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Deep fluid release in warm subduction zones from a breached slab seal

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Abstract

Petrological models and seismic data from subduction zones with geotherms of 7 K km^{-1} or higher suggest that slabs in these systems dehydrate effectively in the forearc. A large fluid flux is nevertheless released from these slabs at and beyond subarc depth, suggesting that large amounts of H_2O can remain slab-bound to much greater depth than expected. We propose that this is due to a transient sealing effect exerted by the subducting lower crust. To test this concept, the petrological and geochemical evolution of such gabbroic crust is investigated through a textural, petrological and Li-chronometric analysis of eclogitized gabbros from an exhumed ultrahigh-pressure terrane. The samples record pristine transitions from dry, rigid gabbro to hydrated eclogite and eclogite mylonite, which occurred when these rocks resided at 90-110 km depth. The observations characterize step-by-step the deformation and overstepped mineral reactions that following the influx of external fluids along a developing network of permeable shear zones. Lithium chronometry indicates that the gabbroic rocks were breached and permeated within a few months at a very specific depth within the 90-110 km interval—depths where, in warm subduction zones, large fluid-filled channel system emanate from the slab. The data support a model in which fluids produced in the deserpentinizing slab mantle are trapped at very high pore pressure beneath the slab Moho and are ultimately released at subarc depth where the lower crust fails and develops highly permeable fluid vents. The subducting lower crust thus may play an important role in regulating H_2O and element budgets, and controlling slab rheology in warm subduction zones.

Keywords: Eclogite, warm subduction zones, arc magmatism, gabbro-to-eclogite transition, earthquakes

1. Introduction

Subduction zones with an average trench-to-arc geotherm of 7 K km^{-1} or higher are common among active margins (England et al., 2004; Syracuse et al., 2010). These typically show low thermal parameters and rates of descent (Syracuse et al., 2010), and, with only few exceptions, consume oceanic lithosphere from intermediate- and fast-spreading centers (e.g., Izanagi-Pacific Ridge, the East Pacific Rise, Juan de Fuca Ridge). Arc lavas produced in these ‘warm’ subduction zones do not differ chemically from those from ‘normal’ subduction zones; all commonly show a distinct ‘slab signature’ (e.g., high Ba, Rb, Sr, Pb, U) that signifies contributions of slab-derived fluids on mantle melting (Grove at el., 2012; Spandler and Pirard, 2013). These fluids are released via dehydration reactions in the hydrated oceanic upper crust and mantle, which can be seismogenic and can raise pore pressure in the slab to near-lithostatic (Hacker et al., 2003; John et al., 2009; Audet et al., 2009; Peacock et al., 2011; Taetz et al., 2018). The fluids are released continuously throughout the forearc and advect upward through self-organizing channel networks and are ultimately released to the slab-mantle interface where they flux the plate interface and—upon breaching the plate interface—migrate into the mantle hanging wall (Hacker et al., 2003; Audet et al., 2009; Peacock et al., 2011; John et al., 2012; Plümper et al., 2017). Magnetotelluric sounding in warm subduction zones reveals the pathways of these fluids through the mantle wedge and into the hanging-wall crust, implicating them in upper-plate seismicity (Wannamaker et al., 2009). Magmatic arcs are fed by fluids that tap from a slab segment just beyond subarc depth and migrate towards the arc source via confined, (near-)vertical or trenchward-oriented advective transport channels (Zhao, 2001; Wannamaker et al., 2009; Gerya et al., 2006; McGary et al., 2014). The δD, $\delta^{18}O$ and trace-element signatures of arc lavas indicate that these fluids are, at least in part, sourced in the oceanic mantle (Singer et al., 2007; Kodolányi et al., 2012; Walowski et al., 2015; Marshall et al., 2017; Spandler and Pirard, 2013). The role of mantle-sourced H_2O in arc magmatism is indirectly indicated by the occurrence of magmatic gaps in subduction zones where, instead of normal oceanic crust, oceanic plateaus containing H_2O-poor mantle are subducted (e.g., the Denali Volcanic Gap in Alaska; Chuang et al., 2017).

Although slabs appear to generate large amounts of H_2O beneath arcs, it is not clear why this is the case. Unlike cold slabs, which still release large amounts of H_2O at and beyond subarc depth (Rüpke et al., 2004; van Keken et al., 2011), warm slabs are predicted to already dehydrate within the forearc (Abers et al., 2017). This is consistent with observations from the seismic record; slab-hosted earthquakes, which are attributed to transformational faulting and dehydration embrittlement caused by hydrous mineral breakdown or the ensuing channelized fluid flow (Peacock, 2001; Hacker et al., 2003; Incel et al., 2017), are frequent within the forearc, yet gradually die out towards subarc depth, occasionally terminating in an earthquake cluster (Fig. 1; England et al., 2003; Armijo et al., 2010; Nakajima et al., 2013; Chuang et al., 2017). A mechanism for deeper retention and release of H_2O may be sought in chlorite within the oceanic mantle, which is stable to greater depth than serpentine minerals and may thus be able to cause re-hydration and flux melting of the eclogitic upper crust beyond the stability of such minerals (Walowski et al., 2015). This process, however, should not only occur at subarc depth and via chlorite breakdown, and should thus trigger flux melting in a far less discrete domain than is observed; serpentine dehydration occurs throughout the forearc and slab-top temperatures are likely to already far exceed those of the wet-basalt solidus in most of this domain (Syracuse et al., 2010).

Models invoking transient low-permeability seals within subducting slabs may provide a reliable explanation for the apparent ability of fluids to remain slab-bound to greater depth than expected. One such seal is inferred to occur in the low velocity zone at the slab-top at shallow depths in Cascadia, where fluids released in the subducting upper crust are trapped at (near-)lithostatic pressure between the underlying low-porosity gabbros and an impermeable plate boundary (Audet et al., 2009; Peacock et al., 2011). Hydrofracturing from critical tension in the reacting upper crust at 40-50 km depth causes the interface to become permeable, permitting the release of the H_2O trapped within the dehydrating crust to the plate interface and forearc mantle (e.g., Taetz et al., 2018). A similar, yet far longer-lived permeability barrier may occur in the gabbroic lower oceanic crust. This crustal section caps the largest H_2O reservoir of subducting slabs, the serpentinized oceanic mantle (Rüpke et al., 2004), and is orders of magnitude less permeable than rocks of that reservoir (Katayama et al., 2012). Within the subduction hanging wall, this permeability contrast between gabbro and serpentinite may impede the escape of H_2O into the overlying crust, leaving large amounts of H_2O trapped at near-lithostatic pore pressure within the corner of the mantle wedge (Katayama et al., 2012). This same effect may also occur within the subducting slab. Field and teleseismic observations from gabbroic rocks buried along a geotherm of 7 K km^{-1} or higher show that such rocks may persist to much greater depth than equilibrium thermodynamic would predict; the transition of gabbro to eclogite is extremely fluid-limited and remains overstepped until the lower crust undergoes extreme, often seismogenic deformation that results in the infiltration of fluids and the occurrence of far-overstepped mineral reactions (Austrheim, 1987; Yuan et al., 2000; Lund & Austrheim, 2003; John & Schenk, 2003; Hacker et al., 2003; John et al., 2009; Angiboust et al., 2011; Putnis et al., 2017). This behaviour may be particularly pronounced in warm subduction zones, which consume slabs from intermediate- to fast-spreading ridges, which are likely to contain a thick, coherent gabbro section. The geochemical, petrological and geophysical observations from warm subduction zones can be synthesized into a new hypothesis, which is proposed and explored here: the gabbroic lower crust represents a metastable permeability barrier that traps H_2O at high pore pressure in the dehydrating slab mantle below. This barrier persists to approximate subarc depth where the lower crust fails and re-equilibrates along fluid percolation networks that allow the trapped H_2O to escape, traverse the slab, and migrate to the arc source through the imaged advective transport channels. Testing of this ‘deep-seal hypothesis’ requires further characterization of how these rocks deform and react during subduction to sub-arc depth. In addition, the specific depths and durations of these processes need to be investigated.

Obtaining the necessary insight requires data pertaining to the prograde-to-peak history from subducted gabbroic rocks. However, for oceanic crust exhumed from subduction zones such data may be difficult to obtain; fragments of such crust are rare and typically reworked. As such, we investigate a possible analog in the form of gabbros from a subducted hyper-extended continental margin. Such margins subduct along geotherms that are, at least when reaching eclogite-facies conditions, similar to those observed in warm subduction zones. They are thicker, on average less mafic, and possibly less pervasively hydrated than typical oceanic crust. The evolution of the gabbroic rocks in their deeper section may nevertheless, on a body scale, be similar to that of gabbros in proper oceanic crust. This is because the presence of fluids is the main control on their reaction is (Austrheim, 1987; John & Schenk, 2003; Putnis et al., 2017) and dehydration reactions in the intermediate to felsic host rocks of these gabbros are expected to provide these throughout most of the burial history. The fluids are sourced in different rocks than the fluids in oceanic slabs, and may thus interact slightly differently with a given rock. Still, partially eclogitized gabbro bodies enclosed in compositionally different rock matrices are, at least to a first order, similar in terms of their petrological, textural and geochemical features (cf. Mørk et al., 1985; John & Schenk, 2003; Lund & Austrheim, 2003; Terry & Heidelbach, 2006; Angiboust et al., 2014; Putnis et al., 2017, and references therein). One major advantage is that gabbros in exhumed hyper-extended margins may more effectively escape overprinting and hence may preserve a better record of high- and ultrahigh-pressure (*HP*, *UHP*) processes. This is because such margins remain attached to their continental crust and may, upon detachment of the oceanic slab, exhume rapidly and coherently under the influence of the extreme buoyancy of that crust. The remarkable preservation of oceanic-crust-like (*U*)*HP* rocks in hyper-extended margins is exemplified by the meta-gabbros exposed in the Western Gneiss Complex (WGC), Norway—the hyper-extended margin of former Baltica (e.g., Mørk et al., 1985; Terry & Robinson, 2004; John et al., 2009). To test various components of the deep-seal hypothesis, gabbroic rocks from this terrane were subjected to a multi-method analysis, involving field and textural characterization, major- and trace-element analysis, and Li concentration and isotope analysis, and Li chronometry.

2. Geological background and sample description

The analyzed gabbroic rocks are exposed in the northernmost *UHP* domain of the WGC—a large tectonic composite of granitic and metasedimentary gneisses with enclaves of mafic and ultramafic rocks, all of Proterozoic protolith, which represent the hyper-extended margin of the Baltic Shield (Krogh et al., 2011; Andersen et al., 2012). The WGC collided with, and was partially subducted beneath, the Laurentia during the Scandian stage of the Caledonian Orogeny (425-400 Ma). Burial of the terrane caused widespread amphibolite- to eclogite-facies metamorphism of the WGC basement, as well as the many infolded thrust complexes, and led to *UHP* conditions in various parts of the WGC along the Norwegian west coast. Most basaltic rocks were pervasively eclogitized and repeatedly re-equilibrated during burial (Terry & Robinson, 2004; Krogh et al., 2011; Cutts and Smit, 2018). In contrast, gabbroic rocks exhibit only a limited degree of eclogite-facies reaction and deformation (e.g., Mørk et al., 1985; Lund & Austrheim, 2003; John et al., 2009). Both rock types occur in the same reactive, hydrous and, in part, migmatitic gneiss matrix, indicating that the poor reactivity of gabbros is not due to the absence of fluids, but rather due to the inability of these fluids to percolate these bodies and cause widespread re-equilibration. Due to their limited reaction, the gabbroic rocks preserve a variety of lithologies from pristine, anhydrous and rigid gabbro (*G*) and corona gabbro (*CG*) to hydrous eclogite (*E*), and from undeformed, to transitional and mylonitic eclogite (*UE*, *TE*, *ME*). The stranded textural and petrological transformations between these uniquely capture snapshots of the (micro-)tectonic processes and reactions that govern the re-equilibration of lower-crustal rocks during deep subduction.

In this study, we focus on (meta-)gabbro bodies occurring in close geographical proximity in the same unit of the Nordøyane-Moldefjord *UHP* domain—notably the Flem

Gabbro on northwest Flemsøya, the Drynasund Gabbro on westernmost Midøya, and the Haram Gabbro on central-west Haramsøya (further descriptions in Terry & Robinson, 2004; Terry and Heidelbach, 2006; Krogh et al., 2011). The bodies were partially or completely transformed to eclogite during burial at peak conditions of c. 820°C and 3.0-3.6 GPa (Terry & Heidelbach, 2006; Krogh et al., 2011). Collectively and individually, the three bodies provide the full range of gabbroic rocks, from *G* to *ME*. They are nevertheless different in the degree to which they were transformed. The Haram and Drynasund Gabbros still largely comprise (*C*)*G* and only reacted locally to eclogite in and around shear zones, which make up less than 10% of outcrops. In contrast, the Flem Gabbro largely comprises sheared eclogite and is dominated by eclogite mineral assemblages that equilibrated at peak conditions—assemblages that are rare in the other bodies; the Flem Gabbro clearly represents a texturally and petrologically more progressed form of the other bodies (see also Terry & Heidelbach, 2006).

Brittle structures were observed only in (*C*)*G*. These structure are relict micro-faults filled with extremely fine-grained dark eclogite-facies material, which are typically interpreted to represent eclogite-facies pseudotachylite (Fig. 2a; see also Lund & Austrheim, 2003; Terry and Heidelbach, 2006; John et al., 2009). Characteristically, the (*C*)*G* around such structures is entirely intact and shows no reaction to eclogite. Brittle structures are very rare; the large majority of shear zones reflect ductile simple shear and some degree of wall-rock reaction (Fig. 2b). Termini were observed on some smaller shear zones in this network, as well as on small shear zones that appear isolated. These show that the shear zones end by localizing into small-offset faults, which themselves die out in an array of fanning cracks. Crosscutting relationships between shear zones were not observed; the shear zones are interconnected in large networks that exceed outcrop scale and connect by converging continuously into single shear zones (see also

Terry & Heidelbach, 2006). The shear zones thus can be considered as representing coeval shear during a single stage of deformation. In various places of the Drynasund Gabbro, compositional layering is still preserved. Using this as a marker, it is clear that the (*C*)*G*-dominated blocks between the shear zones are chaotically rotated relative to one another, giving the outcrops an overall cataclastic lay-out.

Only the wall-rock of the ductile shear zones reflects eclogitization (Fig. 2). The (*C*)*G*-*UE* transformation in the environs of the ductile shear zones was largely pseudomorphic and the reacted rocks still preserve the ophitic texture of the protolith (Fig. 2, 3). The (*C*)*G* precursor around the reacted zone is bound from *UE* by a narrow transition zones of a few centimetres thick. The occurrence of olivine relicts in *UE* provides a reliable field indicator of the proximity of this transition. Towards *ME*, the eclogitized rock volume comprises texturally transitional eclogite *TE* with gradually increasing drag (Fig. 2c-e, 3). The omphacite grains in *TE* provide a strain marker; in spite of these grains being dragged into the shear zone, their height on average is approximately the same as that of omphacite grains in *UE*. This indicates that deformation largely, if not entirely, involved simple shear. In well-developed shear zones within the Flem Gabbro, *ME* developed foliation-parallel dynamic banding defined by cm-thick layers of pyroxenite and garnetite. Such banding envelopes the individual sub-bodies that make up the outcrop (Fig. 2b), indicating that these represent fragments of a larger parental body that was dissected and disaggregated by the *ME* shear zones. Although common in the other bodies, large (meter-sized) low-strain enclaves are rare in the Flem Gabbro. One such enclave was nevertheless encountered on the holm Seiholmen (N 62°41’33.57”, E 06°14’40.58”; Fig. 2e). The enclave exhibits the continuous transition from *ME* to *UE* in its margin (Fig. 3) and even exhibits relict olivine in its core. The deformed margin captures the progression of deformation and reaction in its most advanced stage and thus was sampled for the purpose of this study.

3. Methods

A large (c. 18 kg) sample of the margin of the Seiholmen low-strain enclave was taken (Fig. 3) and sub-samples were subjected to detailed textural analysis. For chemical analysis, pieces of each sub-sample were cut using a diamond-studded saw, and sandblasted and rinsed several times with deionized water to remove any material introduced by sawing. The cleaned fragments were pulverized using an agate mortar and pestle. Aliquots of these powders were digested by fusion using a $LiBO_3$ flux and subjected to major element analysis by X-ray fluorescence spectroscopy. Separate 100-mg aliquots of the powders were subjected to trace-element analysis. Major- and trace-element compositions are provided in Fig. 4 and in Supplementary Table S1, whereas the [Li] data are provided in Fig. 5 and Table 1. A separate set of aliquots from the powders were subjected to Li isotope analysis. The Li isotope data are provided in Fig. 5 and Table 1. Detailed descriptions of methods for these analyses, including dissolution routines and data for reference materials, are provided in the Supplementary Materials. Lithium diffusion chronometry was applied to estimate the duration of ductile shearing. The length scale of diffusive Li zoning d_{Li} was estimated by iteratively regressing error functions obeying Fick's law in 1D through the [Li] and δ^7Li profiles until the coefficient of determination (R^2) was maximized. Estimates of d and its uncertainty were obtained using least-squares fitting as applied by Smit et al. (2016).

4. Results

The transformation from *CG* to *UE* in the analyzed sample involved the replacement of primary plagioclase, forsterite, and augite phenocrysts by aggregates of garnet ± kyanite (after plagioclase), enstatite + omphacite (after forsterite), and omphacite (after augite) containing cleavage-parallel rods of garnet and rutile and grains of phlogopite (Fig. 3; mineral compositions for *CG* and *UE* reported by Mørk, 1985). Although ophitic textures are largely preserved in *UE*, the eclogitic grains within this relict texture are much smaller than their coarse igneous precursors, typically by at least an order of magnitude (3c, d). Olivine occurs in the outermost sample. The replacement of forsterite by enstatite was pseudomorphic and interface-coupled, producing fine-grained enstatite aggregates with palisade-like textures (Fig. 3d). The omphacite replacements after augite aggregates show abundant tensile cracks and veins filled with common eclogitic minerals, as well as phlogopite, dolomite and sulfides; in omphacite, these veins are ghosts healed by the ingrowth of omphacite that, unlike its omphacite host, is free of garnet and rutile rods (Fig. 3e-g).

The textural modification from *UE* to *ME* occurs progressively in narrow zones of continuously increasing shear. The fabrics in this zone indicate that strain was accommodated through grain- and sub-grain rotation, bending and rupturing (omphacite), grain boundary sliding (garnet), and grain growth (phlogopite). Deformed omphacite grains or grain domains show undulose extinction and sub-grain formation, are 'cleansed of' garnet rods similar to the healed veins, and exhibit a relatively high abundance of phlogopite inclusions (Fig. 3e). These grains typically show an aspect ratio of 2 and are aligned in the mylonitic foliation (Fig. 3b, h). Although omphacite grains have a higher aspect ratio with increasing strain, the number of omphacite nuclei—that is, grains and individual clusters of small grains that represent a single clinopyroxene precursor—per given surface area remains similar. This is consistent with observations made from the outcrop-scale that these shear zones largely accommodated simple shear. Ultramylonitic eclogite comprise alternations of garnet- and omphacite-rich layers (Fig. 3i). These layers are near-equigranular, especially in the shear zone core, which can be clearly identified as the part of the *ME* domain where all grains have become fully rotated into parallelism. The grain size becomes increasingly more homogeneous towards the shear zone core. Separate grain size populations that could indicate shear zone reactivation were not observed, which is consistent with the inference from outcrop-scale observations that the rocks represent a single continuous stage of shear. Locally within *ME*, omphacite has coarsened by grain boundary migration, and garnet has grown larger and more facetted towards omphacite, indicating minor post-kinematic grain growth. The modal abundance of hydrous phases (phlogopite and minor phengite), which is already higher in *UE* relative to (*C*)*G*, increases across *TE* and is highest in *ME*, which contains up to 10 vol% phlogopite.

The sub-samples of the Seiholmen transitional gabbro-eclogite rock reveal various compositional trends. Towards the core of the mylonitic shear zone, the sub-samples are broadly enriched in SiO_2 and Na_2O, and depleted in Al_2O_3 and Fe_2O_3; these changes are irregular rather than smooth and diffusive. Along the same vector and following a similarly irregular pattern, concentrations of light Rare Earth Elements (LREE) increase, as do concentrations of Th, Cu and fluid-mobile, large-ion lithophile elements such as Cs, Sr, Ba and Pb, and Th/U values; the concentrations of heavy REE, Zn, Nb, Sb, and W decrease towards the shear zone core.
The sub-samples contain 2-6 ppm Li and show relatively high δ^7Li values between +10 and +20‰ (Fig. 5). The concentration of Li increases and δ^7Li decreases gradually and continuously towards the core of the shear zone.

5. Discussion

5.1 Re-equilibration along dynamic fluid vents

The extremely fluid-limited and pseudomorphic nature of the reaction of gabbro to eclogite is reflected in the many occurrences of partially eclogitized gabbros in terranes exhumed from subduction zones worldwide (Putnis & John, 2010, and references therein). The series of evolutionary steps that allow this transformation to occur is clearly recorded in the rocks analyzed here as well and is summarized below together with references to studies that report similar findings from other well-studied gabbro bodies: 1) During most of its burial, the gabbroic protolith remained largely metastable and transformed only locally by corona reactions that were far too sluggish to allow significant transformation (e.g, Mørk et al., 1985; Austrheim, 1987; Putnis et al., 2017). Fluids circulating in the hydrous matrix of the gabbro bodies were unable to permeate and mediate re-equilibration throughout this stage. The *CG* relicts still reflect this original state (e.g., Fig. 2a). 2) At 90-110 km depth, the gabbros underwent failed and developed brittle faults (e.g., Lund & Austrheim, 2003; John et al., 2009). The faults linked up to form permeable networks of anastomosing shear zones that fed reactive fluids to the cracked wall rock, which caused static eclogitization of the wall-rock around shear zones (Austrheim, 1987). The migration of the reactive fluid front was accelerated by inter- and intra-granular tensile cracking and formation of hydrated and carbonated veins (Fig. 3e-g; John & Schenk, 2003)—a phenomenon that may reflect local tension from the 10-15% volume loss that is incurred upon eclogitization (e.g., Hacker et al., 2003). The reactions were largely pseudomorphic (Fig. 3b, c; e.g., John & Schenk, 2003; Angiboust et al., 2014) and proceeded by local fluid rock interaction along dissolution-reprecipitation fronts emanating from these cracks (Putnis & John, 2010). 4) Widening of the shear zones into the statically eclogitized rocks, as evidenced by the recrystallization of eclogite-facies minerals (e.g., Fig. 3b, f), indicates reaction-induced weakening. This could be ascribed to grain size reduction of the (*C*)*G-UE* transition and resulting enhancement of grain boundary sliding and diffusion creep kinetics, as well as the introduction of weak and aligned micas, and possible shear heating from episodes of thermal runaway within the shear zones (John et al., 2009; Fusseis et al., 2009). The metasomatic effect of the syn-tectonic fluids within the sheared *ME* and *TE* rock volume and the surrounding *UE* is reflected in the enrichment in LREE and fluid-mobile elements (Fig. 4; see also John & Schenk, 2003), and the increasing abundance and alignment of mica towards the shear zone (e.g., Fig. 3i). 5) Static grain growth occurred only locally and only in the most strained sub-samples. The drastic reduction in diffusion distances upon the cessation of deformation is ascribed to the cessation of syn-tectonic porous flow (Terry and Heidelbach, 2006). The latter further illustrates that mineral reactions and deformation occurred as long as pore fluids were present in the eclogitized rock volume and stagnated upon their escape or sequestration in hydrous minerals.

5.2 Duration and depth of gabbro-to-eclogite transitions

A testable prediction from the deep-seal hypothesis is that the gabbro-to-eclogite transition and the emergence of metasomatic vents within the lower crust is punctuated and localized at a very specific slab interval that lies at, or slightly beyond, sub-arc depth. The *P-T* conditions estimated for the eclogite-facies assemblages in the analyzed rocks do not provide a valid test for this concept. They are uncertain and technically only represent the final eclogite-facies re-equilibration, i.e., they provide a *maximum* constraint on the conditions at which the rocks underwent reaction and strain. On the basis of these data alone, it cannot be excluded that these processes occurred over a large depth interval that starts much shallower than the hypothesis predicts. To test this, we investigate the depth interval of reaction through the duration of reaction in the Seiholmen shear zone sample, which represents the furthest state of progression in the transformation of (*C*)*G* to *ME*. The deep-seal hypothesis would be disproved if the duration of this transformation would be on the order of a million of years—the time it would take at an average subduction angle of 30° and convergence rate of 3 cm yr^{-1} to bury the subducted slab by c. 20 km. *Vice versa* the hypothesis would be supported if durations were much shorter; short durations would support the inferred pulsed nature of lower crustal breaching and would place this process at the depths indicated by the peak *P-T* conditions. Reaction timescales may be smaller than what can be resolved by conventional radiometric dating. Lithium diffusion chronometry provides a powerful alternative to constraining reaction timescales and is particularly efficacious in the case of fluid-mediated mass transfer on very small timescales (e.g., John et al., 2012; Taetz et al. 2018).

The Li zoning towards the shear zone is clearly smooth and non-linear (Fig. 5), as is typically observed for diffusion-limited Li transport in rocks mediating porous flow (John et al., 2012; Taetz et al., 2018). The high Li concentrations and low δ^7Li values (c. +10‰) of the shear zone core characterize the externally derived Li that entered the system during deformation and syn-tectonic fluid flow. Strikingly, a substantial domain of the rock away from the shear zone was subjected to porous flow and transformed to hydrous eclogite, yet did not yet develop zoning for Li concentration and δ^7Li as a result of chemical interaction with the shear zone. The Li signature of this domain actually still appears to reflect that of the *CG* protolith; it is identical to that observed for the lower continental crust (Teng et al., 2008; Penniston-Dorland et al., 2017) that *CG* would represent, and is the same regardless of whether sub-samples contain *CG* relicts. The presence of this domain crucially shows that the migration of the reaction front outpaced the diffusive invasion of externally derived Li. This implies that 1) the system in which Li exchange occurred can be considered as comprising two fluid-filled porous system: the shear zone that provided externally derived Li and the hydrous eclogite; 2) the transport of Li within this system was not limited by the factors that controlled reaction rates—i.e., the rate of fluid advection into the dry *CG* protolith, dissolution kinetics and element diffusion to and from the reactive interface—but rather by Li diffusion in the pore fluid. The apparent inability of the externally derived Li to reach the reaction front at any given time implies that the development of diffusive Li zoning was not influenced by the location of this front relative to the Li diffusion penetration depth d_{Li}, or the degree to which this may have changed with time. As such, the eclogitized rock volume can, for Li and for the duration of its exchange, be considered a semi-infinite reservoir. For such specific systems, the chemical and isotope zoning of Li can be treated and modelled in terms of bulk diffusion between two fluid-filled reservoirs (John et al., 2012). The model of John et al. (2012) allows estimates of the duration of this process from diffusion zoning profiles via calculations that take into account changes in partitioning behaviour and porosity during fluid-rock interaction. Non-dimensional time for the exchange process (Ω) can be obtained from fitting modelled diffusion profiles at a given reaction-induced over background porosity ($\varphi_{\mathrm{R}}/\varphi_0$) to the data. Differences in modelled profiles for $\varphi_{\mathrm{R}}/\varphi_0$ values between 1 and 1000 are insignificant at the analytical uncertainty of the data obtained here; $\varphi_{\mathrm{R}}/\varphi_0$ was hence set at a value of 10, which is in the range of values determined for metamorphic rocks so far (John et al., 2012; Taetz et al., 2018). Using these values, we obtained identical values of $(7.1 \pm 2.7)\cdot 10^{-4}$ (δ^7Li) and $(9.8 \pm 3.2)\cdot 10^{-4}$ ([Li]) for Ω. These values can be scaled to estimate the duration of fluid-rock interaction using estimates of d_{Li}, the diffusivity of Li in aqueous fluid (D_{Li}^{f}), and the bulk solid-fluid Li partition coefficient ($K_{\mathrm{d,Li}}^{ecl\text{-}f}$) and background porosity (φ_0^{ecl}) of eclogite—the background lithology of the system in which the modelled diffusion process occurred. The duration of diffusive Li exchange that the model would provide in this case (t_{Li}) would represent the time during which dynamic fluid-filled interconnected porosity existed across the system and sustained diffusive exchange of Li between the hydrous eclogite wall rock and the external fluids fluxing the shear zone.

One of the model assumptions is that the reactive interface between the two Li reservoirs undergoing exchange—in this case the shear zone core—remained stationary during the exchange process. This assumption appears to be warranted in spite of intense strain. Deformation largely, if not entirely, involved simple shear; the reactive interface—the shear zone core as defined by textural analysis—thus appears to have remained fixed in the direction of fluid flow. Another assumption is that the mineral content and *P-T* conditions remained fixed during exchange, such that D_{Li}^{f} and $K_{d,Li}^{ecl\text{-}f}$ can be considered constant within their limits of uncertainty. The main change in mineral content is in the modal abundance of micas. Its effects on $K_{d,Li}^{ecl\text{-}f}$, however, should be well within the large extrapolation uncertainty. Changes in ambient *P-T* during fluid-rock interaction may be considered if t_{Li} proves to be on the order of millions of years. In contrast, shear heating occurs on much shorter time scales and is likewise not accounted for. This effect may cause underestimation of D_{Li}^{f} and, by extension, overestimation of t_{Li} (c. 10% for 200°C shear heating). Further overestimation may come from the possibility of post-tectonic solid-state Li diffusion. The effects of this may not be substantial, considering that intra-crystalline diffusive transport of Li in minerals such as clinopyroxene (10^{-15}-10^{-16} at given conditions; Coogan et al., 2005) is at least 7 orders of magnitude slower than diffusion of Li in fluids at given *P-T* conditions (10^{-8} m^2s^{-1}; e.g., John et al., 2012) and kinetic barriers to inter-crystalline diffusion would suppress transport further. Time differences of similar magnitude may nevertheless exist in duration between reaction and subsequent residence at high temperature. The effects from solid-state diffusion thus can not be *a priori* excluded, suggesting that t_{Li} estimates may represent maximum values.

Solving Fick's Law in one dimension for d_{Li} yielded identical estimates of $0.11_{-0.03}^{+0.07}$ (δ^7Li) and $0.13_{-0.04}^{+0.11}$ m ([Li]). A quantitative estimate of $(1.98 \pm 0.28) \cdot 10^{-8}$ m^2s^{-1} was obtained for D_{Li}^{f} by extrapolating the results of laboratory experiments to the *P*-*T* conditions of the investigated reaction (c. 820°C and 3.0-3.6 GPa; following John et al., 2012; Taetz et al., 2018; and references therein). Estimates of $K_{\mathrm{d,Li}}$ between rocks and fluids are typically estimated on the basis of $K_{\mathrm{d,Li}}$ and modal abundances of given mineral constituents, extrapolated to the appropriate temperature. A recent experimental study, done at temperatures (800°C) and MORB compositions that represent the system investigated here (Rustioni et al., 2019), removed the need for this approach in this case. The discrete $K_{\mathrm{d,Li}}^{ecl\text{-}f}$ estimates obtained from these experiments ($n = 8$) yielded a weighted mean of 1.19 ± 0.25, which is used here. Estimates of φ_0^{ecl} are uncertain and not yet reliably made via (ultra)high-pressure experiments. A laboratory value of $3 \cdot 10^{-3}$ may be applied to natural rock systems (John et al., 2012). This value is similar to $(4.9 \pm 0.1) \cdot 10^{-3}$, which is what would be obtained at given pressures for blueschist with similar grain size as the eclogite analyzed here using $\varphi_0(P)$ as determined via experiments (Taetz et al., 2018). We conservatively apply the laboratory value, recognizing that φ_0^{ecl} may have effectively been higher.

Using the given values and propagating the uncertainties assuming they are entirely uncorrelated, yields identical estimates of $7.1_{-3.9}^{+4.3}$ days (δ^7Li) and $13.6_{-5.4}^{+6.0}$ days ([Li]) for t_{Li}. This provides further strong evidence that fluid release in slabs occurs during extremely short pulses of canalized fluid flow (John et al., 2012; Plümper et al., 2017; Taetz et al., 2018; and references therein). The uncertainties on the time estimate are large, as *d* estimates are uncertain, and extrapolation and *P*-*T* uncertainties are set conservatively large compared to other studies applying Li chronometry. The true uncertainty may nevertheless be larger still. Excess uncertainty may be expected in φ^{ecl}, for which uncertainties may be substantial and are difficult to realistically quantify (e.g., Taetz et al., 2018), and in D_{Li}^{f}, for which uncertainties only account for analytical and extrapolation uncertainties from the experimental data, not for disparities between the experimental and natural system. Possible underestimation of φ^{ecl} and D_{Li}^{f} add to the possibility that t_{Li} was actually shorter than is estimated here. At the same time, t_{Li} estimates may underestimate the total deformation history, as it does not account for any delays in the supply of Li via the shear zone. All aspects considered, it can be reasonably concluded that shearing, and by extension the associated reactions, occurred on very short timescales of days, perhaps months, and can be considered essentially instantaneous on the timescale of plate burial. The reaction of these rocks to weak and permeable eclogite thus would have been localized at a very specific depth in the 90-100 km depth interval that the *P*-*T* estimates indicate.

5.3 Lower crust as a potential transient permeability barrier in slabs

The observations made in this study provide important insight into the behaviour of subducted gabbros and lower oceanic crust in warm subduction zones. Conform to the expectations from the deep-seal hypothesis (Fig. 6), such crust can persist as a metastable and impermeable layer down depths beyond 80 km and then rapidly develop a permeable shear zone network that acts as a long-range metasomatic vent system. Through its unique behaviour, the lower crust may exert a strong control on regulating fluid transfer in warm subduction zones, especially considering that this crust caps the slab mantle where most of the slab-bound H_2O is stored. This H_2O is progressively released via serpentine breakdown which commences from c. 50 km depth onward along the upward-migrating 600°C isotherm (Peacock, 2001; Rüpke et al., 2003; Peacock et al., 2011; Walowski et al., 2015; Plümper et al., 2017). The aqueous fluids that are produced in the oceanic mantle throughout this domain are non-stationary and rapidly migrate towards the slab Moho by channelized flow in self-organizing vein networks (Plümper et al., 2017), possibly undergoing several cycles of sequestration in, and release from, transient serpentine minerals. The observations made in this study indicates that these fluids—neither those that are far-travelled, nor those produced at equilibrium in the forearc where the 600°C geotherm reaches the Moho—will not be able to advect beyond the Moho; they are expected to become trapped at high pore pressure until the slab reaches approximate subarc depth and the lower crust is finally breached (Fig. 6). Considering the large amount of H_2O that may have accumulated below the Moho at this point, it is possible that this breach represents the largest fluid pulse from warm subducting slabs since the removal of their plate-interface seal. The large feeder channels of the subarc mantle in warm subduction zones, which tap from the slab at the same approximate depth (Zhao, 2001; McGary et al., 2014), may thus represent extremely fluid-rich cold plumes that emanate from sites where the slab mantle can finally be drained.

Warm slabs show various changes in slab properties and behaviour approximately at subarc depth and this spatial correlation may be re-evaluated through the concepts presented here. For instance, slabs typically exhibit earthquake clusters at such depth, which are localized in the lower crust and oceanic mantle, and are associated with normal faulting, possibly due to declined coupling and a stronger effects of slab pull (e.g., Nankai, Alaska, Cascadia, Andes; Yuan et al., 2000; England et al., 2004; Nakajima et al., 2013; Chuang et al., 2017; Bloch et al., 2018). The failure and transformation of the lower crust are most likely part of these processes and likewise explain why Wadati-Benioff earthquakes and a resolvable slab Moho do not extend beyond such depth (Yuan et al., 2001; Rondenay et al., 2008; Bostock, 2013). Beyond these seismic clusters, slabs typically exhibit an increase in dip (England et al., 2004; Klemd et al., 2011; Nakajima et al., 2013), suggesting these are no longer able to withstand flexural bending. This is typically attributed to phase changes in upper crust and mantle (e.g., Bloch et al., 2018). However, it may alternatively be suggested that this effect is due to the reaction-induced weakening and disaggregation of the lower crust, which would deprive the slab of the last component that was still relatively competent. The failure and delayed transformation of the lower crust in this depth interval could be tied in with each of these phenomena and the deep-seal hypothesis would uniquely explain why arc fronts developed at approximately the same trench-ward distance. Further investigation of the slab drainage that is proposed to occur in these parts of warm subduction zones requires the analysis of the slab rocks that were exposed to this process. Examples of these may be found among lower-crustal relicts of exhumed ophiolites, such as the Monviso Ophiolite in the western Alps. The lower crust of this ophiolite locally underwent eclogite-facies co-seismic rupturing, brecciation and fluid-induced reaction, and these processes ultimately led to the focussed release of about 90% of the 12 wt% of H_2O stored in the underlying serpentinized mantle (Angiboust et al., 2011; Spandler et al., 2011; Angiboust et al., 2014). Future research on such occurrences may further characterize the mechanisms and rates of fluid flow at these sites, and so develop the concepts presented here.

6. Conclusions

The data presented in this study characterize the unique role of the lower crust in slabs subducting in warm subduction zones and demonstrates how this crust allows warm slabs to retain H_2O to much greater depth than expected on the basis of equilibrium calculations. Instead of reacting at equilibrium, the lower crust persists metastably as a transient permeability barrier throughout the forearc, trapping H_2O released by serpentine breakdown reactions in the mantle below. A highly effective feedback among deformation, fluid flow and reaction ultimately operates to breach this barrier. This process occurs essentially instantaneously in a depth interval that corresponds to common subarc depths and is ultimately triggered by lower-crustal failure, the expression of which may be observed in seismic records from various subduction systems. The permeable shear zone networks that develop in the wake of failure allow the rapid focused escape of the deeply stored H_2O to the subarc mantle. This drainage can explain why warm slabs, which are expected to be dehydrated at subarc depth, are able to still supply large amounts of fluid to the advective channels of the subarc mantle. Besides possibly controlling the elemental flux of arcs, the rapid breaching and transformation of the lower crust may play a role in the rheological and physical changes that occur in slabs beyond subarc depth. Combined geochemical and geophysical observations may allow further testing for deep-sealing effects for slabs with higher descent rates, subduction angles and thermal parameters, and for slabs that were produced along slow-spreading centers.

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Figures

Fig. 1: Earthquake distributions in various warm subduction zones (data from Hacker et al., 2003, and references therein; earthquakes in the hanging wall not shown). Seismicity is largely restricted to the forearc and declines towards the subarc, in some cases culminating in earthquake clusters.

Fig. 2: Field observations of meta-gabbro occurrences. a) Fractures containing fine-grained black fracture infill in meta-stable corona gabbro at Drynasund (*CG*;). b) Zone of banded and largely mylonitic eclogite on the margins of a pervasively eclogitized meta-gabbro body in gneiss country rock at Flemsøy (*CR*). c) Meta-gabbro (*CG*) with minor reaction exhibiting a dextral *ME* shear zone with a small drag zone comprising texturally transitional eclogite (Drynasund). d) Mylonitic eclogite in a partly eclogitized (*C*)*G* of the Haram gabbro. The reaction is strongest close to the shear zone. The shear zone anastomozes and minor branches can be seen in the lower part. e) The low-strain *UE* enclave in *ME*-dominated meta-gabbro on Flemsøy, which was analyzed in this study.

Fig. 3: Photos (a, b), crossed-polarizer photomicrographs (d, f-i) and backscatter-electron images (c, e) of the large sample of the Seiholmen transitional eclogite occurrence (sampled from outcrop in Fig. 2e), showing the transitions from undeformed eclogite with minor relicts of coronitic gabbro (*UE-CG*; d), to undeformed eclogite (c, e, g), texturally transitional eclogite (*TE*; f, h), and mylonitic eclogite (*ME*; g). The textural transition from static to deformed rock can be traced through the loss of the ophitic texture of the protolith (b). (c) Fine-grained enstatite

(*en*) with omphacite (*omp*) rims after forsterite. (d) Forsterite (*fo*) relic in a partially eclogitized meta-gabbro. The replacing enstatite shows palisade-like textures. (e) Vein inside *omp* composed of *omp* and other phases such as dolomite (*dol*). The veins are free of rutile (*rt*) and garnet rods, which are common in the omphacite host. (f) Strained omphacite showing undulose extinction. This distorted region shows a high abundance of phlogopite inclusions and is, just like the vein here and in (e), free of garnet rods. (g) omphacite with garnet and rutile lamellae after augite in garnet corona. Vein networks filled with eclogitic phases can be observed. (h) omphacite porphyroclast with fine-grained *omp* strain tails enclosed in garnet ribbons. (i) Dynamic layering (omphacite -rich versus garnet -rich) from the core of a mylonite zone.

Fig. 4: Trace-element data for the six sub-samples of the transitional eclogite shown in figure 3. The sub-samples show increasing concentrations in LREE and fluid-mobile elements (e.g., Sr, Cs, Ba, and Pb), and decreasing concentrations of HREE towards the shear zone core.

Fig. 5: Lithium concentration and isotope profiles across the slab shown in figure 3. The distance *x* represents that distance to the core of the shear zone as defined in the field. The data represent the sub-samples as shown in Fig. 3. The outer sub-sample is only partially eclogitized and contains olivine and augite relicts of the *CG* protolith.

Fig. 6: Schematic diagram illustrating the deep-seal hypothesis for warm subduction zones. The release of H_2O (arrows) occurs in the two hydrated parts of the slab, the oceanic crust and the serpentinized mantle. a) H_2O released from the upper crust (light blue) feeds the plate interface and the shallow mantle wedge after breaching the plate-interface seal. H_2O produced in mantle

sections above 600°C (dark blue) is mostly consumed by serpentinization upon ascent. b) The upper crust is largely dehydrated. H_2O released from the dehydrating mantle migrates upward and ends up trapped below the dry and impermeable lower crust. c) Critical tension from an increasingly intense slab pull force causes the dry metastable lower crust to fail and develop fluid-filled shear zones. These compromise lower-crustal integrity and act as pathway for trapped H_2O to be released to the subarc mantle. Schematic thermal structure is based on a thermo-mechanical model for the Alaska Range following the assumption of isoviscous rheology (Abers et al., 2006).

Table 1: Lithium concentration and isotope data for the sub-samples of the slab shown in Figure 3.