Invited Research Article

Trace element and oxygen isotope study of eclogites and associated rocks from the Münchberg Massif (Germany) with implications on the protolith origin and fluid-rock interactions

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A B S T R A C T

Eclogites, metagabbros, and paragneisses from the Variscan Münchberg Massif record a complex succession of igneous, hydrothermal and metamorphic processes. The geodynamic setting related to the protolith formation and the impact of different types of fluid-rock interactions have been uncertain up to now. We use major and trace element chemistry as well as oxygen isotopes to disentangle the geochemical signatures related to the different stages of the rocks' history.

In the Münchberg Massif, dark eclogites (kyanite-free; Fe-Ti-MORB signature) are distinguished from light eclogites (kyanite-bearing; higher Mg#, Al₂O₃, and Cr; lower incompatible element contents; positive Eu anomalies; MORB to arc basalt signature). The δ¹⁸O values for both types (+5.0 to +10.8‰) are equal to, or higher than those of MORB. Amphibolite facies metagabbros have a more enriched, almost OIB-like trace element signature and high δ¹⁸O values (+9.4 to +10.3‰).

Good linear correlations between fluid-immobile elements throughout the eclogite types confirm their derivation from a common, N-MORB to E-MORB-like parental magma. We interpret the light eclogites as former plagioclase-rich cumulates and the dark eclogites as their complementary differentiates. This relationship is partly obscured by variable degrees of magma contamination by sediments, which also affected the metagabbros. However, the metagabbros originated from a more enriched mantle source than the eclogites. Following intrusion, the eclogites were subjected to hydrothermal alteration under the influence of seawater, as indicated by positive correlations between Li, B, Sb, and δ¹⁸O. Metamorphic fluid-rock interactions appear to be mostly of limited extent, probably due to the lack of lawsonite dehydration as a fluid source. Nevertheless, the contents at least of some fluid-mobile elements, such as LILE, Li, and Pb, were probably modified during the subduction-exhumation cycle of the eclogites.

The crustal contamination of the protolith magmas argues against derivation of the eclogites and metagabbros from typical oceanic crust. Instead, a rift-drift transition setting related to the opening of the Rheic or Saxothuringian Ocean seems most likely. The eclogites and metagabbros, alongside with similar rocks in the Mariánské Lázně complex and other resembling high-pressure massifs, may record different stages of this rift-drift transition.

1. Introduction

Eclogite facies massifs are the only parts of the accessible Earth that directly sample rocks from ancient subduction zones at depths >60 km. Geochemical data of metabasic rocks provide a wealth of information not only about subduction- and exhumation-related processes, but also...
about the mantle source, tectonic setting, and igneous and hydrothermal processes that shaped the protolith (e.g., Cann, 1976; Jochum and Verma, 1996; Spandler et al., 2004; Pearce, 2014). Because the geochemical signatures of these processes overlap, interpreting geochemical data from metabasic rocks can be challenging. Nevertheless, these overlapping signatures can be disentangled by analyzing a large set of diverse samples, as well as including a large range of geochemical proxies that are specific for a particular process. Most major elements, many incompatible elements such as high field strength elements (HFSE: Nb, Ti, Ta, Zr, Hf, Th, U) and heavy rare earth elements (HREE), and the compatible Cr and Ni, are relatively immobile during fluid-rock interactions and are thus powerful in unravelling magmatic origin and evolution (e.g., Pearce, 2014). On the other hand, elements such as Li, B, LILE (large ion lithophile elements: Rb, Sr, Cs, Ba), LREE (light rare-earth elements), Sb, Sn, and Pb are easily transported by fluids and tend to be strongly enriched in rocks that underwent seawater alteration (e.g., Jochum and Verma, 1996; Pearce, 2014). The oxygen isotope composition of mafic rocks ($\delta^{18}O = +5.7 \pm 0.2\%$ in fresh mid-ocean ridge basalt, MORB; Harmon and Hoefs, 1995) is typically modified towards higher $\delta^{18}O$ values by low-temperature (LT) hydrothermal alteration (<ca. 400 °C) whereas high-temperature alteration causes the opposite (Gregory and Taylor Jr, 1981). Both oxygen isotopes and fluid-mobile trace elements are also commonly metasomatized during later subduction-related fluid-rock interactions, and by retrogression during exhumation, requiring careful interpretation (e.g., Miller and Cartwright, 2000; Spandler et al., 2004; Bebout, 2007).

The Münchberg Massif is one of the most prominent high-pressure (HP) massifs in the Central European Variscan Belt. It contains several petrographically distinct types of metabasic rocks, most notably dark eclogites (kyanite-free), light eclogites (kyanite-bearing), and amphibolite facies metagabbros (Matthes et al., 1974, 1975; Matthes and Seidel, 1977). Despite extensive study in the 20th century (see Section 2), many questions regarding the paleogeographic location, geodynamic setting, igneous processes and metamorphic fluid-rock interactions remained unanswered or controversial: 1) Is the Münchberg Massif derived from oceanic or continental crust? (Klemd et al., 1991); 2) Are the dark and light eclogites derived from a single igneous suite, or genetically unrelated? (Müller-Sohnius et al., 1987; Stoch and Lugmair, 1990; Bosbach et al., 1991; Scherer et al., 2002); 3) Are the geochemically similar light eclogites and metagabbros derived from the same protolith suite, or from two unrelated suites? (Matthes and Seidel, 1977; Bosbach et al., 1991); 4) Were the eclogites affected by hydrothermal alteration (e.g., seafloor metamorphism) prior to subduction, and how did this influence their chemical compositions?; 5) Did HP metamorphic and retrograde fluids alter the chemical composition of the eclogites?; 6) How do the eclogites and metagabbros fit into the tectonic framework of the Variscan Orogeny? (Waizenhofer, 2017).

In order to resolve such questions, trace element studies, including oxygen isotopes and radiogenic isotopes, often prove powerful (e.g., Spandler et al., 2004; El Koth et al., 2013; Zhang et al., 2017). Although abundant major element and radiogenic isotope data exist for the Münchberg Massif, only small selections of trace elements were analyzed (e.g., Matthes et al., 1974, 1975; Stoch and Lugmair, 1990; Bosbach et al., 1991; Scherer et al., 2002) and the oxygen isotope systems are unknown. This study presents a large dataset of major and trace elements, and oxygen isotope compositions, for eclogites, metagabbros and paragneiss sampled throughout the Münchberg Massif. Based on these data and geochemical model calculations, we discuss the different igneous, hydrothermal and metamorphic processes recorded in the samples in order to resolve questions 1–6 above. The insights from our geochemical database provide not only a case study on the origin and evolution of HP massifs, but also clues for the reconstruction of Paleozoic tectonics up to the Variscan Orogeny.

2. Geologic background and petrography

The Münchberg Massif is an allochthonous nappe pile within the Saxothuringian zone of the Variscan Orogen (e.g., Behr et al., 1982). It records metamorphic events of the Variscan Orogeny, which resulted from the collision of Laurussia and Gondwana during the Devonian and Carboniferous (e.g., Kroner and Romer, 2013; Franke et al., 2017). The Münchberg nappe pile shows inverted metamorphic zonation (Fig. 1): from bottom to top, it comprises Paleozoic sedimentary rocks of the Bavarian lithofacies (anchimetamorphic), the Prasnitz-Phyllit-Serie (greenschist facies), Randamphibolit-Serie (amphibolite facies), Liegendserie (amphibolite facies), and Hangendserie (Behr et al., 1982; Klemd, 2010). A large part of the Hangendserie consists of amphibolites and felsic orthogneisses that are thought to represent bimodal magmatism at ca. 550 Ma within a Cadomian arc terrane (Koglin et al., 2018). The numerous eclogite occurrences are located in the lower part of the Hangendserie and in the Hangendserie-Liegendserie boundary, and viewed either as parts of the Hangendserie (e.g., Matthes et al., 1974) or as tectonic bodies (e.g., Stettner, 1960). Due to poor exposure, a 230 m drill core was taken from the largest eclogite occurrence at the Weissenstein locality (Fig. 1). It demonstrated that individual eclogite bodies can be at least 57 m thick and are separated by paragneisses, whose contacts with the eclogites are often tectonically undisturbed (Matthes et al., 1974; Klemd et al., 1991).

The eclogites of the Münchberg Massif can be classified macroscopically into dark eclogites (kyanite-free, low Al2O3 and Mg# ((100 × Mg/(Mg + Fe2+)); atomic %) estimating that 85 atomic % of all Fe was Fe2+ before alteration, see Cottrell and Kelley, 2011)) and light eclogites (kyanite- and/or zoisite-bearing, higher Al2O3 and Mg#; Matthes et al., 1974, 1975; Stoch and Lugmair, 1990; Bosbach et al., 1991). Dark and light eclogites are in direct contact at Weissenstein (Matthes et al., 1974), but most localities in the Münchberg Massif display only one of the two types (Fig. 2). Despite exceptions, the light eclogites tend to be more coarse-grained than the dark eclogites, which led to early speculations about gabbroic protoliths for the light eclogites and basaltic protoliths for the dark eclogites (Matthes et al., 1975). Eclogite layers (<1 m) and hornblende-plagioclase bands (several cm thick) within the Weissenstein paragneisses indicate that at least some of the dark eclogites represent original basaltic flows and/or tuff layers (Matthes et al., 1974, 1975; Gebauer and Grünenfelder, 1979). The dark eclogites have MORB-like major and trace element compositions (Stoch and Lugmair, 1990), whereas the light eclogites resemble high-alumina basalts (Matthes et al., 1975). Due to positive Eu anomalies, however, the light eclogites have been interpreted as former gabbroic cumulates, in contrast to the dark eclogites, which lack notable positive Eu anomalies (Stoch and Lugmair, 1987, 1990; Bosbach et al., 1991). Differences to the dark eclogites in Nd and Hf isotope compositions were interpreted as a distinct, possibly subduction-related origin of the light eclogites (Scherer et al., 2002) or the result of crustal contamination by sedimentary wall rocks (Stoch and Lugmair, 1990; Bosbach et al., 1991). Therefore, it has been debated whether the dark and light eclogites are genetically linked via a cumulative-differentiate relationship (Stoch and Lugmair, 1990; Bosbach et al., 1991) or unrelated (Scherer et al., 2002). Dark and light eclogites from Weissenstein yielded late Cambrian to early Ordovician protolith ages, as suggested by a Sm-Nd whole-rock isochron (480 ± 23 Ma; Stoch and Lugmair, 1990). Within uncertainty, this age is broadly coeval with other igneous events throughout the peri-Gondwanan realm (von Raumer et al., 2015), and the opening of the Rheic Ocean (in the early Ordovician, e.g., Nance et al., 2010).

Eclogite facies metamorphism was dated by Sm-Nd and Lu-Hf mineral isochrons between 400 and 380 Ma (Stoch and Lugmair, 1990; Scherer et al., 2002), in agreement with U-Pb zircon and Rb-Sr white mica age data (Gebauer and Grünenfelder, 1979). Being among the oldest eclogitization ages in the European Variscides, this age coincides with an early subduction of the Rheic Ocean at the onset of the Variscan Orogeny (e.g., Kroner and Romer, 2013). Numerous studies (compiled in...
Massonne and O’Brien, 2003, and Liebscher et al., 2007; Waizenhöfer, 2017) estimated pressure-temperature conditions of eclogite facies metamorphism in the range of 20–34 kbar and 550–750 °C. The adjacent paragneisses contain relics of this HP metamorphism, but were subjected to amphibolite facies retrogression (Klemd et al., 1991, 1994). Retrogression had a variable effect on the eclogites: observations range from essentially unaffected eclogites, and local amphibole/diopside + plagioclase + biotite + plagioclase symplectites around omphacite and phengite, to almost complete retrogression, whereby garnet, often partially replaced by hornblende or chlorite, is generally better preserved than omphacite and phengite (Matthes et al., 1974; Franz et al., 1986). P-T conditions of symplectite formation were estimated to be 8.5–12 kbar and 600 ± 50 °C (Franz et al., 1986). This amphibolite facies stage is dated between 370 and 390 Ma (see Okrusch et al., 1991). Garnet diffusion modelling demonstrated that initially rapid exhumation slowed down considerably at crustal depths (Duchene et al., 1998).

The Liegendserie contains metasediments (deposition age ≤ 483 Ma; Koglin et al., 2018) and orthogneisses (ca. 505 and 480 Ma old protoliths; Koglin et al., 2018), but also metagabbros with largely preserved igneous minerals that are partly replaced by an amphibolite facies paragenesis along grain boundaries (Matthes and Seidel, 1977; Bosbach et al., 1991). It has been debated whether the metagabbros are genetically related to the eclogites, but were subducted to different depths and therefore underwent a different metamorphic evolution. The metagabbros resemble the light eclogites in their high Al₂O₃ contents, mostly positive Eu anomalies, and similar Nd isotopic compositions, indicating that both rock types represent plagioclase-rich cumulates with varying degrees of crustal contamination during emplacement (Bosbach et al., 1991). U-Pb zircon data of 525 ± 40/31 Ma (Gebauer and Grünenfelder, 1979; recalculated to 496 ± 3 Ma by Timmermann et al., 2004) for the metagabbros allow to consider a roughly coeval protolith formation of the eclogites and metagabbros. However, Matthes and Seidel (1977) argued that differences in some element contents (Ti, K, Zr, Rb, Ba) are too large to advocate a genetic link between the two rock units.

### 2.1. Sample characterization

We adopt the classification of dark and light eclogites following previous studies, but instead of relying on macroscopic color and mineral assemblage, which are influenced by numerous factors, we introduce a geochemical classification. By means of the Al₂O₃ content and Mg# (Fig. 2), we define dark and light eclogites and, to better illustrate chemical relations among them, an intermediate group of “transitional eclogites”. Eight dark eclogites, two transitional eclogites, and four paragneiss samples from the Weissenstein drill core were used (see Matthes et al., 1974), labelled according to their corresponding drilling depth (0.13 to 154.01 m). The remaining samples were collected from

![Fig. 1. Geologic map of the Münchberg Massif, modified after Klemd et al. (1991). Stars represent the sampling localities indicated in Table 1.](image1)

![Fig. 2. Geochemical discrimination of dark, transitional, and light eclogites.](image2)
in a medium-grained matrix of plagioclase, K-feldspar, quartz, biotite, rather than pelitic in composition (Klemd et al., 1991). They contain banded gneisses (gn) of Matthes et al. (1974), which are psammitic eclogites to strongly retrogressed eclogites (ATG5: 50 vol%; EPP2: 40 vol%). The paragneiss samples from the Weissenstein core belong to the Steinhügel locality (Liegendserie). They consist of 1
vol%). The paragneiss samples from the Weissenstein core belong to the banded gneisses (gn) of Matthes et al. (1974), which are psammitic rather than pelitic in composition (Klemd et al., 1991). They contain garnet crystals (5 mm in average, often with minor retrograde chlorite) in a medium-grained matrix of plagioclase, K-feldspar, quartz, biotite, phengite, ilmenite, and accessory pyrite. Sample 154.01 also contains mm-scale hornblende-rich bands.

Samples STH1 and STH4 are amphibolite facies metagabbros from the Steinbüchel locality (Liegenserie). They consist of 1–2 mm sized plagioclase and pyroxene crystals with only minor metamorphic minerals, as described in Matthes and Seidel (1977) and Bosbach et al. (1991). In contrast, samples STH8 and STH9 are pegmatoidal meta- quartz diorites (specified as metapegmatoids hereinafter) that crop out in close association with the metagabbros. Matthes and Seidel (1977) mentioned pegmatoid samples, but did not provide a description of their petrography and geochemistry. The metapegmatoids contain pseudomorphs of hornblende and quartz after primary pyroxene/amphibole beneath strongly sericitized plagioclase phenocrysts. Euhedral to subhedral garnet is abundant at the grain boundaries, often forming girdles of multiple crystals. Interstitial quartz pockets are present, as well as rutile crystals that are typically rimmed by ilmenite.

3. Methods

3.1. Analytical geochemistry

Major elements and selected trace elements (V, Ni, Cu, Zn, Zr) were analyzed by wavelength-dispersive X-ray fluorescence spectrometry (WD-XRF) using a Zetium XRF spectrometer (Malvern PANalytical, Netherlands) at the Unit of Earth Sciences of the University of Fribourg. Forty-five trace elements were analyzed by laser ablation inductively-coupled plasma mass spectrometry (LA-ICP-MS) at the Institute of Geological Sciences of the University of Bern, using nanoparticulate pressed powder pellets following procedures described in Peters and Pettke (2017). Oxygen isotope compositions (δ18O vs SMOW) were determined by laser fluorination analysis at the Institute of Earth Sciences of the University of Lausanne. Details of all methods are given in Appendix A.

3.2. Modelling of igneous processes

We assessed the influence of fractional crystallization on the eclogite protolith compositions with a simple model. The composition of the initial melt was matched to that of the eclogites which are thought to be least affected by fractional crystallization, i.e. closest to 8 wt% MgO, 275 μg/g Cr and a Eu/Eu* (2 × Eu/Sm + Gd), contents normalized to N-MORB after Sun and McDonough, 1989) of 1 (see Pearce, 2014). Using this initial melt composition and MELTS, Excel (Guada and Ghiroso, 2015), we modelled phase equilibria during cooling from 0% to 50% crystal fractions at various pressures. Among the obtained results for the bulk solids, those which yielded a composition close to that of the light eclogites were used as the cumulate component. The entire spectrum of

| Sample     | locality       | Rock type       | Petrologic characteristics               | vol% retrogression products |
|------------|----------------|-----------------|------------------------------------------|----------------------------|
| 10.1       | Weissenstein drill core | Dark eclogite | Fe-Ti-rich MORB-type eclogite            | 4                          |
| 10.3       | Weissenstein drill core | Dark eclogite | Fe-Ti-rich MORB-type eclogite            | 2                          |
| 11.41      | Weissenstein drill core | Dark eclogite | Fe-Ti-rich MORB-type eclogite            | 15                         |
| 11.55      | Weissenstein drill core | Dark eclogite | Fe-Ti-rich MORB-type eclogite            | 10                         |
| 19.6       | Weissenstein drill core | Dark eclogite | Fe-Ti-rich MORB-type eclogite            | 17                         |
| 42.15      | Weissenstein drill core | Dark eclogite | Fe-Ti-rich MORB-type eclogite            | 15                         |
| 42.35      | Weissenstein drill core | Dark eclogite | Fe-Ti-rich MORB-type eclogite            | 30                         |
| 118.41     | Weissenstein drill core | Dark eclogite | Fe-Ti-rich MORB-type eclogite            | 25                         |
| 8.28       | Weissenstein drill core | Dark eclogite | Fe-Ti-rich MORB-type eclogite            | 3                          |
| 159        | Fattigau        | Dark eclogite   | Fe-Ti-rich MORB-type eclogite            | 5                          |
| M1         | Fattigau        | Dark eclogite   | Fe-Ti-rich MORB-type eclogite            | 30                         |
| FAT1       | Fattigau        | Dark eclogite   | Fe-Ti-rich MORB-type eclogite            | 25                         |
| FAT2       | Fattigau        | Dark eclogite   | Fe-Ti-rich MORB-type eclogite            | 12                         |
| FAT6       | Fattigau        | Dark eclogite   | Fe-Ti-rich MORB-type eclogite            | 1                          |
| UPF2       | Unterperfdt     | Dark eclogite   | Fe-Ti-rich MORB-type eclogite            | 30                         |
| SIL4       | Silberbach      | Dark eclogite   | Fe-Ti-rich MORB-type eclogite            | 15                         |
| ATG5       | Autengrün       | Dark eclogite   | Fe-Ti-rich MORB-type eclogite            | 50                         |
| ATG17      | Autengrün       | Dark eclogite   | Fe-Ti-rich MORB-type eclogite            | 15                         |
| 0.13       | Weissenstein drill core | Transitional eclogite | Moderately Al-Mg-rich eclogite         | 23                         |
| 0.35       | Weissenstein drill core | Transitional eclogite | Moderately Al-Mg-rich eclogite         | 12                         |
| WEI3       | Weissenstein drill core | Transitional eclogite | Moderately Al-Mg-rich eclogite         | 25                         |
| SIL1       | Silberbach      | Transitional eclogite | Moderately Al-Mg-rich eclogite         | 5                          |
| MR2        | Martinsreuth    | Transitional eclogite | Moderately Al-Mg-rich eclogite         | 10                         |
| WEI        | Weissenstein drill core | Light eclogite  | Kyanite-bearing Al-Mg-rich eclogite      | 0                          |
| S1861      | Oberkotzau      | Light eclogite  | Kyanite-bearing Al-Mg-rich eclogite      | 0                          |
| S1864      | Oberkotzau      | Light eclogite  | Kyanite-bearing Al-Mg-rich eclogite      | 0                          |
| Ober 1     | Oberkotzau      | Light eclogite  | Kyanite-bearing Al-Mg-rich eclogite      | 0                          |
| Ober 2     | Oberkotzau      | Light eclogite  | Kyanite-bearing Al-Mg-rich eclogite      | 2                          |
| Ober 3     | Oberkotzau      | Light eclogite  | Kyanite-bearing Al-Mg-rich eclogite      | 1                          |
| Sil6B      | Silberbach      | Light eclogite  | Kyanite-bearing Al-Mg-rich eclogite      | 15                         |
| ATG2       | Autengrün       | Light eclogite  | Kyanite-bearing Al-Mg-rich eclogite      | 5                          |
| EPP2       | Eppenreuth      | Light eclogite  | Kyanite-bearing Al-Mg-rich eclogite      | 40                         |

(continued on next page)
4. Results

A summary of major and trace element data is given in Table 2, and the full dataset in Appendix A.

4.1. Major and trace element compositions

4.1.1. Dark eclogites

Major and trace element compositions of the dark eclogites are in the usual range of mafic rocks, generally subalkaline and mostly tholeiitic (Fig. S1). They are low in Al₂O₃ (13.0–15.0 wt%) and Mg# (46–62; Fig. 2), and high in Fe₂O₃TOT (9.9–16.6 wt%).

Concentrations of typical incompatible elements in mafic magmas (TiO₂, P₂O₅, Zr, Hf, Nb, Ta, Y, rare earth elements (REE), Sc, V), but largely also Cu, Zn, Ga, Ge, In, and Sn, are higher than in the light eclogites (Figs. 3 and S2), whereas the compatible Cr is lower. The CI chondrite-normalized REE patterns (Fig. 4) are similar to those of N-MORB, although often shifted towards higher REE contents and/or Ce/Yb ratios. Slightly negative Eu anomalies are common among dark eclogites. The N-MORB-normalized multi-element spider diagrams for all eclogite samples show E-MORB-like slopes (Fig. 5). Most samples plot between the N- and E-MORB domains in the Th/Yb vs. Nb/Yb and Hf–Th–Ta discrimination diagrams and within the N-MORB field of the Nb–Zr–Y diagram (Fig. 6), but samples SIL4 and UPF2 have a more depleted signature due to very low Th contents.

4.1.2. Light and transitional eclogites

Like the dark eclogites, the light and transitional eclogites show characteristics of a subalkaline-tholeiitic trend (Fig. S1), but are higher in Al₂O₃ (15.7–20.5 wt%) and Mg# (58–84) and lower in Fe₂O₃TOT (4.9–10.1 wt%).

In contrast to the dark eclogites, the light and transitional eclogites are poor in incompatible elements, high in Cr, and have positive Eu anomalies. Sample EPP2 displays an exceptional composition with high Al₂O₃ and rather low Zr, Nb, Y, REE, Sc, and V contents as other light eclogites, but low Mg# and Cr contents like dark eclogites. The transitional eclogites generally have major and trace element compositions plotting between those of the dark and light eclogites. Typical fluid-mobile elements (e.g., LILE, L, U, Pb) do not correlate with differentiation proxies such as Zr (Fig. 5 and S2). The REE patterns (Fig. 4) generally show LREE enrichment. Except for the transitional eclogite WEI3, which is LREE-depleted. Multi-element spider diagram patterns of the light eclogites are similar to those of the dark eclogites (Fig. 5), but tend to be shifted towards lower contents, except for LILE. The light and transitional eclogites tend to have arc basalt-like signatures with higher Th/Yb, Th/Hf and Th/Ta ratios compared to the dark eclogites (Fig. 6).

4.1.3. Steinhiügel metagabbros

The Steinhiügel metagabbros (including metapegmatoids) have Al₂O₃, Na₂O, P₂O₅, Sc, and V contents similar to the light eclogites, but they are mostly richer in K₂O, LILE, and Ga and poorer in MgO, CaO, and Ni than all the eclogites. TiO₂, Y, and REE are linearly correlated with Zr, but with a different slope than the eclogites (Fig. 3). The CI chondrite-normalized REE diagrams of the metagabbros show high LREE contents and positive Eu anomalies, more strikingly than the eclogites (Fig. 4d). The slopes in multi-element spider diagrams are comparable to those of the light eclogites, and all four samples have positive K, Pb, Sr, Eu, and Li anomalies (Fig. 5d). Ratios of immobile trace elements show mixed characteristics of ocean island basalts (OIB), E-MORB, and arc basalts (Fig. 6).

4.1.4. Weissenstein paragneisses

The paragneisses of the Weissenstein drill core have major and trace element compositions similar to the upper continental crust and Post-Archean Australian Shales (PAAS), although CaO, K₂O, Pb, Sr, and Sn contents are lower (Fig. 5e). Sample 87.0 is enriched in REE, Zr, Nb, and Th.

4.2. O isotope compositions

The δ¹⁸O values of dark, transitional, and light eclogites range from +5.0 to +10.8‰, +5.5 to +8.9‰, and +6.0 to +8.3‰, respectively. Rather than between the different types of eclogites, the δ¹⁸O values of the Münchberg eclogites vary between localities (Fig. 7). The dark Weissenstein eclogites yielded the lightest values (+5.0 to +5.6‰), being equal to, or slightly lower than those of fresh MORB (+5.7 ± 0.2‰; Harmon and Hoefs, 1995). The dark Fattigau eclogites yielded the heaviest δ¹⁸O values (+10.0 to +10.8‰). All eclogites from the remaining localities fall between these two extremes (+6.0 to +9.0‰). Although δ¹⁸O values of samples within a given locality tend to be uniform, individual samples can be distinct, e.g., the transitional eclogite WEI3 (+8.9‰), which differs also texturally from other Weissenstein samples, containing cm-sized, euhedral garnet crystals. The eclogite δ¹⁸O values correlate positively with Li, B, Bi, and Sb contents (Fig. 7). The δ¹⁸O values of the Steinhiügel metagabbros (+9.4 to +10.3‰) and Weissenstein paragneisses (+10.0 to +11.0‰) resemble those of the isotopically heaviest eclogites.
Table 2
Summary of all major and trace element and oxygen isotope data.

| Element | Dark eclogites | Transitional eclogites | Light eclogites | Steinhügel metabasal | Weissenstein paragneisses |
|---------|----------------|------------------------|----------------|----------------------|--------------------------|
| wt% XRF | min. | max. | average | min. | max. | average | min. | max. | average | min. | max. | average |
| SiO2    | 47.63 | 54.69 | 50.23 | 50.45 | 53.80 | 51.59 | 48.24 | 53.69 | 50.88 | 52.61 | 55.78 | 54.30 | 62.11 | 64.05 | 63.14 |
| MgO     | 1.36 | 2.97 | 1.70 | 0.85 | 1.14 | 1.01 | 0.26 | 1.16 | 0.58 | 0.92 | 1.86 | 1.27 | 0.88 | 1.15 | 1.05 |
| Al2O3   | 12.96 | 15.43 | 14.53 | 15.73 | 16.66 | 16.18 | 17.86 | 20.47 | 18.75 | 16.98 | 19.49 | 18.35 | 15.19 | 17.17 | 16.03 |
| Fe2O3T  | 9.90 | 16.57 | 12.93 | 8.21 | 10.9 | 9.89 | 4.88 | 9.83 | 7.55 | 8.91 | 11.68 | 10.57 | 5.66 | 7.94 | 7.15 |
| MnO     | 0.18 | 0.27 | 0.20 | 0.13 | 0.14 | 0.14 | 0.10 | 0.18 | 0.14 | 0.15 | 0.21 | 0.19 | 0.11 | 0.17 | 0.14 |
| MgO     | 5.83 | 8.21 | 7.24 | 7.13 | 9.77 | 8.16 | 5.93 | 11.48 | 9.51 | 3.83 | 6.10 | 4.96 | 2.29 | 2.51 | 2.42 |
| CaO     | 9.05 | 12.33 | 10.89 | 9.63 | 12.84 | 11.10 | 9.55 | 13.07 | 10.63 | 6.57 | 8.04 | 7.27 | 1.83 | 3.48 | 2.89 |
| Na2O    | 1.82 | 1.85 | 2.03 | 2.13 | 4.64 | 3.34 | 1.59 | 3.95 | 2.38 | 2.64 | 3.56 | 3.05 | 2.53 | 3.16 | 2.96 |
| K2O     | 0.02 | 0.42 | 0.14 | 0.10 | 0.63 | 0.30 | 0.07 | 0.51 | 0.20 | 0.42 | 0.71 | 0.53 | 1.76 | 2.69 | 2.23 |
| P2O5    | 0.09 | 0.46 | 0.16 | 0.06 | 0.14 | 0.09 | 0.01 | 0.23 | 0.07 | 0.08 | 0.17 | 0.10 | 0.13 | 0.24 | 0.17 |
| Mg#     | 45 | 62 | 57 | 56 | 70 | 68 | 58 | 80 | 74 | 49 | 56 | 52 | 100 | 100 | 100 |
| SiO2 (%) | 5.0 | 10.8 | 7.2 | 5.8 | 10.9 | 8.3 | 6.0 | 8.3 | 6.9 | 6.0 | 10.3 | 9.8 | 10.0 | 11.0 | 10.6 |

4.3. Fractional crystallization modelling

Plagioclase-rich cumulates resembling the composition of light eclogites are observed at pressures below 2 kbar, where plagioclase (−An50), olivine (−Fo80), clinopyroxene (−En39D48H20T152), often also pigeonite (−En35Fs55D135T152) and accessory spinel are in equilibrium with the residual melt (see Table S1). The trends in major elements and incompatible trace elements observed from our samples (Fig. 3 and S2) are best reproduced with a cumulate component that consists of ca. 60 wt% plagioclase, 40 wt% mafic phases (olivine, augite and pigeonite in a ratio of roughly 3:4:1) and 0.5 wt% Cr-spinel. The overall good agreement of the model with the eclogite data is worse for strongly differentiated samples, for which changing liquid lines of descent can be expected, and for elements with strongly scattered patterns in Figs. 3 and S2. The estimated impact of cumulate formation on immobile trace element ratios is notable, but moderate (Fig. 6).

5. Discussion

5.1. Protoplast characterization and igneous processes

In subduction-related mafic rocks, the chemical composition of the protolith (especially LILE, but often also many other elements) is commonly found to be modified by seafloor alteration and/or metamorphism by metamorphic fluids (e.g., Miller and Cartwright, 2006; Spandler et al., 2007; Bebout, 2007; El Korb et al., 2012). Hence, geochemical fingerprinting of protolith characteristics needs to use elements that show fluid-immobile behavior, as commonly seen for HFSE (e.g., Cann, 1970; Pearce, 2014), although with exceptions (e.g., Rubatto and Hermann, 2003; Gao et al., 2007). Element mobility can be tested by plotting against a strongly immobile element (Cann, 1970), such as Zr (see Figs. 3 and S2). Good correlations with Zr were obtained for Mg#, Y, HFEE, and Cr. This points to immobile behavior of these elements as well as to a genetic relationship of the protoliths via differentiation of a common parental magma. Correlations with Zr are also good for TiO2, P2O5, Sc, V, Ge, Nb, Ta, In, and Sn, but with a few outliers. At least for TiO2, P2O5, Sc, and V, outliers can be explained by magmatic
Fe-Ti oxide and apatite fractionation, whereas occasionally mobile behavior of the other elements cannot be excluded. Crustal contamination is identified as a process that probably scattered LREE–Zr and Th–Zr trends (see below), but additional changes of LREE and Th concentrations in the slab cannot be excluded. Increased scattering for Be, Ga, Cu, Zn, and U, and a complete lack of correlations with Zr for K₂O, LILE, Li, B, As, Sb, Bi, and Pb (Fig. S2) reflect a fluid-mobile behavior of these elements.

Several previous studies of the Münchberg eclogites have demonstrated a MORB-like chemical composition of the dark eclogites, while the light eclogites have been characterized as former high alumina cumulate gabbros with more arc basalt-like trace element compositions.

Fig. 3. Major and trace concentrations of samples from the Münchberg Massif (see Fig. S2 for remaining elements). Reference compositions of fresh MORB are constructed from the data set of Jenner and O’Neill (2012), excluding strongly differentiated samples (<6 wt% MgO) and distinguishing between N-MORB (Nb/Yb = 0.4–2) and E-MORB (Nb/Yb = 2–10). The results of fractional crystallization modelling (see Sections 3.2 and 4.3) are drawn as evolution lines, with the initial melt (L0) and cumulate (C) components indicated. R² values are given as an indication of the quality of correlations.
Our new data confirm these patterns and reveal them in more detail: the dark eclogites show mixed N-MORB and E-MORB characteristics, whereas the light eclogites form a transition between the dark eclogites and arc basalts (Figs. 4–6). The immobile incompatible trace element contents (HFSE, REE, Y) vary by more than one order of magnitude between the most enriched dark eclogites (11.41, 8–28, SIL4) and the most depleted light eclogites (SIL6B, WMH2, ATG2). For the latter, the contents are so low that they are probably derived from gabbros with extreme cumulate characteristics. But also some dark eclogites are probably derived from gabbros rather than basalts. This is indicated by the coarse-grained textures of some dark eclogites, although these could also reflect fast diffusive transport by fluids during metamorphism. Furthermore, dark and light eclogites at Weissenstein yielded igneous zircon cores (Gebauer and Grünenfelder, 1979), which can only have formed from more differentiated interstitial melt within coarse-grained rocks. Some dark Weissenstein eclogites, however, were likely derived from tuffs and/or basaltic lava flows.

Fig. 3. (continued).
There has been a controversy whether the distinct chemical compositions of the dark and light eclogites represent two genetically unrelated magmas (Müller-Sohnius et al., 1987; Scherer et al., 2002), or plagioclase-rich cumulates (light eclogites) and their corresponding differentiates (dark eclogites) derived from a single parental magma (Stosch and Lugmair, 1990; Bosbach et al., 1991). Arguments for the former hypothesis are differences in Nd and Hf isotopes (Scherer et al., 2002) and the rareness of juxtaposition of these two eclogite types (see Matthes et al., 1975). However, as the light eclogites have clear cumulate characteristics (high Al₂O₃, Mg#, and Cr; low incompatible element contents; positive Eu anomalies), their corresponding differentiates should have opposing geochemical characteristics, which essentially correspond to those observed in the dark eclogites. It is difficult to envisage that the differentiated counterparts of the light eclogites are not represented by the closely associated dark eclogites, but by similar hypothetic rocks that were never found in the accessible geologic record. Furthermore, our new dataset contains transitional eclogite samples which represent chemical links between dark and light eclogites for all major and immobile trace elements (Figs. 3 and S2), strengthening the single parental magma hypothesis. Throughout the dark, transitional and light eclogites, the Zr content (indicative of magmatic differentiation) shows positive correlations with other immobile incompatible elements (e.g., TiO₂, P₂O₅, Y, Nb, HREE), and negative correlations with Eu/Euᵣ (indicating plagioclase fractionation) and Cr (indicating clinopyroxene and/or spinel group mineral fractionation). These trends with their gradual transitions, both within and throughout single eclogite localities, convince us to argue for a single parental magma for the dark and light eclogites. This is further corroborated by our fractional crystallization model (Section 3.2, 4.3 and Appendix A; Figs. 3 and S2), which reproduces the chemical compositions (except fluid-mobile elements) of almost all our eclogite samples through fractional crystallization of a MORB-like parental magma at <2 kbar. Only samples UPF2, according to field observations (see Matthes et al., 1974, 1975; Gebauer and Grünenfelder, 1979).
Fig. 5. Multi-element spider diagrams of the Münchberg samples. Normalization to N-MORB and reference data for E-MORB and OIB after Sun and McDonough (1989). Values for bulk and upper continental crust are from Rudnick and Gao (2003), and values of Post-Archean Australian Shale (PAAS) after Taylor and McLennan (1995).
SIL4 (dark eclogites), and WEI3 (transitional) plot outside of the general trend, being exceptionally poor in Nb, Ta, and Th compared to other immobile elements (Fig. 6). Instead of reflecting derivation from different parental magmas, however, these differences may alternatively be a consequence of fluid-rock interactions (see Section 5.2).

Significant differences in Nd and Hf isotope compositions and ratios of mantle source-specific trace elements in the dark and light eclogites (Bosbach et al., 1991; Scherer et al., 2002), previously applied to argue against their derivation from a common parental magma, may instead be a consequence of magma contamination (see Section 5.2).

Fig. 6. Immobile element discrimination diagrams for samples from the Münchberg Massif: (a) Th/Yb vs. Nb/Yb plot after Pearce (2008); (b) Hf–Th–Ta plot after Wood (1980); (c) Nb–Zr–Y plot after Meschede (1986). Note that samples with cumulate characteristics (i.e. light eclogites and metagabbros, see Section 5.1) cannot be reliably classified using basalt discrimination diagrams, but they are plotted here to enable comparison with the dark eclogites and to demonstrate how magma contamination by sediments (paragneisses) distorted their trace element signature (shown by qualitative arrows). A quantitative assessment of the bias related to cumulate formation for the light eclogites (arrows representing changes from 0% to 90% cumulate component) is based on fractional crystallization modelling (see Section 3.1).

Stosch and Lugmair, 1990). The light eclogites have lower εNd values than most of the dark eclogites (Bosbach et al., 1991). The gneisses have high Th/Yb ratios (Fig. 8), suggesting that the Th/Yb ratio is a possible proxy of magma contamination for our dataset. The Th/Yb ratios of the light and transitional eclogites are often much higher than those of the dark eclogites (Fig. 6a), misleadingly suggesting an arc magma signature of the light eclogites. However, some light eclogites have low Th/Yb ratios similar to the dark eclogites, which rather argues that both eclogite types originated from a MORB-type parental magma, and that the Th/Yb ratios of many light eclogites were raised by magma contamination. The dark eclogites are richer in incompatible elements, hence their εNd, εHf, and Th/Yb values are less influenced by late magmatic crustal contamination than in the light eclogites. The different
radiogenic isotope signatures of the dark and light eclogites may have formed via contamination after their separation into different chemical systems. Assuming that the cumulates formed at the margins of the magma chamber, these protoliths of the light eclogites may have been more prone to sediment contamination. Conversely, dark eclogites with extrusive protoliths may have avoided strong sediment assimilation by fast ascent and cooling. We have performed modelling calculations to test the magma contamination hypothesis with actual data, and to assess it quantitatively (Section 3.2 and Appendix A). As displayed in Fig. 8, ≤4–5 wt% of sediment with the composition of the Weissenstein paragneisses need to be added to the magma to generate the initial Nd isotope compositions of the dark and light eclogites in Stosch and Lugmair (1990) and Bosbach et al. (1991). The impact of 4–5 wt% of sediment contamination on the major element composition of the eclogites is small (except for K2O), but strong increases in Rb, Ba, LREE, Pb, Th, and U are predicted (Appendix A). Except for Th, however, these trace elements are very mobile during hydrothermal and metamorphic fluid-rock interactions. Some degree of Nd mobilization by fluids cannot be excluded, but it is unlikely that this could have created the range of εNd values and Th/Yb ratios in the eclogite samples. Moreover, the Lu-Hf system is too fluid-immobile to allow the different Hf isotope signatures (Scherer et al., 2002) to be explained by fluid-rock interactions. Consequently, we infer that most, if not all Münchberg eclogites are derived from the same igneous suite with N- to E-MORB composition that experienced variable degrees of crystal fractionation and sediment contamination.

The Steinhügel metagabbros have been viewed either as a possible equivalent of the light eclogites that did not experience eclogite facies metamorphism (Bosbach et al., 1991), or as the member of a magmatic
suite unrelated to the eclogites (Matthes and Seidel, 1977). Our data show that the metabasalts and light eclogites are similar in their high Al$_2$O$_3$ contents and positive Eu anomalies that characterize plagioclase-rich cumulates (Fig. 3). Moreover, their Nd isotope composition also suggests crustal contamination (see Bosbach et al., 1991). However, the metabasalts have substantially higher Nb/Yb, Nb/Y, and Ta/Hf ratios (Fig. 6) than both the dark and light eclogites. These differences can hardly be explained by crustal contamination or fluid-rock interactions alone. Instead, the metabasalts seem to have originated from a more enriched mantle source without arc affinity, as also suggested by the almost OIB-like patterns in the multi-element spider diagram (Fig. 5d). We therefore conclude that the metabasalts are not sourced from the same igneous suite as the light eclogites, but originated from a more enriched mantle source.

5.2. Extent of bulk-rock chemical modifications by fluid-rock interactions

Lacking correlations between fluid-mobile elements and Zr (Section 5.1) indicate that fluid-rock interactions related to hydrothermal alteration and/or subduction zone metamorphism had a significant influence on the chemical composition of the Münchberg eclogites. The large range (+5.0 to 10.8‰) and geographic distribution of $\delta^{18}$O values in the Münchberg eclogites provide important information about the type, conditions, and impact of fluid-rock interactions. Fresh MORB have very homogeneous, mantle-like $\delta^{18}$O values of $+5.7 \pm 0.2$‰ (Harmon and Hoefs, 1995). The crustal contamination for the protoliths (Section 5.1) can only explain a < 0.5‰ increase in $\delta^{18}$O, as inferred from the $\delta^{18}$O values of the paragneisses. In ophiolitic crust, however, $\delta^{18}$O values are typically raised by LT alteration in the extrusive section and lowered by high-temperature alteration in the gabbro section (e.g., Gregory and Taylor Jr, 1981; Cartwright and Barnicoat, 1999). Fluid-rock interactions can also modify $\delta^{18}$O values during subduction when rocks react with externally-derived fluids, but this effect is thought to be modest unless the rocks were located close to a fluid flow pathway and/or strongly retrogressed (e.g., Cartwright and Barnicoat, 1999; Miller and Cartwright, 2000). For our samples, widespread subduction metamorphism is not supported by field and petrographic evidence, and retrogression is mostly modest or absent. Externally-derived fluids in subduction zones flow along high-permeability zones such as fractures and shear zones and can interact with the adjacent host rock, but the fluid reaction fronts rarely propagate more than several decimeters into the surrounding rocks (e.g., John et al., 2008; Taetz et al., 2018). The fairly uniform $\delta^{18}$O values for Fattigau (+10.0 to +10.8‰, $n$ = 5) and Oberkotzau (+6.0 to +6.9‰, $n$ = 5) eclogites indicate that the exchange of oxygen isotopes is pervasive and thus essentially happened during pre-subduction LT alteration. Most of the Weissenstein eclogites still display MORB-like $\delta^{18}$O values (+5.0 to 5.6‰), either due to (1) very limited impact of seafloor alteration (low water/rock ratios), (2) alteration at intermediate temperatures that did not cause prominent oxygen isotope fractionation, or (3) resetting of the $\delta^{18}$O values during metamorphic fluid-rock interactions. Explanation (1) seems most likely because the mobile elements Li, B, and Sb, which are typically strongly enriched in altered MORB (e.g., Spivack and Edmond, 1987; Jochum and Verma, 1996; Coogan et al., 2017), show contents comparable to fresh MORB (Fig. 7). The entirety of the eclogites from all localities shows positive correlations with $\delta^{18}$O for Li, B, Sb, and Bi (Fig. 7). For B, such correlations were identified in basalts that experienced seawater alteration (Spivack and Edmond, 1987). Lithium and Sb also tend to be drastically enriched during seawater alteration (Jochum and Verma, 1996; Coogan et al., 2017). We thus conclude that the Li, B, and Sb signatures are dominated by the effects of seawater alteration. The same should apply to Bi, whose behaviour during seawater alteration is less understood. The preservation of their correlations with $\delta^{18}$O suggests that the otherwise commonly fluid-mobile elements Li, B, Sb, and Bi were not or only moderately affected by subduction metamorphic overprint. Magna et al. (2006) and Marschall et al. (2007) reported Li contents in Münchberg eclogites similar to our study, and $\delta^{7}$Li values of −8.3 to +2.9‰. The negative $\delta^{7}$Li values, which can hardly be explained by seafloor alteration alone, have instead been interpreted to reflect Li addition by metamorphic fluids after kinetic isotope fractionation (Marschall et al., 2007).

The remaining elements do not show a clear correlation with $\delta^{18}$O, indicating that they either still reflect igneous concentrations, or that they were mobilized by metamorphic fluids. For Sn, which can actually be affected easily by hydrothermal reactions (e.g., Jochum and Verma, 1996), the igneous pattern (Fig. 3e) seems to be well-preserved (the negative Sn anomalies reported here are probably due to a too low Sn concentration in the GSD-1G standard used for calibration of the LA-ICP-MS data, see Peters and Pettke, 2017). K$_2$O and LILE concentrations do not correlate with Zr or $\delta^{18}$O, and ratios of different LILE vary greatly between the eclogite samples (Figs. 9 and S2). Most of our eclogite samples follow the trendline of metamorphic enrichments and losses in the K/Th vs. Ba/Th diagram of Bebout (2007; Fig. 9). Higher K/Ba ratios typical for seafloor alteration are generally absent, although other geochemical modifications that are typical for seafloor alteration are evident (see above). This may indicate that seawater alteration does not always influence the LIL budget in the same way. On the other hand, as discussed in Section 5.3, the Münchberg Massif does not represent typical oceanic crust, hence differences to the typical seafloor alteration pattern are not overly surprising. In case of the Münchberg eclogites, the LILE compositions seem to be dominated by metamorphic fluid-rock interactions. This is corroborated by petrographic observations that phengite is often enriched in bands along the foliation, probably due to K mobility in metamorphic fluids. Some samples with very low K/Ba ratios may suggest that the LILE budgets of metabasic rocks are sometimes affected in different patterns than that proposed by Bebout (2007). Variable source lithologies of metamorphic fluids, and variable stability of LILE-hosting minerals may cause complex patterns as a result of fluid-rock interactions at different stages of the P-T path. In addition to LILE mobility, fluid transport of more immobile elements needs to be considered, at least for the Nb, Ta-, and Th-depleted samples WEI3, UPF2, and SIL4. Samples UPF2 and SIL4 plot beneath the MORB-OIB array in Fig. 6a and may belong to the same magmatic suite as the bulk of the eclogites, but possibly lost Th to fluids or melts. If this explanation is applied to WEI3, these fluids must have removed Nb and Ta as well. This is generally possible (see Gao et al., 2007), and intense fluid-rock interaction in WEI3 is petrographically indicated (cm-sized euhedral garnet crystals; mm-sized quartz-rich pockets; large degree of retrogression). Samples WEI3, UPF2, and SIL4 all show LREE depletion.

Fig. 9. Chemical discrimination diagram of seafloor alteration and metamorphic fluid-rock interactions after Bebout (2007). Average fresh N-MORB and E-MORB after Sun and McDonough (1989).
supporting loss of LREE, and possibly HFSE, to fluids or melts.

Besides geochemical evidence, the eclogites of the Münchberg Massif contain petrographic evidence for metamorphic fluid-rock interactions. Quartz veins are locally common (e.g., at Autengrün) and often contain kyanite and/or phengite whereas garnet-, omphacite-, or rutile-bearing veins are rare (see also Okrusch et al., 1990). However, compared to the large interconnected vein networks of some eclogite occurrences of oceanic crustal origin, such as in the Tianshan (e.g., John et al., 2008) or in New Caledonia (e.g., Taetz et al., 2018), the Münchberg eclogites are poor in metamorphic veins. This argues for limited fluid release during high-pressure metamorphism of the Münchberg eclogites, which can be explained by limited water availability and/or the stability of hydrous phases along the P-T path. Despite oxygen isotope evidence for hydrothermal alteration, the eclogite protoliths may have been only modestly hydrated, which is not surprising if they represent continental crust as discussed in Section 5.3. This may explain why metamorphic veins are especially rare in the Weissenstein eclogites, which mostly have unmodified oxygen isotope compositions. Pseudosection modelling of the Münchberg eclogites (Waiženhofer, 2017) suggests that during prograde (amphibolite facies) and eclogite facies metamorphism, temperatures were too high for lawsonite stability. Lawsonite dehydration is the major fluid source from HP rocks (e.g., Barnicoat and Fry, 1986), and without it, only a limited amount of fluids may have formed via amphibole and/or zoisite dehydration. Whatever the cause for the scarcity of metamorphic veins, it suggests that HP fluid-rock interactions shaped the chemical composition of the Münchberg eclogites only to a moderate degree.

Petrographic evidence for retrograde (mostly amphibolite facies) fluid-rock interactions is much more widespread, especially in the form of amphibole/diopside + plagioclase symplectites, which can be ascribed to retrograde influx of low to moderate salinity aqueous fluids (Klemd, 1989). Such fluids may have introduced fluid-mobile elements into the eclogites. Throughout the localities, the eclogites tend towards higher Pb contents with the percentage of retrogression (Fig. 10a). Although only for the Weissenstein locality, a similar pattern can be observed for Ba (Fig. 10b). The Weissenstein paragneisses are rich in Pb and Ba, suggesting that these elements were potentially introduced into the eclogites by paragneiss-derived fluids. However, besides these two weak correlations, there is no clear link between retrogression and the concentration of any element, implying that the influence of retrogression on the eclogite composition is weak compared to igneous, hydrothermal, and HP metamorphic processes.

5.3. Origin of the eclogites and metagabbros

The question whether the Münchberg Massif represents oceanic or continental crust had been discussed without a clear outcome (Klemd et al., 1991). MORB-type eclogites as in the Münchberg Massif most commonly originate from oceanic crust which, at least ideally, consists of a gabbro section overlain by a 1 km section of sheeted dikes and pillow lavas, and sediments that were accumulated after the formation of the igneous rocks (Penrose Conference Participants, 1972). However, previous Nd isotope data (Stosch and Lugmair, 1990; Bosbach et al., 1991) and our trace element data indicate sediment contamination during intrusion of the eclogite and metagabbro protoliths (Section 5.1). We argue that this is improbable in typical oceanic crust. A continental rather than oceanic origin is further supported by field and petrologic evidence. The paragneisses of the Weissenstein drill core are the only rocks for which a direct lithologic association of protoliths with the eclogites is ascertained (Section 2). Chemically, these gneisses largely represent metapsammites rather than metapelites (SiO₂ contents around 65 wt%; Matthes et al., 1974; this study), which argues for deposition proximal to a continental source of coarse-grained detritus. Thin amphibolite layers and hornblende-plagioclase bands intercalated with the gneisses indicate a deposition of basaltic tuffs (Matthes et al., 1974, 1975; Gebauer and Grünenfelder, 1979). The explosive nature of this

volcanism precludes a deep-water setting (see Honnorez and Kirst, 1975). Furthermore, the Münchberg eclogites experienced fast exhumation rates (Duchêne et al., 1998) and can thus be classified as continental type eclogites after Guillot et al. (2009).

Even though the Münchberg eclogite protoliths were formed within a continental context, our geochemical data (e.g., correlations of Li, B, and Sb contents with δ¹⁸O values; Section 5.2) indicate hydrothermal alteration dominated by seawater. The tectonic environment of protolith formation may have been a rift-drift transition setting, i.e. the formation of new passive margins during continental breakup, accompanied by mafic magmatism intruding continental wall rocks proximal to seawater. This currently happens in the Gulf of California, which has already been applied as a recent analogue for the opening of the Rheic Ocean and separation of Avalonia from Gondwana (Nance et al., 2002).

The large-scale extensional tectonic setting related to the opening of the Rheic Ocean continued eastwards, where Central European terranes like the Saxothuringian and Teplá-Barrandian zones were located (see Kroner and Romer, 2013; Franke et al., 2017; Fig. 11). The protolith age of the Münchberg eclogites (480 + 23 Ma; Stosch and Lugmair, 1990) coincides with the opening of the Rheic Ocean in the early Ordovician, making it a viable scenario that the eclogite and gneiss protoliths represent transitional continental–oceanic crust at the margin of the newly-forming Rheic Ocean. However, it is still debated whether the Rheic Ocean was the only ocean between Laurussia and Gondwana that opened in the early Paleozoic (e.g., Kroner and Romer, 2013; Stephan
Münchberg eclogites. Uniformly NE-SW oriented ductile mineral lineations in the Münchberg Massif, as already proposed by Waizenfenks Klippen suggest that these units represent a formerly coherent tectonic setting in the Münchberg Massif as well as in the Frankenberg and Wildenstein basins, are therefore also a potential setting for the formation of the Mariánské Lázně Complex at the western margin of the Teplá-Barrandian Zone has already been proposed as a possible equivalent of the Hangenserie and Liegendserie (e.g., Beard et al., 1995; Timmermann et al., 2004). Both complexes comprise similar lithologies (eclogites, amphibolites, coronitic metagabbros, orthogneisses, paragneisses with calc-silicate lenses). The eclogites of the Mariánské Lázně Complex resemble those of the Münchberg Massif in their major element contents, paragenesis (kyanite-bearing and kyanite-free eclogites), trace element characteristics, and comparable initial εNd values (Beard et al., 1995; Timmermann et al., 2004). The lowest εNd values in the Mariánské Lázně eclogites (down to ~0.8) were explained with a more enriched mantle source by Beard et al. (1995). However, this could also be a consequence of crustal contamination, as we suggest for the Münchberg eclogites, which would allow for a common tectonic setting of both complexes. Only for intrusion ages, conflicting data were obtained (Mariánské Lázně eclogites: ca. 540 Ma, U-Pb dating of zircon cores, Timmermann et al., 2004; Münchberg eclogites: 480 ± 23 Ma, Sm-Nd, Stosch and Lugmair, 1990). However, the metagabbros of the Münchberg Massif, Mariánské Lázně Complex, and Zone of Erbendorf-Vohenstrauß all yielded U-Pb zircon intrusion ages around 496 Ma (Timmermann et al., 2004, and references therein). Furthermore, geochemical data of the Mariánské Lázně metagabbros (Timmermann et al., 2004) show a similar trace element enrichment as the metagabbros in the Münchberg Massif. This allows to correlate metagabbro formation in the three complexes with the large-scale extensional tectonics preceding the opening of the Rheic Ocean, although the data for the eclogites are still too ambiguous to elucidate whether the Saxothuringian Ocean also opened in this setting. Within the Münchberg Massif, the different trace element compositions of the eclogites and metagabbros, and further differences between the lithologic inventories of the Hangendserie and Liegendserie (e.g., Behr et al., 1982), emphasize that these two nappes are not derived from the same original geologic unit. However, the Münchberg eclogites and metagabbros are probably related to the opening of the same ocean and may represent different stages of the rift-drift transition.

6. Conclusions

1. The protoliths of the Münchberg eclogites appear to be derived from a common parental magma with N-MORB to E-MORB composition. The light eclogites represent former plagioclase-rich cumulates and the dark eclogites their complementary differentiates. Crustal contamination via sediment assimilation affected the trace element composition especially of the light eclogites, misleadingly suggesting derivation from an arc magma.

2. Crustal contamination argues against derivation of the Münchberg eclogites from typical oceanic crust. The most likely geotectonic setting is a transition between continental rifting and the spreading of the Rheic or Saxothuringian Ocean. Geochemical similarities support a correlation with the Mariánské Lázně eclogites, although the protoliths of the latter may be older.

3. The Steinhügel metagabbros represent plagioclase-rich cumulates, as do the light eclogites, but have a more enriched, almost OIB-like trace element signature. The two rock units therefore represent different igneous suites from different mantle sources, even though they were probably emplaced in some spatiotemporal association within the same large-scale extensional tectonic setting.

Fig. 11. Paleogeographic sketch to illustrate two different hypotheses for the tectonic history of the Münchberg Massif at the stages of (a) protolith formation, (b) the opening of mature oceans, and (c) subduction (and later nappe emplacement in dashed arrows): (1) related to the opening of the Rheic Ocean, or (2) related to the opening of the Saxothuringian Ocean. The paleogeographic reconstruction broadly follows the model of Franke et al. (2017) and serves here only as an example, while not rejecting other models (e.g., Kroner and Romer, 2013; von Raumer et al., 2015; Stephan et al., 2019) which, however, would only permit the illustration of (1) as they do not postulate the presence of the Saxothuringian Ocean.
4. Most of the eclogites were variably affected by pre-subduction hydrothermal alteration that seems to be dominated by seawater. This alteration had a large impact on oxygen isotope compositions and enriched the eclogites in some fluid-mobile elements, especially Li, B, Sb, and Bi.

5. Metamorphic fluids probably caused a substantial redistribution of LILE in the Münchberg eclogites. Locally, LREE, Tb, Nb, and Ta were probably mobilized by fluids and/or melts. However, the lack of lawsonite dehydration as a fluid source and absence of interconnected vein networks indicate a rather limited extent of metamorphic fluid-rock interactions. Retrograde fluids may have introduced metasediment-derived Pb and Ba into the eclogites.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Appendix A. Supplementary data

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