron back-scatter diffraction (EBSD) measurements were carried out in the Environmental SEM Laboratory, China University of Geosciences (Wuhan) (Nordlys-II & Channel 5.0 HKL EBSD system and Quanta 2000-type SEM) and Tectonophysics Laboratory, SEES of Seoul National University (Nordlys-II & Channel 5.0 HKL EBSD system and JEOL JSM-6380 SEM), using 20 kV accelerating voltage, 60 nm spot size and 15.0 mm working distance with 70° tilting of specimen. Electron backscattered patterns (EBSPs) were captured by Nordlys



**Fig. 4.** (a) TAS diagram and (b) primitive mantle normalized multi-element diagram for massive and foliated eclogite.

II Detector and indexed with HKL CHANNEL 5.0 software. In order to acquire LPO precisely, EBSPs were indexed manually.

For the measurement of LPOs, crystals only in the relatively flat foliation were chosen for indexing; inclusions in garnet or grains near garnet where the foliation is disturbed were excluded. In addition, it is difficult and not necessary to distinguish clinozoisite from epidote because of their structural similarity and compositional miscibility. Therefore, the LPO of clinozoisite is used to represent LPOs of both clinozoisite and epidote. Representative LPOs of garnet, omphacite, clinozoisite/epidote and rutile are obtained from one massive eclogite (Q05-1) and four foliated eclogites. They are shown in the pole figures (Figs 5–8). The misorientation index (*M*-index) is applied to compare their LPO strengths (see appendix for details).

## **Garnet fabrics**

The limited observations make it difficult to provide reliable LPO data for garnet in the foliated eclogite. The random LPO of garnet in massive eclogite (Fig. 5), like published garnet LPO data (e.g. Zhang *et al.*, 2006; Zhang & Green, 2007a,b), is consistent with high rigidity and isometric crystal structure of garnet.

## **Omphacite fabrics**

Omphacite from the massive eclogite (sample Q05-1) shows a weak SL-type fabric (*M*-index = 0.07) that is characterized by <001 > axis weakly distributed in a girdle in the foliation with maximum subparallel to lineation, and by <010 > axis distributed subnormal to the foliation (Fig. 6a). This weak LPO is consistent with the weak foliation observed microscopically, but could be influenced to some extent by deformed, locally distributed Omp<sup>2</sup> of the second generation (Fig. 2). The {110} and <100 > of omphacite are fairly scattered with a weak concentration normal to the foliation and perpendicular to lineation within the foliation, respectively. This weak SL-type fabric can be regarded as the product of deformation in transpression with a large component of axial compression (Bascou *et al.*, 2002).

In the four foliated eclogite samples (2Q19, 2Q22, 08BSY-38 & 08BSY-39), omphacite exhibits much stronger SL-type LPOs than that in the massive eclogite (Fig. 6b–e). These characteristics indicate that simple shear is prevalent in the deformation process (Bascou *et al.*, 2002; Zhang *et al.*, 2006). The M-indices of omphacite in the four foliated samples range from 0.20 to 0.22, which are significantly higher than that of massive eclogite and suggest stronger shear strain in foliated eclogites.

#### Clinozoisite/epidote fabrics

The epidote group minerals (monoclinic, with a space group  $P2_1/m$ ) are ubiquitous hydrous phases in low-T eclogite. However, the rheological property and LPO of epidote group minerals are rarely studied. We measured LPOs of clinozoisite from the one massive and four foliated eclogite samples (Fig. 7). All samples show that <010> axis is concentrated nearly parallel to the lineation with a weak girdle along the foliation, <001> axis aligned normal to the foliation with a strong maximum, and {110} aligned nearly parallel to the foliation. This LPO can also be termed as SL-type fabric. Similar to omphacite, the M-index of clinozoisite in massive eclogite (M = 0.07) is significantly lower than that in the four foliated eclogites (M = 0.18 - 0.20). These values are also able to suggest the increment of shear strain from massive to foliated eclogite. However, most clinozoisite in the massive eclogite occurs as the second generation in the deformation bands with Omp<sup>2</sup> (Fig. 2e,f) and the SL-type fabric may reflect the shear deformation overprinting.

	Q98-114 <sup>a</sup>	Q98-116 <sup>a</sup>	Q98-118 <sup>a</sup>	Q05-1	2Q19	2Q22	8BSY38	8BSY39	Q98-126 <sup>a</sup>	Q98-128ª
Sample	Massive	Massive	Massive	Massive	Foliated	Foliated	Foliated	Foliated	Foliated	Foliated
Major elem	ents (wt%)									
SiO <sub>2</sub>	47.04	48.33	49.42	50.39	48.07	48.57	44.95	45.45	47.99	45.78
TiO <sub>2</sub>	0.90	1.15	0.82	0.61	1.78	2.71	2.93	1.44	1.64	1.56
$Al_2O_3$	17.22	18.01	17.19	17.81	14.62	13.46	13.36	15.66	15.87	16.14
Fe <sub>2</sub> O <sub>3</sub> t	8.70	8.50	8.77	9.38	10.73	10.58	10.59	10.48	10.91	12.19
MnO	0.18	0.17	0.25	0.13	0.17	0.16	0.19	0.16	0.21	0.20
MgO	8.29	7.80	8.49	7.47	5.92	7.64	8.95	6.65	6.63	7.15
CaO	8.77	8.85	7.38	8.07	11.88	9.69	9.13	12.20	8.61	7.95
Na <sub>2</sub> O	2.90	1.69	2.70	3.56	2.68	3.10	3.22	3.35	2.80	1.80
K <sub>2</sub> O	0.46	2.25	1.12	0.12	0.34	1.95	2.04	0.20	1.15	1.00
$P_2O_5$	0.13	0.22	0.18	0.14	0.11	0.61	0.73	0.21	0.25	0.25
LOI	n.a.	n.a.	n.a.	2.38	4.27	2.11	3.65	4.62	n.a.	n.a.
Total	94.61	96.97	96.32	100.06	100.56	100.59	99.75	100.41	96.06	94.01
Trace eleme	nts (ppm)									
Li	33.09	20.93	40.03	2.98	28.06	10.27	8.70	15.07	20.87	42.75
Sc	29.11	30.09	28.73	11.19	28.02	20.06	21.20	34.76	33.95	38.20
V	187.62	201.17	198.61	150.18	221.20	224.00	245.60	271.80	218.75	247.94
Cr	446.42	327.42	313.27	7.74	181.87	198.38	212.80	225.20	277.61	162.38
Co	63.74	42.59	34.75	17.69	36.16	41.42	53.02	34.68	43.89	41.21
Ni	168.87	122.12	90.33	8.74	79.30	99.90	190.78	87.76	115.69	56.26
Cu	125.72	52.23	674.12	29.40	52.86	45.80	54.26	56.96	52.76	51.71
Zn	95.96	76.79	131.87	72.42	86.52	124.98	137.44	86.92	81.95	76.29
Ga	14.55	16.84	14.95	16.42	16.00	20.96	20.72	20.30	14.78	15.58
Rb	6.07	37.56	20.93	3.46	3.79	31.30	43.12	4.81	17.32	22.18
Sr	256.62	372.24	210.63	466.00	255.90	751.40	702.20	254.80	124.15	301.09
Y	18.46	20.14	17.84	15.68	28.22	19.85	23.30	26.96	29.32	30.97
Zr	54.62	79.19	64.21	118.60	86.56	203.00	249.60	96.86	116.02	102.63
Nb	4.47	8.88	4.31	3.82	3.01	56.18	71.98	4.42	4.15	3.95
Cs	0.19	1.04	0.67	0.05	1.09	0.49	0.82	0.35	1.74	1.40
Ba	66.63	389.94	167.87	135.82	43.88	596.00	589.40	45.46	87.80	261.90
La	5.98	10.72	10.61	12.14	4.60	25.08	33.58	4.41	4.90	7.12
Ce	13.90	23.80	22.85	23.16	13.47	48.10	65.34	12.29	15.18	18.24
Pr	1.90	3.10	2.88	2.75	2.18	5.89	7.70	1.88	2.49	2.68
Nd	8.27	12.70	11.35	11.42	11.51	25.82	32.98	9.85	11.99	12.39
Sm	2.28	3.16	2.71	2.28	3.48	5.76	7.04	3.01	3.69	3.66
Eu	0.85	1.12	0.97	0.80	1.23	2.02	2.24	1.08	1.32	1.29
Gd	2.87	3.73	3.15	2.27	4.28	5.40	6.42	3.84	4.67	4.80
Tb	0.50	0.63	0.53	0.36	0.74	0.73	0.86	0.65	0.82	0.84
Dy	3.35	4.00	3.36	2.25	4.72	3.78	4.44	4.21	5.31	5.62
Но	0.74	0.81	0.71	0.46	0.95	0.63	0.73	0.83	1.18	1.25
Er	2.15	2.23	1.98	1.42	2.78	1.58	1.85	2.46	3.43	3.70
Tm	0.33	0.33	0.29	0.22	0.39	0.20	0.23	0.35	0.50	0.58
Yb	2.10	2.07	1.82	1.48	2.45	1.14	1.29	2.24	3.14	3.79
Lu	0.32	0.31	0.27	0.23	0.35	0.15	0.17	0.32	0.46	0.57
Hf	1.53	2.08	1.76	2.68	2.72	4.67	5.30	2.33	2.91	2.73
Та	0.25	0.52	0.24	0.19	0.18	2.67	3.46	0.28	0.27	0.23
W	0.17	0.39	0.18	0.85	0.41	1.17	1.88	0.58	0.08	1.17
Pb	4.85	13.57	3.50	1.78	2.26	2.31	3.29	1.35	0.73	25.36
Th	1.01	1.83	2.30	1.75	0.23	3.05	4.31	0.24	0.34	1.02
U	0.32	0.50	0.56	0.48	0.12	0.85	1.10	0.09	0.11	0.31

Table 1. Major and trace elements for massive and foliated eclogites.

<sup>a</sup>Data from Song et al. (2009).

## **Rutile fabrics**

Rutile is volumetrically minor, but geochemically significant in eclogite for understanding high field strength elements geochemistry during subduction zone metamorphism. Rutile belongs to the tetragonal crystal system and to the highest symmetry point group (4/mmm) in which <100> and <010> axes are equivalent. All samples show SL-type fabric in which <001> axes are distributed as a girdle with the maxima close to the lineation, whereas <110> and <100> axes are distributed nearly normal to the foliation (Fig. 7). With the increment of shear strain confirmed by a gradual increase of the *M*-index (M = 0.08 for Q05-1, 0.10 for 2Q22, 0.10 for

08BSY-39, 0.15 for 08BSY-38 and 0.17 for 2Q19), <001> axes becomes more concentrated close to the lineation.

## CHEMICAL VARIATIONS OF ECLOGITIC MINERALS AND $P-T-F_{O_2}$ CONDITIONS

The chemical compositions of minerals are acquired using a JEOL JXA-8100 electron probe micro-analyzer (EPMA) at Peking University. The EPMA was operated at 15 kV acceleration voltage, 10 nA beam current and  $1-5 \,\mu\text{m}$  spot size. Relevant standard minerals (from SPI corp. US) were used for calibration. Representative mineral analyses and calculation details are given in Tables 2–5.



**Fig. 5.** Lattice preferred orientation of garnet from the massive eclogite (Q05-1). E-W is lineation; N-S is foliation normal. (a) Scattered figure. (b) Contoured figure in lower hemisphere. Contours in pole figures and numbers in legend correspond to the multiples of uniform distribution. Pole figure was drawn using the half width of 20° and cluster size of 5°.

### Garnet

Almost all porphyroblastic garnet in low-T eclogite possesses core-to-rim compositional variation (Song et al., 2007; Zhang et al., 2007; Wei et al., 2009). Garnet in the massive eclogite is rounded and euhedral with randomly oriented inclusions of Omp + Lws (or pseudomorphs) + Cz + Ph + Qtz. Garnet in the foliated eclogite, on the other hand, is elongated and has orientated inclusions of Omp + Cz (Fig. 3g,h), suggesting its syntectonic growth during shear deformation. The representative compositions of garnet in both massive and foliated eclogite are given in Table 2. As shown in Fig. 9, garnet from the massive eclogite has higher CaO (or grossular; Grs 24.3–27.5 mol.%) than garnet in the foliated eclogite (Grs 19.7–22.6 mol.%). whereas FeO + MgO decreases accordingly with a similar Mg/Fe range. Similar compositional trends have also been observed between massive and foliated lawsonite eclogites from Guatemala (Tsujimori et al., 2006).

Some garnet in the foliated eclogite exhibits compositional profiles of decreasing CaO from the core (Grs 26.92–24.02 mol.%) to the rim (Grs 20.66–22.49 mol.%) with increasing MgO + FeO (Fig. 9). This feature is opposite to the high CaO bulk composition of the foliated eclogite (Table 1), suggesting that chemical variation of garnet from massive to foliated eclogite is not controlled by whole-rock composition but by metamorphic equilibrium (more proportion of epidote occurs) during shear deformation.

#### Omphacite

The compositions of omphacite in massive and foliated eclogites are much different (Table 3). Figure 10

shows a systematic and distinct compositional variation with increasing shear strain is observed. With increasing intensity of deformation, both  $X_{\rm Fe}$  [Fe/ (Fe + Mg)] and  $Y_{\rm Fe}$  [Fe<sup>3+</sup>/(Fe<sup>3+</sup> + Al<sup>VI</sup>)] in omphacite increase significantly from massive to foliated eclogites (Fig. 10a). In massive eclogite samples (e.g. Q05-1),  $Omp^2$  of the second generation that is elongated along the strongly deformed micro-shear zones contains much higher total FeO content (5.69-6.93 wt%) than that in the weakly deformed  $Omp^1$  grains (FeO 3.36-4.86 wt%) (Fig. 10b). In the foliated eclogite samples (e.g. 2Q22), the fine-grained  $Omp^2$  in the matrix contains higher contents of FeO and aegirine (Ae)-component than coarse-grained Omp<sup>1</sup> that retain in the lens-shaped, weakly strained domains (Fig. 10c). The increasing  $Fe^{3+}$  or Ae component in the Omp<sup>2</sup> implies an increase in oxygen fugacity of the system during shear deformation and retrograde metamorphism.

### Clinozoisite/epidote

Chemical compositions of clinozoisite are given in Table 4 and mainly belong to clinozoisite with  $X_{\text{Fe}}$  [Fe<sup>3+</sup>/(Fe<sup>3+</sup> + Al<sup>VI</sup>)] <0.2. Ferric iron in epidote group minerals was recalculated assuming two-site ordering with a total of six silicon, aluminium and ferric cations per 12.5 oxygen. Similar to omphacite, Fe<sup>3+</sup> or  $X_{\text{Fe}}$  in clinozoisite also increases from massive to foliated eclogite (Fig. 11).

## Sodic amphibole

Sodic amphibole is a ubiquitous hydrous phase and is glaucophane (Gln) in all studied samples. Some glaucophane grains invariably possess compositional



**Fig. 6.** Lattice preferred orientations of omphacite in one massive and four foliated eclogites. Pole figures are presented in the upper hemisphere and equal area projection where E-W direction is lineation and foliation is shown as horizontal line. (a) Weak SL-type fabric of the massive eclogite (Q05-1). (b–e) Strong SL-type fabrics of the four foliated eclogites. *N* is the number of grains, *M* is the *M*-index of LPO. Contours in pole figures and numbers in legend correspond to the multiples of uniform distribution. Pole figure was drawn using the half width of 20° and cluster size of 5°.



**Fig. 7.** Lattice preferred orientation of clinozoisite/epidote for one massive and four foliated eclogites. Pole figures are presented in the upper hemisphere and equal area projection where E-W direction is lineation and foliation is shown as horizontal line. The LPOs of <010 > and <001 > axes in clinozoisite show a remarkable contrast with the LPOs of those axes in omphacite. N is the number of grains, M is the M-index of LPO. Contours in pole figures and numbers in legend correspond to the multiples of uniform distribution. Pole figure was drawn using the half width of 20° and cluster size of 5°.



**Fig. 8.** Lattice preferred orientation of rutile for the one massive and four foliated eclogites. Pole figures are presented in the upper hemisphere and equal area projection where E-W direction is lineation and foliation is shown as horizontal line. SL-type fabric and strengthening of <001> maxima parallel to lineation are clearly exhibited, which agrees with shear strain increases from Q05-1 through 2Q22, 08BSY-39, 08BSY-38 to 2Q19. *N* is the number of grains, *M* is the *M*-index of LPO. Contours in pole figures and numbers in legend correspond to the multiples of uniform distribution. Pole figure was drawn using the half width of 20° and cluster size of 5°.

Table 2.	Representative	composition o	f garnet in	both massive	and foliated eclogite.
----------	----------------	---------------	-------------	--------------	------------------------

Sample				М	assive ecl	ogite								Fo	liated ec	logite				
_		Q05-1			Q05-3			2Q27			2Q22			2Q19			2Q24		Q	L-33
SiO <sub>2</sub>	38.48	39.01	38.51	38.04	39.04	38.73	38.57	38.52	38.34	38.97	38.47	38.10	38.31	38.17	38.14	38.74	38.28	38.90	38.47	39.25
TiO <sub>2</sub>	0.06	0.18	0.09	0.00	0.04	0.00	0.03	0.12	0.12	0.01	0.04	0.01	0.01	0.02	0.05	0.00	0.08	0.02	0.08	0.08
$Al_2O_3$	21.59	21.52	21.70	21.74	21.54	22.19	21.90	21.49	22.00	21.63	21.75	21.54	21.40	21.05	21.72	21.85	21.53	21.45	21.20	21.59
$Cr_2O_3$	0.00	24.80	0.04	0.04	25.68	0.05	26.50	0.03	25.00	25.67	26.51	0.00	0.02	24.54	0.03	26.26	20.00	0.14	0.07	25.21
MnO	23.24	24.80	0.20	23.15	23.08	23.04	20.30	0.51	23.99	23.07	20.31	20.48	20.70	24.34	20.15	0.20	20.00	0.16	0.42	0.25
MgO	5.02	5 4 5	4.81	5.51	5.16	4 61	4 50	3 13	4 15	6.12	4 99	4 50	4 43	6 49	5.14	5 43	3 49	4 85	4 35	5.91
CaO	9.03	8.77	9.72	8.98	9.10	9.29	9.06	9.59	9.16	7.13	7.66	8.18	7.67	7.76	7.81	8.09	7.97	7.72	7.56	7.92
Na <sub>2</sub> O	0.00	0.06	0.04	0.07	0.04	0.02	0.01	0.05	0.04	0.08	0.11	0.03	0.00	0.07	0.03	0.07	0.00	0.00	0.02	0.05
K <sub>2</sub> O	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.03	0.01
Total	99.93	100.33	100.73	99.80	100.93	100.97	100.94	100.70	100.19	100.20	100.10	99.57	99.45	99.06	99.73	100.69	100.69	101.15	99.41	100.27
Calculat	ion using	12 oxyge	m																	
Si	3.00	3.02	2.99	2.97	3.01	2.99	2.99	3.01	2.99	3.02	3.00	3.00	3.02	2.99	2.99	3.00	3.00	3.01	3.03	3.03
Ti	0.00	0.01	0.01	0.00	0.00	0.00	0.00	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Al	1.98	1.96	1.98	2.00	1.96	2.02	2.00	1.98	2.02	1.97	2.00	1.99	1.99	1.95	2.00	1.99	1.99	1.96	1.97	1.96
Cr E-3+	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00
Fe Eo <sup>2+</sup>	0.01	0.00	0.00	0.10	0.02	0.00	0.03	1.78	0.00	0.00	1.72	0.02	0.00	0.11	0.03	0.03	1.80	1.80	0.00	0.00
Mn	0.03	0.04	0.02	0.02	0.02	0.03	0.02	0.03	0.02	0.04	0.03	0.04	0.06	0.06	0.04	0.01	0.03	0.01	0.03	0.02
Mo	0.05	0.63	0.56	0.62	0.52	0.05	0.52	0.05	0.02	0.04	0.05	0.53	0.52	0.00	0.60	0.63	0.05	0.56	0.05	0.62
Ca	0.75	0.73	0.81	0.75	0.75	0.77	0.75	0.80	0.77	0.59	0.64	0.69	0.65	0.65	0.66	0.67	0.67	0.64	0.64	0.66
Na	0.00	0.01	0.01	0.01	0.01	0.00	0.00	0.01	0.01	0.01	0.02	0.00	0.00	0.01	0.00	0.01	0.00	0.00	0.00	0.01
К	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Sum	8.00	7.99	8.02	8.03	8.01	8.00	8.01	7.99	7.99	8.00	8.00	8.01	7.99	8.04	8.01	8.01	8.00	8.00	7.98	7.99
	Ga	rnet profi	le of folia	ted eclog	gite 2Q22	from cor	e (left) to	rim (righ	it)											
SiO <sub>2</sub>	37	7.73	38.07	3	7.87	37.98	3'	7.89	37.84	37	.94	38.06		38.05	37	.92	38.10	38.	37	38.94
TiO <sub>2</sub>	C	0.10	0.17		0.15	0.18	(	0.14	0.09	0	0.10	0.13		0.10	0	.10	0.00	0.	06	0.00
$Al_2O_3$	21	.19	21.11	2	1.09	21.45	2	1.26	21.60	21	.06	21.26		21.22	21	.18	21.71	21.	82	22.11
$Cr_2O_3$	C	0.03	0.06		0.01	0.06	(	0.04	0.05	0	0.05	0.06		0.00	0	.08	0.01	0.	01	0.02
FeO	27	7.73	28.32	2	8.69	28.32	2	8.65	28.99	28	3.88	29.53		29.49	29	.10	27.84	27.	53	25.17
MnO	1	.76	1.65		1.43	1.41		1.01	0.75	0	0.61	0.50		0.41	0	.49	0.51	0.	28	0.10
MgO	2	2.06	2.34		2.47	2.17	-	2.30	2.22	2	2.69	2.85		3.17	3	.37	4.13	4.	67	6.39
CaO	9	0.38	9.13		8.54	9.30		9.01	8.61	ý	0.02	8.04		7.94	7	.84	7.85	7.	95	7.48
Na <sub>2</sub> O		0.05	0.14		0.04	0.07		0.01	0.04		0.00	0.07		0.05	0	.04	0.00	0.	03	0.02
R <sub>2</sub> O Total	100	).01 ).04	101.05	10	0.00	100.96	10	0.33	100.2	100	).43	100.49	1	0.00	100	.15	100.15	100.	75	100.25
Calculat	ion using	12 oxyge	'n																	
Si	3	3.00	2.99		3.00	2.99		3.00	3.00	2	2.99	3.00		3.00	2	.99	2.99	2.	98	3.00
Ti	0	0.01	0.01		0.01	0.01	(	0.01	0.01	0	0.01	0.01		0.01	0	.01	0.00	0.	00	0.00
Al	1	.98	1.95		1.97	1.99		1.98	2.02	1	.96	1.98		1.97	1	.97	2.01	2.	00	2.01
Cr	C	0.00	0.00		0.00	0.00	(	0.00	0.00	0	0.00	0.00		0.00	0	.00	0.00	0.	00	0.00
Fe <sup>3+</sup>	C	0.02	0.07		0.02	0.02	(	0.00	0.00	0	0.05	0.01		0.02	0	.03	0.01	0.	04	0.00
Fe <sup>2+</sup>	1	.83	1.80		1.89	1.84		1.90	1.92	1	.86	1.94		1.92	1	.90	1.81	1.	75	1.62
Mn	0	0.12	0.11		0.10	0.09	(	0.07	0.05	0	0.04	0.03		0.03	0	.03	0.03	0.	02	0.01
Mg	0	0.24	0.27		0.29	0.25	(	0.27	0.26	0	0.32	0.34		0.37	0	.40	0.48	0.	54	0.73
Ca	0	0.80	0.77		0.72	0.78	(	0.76	0.73	0	0.76	0.68		0.67	0	.06	0.66	0.	00	0.62
ina V	0	0.01	0.02		0.01	0.01	(	0.00	0.01	(	101	0.01		0.01	0	.01	0.00	0.	00	0.00
K. Sum		2.00	0.01		0.00 8.00	0.00		0.00 8.00	0.00		2.00	0.00		0.00	0	.00	0.00	0.	00	0.00
Sun	8	.00	0.00		0.00	8.00		0.00	8.00	5	.00	8.00		0.00	8	.00	0.00	8.	00	8.00

Ferric iron was determined by stoichiometry following Droop (1987).

zoning with increased  $X_{\text{Fe}}$  [Fe<sup>2+</sup>/(Fe<sup>2+</sup> + Mg)] and  $Y_{\text{Fe}}$  [Fe<sup>3+</sup>/(Fe<sup>3+</sup> + Al<sup>VI</sup>)] towards the rim (Song *et al.*, 2007). In addition, both Fe<sup>2+</sup> and Fe<sup>3+</sup> in glaucophane increase slightly from massive to foliated eclogite with increasing shear strain (Table 5).

## Phengite

Phengite is the only potassium-rich mineral in eclogite and can be stable over a wide range of P-T conditions, partly because it is not involved in most of the metamorphic reactions except for Si increase with pressure. The Si content of phengite in our studied samples (both massive and foliated) shows no statistically significant difference (e.g.  $3.41 \pm 0.03$  within analytical error), suggesting the possibility that the chemical composition of phengite may have not been re-equilibrated during retrograde metamorphism and shear deformation (Table 5).

## **P**–**T** estimates

Imperfect equilibrium relationships among coexisting minerals in a given rock sample will inevitably result in large uncertainties in estimated 'equilibrium' P-T conditions. Nevertheless, the Grt-Omp-Ph (-Ky)

																											1
Sample				1	Massive ε	sclogite												Foliat	ed eclogit	0							
	Q.	15-1 Omp		Q0	5-1 Omp	5	Q05	-3	2Q2	4	200	22 Omp <sup>1</sup>		2Q2	2 Omp <sup>2</sup>		5	219		QL-33		08BSY-3	0	8BSY-38		2Q24	
SiO <sub>2</sub>	55.68	56.42	56.25	55.35	55.96	55.60	56.72	56.77	56.36	56.20	56.99	56.03	56.16	55.86 5	5.95 5	5.71 5.	5.72 5	4.95 55	.62 55	.77 55	.96 55	.67 55	27 55.	90 55.	33 56.2	53 57.	48
$TiO_2$	0.05	0.05	0.10	0.11	0.10	0.08	0.00	0.02	0.12	0.05	0.07	0.05	0.12	0.03	0.14	0.04	0.04	0.01 (	.03 0	.07 0	.06 0	.05 0	.04 0.	02 0.	0.0	0.0	00.
$Al_2O_3$	10.14	10.08	10.32	9.55	10.89	9.87	10.76	10.62	11.15	8.72	11.37	10.25	10.34	9.99	9.95	9.83 10	0.78 1	0.93 5	.83 8	.01 10	9 00.	.71 9	.78 9.	13 9.	57 10.0	9 11.9	.90
$Cr_2O_3$	0.00	0.02	0.00	0.06	0.00	0.01	0.00	0.03	0.04	0.01	0.00	0.06	0.03	0.08	0.04	0.03	0.05	00.00	.06 0	00.00	.04 0	.01 0	.04 0.	02 0.	0.0	.0 0.1	01
FeO	3.85	4.33	3.92	6.55	6.48	6.57	3.29	3.08	3.88	5.04	5.05	4.75	4.93	7.21	6.41	6.90	6.26	9.06 6	.52 6	.82 5	29 7	.15 7	01 7.	00 6.	54 8.2	21 5.2	.23
MnO	0.00	0.00	0.00	0.04	0.00	0.01	0.05	0.04	0.02	0.09	0.06	0.10	0.02	0.05	0.00	0.04	0.03	00.00	.03 0	.02 0	.03 0	00.	.00	00 00.	0.0	.0	.03
MgO	9.04	8.59	8.58	7.48	6.66	7.54	8.83	8.44	8.09	14.37	7.34	8.34	7.89	6.76	7.54	7.24	7.26	5.26 7	8 11. 8	69.	.80	.23 7	32 7.	92 7.	54 6.2	23 7.	24
CaO	13.13	13.00	13.36	13.29	11.03	12.95	13.07	12.91	12.74	0.00	11.01	11.99	11.89	10.41 1	1.43 1	0.64 1	1.24	0.78 11	.38 13	.53 12	.50 10	11 26.	.18 11.	68 11.	39 10.0	5 10.9	96
$Na_2O$	7.56	7.10	7.44	7.63	8.20	7.17	7.04	7.13	7.69	60.9	7.74	7.80	7.92	9.04	8.19	8.91	8.12	3.68 8	59 7	.14 8	.06 8	.53 8	.43 8.	22 8.	53 8.	7. 7.	.98
$K_2O$	0.01	0.00	0.02	0.00	0.00	0.02	0.02	0.00	0.00	0.00	0.02	0.02	0.02	0.00	0.01	0.00	0.04	0.02 (	.01 0	00.	00.00	00.	.01 0.	01 0.	0.0	0.0	.02
Total	99.52	99.62	100.00	100.06	99.32	06.66	99.78	99.04	100.09	79.62	99.73	99.47	99.33 9	99.48 9	9.71 9	9.34 9	9.66 9	56 69.6	.84 100	.07 99	96 LL:	.37 99	.19 99.	94 99.	73 100.0	51 100.3	.85
Calculation	nsing 6 (	μοσιχυ																									
Si	1.99	2.01	2.00	2.00	2.02	2.00	2.01	2.02	1.99	2.02	2.03	2.01	2.02	2.02	2.02	2.02	2.01	2.00	.01 2	.02	.01	.02	01 2.	02 2.	02 2.0	02 2.0	.02
Al <sup>IV</sup>	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	00.00	00.	00.	00.00.	.00	00 00.	0.0	0.0	00.
$Al^{VI}$	0.42	0.42	0.44	0.41	0.46	0.42	0.45	0.45	0.45	0.37	0.48	0.43	0.44	0.43	0.42	0.42	0.46	.47 (	.42 0	.34 0	.42 0	.42 0	.42 0.	39 0.	41 0.4	45 0. <sup>,</sup>	49
Ti	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	00.00	00.00	00.00	00.00	00.	.00 00.	00 00.	0.0	0.0	00.
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	00.00	00.00	00.00	00.00	00.	.00 00.	00 00.	0.0	0.0	00.
$Fe^{3+}$	0.13	0.06	0.09	0.16	0.11	0.09	0.03	0.02	0.08	0.01	0.03	0.12	0.12	0.22	0.16	0.22	0.13	0.18 0	.21 0	.17 0	.15 0	.20 0	20 0.	20 0.	21 0.	15 0.1	.03
$Fe^{2+}$	0.00	0.07	0.02	0.04	0.09	0.10	0.07	0.07	0.03	0.14	0.12	0.02	0.03	0.00	0.03	0.00	0.06	0.10 (	00.	.04	00.	.02 0	.01 0.	01 0.	0.0	.0 0.	.12
Mn	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	00.0	00.00	00.	00.	00.	00.	.00	00 00.	0.0	0.0	00.
Mg	0.48	0.46	0.46	0.40	0.36	0.41	0.47	0.45	0.43	0.77	0.39	0.45	0.42	0.37	0.40	0.39	0.39	).34 (	.42 0	.47 0	.42 0	.39 0	.40 0.	43 0.	41 0.3	33 0.	.38
Ca	0.50	0.50	0.51	0.51	0.43	0.50	0.50	0.49	0.48	0.00	0.42	0.46	0.46	0.40	0.44	0.41	0.43	).38 (	.44	.52 0	.48 0	.43 0	.0	45 0.	14	9 0.	.41
Na	0.52	0.49	0.51	0.53	0.57	0.50	0.48	0.49	0.53	0.42	0.53	0.54	0.55	0.63	0.57	0.63	0.57	).61 (	0 09.	.50 0	.56 0	.60 0	.59 0.	58 0.	50 0.0	51 0.	54
К	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	00.0	00.00	00.00	00.00	00.00	00.00	.00 00.	00 00.	0.0	0.0	00.
$X_{\mathrm{Fe}}$	0.21	0.22	0.19	0.33	0.36	0.32	0.18	0.17	0.20	0.16	0.28	0.24	0.26	0.37	0.32	0.36	0.33	.45 (	.33 0	.31 0	.26 0	.36 0	34 0.	33 0.	34 0.4	12 0.7	.28
$Y_{\rm Fe}$	0.24	0.13	0.17	0.28	0.19	0.18	0.06	0.04	0.15	0.03	0.06	0.22	0.21	0.34	0.28	0.34	0.22	).28 (	.33 0	.33 0	.26 0	.32 0	.32 0.	34 0.	34 0.2	25 0.1	.06
Cations	4.05	4.02	4.04	4.06	4.04	4.03	4.01	4.00	4.00	4.00	4.00	4.04	4.04	4.08	4.06	4.08	4.05	1.07	.08	.06	.06	.07 4	.08	07 4.	9.4.(	)5 4.1	01
Ferric iron	was dete	ermined t	by stoichi	ometry fc	llowing i	Droop (1	987).																				I I

Table 3. Representative composition of omphacite in both massive and foliated eclogite.

Table 4. Re	epresentative co	omposition	of clino	zoisite/e	pidote ir	ı both	massive and	l foliated	eclogite.
-------------	------------------	------------	----------	-----------	-----------	--------	-------------	------------	-----------

Sample		М	assive eclog	gite							Foliated	l eclogite					
		Q05-1		20	227		2Q22			2Q19		08BS	SY-38	08B	SY39	QI	_33
SiO <sub>2</sub>	38.39	38.52	38.83	38.11	38.09	38.14	38.20	37.87	37.93	38.09	38.13	38.32	37.85	38.26	38.31	38.16	38.25
TiO <sub>2</sub>	0.09	0.09	0.13	0.12	0.19	0.09	0.06	0.06	0.00	0.11	0.05	0.07	0.06	0.03	0.09	0.11	0.16
$Al_2O_3$	28.65	28.84	28.66	29.55	28.68	25.80	26.25	25.37	24.54	25.95	26.19	27.40	27.23	25.69	25.97	27.57	26.18
Cr <sub>2</sub> O <sub>3</sub>	0.01	0.03	0.03	0.07	0.03	0.05	0.02	0.03	0.06	0.08	0.06	0.09	0.00	0.03	0.03	0.03	0.00
FeO	6.24	5.39	5.69	5.23	5.59	9.76	8.65	10.05	10.78	9.49	8.97	7.45	7.63	9.74	8.90	7.22	8.92
MnO	0.00	0.00	0.00	0.04	0.17	0.09	0.07	0.07	0.15	0.12	0.04	0.09	0.06	0.01	0.03	0.00	0.19
MgO	0.13	0.08	0.15	0.24	0.07	0.04	0.06	0.14	0.00	0.08	0.11	0.15	0.11	0.05	0.03	0.15	0.02
CaO	23.58	24.00	23.74	23.43	24.01	23.13	23.35	23.18	23.24	23.25	23.40	23.62	23.53	23.46	23.63	23.47	23.37
Na <sub>2</sub> O	0.00	0.02	0.07	0.03	0.01	0.01	0.04	0.01	0.00	0.02	0.02	0.04	0.03	0.03	0.01	0.00	0.02
$K_2O$	0.00	0.00	0.01	0.00	0.00	0.02	0.00	0.00	0.00	0.03	0.03	0.01	0.01	0.02	0.00	0.00	0.01
Total	97.09	96.97	97.31	96.82	96.84	97.13	96.70	96.78	96.70	97.22	97.00	97.24	96.51	97.32	97.00	96.71	97.12
Calculatio	n using 12.	5 oxygen															
Si	2.98	2.99	3.01	2.96	2.97	2.99	3.00	2.99	3.00	2.99	2.99	2.99	2.98	3.00	3.01	2.99	3.00
Al <sup>IV</sup>	0.02	0.01	0.00	0.04	0.03	0.01	0.00	0.01	0.00	0.01	0.01	0.01	0.02	0.00	0.00	0.01	0.00
Al <sup>VI</sup>	2.61	2.63	2.61	2.67	2.61	2.38	2.43	2.35	2.29	2.38	2.41	2.50	2.50	2.37	2.40	2.53	2.42
Ti	0.01	0.01	0.01	0.01	0.01	0.01	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.01	0.01	0.01
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.01	0.00	0.00	0.00	0.00	0.00
Fe <sup>3+</sup>	0.40	0.35	0.37	0.34	0.36	0.64	0.57	0.66	0.71	0.62	0.59	0.49	0.50	0.64	0.58	0.47	0.58
Mn	0.00	0.00	0.00	0.00	0.01	0.01	0.01	0.01	0.01	0.01	0.00	0.01	0.00	0.00	0.00	0.00	0.01
Mg	0.02	0.01	0.02	0.03	0.01	0.01	0.01	0.02	0.00	0.01	0.01	0.02	0.01	0.01	0.00	0.02	0.00
Ca	1.96	2.00	1.97	1.95	2.01	1.95	1.97	1.96	1.97	1.95	1.97	1.97	1.98	1.97	1.99	1.97	1.96
Na	0.00	0.00	0.01	0.01	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.01	0.01	0.01	0.00	0.00	0.00
K	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Cations	7.99	8.00	8.00	8.01	8.01	7.99	7.99	7.99	7.99	7.99	8.00	8.01	8.01	8.00	7.99	7.99	7.99
$X_{\rm Fe}$	0.13	0.12	0.12	0.11	0.12	0.21	0.19	0.22	0.24	0.21	0.20	0.16	0.17	0.21	0.19	0.16	0.19

Total Fe as Fe<sup>3+</sup>.

geothermobarometry of Ravna & Terry (2004) is a efficient method to compare calculated P-T values among samples (Table 6). Interestingly, the massive eclogites yield 23–26 kbar and 485–510 °C, which is within the lawsonite stability field. On the other hand, the foliated eclogites yield 18–22 kbar and 450–480 °C, which is well within the epidote-eclogite field. These differences are statistically significant and are consistent with the petrography showing an apparent P-T decrease with increasing plastic deformation from massive to foliated eclogites.

### **Oxygen fugacity**

Figure 12 suggests that, to a first order, there must be a systematic increase in oxygen fugacity ( $f_{O_2}$ ) in response to the plastic deformation as manifested by the texture and mineral compositional change from massive to foliated eclogites. This feature is also observed in other HP terranes, for example, zoned omphacite, glaucophane, and/or epidote in eclogite from Turkey (Davis & Whitney, 2006). We thus estimated  $f_{O_2}$  by considering the garnet-epidote equilibrium reaction:

$$8Ca_{3}Al_{2}Si_{3}O_{12} + 4Fe_{3}^{2+}Al_{2}Si_{3}O_{12} + 6H_{2}O + 3O_{2}$$
  
grossular almandine fluid  
$$=\frac{12Ca_{2}Fe^{3+}Al_{2}Si_{3}O_{12}(OH)}{epidote(X_{Fe} = 0.33)}$$

using THERMOCALC v.3.33 with updated database (file tcds55.txt) of Holland & Powell (1998), and assuming unit  $H_2O$  activity. The above reaction is a valid

description that is consistent with the production of epidote at the expense of garnet (Figs 2 & 3) with an associated Ca/Fe (precisely Ca/[Fe+Mg]) reduction in garnet (Fig. 9). The calculation procedure is the same as described by Donohue & Essene (2000). These latest updated solid-solution models, database and program, as well as the extremely low  $Fe^{3+}$  content in garnet allow this oxygen barometer to be potentially applicable to our samples. It is, however, not straightforward to obtain precise  $f_{O_2}$  values for the two types of the eclogites and to compare  $f_{O_2}$  values at different pressures because of temperature and pressure uncertainties inherited from thermobarometric estimates (Table 6), on which the estimated  $f_{O_2}$  values depend. Nevertheless, the differences in the calculated  $f_{\rm O}$ , values between the two eclogite types are significant. Figure 12 shows the curves for both massive and foliated eclogites in  $\log f_{O_2} - T$  space at the temperature range of 400-600 °C for two fixed pressures of 24 kbar and 19 kbar, which correspond to the calculated pressures for massive and foliated eclogites (Table 6). The foliated eclogites are obviously more oxidized than the massive eclogites by  $\sim 2 \log f_{O_2}$  units for the two pressures in each diagram, and both are 2–4 log  $f_{O_2}$ units above the fayalite-magnetite-quartz oxygen buffer.

This highly oxidized condition in eclogite was also reported in Qinglongshan of Sulu UHPM terrane (Mattinson *et al.*, 2004). The reaction of almandine and lawsonite (Fig. 12) occurs in higher  $f_{O_2}$  condition than that of pure almandine (Fig. 12a,b). This difference is remarkable at low temperature and decreased

	М	assive ec	logite (G	ln)		Fo	pliated ec	logite (G	ln)		М	assive ec	logite (Ph	in)		F	oliated ec	ogite (Ph	in)	
Sample	Q0	5-1	20	227	20	222	2Q	219	QI	_33	Q	5-1	2Q	227	20	222	20	219	QL	33
SiO <sub>2</sub>	58.00	58.08	57.85	57.47	56.98	57.18	57.26	57.76	57.49	57.95	50.67	51.34	52.48	49.86	51.36	50.69	51.21	51.89	50.27	50.72
$TiO_2$	0.01	0.05	0.00	0.00	0.05	0.00	0.07	0.00	0.04	0.07	0.26	0.22	0.30	0.10	0.24	0.22	0.26	0.21	0.20	0.12
$Al_2O_3$	11.36	11.61	11.44	11.58	11.10	10.88	10.66	11.09	11.25	11.42	27.38	26.79	28.35	28.12	26.12	25.53	26.89	27.72	26.86	27.73
$Cr_2O_3$	0.06	0.03	0.05	0.15	0.01	0.08	0.04	0.00	0.06	0.05	0.02	0.02	0.04	0.17	0.01	0.14	0.00	0.02	0.06	0.10
FeO	6.11	5.95	6.11	5.94	9.76	8.82	9.74	8.81	8.40	6.81	1.43	1.52	1.21	1.29	3.17	2.57	1.68	2.55	2.58	1.46
MnO	0.07	0.11	0.06	0.05	0.00	0.09	0.03	0.02	0.02	0.05	0.00	0.04	0.04	0.08	0.00	0.04	0.03	0.04	0.05	0.01
MgO	13.12	13.12	13.02	13.33	10.84	11.91	12.25	11.90	11.61	12.81	3.52	3.60	3.63	3.22	3.11	3.39	3.42	3.33	3.33	3.52
CaO	1.35	1.04	1.32	1.35	0.74	1.41	0.80	0.84	1.10	1.40	0.02	0.02	0.01	0.01	0.01	0.01	0.00	0.02	0.00	0.00
Na <sub>2</sub> O	7.49	7.27	7.02	6.94	7.51	6.96	7.20	7.11	7.12	7.16	0.80	0.71	0.94	0.58	0.54	0.79	0.87	0.79	0.49	0.86
$K_2O$	0.05	0.02	0.02	0.03	0.03	0.03	0.02	0.02	0.03	0.03	10.37	10.47	9.77	10.24	10.34	10.33	10.19	9.91	10.73	10.17
Total	97.62	97.28	96.89	96.84	97.02	97.36	98.07	97.55	97.12	97.75	94.47	94.73	96.75	93.67	94.90	93.71	94.55	96.48	94.57	94.69
0	23	23	23	23	23	23	23	23	23	23	11	11	11	11	11	11	11	11	11	11
Si	7.84	7.83	7.84	7.77	7.85	7.80	7.73	7.83	7.85	7.82	3.40	3.43	3.41	3.37	3.45	3.45	3.43	3.41	3.40	3.39
Al	1.81	1.84	1.83	1.85	1.81	1.75	1.69	1.77	1.81	1.82	2.16	2.17	2.17	2.24	2.07	2.05	2.12	2.15	2.14	2.18
Ti	0.00	0.01	0.00	0.00	0.01	0.00	0.01	0.00	0.00	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Cr	0.01	0.00	0.01	0.02	0.00	0.01	0.00	0.00	0.01	0.01	0.00	0.00	0.00	0.01	0.00	0.01	0.00	0.00	0.00	0.01
Fe <sup>3+</sup>	0.14	0.28	0.26	0.38	0.27	0.38	0.71	0.46	0.26	0.24										
Fe <sup>2+</sup>	0.55	0.39	0.43	0.30	0.85	0.63	0.39	0.54	0.70	0.53	0.08	0.09	0.07	0.07	0.18	0.15	0.09	0.14	0.15	0.08
Mn	0.01	0.01	0.01	0.01	0.00	0.01	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Mg	2.64	2.64	2.63	2.69	2.23	2.42	2.46	2.40	2.36	2.58	0.35	0.36	0.35	0.32	0.31	0.34	0.34	0.33	0.34	0.35
Ca	0.20	0.15	0.19	0.20	0.11	0.21	0.12	0.12	0.16	0.20	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na	1.96	1.90	1.81	1.80	2.01	1.84	1.88	1.87	1.89	1.87	0.10	0.09	0.12	0.08	0.07	0.10	0.11	0.10	0.06	0.11
K	0.01	0.00	0.00	0.01	0.01	0.01	0.00	0.00	0.01	0.01	0.89	0.89	0.81	0.88	0.89	0.90	0.87	0.83	0.92	0.87
Cations	15.17	15.05	15.04	15.02	15.12	15.05	15.00	14.99	15.05	15.08	7.00	6.99	6.95	6.98	6.98	7.01	6.99	6.97	7.02	7.00

Table 5. Representative composition of glaucophane (Gln) and phengite (Phn) in both massive and foliated eclogite.

Fe<sup>3+</sup> in Gln is maximum ferric recalculated on the basis of 13 total tetrahedraly and octahedraly co-ordinated cations and 46 charges (Robinson et al., 1982).



**Fig. 9.** Diagram showing rim compositions of garnet between massive and foliated eclogite. A compositional profile of garnet from the foliated eclogite (2Q22) is also presented.

to higher temperature. In addition, the  $f_{O_2}$  condition for the reaction of almandine, kyanite and diopside approximates to that of pure almandine. Both reactions show higher oxidized conditions for the foliated eclogite than for the massive eclogite. However, lawsonite may be largely decomposed to form clinozoisite, hence, the oxidation reaction of almandine and lawsonite may be limited in a lesser extent.

## DISCUSSION AND CONCLUSIONS

#### Deformation mechanisms of minerals in Low-T eclogite

The deformation mechanism of low-T eclogite is expected to be much more complex than that of medium- to high-T eclogites due to the presence of hydrous phases and its multiphase assemblages. As the low temperature at eclogite facies conditions could also severely restrict the dislocation activity (Godard & van Roermund, 1995), other deformation mechanisms such as fluid-facilitated DMT process may compete with dislocation creep.

Both field observations and experimental studies have shown that fluids play an essential role in the deformation of rocks and minerals in both brittle and ductile regimes (Green & Jung, 2005). For example, in the case of brittle deformation of rocks in deep subduction zones, fluids from the dehydration of serpentine can serve as a medium to reduce the effective stress by elevating the pore-pressure and triggering faulting (i.e. earthquakes) even at high pressures up to 6 GPa (Jung et al., 2004; Jung, 2009), via dehydration embrittlement (Jung & Green, 2004). In the case of ductile deformation, the dissolved 'fluids', especially water, can reduce the flow strength of minerals significantly through water- or hydrolytic-weakening, and thus facilitating plastic deformation (Green & Jung, 2005 and references therein). Water is also known to enhance dynamic recrystallization of olivine (Jung & Karato, 2001) and significantly influence LPO of olivine and orthopyroxene (Jung et al., 2006, 2010; Jung, 2009).



**Fig. 10.** Variation of omphacite chemical compositions among eclogites. Both Fe-Mg exchange and  $\text{Fe}^{3+}$ -Al substitution are expressed in this diagram (Fe<sup>3+</sup> = Na-Al<sup>VI</sup>). The arrow shows a trend of increasing  $X_{\text{Fe}}$  and  $Y_{\text{Fe}}$  along with increasing shear strain in eclogites.



**Fig. 11.** Histogram showing the increase of  $X_{\text{Fe}}$  of clinozoisite from massive to foliated eclogites. All Fe is Fe<sup>3+</sup> and Fe<sup>3+</sup>-Al substitution is considered.

In our samples, the presence of large amounts of hydroxyl-bearing minerals like clinozosite and glaucophane indicates that these low-*T* eclogite samples are rather 'wet', containing at least 3 wt% of water as inferred from lawsonite inclusions (Song *et al.*, 2007; Zhang *et al.*, 2007; Wei *et al.*, 2009).

#### Omphacite deformation mechanisms

Glide- or recovery-controlled dislocation creep with Burgers vector (100) < 001 >,  $\{110\} < 001 >$  and  $\{110\}1/2 < 110 >$  is considered to be the predominant deformation mechanism in omphacite (Godard & van

**Table 6.** P-T estimates of eclogites calculated using the Grt-Omp-Phn geothermobarometry of Ravna & Terry (2004).

Sample	Rock type	T (R-2000) F Al-0	$e^{3+} = Na-$ Cr
		P (kbar)	<i>T</i> (°C)
Q05-1	Massive eclogite	23.0	480
Q05-3	Massive eclogite	25.6	497
2Q27 <sup>a</sup>	Massive eclogite	25.9	508
Q01	Massive eclogite	24.5	498
2Q22-1	Coarse relics in foliated eclogite	23.4	454
2Q22-4	Foliated eclogite	19.4	476
2Q19-1	Foliated eclogite	19.9	431
2Q19-2		19.8	493
2Q19-3		17.3	476
2Q24	Foliated eclogite	20.2	443
Q5-45	Foliated eclogite	18.4	453

<sup>a</sup>Data from Song et al. (2007).

Roermund, 1995; Zhang *et al.*, 2006). This mechanism can produce many micro-structures of intra-crystalline plasticity including optically observed deformation twins, undulatory extinction, deformation bands, kink bands, subgrains, and dynamic recrystallization, as well as dislocation substructures consisting of free dislocations, loops, networks, dislocation walls recognized in TEM images (Buatier *et al.*, 1991; Philippot & Van Roermund, 1992; Godard & van Roermund, 1995). Some of these features were observed in our massive eclogite samples, including undulatory extinction and dynamic recrystallization (Fig. 2c,d). These observations and the LPOs of omphacite



**Fig. 12.** Log  $f_{O_2}-T$  diagrams calculated assuming (a) 24 kbar and (b) 19 kbar conditions. The solid and dashed lines show the massive and foliated eclogite, respectively, which stand for the reaction  $Ep = Grs + Alm + H_2O + O_2$  with their specific mineral activities. The dotted-dashed lines represent the hematite–magnetite (HM) oxygen buffer and fayalite–magnetite–quartz (FMQ) oxygen buffer as labelled in diagrams.

(Fig. 6a) suggest that dislocation creep is expected to be a dominant deformation mechanism of omphacite in the massive eclogite. However, the foliated eclogites showed elongated and orientated crystals of omphacite, glaucophane and epidote, and these minerals commonly possess chemical zoning and lack intracrystalline dislocation characters such as undulatory extinctions, kink bands and subgrains. The straight foliation parallel grain boundaries are firmly indicative of dissolution and removal of material along the grain contacts (Fig. 3i, also see Blenkinsop, 2000). The increased amount of hydrous minerals (e.g. epidote and glaucophane) from massive to foliated eclogite suggests that water is pervasively participated in the deformation processes. Therefore, the SL-type LPOs are likely developed by fluid-assisted DMT creep with dissolution and concomitant anisotropic growth (see Godard & van Roermund, 1995; Mauler et al., 2001).

Nonetheless, it remains unclear whether this mechanism can effectively produce the LPOs as dislocation creep does, as the latter is generally regarded as the most likely mechanism to produce the LPO. In addition, as dislocation microstructures can be much more readily be changed or eradicated than LPO fabrics with as little as 1% strain (Durham et al., 1977), the rare dislocation creep features in foliated eclogites could be due to dislocation recovery during the final stage of deformation. We thus cannot rule out the possibility of dislocation creep as another important deformation mechanism of omphacite in the foliated eclogites. Indeed, water can enhance both dislocation creep by hydrolytic weakening and diffusion creep by grain boundary fluid film to transport dissolved materials. These two processes are likely to take place in parallel. At low-temperature conditions (e.g. in our case), DMT creep seems to be more favoured because

it requires lower stress compared to temperaturedependent dislocation creep. Also, the presence of abundant fluids can facilitate the DMT and mineral growth to induce a higher strain rate.

On the other hand, it needs confirmation whether deformation regimes can indeed be inferred from the observed LPOs produced by DMT creep. Bascou *et al.* (2002) used an anisotropic viscoplastic self-consistent model to simulate the development of various omphacite LPOs controlled by dislocation glide in response to the change in deformation regime. Kurz *et al.* (2004) and Zhang *et al.* (2006) have further confirmed that dislocation creep is the predominant deformation mechanism. Nevertheless, there seems to be no difference between fabric types produced by dislocation creep and DMT creep under the same deformation geometry, if the type of LPO is essentially controlled by deformation and finite strain geometry (Mauler *et al.*, 2001).

Moreover, the observed microstructures suggest the possible change of deformation mechanism from dislocation creep (dynamic recrystallization) in massive eclogites to DMT creep (syntectonic growth along the foliation) in foliated eclogites during exhumation. Dynamic recrystallization has decreased the grain size of omphacite, and further facilitates the later grainsize-dependent DMT creep through the effect of water most probably released from lawsonite breakdown during decompression-dominated retrograde metamorphism. As shear progresses, the syntectonic anisotropic growth can further enhance grain aspect ratios (Fig. 3c–f).

# Clinozoisite/epidote deformation mechanism

As an analogue to omphacite in monoclinic crystal symmetry, clinozoisite/epidote bears edge-sharing  $[AlO_6]$  octahedral chains along <010> axis bridged by single tetrahedral  $[SiO_4]$  and double tetrahedral  $[Si_2O_7]$ in the <001> axis. These structural properties imply that the octahedral chains in <010> direction and single and double tetrahedra in <001> direction of clinozoisite/epidote would correspond to those tetrahedral chains in <001> direction and octahedra in <010> direction of omphacite, respectively. Therefore, we can interpret the apparently reverse LPOs of <010> and <001> axis between omphacite and clinozoisite (Figs 6 & 7) as resulting from their opposite crystallographic axis definitions. In both massive and foliated eclogites, clinozoisite displays features of chemical zoning and lack of intra-crystal plasticity, which also implies that the fluid-assisted DMT creep may enhance the deformation and anisotropic growth of water-bearing clinozoisite grains with increasing shear strains (Fig. 3c-f). In addition, for the same reasons as for omphacite, we cannot totally rule out dislocation creep as another important deformation mechanism for clinozoisite, and further work is needed.

### Rutile deformation mechanism

A series of experimental results (Blanchin *et al.*, 1990) shows that the dominant slip systems such as  $<101>\{101\}$  and  $<001>\{110\}$  in rutile become important above  $\sim600$  °C. It is nonetheless uncertain whether these slip systems operate in low-*T* eclogites. However, for the similar reasons discussed above, we prefer DMT creep as probably an important deformation mechanism for rutile in equilibrium with other minerals.

# Variation of mineral composition and oxygen fugacity during shear deformation

In general, prograde metamorphism is accompanied by a series of dehydration reactions that release volatiles (e.g. H<sub>2</sub>O and CO<sub>2</sub>), whereas retrograde metamorphism involves hydration. The weakly deformed massive eclogite may represent the weak transpression region, probably owing to heterogeneous strain during shear deformation. In some samples, two generations of mineral assemblages can be distinguished. The first generation consists of  $\operatorname{Grt}^1$  (up to 30 vol.%) +  $\operatorname{Omp}^1$ with minor Ph + Rt + Lws (inclusions in garnet) showing a coarse-grained granoblastic texture. Clinozoisite is rare ( $< 2 \mod .\%$ ) and has relatively low FeO, which is believed to be transformed from lawsonite by decompression (Song *et al.*, 2007). The P-T and oxygen fugacity calculation yields relatively higher pressure (lawsonite-eclogite facies) and lower oxidizing condition than for the foliated eclogites (Fig. 12). The second generation of high-Fe  $Cz + Omp^2$  occurs when micro-shear zones/bands locally overprint the first generation during retrograde process, while the deformation of the massive eclogite eventually produces weak SL-type fabrics of omphacite (with overprinting of  $Omp^1$  and  $Omp^2$ ), clinozoisite and rutile.

The foliated eclogites, on the contrary, exhibit much stronger deformation by microstructures and LPO fabrics. The progressive increment of shear strain is apparent petrographically and is also obvious from both SPOs and LPOs of omphacite, clinozoisite and rutile with SL-type fabrics. Besides, mineral assemblages, mineral compositions and oxygen fugacity of the foliated eclogites also differ from the massive eclogites. The decrease of modal omphacite and garnet is apparently compensated by the increase of hydrous phases such as clinozoisite and glaucophane through hydration processes during shear deformation (Fig. 3a,b). The syn-tectonic growth of garnet (Fig. 3g,h) reveals its equilibrium with the fine-grained matrix through decreasing CaO and increasing total FeO. P-T estimation (1.7-2.2 GPa) and redox calculations (more oxidized condition than massive eclogite) thus reflect retrograde epidote-eclogite facies conditions. The higher total FeO and  $Fe^{3+}$  in  $Omp^2$  and clinozoisite from the foliated eclogites suggests that (i) the retrograde metamorphism was effectively promoted by plastic deformation and (ii) the apparent oxygen fugacity increase was likely associated with fluid participation during exhumation.

### Implications for exhumation through subduction channels

As documented by previous studies, deformation of minerals in eclogite could happen either during burial or exhumation processes. The LPO and SPO of omphacite can help reconstruct slab-deformation evolution and exhumation processes (Kurz et al., 2004; Kurz, 2005). The prevalent SL-/LS-/L-type LPO of omphacite in eclogites probably implies that exhumation is related to low-angle corner ('subduction channel') sub-parallel extrusion. The S-type fabric of omphacite in the massive eclogite is generally associated with uniaxial-flattening or transpression strain during slab subduction, which has been demonstrated by previous studies (Boundy et al., 1992; Kurz et al., 2004; Neufeld et al., 2008). A recent experimental study on the deformation of omphacite (Zhang et al., 2006) shows that S-type LPO was developed when the sample was uniaxially compressed and SL-type LPO was developed when the sample was deformed in simple shear. The SL-type fabric of omphacite, clinozoisite and rutile of the North Qilian eclogites can thus be interpreted as the development of LPO in a simple shearing process along the 'subduction channel'. The random distributed, rounded euhedral crystals of garnet from the massive eclogite (Fig. 2a), combining its prograde growth at increasing P-T conditions (Wei et al., 2009), suggests a weak stress/strain during subduction. Therefore, the variation of omphacite, clinozoisite and rutile fabrics from weak SL-type in the massive eclogite to strong SL-type in the foliated eclogite is more likely to represent the increment of shear strain during exhumation, while a transition of deformation geometry from subduction to exhumation could be expected.

#### ACKNOWLEDGEMENTS

We thank H. J. Xu for his help with the EBSD measurement and G. M. Shu for his help with the EPMA analysis. We also thank D. Whitney and T. Tsujimori for their detailed and constructive review comments, which led to a better presentation of the final product. This study was supported by National Natural Science Foundation of China (grant nos. 40825007, 40773012, 40821002, 40730314), the Major State Basic Research Development Projects (2009CB825007) and by the grant CATER 2008-5112 (H. J.) and BK21 at SEES in SNU (H. J.).

#### REFERENCES

Avé Lallemant, H.G., 1978. Experimental deformation of diopside and websterite. *Tectonophysics*, 48, 1–27.

- Bascou, J., Barruol, G., Vauchez, A., Mainprice, D. & Egydio-Silva, M., 2001. EBSD-measured lattice-preferred orientations and seismic properties of eclogites. *Tectonophysics*, 342, 61–80.
- Bascou, J., Tommasi, A. & Mainprice, D., 2002. Plastic deformation and development of clinopyroxene lattice preferred orientations in eclogites. *Journal of Structural Geology*, 24, 1357–1368.
- Blanchin, M.G., Bursill, L.A. & Lafage, C., 1990. Deformation and microstructure of rutile. *Proceedings of the Royal Society* of London. Series A: Mathematical Physical and Engineering Sciences, 429, 175–202.
- Blenkinsop, T.G., 2000. *Deformation Microstructures and Mechanisms in Minerals and Rocks*. Kluwer Academic Publishers, Dordrecht, the Netherlands, pp. 25–27.
- Boundy, T.M., Fountain, D.M. & Austrheim, H., 1992. Structural development and petrofabrics of eclogite facies shear zones, Bergen Arcs, Western Norway – implications for deep crustal deformational processes. *Journal of Metamorphic Geology*, 10, 127–146.
- Brenker, F.E., Prior, D.J. & Muller, W.F., 2002. Cation ordering in omphacite and effect on deformation mechanism and lattice preferred orientation (LPO). *Journal of Structural Geology*, 24, 1991–2005.
- den Brok, B. & Kruhl, J.H., 1996. Ductility of garnet as an indicator of extremely high temperature deformation: discussion. *Journal of Structural Geology*, **18**, 1369–1373.
- Buatier, M., Van Roermund, H.L.M., Drury, M.R. & Lardeaux, J.M., 1991. Deformation and recrystallization mechanisms in naturally deformed omphacites from the Sesia-Lanzo Zone – geophysical consequences. *Tectonophysics*, 195, 11–27.
- Cordier, P., Raterron, P. & Wang, Y., 1996. TEM investigation of dislocation microstructure of experimentally deformed silicate garnet. *Physics of the Earth and Planetary Interiors*, **97**, 121–131.
- Davis, P.B. & Whitney, D.L., 2006. Petrogenesis of lawsonite and epidote eclogite and blueschist, Sivrihisar Massif, Turkey. *Journal of Metamorphic Geology*, **24**, 823–849.
- Donohue, C.L. & Essene, E.J., 2000. An oxygen barometer with the assemblage garnet-epidote. *Earth and Planetary Science Letters*, **181**, 459–472.
- Droop, G.T.R., 1987. A general equation for estimating Fe<sup>3+</sup> concentrations in ferromagnesian silicates and oxides from microprobe analyses, using stoichiometric criteria. *Mineralogical Magazine*, **51**, 431–435.
- Durham, B., Goetze, C. & Blake, B., 1977. Plastic flow of oriented single crystals of olivine: 2, Observations and interpretations of the dislocation structures. *Journal of Geophysical Research*, 82, 5755–5770.
- Godard, G. & van Roermund, H.L.M., 1995. Deformation-induced clinopyroxene fabrics from eclogites. *Journal of Structural Geology*, **17**, 1425–1443.
- Green, H.W. II & Jung, H., 2005. Fluids, faulting, and flow. *ELEMENTS*, 1, 31–37.
- Helmstaedt, H., Anderson, O.L. & Gavasci, A.T., 1972. Petrofabric studies of eclogite, spinel-websterite, and spinel-Iherzolite xenoliths from kimberlite-bearing breccia pipes in southeastern Utah and northeastern Arizona. *Journal of Geophysical Research*, 77, 4350–4365.
- Holland, T.J.B. & Powell, R., 1998. An internally consistent thermodynamic data set for phases of petrological interest. *Journal of Metamorphic Geology*, **16**, 309–343.
- Ingrin, J. & Madon, M., 1995. TEM observations of several spinel-garnet assemblies: toward the rheology of the transition zone. *Terra Nova*, 7, 509–515.
  Ingrin, J., Doukhan, N. & Doukhan, J.C., 1991. High-temper-
- Ingrin, J., Doukhan, N. & Doukhan, J.C., 1991. High-temperature deformation of diopside single-crystal. 2. Transmission electron-microscopy investigation of the defect microstructures. *Journal of Geophysical Research-Solid Earth and Plan*ets, 96, 14287–14297.

- Ji, S. & Martignole, J., 1994. Ductility of garnet as an indicator of extremely high temperature deformation. *Journal of Structural Geology*, 16, 985–996.
- Ji, S., Saruwatari, K., Mainprice, D., Wirth, R., Xu, Z. & Xia, B., 2003. Microstructures, petrofabrics and seismic properties of ultrahigh-pressure eclogites from Sulu region, China: implications for rheology of subducted continental crust and origin of mantle reflections. *Tectonophysics*, **370**, 49–76.
- Jung, H., 2009. Deformation fabrics of olivine in Val Malenco peridotite found in Italy and implications for the seismic anisotropy in the upper mantle. *Lithos*, **109**, 341–349.
- Jung, H. & Green, H.W., 2004. Experimental faulting of serpentinite during dehydration: implications for earthquakes, seismic low-velocity zones, and anomalous hypocenter distributions in subduction zones. *International Geology Review*, 46, 1089–1102.
- Jung, H. & Karato, S.-I., 2001. Effects of water on dynamically recrystallized grain-size of olivine. *Journal of Structural Geol*ogy, 23, 1337–1344.
- Jung, H., Green, H.W. & Dobrzhinetskaya, L.F., 2004. Intermediate-depth earthquake faulting by dehydration embrittlement with negative volume change. *Nature*, 428, 545–549.
- Jung, H., Katayama, I., Jiang, Z., Hiraga, I. & Karato, S., 2006. Effect of water and stress on the lattice-preferred orientation of olivine. *Tectonophysics*, 421, 1–22.
- Jung, H., Park, M., Jung, S. & Lee, J., 2010. Lattice preferred orientation, water content, and seismic anisotropy of orthopyroxene. *Journal of Earth Science*, 21, 555–568.
- Karato, S., Wang, Z. & Fujino, K., 1994. High-temperature creep of Yttrium-Aluminum-Garnet single-crystals. *Journal of Materials Science*, 29, 6458–6462.
- Karato, S., Wang, Z.C., Liu, B. & Fujino, K., 1995. Plasticdeformation of garnets – systematics and implications for the rheology of the mantle transition zone. *Earth and Planetary Science Letters*, **130**, 13–30.
- Kurz, W., 2005. Constriction during exhumation: evidence from eclogite microstructures. *Geology*, **33**, 37–40.
- Kurz, W., Jansen, E., Hundenborn, R., Pleuger, J., Schafer, W. & Unzog, W., 2004. Microstructures and crystallographic preferred orientations of omphacite in Alpine eclogites: implications for the exhumation of (ultra-) high-pressure units. *Journal of Geodynamics*, 37, 1–55.
- Liou, J.G., Wang, X.M., Coleman, R.G., Zhang, Z.M. & Maruyama, S., 1989. Blueschists in major suture zones of China. *Tectonics*, 8, 609–619.
- Liu, Y.J., Neubauer, F., Genser, J., Takasu, A., Ge, X.H. & Handler, R., 2006. <sup>40</sup>Ar/<sup>39</sup>Ar ages of blueschist facies pelitic schists from Qingshuigou in the northern Qilian mountains, western China. *Island Arc*, **15**, 187–198.
   Lorimer, G.W., Spooner, E.T.C. & Champnes, P., 1972.
- Lorimer, G.W., Spooner, E.T.C. & Champnes, P., 1972. Dislocation distributions in naturally deformed omphacite and albite. *Nature-Physical Science*, 239, 108–109.
- Mattinson, C.G., Zhang, R.Y., Tsujimori, T. & Liou, J.G., 2004. Epidote-rich talc-kyanite-phengite eclogites, Sulu terrane, eastern China: P-T-f( $O_2$ ) estimates and the significance of the epidote-talc assemblage in eclogite. *American Mineralogist*, **89**, 1772–1783.
- Mauler, A., Bystricky, M., Kunze, K. & Mackwell, S., 2000. Microstructures and lattice preferred orientations in experimentally deformed clinopyroxene aggregates. *Journal of Structural Geology*, 22, 1633–1648.
- Mauler, A., Godard, G. & Kunze, K., 2001. Crystallographic fabrics of omphacite, rutile and quartz in Vendée eclogites (Armorican Massif, France). Consequences for deformation mechanisms and regimes. *Tectonophysics*, **342**, 81–112.
- Neufeld, K., Ring, U., Heidelbach, F., Dietrich, S. & Neuser, R.D., 2008. Omphacite textures in eclogites of the Tauern Window: implications for the exhumation of the Eclogite Zone, Eastern Alps. *Journal of Structural Geology*, **30**, 976– 992.
- Philippot, P. & Van Roermund, H.L.M., 1992. Deformation processes in eclogitic rocks – evidence for the rheological

delamination of the oceanic-crust in deeper levels of subduction zones. *Journal of Structural Geology*, **14**, 1059–1077.

- Piepenbreier, D. & Stöckhert, B., 2001. Plastic flow of omphacite in eclogites at temperatures below 500 °C – implications for interplate coupling in subduction zones. *International Journal* of Earth Sciences, **90**, 197–210.
- Prior, D.J., Wheeler, J., Brenker, F.E., Harte, B. & Matthews, M., 2000. Crystal plasticity of natural garnet: new microstructural evidence. *Geology*, 28, 1003–1006.
  Raterron, P. & Jaoul, O., 1991. High-temperature deformation
- Raterron, P. & Jaoul, O., 1991. High-temperature deformation of diopside single crystal, 1, Mechanical data. *Journal of Geophysical Research*, 96, 14277–14286.
- Raterron, P., Doukhan, N., Jaoul, O. & Doukhan, J.C., 1994. High temperature deformation of diopside IV: predominance of {110} glide above 1000°C. *Physics of the Earth and Planetary Interiors*, 82, 209–222.
- Ravna, E.J.K. & Terry, M.P., 2004. Geothermobarometry of UHP and HP eclogites and schists – an evaluation of equilibria among garnet-clinopyroxene-kyanite-phengitecoesite/quartz. *Journal of Metamorphic Geology*, 22, 579–592.
- Robinson, P., Spear, F.S. & Schumacher, J.C., 1982. Phase relations of metamorphic amphiboles: natural occurrence and theory. In: *Amphiboles, Petrology and Experimental Phase Relations, Reviews of Mineralogy*, Vol. 9b (eds Veblen, D.R. & Ribbe, P.H.), pp. 1–227. Minerological Society of America, Washington, DC.
- van Roermund, H., 1983. Petrofabrics and microstructures of omphacites in a high temperature eclogite from the Swedish Caledonides. *Bulletin of Mineralogy*, **106**, 709–713.
- van Roermund, H.L.M. & Boland, J.N., 1981. The dislocation substructures of naturally deformed omphacites. *Tectonophysics*, **78**, 403–418.
- Skemer, P., Katayama, I., Jiang, Z. & Karato, S.-I., 2005. The misorientation index: development of a new method for calculating the strength of lattice-preferred orientation. *Tectonophysics*, 411, 157–167.
- Song, S.G., Zhang, L.F., Niu, Y.L., Song, B., Zhang, G.B. & Wang, Q.J., 2004. Zircon U-PbSHRIMP ages of eclogites from the North Qilian Mountains in NW China and their tectonic implication. *Chinese Science Bulletin*, **49**, 848–852.
- Song, S.G., Zhang, L.F., Niu, Y.L., Su, L., Song, B.A. & Liu, D.Y., 2006. Evolution from oceanic subduction to continental collision: a case study from the Northern Tibetan Plateau based on geochemical and geochronological data. *Journal of Petrology*, 47, 435–455.
- Song, S.G., Zhang, L.F., Niu, Y.L., Wie, C.J., Liou, J.G. & Shu, G.M., 2007. Eclogite and carpholite-bearing metasedimentary rocks in the North Qilian suture zone, NW China: implications for early Palaeozoic cold oceanic subduction and water transport into mantle. *Journal of Metamorphic Geology*, 25, 547–563.
- Song, S.G., Niu, Y., Zhang, L.F., Wei, C.J., Liou, J.G. & Su, L., 2009. Tectonic evolution of Early Paleozoic HP metamorphic rocks in the North Qilian Mountains, NW China: new perspectives. *Journal of Asian Earth Sciences*, 35, 334–353.
- Song, S.G., Su, L., Li, X.-H., Niu, Y.L., Zhang, G.B. & Zhang, L.F., 2010. Tracing the 850-Ma continental flood basalts from a piece of subducted continental crust in the North Qaidam UHPM belt, NW China. *Precambrian Research*, **183**, 805–816.
- Storey, C.D. & Prior, D.J., 2005. Plastic deformation and recrystallization of garnet: a mechanism to facilitate diffusion creep. *Journal of Petrology*, 46, 2593–2613.
- Su, W., You, Z.D., Cong, B.L. & Ye, K., 2003. Role of water in deformed omphacite in UHP eclogite from the Dabie Mountains, eastern China. Acta Geologica Sinica-English Edition, 77, 320–325.
- Sun, S.S. & McDonough, W.F., 1989. Chemical and isotopic systematics of oceanic basalt: implications for mantle composition and processes. In: *Magmatism in the Ocean Basins* (eds Saunders, A.D. & Norry, M.J.), *Geological Society*, *London, Special Publications*, 42, 313–345.

- Terry, M.P. & Heidelbach, F., 2004. Superplasticity in garnet from eclogite facies shear zones in the Haram Gabbro, Haramsoya, Norway. *Geology*, **32**, 281–284.
- Terry, M.P. & Heidelbach, F., 2006. Deformation-enhanced metamorphic reactions and the rheology of high-pressure shear zones, Western Gneiss Region, Norway. *Journal of Metamorphic Geology*, 24, 3–18.
- Tsujimori, T., Sisson, V.B., Liou, J.G., Harlow, G.E. & Sorensen, S.S., 2006. Petrologic characterization of Guatemalan lawsonite eclogite: eclogitization of subducted oceanic crust in a cold subduction zone. In: Ultra High Pressure Metamorphism: Deep Continental Subduction (eds Hacker, B.R., McClelland, W.C. & Liou, J.G.), Geological Society of America Special Paper, 403, 147–168.
- Voegelé, V., Ando, J.I., Cordier, P. & Liebermann, R.C., 1998a. Plastic deformation of silicate garnets: I. High-pressure experiments. *Physics of the Earth and Planetary Interiors*, 108, 305–318.
- Voegelé, V., Cordier, P., Sautter, V., Sharp, T.G., Lardeaux, J.M. & Marques, F.O., 1998b. Plastic deformation of silicate garnets: II. Deformation microstructures in natural samples. *Physics of the Earth and Planetary Interiors*, **108**, 319–338.
- Wang, Z.C. & Ji, S.C., 1999. Deformation of silicate garnets: brittle-ductile transition and its geological implications. *Canadian Mineralogist*, 37, 525–541.
- Wang, Q., Ji, S.C., Xu, Z.Q., Salisbury, M.H., Xia, B. & Pan, M.B., 2005. Seismic properties of eclogites: implications on crustal composition and exhumation mechanism of the Sulu ultrahigh-pressure metamorphic belt. *Acta Petrologica Sinica*, 21, 451–464.
- Wei, C.J. & Song, S.G., 2008. Chloritoid-glaucophane schist in the North Qilian orogen, NW China: phase equilibria and P-T path from garnet zonation. *Journal of Metamorphic Geology*, 26, 301–316.
- Wei, C.J., Yang, Y., Su, X.L., Song, S.G. & Zhang, L.F., 2009. Metamorphic evolution of low-*T* eclogite from the North Qilian orogen, NW China: evidence from petrology and calculated phase equilibria in the system NCKFMASHO. *Journal of Metamorphic Geology*, 27, 55–70.
- Whitney, D.L. & Evans, B.W., 2010. Abbreviations for names of rock-forming minerals. *American Mineralogist*, 95, 185– 187.
- Wu, H.Q., Feng, Y.M. & Song, S.G., 1993. Metamorphism and deformation of blueschist belts and their tectonic implications, North Qilian Mountains, China. *Journal of Metamorphic Geology*, **11**, 523–536.

- Zhang, J.F. & Green, H.W., 2007a. Experimental investigation of eclogite rheology and its fabrics at high temperature and pressure. *Journal of Metamorphic Geology*, 25, 97–115.
- Zhang, J.F. & Green, H.W., 2007b. On the deformation of UHP eclogite: from laboratory to nature. *International Geology Review*, 49, 487–503.
- Zhang, J.X. & Meng, F.C., 2006. Lawsonite-bearing eclogites in the north Qilian and north Altyn Tagh: evidence for cold subduction of oceanic crust. *Chinese Science Bulletin*, **51**, 1238–1244.
- Zhang, J.X., Xu, Z.Q., Chen, W. & Xu, H.F., 1997. A Tentative discussion on the ages of the subduction – accretionary complex/volcanic arcs in the middle sector of North Qilian Mountain. Acta Petrologica et Mineralogica, 16, 112– 119.
- Zhang, J.F., Green, H.W. & Bozhilov, K.N., 2006. Rheology of omphacite at high temperature and pressure and significance of its lattice preferred orientations. *Earth and Planetary Science Letters*, 246, 432–443.
- Zhang, J.X., Meng, F.C. & Wan, Y.S., 2007. A cold early Palaeozoic subduction zone in the North Qilian Mountains, NW China: petrological and U-Pb geochronological constraints. *Journal of Metamorphic Geology*, 25, 285–304.

#### **APPENDIX**

M-index (misorientation index) is a method for quantifying the LPO fabric strength based on distribution of uncorrelated misorientation angles proposed by Skemer *et al.* (2005). *M*-index is defined by

$$M = \frac{1}{2} \int \left| R^{T}(\theta) - R^{0}(\theta) \right| d\theta$$

to describe the difference between the observed distribution of uncorrelated misorientation angles  $R^{0}(\theta)$  (obtained from EBSD data) and distribution of uncorrelated misorientation angles for a random fabric  $R^{T}(\theta)$ . It ranges from 0 (random fabric) to 1 (single crystal fabric) and is insensitive to and less dependent on the parameters specified for its calculation and grain number (>150), respectively.

Received 8 May 2010; revision accepted 22 February 2011.