Paleoproterozoic high-pressure metamorphic history of the Salma eclogite on the Kola Peninsula, Russia

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INTRODUCTION

One of the most important questions in Earth sciences involves the initiation and evolution of subduction during the Precambrian (e.g., Cawood et al., 2006; van Hunen and Moyen, 2012, and references therein). Many researchers have inferred that subduction began during the Archean (e.g., Komiyama et al., 1999; Brown, 2006, 2009; Cawood et al., 2006; Van Kranendonk et al., 2007), based on the presence of indicators of plate tectonics, such as accretionary prisms, orogens, and paired metamorphic belts. Brown (2006) suggested that the first appearance of Neoarchean high-pressure granulite reflects the initiation of subduction, which has a geothermal gradient higher than that observed in modern subduction characterized by blueschist. Blueschist facies metamorphism is believed to have begun during the Neo-Proterozoic (Maruyama et al., 1996; Stern, 2005; Tsujimori and Ernst, 2014). These occurrences indicate that an early style of subduction without blueschist may have dominated between the Neoarchean and the Neoproterozoic (Brown, 2006, 2009). However, the evolution of this early style of subduction prior to the Neoproterozoic remains unclear.

Because eclogite is a typical high-pressure rock formed within subduction zones, unraveling the metamorphic evolution of Precambrian eclogites with pre-Neoproterozoic ages is important in order to study the evolution of the early style of subduction occurring prior to the Neoproterozoic. Precambrian eclogites are rare worldwide. The Paleoproterozoic eclogites from Tanzania (2.0 Ga in the Usagaran belt; 1.89–1.86 Ga in the Ubendian belt) are well-known examples that are considered to represent remnants of the subducted Paleoproterozoic oceanic lithosphere (e.g., Möller et al., 1995; Collins et al., 2004; Boniface et al., 2012). However, the existence of Archean eclogites remains controversial. Mints et al. (2010, 2014) suggested that the eclogite in the Salma area of the Kola Peninsula is an Archean eclogite, based on the 2.87–2.82 Ga zircon age recorded in the eclogite. The 2.72–2.70, 2.4, and 1.9 Ga zircon ages obtained from the Salma eclogite were interpreted to be retrograde metamorphic ages (Mints et al., 2010, 2014). In contrast, Skublov et al. (2010a, 2011) suggested that
the Salma eclogite is a Paleoproterozoic eclogite, based on the presence of metamorphic zircons recording ages of ca. 1.92–1.88 Ga, low Th/U ratios, and flat heavy rare earth element (HREE) patterns. Lu-Hf and Sm-Grt ages (1.90–1.88 Ga) obtained from eclogite and eclogitized ultrabasite were also interpreted to reflect eclogite facies metamorphic ages (Skublov et al., 2010b; Herwartz et al., 2012; Mel’nik et al., 2013). However, garnet from massive eclogite (sample 46, Table 1) exhibits prograde zonning (Skublov et al., 2011), implying that these ages may represent prograde metamorphic ages rather than ages of peak metamorphism. These previous studies presented no direct evidence with which to determine which zircons formed during eclogite facies metamorphism, such as omphacite inclusions in zircon. Therefore, the timing of the eclogite facies metamorphism of the Salma eclogite is still uncertain, and it is necessary to confirm whether the Salma eclogite is Archean or Paleoproterozoic in age based on direct evidence. In addition, the study of the pressure-temperature (P-T) conditions of eclogites in the Salma area is necessary, because only minimum pressure conditions have been determined using geothermobarometry (Mints et al., 2010, 2014; Shchipansky et al., 2012). Determining the age and petrogenesis in zircon. The metamorphic zircons recording ages of ca. 1.92–1.88 Ga, low Th/U ratios, based on zircon U-Pb dating coupled with analyses of REEs and inclusions in zircon. The pressure conditions have been determined using geothermobarometry and pseudosection modeling. In addition, whole-rock chemistry was analyzed to characterize the tectonic setting in which these eclogites originally formed. We suggest that Paleoproterozoic subduction zones were relatively warmer than Phanerozoic subduction zones but colder than Neoproterozoic subduction zones.

### GEOLOGICAL BACKGROUND

The Fennoscandian shield records a general trend in which the age of geological activity decreases toward the southwest. The northern part of the shield is dominated by Archean rocks, whereas the major part of the shield comprises the Paleoproterozoic 2.0–1.8 Ga Svecofennian Province and the 1.8–1.65 Ga Transscandinavian Igneous Belt. The 1.2–0.9 Ga Sveconorwegian Province is farther to the southwest (Daly et al., 2006; Fig. 1A). In the northern part of the shield, the Kola-Karelian orogen is located between the Kola and Karelian cratons. The Kola-Karelian orogen mainly consists of three Paleoproterozoic tectonic belts (the Kola suture belt, the Tanaelv belt, and the Lapland and Umbra granulite belts), the Neoarchean Inari microcontinent, and the Belomorian mobile belt (Fig. 1B).

The Belomorian mobile belt is principally composed of 2.9–2.6 Ga tonalite-trondhjemite-granodiorite (TTG) gneisses (Holttä et al., 2008; Mints et al., 2014) and includes a ca. 2.9 Ga paragneiss complex and 2.9–2.8 Ga greenstone belts (Slabunov et al., 2006). The available geological, isotopic, and geochemical data from the mafic-ultramafic rocks of the greenstone complex are compatible with their interpretation as the tectonically disrupted and metamorphosed remnants of a Mesoproterozoic ophiolitic association (Slabunov et al., 2006). This belt underwent multiple deformation and metamorphic events during both the Archean and Paleoproterozoic (Daly et al., 2001, 2006; Mints et al., 2014). The Paleoproterozoic

### TABLE 1. AGE CONSTRAINTS AND INTERPRETATIONS FROM THE SALMA ECLOGITES AND RELATED ROCKS

<table>
<thead>
<tr>
<th>Lithology (sample identification)</th>
<th>Method</th>
<th>Age (Ma)</th>
<th>Interpretation of source given</th>
<th>References</th>
</tr>
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<tr>
<td>Eclogite (S-198/107)</td>
<td>U-Pb Zrn</td>
<td>2703 ± 9</td>
<td>Retrograde granulite facies metamorphism</td>
<td>Mints et al. (2010)</td>
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<td>Fe-Ti eclogite (S-204-2B)</td>
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<td>Magmatic protolith age</td>
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<td>Garnetite (S-204-23B)</td>
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<td>Partially reset age</td>
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<td>Plagiogranite vein (S-204-26)</td>
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<td>2781 ± 15</td>
<td>Eclogite facies metamorphism to ca. 2.87 Ga or older</td>
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<td>Eclogite (S-198/107)</td>
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<td>2724 ± 35</td>
<td>Granulite facies metamorphism</td>
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<tr>
<td>Eclogite (Ex198)</td>
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<td></td>
<td></td>
<td>(2079 ± 34)</td>
<td>Magmatic protolith event</td>
<td>Skublov et al. (2011)</td>
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<tr>
<td></td>
<td></td>
<td>1923 ± 75 (1878 ± 36)</td>
<td>Eclogite facies metamorphism</td>
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<td>Elogitized ultrabasic (sample 21)</td>
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<td>1907 ± 11</td>
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<td>Skublov et al. (2010a)</td>
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<td>Pegmatite vein (sample 62)</td>
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<td>Massive eclogite (sample 46)</td>
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<td>Metamorphic event</td>
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<td>This study</td>
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<td>Eclogite facies metamorphism</td>
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<td>Eclogite (RPB3A)</td>
<td>U-Pb Zrn</td>
<td>2727 ± 8</td>
<td>Amphibolite facies metamorphism</td>
<td>This study</td>
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<td>1868 ± 17</td>
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<td></td>
<td></td>
<td>1720 ± 79</td>
<td>Retrograde amphibolite facies metamorphism</td>
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</table>

**Note:** Ages in parentheses are data from Skublov et al. (2011). Zrn—zircon; Grt—garnet.
Reworked Archean TTG (including eclogite) occur within TTG gneisses, which mainly consist of LITHOSPHERE | Volume 9 | Number 6 | www.gsapubs.org

division of the northeastern Fennoscandian shield region. Sample locations: 3A, 3B, and 3C: N67°28 ′03 ″; samples 1A and 1B: N67°31′07 ″/E32°22′39 ″; modified from Berthelsen and Marker (1986), Zhao et al. (2002), and Daly et al. (2006). TTG—tonalite-trondhjemite-granodiorite.

In the Belomorian mobile belt, several eclogite exposures occur (Fig. 1) in the Salma area (Kaulina et al., 2010; Mints et al., 2010; Skublov et al., 2010a, 2010b, 2011), in the Kuru-Vaara quarry (Shchipansky et al., 2012; Balagansky et al., 2014), and in the Gridino area in the northeastern region of Karelia (Volodichev et al., 2004, 2012; Dokukina et al., 2014). The Salma and Kuru-Vaara eclogites are related to the high-pressure metamorphism of oceanic lithosphere and exhibit peak P-T conditions of 13–14 kbar and 700–750 °C (e.g., Mints et al., 2010, 2014; Shchipansky et al., 2012). The Gridino eclogites were originally gabbroic dike swarms that intruded felsic gneiss and were metamorphosed under peak P-T conditions of 17–18 kbar (probably to 22 kbar) and 740–865 °C (Volodichev et al., 2004; Dokukina et al., 2014).

In this study we focus on eclogite and granulite samples collected from two outcrops in the Salma area of the Belomorian mobile belt, which is located near the Tanaelv belt (Fig. 1B). In the Salma area, mafic bodies (including eclogite) occur within TTG gneisses, which mainly consist of quartz diorite and trondhjemite. In the mafic bodies, eclogites occur as layers that are intercalated with garnet granulite, garnet amphibolite, amphibolite, and garnetite (Fig. 2A). Although some eclogite layers are fresh and coarse grained, most eclogite layers are medium grained and have retrograded into granulite and amphibolite (Figs. 2B–2D). In the coarse-grained eclogites, garnets occur within a bright green matrix, which mainly consists of clinopyroxene (omphacite and calcic clinopyroxene) with minor amphibole (Fig. 2C). In the medium-grained eclogite, less garnet occurs within a matrix that is dark green, due to the presence of abundant amphiboles that formed during retrograde metamorphism (Figs. 2B, 2D). The interlayered granulite and amphibolite can be considered to represent strongly retrograded eclogite. During retrograde metamorphism, omphacite was first retrograded into calcic clinopyroxene, thus forming symplectite around garnet and in the matrix; then, clinopyroxene was retrograded to amphibole. Mints et al. (2014) inferred that the protolith of the layered mafic body was originally a series of normal gabbro norite, olivine gabbro, and Fe-Ti oxide gabbro intercalated with local troctolite, which resembled the gabbroic suite from the modern oceanic crust of the slow-spreading Southwest Indian Ridge (Dick et al., 2000). The different degrees of retrograde metamorphism may be due to compositional differences or varying amounts of water infiltration. The retrograde eclogite includes a few leucosomes (Figs. 2B, 2D) consisting of quartz, plagioclase, K-feldspar, and relict garnet, thus indicating that partial melting occurred during metamorphism.

**PETROGRAPHY AND MINERAL CHEMISTRY**

The mineral compositions of the five samples were analyzed using a Shimadzu electron probe microanalyzer (EPMA-1600) at the Jeonju branch of the Korea Basic Science Institute (South Korea). Multiple eclogite samples (RPB3A, RPB1A, and RPB1B) and garnet-clinopyroxene granulite samples (RPB3B and RPB3C) were collected. The operating conditions used for the analyses included an accelerating voltage of 15 kV and a beam current of 20 nA. The probe diameter for the mineral composition spot analyses was 3 µm. Natural and synthetic silicates and oxides were used as standards. The ZAF method was employed for matrix correction. Representative mineral compositions are listed in Data Repository Table DR1.

**Retrograded Eclogite**

Sample RPB3A is a retrograde eclogite mainly consisting of garnet, clinopyroxene, amphibole, biotite, plagioclase, quartz, and rutile. Garnet is characterized by an inclusion-rich core and an inclusion-free

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Figure 1. (A) Simplified geological map of the Fennoscandian shield showing the locations of the eclogites in the Belomorian mobile belt. (B) The tectonic division of the northeastern Fennoscandian shield region. Sample locations: 3A, 3B, and 3C: N67°28′32″/E32°22′39″; samples 1A and 1B: N67°31′07″/E32°21′03″. Modified from Berthelsen and Marker (1986), Zhao et al. (2002), and Daly et al. (2006). TTG—tonalite-trondhjemite-granodiorite.
rim (Fig. 3A). The main inclusions in the garnet core are amphibole, clinopyroxene, quartz, and rutile (Figs. 3B, 3C), with minor biotite. Most clinopyroxene crystals in the matrix are symplectite with plagioclase (Fig. 3D) and are calcic clinopyroxene (commonly diopside and minor augite). Omphacite is found as relics (Fig. 3D). Omphacite records jadeite contents of as much as 21%, whereas the symplectic diopside records low jadeite contents (3%–10%; Fig. 4A). Garnet is characterized by homogeneous cores [almandine (Alm41–44), pyrope (Prp33–34), grossular (Grs20–26), spessartine (SpS1)] and Mg-rich inner rims (Alm39–40Prp36–39Grs22–23SpS1) and outer rims (Alm41–42Prp35Grs22–23SpS1; Fig. 5A). A decrease in $X_{Fe}$ from 0.54 to 0.55 to 0.50–0.52 is observed toward the inner rim, and $X_{Fe}$ increases to 0.54 at the outer rim. All amphiboles within garnets and the matrix plot within the compositional field of pargasite (Fig. 4B) with an $X_{Fe}$ ratio of 0.08–0.26. Plagioclase has an anorthite content ($X_{An}$) of 0.21–0.35. Clinopyroxene records low Fe$^{3+}$/(Fe$^{3+}$ + Al) ratios (0.02–0.03).

Samples RPB1A and RPB1B are also retrogressed eclogites that are relatively dark in color and consist of garnet, clinopyroxene, amphibole, plagioclase, quartz, and rutile with minor biotite. The retrogressed eclogites include many subhedral garnet porphyroblasts ranging in size from 3 to 10 mm. The main inclusions in the garnet are amphibole, clinopyroxene, quartz, and rutile. Subhedral clinopyroxene is dominant within the matrix, and plagioclase lamellae were found within the clinopyroxene in sample RPB1A (Fig. 3E). The symplectites of calcic clinopyroxene + plagioclase and amphibole + plagioclase around garnet due to the breakdown of omphacite during granulite facies metamorphism (Fig. 3F). In contrast, the formation of coronitic plagioclase is related to the replacement of garnet rims during amphibolite facies metamorphism (Fig. 3F). In sample RPB1A, the jadeite contents of the clinopyroxene in the garnet and matrix range from 11% to 18% and 15% to 21%, respectively, whereas the jadeite content of symplectic clinopyroxene is 9% (Fig. 4C). In sample RPB1B, clinopyroxene in garnet records jadeite contents ranging between 14% and 27%, whereas the clinopyroxene in the symplectite records jadeite contents ranging between 5% and 10% (Fig. 4C). The garnets in both samples have homogeneous cores (Alm41–43Prp36–38Grs18–21SpS1 with $X_{Fe}$ values of 0.52–0.54 in sample RPB1A; Alm44–46Prp29–32Grs23–25SpS1 with $X_{Fe}$ values of 0.59–0.61 in sample RPB1B) and relatively Fe-rich outermost rims (with $X_{Fe}$ ratios of 0.57–0.58 in sample RPB1A and $X_{Fe}$ ratios of 0.63–0.64 in sample RPB1B) (Fig. 5B). The $X_{Fe}$ ratio in garnet increases toward the rim, thus reflecting retrograde metamorphism (Fig. 5B). The amphibole inclusions in garnet are magnesiohornblende with $X_{Fe}$ contents of 0.11–0.12 in sample RPB1A and paragisite with $X_{Fe}$ values of 0.23–0.29 in sample RPB1B (Fig. 4D). In contrast, amphibole in contact with garnet records a wide compositional range, from magnesiohornblende to paragisite, with $X_{Fe}$ contents of 0.26–0.38 in sample RPB1A and 0.21–0.29 in sample RPB1B (Fig. 4D). The $X_{Fe}$ values of plagioclase are 0.18–0.31 in sample RPB1A and 0.23–0.30 in sample RPB1B.
Garnet-Clinopyroxene Granulite

Samples RPB3B and RPB3C are garnet-clinopyroxene granulites consisting of garnet, clinopyroxene, amphibole, plagioclase, and quartz. Clinopyroxene crystals in the 2 granulites are diopside with jadeite contents of <5% (Fig. 4A). In sample RPB3B, the composition of the garnet core is Alm$_{34-36}$Prp$_{13-15}$Grs$_{26-29}$Sps$_{1}$; the almandine component increases at the rim (Alm$_{42-44}$Prp$_{27-29}$Grs$_{25-27}$Sps$_{2}$), thus exhibiting retrograde zoning. The garnet in sample RPB3C also exhibits retrograde zoning. The garnet core has a composition of Alm$_{42-44}$Prp$_{27-29}$Grs$_{26-29}$Sps$_{1}$, whereas the garnet rim has an Fe-rich composition of Alm$_{42-44}$Prp$_{27-29}$Grs$_{26-29}$Sps$_{2}$. The values of X$_{Fe}$ increase toward the rim from 0.59 to 0.62 and to 0.66–0.68 in RPB3B and RPB3C. The amphiboles in sample RPB3B plot on the boundary between the compositional fields of magnesioclorite and pargasite (Fig. 4B). The X$_{Fe}$ values of amphibole in garnet (0.11–0.22) are slightly lower than those in amphiboles in contact with garnet (0.26–0.38). The amphiboles in sample RPB3C is magnesioclorite (Fig. 4B) with X$_{Fe}$ values of 0.19–0.30. The X$_{Fe}$ values of plagioclase in samples RPB3B and RPB3C are 0.43–0.65 and 0.33–0.36, respectively.

P-T ESTIMATES

Metamorphic Stages Based on Petrography

On the basis of microstructural observations and mineral relationships, several metamorphic stages have been recognized from the Salma eclogites. (1) Evidence of prograde metamorphism is preserved within garnet in sample RPB3A (Fig. 3A). The prograde assemblage is clinzoisite (Czo) + amphibole (Amp) + garnet (Grt) + biotite (Bt) + quartz (Qz) + rutile (Rt) (Whitney and Evans, 2010). (2) The omphacite in garnet in
sample RPB1B (Fig. 3F) and the relict omphacite in the matrix of sample RPB3A (Fig. 3D) formed during eclogite facies metamorphism. The Mg-rich garnet rim in sample RPB3A and the omphacite-bearing garnet core in sample RPB1B grew during this stage. The omphacite-bearing garnets from sample RPB1B also include amphibole, quartz, and rutile. The leucoxosome associated with eclogite within the outcrop (Fig. 2D) indicates that melt existed during metamorphism. The inferred peak assemblage consists of omphacite (Omp) + Amph + Qtz + Rt + melt ± Bt. The first overprint resulted in the growth of plagioclase and Ca-clinopyroxene, producing the mineral assemblage Ca-Cpx + Amph + Grt + Qtz + Rt + Pl + melt ± Bt. The symplectite of Ca-clinopyroxene + plagioclase observed in all samples (e.g., Figs. 3D, 3F), as well as the plagioclase lamellae observed in the Ca-clinopyroxene of sample RPB1A (Fig. 3E), developed during this stage. (4) The secondary amphiboles replacing garnet rims (Figs. 3A, 3F) and surrounding the symplectites (Fig. 3D) represent an amphibolite facies overprint that formed during cooling. The coronitic plagioclase surrounding the garnet rims also formed during this stage, along with secondary amphibole.

Figure 4. Clinopyroxene compositions plotted on the wollastonite + enstatite + ferrosilite (WoEnFs)–jadeite–aegirine (Jd-Aeg) diagram of Morimoto (1988) (grt—garnet) for (A) samples RPB3A–RPB3C and (B) samples RPB1A, RPB1B. Classification of Ca-amphiboles according to Hawthorne et al. (2012) based on a (Na + K + 2Ca) in A site (Al + Fe$^{3+}$ + 2Ti) in C site diagram for (C) samples RPB3A–RPB3C and (D) samples RPB1A, RPB1B.
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Figure 5. Representative compositional zoning profiles of garnet in (A) sample RPB3A and (B) sample RPB1B. Abbreviations: alm—almandine; pyp—pyrope; grs—grossular; sps—spessartine.

Conventional Geothermobarometry

The $P$-$T$ conditions of the different metamorphic stages of the metabasites were estimated using the garnet-clinopyroxene and garnet-amphibole geothermometers (Ellis and Green, 1979; Graham and Powell, 1984) and the garnet-clinopyroxene-plagioclase-quartz and garnet-amphibole-plagioclase-quartz geobarometers (Powell and Holland, 1988; Kohn and Spear, 1990). The $P$-$T$ results obtained using conventional geothermobarometry are listed in Table 2.

The homogeneous garnet cores from sample RPB3A include clinzoisite; their compositions are relatively Mg poor compared to their inner rim compositions. These data imply that the homogeneous cores formed during the prograde stage. The metamorphic temperatures estimated using the compositions of the amphibole inclusions in garnet and the adjacent garnet cores in sample RPB3A are ~610–660 °C, which represent the $P$-$T$ conditions during prograde metamorphism. However, the uncertainties of the estimated temperatures may be high, due to the possibility of compositional changes occurring during peak or retrograde metamorphism. Sample RPB3A records Mg-rich inner rims in garnet and relict omphacite, and the garnet core in sample RPB1B includes omphacite. Thus, we infer that these garnet and omphacite compositions are related to eclogite facies metamorphism. They yield peak eclogite-stage temperatures ranging from 730 to 810 °C, assuming a pressure of 17 kbar, as obtained using compositional isopleths in the pseudosection (see later section). The $P$-$T$ conditions of the first overprint (11.5–12.5 kbar and 770 °C), which indicate granulite facies metamorphism, were inferred from the compositions of garnet rims and the Ca-clinopyroxene and plagioclase located near the garnet in sample RPB1A. The retrograde $P$-$T$ conditions of amphibolite facies metamorphism (8.0–10.0 kbar and 590–610 °C) were calculated using the compositions of garnet rims and the plagioclase and amphibole surrounding the garnet in samples RPB3A, RPB1A, and RPB1B.

$P$-$T$ Pseudosection

The $P$-$T$ pseudosections for samples RPB3A and RPB1B were calculated in the modal chemical system NCKFMASHTO (Na$_2$O-CaO-K$_2$O-Fe$_2$O$_3$-MgO-Al$_2$O$_3$-SiO$_2$-TiO$_2$-O$_2$) using the Perplex X program (Connolly, 1990) with an internally consistent thermodynamic data set (Holland and Powell, 1998; updated in 2002). The following solid-solution models were used in these calculations: garnet (White et al., 2000), biotite (Tajcmanová et al., 2009), plagioclase (Newton et al., 1980), K-feldspar (Waldbaum and Thompson, 1968), clinopyroxene and orthopyroxene (Holland and Powell, 1996), phengite (parameters from thermodynamic dataset of Powell and Holland, 1988), amphibole (Dale et al., 2005), and melt (White et al., 2001). All fluid was assumed to be H$_2$O; its content was obtained from the values of weight loss on ignition. The ferrous/ferric ($\text{Fe}^2+/\text{Fe}^{3+}$) ratio was determined by using the titration of FeO to calculate the O$_2$ component. The haplogranitic melt of White et al. (2001) may not always be appropriate for modeling partial melting in metabasites. As a result, it is possible that the assemblages containing melt in the calculated pseudosections may be metastable. However, the topology of the phase relationship between amphibolite facies and granulite facies metabasites does not significantly change when mineral assemblages coexist with fluids or melts (Pattison, 2003). The pseudosection approach is useful for inferring mineral assemblages including melt for the Salma eclogites. The bulk-rock compositions used in the pseudosection calculations were analyzed using inductively coupled plasma–mass spectrometry (ICP-MS; Perkin Elmer Optima 3000) at Activation Laboratories Ltd. (Canada).

The effective bulk composition was possibly modified by the growth of zoned garnet due to crystal fractionation (e.g., Evans, 2004). Because garnet in sample RPB3A displays prograde zoning with homogeneous cores and Mg-rich inner rims, its effective bulk composition was calculated. First, the pseudosection used to estimate prograde $P$-$T$ conditions was constructed using the bulk composition determined from the ICP-MS analyses (Fig. 6). Second, the modal percentage of garnet cores within the rock was estimated (~10 vol%), and the composition of the garnet cores was subtracted from the bulk chemical data (Konrad-Schmolke et al., 2008). The recalculated bulk composition was used for the pseudosection in order to estimate the peak and retrograde $P$-$T$ conditions (Fig. 7). The molar proportion of the unfractiected and effective bulk rock composition used for the pseudosection modeling is shown in the captions for Figures 6 and 7.

The $P$-$T$ pseudosection constructed using the actual measured bulk composition indicates that the prograde assemblage of zoisite (Czo) + Amp + Grt + Bt + Qz + Rt + H$_2$O occurs at $P$-$T$ conditions of 13–18 kbar and 640–720 °C (Fig. 6). This $P$-$T$ range can be considered reasonable...
Figure 6. Pressure-temperature (P-T) pseudosection of sample RPB3A calculated in the NCKFMASHTO (+ Amp, Rt) system. The bulk compositions (mol%) used are SiO₂ (45.15), TiO₂ (0.33), Al₂O₃ (8.95), FeO (7.46), MgO (16.85), CaO (19.59), Na₂O (2.01), K₂O (0.13), O₂ (1.03), and H₂O (7.49). The prograde assemblage is shown in italics. Bt—biotite, Chl—chlorite, Ilm—ilmenite, Opx—orthopyroxene, Zo—zoisite, Ky—kyanite, Ph—phengite. Abbreviations as in Figure 3.
because it matches with the temperature estimate (610–660 °C) in the conventional geothermobarometry section and overlaps that of the epidote-amphibolite facies on the metamorphic facies diagram delineated by Oh and Liou (1998). The eclogite facies assemblage of Omp + Amp + Grt + Bt + melt + Qz + Rt yields P-T conditions of 13–20 kbar and 730–820 °C (Fig. 7). The compositional isopleths of \( X_{Na} \) in clinopyroxene and \( X_{Fe} \) in garnet are shown in Figure 7. The isopleths of the inner rims of garnet are shown in Figure 7. The isopleths of the inner rims of garnet (\( X_{Na} = 0.50–0.52 \)) and relict omphacite (\( X_{Na} = 0.21–0.23 \)) constrain the peak P-T conditions to 17–18 kbar and 750–770 °C, which are within the range inferred from the mineral phase assemblages and within the temperature estimate (730–800 °C) in the conventional geothermobarometry section. The plagioclase-forming reaction occurred at conditions of 9–14 kbar and 700–850 °C, thus indicating that the growth of plagioclase with Ca-clinopyroxene occurred during decompression under granulate facies conditions. The granulate facies assemblage of Ca-Cpx + Amp + Grt + Qz + Rt + plagioclase (Pl) + melt + Bt constrains the P-T conditions to 12–14 kbar and 780–820 °C (Fig. 7). As a result, a clockwise P-T path including rapid uplift was estimated.

The garnet core in sample RPB1B is homogeneous, and the pseudosection modeling was done without considering the effective bulk-rock composition (Fig. 8). This bulk composition has relatively higher \( \text{SiO}_2 \), \( \text{Na}_2\text{O} \), and \( \text{CaO} \) contents than sample RPB3A. The calculated P-T pseudosection shows an increase in the stability of plagioclase, which is stable at P-T conditions below 12–16 kbar at 700–850 °C (Fig. 8). The melt stability field occurs at temperatures above ~680–700 °C. The eclogite facies assemblage of Omp + Grt + Bt + melt + Qz + Rt + Amp yields P-T conditions of 10–15 kbar and 720–800 °C. The melt assemblage at peak P-T conditions of 15–18 kbar and 740–790 °C (Fig. 8). The isopleths of \( X_{Na} = 0.27–0.28 \) for omphacite in garnet and \( X_{Fe} = 0.58–0.60 \) for garnet cores yield peak P-T conditions of 16–17 kbar and 750–770 °C (Fig. 8), consistent with those of sample RPB3A. The granulate facies assemblage of Ca-Cpx + Amp + Grt + Qz + Rt + Pl + melt plots within the P-T conditions of 10–15 kbar and 720–800 °C. The isopleths of the \( X_{Na} = 0.63–0.64 \) of garnet cores and the maximum \( X_{Fe} = 0.10 \) of clinopyroxene symplectite yield P-T conditions of 10–11 kbar and 790–820 °C, indicating that they reflect a rapid uplifting P-T path (Fig. 8).

**WHOLE-ROCK CHEMISTRY**

The major and trace element contents of the five studied metabasites were analyzed using an ICP-MS (Perkin Elmer Optima 3000, Activation...
were originally low-alkali tholeiitic basalts, as shown in Figure 9A. They fields on the Ti/100-Zr-Y and Nb-Zr/4-Y diagrams (Figs. 9B, 9C).

O₂ (0.86), and H₂O (3.77). The peak and retrograde assemblages are shown in bold and white letters, respectively. Compositional isopleths of garnet for XFe and clinopyroxene for XNa from sample RPB1B are also shown. The bold and dashed circles represent the peak and retrograde respectively. Abbreviations are as in Figures 3 and 6.

Pseudosection of sample RPB1B calculated in the NCKFMASHTO (Na₂O-CaO-K₂O-FeO-MgO-Al₂O₃-SiO₂-H₂O-TiO₂-O₂) system. The bulk compositions (mol%) used are SiO₂ (52.81), TiO₂ (1.14), Al₂O₃ (8.46), FeO (9.87), MgO (11.14), CaO (10.45), Na₂O (1.46), K₂O (0.04), O₂ (0.86), and H₂O (3.77). The peak and retrograde assemblages are shown in bold and white letters, respectively. Compositional isopleths of garnet for XFe and clinopyroxene for XNa from sample RPB1B are also shown. The bold and dashed circles represent the peak and retrograde P-T conditions, respectively. Abbreviations are as in Figures 3 and 6.

Figure 8. Pressure-temperature (P-T) pseudosection of sample RPB1B calculated in the NCKFMASHTO (Na₂O-CaO-K₂O-FeO-MgO-Al₂O₃-SiO₂-H₂O-TiO₂-O₂) system. The bulk compositions (mol%) used are SiO₂ (52.81), TiO₂ (1.14), Al₂O₃ (8.46), FeO (9.87), MgO (11.14), CaO (10.45), Na₂O (1.46), K₂O (0.04), O₂ (0.86), and H₂O (3.77). The peak and retrograde assemblages are shown in bold and white letters, respectively. Compositional isopleths of garnet for XFe and clinopyroxene for XNa from sample RPB1B are also shown. The bold and dashed circles represent the peak and retrograde P-T conditions, respectively. Abbreviations are as in Figures 3 and 6.

Laboratories Ltd., Canada). The whole-rock data are listed in Data Repository Table DR2. The five metabasites have basaltic compositions, with SiO₂ contents of 47.0–52.4 wt% and low alkali concentrations in terms of Na₂O (1.49–2.16 wt%) and K₂O (0.02–0.33 wt%). These metabasites were originally low-alkali tholeiitic basalts, as shown in Figure 9A. They plot in the island arc basalt (IAB) and mid-oceanic ridge basalt (MORB) fields on the Ti/100-Zr-Y and Nb-Zr/4-Y diagrams (Figs. 9B, 9C).

The REE patterns normalized using C1 chondritic values are basically depleted in light REEs, thus yielding flat patterns showing an affinity to normal (N) MORB (Figs. 10A, 10B). Sample RPB1B exhibits a negative Eu anomaly. The incompatible trace element abundances normalized using primitive mantle values are shown in Figures 10C and 10D. The elements ranging from Nd to Yb produce low and flat trends in most samples, except for sample RPB1B. In contrast, the elements from Rb to Sr, which are highly incompatible, record variable values, which were likely produced by disturbances during subduction. These whole-rock data indicate that the origin of the Salma eclogites is likely subducted oceanic crust that originated at a spreading center. This interpretation agrees well with those of previous studies (Mints et al., 2010, 2014; Konilov et al., 2011).

ZIRCON U-Pb AGES AND GEOCHEMISTRY

Analytical Procedure

Zircon grains from the two eclogite samples (RPB3A and RPB1B) were separated using the standard heavy liquid technique and were then hand-picked under a binocular microscope. Cathodoluminescence (CL) and backscattered electron (BSE) images were obtained using the JEOL 6610LV scanning electron microscope at the Korea Basic Science Institute (KBSI, Ochang, South Korea). The CL and BSE images from samples RPB3A and RPB1B are shown in Figures 11 and 12, respectively. Microinclusions in zircons were identified using the scanning electron
Figure 9. The whole-rock compositions of retrograded eclogite and granulite in the Salma area. (A) Plotted on the FeO_total-(Na_2O + K_2O)-MgO classification diagram. (B) Plotted on the Ti/100-Zr-Y Y^3 tectonic discrimination diagram. IAT—island-arc tholeiites; CAB—calc-alkaline basalts; WPB—within-plate basalts; MORB—mid-oceanic ridge basalt. (C) Plotted on the Nb Zr/4-Y tectonic discrimination diagram. AI—within-plate alkali basalts; AII—within-plate alkali basalts and within-plate tholeiites; B—enriched-type MORB; C—within-plate tholeiites and volcanic-arc basalts; D—normal-type MORB and volcanic-arc basalts (Rollinson, 1993, and references therein).

Figure 10. Chondrite-normalized rare earth element patterns. MORB—mid-oceanic ridge basalt (N—normal; E—enriched); OIB—oceanic island basalts for (A) samples RPB3A–RPB3C and (B) samples RPB1A, RPB1B. Primitive mantle-normalized trace element patterns for (C) samples RPB3A–RPB3C and (D) samples RPB1A, RPB1B. Data are normalized using the values of chondrite and primitive mantle from Sun and McDonough (1989).
Figure 11. (A–F) Representative cathodoluminescence (CL) and backscattered electron (BSE) images of dated zircon crystals in sample RPB3A. The analyzed spots are shown with their $^{207}\text{Pb}/^{206}\text{Pb}$ ages and spot numbers. Abbreviations as in Figure 3.

Figure 12. (A–I) Representative cathodoluminescence (CL) and backscattered electrons (BSE) images of dated zircon crystals in sample RPB1B. The analyzed spots are shown with their $^{207}\text{Pb}/^{206}\text{Pb}$ ages and spot numbers. Abbreviations as in Figure 3.
microscope with energy-dispersive X-ray spectroscopy (SEM-EDX) detector at the KBSI and the Thermo Scientific DXR micro-Raman microscope equipped with a 532 nm laser at the Tectonophysics Laboratory in the School of Earth and Environmental Sciences (Seoul National University). The EDX spectra obtained from the inclusions in the zircon were used to identify mineral inclusions by comparing them with those of minerals in thin sections within the same samples. The mineral inclusion assemblages in the zircons are listed in Table 3, and the EDX spectra of garnet, omphacite, and Ca-clinoxyroxene inclusions in zircons are shown in Data Repository Figure 1.

The REE composition of zircon was analyzed using laser ablation (LA)-ICP-MS at the KBSI. The LA-ICP-MS system consists of a laser ablation system (213 nm Nd-YAG [neodymium-doped yttrium aluminum garnet laser] UP213, New Wave Research, a division of ESI), ICP, and a quadrupole mass spectrometer (X2 series, Thermo Scientific). The analytical procedures for the REE analyses of the zircons followed those of Yuan et al. (2004) and Liu et al. (2007). Ablation signals were collected by ICP-MS using a time-resolved analysis of 45 s. The Nd-YAG laser was operated at a repetition rate of 10 Hz, a spot size of 55 mm, and an energy level of 80% (27 J/cm²). NIST 612 glass was used as an external calibration standard, and each analysis was normalized to the silicon content (29Si) as an internal standard. GLITTER software (http://www.glitter-gemoc.com/) was used for data reduction. Zircon REE data are given in Data Repository Table DR3.

The zircon U-Pb ages were analyzed using the SHRIMP (sensitive high-resolution ion microprobe) IIe ion microprobe at the KBSI. The analytical procedures for SHRIMP dating were mainly based on those proposed by Williams (1998). A spot size of 15–20 µm and a 1.5–2 nA negative ion oxygen beam (O−) were used for all analyses. The measured 206Pb/238U ratio was calibrated using the FC1 zircon standard (ca. 1099 Ma; Paces and Miller, 1993). The SL13 zircon standard was also used for the calibration of U concentrations (238 ppm; Hoskin, 1998). Data reduction, age calculations, and common Pb corrections were conducted using SQUID 2.50 (Ludwig, 2009) and Isoplot 3.6 software (Ludwig, 2008). The zircon U-Pb ages are listed in Data Repository Table DR4.

**Results**

**Sample RP3A**

The zircon grains from sample RP3A are subhedral and have sub-rounded edges (Figs. 11A–11F). Most zircon grains have dark CL cores surrounded by pale gray CL mantles with sector or patchy zoning (Figs. 11A, 11B). In the BSE images of these grains, the cores are brighter than the mantles (Fig. 11C). However, some zircons have pale gray CL cores with sector or patchy zoning that are similar to the mantles surrounding the dark CL cores (Figs. 11D, 11E). A few grains have bright CL cores surrounded by pale gray CL mantles (Fig. 11F). Thin, bright CL rims are locally observed surrounding pale gray CL cores and mantles (Figs. 11E, 11F). The dark CL cores contain inclusions of garnet, amphibole, plagioclase, quartz, biotite, and rutile; the pale gray CL cores and mantles contain inclusions of garnet (Fig. 11F), amphibole, biotite, and quartz (Table 3; Data Repository Item). K-feldspar occurs along fractures in garnet laser analysis by SQUID 2.50 (Ludwig, 2009) and Isoplot 3.6 software (Ludwig, 2008). The zircon U-Pb ages are listed in Data Repository Table DR4.

![Chondrite-normalized rare earth element patterns for different zircon domains. CL—cathodoluminescence.](image)

**Figure 13.** Chondrite-normalized rare earth element patterns for different zircon domains. CL—cathodoluminescence. (A) Sample RP3A. (B) Sample RP1B. Data are normalized using the chondritic values of Sun and McDonough (1989).
Figure 14. Concordia diagrams for the sensitive high-resolution ion microprobe (SHRIMP) U-Pb analyses of zircon. The dashed line represents the discordia line. All error ellipses are quoted at the 1σ level. MSWD—mean square of weighted deviates. The mean and discordia ages are shown at the 2σ level. CL—cathodoluminescence. (A) Sample RPB3A. (B) Sample RPB1B.
The omphacite-bearing zircon domains yield ages of 1863 ± 56 Ma (Fig. 12D), omphacite (Fig. 12H), quartz, amphibole, rutile, and apatite occur in Sample RPB1B at 2.87 and 2.82 Ga based on the intrusions of mafic dikes at 2.86–2.83 Ga and felsic veins at 2.82–2.78 Ga. Mints et al. (2014) believed that high-pressure metamorphism, which formed the Salma eclogite, may have occurred during the subduction stage; however, the Archean age of the eclogite facies metamorphism in the Salma area is uncertain, because they did not provide direct evidence for it.

The samples obtained from the dark CL zircons in samples RPB3B and RPB1B are interpreted to represent an Archean amphibolite facies metamorphic event. The unzoned regions present in the dark CL zircons are generally produced by metamorphism. We also found a representative metamorphic mineral assemblage (i.e., garnet) in the dark CL zircons from sample RPB3A. Although high to moderate Th/U ratios in the dark CL zircons (0.2–5.1 for RPB3A, 0.2–0.3 for RPB1B) may indicate their magmatic origins (cf. Skublov et al., 2010a), metamorphic zircons with high Th/U ratios have also been reported in high-grade rocks (e.g., Harlow et al., 2007).

Some researchers reported 2.72–2.70 Ga (retrograde) granulite facies metamorphism (Kaulina et al., 2010; Mints et al., 2010). Although there is no direct petrological evidence to link these ages to granulite facies metamorphism, the REE pattern with HREE enrichment in zircons was explained by zircon growth in equilibrium with melt during granulite facies metamorphism (Kaulina et al., 2010). However, melt can be produced from temperatures of 680–700 °C, which correspond to conditions of upper amphibolite facies metamorphism in bulk-rock composition of the Salma eclogites (Figs. 7 and 8). Metamorphic zircons have been reported in amphibolites in the orogeny (e.g., Oh et al., 2014), thus implying that (upper) amphibolite facies metamorphism can produce abundant zircons. In this study, the mineral inclusions within the dark CL zircons with 2.73–2.72 Ga ages are characterized by an amphibolite facies mineral assemblage (Grt + Amp + Pl + Qz + Rt ± Bi). The relatively enriched HREE patterns and remarkably negative Eu anomalies observed in the dark CL zircons, compared to the pale gray zircons, can be explained by the presence of less garnet and abundant plagioclase. These indicate that amphibolite facies metamorphism occurred ca. 2.73–2.72 Ga.

Interpretation of Paleoproterozoic Zircon Ages

The pale CL zircons from samples RPB3A and RPB1B that contain garnet and omphacite yield 207 Pb/206 Pb age mean ages of 1865 ± 15 Ma and 1868 ± 17 Ma, respectively. Direct evidence for Paleoproterozoic eclogite is provided by 207 Pb/206 Pb ages of 1863 ± 56 and 1850 ± 36 Ma from omphacite-bearing zircon domains. These results indicate that eclogite facies metamorphism (16–18 kbar and 750–770 °C) occurred ca. 1.87 Ga. The flat HREE patterns indicate that these metamorphic zircons formed in equilibrium with garnet during eclogite facies metamorphism (e.g., Rubatto, 2002; Whitehouse and Platt, 2003; Imayama et al., 2012). Metamorphic zircons characterized by flat HREE patterns have been obtained from many eclogites around the world (e.g., northeast Greenland, Gilotti et al., 2004, McClelland et al., 2006; South Korea, Kim et al., 2006; Central Alps, Liati et al., 2009; Bohemian Massif, Bröcker et al., 2010). High HREE abundances in garnet produce flat HREE patterns in zircon. In addition, the very weak Eu anomalies in the pale gray CL zircons, compared to the dark CL zircons, indicate that plagioclase-free mineral assemblages exist in the eclogite. The presence of small Eu anomalies and Ca-clinopyroxene (instead of only omphacite) in the pale gray CL zircons may indicate that zircon growth continued to granulate facies during decompression. Nevertheless, the absence of plagioclase inclusions in the pale gray CL zircons means that the zircons yielding ages of ca. 1.87 Ga mainly grew during eclogite facies metamorphism.

Skublov et al. (2010a, 2011) reported that eclogite facies metamorphism occurred ca. 1.92–1.88 Ga, based on the analysis of metamorphic rock types from the Salma eclogite. These ages are consistent with the 2.73–2.72 Ga ages obtained from the dark CL zircons, suggesting that the Salma eclogite experienced a single phase of Paleoproterozoic high-pressure metamorphism.

Sample RPB1B

The zircon grains from sample RPB1B are subhedral with rounded or subrounded edges (Figs. 12A–12D). Three domains are observed on the basis of CL and BSE images. Most crystals have dark CL cores, which are surrounded by pale gray CL mantles (Figs. 12A, 12B, 12D). The BSE images show brighter cores and relatively darker mantles (Fig. 12C). Several zircons have pale gray CL cores with sector or patchy zoning that are similar in their CL brightness to the mantles surrounding the dark CL cores (Figs. 12E, 12F). In some zircons, the internal zoned structure of the pale gray cores resembles a cloudy zoning pattern (Figs. 12G–12I). These pale gray cores are surrounded by brighter CL outer rims, which vary in thickness from narrow (Figs. 12E–12H) to broad (Fig. 12I). Apatite occurs as inclusions in the dark CL cores, whereas Ca-clinopyroxene (Fig. 12D), omphacite (Fig. 12H), quartz, amphibole, rutile, and apatite occur in the pale gray CL domains (Table 3; Data Repository Fig. 1). The dark CL zircons display HREE-enriched patterns with a steep slope from the middle to the concordant data is 2727 ± 8 Ma (MSWD = 0.6, n = 8, 2σ). The dark CL zircons have flat HREE patterns with moderate to shallow slopes from the middle REEs to the HREEs (Lu66/ Gand82 = 23.3–34.3). They also record negative Eu anomalies (Eu/Eu* = 0.21–0.30; Fig. 13B). The 207 Pb/206 Pb ages of the dark CL cores are 2757 ± 2324 Ma. The weighted mean 207 Pb/206 Pb age of the concordant data is 2727 ± 8 Ma (MSWD = 0.6, n = 8, 2σ). The pale gray CL zircons from samples RPB3A and RPB1B are interpreted to represent an Archean amphibolite facies metamorphic event. The unzoned regions present in the dark CL zircons are generally produced by metamorphism. We also found a representative metamorphic mineral assemblage (i.e., garnet) in the dark CL zircons from sample RPB3A. Although high to moderate Th/U ratios in the dark CL zircons (0.2–5.1 for RPB3A, 0.2–0.3 for RPB1B) may indicate their magmatic origins (cf. Skublov et al., 2010a), metamorphic zircons with high Th/U ratios have also been reported in high-grade rocks (e.g., Harlow et al., 2007).

Some researchers reported 2.72–2.70 Ga (retrograde) granulite facies metamorphism (Kaulina et al., 2010; Mints et al., 2010). Although there is no direct petrological evidence to link these ages to granulite facies metamorphism, the REE pattern with HREE enrichment in zircons was explained by zircon growth in equilibrium with melt during granulite facies metamorphism (Kaulina et al., 2010). However, melt can be produced from temperatures of 680–700 °C, which correspond to conditions of upper amphibolite facies metamorphism in bulk-rock composition of the Salma eclogites (Figs. 7 and 8). Metamorphic zircons have been reported in amphibolites in the orogeny (e.g., Oh et al., 2014), thus implying that (upper) amphibolite facies metamorphism can produce abundant zircons. In this study, the mineral inclusions within the dark CL zircons with 2.73–2.72 Ga ages are characterized by an amphibolite facies mineral assemblage (Grt + Amp + Pl + Qz + Rt ± Bi). The relatively enriched HREE patterns and remarkably negative Eu anomalies observed in the dark CL zircons, compared to the pale gray zircons, can be explained by the presence of less garnet and abundant plagioclase. These indicate that amphibolite facies metamorphism occurred ca. 2.73–2.72 Ga.

DISCUSSION

Meaning of Archean Zircon Ages

Petrographic, geochemical, and geochronological data from the Salma eclogites in the Kola Peninsula reveal the polymetamorphic history of this area. The ages of the magmatic protoliths of the Salma eclogites are known to be 2.94–2.92 Ga (Kaulina et al., 2010) and 2.88–2.82 Ga (Mints et al., 2010; Skublov et al., 2010a, 2011; Mel’nik et al., 2013). However, the 2.88–2.87 Ga zircon age obtained from Skublov et al. (2010a, 2011) should be interpreted as a metamorphic age, rather than as a magmatic age, because these zircons do not record zoning patterns that are typical for igneous zircons, such as concentric or banded zoning (Hoskin and Schaltegger, 2003); instead, they exhibit unzoned patterns, which are characteristic of metamorphic zircons. Mints et al. (2014) also interpreted 2.82 Ga to be the earliest age of metamorphism based on 176 Hf/177 Hf isotopic ratios, and suggested that subduction occurred in the Salma area between 2.87 and 2.82 Ga based on the intrusions of mafic dikes at 2.86–2.83 Ga and felsic veins at 2.82–2.78 Ga. Mints et al. (2014) believed that high-pressure metamorphism, which formed the Salma eclogite, may have occurred during the subduction stage; however, the Archean age of the eclogite facies metamorphism in the Salma area is uncertain, because they did not provide direct evidence for it.
zircon rims surrounding Archean magmatic zircon cores (ca. 2.88–2.87 Ga) within the massive eclogite (sample 46, Table 1). However, the ca. 1.92–1.88 Ga metamorphic zircons include zoisite and quartz (Skublov et al., 2010a, 2011), which appear to represent prograde metamorphism, rather than peak eclogite facies metamorphism, which is represented by the presence of garnet and omphacite. Garnet in the analyzed eclogite records prograde zoning, with increasing pyrope contents from core to rim (Skublov et al., 2011). Studies of Lu-Hf and Sm-Nd garnet geochemistry from the same eclogite yielded garnet–whole-rock isochron ages of 1901 ± 5 Ma and 1897 ± 16 Ma, respectively (Herwartz et al., 2012; Mel’nik et al., 2013). Although these ages were interpreted to reflect the timing of peak eclogite facies metamorphism, the Lu-Hf and Sm-Nd ages of garnets with growth zoning are generally interpreted to represent prograde metamorphic ages, due to the low diffusivities of REEs (Baxter and Scherer, 2013). The omphacite-bearing metamorphic zircons that formed ca. 1.87 Ga found in this study provide the first clear age of peak eclogite facies metamorphism in the Salma eclogite.

**P-T Path During Paleoproterozoic Metamorphism**

In the Salma eclogites, the identification of epidote-amphibolite facies prograde metamorphism was based on the presence of clinzoisite and amphibole inclusions in garnet cores and prograde zoned garnets with homogeneous cores and Mg-rich inner rims in sample RPB3A. Because the garnet in sample RPB1A only contains amphibole inclusions but lacks epidote (Fig. 4D), some Salma eclogites may have undergone amphibolite facies metamorphism prior to eclogite facies metamorphism. The P-T conditions of the prograde stage are estimated to be 13–18 kbar and 640–720 °C. These results closely match with the boundary between the amphibolite, epidote-amphibolite, and eclogite facies on the metamorphic facies diagram developed by Oh and Liou (1998). This prograde metamorphism likely occurred ca. 1.92–1.88 Ga (Skublov et al., 2010a, 2011; Herwartz et al., 2012; Mel’nik et al., 2013), as mentioned herein.

The granulite facies overprinting (10–14 kbar and 770–820 °C) occurred during the subsequent exhumation stage from 17 to 18 kbar, leading to the breakdown of omphacite to Ca-clinopyroxene and plagioclase. Because the zircon growth ca. 1.87 Ga could have continued to undergo granulite facies metamorphism, this probably occurred soon after the eclogite facies metamorphism, thus implying that rapid decompression occurred. Upon cooling, the amphibolite facies overprint occurred at conditions of 8–10 kbar and 590–610 °C.

**Tectonic Implications**

Some researchers have interpreted the eclogites in the Belomorian mobile belt to represent evidence of Archean subduction followed by collision, leading to the amalgamation of the Karelia craton, the Kola craton, and the microcontinent between them (Mints et al., 2010, 2014). However, in this study, an age of ca. 1.87 Ga for eclogite facies metamorphism was obtained from zircons with omphacite inclusions and flat HREE patterns within the Salma eclogite. The Paleoproterozoic zircons collected from the eclogites from the Kuru-Vaara quarries also exhibit flat HREE patterns (Skublov et al., 2011) and the P-T paths of the eclogites from the Grindino and Salma areas are similar (Mints et al., 2014), indicating that the eclogites in the Belomorian mobile belt formed during the Paleoproterozoic; these findings do not support the Archean subduction-collision model. The Archean subduction-collision model is not able to explain the regional occurrence of Paleoproterozoic granulite facies metamorphism (i.e., the Lapland and Umba belts) in the Kola-Karelian collisional zone (cf. Daly et al., 2006). The 1.87 Ga metamorphic age of the Salma eclogite in this study supports the model of Paleoproterozoic collision between the Kola and Karelian cratons suggested by Berthelsen and Marker (1986), Zhao et al. (2002), and Daly et al. (2006).

Eclogites that formed within a transitional eclogite-granulite facies P-T range could have formed in a deep continental crustal root zone (e.g., De Paoli et al., 2009). However, the whole-rock chemistry of the Salma eclogites in this study is characterized by depleted light REEs, which reflects their origins as N-MORB and is consistent with the results of previous studies (Konilov et al., 2011; Mints et al., 2014). It is likely that the 2.94–2.93 Ga ages inferred from the magmatic zircon in the Salma eclogite represent the protolith age of the N-MORB (Kaulina et al., 2010). This study also indicates that the Salma eclogites underwent an amphibolite facies metamorphic event ca. 2.73–2.72 Ga. During the Paleoproterozoic subduction of the unit including the protolith of the Salma eclogites, these rocks underwent progressive metamorphism from epidote-amphibolite facies ca. 1.92–1.88 Ga to eclogite facies ca. 1.87 Ga. This subduction stage was followed by the collision of the Kola and Karelian cratons; the unit including the Salma eclogites was rapidly uplifted and recorded strong overprinting produced first by granulite facies metamorphism and then by amphibolite facies metamorphism during or after the collision.

**Secular Changes in the Geothermal Gradients of Subduction Zones**

Subduction in the Precambrian may have proceeded differently than modern subduction, due to the hotter conditions in the mantle of the early Earth (e.g., Davies, 1992); however, it is debatable when ancient subduction changed into modern subduction (e.g., Cawood et al., 2006; van Hunen and Moyen, 2012, and references therein). Determining the changes in the patterns of metamorphism at plate boundary zones allows us to understand the evolutions and geodynamics of subduction zones (e.g., Brown, 2006, 2009). Determining the timing of the first appearances of high-pressure granulite, eclogite, and blueschist is thus very important for understanding the changes in the patterns of metamorphism and the geothermal gradient in Precambrian subduction zones over time.

The oldest high-pressure granulite is Neoarchean (ca. 2.5 Ga) and is present in the Jiaoping complex; it was produced by a subduction-collision event in the eastern region of the North China Craton (Wei et al., 2001; O’Brien and Rötzler, 2003; Liu et al., 2011; Lu et al., 2017). The P-T conditions of this high-pressure granulite metamorphic event were 10–13 kbar and 780–850 °C (Wei et al., 2001; Wang and Cui, 1992; Lu et al., 2017; Fig. 15). Eclogite facies metamorphism occurred at ca. 2.0 Ga in the Usagaran orogen of Tanzania, with peak metamorphic conditions of ~750 °C and 18 kbar (Möller et al., 1995; Collins et al., 2004). The peak metamorphic conditions and P-T paths of the Tanzanian eclogites are similar to those of the Salma eclogites (Fig. 15). These rocks underwent rapid decompression after peak metamorphism and were then retrograded into first granulites and then amphibolites. Because the eclogite facies metamorphism in the Salma eclogite and the Usagaran eclogite occurred during the Paleoproterozoic, subduction accompanying the development of eclogite likely began during or prior to the Paleoproterozoic. The oldest reported blueschist (ca. 750–730 Ma) is from the Aksu Group of western China (Liou et al., 1989, 1996; Nakajima et al., 1990; Maruyama et al., 1996; Zhu et al., 2011; Yong et al., 2013); the P-T conditions of this blueschist facies metamorphism were 4–10 kbar and 300–400 °C (Liou et al., 1989, Zhang et al., 1999; Fig. 15). The geothermal gradient required for a formation of high-temperature eclogite (>~750 °C) is higher than that of high-pressure granulite. Consequently, the first appearance of high-pressure granulite, eclogite, and blueschist in the Neoarchean, Paleoproterozoic, and Neoproterozoic, respectively,
may reflect a decrease in the geothermal gradients of subduction zones from the Neoarchean to the Neoproterozoic due to the cooling of the Earth.

Changes in metamorphic facies during prograde metamorphism in the Phanerozoic to Paleoproterozoic eclogites and Neoarchean high-pressure granulites can provide clearer information about changes in the geothermal gradients at subduction zones from the Neoarchean to the Phanerozoic. This study provides evidence that there was a prograde metamorphic event that involved epidote-amphibolite facies and/or amphibolite facies during the formation of the Paleoproterozoic Salma eclogites. Eclogites that have undergone low-grade epidote-amphibolite facies conditions during prograde metamorphism are known from several Phanerozoic subduction zones, such as the high- and ultrahigh-pressure eclogites of Sambagawa (Takasu, 1984; Enami et al., 1994; Itaya et al., 2011; Fig. 15). However, blueschist facies metamorphism mainly occurs prior to high- and ultrahigh-pressure eclogite facies metamorphism in most Phanerozoic subduction zones, such as the western Alps and New Caledonia (e.g., Oh and Liu, 1998; Rubatto and Hermann, 2001; Fig. 15).

Moreover, eclogites that formed via prograde metamorphism from amphibolite facies are almost absent in Phanerozoic subduction zones (Oh and Liu, 1998). However, the prograde mineral assemblage of the Neoarchean granulite in the Jianping complex is Amp + Pl ± Qz ± Bt, which represents amphibolite facies metamorphism (Wang and Cui, 1992; Liu et al., 2011). These data imply that Paleoproterozoic subduction zones were relatively warmer than Phanerozoic subduction zones but colder than Neoarchean subduction zones.

CONCLUSIONS

Based on the petrologic, thermobarometric, geochemical, and geochronological data presented in this study and previous studies, the following conclusions are proposed for the tectonothermal evolution of the Salma eclogites in the Kola Peninsula.

1. The source rocks for the Salma eclogites formed in a mid-ocean ridge ca. 2.94–2.93 Ga and first underwent amphibolite facies metamorphism ca. 2.73–2.72 Ga. This amphibolite facies metamorphism is confirmed by inclusions of Grt + Amp + Pl + Qz + Rt ± Bt in 2.73–2.72 Ga unzoned dark CL zircons, which are characterized by enriched HREE patterns and remarkably negative Eu anomalies.

2. The Salma eclogites may have undergone epidote-amphibolite facies or amphibolite facies prograde metamorphism at ca. 1.92–1.88 Ga. The ca. 1.87 Ga peak eclogite facies metamorphism can be inferred from the U-Pb age dating of pale gray CL metamorphic zircons with inclusions of Grt + Omp + Ca-Cpx + Amp + Qz + Rt ± Bt. These zircons record flat HREE patterns and nearly lack Eu anomalies. The peak metamorphic P-T conditions were ~16–18 kbar and 750–770 °C. Soon after this peak metamorphism, granulite facies metamorphism occurred after decompression at 10–14 kbar and 770–820 °C and was followed by amphibolite facies overprinting at 8–10 kbar and 590–610 °C during cooling.

3. The Paleoproterozoic subduction and subsequent continent-continent collision between the Kola and Karelian continents is supported by the occurrence of eclogite facies metamorphism at ca. 1.87 Ga.

4. The oldest appearances of high-pressure granulite, eclogite, and blueschist occurred in the Neoarchean, Paleoproterozoic, and Neoproterozoic, respectively, which may reflect a decrease in the geothermal gradients of Precambrian subduction zones due to the cooling of the Earth. The prograde metamorphism from epidote-amphibolite facies or amphibolite facies to eclogite facies in the subduction zone during the Paleoproterozoic also implies that the Paleoproterozoic subduction zone was relatively warmer than the Phanerozoic subduction zone but was colder than the eclogite-free Neoarchean subduction zones.

ACKNOWLEDGMENTS

We thank K. de Jong, Seoul National University, Korea, for helpful discussion; Y. Park, Seoul National University, Korea, for assistance with micro-Raman analyses; and Juhn G. Liu and two anonymous reviewers for constructive and critical reviews that significantly helped to improve the manuscript. We also thank R. Damian Nance for careful editorial handling. This work was supported by National Research Foundation of Korea (NRF) grants 657 NRF-2013R1A1A2005629, NRF-2014R1A2A2A01003052, and NRF-2017R1A2B2011224.

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