Multistage Magmatism in Ophiolites and Associated Metavolcanites of the Ulan-Sar’dag Mélange (East Sayan, Russia)

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Abstract: We present new whole-rock major and trace element, mineral chemistry, and U-Pb isotope data for the Ulan-Sar’dag mélange, including different lithostratigraphic units: Ophiolitic, mafic rocks and metavolcanites. The Ulan-Sar’dag mélange comprises a seafloor and island-arc system of remnants of the Paleo-Asian Ocean. Detailed studies on the magmatic rocks led to the discovery of a rock association that possesses differing geochemical signatures within the studied area. The Ulan-Sar’dag mélange includes blocks of mantle peridotite, podiform chromitite, cumulate rocks, deep-water siliceous chert, and metavolcanic rocks of the Ilchir suite. The ophiolitic unit shows overturned pseudostratigraphy. The nappe of mantle tectonites is thrusted over the volcanic-sedimentary sequence of the Ilchir suite. The metavolcanic series consist of basic, intermediate, and alkaline rocks. The mantle peridotite and cumulate rocks formed in a supra-subduction zone environment. The mafic and metavolcanic rocks belong to the following geochemical types: (1) Ensimatic island-arc boninites; (2) island-arc calc-alkaline andesitic basalts, andesites, and dacites; (3) tholeiitic basalts of mid-ocean ridges; and (4) oceanic island basalts. U–Pb dating of zircons from the trachyandesite, belonging to the second geochemical type, yielded a date of 833 ± 4 Ma which is interpreted as the crystallization age during mature island-arc and intra-arc rifting stages. The possible influence of later plume magmatic-hydrothermal activities led to the appearance of moderately alkaline igneous rocks (monzogabbro, trachybasalt, trachyandesite, subalkaline gabbro, and metasomatized peridotites) with a significant subduction geochemical fingerprint.

Keywords: ophiolite; volcanic rocks; geochemistry; subduction; plume magmatic-hydrothermal activity

1. Introduction

The studied area is located in the southeastern region of Eastern Sayan. This region was tectonically juxtaposed during the Neoproterozoic accretionary orogeny of the Central Asian Orogenic Belt (CAOB), which consists of a complex amalgamation of geologic terranes comprised of rocks of Archaean to Mesozoic ages [1–7]. The earliest stages in the CAOB formation are associated with the development of the Paleo-Asian Ocean, which was opened during the Late Riphean split of the Laurasia supercontinent into the Siberian and Laurentian cratons and a number of microcontinents [2,5,8–13]. Continental rifting, ocean subduction, and marginal basin formation events began prior to 1000 Ma and continued until 570 Ma [7]. Neoproterozoic dikes, sills, and small stocks of gabbro-dolerites with rift-related
geochemical affinities have ages of 950–1000 Ma and are abundant throughout the southwestern margin of the Siberian craton [2]. During the early Paleozoic to late Mesozoic periods, the island arcs, ophiolites, oceanic islands, seamounts, accretionary wedges, oceanic plateau, and microcontinents were joined to the Siberian craton (Figure 1a) [2,7,14–16].

The southeastern domain of the CAOB includes Archean-Proterozoic terranes—Gargan block (microcontinent), parautochthon of Proterozoic carbonate cover (Irkut and Ilchir suites), and ophiolitic belts associated with volcanic complexes and Paleozoic terranes: Khamardaban and Khamsara belts (Figure 1b).

The basement of the Gargan block (microcontinent) is composed of Archean to Paleoproterozoic high-grade metamorphic rocks—amphibolites, gneisses, and crystalline schists [2,10,17]. The gneisses have an Rb–Sr whole-rock isochron age of 3153 ± 57 Ma and a U–Pb zircon with an age near 2000 Ma [18]. The basement is unconformably overlain by the Proterozoic carbonate cover of the Irkut suite [19,20] and the rocks of the Ilchir suite, which are composed of dark grey shales, sandstones, and limestones. Olistostromes, including ophiolite and Irkut dolomite exotic clasts exist among the rocks of the Ilchir suite. This sedimentary succession is intruded by tonalities of the Sumsunur complex with an age of 790 Ma [21]. According to their lithological and geochemical characteristics, the carbonate rocks of the Irkut suite belong to sub-platform shelf zone facies [19,20]. Irkut and Ilchir suites are interpreted as passive margin shelf deposits. Thus, the Gargan block represents the detached continental block within the Paleo-Asian ocean in the Riphean (1600–650 Ma).

The ophiolitic belts exposed in the Eastern Sayan are Southern Siberia’s largest and best-preserved relics of the ancient oceanic crust of the Paleo-Asian Ocean. Several tectonically dismembered ophiolite complexes are exposed along the margin of the Gargan block and tectonically thrust over this block [1,4–6,8,22–25]. Ophiolites are associated with the sedimentary and volcanic units of the Dunzhugur island arc of 850–800 Ma [4]. The Ulan-Sar’dag ophiolitic mélangé is located in the inner part of the Gargan block. The Osipa-Kitoi ophiolite is situated at the eastern edge of the Gargan block. The northern part of the Osipa-Kitoi ophiolite is composed of mantle peridotites, containing podiform chromitite [6,26,27], fragments of the sheeted diabase dike complex, and a basaltic lava sequence [8,28]. The ophiolitic serpentinite mélangé is unconformably overlain by dolomites of the Gorlyk suite, with a late Neoproterozoic–early Cambrian age with the contact between them marked by ophiolcates. This is the only stratigraphic evidence for a late Neoproterozoic age of the ophiolitic assemblage. The southern part of the Osipa-Kitoi ophiolite unit is comprised by the mantle peridotite with podiform chromitites, cumulate sequence with the layered and isotropic gabbros, and pegmatitic amphibole gabbros [4,6,8,26–28]. In the Dunzhugur area, the ophiolite association and island-arc rock complexes comprise of the Dunzhugur massif, which is situated at the western margin of the Gargan block. They consist of mantle peridotite, layered cumulate sequences, gabbroic sections, sheeted diabase dike complexes, and basaltic pillow lavas, with the flysch-type sedimentary sequences overlying the upper part by the ophiolite massif [8,21,24,25,29–31]. Two generations of basic dikes were identified in the massif. The first generation geochemically corresponds to high-Mg, low-Ti basalts and andesitic basalts of the tholeiitic series. The second generation is represented by geochemically matching the high-Mg member of the boninite series [9,24,30,31]. The volcanic complex is represented by extremely low-Ti and low-Fe basaltic andesites and andesites with elevated MgO and Cr₂O₃. Typical boninites and manarites were discovered among these volcanic rocks. The trace rare element (TRE) patterns reveal negative Ta, Nb, and Ti anomalies [24,30,31]. The island-arc geochemical affinities correspond to the suprasubduction setting [24,30,31]. Diabase sills and dikes intrude the ophiolite series [24,30–32] covered by sedimentary rocks. They correspond to high-Ti/Fe/P subalkaline tholeiites. The presence of these igneous rocks indicates the extension of the island arc that can be attributed to intra-arc rifting. The plagiogranite and volcano-clastic rocks in the Dunzhugur island-arc complex yielded an age of 1020–850 Ma [30,31] using U/Pb and Pb/Pb zircon dating. The evolution of the Dunzhugur island-arc complex continued during the period of 850–760 Ma and was followed by the formation of the Sarkhoi
island arc. A collision of the Dunzhugur island arc and the continental margin of the Gargan block occurred at about 810 Ma [30].

![Figure 1](image-url)

**Figure 1.** (a) Fossil island oceanic arc systems of the Central Asian Orogenic Belt (CAOB). Siberian Group: Neoproterozoic to early Cambrian island oceanic arcs of Siberia and Mongolia (pink). Kazakhstan Group: Early Paleozoic arcs of Kazakhstan (blue). Southern Group: Middle Paleozoic arcs of Tienshan, Chinese Altai, and Mongolia (yellow). Lines marking fossil intra-oceanic arcs made according to [5,10,33] (and references therein). The scheme was taken from [33]. Eastern Tuva-Sayan, Russia—NW Mongolia: 1—Dunzhugur arc; 2—Shishkhid arc; 3—Ilichir arc; Western Tuva-Sayan, Russia: 4—Agardag arc; 5—Shatskii arc; 6—Tannu-Ola arc; 7—Kurtushibin arc; Western Mongolia: 8—Dariv arc; 9—Khan-Taishirin arc; Mongolia-Transbaikalia: 10—Dzhida arc; Russian Altai: 11—Kurai-Ulagan arc; Northern Kazakhstan: 12—Selety-Urumbai arc; East Kazakhstan: 13—Bozshakol-Chingiz arc; 14—Baydaulet-Aqbastau arc; Southern Tienshan, Kyrgyzstan: 15—Chatkal-Atbashii arc; Southern Tienshan, Tajikistan: 16—Fan-Karategin arc; SW Mongolia: 17—Gurvansayhan-Zoolen arc; Chinese Altai: 18—19—Saerbulake arc. (b) Major geological units of Eastern Tuva-Sayan (modified from [34]) with new information from the authors. Captions of the ophiolite massifs: 1—Dunzhugur; 2—Holbin-Hairhan; 3a—northern branch of Osipa-Kitoi, 3b—southern branch of Osipa-Kitoi; 4—Ulan-Sar’dag; 5—Ehe-Shigna Shishkhid branch.

The Sarkhoi volcanic belt is composed of the volcanic-sedimentary arc-like rock association named the Sarkhoi island arc. This association comprises of volcanic coarse-grained sandstones, conglomerates, and greenstones with clasts of basalts, andesites, dacites, and ignimbrites. This complex was formed by suprasubduction magmatism, which includes the differentiation and assimilation of crustal material at the active continental margin setting. The age of the volcanic rocks is 760–700 Ma [23,35]. At ~750 Ma, the back-arc rifting was developed at the Sarkhoi island arc. It is likely that at approximately the same time (750 Ma), the Shishkhid island arc collided with the margin of the Sarkhoi island arc [30].
The Sumsunur tonalite-trondhjemite complex comprises of synollisional intrusions. It consists of tonalities, diorites, trondhjemites, and granodiorites. It appears that the complex was coeval to the Sarkhoi island arc. Sumsunur tonalites were dated by the zircon U–Pb method at 700–785 Ma and in situ Rb–Sr dating of the K-bearing minerals by the isochron method, which yielded 812 Ma [4,5,30]. The Oka belt is an ancient accretionary prism at the late Neoproterozoic accretionary prism consisting of shales, sandstones, greenschists, and blueschists. The age and tectonic setting of the Oka belt were controversial for a long time. Kuzmichev [36] suggested that the belt was formed in the late Neoproterozoic time as an accretionary prism. Blueschist-bearing tectonic sheets reveal a mineral isochrone Rb–Sr age of 620–640 Ma [4,5,36]. The imbricated structure, occurrence of the oceanic rocks and blueschists, and some other features of the Oka belt are typical for modern accretionary prisms. The Sarkhoi island arc, Sumsunur tonalities, and Oka accretionary prism are interpreted as a continental arc that resulted from the collision of the Dunzhuger island arc with the Gargan block.

The Shishkhid ophiolitic and island-arc complexes, named Shishkhid ophiolite [34], are represented by a suprasubduction ophiolite and volcanic island-arc association. Ophiolitic association corresponds to the uppermost oceanic lithosphere, comprising of mantle peridotites and cumulate ultramafic-mafic sequence, gabbro, and a sheet-dike complex. Volcanic association consists of basic to felsic and pyroclastic rocks. Shishkhid ophiolite formed as a result of island-arc rifting. The ophiolite was thrust eastward onto the Oka belt and is underlain by the Oka mélangé zone comprising of serpentinite lenses intercalated with the shales of the Oka formation. In the west, the ophiolite is covered by a sedimentary sequence, showing progressive subsidence of the volcanic edifice after volcanism cessation. Ediacaran–Cambrian platform sediments unconformably overlie this sequence. Therefore, the Shishkhid ophiolite was thrust upon the Oka prism before the end of the Neoproterozoic time [32,36,37]. After the oceanic lithosphere of the basin, dividing the Sarkhoi and Shishkhid island arcs had been completely subducted, and the Shishkhid island arc was accreted to the Sarkhoi island arc. The Gargan block evolved as a carbonate platform fringed by the passive margin. [30].

The ophiolitic units form three long belts (Figure 1): (1)—The southern (Ilchir) branch was formed at the oceanic spreading stage, >1200–1100 Ma [27,37,38]; (2)—the northern Dunzhuger branch was formed at the island-arc stage at 1035–850 Ma [24,25]; and (3)—the Ehe-Shigna-Shishkhid branch was formed in a back-arc setting at 850–800 Ma [25,34]. In general, geochemical signatures for the volcanic rocks associated with Eastern Sayan ophiolites reveal wide compositional series mid-ocean ridge basalt (MORB)-type basalts (tholeiites), boninites, island-arc andesites and dacites, and ocean island basalt (OIB)-type basic rocks. In the ophiolite units, the alkaline rocks, such as amphibole-plagioclase, plagioclase-phlogopite, and lamprophyre-like (mica peridotite) rocks, were exposed within the mantle peridotite rock series [39]. In addition, basalts and mafic dikes corresponding to oceanic plateau basalts are localized within the basement of the Gargan block and within the ophiolitic branches [40]. The origin of the OIB-type mafic rocks in the accretionary complexes has been widely discussed [41–43]. It is known that OIB-type melts cannot be generated directly by the partial melting of the arc type mantle wedge overlying subduction zones. The intraplate basalts require a specific mantle source that is significantly enriched in large-ion lithophile elements (LILE), light rare earth elements (LREE), and high-field-strength elements (HFSE) relative to the MORB-type athespheric mantle, but not depleted in HFSE relative to the arc-type mantle wedge [41,42]. Many previous studies have presented the geochemical data on mafic volcanic rocks, such as those from the spreading or island-arc type, among which only 5 to 10% comprise OIB-type basalts. The OIB-type basalts may carry critical information related to either continuous or episodic mantle plume magmatism [43].

The Ulan-Sar’dag melange (USM) occurs as an ophiolitic mélangé, composed of a variety of structurally mixed different lithostratigraphic units. These units are represented by ultramafic-mafic rocks with podiform chromitites, foliated serpentinites, cherts, volcanic-sedimentary rocks of the Ilchir suite with MORB, and island-arc and OIB geochemical affinities [44,45]. The units are tectonically dispersed in a matrix of metamorphic rocks (amphibolites, gneisses, and crystalline schists) of the Gargan block. The USM is located in the inner part of the Gargan block between the ophiolites of
the southern and northern ophiolitic branches (Figure 1). The Ulan-Sar’dag ophiolitic mélange may constitute an independent branch [38]. At the present time, this ophiolitic mélange is still poorly studied and only briefly described in Russian and international publications [44–46]. The focus of this paper is to describe the petrography, mineralogy, geochemistry, and geochronology of the mantle and cumulative peridotites, gabbros, and volcanic-sedimentary sequence of the USM. The purpose of the paper is to estimate the geochemical characteristics of the magmatic sources and the geodynamic setting of the different segments of the Ulan-Sar’dag ophiolitic mélange as well as their relations to the units of the Paleo-Asian Ocean and Central Asian orogenic belt.

2. Geological Setting

The Ulan-Sar’dag ophiolitic mélange is a lenticular body elongated in an east–west direction with dimensions of 2 km × 5 km. The Ulan-Sar’dag ophiolitic melange (USM) is composed of a variety of structurally mixed different units: Ophiolitic ultramafic-mafic rocks with podiform chromitites, foliated serpentinites, deep-sea sedimentary cherts, shaled volcanic-sedimentary island-arc rocks (Ilchir suite), and limestones (Irkut suite) (Figures 2–4) [44,45]. In some places in the studied area, the exotic blocks of the sheeted volcanic-sedimentary rocks are emplaced into the limestones.

Figure 2. Photographs of field relationship between mantle peridotites and volcanic-sedimentary sequence (Ilchir suite), limestone (Irkut suite), Gargan gneisses, and Sumsunur tonalites: (a) View of the south side of the Ulan-Sar’dag melange (USM) and (b) view at the north side of the Ulan-Sar’dag melange.
Figure 3. (a) Geological scheme of the Ulan-Sar’dag mélangé with two sections (b) AB and CD.

Figure 4. Photograph of the field relationship between the gabbro-diabase dike and serpentinized mantle peridotites (east flank of the USM).

On the eastern flank of the Ulan-Sar’dag mélangé, the gabbro-diabase dikes crosscut the serpentinized dunites (Figure 4). Dikes cross the Ulan-Sar’dag mélangé and have a northeastern extension, with a length of 650–700 m and a thickness of up to 10 m (azimuth 320–330°, angle 75°). At the border with the dunite, chilled margins exist with a thickness of up to 1.5 m. The block of mantle peridotites are thrusted over the volcanic-sedimentary sequence. The rocks at the base of the
3. Materials and Methods

A total of 70 samples of ophiolitic and volcanic-sedimentary rocks from the USM were studied. The analyses of the whole-rock major, trace, and rare-earth element compositions were carried out at the Analytical Center for Multi-Elemental and Isotope research (VS Sobolev Institute of Geology and Mineralogy, Novosibirsk, Russia). Mineral chemistry was determined by wavelength-dispersive analysis using electron probe microanalyses (JEOL JXA-8100) at the Sobolev Institute of Geology and Mineralogy, Russian Academy of Science, Novosibirsk, Russia (Analytical Center for Multi-Elemental and Isotope Research, SB RAS). The accelerating voltage was 20 kV, the probe current was 50 nA, the beam size was 3–5 µm, and the signal accumulation time was 10 s. The standards used were natural and synthetic silicates and oxides. The detection limit for oxides was 0.01–0.05 wt.%. The major element composition of whole-rock samples was determined using VRA-20R X-ray fluorescence. The analytical errors were generally less than 5%. Trace elements (including rare-earth elements) were analyzed in solutions by inductively coupled plasma mass spectrometry (ICP-MS) using a Finnigan Element mass spectrometer [47]. The detection limits for trace elements were in the range 0.01–0.2 µg/L. Zircon U–Pb isotopic analyses were performed using a Sensitive High-Resolution Ion Microprobe (SHRIMP-II) at the Russia Geological Research Institute (VSEGEI, St. Petersburg, Russia). The zircons were photographed in reflected and transmitted light using an Axio Scope A1 Zeiss microscope (Axio Scope A 1, Carl Zeiss Microlmaging GmbH, Göettingen, Germany). The internal structure of the zircons was studied by SEM-cathodoluminescence (CL) images at the Center of Isotopic Research, VSEGEI, St. Petersburg. Instrumental conditions and measurement procedures were described previously [48]. Spots of approximately 20 microns were analyzed. Data for each spot were collected in sets of 5 scans. At VSEGEI, TEMORA zircon from leucogabbro Middledale (Lachlan fold belt, Eastern Australia) was used as an age standard. Age calculations and concordia plots were drawn using Isoplot software (ver. 3.0, Berkeley Geochronology Center, Berkeley, CA) [49].

4. Results

4.1. Petrography

Most ophiolitic peridotites, gabbros, and volcanic-sedimentary rocks of the Ilchir suite had undergone hydrothermal alteration (serpentinites, talcites, and actinolite-tremolite rocks) and greenschist to amphibolite facies metamorphism (Figures 5–7).
The mantle peridotites are mainly serpentinized dunites, with minor harzburgites and podiform chromitites [45]. The primary minerals in the serpentinized dunites and harzburgites are olivine (80–94 vol. %), rarely subidiomorphic orthopyroxene (3–10 vol. %), accessory subidiomorphic Cr-spinel (3–6 vol. %) (Figure 5), and secondary minerals—serpentine, chlorite, talc, carbonate, and magnetite. Dunites and harzburgites have a porphyroblastic and granoblastic texture with subidiomorphic porphyroblasts of orthopyroxene in the olivine matrix.

The cumulate rocks include pyroxenite and gabbro. These consist mainly of amphibolized pyroxenites, isotropic amphibolized gabbros, rare olivine-bearing pyroxenites, and bimineral hornblende-anorthosite rocks. Pyroxenite has a medium-grained and equigranular texture (Figure 6a). The primary minerals are olivine, Cr-spinel, orthopyroxene, clinopyroxene, and minor feldspar; accessory—ilmenite and apatite; and secondary—amphibole, biotite, serpentine, talc, chlorite, epidote, titanite, and newly-formed zircon. The ore minerals are awaruite, Co-bearing pentlandite, chalcopyrite, and pyrite. Olivine-bearing pyroxenites are medium-grained rocks with porphyritic and cataclastic textures. Phenocrysts are partially fragmented primary olivine, Cr-spinel with orthopyroxene inclusions, and orthopyroxene. These grains are dissected by the cracks filled with secondary serpentine and talc. In the serpentinized matrix, small magnetite grains and awaruite are often found (Figure 6b). Gabbro are medium-grained with equigranular or porphyritic textures (clinopyroxene phenocrysts). Gabbro is composed of primary clinopyroxene—diopside or augite, plagioclase. Accessory phases are ilmenite and apatite; secondary—biotite, titanite, albite, epidote, scapolite, muscovite, chlorite, and hematite (Figure 6c,d). Hornblende-anorthosite is an exotic bimineral rock consisting of hornblende and anorthite.
Figure 6. Back scattered electron (BSE) microphotographs of the cumulate rocks and mafic rocks associated with ophiolite: (a) Amphibolized pyroxenite with relics clinopyroxene and plagioclase replaced by epidote; (b) olivine-bearing amphibolized pyroxenite–olivine phenocrysts with orthopyroxene inclusion in an olivine-amphibole-talc-serpentine matrix; (c) medium-grained amphibolized gabbro with equigranular texture; (d) amphibolized gabbro with porphyritic texture; (e) monzogabbro with veinlets of pyrite-chalcopyrite; and (f) amphibolized gabbro-diabase dike with accessory ilmenite, secondary titanite, and new-formed zircon. Abbreviations: Ol—olivine, Opx—orthopyroxene, Cpx—clinopyroxene, Amp—amphibole, Pl—plagioclase, Kfsp—K-feldspar, Mag—magnetite, Srp—serpentine, Ap—apatite, Ep—epidote, Ilm—ilmenite, Ttn—titanite, Zrn—zircon, Awr—awaruite, Py—pyrite, and Ccp—chalcopyrite.
Among the cumulate rocks, monzogabbro occurs rarely. It has a medium-grained structure with hypidiomorphic texture. The essential mineral assemblage is represented by plagioclase—40 vol. %, orthoclase feldspar—10 vol. %, biotite—15 vol. %, amphibole—5 vol. %; accessory—apatite, titanite, ilmenite, zircon, pyrite, pyrrhotite, and chalcopyrite; and secondary—epidote and chlorite (Figure 6e). The gabbro-diabase dike does not belong to the ophiolitic dike complex. It has a phaneritic texture. Primary minerals are clinopyroxene–diopside and plagioclase andesine–labradorite. Accessory minerals are biotite, apatite, ilmenite, and zircon. Pyroxenes were replaced by amphiboles of varying composition and plagioclase was saussuritized and albitized. The secondary phases are epidote, scapolite, chlorite, sericite, and titanite. Sulfide mineralization is pyrite, cubanite, chalcopyrite, molybdenite, and sphalerite (Figure 6f).

Metamorphosed volcanic (metavolcanic) rocks have a chemical composition corresponding to basalts, trachybasalts, andesitic basalts, andesites, trachyandesites, and dacites. Basalt has a hypidiomorphic texture of groundmass consisting of amphibolized pyroxene and laths of idiomorphic plagioclase. Accessory minerals are apatite. Secondary minerals are amphibole, biotite, titanite, new-formed zircon, chlorite, albite, sericite, talc, and calcite (Figure 7a,b).

Andesitic basalt has an ophitic-subophitic texture with relics of clinopyroxene diopside–augite and plagioclase. Clinopyroxene is replaced by actinolite and chlorite. Plagioclase is replaced by epidote and albite (Figure 7c,d). Andesites and dacites have a microporphyritic texture with phenocrysts of andesine placed in the microcrystalline groundmass consisting of altered plagioclase and actinolite (Figure 7e,f). Trachyandesite has porphyritic textures with fine-grained groundmass and preferentially oriented lathes of feldspar. Primary minerals are biotite and small quantities of orthoclase and quartz. Accessory minerals are ilmenite, apatite, and zircon. Secondary minerals are albite, epidote, titanite, and sericite. Groundmass consists of altered plagioclase and metamorphogenic porphyroblasts of the Ca–Fe–Mn
garnet grossular–almandine–spessartine group. Groundmass comprises of secondary phases—biotite,
albite, epidote, and sericite. Trachyandesite hosts the sulphidized zone. Sulfide mineralization is
represented by pyrrhotite and pyrite. Dacites have porphyritic textures with fine-grained groundmass.
Phenocrysts are primary oligoclase and potassium feldspar. Groundmass consist of primary—biotite,
feldspars, and quartz; accessory phases include apatite, zircon, and monazite; and secondary phases
are epidote, titanite, chlorite, and sericite.

Trachybasalt has not preserved primary minerals. It has a fine-grained structure and consists
of secondary minerals—amphibole, albite, epidote, biotite, titanite, sericite, and chlorite; and
accessory—ilmenite and apatite.

4.2. Mineral Chemistry

The chemical composition of the minerals is given in Tables S1–S5.

Olivine in the dunites and the harzburgites is represented by high-Magnesium forsterite (Fo 90–91).
It is unzoned and homogeneous. Olivine in the olivine-bearing pyroxenite is more ferrous, Mg# of 85–86,
and in dunite and harzburgite the Mg# of 91–93 (Table S1) (Mg# = 100xMg²⁺/(Mg²⁺+Fe³⁺)).

Cr-spinel is chromite. It occurs as an accessory mineral in the dunites, harzburgites, and olivine-
bearing pyroxenite. Cr-spinel has the same Mg# of 54–56 as that for dunite and olivine-bearing
peridotite, but differs in Al# (17–18), (32–34) (Al# = 100xA³⁺/(A³⁺+Cr³⁺+Fe³⁺)), and Cr# (82–83), (66–68) (Cr# = 100xCr³⁺/(A³⁺+Cr³⁺+Fe³⁺)), respectively (Table S1).

Pyroxene in the olivine-bearing pyroxenites is enstatite (Mg# of 92–94) (Figure 8). Relics of the
pyroxene from the gabbro are augite with Mg# 80–85. Relics of the pyroxenes in the gabbro-diabase
dike are diopside in composition with Mg# of 52–60 (Table S1).

![Figure 8. A pyroxene nomenclature diagram [50].](image)

Feldspar: Plagioclase composition is illustrated in a ternary plot of the An–Ab–Or system (Figure 9)
and is listed in Table S3. Plagioclase (An. 10–70%) partially or almost completely was saussuritized
by an aggregate of albite, epidote, and sericite. Plagioclase in the peridotite and monzogabbro has
a composition of albite-oligoclase (secondary) and albite-andesine-orthoclase, respectively. In the
hornblende-anorthosite, plagioclase is pure anorthite. In the gabbro, the composition of plagioclase
varies from bytownite to oligoclase. Plagioclase in the basalt varies from bytownite–labradorite to
albite. Plagioclase in the andesitic basalt varies from andesine to albite. In the andesite, plagioclase
varies from andesine to albite. Plagioclase of the trachyandesite is oligoclase-andesine and a small
amount of K-feldspar.
Biotite is a primary mineral in the monzogabbro, andesites, and dacites. In the gabbros and basic volcanic rocks, biotite is a secondary mineral. The chemical composition of biotite is listed in Table S4. There is wide data scattering for the biotite of basic rocks (Figure 10). Andesitic basalt has Mg\# = 0.45 and a low content of TiO$_2$ of 0.02–0.8 wt.%. The biotite of the andesites has a similar Mg\# as in the andesitic basalt, but a higher TiO$_2$ content. The most ferrous biotite is in dacite. In the monzogabbro, biotite is ferrous, Mg\# = 0.3, Fe$_{\text{total}}$ (in the monzogabbro) is 25–26 wt.%, and TiO$_2$ is high at 1.5–3.5 wt.%

Amphibole is a secondary mineral in the ophiolitic and metavolcanic rocks. The chemical composition is listed in Table S2. The amphiboles belong to the calcic and sub-alkaline (Figure 11a,b) amphibole group [50]. Amphiboles in peridotites vary from pargasite–hornblende–tschermakite to actinolite. Amphiboles in the basalts vary in the composition of pargasite–hornblende–tschermakite. In the andesitic basalts, amphiboles are actinolite. Amphiboles in the monzogabbro show a composition from hornblende to actinolite. Amphiboles in the gabro-diabase dike are pargasite, edenite, hornblende, and actinolite. The subalkaline amphiboles such as edenite and pargasites belong to the ferrous subgroup.
Figure 11. A nomenclature diagram [52] for amphiboles.

Ti-bearing phases include ilmenite and titanite (Table S5). These phases are found in almost all ophiolitic and metamorphosed volcanic rocks with elevated TiO$_2$ content. Most often, ilmenite is an aggregate of scattered microparticles in a titanite (replaced by titanite), or it is a number of microns (not more than 20 µm) of grains in the titanite. Ilmenite also occurs as inclusions in biotite. A distinctive feature of ilmenite is a high content of MnO, from 4 to 10 wt.%. Ilmenite has a negative correlation with FeO wt.% and MnO wt. % (Figure 12a,b). It is noteworthy that ilmenite with the same composition from the rutile-ilmenite-titanite-apatite accessory assemblage of alkaline lens-shaped rock bodies (plagioclase-amphibole, plagioclase-phlogopite, and lamprophyre-like rocks) intruded the Ospa-Kitoi mantle peridotites [34]. Titanite in all studied rocks forms rims around the ilmenite or xenomorphic grains in plagioclase-amphibole-epidote groundmass. In both cases, titanite associates with biotite.

Zircon: Two types of zircons were found in the ophiolitic and metavolcanic rocks. The first type is primary magmatic zircon in the trachyandesite, andesite, dacites, and gabbro-diabase dike. The second is newly formed micro-aggregate (3–10 µm) in the metavolcanic rocks and the gabbro-diabase dike. Zircons suitable for U–Pb dating are found in the trachyandesite. A detailed description is given in Section 4.4.
4.3. Whole-Rock Major and Trace Elements Composition

4.3.1. Major Elements

The chemical composition of the peridotites, cumulate, and metavolcanic rocks is given in Table S6. Intensively altered magmatic rocks that did not follow a typical alkalinity magmatic trend [53] were excluded from analytical results. If the values of $K_2O/(K_2O + Na_2O) \times 100/(K_2O + Na_2O)$ lie outside a typical alkalinity magmatic compositional trend, then the rocks have undergone intense hydrothermal alteration and metamorphism [53].

The geochemical character in terms of the distinction between alkaline, tholeiite, and calc-alkaline magmatic series has traditionally been based on the total alkali-silica (TAS) diagram (Figure 13a). According to the $SiO_2/Na_2O + K_2O$ ratio, the ophiolitic and associated metavolcanic rocks have a range of compositions from ultrabasic (dunites, harzburgites, pyroxenites), basic (gabbros, basalts), high-Mg andesitic basalts, and andesitic basalts to intermediate (andesites) and felsic (dacites) rocks of a normal alkaline rock series and basic alkaline rock series. High field strength elements (such as Ti, Y, Zr, Nb, and Ta) are thought to remain relatively immobile during a wide range of metamorphic conditions [54–56].
Figure 13. Geochemical diagrams for mafic, intermediate, and felsic of the USM rocks: (a) Total alkali-silica (TAS) diagram for magmatic rocks [57]; (b) Nb/Y vs. Zr/Ti diagram to distinguish between subalkaline and alkaline basalts for altered rocks [58]; and (c) Zr/Y vs. Th/Yb diagram to distinguish between the tholeiite and calc-alkaline series for altered rocks [59].

We used HFSE (Ti, Zr, Y, Nb, Th, and Yb) to characterize the magmatic rocks (Figure 13b,c). The rocks correspond to the sub-alkaline basalt, andesitic basalt, andesite, dacite, and rhyolite consistent
with the Nb/Y–Zr/Ti discrimination diagram. A few of the samples display trachyandesite, trachyte, and alkali basalt composition. The samples of the gabbro-diabase dike belong to the alkaline basalt field.

4.3.2. Trace Elements

Four groups of magmatic rocks were identified according to major and trace element compositions (Figure 14a–h). Each group is characterized by its own concentration of LILE, REE, and HFSE elements, and trace element ratios.

Group I: High-Mg andesitic-basalt (named further as boninites) shows the distribution of REE (rare earth elements) and the trace elements on the spidergram (Figure 14a,b) which can be referred to as boninites [60]. It is characterized by low concentrations of REE and negative anomalies of HFSE (Nb, Ta, and Ti). It has a pronounced positive anomaly in Sr, low \((\text{La}/\text{Yb})_n = 1.41\) (normal-type mid-ocean ridge basalt (N-MORB)–0.82), \((\text{Ce}/\text{Y})_n = 0.33\), \((\text{Th}/\text{Yb})_n = 1.1\), \((\text{Ta}/\text{Nb})_n = 0.07\), and \((\text{Zr}/\text{Nb})_n = 29\) (Table S6). According to the elemental ratios, the andesitic-basalts are similar to those of N-MOR basalts and boninites.

Group II: Gabbros, andesitic basalts, andesites, dacites, and rhyolites have a slightly pronounced negative slope in the REE pattern (Figure 14c,d) and the distribution of REE, HFSE, and LILE which fully correspond to the distributions of the island-arc rocks. All rocks from this group have a negative anomaly for Ta, Nb, and Ti. In a number of samples, positive anomalies in Sr, Ba, Rb, and Zr are presented. The ratios \((\text{La}/\text{Nb})_n (1–4), \text{Ce}/\text{Y} (2–3), \text{Th}/\text{Yb})_n (1–5),\) and \((\text{Ta}/\text{Yb})_n (0.2–1)\) in these rocks correspond to values in the upper and lower continental crust [61]. The elevated abundance of Nb (41 ppm), Ti (4000–8000 ppm), Y (23–64 ppm), and Zr (450–700 ppm) in some andesites and dacites are not typical for intermediate island-arc rocks.

Group III: E-MOR basalts. The basalts and one sample of gabbro show the distribution of REE and HFSE corresponding to enriched-type MORB (E-MORB), with the exception of LILE elements (Cs, Rb, Ba, and U) (Figure 14e,f). The presence of a subduction component is noted and indicated by the elevated content of Rb compared to the MORB. According to trace element ratios, basalts show hybrid characteristics of N-MOR, E-MOR basalts, and continental crust components (Table S6). These are slightly enriched in LREE relative to E-MORB, and have a weak positive anomaly in Sr.

Group IV: Basic and intermediate rocks of the sub-alkaline and alkaline series. These have concentrations and patterns of REE and HFSE identical to those of the OIB (Figure 14g,h). The rocks show a negative anomaly in Sr and high \((\text{La}/\text{Yb})_n = 13–43\) (OIB–17.43) and \((\text{Nb}/\text{Y})_n = 1.38–1.65\).

To identify the mantle-plume component in the ophiolitic and volcanic rocks, we used the \((\text{Zr}/\text{Nb})_n\) ratio proposed by J.G. Fitton [62] for the subdivision of Icelandic basalts Equation (1):

\[
\Delta\text{Nb} = \log (\text{Nb}/\text{Y}) + 1.74 - 1.92 \times \log (\text{Zr}/\text{Y})
\]

In nature, there is a mixture of various mantle sources (enriched and depleted reservoirs). To separate the plume component, an analysis of the relationship of variations in trace elements and their ratio was proposed, which in many cases makes it possible to estimate the plume component in magmatic products of mixed origin [62–64]. The \((\text{Zr}/\text{Nb})_n\) ratio is the most informative, which in all OIB types of basalts is the most distinctive from the values typical for depleted mantle. A deficiency or excess of Nb, relative to the lower limit of the Iceland array, may be expressed as \(\Delta\text{Nb} = 1.74 + \log (\text{Nb}/\text{Y}) - 1.92 \log (\text{Zr}/\text{Y})\). Icelandic OIB plume basalt has \(\Delta\text{Nb} > 0\) and N-MORB has \(\Delta\text{Nb} < 0\), and these are the fundamental source characteristics that are insensitive to the effects of variable degrees of mantle melting. However, this may characterize the source depletion through melt extraction, crustal contamination of the magmas, or subsequent alteration [62]. Positive \(\Delta\text{Nb}\) values indicate the contribution of the plume component in the mantle source. Contamination by components released from the subducted slab can significantly decrease \(\Delta\text{Nb}\) in the rocks. Ophiolitic rocks of the first and second groups have negative \(\Delta\text{Nb}\) values. The rocks of the third and fourth groups have positive \(\Delta\text{Nb}\)
values from (+0.08) to (+0.5), which indicate the contribution of the plume component to the magmatic source (Table S6).

Figure 14. C1 chondrite normalized REE (rare earth elements) patterns and primitive mantle normalized trace element spidergram; normalizing values are from [65]: (a,b) Group I—boninites (293-16), literature data [60]: Low calcium (LCB), medium calcium (ICB), high calcium (HCB) boninites; (c,d) group II—rside-arc rocks: 1—gabbro, 2—gabbro-diorite, 3—basalt, 4—andesitic basalt, 5—andesite, 6—trachyandesite; (e,f) group III—E-MOR basalts: 1—gabbro, 2—basalts; (g,h) group IV, basic rocks of sub-alkaline and alkaline series: 1—monzogabbro, 2—trachybasalt, 3—gabbro-diabase dike.

4.4. Zircon U-Pb Dating

Zircons from a trachyandesite were analyzed for U–Pb to establish its crystallization age. Representative CL images of eight zircons are shown in Figure 15a–h. The U–Pb dating results
are listed in Table 1 and presented as concordia diagrams (Figure 16a,b). The inner domain of some zircon grains has a weakly sectorial texture with concentric zonation, which is bordered by a contrasting striped zonation of different rhythmicity. Some grains show igneous oscillatory zoned rims and others possibly thin metamorphic rims. They also preserve an igneous texture. All analyses yield close $^{206}\text{Pb}/^{238}\text{U}$ apparent ages of 829–840 Ma, with a weighted mean age of 833 ± 4.0 Ma (MSWD = 0.006, n = 18, Figure 16), which is interpreted as the crystallization age during the mature island-arc and intra-arc rifting stages [23,27,32].

Figure 15. Representative cathodoluminescence (CL) images of zircons (a–h) from trachyandesite (ES-314-16) of the Ulan-Sar‘dag mélangé. The yellow circles show the analyzed areas and the $^{206}\text{Pb}/^{238}\text{U}$ ages (Ma).
Table 1. Sensitive High-Resolution Ion Microprobe (SHRIMP) zircon U–Pb dating results for trachyandesite (ES-314-16).

| Spot  | % 206Pb | U ppm | Th ppm | 232Th/238U | 206Pb/206Pb | (t) 206Pb/207Pb | (t) 207Pb/206Pb | % Dis-cor-dant | 206U/206Pb* | ± | 207Pb/206Pb* | ± | 207Pb/235U | ± | 206Pb/238U | ± | err corr |
|-------|---------|-------|--------|------------|-------------|---------------|----------------|----------------|------------|-----------|-----|-------------|-----|-------------|-----|----------|-----|---------|
| 314_1 | 0.07    | 350   | 156    | 0.46       | 41.3        | 828.9         | ±7.3           | 842            | ±26        | 2         | 7.267 | 0.9        | 0.00715 | 1.2          | 1.271 | 1.6      | 0.1572 | 0.9      |
| 314_2 | 0.16    | 69    | 20     | 0.30       | 9.25        | 838.0         | ±10            | 884            | ±53        | 6         | 7.207 | 1.3        | 0.06550 | 2.6          | 1.310 | 2.9      | 0.1887 | 1.3      |
| 314_3 | 0.19    | 119   | 45     | 0.37       | 14.1        | 827.5         | ±8.6           | 859            | ±53        | 4         | 7.303 | 1.1        | 0.06770 | 2.6          | 1.278 | 2.8      | 0.1869 | 1.1      |
| 314_4 | 0.07    | 147   | 51     | 0.36       | 17.3        | 831.0         | ±8.3           | 809            | ±39        | -3        | 7.268 | 1.1        | 0.06610 | 1.9          | 1.253 | 2.1      | 0.1576 | 1.1      |
| 314_5 | 0.08    | 167   | 51     | 0.34       | 19.8        | 831.6         | ±6.7           | 859            | ±45        | 3         | 7.244 | 1.1        | 0.06770 | 2.2          | 1.238 | 2.5      | 0.1380 | 1.1      |
| 314_6 | 0.04    | 256   | 94     | 0.38       | 30.5        | 835.7         | ±7.7           | 804            | ±30        | -4        | 7.224 | 1.0        | 0.06593 | 1.4          | 1.258 | 1.7      | 0.1384 | 1.0      |
| 314_7 | 0.27    | 134   | 35     | 0.27       | 15.9        | 832.1         | ±9.5           | 940            | ±57        | 13        | 7.258 | 1.2        | 0.07040 | 2.8          | 1.337 | 3.0      | 0.1578 | 1.2      |
| 314_8 | 0.01    | 1108  | 49     | 0.46       | 131         | 828.7         | ±6.7           | 826            | ±14        | 0         | 7.290 | 0.9        | 0.06661 | 0.7          | 1.260 | 1.1      | 0.1572 | 0.9      |
| 314_9 | 0.35    | 145   | 49     | 0.35       | 17.2        | 831.4         | ±8.5           | 899            | ±54        | 7         | 7.265 | 1.1        | 0.06070 | 2.6          | 1.303 | 2.8      | 0.1577 | 1.1      |
| 314_10| 0.02    | 688   | 712    | 1.07       | 81          | 826.3         | ±7.1           | 831            | ±18        | 0         | 7.295 | 0.9        | 0.06777 | 0.9          | 1.262 | 1.2      | 0.1571 | 0.9      |
| 314_11| 0.66    | 115   | 36     | 0.33       | 13.6        | 829.0         | ±3.1           | 911            | ±86        | 10        | 7.280 | 1.4        | 0.06940 | 4.2          | 1.314 | 4.4      | 0.1573 | 1.4      |
| 314_12| 0.03    | 353   | 224    | 0.69       | 39.3        | 829.2         | ±7.3           | 821            | ±25        | -1        | 7.285 | 0.9        | 0.06845 | 1.2          | 1.258 | 1.5      | 0.1573 | 0.9      |
| 314_13| 0.00    | 134   | 39     | 0.30       | 16          | 841.0         | ±9.3           | 822            | ±40        | -2        | 7.176 | 1.2        | 0.06550 | 1.9          | 1.278 | 2.3      | 0.1393 | 1.2      |
| 314_14| 0.14    | 135   | 76     | 0.58       | 16          | 835.3         | ±6.7           | 822            | ±45        | -2        | 7.226 | 1.1        | 0.06650 | 2.2          | 1.268 | 2.4      | 0.1583 | 1.1      |
| 314_15| 0.04    | 229   | 155    | 0.70       | 27.3        | 837.1         | ±7.6           | 812            | ±31        | -3        | 7.211 | 1.0        | 0.06516 | 1.5          | 1.265 | 1.8      | 0.1887 | 1.0      |
| 314_16| 0.02    | 346   | 182    | 0.54       | 41.3        | 838.8         | ±7.4           | 827            | ±25        | -1        | 7.196 | 1.0        | 0.06667 | 1.2          | 1.277 | 1.5      | 0.1390 | 1.0      |
| 314_17| 0.04    | 223   | 110    | 0.51       | 26.5        | 835.7         | ±7.9           | 836            | ±32        | 0         | 7.225 | 1.0        | 0.06690 | 1.5          | 1.277 | 1.8      | 0.1384 | 1.0      |

*: ppm.
5. Discussion

5.1. Mineral Assemblage Evolution

The magmatic assemblages in the ophiolitic and volcanic rocks are poorly preserved. The calculated pressure and temperature conditions for the relics of orthopyroxene and Cr-bearing clinopyroxene from peridotites yield a range of 1.3–0.8 GPa and temperatures of 880–950 °C at the shallow mantle depths [66].

The presence of amphibole-anorthosite in the mantle peridotites of the Ulan-Sar’dag mélange may indicate metasomatic processes in the upper mantle. Plagioclase with a composition of almost pure anorthite in mantle peridotite from Papua New Guinea and the Yangbwa suprasubduction zone ophiolites (Southwestern Tibetan Plateau) has been previously reported and interpreted by the effect of a metasomatizing agent on the mantle wedge’s mantle peridotites [67,68].

Ophiolitic and volcanic rocks underwent hydrothermal alteration and metamorphism from greenschist to amphibole facies. The metamorphosed igneous rocks of amphibolite facies have an indicative metamorphic mineral assemblage in the studied rocks: Amphibole–edenite, pargasite, tschermakite, and hornblende, high-Ti biotite, and titanite. A wide range in the composition of amphibole is due to a different primary composition of the magmatic rocks. Evaluating the P-T parameters according to the $\text{Al}_2\text{O}_3$, the $\text{Al}^{IV}$ content in amphibole [69,70] shows values for: Edenite in the gabbro-diabase dike—$T \approx 770–860$ °C, $P \approx 0.6–0.7$ GPa; tschermakite in the pyroxenite, gabbro, basalts—$T \approx 840–930$ °C, $P \approx 0.6–0.9$ GPa; and pargasite in the pyroxenite, gabbro, basalt, gabbro-diabase dike—$T \approx 850–930$ °C, $P \approx 0.7–1.0$ GPa (potassian-alumino-ferro-pargasite $T \approx 1000$ °C, $P \approx 1.3$ GPa) (Table S2).

Ilmenite in all rocks is replaced by titanite. A high content of MnO and low content of MgO in ilmenite indicate that it formed as a result of diffusion re-equilibrium with coexisting metamorphic silicates [71–74]. The metamorphosed igneous rocks of epidote-amphibolite facies have an indicative metamorphic mineral assemblage: Amphibole tremolite-actinolite, epidote, biotite, and albite. This association was formed at $T \approx 500–650$ °C and $P \approx 0.2–0.7$ GPa [75].

The metamorphosed igneous rocks of greenschist facies have an indicator metamorphic mineral assemblage: Tremolite-actinolite, chlorite, epidote, albite, and sericite. This association was formed at $T \approx 400–500$ °C [73].

The study results from the studied rocks suggests several stages of metamorphism occurred. The first stage is low-grade sub-sea ocean floor metamorphism. The peridotites are affected by a pervasive, partial serpentinization of olivine. The second stage is retrograde metamorphism which occurred as a result of the obduction of the ophiolites onto the continental crust. Retrograde metamorphism caused peridotites, gabbros, basalts to change lithologically into amphibolites and greenschists.

The characteristics of the retrograde metamorphic stage for peridotites and basic rocks are:

1. Mid-pressure amphibolite facies $P \approx 0.6–1.0$ GPa and $T \approx 770–930$ °C;
(2) Low-pressure greenschist facies $P \approx 0.2-0.7$ GPa and $T \approx 500-650$ °C.

5.2. Tectonic Setting

In recent decades, a number of indicator rare elements and their ratios have been identified and used to determine magmatic sources and geodynamic settings. We assume that HFSE are relatively immobile components in hydrothermal and metamorphic processes. In numerous works, tectonic discrimination diagrams are used based on trace element relationships, including for Proterozoic ophiolitic complexes, and Archean and Proterozoic greenstone belts [55,56]. This work presents a new classification of ophiolite and the implications for the origin of the Precambrian oceanic crust, particularly for some Archean greenstone belts. Based on trace elements, this discrimination was used for the Isua supracrustal belt, Wawa greenstone belt, and Jormua complex. The Isua supracrustal belt plots both to the subduction-related and subduction-unrelated fields. The Wawa greenstone belt and Jormua complex plot entirely within the plume and continental margin types, respectively, of subduction-unrelated ophiolite [56]. The compositions of studied ophiolitic and metavolcanic rocks are plotted on trace element discrimination diagrams for the tectonic interpretation (Figure 17a–e).

The first group is the boninite (high-Mg andesitic basalt) on the discrimination diagram plotted in the fields of island-arc volcanic rock and boninite (Figure 17a–e). Low concentrations of REE and negative anomalies of HFSE (Nb, Ta, and Ti) are characteristic of boninites, and indicate a strong depletion of their mantle source due to one or several episodes of extraction of basalt melts [81–84]. This was generated from a depleted source (N-MORB). Enrichment in large-ion lithophile elements (LILE) Rb, Ba, Cs, Sr, and U in comparison with N-MORB reflects the contribution of the fluid extracted from the subducted slab [81–84]. Boninites are a type of volcanic rock, which clearly indicates their origin in the subduction setting of ensimatic island arcs [33,64,81–84].

The second group contains gabbros, andesitic basalts, andesites, dacites, and rhyolites, which have a wide spectra of island-arc geochemical affinities (Figure 14c,d). On the tectonic discrimination diagram, they plotted in the fields of volcanic arc, island-arc tholeiite, calc-alkaline basalts, and continental arc (Figure 17a–e). They have geochemical characteristics identical to island-arc volcanites. They were generated from island-arc magmas with a significant contribution of crustal components.

The third group contains basalts belong to N-MOR and E-MOR basalts (Figure 14e,f and Figure 17a–e). The presence of a subduction component is noted and indicated by the elevated contents of Rb (7–23 ppm), Ba (56–200 ppm), and Sr (150–370 ppm) compared with E-MORB [65]. The values of trace element ratios ($Zr/Nb)_n$, ($La/Nb)_n$, and ($Th/Nb)_n$ (Table S6) indicate an enriched mantle source [76–80]. The basalts appear to be formed in the mid-ocean ridge setting from an enriched magmatic source. These basalts may be relics of basalt formed at the stage of the opening of the Paleo-Asian Ocean or a newly formed back-arc basin [24,30].

The fourth group contains sub-alkaline and alkaline rocks on the discrimination diagrams plot (Figure 17) in the OIB field. According to the ratio of rare elements ($La/Yb)_n$, ($Zr/Nb)_n$, ($La/Nb)_n$, and ($Th/Nb)_n$ (Table S6), and positive $\Delta$N\text{b values}, the source of the sub-alkaline and alkaline-type correspond to an enriched deep mantle reservoir [63,85,86].
Figure 17. Tectonic discrimination diagrams for the Ulan-Sar’dag mélangé. (a) Nb vs. La (ppm) after [76]; (b) Ta-Hf/3-Th [77]; (c) Zr vs. Nb [78]; (d) (Nb/Yb)n vs. (Th/Yb)n [79]; and (e) Nb-Th, N-MORB normalized [80]. Abbreviation for rock types: N-MORB—normal-type mid-ocean ridge basalt, E-MORB—enriched-type MORB, P-MORB—plume-type MORB, AB alkaline ocean-island basalt, IAT—island-arc tholeiite, OIB—ocean island basalt, CAB—calc alkaline basalt, and BABB—back arc basin basalt.

In the (Nb/Yb)n vs. (Th/Yb)n and Nb, vs. Th, diagrams (Figure 17d,e), which are used for the determination of the tectonic setting for the post-Archaean ophiolites [80], USM rocks of the first and second groups lie within the volcanic arc array. The USM rocks of the third and fourth groups plot in the MORB-OIB array. Part of the andesites (the second group) plot in the P-MORB field.

The Sm-Nd isotope systematics in representative samples of the Dunzhugur ophiolites were completed by Sklyarov [31]. The analyzed diabase dikes and gabbros (according to the first and the second groups) have negative or positive εNd(T) ranging from −1.0 to +1.5 and late Paleoproterozoic model ages TNd(DM) = 1.8–1.6 Ga, which are much older than the 1020 Ma assumed for plagiogranite crystallization. A higher positive εNd(T) of +2.8 in a later dolerite dike with a moderate titanium content (according to the forth group) may record an event of basaltic magmatism during the opening of a back-arc basin in a complementary system oceanic island arc-back-arc basin maintained by plume activity [31]. Sub-alkaline and alkaline igneous rocks (group IV) are also localized within the area adjacent to the Ulan-Sar’dag tectonic mélangé. At the boundary of the oceanic basin and the Gargan...
block evidence for local ocean spreading or mantle plume activity \cite{40,87–89}, is represented by the intrusion of OIB-type mafic riftogenic dikes (Barun–Kholba volcanoplutonic edifice) with \(^{40}\text{Ar}/^{39}\text{Ar}\) ages of 828 ± 7 Ma \cite{40,88}. The occurrence of OIB-type basalts \cite{1,31,44,88,90}, hornblende-anorthosite, albitized phlogopite-anorthosite, and lamprophyre-like \cite{39} rocks among the ophiolite complexes in the south-eastern fragment of CAOB indicate a generation of alkaline melts during magmatic processes.

5.3. The Stages of Magmatism of the Ulan-Sar’dag Mélange

The stages of magmatism according to the evolution of the subduction zone in the studied segments of CAOB are presented in a simplified tectonic model (Figure 18).

![Figure 18. Simplified model of multistage magmatism at Neoproterozoic to Vendian-early Paleozoic time frame of a Pale-Asian Ocean segment.](image)

Stage I. Group I: In the Meso-Neoproterozoic, in connection with the dismemberment of the Rodinia supercontinent along the margin of the Siberian Craton and the emerging Paleo-Asian Ocean, island-arc systems were formed \cite{3,5,6,9–13,90}. At an early stage, an ensimatic island arc was formed on the oceanic-type crust characterized by a generation of boninitic melts. Boninites of the first geochemical group were generated in the intra-oceanic subduction setting (Figure 18a). The shallow melting of the restitic peridotites occurs under conditions of high heat flow and water fluid infiltration from water-rich components released from the subducted slab where the temperature of the boninite melt reaches 1500 °C, which requires special conditions for its genesis and is realized only in the hot mantle wedge of ensimatic island arcs \cite{83}.

Stage II. Group II. During the development to an ensialic island arc, boninite magmatism turned to calc-alkaline andesite magmatism. The emergence of calc-alkaline magmatism in ensialic mature
arcs in the framework of the model [76,91] is associated with the dehydration of serpentine at depths of about 100 km, as a result of which quartz eclogites in the upper part of the subducted plate undergo melting to form felsic (dacite and rhyodacite) melts. Fluid and melt infiltration into the lherzolite layer of the mantle wedge, and processes at the lower crust, caused the generation of basaltic andesite and andesite magmas of the calc-alkaline front of a mature arc coeval with the intrusion and crystallization of calc-alkaline plutonic rocks.

Stage III: Group III and IV are formed during the third stage. The presence of E-MORB-type basalts in the studied area is related to the island-arc rifting and the initial stage of back-arc basin formation (initiation of Shishkhid island arc) [30]. For the formation of OIB-type basalts, a deep mantle source of magmas is required, in which the residual phases are dominated by garnet rather than plagioclase [92–94].

A common model of the origin of OIB-type rocks in an island-arc setting is the slab window model accompanied by the rifting, commonly in back arc environment. The interaction of asthenospheric melts and depleted mantle lithosphere leads to the formation of polygenic magmatic sources with the involvement of metasomatized suprasubduction mantle. The modern analogue of such settings is the active continental margin of the Kamchatka island arc [95–100]. Magma generation during the intra-arc spreading occurs at the garnet-spinel facies in the mantle accompanied by the thinning of the crust because of upwelling of asthenospheric mantle material [95–100]. The other model of the origin OIB-type rocks in an island-arc setting is the tectonic model when fore-arc volcanics are accreted with the oceanic fragments forming the ophiolitic mélangé.

We envisage that the deep mantle source is related to the melting of the asthenosphere due to slab window formation. The collision of Paleo-Asian island arcs with continental plates is a major cause of convergence geometry that can lead to slab window development. Once the collision of a divergent ridge enables the coupling of the divergent ocean floor to the continental crust through a transform fault, then the subducted plate detaches (or unzips), leaving a slab window [88–90,101]. The melting of slabs plunging into the subduction zones of the Neoproterozoic island arcs, and the formation of slab-window, leads to the formation of OIB magmas derived from the deep mantle [39,88–90,101–104]. At the convergent boundary of the Paleo-Asian Ocean and the Siberian paleocontinent, the island-arc volcanic activity and plume magmatism coincided [90,102]. Sub-alkaline and alkaline igneous rocks are also localized in the area adjacent to the Ulan-Sar’dag melange. U–Pb dating of the zircon from the trachyandesite has estimated its age at 833 ± 4 Ma, which likely corresponds to a global tectono-magmatic event correspondent to mantle plume activity [87,89,90,102]. The manifestation of the subalkaline dolerite dikes with an age of 829 Ma confirms these events.

6. Conclusions

(1) The Ulan-Sar’dag melange (USM) is composed of a variety of structurally mixed different units – ophiolitic ultramafic-mafic rocks with podiform chromitites, foliated serpentinites, cherts, volcanic-sedimentary rocks with MORB, and island-arc and OIB geochemical affinities, and metasedimentary rocks.

(2) Igneous rocks, according to geochemical affinities, are divided into four groups. Group I belongs to ensimatic island-arc boninite. It is characterized by low concentrations of REE and negative anomalies of HFSE (Nb, Ta, and Ti). It has a pronounced positive anomaly in Sr and low relationships of (La/Yb)$_n$, (Ce/Y)$_n$, (Th/Y)$_n$, (La/Ta)$_n$, and (Zr/Nb)$_n$. According to the ratios of trace elements, boninite is close to N-MOR basalts. On the tectonic discrimination diagrams, it lies within the island-arc volcanic rock and boninite fields. Group II belongs to island-arc rocks. The igneous rocks of this group have a slightly pronounced negative slope of the REE pattern and a distribution of REE, HFSE, and LILE, which is similar to a continental crust. The rocks have a negative anomaly for Ta, Nb, and Ti. In a number of samples, positive anomalies in Sr, Ba, Rb, and Zr are recorded. The (La/Nb)$_n$, (Ce/Y)$_n$, (Th/Y)$_n$, and (La/Yb)$_n$ ratios in these rocks correspond to values of the upper and lower continental crust. On the tectonic diagrams, these rocks plot in fields
of volcanic arc, island-arc tholeiite, calc-alkaline, basalts, and continental arc. It is worth noting the elevated abundance of Nb, Ti, Y, and Zr in some samples, which is not typical for intermediate island-arc rocks. The rocks of the second group were generated from island arc-magmas with a significant contribution of a crustal component. Group III corresponds to E-MOR basalts. These rocks show a distribution of REE and HFSE that corresponds to E-MORB. Furthermore, the presence of a subduction component is noted and indicated by the elevated contents of LREE, Rb, Ba, and Sr, which differs from contents E-MORB. According to trace element ratios, metabasalts show hybrid characteristics of N-MORB, E-MORB basalts, and continental crust components. They are slightly enriched in LREE relative to E-MORB. On the tectonic discrimination diagrams, these rocks plot in the N-MORB–E-MORB field and on an N-MORB–OIB trend. The basalts appear to have formed in the mid-ocean ridge setting from an enriched magmatic source. The igneous rocks of group IV have contents and patterns of REE and HFSE identical to those of OIB basalts. The rocks show a negative anomaly in Sr, and high ratios of (La/Yb)$_n$ and (Nb/Y)$_n$. On the tectonic discrimination diagrams, these rocks plot in the OIB field. They could have been formed at the plume-magmatism stage, from an enriched deep magmatic source.

(3) Three stages of magmatism are proposed according to the evolution of the subduction zone in the studied segments of CAOB. Stage I: In the Meso-Neoproterozoic at the closure of the Paleo-Asian Ocean, intra-oceanic island-arc systems were formed. At an early stage, an ensimatic island arc on oceanic-type crust was formed with generation boninitic melts. Stage II: During the development to an ensialic island arc, boninite magmatism turned to calc-alkaline andesite-dacite magmatism. Stage III: At a late Neoproterozoic–early Paleozoic period, island-arc rifting results in the initial stage of back-arc basin formation (initiation of Shishkhid island arc). The third stage is likely associated with plume-magmatic activity that led to the slabs plunging into the subduction zones and the formation of a slab-window. Mixed enriched magmas were generated.

(4) U–Pb dating of zircons from the trachyandesite, belonging to the second geochemical type, yielded a crystallization age of 833 ± 4 Ma, which is interpreted as the crystallization age during mature island-arc and intra-arc rifting stages.

Supplementary Materials: The following are available online at [link], Table S1: Representative analyses of olivine, chromspinel, pyroxene (wt.%); Table S2: Representative analyses of amphiboles (wt.%) and calculated values of temperature and pressure; Table S3: Representative analyses of plagioclase chemical composition (wt.%); Table S4: Representative chemical composition of the biotites (wt.%); Table S5: Representative chemical analyses of the ilmenite, titanite (wt.%); Table S6: Representative whole-rock and trace element analyses of cumulates, diorites and metavolcanic rocks of Ulan-Sar’dag ophiolitic mélangé.


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