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Lawsonite blueschists and lawsonite eclogites as proxies for palaeo-subduction zone processes: a review

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ABSTRACT Lawsonite-bearing metamorphic/metasomatic rocks form at high-pressure–ultrahigh pressure (H*P*–UH*P*) and low-temperature (L*T*) conditions, commonly in Pacific-type subduction zones. The *P–T* stability fields of lawsonite blueschists and lawsonite eclogites represent subfacies of the blueschist and eclogite facies, respectively. Although the lawsonite–epidote transition boundary has a positive Clapeyron (d*P*/d*T*) slope, the blueschist-to-eclogite transformation within the lawsonite stability field in metabasaltic rocks is gradual and cannot be defined by a specific discontinuous reaction in *P–T* space. The oldest occurrences of lawsonite-bearing blueschists are latest Neoproterozoic, suggesting that subduction-zone thermal structures evolved towards the necessary L*T* conditions for lawsonite formation only by the late Neoproterozoic. A clear difference in frequency between Phanerozoic lawsonite and epidote blueschists does not exist, but our new compilation found a global lawsonite hiatus in the Permian that is a robust indication of relatively warm subduction-zone thermal regimes. Lawsonite eclogites have been confirmed from at least 19 localities; they are classified as L- (lawsonite only), E- (lawsonite + epidote), and U-type (lawsonite + coesite). Complete preservation of L-type lawsonite eclogites attending their return to the surface is uncommon. Rare evidence of progressive eclogitization within the lawsonite stability field is preserved in some zoned garnets, as growth isolates a significant volume of precursor phases and textures during incipient eclogitization. Brittle fracturing and fluid infiltration are common during prograde eclogite facies metamorphism. Certain lawsonite-bearing metasomatic rocks record multiple fluid-infiltration events. Significant cooling and continuous H_2O supply from the dehydrating oceanic plate to exhuming H*P* serpentinite mélange may cause lawsonite blueschist facies overprinting and prevent breakdown of lawsonite during decompression. The subduction records of lawsonite blueschists and eclogites agree with numerical modelling of subduction zones.

Key words: H*P*–UH*P* metamorphism; lawsonite blueschist; lawsonite eclogite; Pacific-type convergent margins; progressive eclogitization.

INTRODUCTION

Blueschists and glaucophane-bearing low-temperature eclogites have been documented for almost half a century in the Earth's Neoproterozoic–Phanerozoic orogenic belts. As a key proxy of oceanic plate subduction along a cold geothermal gradient marking ancient convergent margins, these metamorphic rocks have been intensively studied from various viewpoints, and at scales ranging from single crystals to mountain belts. In particular, understanding of metamorphic processes involving the generation of lawsonite [$CaAl_2Si_2O_7(OH)_2{\cdot}H_2O$] in subduction zones is of considerable importance. Lawsonite possesses a high H_2O content (~11.5 wt% H_2O), a high-pressure stability limit (up to 8.5 GPa), a high shear elastic anisotropy (74%), a high V_P/V_S ratio (1.94), and a high Poisson's ratio (σ = 0.318) (cf., Sinogeikin *et al.*, 2000), so lawsonite can potentially serve as a geochemical and geophysical tracer (Abers *et al.*, 2013; Reynard & Bass, 2014). Research interest in lawsonite-bearing metamorphic rocks has not only included field geology, petrology, geochronology and high-pressure experiments (e.g., Mulcahy *et al.*, 2009; Davis, 2011; Martin *et al.*, 2011; Vitale Brovarone *et al.*, 2011; Chantel *et al.*, 2012; Compagnoni *et al.*, 2012), but current attention also reflects their significance regarding geochemical components of subduction-zone magmas, dynamics of the subduction-zone channel overlying the descending slab, and the rheology + seismology of the subducting oceanic crust (e.g., Hacker *et al.*, 2003; Bebout, 2007; Dorbath *et al.*, 2008; Hacker, 2008; Teyssier *et al.*, 2010; Abers *et al.*, 2013; Cao *et al.*, 2013; Kim *et al.*, 2013; Spandler & Pirard, 2013). However, much still remains to be done for a more interdisciplinary understanding of the lawsonite-forming Phanerozoic-modern style of oceanic subduction.

Many review articles and special issues related to HP–LT metamorphism in Pacific-type subduction zones have been published (e.g., Miyashiro & Banno, 1958; Ernst, 1963, 1972, 1973; Evans & Brown, 1986; Dobretsov *et al*., 1987; Peacock, 1992; Maruyama *et al*., 1996; Agard *et al*., 2009). In 2006, the first author investigated the subgroup of eclogites that contains lawsonite with a specific focus on their very-low temperature record of the subduction process (Tsujimori *et al*., 2006a). Further details on the metamorphic evolution of lawsonite eclogites worldwide were presented by Wei & Clarke (2011) based on calculated phase equilibria for MORB compositions. In this article, we update some new insights regarding lawsonite-bearing eclogites, but we also expand comparisons to lawsonite blueschists from both our own data and the literature. Specific foci in this review include (i) petrological aspects of lawsonite-bearing metamorphic rocks, (ii) spatial and temporal distributions of lawsonite-bearing metamorphic rocks through Earth history, and (iii) specific problems related to lawsonite-bearing rocks.

In this review, the term eclogite is used for any metabasite in which almandine-rich garnet and omphacite are present as an equilibrium assemblage, regardless of their abundance, although the IUGS Subcommision on the Systematics of Metamorphic Rocks recommends that the term eclogite be restricted to a plagioclase-free rock composed of >75 vol% of omphacite + garnet (Desmons & Smulikowski, 2007). Throughout this study, mineral abbreviations are after Whitney & Evans (2010).

PETROLOGICAL ASPECTS OF LAWSONITE-BEARING METAMORPHIC ROCKS

General overview

According to the classic metamorphic facies concept, lawsonite is an index mineral confined to blueschist facies rocks (e.g., Miyashiro, 1973; Turner, 1981). However, the existence of lawsonite eclogites led to their recognition as a subfacies of the eclogite facies based on both natural and experimental evidence (Helmstaedt & Doig, 1975; Poli & Schmidt, 1995; Maruyama *et al*., 1996; Okamoto & Maruyama, 1999). Lawsonite occurs commonly as prismatic and tabular euhedral-to-subhedral crystals in various regional metamorphic rocks – including metasomatic rocks – formed within mainly Pacific-type orogenic belts (cf. Tsujimori *et al*., 2006a); rare HP lawsonite blueschists and UHP lawsonite-bearing eclogites have been reported as xenoliths within rhyolitic tephra deposits and kimberlitic pipes, respectively (e.g., Watson & Morton, 1969; Hoffmann & Keller, 1979; Usui *et al*., 2003). Lawsonite typically shows twinning on the {101} plane and rare sector growth internal structure denoted as the {100}, {010} and {001} sectors. Usually non-pleochroic, it commonly is white, but rarely shows blue or pink colours due to substitution of Fe^{3+} and Cr for Al; the diaphaneity is transparent. Because of its hardness (7.5), high density (3.09 g cm^{-3}), and relatively great corrosion resistance, lawsonite is well known as a detrital heavy mineral useful for identifying provenance-specific features in sandstones (e.g., Mange-Rajctzky & Oberhänsli, 1982; Mange *et al*., 2005).

The P–T stability fields of lawsonite and lawsonite-bearing mineral equilibria have been determined in numerous experimental studies (see Martin *et al*., 2014). The minimum pressure stability of lawsonite is defined by the decomposition reaction of laumontite to Lws + Qz + H_2O (Nitsch, 1968; Liou, 1971). Pistorius *et al*. (1962) first confirmed the high-pressure stability limit of lawsonite; they synthesized Lws + Qz and Lws + Coe equilibria at pressures above 2.3 GPa. The upper pressure stability of lawsonite-bearing mineral assemblages reaches 8–10 GPa at 750–900 °C (Pawley, 1994; Schmidt & Poli, 1994; Poli & Schmidt, 1995; Schmidt, 1995; Okamoto & Maruyama, 1999). The lawsonite stability field covers a wide P–T range and overlaps that of glaucophane (Fig. 1a). Lawsonite-bearing metamorphic rocks mainly belong to the subfacies of the blueschist and eclogite facies. However, lawsonite also occurs in a transition zone between blueschist facies and pumpellyite–actinolite facies rocks (e.g., Banno, 1998). Although Turner (1981) proposed the lawsonite–albite–chlorite facies, we regard this assemblage as the lowest grade of the lawsonite blueschist facies because lawsonite-bearing mineral assemblages without glaucophane (*sensu lato*) can exist in lawsonite blueschist facies metabasalt (e.g., Hirajima, 1983; Tsujimori & Liou, 2007). In the very-low grade lawsonite-bearing rocks of New Zealand, morphological preservation of carbonaceous plant macrofossils has been described (e.g., Galvez *et al*., 2012).

Lawsonite blueschist facies rocks

The Lws + Gln paragenesis defines the lawsonite blueschist facies; in the NCASH system, the discontinuous reaction Lws + Ab (or Jd) = Zo + Pg (or + Qz) + H_2O defines the boundary between the lawsonite blueschist and epidote blueschist facies characterized by the Ep + Gln + Qtz paragenesis (Evans, 1990). The lawsonite–epidote transition boundary in the blueschist facies, including a narrow transition zone, has a positive Clapeyron ($\mathrm{d}P/\mathrm{d}T$) slope. As shown in the calculated phase diagrams for MORB of Fig. 1b,c, if the rock is fully hydrated (i.e., saturated in H_2O-rich fluid), metabasalts in the lawsonite blueschist facies can retain at least 4.5 wt% H_2O in its solid phases; this is more than 1.5 wt% greater than that of epidote blueschist facies rocks at the same depth. However, this infers that formation and preservation of lawsonite-bearing mineral

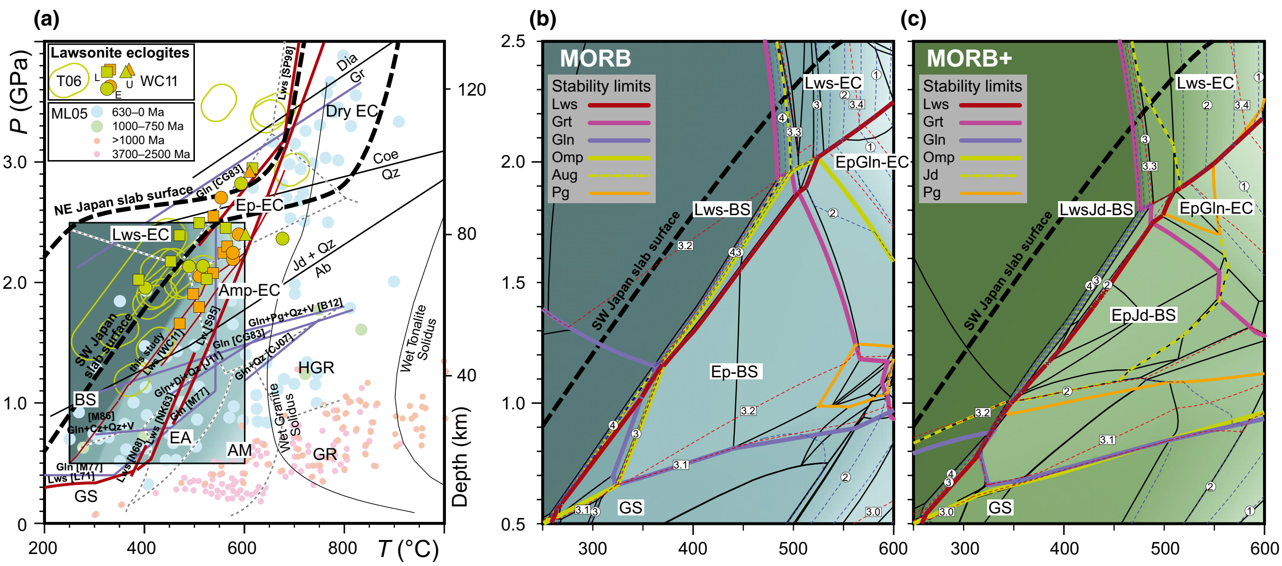
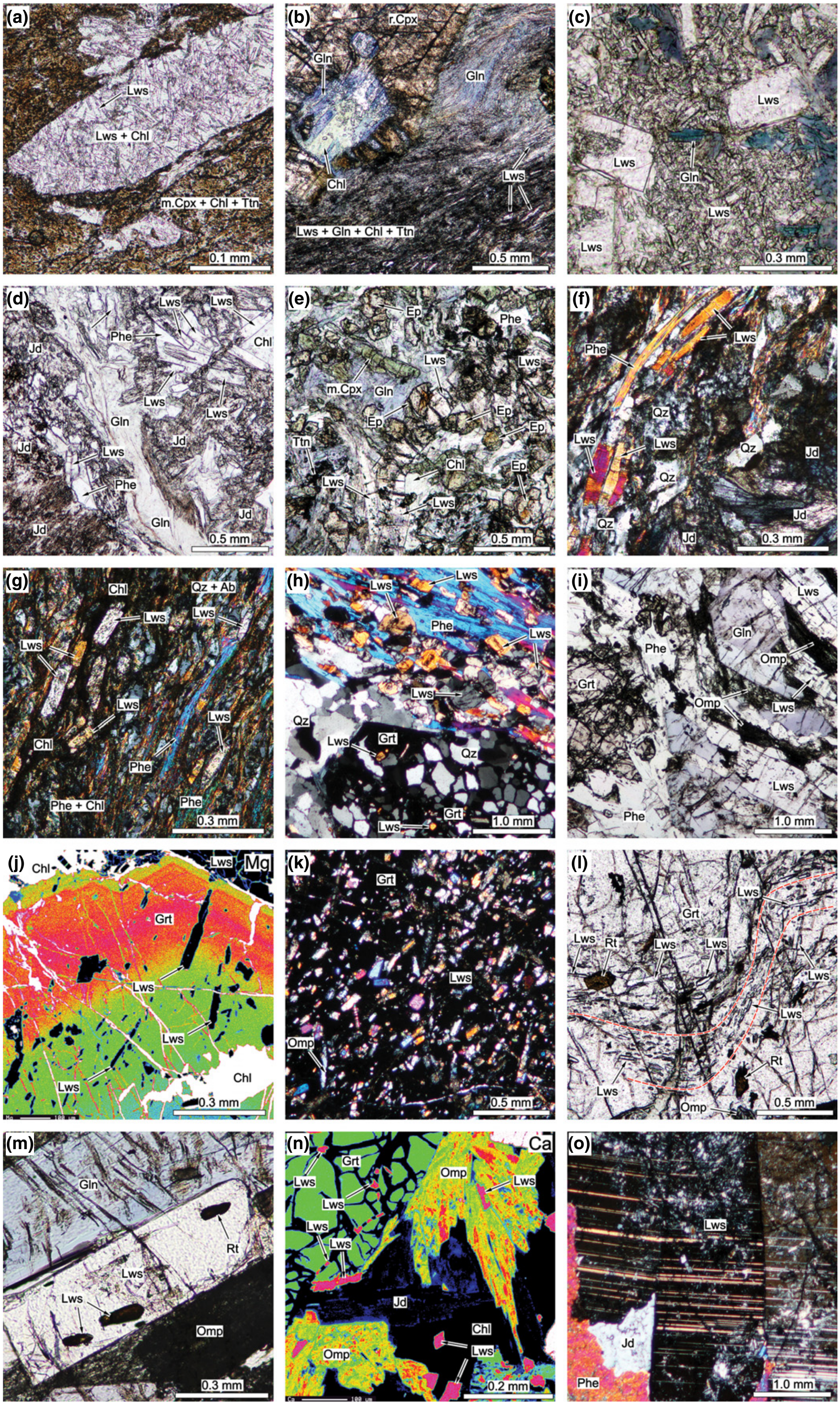


Fig. 1. (a) P–T diagram showing experimentally determined stability limits of lawsonite [Lws]- (red bold lines) and glaucophane [Gln]-bearing (lavender lines) equilibria and modelled P–T paths (model D80 of Syracuse *et al*., 2010) for slab surfaces, both warm (SW Japan/Nankai) and cool (NE Japan/Tohoku) subduction zones. [SP98] = Schmidt & Poli (1998); [S95] = Schmidt (1995); [L71] = Liou (1971); [N68] = Nitsch (1968); [NK63] = Newton & Kennedy (1963); [M77] = Maresch (1977); [M86] = Maruyama *et al*. (1986); [CG83] = Carman & Gilbert (1983); [CJ07] = Corona & Jenkins (2007); [J11] = Jenkins (2011); [B12] = Basora *et al*. (2012). Red thin lines represent calculated high-temperature stability limits of lawsonite for a MORB bulk-rock composition in NCFMASH+Mn system by Wei & Clarke (2011) [WC11] and this study (See the details in text). Metamorphic facies boundaries (dotted lines) are after Maruyama *et al*. (1996) and Liou *et al*. (2004). A boxed P–T space (250–600 °C and 0.5–2.5 GPa) is indicated as equilibrium phase diagrams for panels (b) and (c). P–T estimates of lawsonite eclogites by Tsujimori *et al*. (2006a) [T06] and Wei & Clarke (2011) [WC11] are presented; light green – conventional thermobarometry, orange – garnet isopleths of P–T pseudosection. For comparison, P–T estimates of various subduction/collisional metamorphic rocks (Maruyama & Liou, 2005) [ML05] are also shown. Metamorphic facies abbreviations: BS = blueschist; AM = amphibolite; Lws-EC = lawsonite eclogite; Ep-EC = epidote eclogite; Amp-EC = amphibole eclogite; Dry-EC = dry eclogite; GS = greenschist; EA = epidote–amphibolite; GR = granulite; HGR = high-pressure granulite. (b) Equilibrium phase diagram calculated for the Gale *et al*. (2013) average MORB composition in NCFMASH system with excess H_2O and rutile–ilmenite oxygen buffer. The colour graduation level represents H_2O wt% of solids; blue-dotted lines are contour lines. Contours of density (g cm^{-3}) are also shown as red dotted lines. The thermodynamic modelling was performed using the Theriak/Domino software (de Capitani & Petrakakis, 2010) and the internally consistent thermodynamic data set of Holland & Powell (1998); solution models and oxygen buffer suggested by Konrad-Schmolke *et al*. (2008) were used. (c) P–T diagrams calculated for a hypothetical bulk-rock composition MORB+ (average MORB + 5 mole jadeite).

equilibria at H_2O undersaturation are strongly controlled by H_2O availability (see "lawsonite paradox" by Clarke *et al*., 2006). The stability of lawsonite-bearing mineral assemblages is also controlled by the XCO_2 of the fluid phase (Goto *et al*., 2007; Poli *et al*., 2009). Moreover inasmuch as high-oxidation state stabilizes epidote, the position of the lawsonite–epidote phase boundary shifts to higher pressures at elevated values of fO_2 (e.g., Warren & Waters, 2006; López-Carmona *et al*., 2013).

In very low-grade lawsonite blueschist facies metabasites, lawsonite typically occurs with chlorite, glaucophane (*sensu lato*), and Na- and/or Ca-Na pyroxene (Fig. 2a); lawsonite is also commonly associated with pumpellyite, phengite, stilpnomelane and rare actinolite. The Ti-bearing phase is generally titanite, but rarely ilmenite. Igneous augite is a common relict mineral (Fig. 2b). Irregular-shaped domains/clots or veins consisting of aggregates of lawsonite are common in undeformed/weakly deformed rocks (Fig. 2c). Vein occurrences suggest the precipitation and growth of lawsonite crystals from a metamorphic fluid (e.g., Davis, 1960; Schertl *et al*., 2012). In the Piemont zone of the Western Alps, the Lws + Jd assemblage replacing igneous plagioclase in metagabbro indicates lawsonite blueschist facies metamorphism (Mevel & Kienast, 1980) (Fig. 2d).

The paragenetic sequence of Ca-Al hydrous silicates (lawsonite, pumpellyite and epidote) has been recognized as a tool to trace metamorphic field gradients since Ernst (1963). Coexistence and/or replacement textures among these three Ca-Al hydrous silicates have been described from some lawsonite blueschists. Metamorphic zones based on Ca-Al hydrous silicates were mapped in coherent blueschist units, such as the Cazadero area of the Franciscan Complex (Maruyama & Liou, 1988), the Kamuikotan Belt (Shibakusa, 1989) (Fig. 2e), and the Diahot terrane of NE New Caledonia (Brothers, 1970; Fitzherbert *et al*., 2003; Vitale Brovarone & Agard, 2013). With increasing metamorphic grade, epidote


(a)
Lws
Lws + Chl
m.Cpx + Chl + Ttn
0.1 mm
(b)
r.Cpx
Gln
Gln
Chl
Lws
Lws + Gln + Chl + Ttn
0.5 mm
(c)
Lws
Lws
Gln
Lws
Lws
0.3 mm
(d)
Lws
Lws
Phe
Chl
Jd
Lws
Lws
Gln
Jd
Jd
Lws
Phe
Gln
Jd
Jd
0.5 mm
(e)
Ep
Phe
m.Cpx
Gln
Lws
Lws
Ep
Ep
Ep
Ttn
Chl
Ep
Lws
Lws
0.5 mm
(f)
Phe
Lws
Qz
Lws
Lws
Qz
Qz
Jd
Qz
Jd
Jd
0.3 mm
(g)
Chl
Qz + Ab
Lws
Lws
Lws
Lws
Lws
Chl
Phe
Phe + Chl
Phe
0.3 mm
(h)
Lws
Lws
Phe
Lws
Lws
Qz
Lws
Grt
Qz
Grt
Lws
1.0 mm
(i)
Lws
Gln
Phe
Omp
Grt
Omp
Lws
Lws
Phe
1.0 mm
(j)
Chl
Lws
Mg
Grt
Lws
Lws
Lws
Chl
0.3 mm
(k)
Grt
Lws
Omp
0.5 mm
(l)
Lws
Grt
Lws
Rt
Lws
Lws
Lws
Lws
Rt
Lws
Omp
0.5 mm
(m)
Gln
Rt
Lws
Lws
Omp
0.3 mm
(n)
Ca
Grt
Omp
Lws
Lws
Lws
Lws
Lws
Jd
Chl
Lws
Omp
0.2 mm
(o)
Lws
Jd
Phe
1.0 mm

appears in lawsonite-bearing metabasites; in some case, a pumpellyite zone is present at lower grades than the lawsonite zone and/or between lawsonite and epidote zones. In general, the lawsonite–pumpellyite transition occurs <~0.6 GPa at <~350 °C (e.g., Liou *et al.*, 1985).

Lawsonite also occurs in metamorphosed trench-fill sedimentary rocks such as metagreywackes and Ca-bearing metapelites of the lawsonite blueschist facies (Fig. 2f–h). For instance, the occurrence of Lws + Jd is a characteristic feature of Franciscan H*P* metagreywackes (e.g., Ernst, 1993). Rare paragonite has been identified in very low-grade lawsonite blueschist facies metasedimentary rocks (Dalla Torre *et al.*, 1996; Potel *et al.*, 2006). In continental collisional zones (i.e., Alpine type), a lawsonite blueschist facies mineral assemblage in peraluminous metapelites and impure marbles typically contains Fe-Mg carpholite [$(Mg,Fe)Al_2Si_2O_6(OH)_4$] or chloritoid (e.g., Ganne *et al.*, 2003).

Lawsonite eclogite facies rocks

The Grt + Omp (or Jd) + Lws ± Gln paragenesis defines the lawsonite eclogite facies. Compared with lawsonite blueschist occurrences, the preservation of lawsonite eclogites is rather rare; all lawsonite eclogites belong to the Group C eclogites of Coleman *et al.* (1965). Lawsonite eclogite facies mineral equilibria can be divided into glaucophane-bearing and glaucophane-free fields; those two fields are further divided by the quartz–coesite transition (e.g., Wei & Clarke, 2011). In general, rigid minerals such as garnet are the best mineral containers; in fact, lawsonite crystals in eclogites survive much more completely as inclusions in garnet than in clinopyroxene (cf., Tsujimori *et al.*, 2006a). However, these inclusions are generally pseudomorphed by an aggregate of Czo/Ep + Pg or Czo/Ep + Ky ± Pg, suggesting that lawsonite is decomposed by dehydration during exhumation.

In the lawsonite blueschist-to-lawsonite eclogite transition, lawsonite coexists with glaucophane, omphacite and garnet (Fig. 2i). The preservation of pristine lawsonite within almandine-rich eclogitic garnet serves as a petrographic indicator used to define lawsonite eclogite (Fig. 2j–l). Moreover, eclogite facies lawsonite is generally associated with rutile ± rare ilmenite rather than with titanite (Fig. 2l,m). Lawsonite eclogite facies mineral assemblages are sensitive to bulk-rock composition, especially to H_2O saturation during metamorphism. For example, Corsican lawsonite eclogites occur with garnet-bearing lawsonite blueschist that formed under the same metamorphic conditions of ~520 ± 20 °C and 2.3 ± 0.1 GPa; the contrasts in mineral parageneses are due to different bulk-rock compositions (Vitale Brovarone *et al.*, 2011). Moreover, the occurrence of lawsonite does not always reflect lawsonite eclogite facies metamorphism. Some Franciscan eclogites have been partially overprinted by secondary lawsonite blueschist facies mineral parageneses (Fig. 2n); these are "false lawsonite-eclogites". In some Ca-bearing pelitic schists associated with lawsonite eclogites, the assemblages Lws + Cld + Gln (Okay, 2002; Du *et al.*, 2011) and Grt + Gln + Lws ± Jd ± Cld (Lü *et al.*, 2012) are stable. Lawsonite has been also described from some jadeitites and related metasomatic rocks (e.g., Harlow *et al.*, 2003; Oberhänsli *et al.*, 2007). In such lithologies, lawsonite grew together with jadeite by fluid precipitation (Fig. 2o) and/or by metasomatic replacement of precursor minerals.

Fig. 2. Photomicrographs showing occurrence of lawsonite in various rocks. (a) Lws + Chl pseudomorphs after plagioclase in Lws-BS facies metamorphosed pillow basalt from Renge metamorphic rocks, SW Japan. The sample is characterized by a Na-amphibole–free mineral assemblage: metamorphic pyroxene (m.Cpx) ($jd_{<29}$) + Lws + Chl + Qtz (see Tsujimori & Liou, 2007) [Plane-polarized light (PPL)]. (b) Lawsonite crystals aligned along the schistosity in a lawsonite blueschist from Kamuikotan Belt (KMB), Hokkaido, Japan. The sample contains a Gln + Lws + Chl assemblage with relict igneous augite; jadeitic pyroxene occurs in the same locality (see Takayama, 1986) [PPL]. (c) Lawsonite-rich domain in Lws BS facies metadolerite, South Motagua Mélange (SMM), Guatemala [PPL]. (d) Lws + Jd assemblage pseudomorphing plagioclase of Lws-BS facies metagabbro [sample 336A of Ishiwatari (1985)], Roche Noire, Western Alps, France (see Mevel & Kienast, 1980) [PPL]. (e) Coexisting lawsonite and epidote in the highest grade coherent blueschist unit of KMB. The sample contains Gln + Lws + Ep + Chl + m.Cpx (jd_{36-37}) (see Shibakusa, 1989) [PPL]. (f) Lws + Jd + Qtz assemblage in Franciscan metagreywacke, Pacheco Pass, California, USA (see Ernst, 1993) [Crossed polar light (XPL)]. (g) Euhedral-to-subhedral crystals of lawsonite in a Lws-BSfacies pelitic schist, Osayama serpentinite mélange, Japan (see Tsujimori & Itaya, 1999) [XPL]. (h) Lawsonite in a garnet-bearing pelitic schist from SMM. Skeletal garnet contains lawsonite and rare prograde pumpellyite (see Tsujimori *et al.*, 2006b) [PPL]. (i) Coarse-grained lawsonite in a garnet-bearing lawsonite blueschist from SMM [PPL]. This sample contains a Gln + Lws + Omp + Grt + Phe + Rt assemblage, resembling the Type-II lawsonite eclogite of Tsujimori *et al.* (2006b). (j) The rim of a prograde-zoned garnet porphyroblast containing lawsonite in Type-I lawsonite eclogites from SMM (Tsujimori *et al.*, 2006b) [X-ray (Mg) image]. (k) Abundant lawsonite inclusions within garnet from Type-I lawsonite eclogites from SMM (Tsujimori *et al.*, 2006b) [XPL]. (l) Garnet porphyroblast containing folded inclusion trails of lawsonite from SMM (Tsujimori *et al.*, 2006b) [PPL]. (m) Euhedral lawsonite containing rutile in Type-II lawsonite eclogite of SMM (Tsujimori *et al.*, 2006b) [PPL]. (n) "False lawsonite eclogite" from New Idria, Franciscan Complex, California – eclogite overprinted by lawsonite blueschist facies mineral assemblage. Lawsonite occurs in fractures of garnet and in matrix; the secondary lawsonite and jadeite partially replace matrix eclogite facies omphacite (see Tsujimori *et al.*, 2007) [X-ray (Ca) image]. (o) Parallel growth of coarse-grained lawsonite exhibiting twining in a jadeite-bearing lawsonitite from SMM [XPL].

Calculated phase diagrams for hydrated MORB and hypothetical altered MORB (i.e., MORB+) are shown in Fig. 1a,b. The exclusion of Mn to simplify the system limits the *P–T* stability range of garnet-bearing mineral equilibria. In this simplified system, the garnet-in reaction in the lawsonite stability field at ~450–480 °C separates the lawsonite blueschist facies from the lawsonite eclogite facies. In the latter, a chlorite-consuming reaction to form garnet releases H_2O gradually, and the rock becomes denser than mantle peridotite (~3.4 g cm^{-3}) at ~560–580 °C. It is important that the changing H_2O contents of solid phases across the lawsonite–epidote transition are more substantial than that of the chlorite-consuming *P–T* region. Our modelling predicts that the stability of two-pyroxene pairs (Omp + Aug or Omp + Jd), which have been described in natural rocks (e.g., Tsujimori *et al.*, 2005, 2006b, 2007), are characteristic features of the transition between lawsonite blueschist and lawsonite eclogite. Although natural evidence casts doubt on the reliability of some thermodynamic modelling for L*T* metamorphic rocks – especially modelling with a fixed effective bulk-rock composition, fluid activity, oxidation state, etc. – the chemographic relations in *P–T*–X space are still helpful in order to understand natural metamorphic parageneses. Na- and Al-enrichment (i.e., increased jadeite component) during H*P* metamorphism is suggested by the presence of jadeite-, omphacite- and/or glaucophane-rich H*P* metasomatic veins in some L*T* eclogites (e.g., Bröcker & Keasling, 2006; John *et al.*, 2008); such metasomatic alteration processes likely are promoted by introduced slab fluids. In any case, our modelling shows that the Omp + Jd assemblage appears in metasomatized metabasalt (MORB+) rather than in MORB of normal composition.

Lawsonite geochronology

Although it is not a common target mineral for geochronology, Mulcahy *et al.* (2009) succeeded in dating lawsonite using the Lu-Hf system; they determined an age of 145.5 ± 2.4 Ma for a blueschist facies overprint after eclogite facies metamorphism in the Franciscan Complex. More recently, Vitale Brovarone & Herwartz (2013) reported a Lu–Hf lawsonite age of 37.6 ± 1.3 Ma for lawsonite blue schist facies metagabbro from the Alpine zone of Corsica. Gleissner *et al.* (2007) applied the Rb-Sr method for lawsonite pseudomorphs (Ms + Ep + Ab) in metabasalts from the Tauern Window of the Eastern Alps; they interpreted an internal isochron age of 29.82 ± 0.52 Ma as the time of decompression attending greenschist facies overprinting.

Classification of lawsonite eclogite types

Tsujimori *et al.* (2006a) proposed a classification of lawsonite eclogites based on assemblages of inclusions in prograde-zoned garnet; L-type grew only within the lawsonite stability field, and E-type records maximum temperature in the epidote stability field. Wei & Clarke (2011) extended their classification and added the additional U-type lawsonite eclogite that contains coesite and/or its pseudomorphs (Fig. 1a).

SPATIAL AND TEMPORAL DISTRIBUTIONS

Lawsonite eclogites: Updates

Since the Tsujimori *et al.* (2006a) review of 13 localities, we have become aware of at least six more lawsonite eclogite occurrences; it should be noted that localities with only textural evidence of lawsonite eclogite facies have been also included. The latest recognized include the following: North Qilian (Zhang & Meng, 2006; Wei *et al.*, 2009), Chinese Western Tianshan (Lü *et al.*, 2009, 2012; Tian & Wei, 2013), Atbashi Ridge, Kyrgyzstan (Shatsky & Usova, 1989), Xinxian, Western Dabie (Wei *et al.*, 2010), Monviso, Western Alps (Angiboust & Agard, 2010; Groppo & Castelli, 2010), and Kotsu/Bizan, Eastern Shikoku (Tsuchiya & Hirajima, 2013). Moreover Turkish (Sivrihisar) and Corsican lawsonite eclogites have been extensively studied (e.g., Davis & Whitney, 2006; Vitale Brovarone *et al.*, 2011). New Caledonian lawsonite eclogites were classified as E-type (Tsujimori *et al.*, 2006a), but a recent study by Vitale Brovarone & Agard (2013) inferred that the peak pressure of New Caledonian lawsonite eclogite was at a pressure value exceeding the lawsonite–epidote phase boundary. Lawsonite eclogites for each locality newly recognized since 2006 are described below.

North Qilian

The early Palaeozoic Pacific-type orogenic belt characterized by ophiolites, blueschists, calcalkaline volcanics and granitic plutons occupies the suture zone between the North China craton and the Qilian-Qaidam microcontinent; Qilian H*P* rocks consist of low-grade lawsonite blueschists and high-grade garnet-bearing epidote blueschists with lenses of glaucophane–epidote eclogites (e.g., Song *et al.*, 2007, 2009). Geochemical analysis indicates MORB and OIB compositions of protoliths for both blueschists and eclogites. E-type lawsonite eclogites were reported by Zhang & Meng (2006); lawsonite and its pseudomorphs occur as inclusions in garnet. *P–T* estimates yield peak eclogitic conditions of 2.2–2.6 GPa and 460–540 °C (Song *et al.*, 2007; Wei *et al.*, 2009). Zircon U-Pb and phengite Ar-Ar geochronology suggest *c.* 472 Ma for the eclogite facies metamorphism and 460–450 Ma for the blueschist facies recrystallization (Song *et al.*, 2009).

Chinese Western Tianshan

The late Palaeozoic Tianshan orogenic belt extends for ~1500 km from the Fan-Karategin belt of Tajikistan in the west via the Atbashi belt of Kyrgyzstan to the Chinese Western H*P*–UH*P* Tianshan in the east (e.g., Liou *et al*., 2009); the belt marks the suture zone between the Central Tianshan Arc and the Kazakhstan–Yili microcontinent, and continues to the Central Asian Orogenic Belt. Coesite-bearing UH*P* eclogites with lawsonite pseudomorphs occur in the Chinese Western Tianshan (e.g., Lü *et al*., 2009). Although lawsonite has not yet been confirmed, the presence of its pseudomorphs (Klemd *et al*., 2002) and *P*–*T* estimates suggest the original presence of U-type lawsonite eclogites (Wei & Clarke, 2011). *P*–*T* estimates for the eclogitic assemblage Grt + Omp + Lws + Gln yields peak metamorphic conditions of 2.9–3.0 GPa at 526–540 °C; post-temperature peak (2.4–2.7 GPa at 590 °C) isothermal decompression evidently caused the decomposition of lawsonite (Tian & Wei, 2013). Some UH*P* metapelites contain the assemblage Grt + Gln + Lws ± Jd ± Cld (Lü *et al*., 2012). Omphacite-bearing zircon yielded a U-Pb age of *c*. 320 Ma (Su *et al*., 2010), suggesting the time of the temperature peak or the later decompression stage (Tian & Wei, 2013). Rutile in the glaucophane–epidote eclogite gave a U-Pb age of 318 ± 7 Ma (Li *et al*., 2011).

Atbashi Ridge, Kyrgyzstan

The H*P*–UH*P* belt of Kyrgyzstan is a western extension of the Late Palaeozoic H*P*–UH*P* belts of the Chinese Western Tianshan. Tagiri *et al*. (1995) reported that eclogite from this locality contains quartz pseudomorphs after coesite in garnet and omphacite. Lawsonite inclusions and their pseudomorphs within garnet were described by Shatsky & Usova (1989); in addition, the lawsonite-bearing garnet contains glaucophane. Hegner *et al*. (2010) also confirmed the presence of lawsonite in the garnet cores. Eclogites with N-MORB and OIB compositions occur as lenses within garnet-bearing epidote blueschists with the mineral assemblages Tlc + Gln + Qz + Ph + Grt + Omp and Ep + Grt + Gln + Qz + Ph. The lawsonite-bearing garnet probably crystallized within the lawsonite stability field, thus is likely an L-type lawsonite eclogite. Moreover, the discovery of coesite pseudomorphs in the same metamorphic complex suggests the presence of U-type UH*P* lawsonite eclogite. The *P*–*T* estimate for the Atbashi Ridge eclogites suggests peak metamorphic conditions of 2.3–2.5 GPa at 510–570 °C, and phengite and glaucophane Ar-Ar geochronology yielded 327–324 Ma for the eclogite facies metamorphism (Simonov *et al*., 2008).

Xinxian, Western Dabie

Glaucophane- and kyanite-bearing UH*P* eclogites occur in the Xinxian UH*P* unit of the Western Dabie Mountains, a classic H*P*–UH*P* terrane metamorphosed during Triassic continental collision (cf., Liou *et al*., 2009). The Xinxian eclogite matrix contains rectangular-shaped lawsonite pseudomorphs consisting of Ep/Czo + Ky + Pg (Wei *et al*., 2010). The modelled peak-stage mineral assemblage of Grt + Omp + Lws + Tlc + Ph + Coe ± Gln ± Ky constrains attending *P*–*T* conditions to 605–613 °C and 2.8–3.3 GPa.

Monviso, Western Alps

Eclogite facies Tethyan oceanic lithosphere, the Monviso meta-ophiolite, occurs as a 35 km-long, 8 km-wide body structurally sandwiched between the underlying Dora-Maira Massif and the ocean-derived metasedimentary rocks-dominated blueschist facies units of the Piemonte Zone (cf., Lombardo *et al*., 2002). It is viewed as a classic area of Alpine subduction-zone metamorphism (e.g., Messiga *et al*., 1999). Recent studies have reported the occurrence of lawsonite eclogites (Angiboust & Agard, 2010; Groppo & Castelli, 2010). Jadeitites formed by metasomatic replacement of plagiogranite are associated with eclogite facies metaserpentinites (Compagnoni *et al*., 2012). At Monviso, preservation of lawsonite is rare, but pseudomorphs are ubiquitous. Garnet porphyroblasts of lawsonite eclogite (Fe-Ti metagabbro) contain inclusions of Omp (± Jd) + Lws + Chl + Qz in the cores and Omp + Chl + Tlc + Gln + Qz near the rims; lawsonite is not present in the matrix but its pseudomorphs are quite common (Groppo & Castelli, 2010). The peak metamorphic assemblage of basalt-derived eclogite is characterized by Omp + Grt + Ph ± Gln ± Lws pseudomorphs (Angiboust & Agard, 2010; Angiboust *et al*., 2012a). The Monviso lawsonite eclogite yields peak conditions of ~550 °C and 2.5–2.7 GPa (Groppo & Castelli, 2010; Angiboust *et al*., 2012a,b). Zircon U-Pb geochronology yielded an Eocene age of 45 ± 1 Ma for the metamorphism (Rubatto & Hermann, 2003).

Kotsu/Bizan, Eastern Shikoku (Japan)

Epidote blueschists with rare precursor epidote–glaucophane eclogites occur as a ~1.2 km-thick coherent unit of the Cretaceous Sanbagawa Belt in eastern Shikoku (Matsumoto *et al*., 2003). Recently, lawsonite inclusions in the cores and mantles of zoned garnet were reported from epidote–glaucophane eclogites by Tsuchiya & Hirajima (2013); the inclusion mineralogy of cores and mantles are Lws + Ab + Gln + Ep + Pg and Lws + Omp + Gln + Ep + Pg + Rt, respectively. This new E-type lawsonite eclogite yields lawsonite eclogite facies conditions of ~450–550 °C and 1.3–1.8 GPa (Tsuchiya & Hirajima, 2013).

Similar to many lawsonite eclogite localities, rectangular/rhombic-shape pseudomorphs consisting of mineral aggregates of Czo/Ep + Pg or Czo/Ep + Ky ± Pg constitute a reliable indicator of former lawsonite. Thus, eclogites from the Zermatt-Saas, Western Alps (Bearth, 1973; Oberhänsli, 1982; Barnicoat & Fry, 1986), the As Sifah, NE Oman (Searle *et al*., 1994), the Tauern Window (Hoschek, 2004) and the Ile de Groix (El Korh *et al*., 2009) also can be regarded as lawsonite eclogites (*sensu lato*); these eclogites generally show MORB-like affinities. Our review of the literature undoubtedly has overlooked yet other localities of such lawsonite eclogites.

Spatial and temporal distribution of H*P*–UH*P* glaucophane-bearing rocks

More than 190 blueschist (*sensu lato*) units/complexes have been re-evaluated for this study (Table S1). Our blueschist compilation includes localities of glaucophane (Na-amphibole) schists and/or glaucophane metabasalts and glaucophane-bearing eclogites. We have also included a few localities lacking glaucophane but containing lawsonite; eclogites lacking glaucophanitic amphibole were excluded (i.e., hornblende-bearing eclogites are not considered). The rare, oldest blueschist facies metamorphic rocks are of Neoproterozoic age (Ernst, 1972; Dobretsov *et al*., 1987; Maruyama *et al*., 1996; Brown, 2007) (Fig. 1). Although Ernst (1972) pointed out that lawsonite is uncommon in Palaeozoic blueschists, data accumulated during the last 40 years fail to reflect a distinct difference in frequency between Phanerozoic lawsonite and epidote blueschists (Fig. 3). This suggests that the petrological record of blueschists does not reflect a detectable secular change in thermal regimes of subduction zones since the end of the Proterozoic. However, our compilation found a global absence of lawsonite in rocks of Permian metamorphic age. We postulate that the global lawsonite hiatus is a robust indication of relatively warm subduction-zone thermal regimes at this time. This may be a consequence of the upwelling of sub-supercontinental mantle plumes that attended the breakup of Pangea, in the process generating large igneous provinces such as those of Siberia, Tarim, Emeishan and Central Europe.

The oldest lawsonite-bearing blueschists (560–550 Ma) have been reported from eastern Anglesey, Wales (UK) in the Anglesey–Lleyn accretionary orogen (e.g., Gibbons & Mann, 1983; Kawai *et al*., 2006, 2007). These blueschists formed within a Pacific-type subduction zone along the convergent margin of Avalonia in the latest Neoproterozoic (Kawai *et al*., 2007). At least two pulses of lawsonite blueschist formation are known from the Palaeozoic (Fig. 3). Early Palaeozoic lawsonite eclogites occur in Spitsbergen (Hirajima *et al*., 1988), the North Qilian (Zhang *et al*., 2007), and Eastern Australia (Och *et al*., 2003), as well as in circum-Pacific terranes. Middle Palaeozoic lawsonite blueschists (including those with lawsonite pseudomorphs) are widely distributed in the Variscan orogenic belts of Central

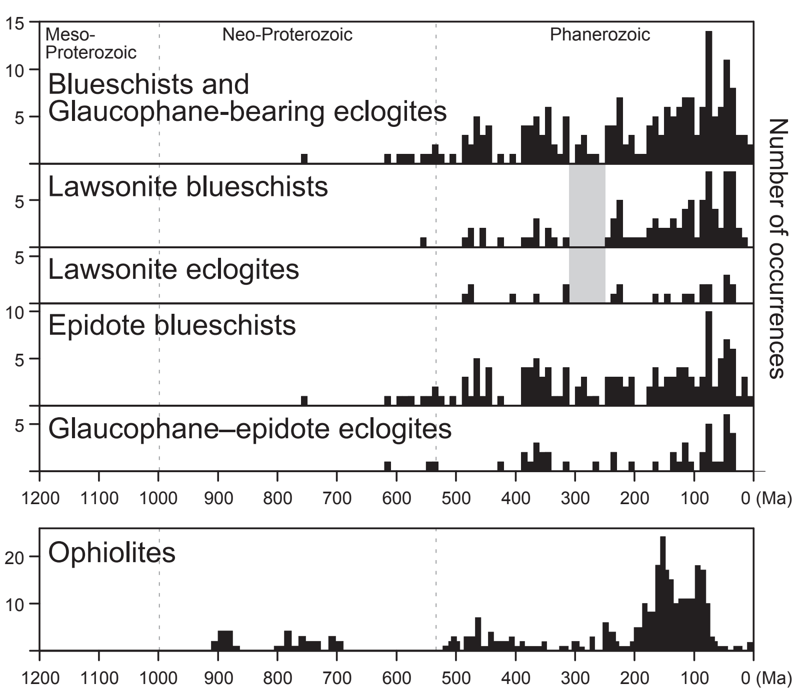


Fig. 3. Histograms showing ages of blueschist and glaucophane-bearing eclogite from the NeoProterozoic to the present. The data include >190 H*P*–UH*P* glaucophane- or lawsonite-bearing metamorphic units and/or domains. Data sources were newly compiled from the literature; see Table S1. Histograms are subgrouped into lawsonite blueschist, lawsonite eclogite (including eclogite with lawsonite pseudomorph), epidote blueschist, and glaucophane–epidote eclogite. Grey area represents the global lawsonite hiatus. For comparison, age distribution of ophiolites from Dilek (2003) are also shown.

Europe, the Urals and the Tianshan, and from circum-Pacific orogenic belts (the Russian Far East, SW Japan and Eastern Australia). Although lawsonite is not preserved in the Variscan and Ural localities, its pseudomorphs are common (e.g., Maksyutov Complex: Schulte & Sindern, 2002; Bean & Liou, 2005). Lawsonite eclogites occur in the South and West Tianshan of the China and Kirgizstan (Shatsky & Usova, 1989); lawsonite eclogites from the Chinese Tianshan reached UH*P* condition inasmuch as they contain coesite.

A distinct peak of lawsonite blueschists typifies the Triassic (Fig. 3). This pulse includes Pacific-type H*P* terranes of the circum-Pacific orogenic belts (North American Cordilleran and Japan) and the continental collision-type UH*P* terranes of east-central China; lawsonite from Chinese collision-type UH*P* terranes occur exclusively as pseudomorphs in kyanite-bearing UH*P* eclogites (Castelli *et al*., 1998). Triassic lawsonite eclogite has been described from British Columbia (Ghent & Erdmer, 2011). Abundant Jurassic–Cretaceous lawsonite blueschists are widely distributed in circum-Pacific orogenic belts (North American Cordilleran, the Andes, the Russian Far East, Japan), the Caribbean margins, the Tethyan orogenic belt, and the South Shetland Islands. Lawsonite eclogites occur in the Cretaceous Franciscan Complex (Shibakusa & Maekawa, 1997), the Sanbagawa Belt (Tsuchiya & Hirajima, 2013), the Caribbean (Guatemala and the Dominican Republic: Tsujimori *et al*., 2006b; Zack *et al*., 2004), and in Turkey (Davis & Whitney, 2006; Okay *et al*., 2006). Eocene and younger lawsonite blueschists extend discontinuously from the Alpine–Himalayan chain, through Indonesia + New Caledonia to New Zealand. Lawsonite eclogites are present in the Alps (Corsica: Vitale Brovarone *et al*., 2011; Monviso: Groppo & Castelli, 2010), Sulawesi Island, Indonesia (Parkinson *et al*., 1998) and New Caledonia (Clarke *et al*., 1997). Lawsonite blueschists dredged from a serpentinite seamount in the Mariana forearc include micaceous pumpellyite-bearing clasts that yielded a phengite K-Ar age of *c*. 48 Ma, coeval with subduction initiation of the Pacific plate beneath the Philippine Sea plate along the Mariana trench (Maekawa *et al*., 1993, 2002). The youngest lawsonite-bearing rock might be from an uplifted segment of the active Cascadia margin; its age of the metamorphism is inferred to be *c*. 7–12 Ma based on apatite fission-track dating (Brandon & Calderwood, 1990).

DISCUSSION

Natural record of lawsonite eclogitization

Modelled *P–T* paths for slab surfaces, both warm (SW Japan/Nankai) and cool (NE Japan/Tohoku) subduction zones (Syracuse *et al*., 2010), suggest that normal subduction can promote the prograde metamorphic transformation of oceanic crust from lawsonite blueschist facies to lawsonite eclogite facies assemblages (Fig. 1). How does the downgoing oceanic crust transform to lawsonite eclogite in a subduction zone? Numerous observations have led to a general consensus on the nature of the blueschist facies to eclogite facies boundary (e.g., Oberhänsli, 1982; Reinsch, 1979; Ridley, 1984; Schliestedt, 1986; Vitale Brovarone & Agard, 2013; Vitale Brovarone *et al*., 2011). Due to the wide transitional *P–T* zone, high-grade blueschists can contain the eclogite facies mineral assemblage Grt + Omp + Qz, whereas low-temperature eclogites can include glaucophane as an eclogite facies mineral. Oh & Liou (1998) suggested that the reaction Gln + Ep = Grt + Omp + Pg + Qz + H_2O serves as the *P–T* boundary between those two facies. However, this reaction is divariant in NCFMASH only if Fe and Mg are independent components of the system (Molina & Poli, 1998). Moreover, the disappearance of Gln + Ep does not take place in the lawsonite stability field. Recent thermodynamic modelling for the MORB composition in the system NCKMnFMASHO (+ Ms + Coe/Qz + H_2O) predicts that no specific discontinuous reaction constrains the blueschist facies to eclogite facies transition in the lawsonite stability field (Wei & Clarke, 2011); both omphacite and garnet can coexist over a wide range of lawsonite blueschist facies equilibria. In calculated phase diagrams, most phase changes in the lawsonite stability field in basaltic rocks are continuous, and dehydration is not restricted to the inferred metamorphic facies transitions (e.g., Hacker *et al*., 2003; Wei & Clarke, 2011).

At the regional scale, metamorphic field mapping along a relict prograde metamorphic gradient in NW New Caledonia suggests a transition from lawsonite blueschist to lawsonite eclogite at ~500–520 °C and *c*.1.8 GPa (Vitale Brovarone & Agard, 2013). Although a remarkable pressure gap (0.6 GPa) defines the tectonic break between eclogite facies meta-ophiolite and the eclogite facies metamorphosed accretionary wedge, the prograde blueschist-to-eclogite transition is gradual. A continuous metamorphic field gradient from lawsonite blueschist to lawsonite eclogite is also preserved in the coherent blueschist sheet at Ward Creek, Franciscan Complex (Maruyama & Liou, 1988). The first appearance of garnet (i.e., lawsonite eclogite facies mineral assemblage) in this blueschist unit was estimated as >~290 °C and ~0.8–0.9 GPa. However, inasmuch as relatively low-*T* reactions during early eclogitization are sluggish in subducting oceanic crust, transition from lawsonite blueschist to lawsonite eclogite is not necessarily gradual, but may be a function of pressure overstepping.

Progressive eclogitization within the lawsonite stability field is also sporadically preserved in the

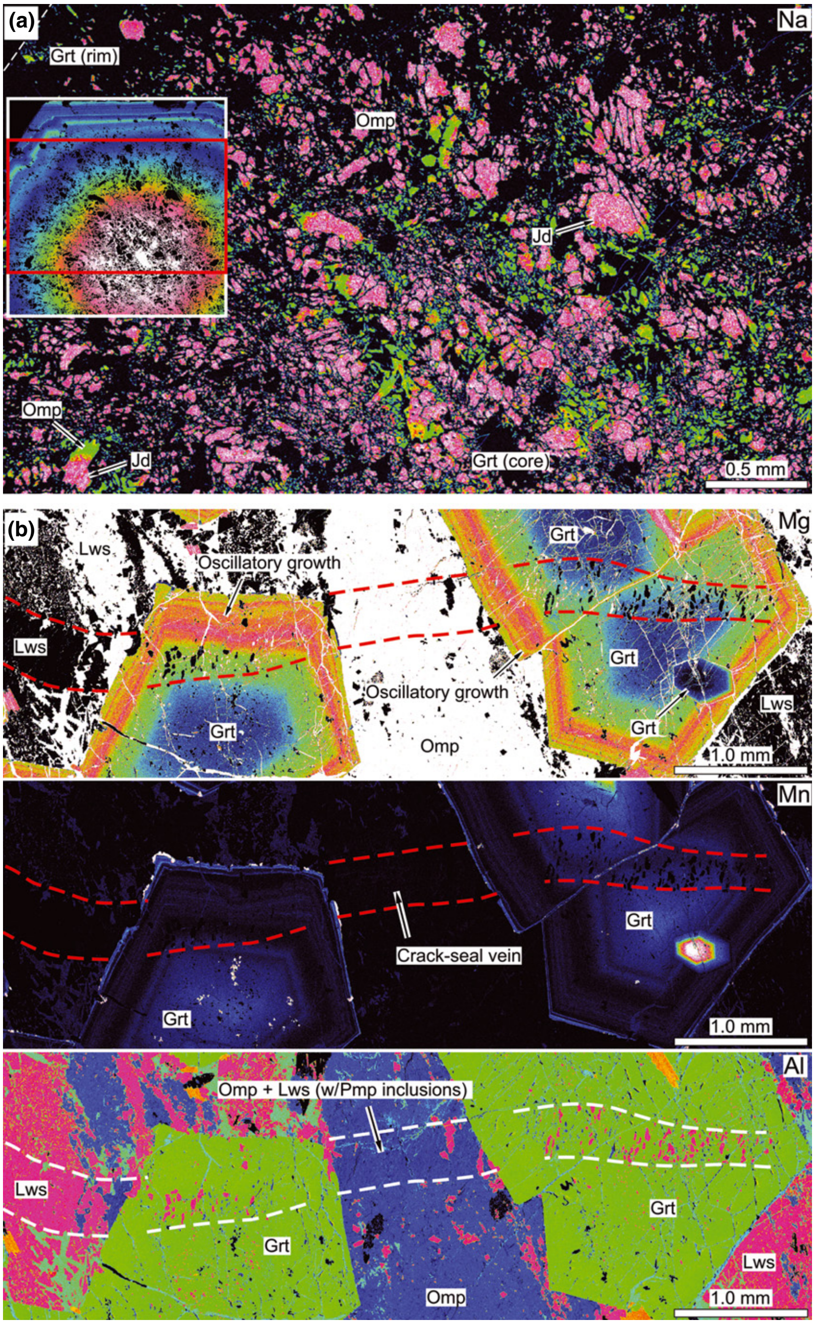


Fig. 4. Brittle deformation features recognized as microtextures in natural lawsonite eclogites. (a) X-ray image (Na) showing pyroxene inclusions within a prograde-zoned garnet in a jadeite-bearing lawsonite eclogite (South Motagua Mélange, Guatemala; Tsujimori *et al.*, 2006b). The textures suggest crystal growth of garnet along brittle fractures in the pyroxene-rich matrix during eclogitization. An index map in the figure is of Mn. (b) X-ray images (Mg, Mn and Al) of the same area of Type-I lawsonite eclogite (Tsujimori *et al.*, 2006b) showing crack-seal omphacite + lawsonite veins crosscutting the matrix and garnet during garnet growth. The three images also show the merging growth of garnet as well as inheritance of a garnet seed crystal.

zoned garnet (e.g., Tsujimori *et al.*, 2006b; Groppo & Castelli, 2010). In lawsonite eclogite from the South Motagua Mélange, lawsonite inclusions within garnet contain rare pumpellyite; its preservation (pumpellyite-(Al) *sensu stricto*) and the presence of synmetamorphic brittle deformation suggest that eclogitization was initiated at ~300 °C (Tsujimori *et al.*, 2006b).

Figure 4 shows two examples of synmetamorphic brittle deformations preserved in the South Motagua lawsonite eclogite. Pre-eclogite facies brittle fracturing is evident by shapes and distribution of irregular-shaped inclusions within a prograde-zoned garnet porphyroblast (Fig. 4a). Garnet growth was initiated along brittle fractures of the jadeitic pyroxene-rich matrix; it is also clear that garnet enlargement strongly fractionates elements in the system, and fossilizes significant volumes of precursor phases + textures during incipient eclogitization. Syn-eclogite facies crack-seal veins also are traced into euhedral garnet as bands (Fig. 4b); fluid-induced oscillatory zoning and abnormal grain growth is apparent in garnet. All these microtextures (Fig. 4) infer not only brittle behaviour in the blueschist-to-eclogite transformation, but also reflect incomplete recrystallization, anisotropic mineral growth, and mass transfer associated with fluid infiltration over a relatively large scale during eclogite facies metamorphism. Such natural features evoke questions regarding the elegant results of thermodynamic modelling for L*T* eclogites assuming simple interfacial equilibrium in a fixed effective bulk-rock composition.

In addition to the above examples, lawsonite eclogite from the Monviso meta-ophiolite contains precursor glaucophane and coexisting Omp + Jd within garnet cores, suggesting a metamorphic condition of < 300 °C and < 1.4 GPa for the pre-eclogite stage (Groppo & Castelli, 2010). Synmetamorphic brittle deformation has been also described in lawsonite eclogite facies metagabbros. Angiboust *et al*. (2012b) described Monviso eclogite breccias cemented by interclast matrix with Grt + Omp + Lws (pseudomorphs); crack-seal omphacite veins and garnet fractures provide additional evidence for polyphase fracturing-healing events during lawsonite eclogite facies metamorphism.

Considering these observations, it is apparent that brittle fracturing and fluid infiltration are common phenomenon in the subducting oceanic crust during prograde lawsonite eclogite facies metamorphism. Further comprehensive study of synmetamorphic brittle behaviour of lawsonite eclogites should lead to a new understanding of intra-slab seismicity at ~30–80 km depths in the forearc region (e.g., Angiboust *et al*., 2012b).

Multiple fluid-infiltration records in lawsonite

Flow of slab-derived fluids enriched in Si, Al, Na and Ca enhances metamorphic + metasomatic reactions and precipitation of H*P* minerals.In some cases, intra-slab flow causes extensive elemental leaching (cf., Spandler & Pirard, 2013). How does lawsonite record subduction-zone fluid flow? An apparent example of lawsonite preserving evidence of multiple fluid infiltration is shown in Fig. 5. A peculiar metasomatic rock consisting of ~70–90 vol% lawsonite occurs as tectonic inclusions within the South Motagua Mélange (Tsujimori *et al*., 2006c; Fig. 5a). The rock consists principally of fluid precipitates and represents the extreme of lithological variation in "P-type" jadeitite that was formed by precipitation directly from a fluid (Tsujimori & Harlow, 2012; Flores *et al*., 2013). Randomly oriented lawsonite crystals up to 5 cm long show sector- and oscillatory-growth zoning on a millimetre to submicrometre scale (Fig. 5b–d); Sr and LREE are enriched in the {010} sector. X-ray images

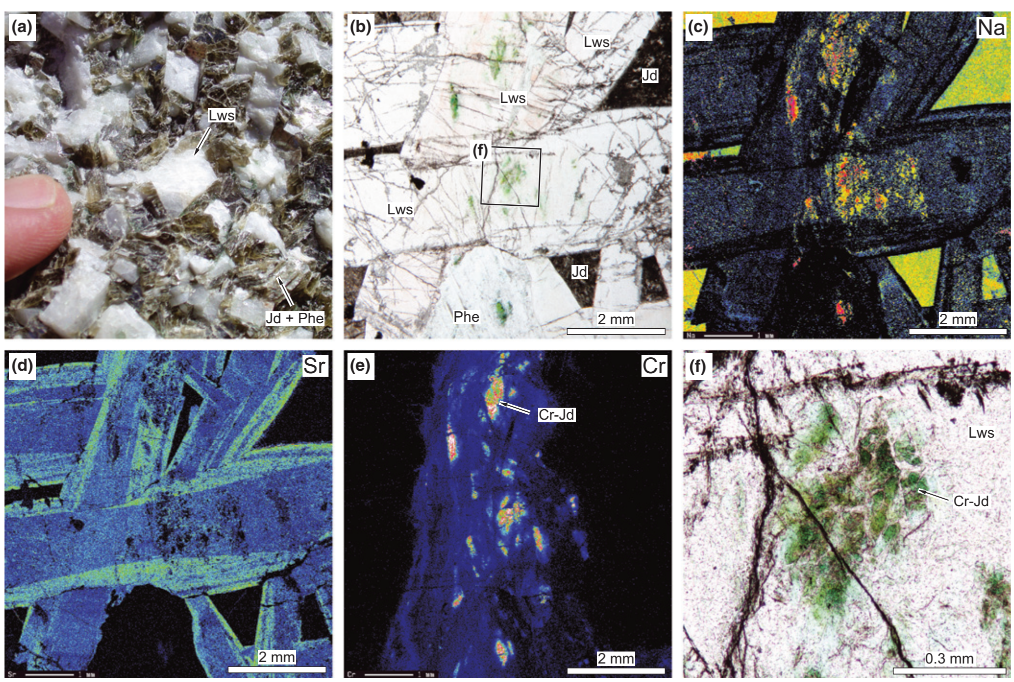


Fig. 5. Lawsonite-rich metasomatic rock preserving a fluid-path way in the high-pressure serpentinite mélange along a subduction-zone channel. (a) Euhedral coarse-grained lawsonite crystals cemented by interstitial phengite and jadeite. (b) Plane-polarized light (PPL) view of lawsonite intergrowth filled with fine-grained aggregates of jadeite and phengite. Cr-bearing pyroxene and lawsonite are green and pale pink, respectively. See the Cr distribution in (e). (c) X-ray (Na) image of (b) showing jadeititic pyroxene within lawsonite crystals and interstitial pockets. (d) X-ray (Sr) image of (e) showing enrichment of Sr and oscillatory zoning at the rims of lawsonite crystals. (e) X-ray (Sr) image of (b) showing a Cr-rich fluid path. (f) Cr-rich jadeite within lawsonite in the boxed area of (b) [PPL].

show millimetre-scale flow paths (<4 mm wide) of Cr-rich fluids crosscutting lawsonite crystals; these textures indicate that fluid infiltration occurred after lawsonite crystallization (Fig. 5e). The fluid flow paths follow the track of Cr-bearing pinkish lawsonite (~0.6 wt% Cr_2O_3) with inclusions of Cr-rich (kosmochloric) jadeite (up to ~22 wt% Cr_2O_3, $ko_{25–63}$) and Cr-rich phengite (up to ~6 wt% Cr_2O_3) (Fig. 5b,f). It is noteworthy that the rock lacks relict Cr-rich minerals, such as chromite and uvarovite, which can act as an internal source of Cr. These microtextures and mineral compositions indicate a chronological sequence of fluid-infiltration events: lawsonite grains together with jadeitic pyroxene precipitated from subduction-zone aqueous fluids veining the overlying mantle wedge, and subsequently a Cr-rich fluid infiltrated the lawsonitite, probably along microfractures, and crystallized Cr-rich jadeite. Lawsonite acted as a sink of Sr and LREE (e.g., Spandler & Pirard, 2013). Moreover, although generally regarded as relatively immobile during metamorphism, petrological observations indicate heightened mobility of Cr in subduction-zone metasomatic fluids. Except for fluid-inclusion studies, direct tracking of a subduction-zone aqueous phase is not simple. However, lawsonite-bearing H*P* metasomatic rocks have a great potential for decoding the origin and history of such fluids.

Lawsonite blueschist facies overprinting during decompression

Blueschist facies overprinting within the lawsonite stability field has been reported from high-grade H*P* metamorphic rocks of some Pacific-type blueschist belts, such as the Franciscan Complex, and the South Motagua Mélange (e.g., Wakabayashi, 1990; Krogh *et al*., 1994; Tsujimori *et al*., 2006b). These parageneses indicate that the rocks were refrigerated and hydrated during exhumation. Such H*P* rocks commonly occur as tectonic blocks within serpentinite-matrix tectonic mélange, and their *P–T* trajectories are characterized by either hairpin-like or counterclockwise paths. For Pacific-type convergent margins, some tectonic slices of the subducting oceanic crust recrystallized to eclogite facies evidently return surfaceward along the subduction channel, propelled by buoyancy of the enveloping serpentinite (e.g., Ernst, 1970; Guillot *et al*., 2000). These slices ascend to blueschist facies levels and appear to have been sequestered at shallower depths for a considerable time interval. The question arises, how long are the exhuming eclogite facies slices retained at blueschist facies conditions? Recent geochronology of Franciscan eclogites revealed a *c*. 7 Ma gap between eclogite facies metamorphism (153 Ma) and lawsonite blueschist facies overprint (146 Ma) (Anczkiewicz *et al*., 2004; Mulcahy *et al*., 2009). Zircon U-Pb ages of lawsonite eclogite xenoliths from the Colorado Plateau diatremes range from 81 to 33 Ma and reveal that these eclogite fragments have been stored in a slab–wedge-mantle interface for *c*. 50 Ma (Usui *et al*., 2003). The Eocene age of blueschist clasts dredged from an active serpentinite diaper in the Mariana forearc suggests that these blueschist clasts remained for *c*. 48 Ma in a serpentinite mélange under the forearc region (Maekawa *et al*., 2002).

Considering these facts, exhumed eclogite tectonic blocks within a H*P* serpentinite mélange can remain at depth for a lengthy time, and consequently can have undergone significant later hydration/recrystallization (Fig. 6). Significant cooling and continuous H_2O supply from the dehydrating oceanic slab to the exhuming H*P* serpentinite mélange enhance the interaction between H*P* fragments and host serpentinite; moreover, they may cause lawsonite blueschist facies overprinting and consequently prevent breakdown of lawsonite during its decompression. In the case of L-type lawsonite eclogites, preservation of prograde and peak lawsonite-bearing mineral assemblages can be retained by rapid exhumation (Whitney & Davis, 2006). Even if the lawsonite eclogites had a long residence time under lawsonite blueschist facies conditions, fluid infiltration without heating and significant deformation allow partial preservation of the precursor lawsonite-bearing mineral assemblage.

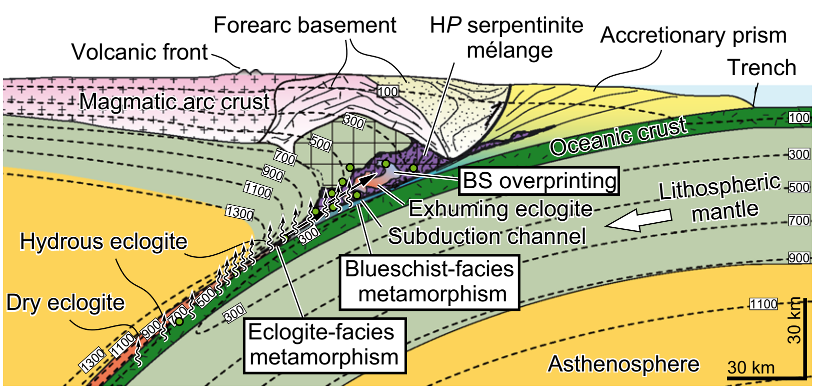


Fig. 6. Cross-section showing a Phanerozoic Pacific-type subduction zone, where lawsonite-bearing blueschists and eclogites form. Lithological variations and thermal structures are modified after numerical modelling by Gerya (2011). High-pressure (H*P*) serpentinite mélange, location of exhuming H*P*-rocks, and environment of jadeitite formation are based on Stern *et al*. (2013).

PERSPECTIVES

Recent recognition of intense material recycling of oceanic/continental crust through deep subduction, mantle upwelling and exhumation to the Earth's surface (cf., Liou & Tsujimori, 2013) requires the reassessment of various geological processes in Pacific-type convergent margins where H*P*–L*T* rocks form. The subduction zones of such margins promote large-scale mass circulation into the deep mantle, and have done so since plate tectonics began. In fact, Usui *et al.* (2003) provided conclusive evidence of coesite-bearing lawsonite eclogite xenoliths in the Colorado Plateau that were derived from the subducted Farallon plate. Hence understanding the extent of eclogitization of subducting oceanic crust within the lawsonite stability field is of crucial importance.

The dynamic thermal regimes recorded in lawsonite blueschists and eclogites are in accord with numerical models of typical subduction zones (Fig. 1). Subduction zone *P*–*T* conditions control H_2O-carrying capacity of the downgoing oceanic slabs. Numerical modelling demonstrates that cold slabs can retain H_2O beyond sub-arc depths (e.g., Hacker, 2008). Although the total amount of H_2O in lawsonite and phengite in eclogitized oceanic crust is less than in fully serpentinized slab peridotites, lawsonite-bearing crust-derived metamorphic rocks have more potential to retain and carry trace elements, such as K, Rb, Cs, U, Pb and Sr, and isotope signatures to beyond sub-arc depths (e.g., Vitale Brovarone *et al.*, 2014). Numerical models also allow the inference that eclogites that return are either sampled contemporaneously from the subduction channel or were transported from greater depths in the developing H*P* mélange (Gerya *et al.*, 2002). Marschall & Schumacher (2012) proposed a model whereby dehydration of diapiric upwelling of H*P* mélange (a mixture of crustal-derived metamorphic rocks, slab fluids-related metasomatic lithologies, and serpentinized wedge-mantle peridotites) at the slab–mantle interface would cause fluid-induced partial melting and consequently would control the geochemical signatures of arc volcanics. Experimental data of trace element partition coefficients for lawsonite show a preference for LREE/HREE and Be; consequently the observed B/Be decrease in Phanerozoic arc magmas with increasing distance from the trench can be explained by lawsonite breakdown (Martin *et al.*, 2011). After decomposition, the remaining trace elements inherited from the precursor lawsonite in some UH*P* phases may return to the Earth's surface via a deep mantle plume, or may be accidentally trapped as peculiar UH*P* phases (such as LREE-rich CaTiSi-perovskites inclusions within Group-2 Brazilian diamonds: Bulanova *et al.*, 2010) in diamonds and/or chromitites in the mantle transition zone (Liou & Tsujimori, 2013).

As documented above, the recent recognition of lawsonite eclogite localities has increased significantly. However, complete preservation of L-type eclogites that return to the surface is rare. Integrated approaches continue to provide better answers for some old but persistent questions, such as: Why are lawsonite eclogites rare? What is a suitable tectonic setting for their preservation? Systematic and precise age data for subduction, prograde blueschist facies and eclogite facies metamorphism and blueschist facies overprint (if refrigerant conditions are maintained on exhumation) together with precise *P*–*T* estimates are essential in order to realistically describe ancient Pacific-type plate convergent margins.

The blueschist-to-eclogite transition within the lawsonite stability field for oceanic crust seems to be strongly controlled by reaction kinetics. Although the concept of metamorphic facies does not take rates into account, in order to evaluate eclogitization and related solid-earth processes, quantitative understanding of the reaction kinetics of eclogite facies mineral growth, reassessment of grain-scale stresses, dynamics of diffusion-limited mineral growth, dynamic domain equilibrium during fluid infiltration, and elemental fractionation and chemical transportation during metamorphic differentiation + mineral growth are crucial topics requiring further research. In any case, the classical and neoclassical multidisciplinary approaches continue to provide opportunities to link H*P*–L*T* metamorphism of ancient subduction zones to geophysical observations of modern analogues, to evaluate hydration and dehydration along the subduction channels, and subsequent slab–mantle interaction.

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SUPPORTING INFORMATION

Additional Supporting Information may be found in the online version of this article at the publisher's web site:

Table S1. Blueschist (H*P*–UH*P* glaucophane-bearing rock) compilation for Fig. 3.

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