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Mafic–Ultramafic Tyulkulam Massif, West–Central Asian Orogenic Belt (Central Kazakhstan): Mineralogical and Geochemical Evidence for Transition from Forearc to Island-Arc Magmatism during the Evolution of the Junggar–Balkhash Ocean

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Mafic–Ultramafic Tyulkulam Massif, West–Central Asian Orogenic Belt (Central Kazakhstan): Mineralogical and Geochemical Evidence for Transition from Forearc to Island-Arc Magmatism during the Evolution of the Junggar–Balkhash Ocean

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ABSTRACT

We present here the first comprehensive mineralogical, textural, and geochemical data and constraints on the tectonomagmatic evolution of ultramafic-mafic lithologies of the Tyulkulam Massif in the eastern North Balkhash ophiolite zone within the West Central Asian Orogenic Belt of Central Kazakhstan. The Tyulkulam Massif represents a large tectonic sliver within a Silurian to Middle Devonian flysch sequence. The massif consists mainly of a serpentinite mélange hosting blocks and tectonic lenses of residual and cumulate harzburgite-lherzolite-pyroxenite, gabbroids, as well as porphyritic granodiorite and plagiogranite bodies, and numerous intrusions of intermediate to felsic dykes and veins. Harzburgites in the Tyulkulam Massif are highly depleted upper mantle residues produced by high degrees of partial melting (17–22%). Geochemical signatures indicate that ultramafic–mafic rocks represent cumulate magmatic series derived from boninitic and tholeiitic melts generated in a suprasubduction zone setting under moderate pressure conditions. Upper mantle rocks and ultramafic–mafic cumulates of the Tyulkulam Massif are interpreted as fragments of different structural levels of oceanic lithosphere formed during the progressive transition in melt evolution from forearc to island-arc magmatism in the early stages of the development of the Junggar-Balkhash Ocean.

KEY WORDS*:* Ophiolite; suprasubduction zone oceanic crust formation; Junggar-Balkhash Ocean; Central Asian Orogenic Belt; Boninitic to tholeiitic melts; Forearc mantle peridotites; Geochemical transition from forearc to island arc magmatism.

INTRODUCTION

Ophiolites occur within a series of narrow, elongated tectonic zones in the western segment of the Central Asian Orogenic Belt, where Palaeozoic orogenic systems of Kazakhstan, the Tien Shan, and northwestern China are exposed. A fully preserved ophiolite sequence displaying primary relationships between all lithological units has been documented only at the Karaulcheku locality in the southwestern Bayanaul ophiolite zone of the NE Central Kazakhstan (Yakubchuk et al., 1988, 1989; Stepanets, 2016). In other tectonic zones, ophiolites are represented only by dismembered fragments of different structural levels occurring either as fault slivers or as blocks enclosed within serpentinite–matrix mélanges (Stepanets, 2016; Degtyarev et al., 2021, 2022, 2024; Guo et al., 2020; Milyukova et al., 2023). Ophiolitic rocks in these fragments display significant differences in their geochemical signatures because they might have likely formed in different stages of the evolution of an intra-oceanic subduction zone system.

The North Balkhash zone of Central Kazakhstan hosts one of the largest ophiolite belts in the western segment of the Central Asian Orogenic Belt, where lithological fragments representing the lower structural levels of ophiolite sequences are widely retained forming large tectonic blocks within serpentinite–matrix mélanges, which are traditionally referred to as ultramafic–mafic massifs (Mikhailov 1971). Despite serpentinization and other secondary alteration processes, ultramafic and gabbroic rocks comprising these massifs commonly preserve primary mineral parageneses and thus serve as important records of their evolution. Our research group has previously conducted detailed studies of ultramafic-mafic massifs situated in the western and central parts of the North Balkhash ophiolite zone (Degtyarev et al., 2021; Milyukova et al., 2023). However, ultramafic-mafic rocks of the Tyulkulam Massif, located in the eastern segment of the North Balkhash ophiolite zone, have remained poorly constrained in terms of their geochemical signatures, ages, and tectonic evolution, despite preserving primary whole-rock and mineral

compositions. In addition, whether the complexes of the Tyulkulam Massif are comparable with those from the western North Balkhash ophiolite zone has remained unresolved. These limitations hinder, therefore, our reconstruction of a coherent tectonic evolution of the Junggar-Balkhash Ocean.

Overall, information on the geology, internal structure, and mineralogical and geochemical compositions of ultramafic-mafic complexes in the lower parts of the ophiolite sequences occurring in the western Central Asian Orogenic Belt remains limited. In this study, we focus on the whole-rock and mineral chemistry of poorly studied ultramafic–mafic rocks of the ophiolitic suites in the eastern North Balkhash zone, with a particular emphasis on establishing possible cogenetic relationships amongst magmatic rocks and reconstructing the tectonic and magmatic settings of their formation.

GEOLOGICAL OUTLINE

In the western part of the Central Asian Orogenic Belt (CAOB), ophiolite zones are mainly confined to the Kokchetav–North Tien Shan (northern and western Central Kazakhstan, southern Kazakhstan, and the North Tien Shan), Boshchekul-Chingiz (northeastern and eastern Kazakhstan), and Junggar-Balkhash (Central Kazakhstan and northwestern China) regions, which display contrasting structural architectures, different ages of the constituent complexes, and discrete timing of continental crust formation. Siliceous, siliceous-basaltic, and siliceous volcaniclastic sequences make up a significant part of all these ophiolite zones, whereas serpentinite–matrix mélanges that contain fragments of plutonic rock suites are less widespread. These siliceous and volcaniclastic rock sequences typically exhibit a fold and thrust–type structural architecture, and primary stratigraphic relationships between them are rarely preserved.

The ages of the plutonic, siliceous-basaltic, and siliceous rock sequences in the ophiolite zones in different parts of the orogenic belt vary significantly. The ones in the Kokchetav–North Tien Shan and Boshchekul–Chingiz regions are Cambrian to Early Ordovician in age, whereas those in the Junggar–Balkhash region display a broader temporal range extending from Cambrian to Late Devonian (Degtyarev and Tretyakov, 2022).

Fragments of mafic and ultramafic rocks within the ophiolite complexes in the western CAOB occur only in discrete zones, where they form tectonic slivers and blocks of limited extent within serpentinite-matrix mélanges. The best-preserved and most representative ophiolite fragments occur in the Bayanaul and Maikain–Kyzyltas zones of the Boshchekul–Chingiz zone, the Dzhalair–Naiman zone of the Kokchetav–North Tien Shan region, and the Tekturmas, North Balkhash, and Tangbale zones within the Junggar-Balkhash area. The ~20–35 km-wide North Balkhash ophiolite zone, which is situated in the central Junggar–Balkhash region of Central Kazakhstan extends for nearly 250 km along-strike in the east–west direction. In its western segment, the assemblages of the North Balkhash ophiolite zone are unconformably overlain by Upper Devonian to Carboniferous extrusive and sedimentary sequences. To the east, this zone is bounded by the Junggar dextral strike-slip fault (**Fig. 1**). The North Balkhash zone consists of a serpentinite-matrix mélange and Middle to Upper Ordovician siliceous, siliceous-basaltic, siliceous volcaniclastic, and extrusive and sedimentary rock sequences forming intensely folded tectonic slivers. To the north, Upper Ordovician volcanogenic lithologies of the North Balkhash zone are conformably overlain by a Silurian flysch, whereas siliceous-basaltic and siliceous sequences to the south are tectonically overlain by a Silurian olistostromal unit that contains blocks of Middle to Upper Ordovician cherts.

The serpentinite–matrix mélange occupies the structurally lowest position and contains fragments of ophiolitic plutonic rocks. Locally, in the western segment of the North Balkhash

zone, the mélange displays a polymictic character, with serpentinites enclosing blocks of gabbro, plagiogranite, dolerite, siliceous and basaltic rocks, and more rarely variably sized bodies of eclogite and garnet glaucophane schist spatially associated with jadeitite and other metasomatic rocks (Mikhailov, 1971; Degtyarev et al., 2021; Skoblenko et al., 2024). The North Balkhash zone is distinguished by the widespread occurrence of large fragments of mafic and ultramafic rocks forming individual ophiolite massifs. In the western part of the zone, there exists a residual dunite-harzburgite complex known as the Itmurundy Massif, a dunite-wehrlite-clinopyroxenite-gabbro complex constituting the Arkharsu Massif, and a gabbro-dolerite complex making up the East Arkharsu Massif (**Supplementary 1**; Mikhailov, 1971; Degtyarev et al., 2021; Milyukova et al., 2023). The plutonic parts of the mafic–ultramafic rock complexes in the western North Balkhash zone represent Lower Cambrian forearc units (**Supplementary 1**; Degtyarev et al., 2021; Milyukova et al., 2023).

The eastern North Balkhash zone comprises the Chulkyzyl Massif in its western part, the Tyulkulam and South massifs in the central segment, and the East Massif in the eastern part. Mafic and ultramafic rocks of the Tyulkulam Massif (**Fig. 2**) show the highest degree of preservation of the primary whole-rock and mineral chemistry features, and thus form the principal focus of our investigation in this study.

INTERNAL STRUCTURE AND STRATIGRAPHY OF THE TYULKULAM MASSIF

Since the prominent rock types in the Tyulkulam Massif are poorly exposed, field investigations involved excavation in numerous trenches during geological mapping in the 1950s (Semenov, 1960b). However, intensive alteration of rock units in the trenches hindered our recognition of the original contact relationships.

137 The Tyulkulam Massif comprises a large, east–west–striking, 23-km-long and 0.5–0.8-km-
138 wide tectonic block, bounded by Silurian to Middle Devonian flysch sequences. The massif
139 includes a chrysotile-bearing serpentinite–matrix mélange hosting blocks of ultramafic rocks of
140 various sizes, diverse gabbroic and granitoid rocks, as well as basalts with lenses of grey cherts
141 (**Fig. 2**) (Semenov, 1960b). Ultramafic rocks in the massif are extensively serpentinized, but a
142 nearly 4-km-long block of a relatively unaltered peridotite near its southern boundary consists
143 mainly of harzburgite with local dunite interlayers. Fine- to medium-grained amphibolized gabbro
144 and gabbro-diorite bodies ranging in length from 2–3 m to several tens of meters widely occur in
145 the massif. Medium-grained gabbro bodies (samples B2111 and B2114) intrude the ultramafic
146 block as dykes in the vicinity of the southern boundary of the massif. In addition, two large blocks
147 of amphibolized gabbro have been observed in the northwestern and central parts of the Tyulkulam
148 Massif. In the central part of the massif, small bodies (up to 5 m across) of coarse-grained lherzolite
149 and olivine-websterite also occur within these gabbros (sample B2108). These ultramafic rocks
150 constitute a distinctive vein-like complex with a symmetrical internal structure: the central parts
151 are composed of lherzolite, whereas the marginal zones consist of pyroxenite forming a transitional
152 zone between the lherzolite and gabbro (Semenov, 1960b). Due to limited exposures in the field,
153 direct contact relationships between these rock types and their orientation were not well
154 established. Therefore, we interpret them as probable xenoliths of older rocks enclosed within
155 gabbro bodies, rather than as separate intrusions across the gabbro.
156 Serpentinite and serpentinized ultramafic rocks of the Tyulkulam Massif are intruded by
157 abundant felsic rocks, mainly concentrated in its western part. These felsic rocks comprise
158 relatively large porphyritic granodiorite and plagiogranite bodies and numerous felsic to
159 intermediate dykes and veins, most of which show sharp intrusive contacts with their host rocks.
160 Although the timing of the formation of the rock bodies within the Tyulkulam Massif is not directly

161 constrained, the age of ca. 300 Ma obtained from granitoids intruding the serpentinite and
162 serpentinized ultramafic rocks delineates the minimum age of the mélange formation
163 (**Supplementary 2, Table S9**).
164
165 **ANALYTICAL METHODS**
166 Petrographic studies were performed with a Carl Zeiss Axioskop 40 polarizing microscope
167 at the Laboratory of Late Precambrian and Phanerozoic Geodynamics, Geological Institute,
168 Russian Academy of Sciences (GIN RAS), Moscow, Russia.
169 Whole-rock major element concentrations (SiO_2, TiO_2, Al_2O_3, Fe_2O_3, FeO, MnO, MgO,
170 CaO, Na_2O, K_2O, P_2O_5, and loss on ignition, LOI) in the samples were determined by X-ray
171 fluorescence (XRF) using a sequential S4 Pioneer spectrometer (Bruker, Germany) operated with
172 Spectra-Plus software at the Chemical Analytical Laboratory of the GIN RAS. The applied
173 analytical procedure allows determination of elements within the following concentration ranges
174 (expressed as oxides, wt.%): Si 0.01–1.0, Ti 0.01–5.0, Al 1.0–60.0, total Fe 1.0–40.0, Mn 0.01–
175 1.0, Ca 1.0–50.0, Mg 0.1–40.0, Na 0.1–10.0, K 0.1–10.0, and P 0.01–5.0. Certified reference
176 materials compositionally similar to the analyzed samples (harzburgite 38A HARZO1, MU-3,
177 OSO, and MSO) were used for calibration and quality control.
178 Whole-rock trace element concentrations (Li, Be, Sc, Ti, V, Cr, Co, Ni, Cu, Zn, Ga, Rb,
179 Sr, Y, Zr, Nb, Cs, Ba, La, Ce, Pr, Nd, Sm, Eu, Gd, Tb, Dy, Ho, Er, Tm, Yb, Lu, Hf, Ta, Pb, Th,
180 and U) were determined by inductively coupled plasma mass spectrometry (ICP-MS) using a
181 Perkin Elmer ELAN 6100 DRC operated in standard measurement mode. Additional analyses
182 were carried out by inductively coupled plasma atomic emission spectrometry (ICP-AES) on an
183 ICAP-61 instrument (Thermo Jarrell Ash). Measurements were performed at the Analytical
184 Certification Testing Centre of the Institute of Microelectronics Technology and High Purity

185 Materials, Russian Academy of Sciences (Chernogolovka, Moscow Region, Russia), in the
186 Laboratory of Nuclear-Physical and Mass-Spectrometric Methods of Analysis. The analytical
187 accuracy was monitored using the following standard samples: gabbro SGD-1A (GSO 521-84P),
188 gabbro SGD-2A (GSO 8670–2005), basalt BHVO-2 (U.S. Geological Survey), and andesite AGV-
189 2 (U.S. Geological Survey).

190 SIMS U–Pb isotope analyses of zircons have been carried out on SHRIMP II ion probe at
191 the Center for Isotopic Research at A.P. Karpinsky Russian Geological Research Institute (St.
192 Petersburg, Russia). U and Pb isotope ratios were measured following the conventional analytical
193 procedure described by Williams (1998) and Larionov et al (2004). The primary beam of molecular
194 negative oxygen ions had an intensity of ~2.5–4 nA, and the analytical spot (crater) size was ~15
195 × 10 μm. Zircon morphology and internal structures were examined using secondary electron and
196 cathodoluminescence imaging on a Camscan MX 2500S scanning electron microscope. The data
197 have been processed with the SQUID v1.12 and Isoplot R (Vermeesch, 2018) programs using the
198 decay constants of Steiger and Jager (1976). The analytical results of dating are listed in Table S9.

199 Representative mineral compositions were determined using an electron probe
200 microanalyzer (EPMA) JEOL JXA-8230 equipped with five wavelength-dispersive spectrometers
201 and one energy-dispersive spectrometer, operated at an accelerating voltage of 20 kV and a beam
202 current of ~20 nA at the Earth Sciences Department, University of Grenoble (France). In addition,
203 scanning electron microscopy investigations were carried out using a TESCAN VEGA3 SBH
204 SEM equipped with an ULTIM MAX 40 X-ray detector (Oxford Instruments) at the Geological
205 Institute of the Russian Academy of Sciences (GIN RAS), and a TESCAN MIRA 3 SEM with an
206 ULTIM MAX 100 EDS detector (Oxford Instruments) at the Vernadsky Institute of Geochemistry
207 and Analytical Chemistry (Moscow, Russia). The EDS analysis was operating at 20 kV
208 accelerating voltage, beam current of 800–900 pA, and $6x10^6$ impulses. The multielement standard

was used for the EDS analyses, C321 (Oxford Instruments set of standards) and Smithsonian Institution set of standards.

Clinopyroxene trace element compositions were measured by LA-ICP-MS using an NWR-213 laser system coupled to an Element-2 high-resolution ICP-MS at GIN RAS (Moscow, Russia). Analyses of the clinopyroxene grains free of exsolution lamellae were performed on polished thin sections previously characterized by EPMA. The laser fluence was set to ~ 10 J/cm^2 at the sample surfaces with laser repetition rate and laser diameter set to 10 Hz and 40 μm, respectively. Each spot analysis consisted of an approximately 30 s background and 60 s sample data acquisition. At the ablation, He flow with subsequent mixing of gasified Ar is used to transport the substance vaporized by the laser before the ionization at the ICP-MS. Fine filtration and mixing of the gas streams are used to reduce the gas background and increase the stability of the analytical signal. The obtained data were processed in Iolite software (Paton et al., 2011) using the "Trace Elements" data reduction scheme. NIST SRM 610 (Norman et al., 1996) was used as primary standard. The clinopyroxene Si contents, measured by EPMA, served as an internal standard. The limits of detection for each measured isotope were determined by Longerich et al. (1996). To control the quality of the results obtained, standard samples of synthetic glass NIST-612 and synthetic basalt glass GSC-2g, GSD-2g, and GSE-2g (USGS) were analyzed along with the routine samples every seven analyses. The measurement values of the elements in the secondary standards are comparable to the certified values. Most elements have no more than 15% relative errors in their content (up to 20 rel % for Ni, Y, and REE).

PETROGRAPHY AND MINERAL CHEMISTRY OF THE KEY LITHOLOGIES IN THE TYULKULAM MASSIF

Representative mineral compositions of the analyzed rocks are presented in **Tables S2–S7**. Mineral abbreviations follow Whitney & Evans (2010).

Harzburgite suite

Among the peridotite bodies examined, only harzburgite has been considered in this study, as no representative varieties suitable for further analysis were identified amongst the sampled dunite. The harzburgite exhibits a porphyroclastic texture bearing clear evidence of plastic deformation (**Fig. 3a**). These rocks underwent intense serpentinization, during which fibrous serpentine formed a mesh texture with relict grains of primary olivine and chromian spinel.

Olivine occurs as rounded anhedral porphyroclasts, reaching up to 2 mm in size, and displays a homogeneous composition close to primary mantle values ($Fo_{90–91}$; NiO = 0.32–0.45 wt.%) (**Fig. 3b**). Olivine is commonly affected by secondary serpentinization, resulting in recrystallization into aggregates of fine sub-grains (≤ 0.2 mm) characterized by irregular undulatory extinction (**Fig. 3c**). Such olivine grains exhibit significant variations in forsterite content as well as in Ni and Mn contents, which are discriminative of metamorphic olivine ($Fo_{83–95}$; NiO = 0.20–0.48 wt.%, MnO = 0.03–0.67 wt.%). Chromian spinel occurs as rare subhedral to fragmented grains up to 1 mm in size (**Fig. 3d**) and as anhedral grains reaching 0.8 mm in length (**Fig. 3c**). Relict primary spinels are compositionally uniform within samples and are characterized by extremely low TiO_2 contents (commonly below the detection limit), low F# ($Fe^{3+}/(Fe^{3+}+Al+Cr)$ = 0.01–0.06), and Cr# (Cr/(Cr+Al) and Mg# (Mg/(Mg+Fe) ranges of 0.50–0.71 and 0.49–0.60, respectively. Porphyroclastic orthopyroxene is also present, but it is commonly replaced by serpentine to form bastite pseudomorphs.

Besides serpentine, magnetite and talc are common secondary minerals. Magnetite forms thin veins within the serpentine mesh textures (**Fig. 3b**) or constitutes fine disseminated grains,

and locally replaces chromian spinel along the grain rims (**Fig. 3c, d**). Small sulfide inclusions also exist within olivine grains.

Lherzolite-pyroxenite suite

Lherzolite occurrences in the Tyulkulam Massif constitute coarse-grained rocks displaying an orthocumulate texture, with euhedral olivine grains enclosed by intercumulus pyroxene and amphibole grains (**Fig. 4a**). The lherzolite consists of olivine (72–75 vol.%), clinopyroxene (15–16 vol.%), orthopyroxene (4–5 vol.%), and amphibole (3–4 vol.%), with minor accessory spinel. Olivine exists as large anhedral grains replaced by mesh-textured serpentine forming a network of veinlets, which contain aggregates of magnetite in their central parts (**Fig. 4a, d).** Clinopyroxene occurs as subhedral grains up to 0.5 mm in size and locally replaced by amphibole along grain margins (**Fig. 4b**). Orthopyroxene grains are commonly anhedral in shape and are locally replaced by amphibole at their rims. Rare relics of primary amphibole are also preserved in interstitial sites between other rock-forming minerals (**Fig. 4a**). Talc is locally present as a secondary phase. Spinel occurs as small, individual grains reaching up to 0.3 mm in size (**Fig. 4a, c**).

The olivine websterite in the Tyulkulam Massif is a coarse-grained rock characterized by a panidiomorphic cumulate texture and a massive fabric (**Fig. 5a**). They consist of olivine (up to 23 vol.%), clinopyroxene (up to 39 vol.%), orthopyroxene (up to 36 vol.%), and minor chromian spinel (up to 2 vol.%). Pyroxene occurs as prismatic grains, locally subhedral, and commonly contains clinopyroxene exsolution lamellae within orthopyroxene (**Fig. 5b**). Spinel presents both as anhedral interstitial grains up to 0.5 mm in size and as euhedral inclusions within pyroxene crystals. Chromian spinel occurs as fine, euhedral to subhedral grains not exceeding 0.2 mm in size. Secondary phases comprise serpentine and talc developed after orthopyroxene and olivine, together with amphibole, replacing clinopyroxene.

Mineral compositions of the lherzolite and olivine websterite are nearly identical. Olivine in these rocks is characterized by high forsterite content ($Fo_{73–84}$) and relatively low Ni contents (NiO = 0.11–0.20 wt.%). Clinopyroxene shows comparatively high Mg# (0.83–0.91), low Al_2O_3 contents (1.23–2.42 wt.%), and moderate TiO_2 (0.10–0.21 wt.%) and Na_2O (0.11–0.26 wt.%) concentrations. Orthopyroxene shows limited variations in Mg# (0.81–0.85), moderate Al_2O_3 contents (1.38–1.70 wt.%), and low TiO_2 (0.07–0.18 wt.%) and Na_2O concentrations (Na_2O <0.01 wt.%). Primary amphibole is characterized by high Mg# values (0.81–0.86) and elevated Al_2O_3 (7.73–11.36 wt.%) and Cr_2O_3 contents (0.71–0.98 wt.%), together with moderate TiO_2 concentrations (0.10–0.50 wt.%). Metamorphic amphiboles (actinolite-tremolite) exhibit variable Mg# (0.66–0.92), Al_2O_3 (1.22–6.07 wt.%), and TiO_2 contents (0.06–0.32 wt.%). Chromian spinel shows relatively high Cr# values (0.54–0.66), elevated TiO_2 contents (0.50–1.01 wt.%), and higher F# (0.19–0.24). Spinels are commonly characterized by pronounced compositional zoning with Mg–Al-rich cores and Cr–Fe-enriched rims. A major part of the spinel grains is represented by ferrichromite rims (Cr# = 0.62–0.66; Fe# = 0.28–0.32; TiO_2 = 0.74–1.02 wt.%; Mg# = 0.23–0.25), whereas the most strongly modified spinels are overgrown by magnetite rims, reflecting late-stage alteration during serpentinization and/or metamorphism. The composition of metamorphic spinel is marked by depletion in Al_2O_3 (4.58–4.78 wt.%) and Mg# (0.06–0.07) coupled with enrichment in Cr# (0.67–0.68) and Fe^{3+}# (0.64–0.67).

Gabbroid suite

Gabbroids hosted by harzburgite blocks in the massif (samples B2111 and B2114) are represented by medium-grained amphibolized gabbro with commonly phaneritic, locally ophitic textures, with weak idiomorphism of plagioclase relative to pyroxene and amphibole grains (**Fig. 6a, b**). The primary mineral assemblage comprises plagioclase (35–44 vol.%), clinopyroxene (45–

303 56 vol.%), and amphibole (6–9 vol.%). Plagioclase forms mainly short prismatic to tabular grains,
304 locally saussuritized and partly replaced by albite along grain margins. Its composition
305 corresponds to bytownite–anorthite ($An_{84–93}$) and is marked by relatively high FeO contents (0.10–
306 0.82 wt.%), typical of primary magmatic plagioclase. Clinopyroxene forms euhedral short tabular
307 grains (Mg# = 0.71–0.86; Al_2O_3 = 1.00–2.14 wt.%; TiO_2 = 0.11–0.26 wt.%; Na_2O = 0.14–0.21
308 wt.%) that are partially or completely replaced by actinolitic amphibole (Mg# = 0.59–0.75; Al_2O_3
309 = 1.68–3.88 wt.%; TiO_2 = 0.18–0.37 wt.%) along grain margins. Primary amphibole fills
310 interstitial spaces between major mineral phases (Mg# = 0.65–0.67; Al_2O_3 = 9.04–9.63 wt.%; TiO_2
311 = 0.96–0.99 wt.%). Secondary amphibole is commonly altered to chlorite. Accessory minerals
312 include magnetite and apatite, the latter occurring as needle-like crystals.

313 Gabbro that is hosting coarse-grained lherzolite and olivine websterite bodies displays a
314 porphyritic texture (sample B2108), defined by large clinopyroxene phenocrysts, which are
315 partially replaced by amphibole along their rims (**Fig. 6c**). Clinopyroxene shows Mg# = 0.79–0.83,
316 Al_2O_3 = 1.88–2.51 wt.%, TiO_2 = 0.28–0.42 wt.%, and Na_2O = 0.17–0.26 wt.%, whereas amphibole
317 rims exhibit Mg# = 0.78–0.81, Al_2O_3 = 2.84–3.52 wt.%, and TiO_2 = 0.25–0.34 wt.%. Plagioclase
318 unaffected by secondary alteration is locally preserved and corresponds to bytownite ($An_{89–90}$).
319 However, in most cases, the central parts of plagioclase crystals are strongly saussuritized, locally
320 replaced by sericite, and rimmed by albite.

321 Gabbro–diorite (sample B2105/1) rocks are represented by fine-grained varieties with an
322 ophitic texture (**Fig. 6d**). Plagioclase of anorthite composition (45–46 vol.%; $An_{97–98}$) forms short
323 tabular crystals with irregular boundaries. In most cases, the crystal cores are saussuritized,
324 whereas the grain margins are replaced by albite. Relict primary clinopyroxene (36–37 vol.%)
325 (Mg# = 0.84–0.85; Al_2O_3 = 1.91–1.97 wt.%; TiO_2 = 0.13–0.14 wt.%; Na_2O = 0.12–0.14 wt.%) is
326 preserved within irregular tabular amphibole crystals. Primary amphibole (17–18 vol.%) (Mg# =

0.55–0.58; Al_2O_3 = 6.97–7.27 wt.%; TiO_2 = 0.76–1.03 wt.%) occurs as short tabular grains reaching up to 0.5 mm in length and is commonly altered to fine-grained chlorite along grain rims. Accessory phases comprise magnetite and apatite.

Thus, the harzburgite of the Tyulkulam Massif displays features reminiscent of those typically observed in a residual mantle: it shows relatively uniform compositions, lacks plagioclase and clinopyroxene, contains accessory chromian spinel, and preserves deformation-related textures. Cumulate rocks are recognized based on their characteristic rock association (lherzolite, olivine websterite, and gabbro), preservation of relict cumulate textures, and the modal predominance of clinopyroxene.

TRACE ELEMENT GEOCHEMISTRY OF THE KEY LITHOLOGIES IN THE TYULKULAM MASSIF

Whole-rock geochemistry

Representative chemical compositions of the analyzed rocks from the Tyulkulam Massif are presented in **Table S1**.

Harzburgite suite

The analyzed harzburgite samples show moderate to high loss-on-ignition values (LOI = 4.81–8.34 wt.%), consistent with the development of serpentinization and talc alteration.

The FeO/SiO_2 (0.108–0.176), Cr_2O_3/SiO_2 (0.006–0.008), and CaO/Al_2O_3 (0.123–1.282) ratios of the studied rocks fall within or are close to the ranges typical of isochemically serpentinized residual mantle peridotites (FeO/SiO_2 = 0.170–0.202; Cr_2O_3/SiO_2 = 0.0065–0.0119; $CaO/Al_2O_3 \leq 2.0$; Bazylev, 1998). However, samples with elevated SiO_2 contents (41.42–46.45 wt.%) show a lower FeO/SiO_2 ratio, suggesting possible silica enrichment during secondary

alteration. In addition, the harzburgite is characterized by a low MgO/SiO_2 ratio (0.85–1.00), which may also reflect silica addition during alteration to talc. All samples show high Mg# values (91–92) and low Al_2O_3 (0.46–0.83 wt.%) and CaO contents (0.06–0.83 wt.%), consistent with the absence of clinopyroxene, and are enriched in compatible elements, including Cr (1504–2585 ppm), Ni (2221–2896 ppm), and to a lesser extent, Co (102–133 ppm).

Harzburgite exhibits strongly depleted REE contents with relative enrichment in light REE (LREE) compared to moderate REE (MREE) ($(La/Sm)_n < 3.8$) and heavy REE (HREE) ($(La/Yb)_n < 2.13$) (**Fig. 7a**). The analyzed samples are enriched in large-ion lithophile elements (LILE; $Rb/Rb_{PM} = 0.13–0.21$; $Cs/Cs_{PM} = 3.31–12.78$) relative to high field strength elements (HFSE; $Zr/Zr_{PM} = 0.04–0.07$; $Y/Y_{PM} < 0.04$) (**Fig. 7b**). Multi-element patterns of the analyzed samples display pronounced positive Pb and Sr anomalies.

Lherzolite-pyroxenite suite

Compared to the mantle peridotites, the cumulate peridotites are characterized by more evolved compositions and higher abundances of mafic components (**Table S1**). In these ultramafic cumulates, FeO contents increase, whereas TiO_2, CaO, and Na_2O contents decrease with increasing magnesian values from olivine websterite towards lherzolite. Moreover, lherzolite and pyroxenite contain significantly lower Ni concentrations (867–871 ppm) than harzburgite (2221–2896 ppm).

Chondrite-normalized REE patterns reveal variable behavior of LREE, showing both depletion and enrichment relative to MREE ($(La/Sm)_n = 0.24–2.18$) and HREE ($(La/Yb)_n = 0.12–1.48$) (**Fig. 7c**). Primitive mantle-normalized multi-element diagrams indicate general enrichment in LILE ($Rb_{PM} = 0.31–0.49$; $Cs_{PM} = 3.05–6.07$) compared to HFSE ($Zr_{PM} = 0.15–0.16$; $Hf_{PM} = 0.10–0.20$), together with distinct positive U, Pb, and Sr anomalies (**Fig. 7d**).

375 ***Gabbro suite***

376 Gabbros exhibit pronounced compositional differences depending on their intrusive nature.
377 Those gabbros occurring as dykes within harzburgite blocks (samples B2111, B2114) are less
378 magnesian (Mg# = 42–64) and contain lower K_2O (0.07–0.36 wt.%), but higher TiO_2 (1.26–1.81
379 wt.%) and P_2O_5 (0.16–0.20 wt.%) contents compared to gabbros associated with lherzolite–
380 pyroxenite bodies (sample B2108; Mg# ≈74; K_2O = 1.44 wt.%; TiO_2 = 0.28 wt.%; P_2O_5 <0.01
381 wt.%). Both gabbro types are enriched in Al_2O_3 and CaO, reflecting the high modal proportion of
382 plagioclase. Gabbro-diorite occurrences are distinguished by even lower Mg# (~36) and CaO, and
383 by elevated FeO and Na_2O contents (**Table S1**).

384 Most gabbro samples display similarly shaped REE patterns, whereas sample B2108 shows
385 a distinctly different signature. Gabbroic dyke rocks within the harzburgite block are characterized
386 by weak LREE enrichment relative to MREE and HREE ($(La/Sm)_n$ = 1.1–1.5; $(La/Yb)_n$ = 1.5–
387 1.6), whereas sample B2108 displays pronounced LREE depletion ($(La/Sm)_n$ = 0.19; $(La/Yb)_n$ =
388 0.16) (**Fig. 7c**). In addition, the dyke-hosted gabbros are significantly enriched in HFSE and Y
389 (Ti_{PM} = 5.62–8.22; Nb_{PM} = 2.77–3.05; Y_{PM} = 5.49–8.40; Zr_{PM} = 4.22–4.41; Hf_{PM} = 4.41–5.54)
390 relative to the gabbros associated with lherzolite bodies (**Fig. 7d**).

391 The gabbro-diorite rock (sample B2105/1) shows a weakly fractionated REE pattern with
392 slightly lower LREE contents relative to HREE [$(La/Yb)_n$ = 0.8] and a negative Eu anomaly (Eu/Eu
393 = 0.769) (**Fig. 7c**). Primitive mantle-normalized multi-element patterns of all analyzed gabbroic
394 rocks display negative Nb–Ta, Zr–Hf, and Ti anomalies, together with positive Pb, U, and Sr
395 anomalies (**Fig. 7d**).

396

397 **Trace element distribution in clinopyroxenes from cumulative ultramafic rocks**

Clinopyroxene REE patterns are weakly fractionated and characterized by relative LREE depletion compared to MREE and HREE [$(La/Sm)_n$ = 0.13–0.19; $(La/Yb)_n$ = 0.09–0.13]. The overall low HREE abundances (e.g., $(Yb)_n \leq 1$) further distinguish these compositions. In addition, the trace-element spectra display pronounced negative Zr and Ti anomalies and a positive Sr anomaly (**Fig. 8; Table S8**).

DISCUSSION

PETROGENESIS AND TECTONOMAGMATIC SETTING OF THE MAFIC – ULTRAMAFIC ROCKS IN THE TYULKULAM MASSIF

The suprasubduction-zone setting of ophiolite formation is largely inferred from the geochemical signatures of associated volcanic rocks that reflect island-arc affinity (Dilek and Furnes, 2011). However, in the absence of preserved volcanic rock suites, residual peridotites in a mafic–ultramafic massif serve as reliable proxies for constraining the tectonomagmatic setting of their formation, as their whole-rock and mineral chemistry retains information on mantle processes and melt–rock interactions. For cumulate rocks, petrogenetic interpretations rely mainly on the compositions of primary minerals, since bulk-rock chemistry does not directly record the composition of parental magmas, and is therefore of limited use for reconstructing tectonic settings. To reconstruct the tectonic environment of formation of the mafic–ultramafic rocks in the Tyulkulam Massif, we conducted a comparative evaluation of some of the well-documented ophiolitic analogues.

Primary mineral assemblages of the ultramafic–mafic complexes in the Tyulkulam Massif

Information on the compositions of primary minerals in a mafic–ultramafic complex is particularly important, as it allows discrimination between rocks of different origins and reconstruction of their formation conditions. The rocks of the Tyulkulam Massif have been variably affected by serpentinization and metamorphism at greenschist to amphibolite facies conditions. Consequently, deciphering their affinities is best constrained by the compositions of relic primary minerals.

Trace element concentrations of olivine in the harzburgites (NiO = 0.36–0.45 wt.%; MnO = 0.12–0.23 wt.%) fall within the range of primary mantle olivine (**Fig. 9**; Takahashi et al., 1987). The chemical compositions of associated spinel grains (TiO_2 <0.05 wt.%; Fe^{3+}# = 0.01–0.06; Cr# = 0.50–0.71; Mg# = 0.49–0.60) suggest that they experienced only limited modification during later stages of alteration and metamorphism (Dick and Bullen, 1984; Ishii et al., 1992; Arai, 1994a), and that they closely resemble primary mantle spinel compositions (Khedr and Arai, 2017). Therefore, olivine and spinel from the studied rocks can be used as reliable petrogenetic indicators.

High Cr# values of spinel are widely regarded as indicative of high degrees of partial melting of the mantle source during the formation of ultramafic rocks in the Tyulkulam Massif (Dick and Bullen, 1984; Hellebrand et al., 2001). The Cr# values of spinels from the Tyulkulam harzburgite (Cr# = 0.50–0.71) suggest formation at relatively high degrees of partial melting (F ≈ 17–21%; Hellebrand et al., 2001). The extent of partial melting in different geodynamic settings, as reflected in the Cr# of primary spinels, varies within characteristic ranges. Peridotite spinels from mid-ocean ridge settings typically show relatively low Cr# values (0.10–0.40; Dick and Bullen, 1984; Arai, 1994a), as do residual spinels from back-arc basin peridotite (Cr# = 0.17–0.29; Ohara et al., 2002, 2003). In contrast, spinels from forearc peridotites are characterized by significantly higher Cr# values (0.38–0.80; Ishii et al., 1992; Parkinson and Pearce, 1998). Primary

446 spinels from the studied harzburgites display Cr# values comparable to those of peridotites from
447 forearc environments. In mantle peridotites formed in different tectonic settings, a well-established
448 correlation exists between olivine Mg# and spinel Cr# (**Fig. 10a**; OSMA – olivine-spinel mantle
449 array; Arai, 1994a), reflecting increasing degrees of partial melting in residual peridotites. The
450 Tyulkulam harzburgite plots within the olivine-spinel stability field, supporting their residual
451 mantle origin, and occupies the compositional range typical of forearc mantle peridotites.

452 The Cr# vs. Mg# relationship in harzburgitic spinels (**Fig. 10b**) shows an increase in Cr#
453 with increasing Mg#, consistent with trends typical of residual mantle peridotites (Dick and Bullen,
454 1984; Ishii et al., 1992). All analyzed samples fall within the suprasubduction zone peridotite
455 fields. In addition, the Cr# range of spinels from the Tyulkulam harzburgite overlaps with that
456 documented for peridotites of other massifs within the North Balkhash ophiolite zone (i.e.,
457 Itmurundy Massif with Spl Cr# 0.63–0.68). These compositional ranges are consistent with spinel
458 compositions from peridotites formed in forearc spreading settings (Pearce et al., 2000).

459 Olivine from the ultramafic cumulates shows a systematic decrease in its forsterite content
460 from lherzolite ($Fo_{81–82}$) to olivine websterite ($Fo_{72–77}$). NiO contents follow a similar trend,
461 decreasing from 0.18–0.20 wt.% in lherzolitic olivine to 0.11–0.20 wt.% in websteritic olivine.
462 The positive correlation between NiO and Fo observed in these rocks (**Fig. 9**) supports the use of
463 olivine composition as an indicator of fractional crystallization during the evolution of cumulate
464 assemblages.

465 Compositions of coexisting olivine and spinel in the lherzolite and websterite fall outside
466 the olivine-spinel mantle array field (OSMA; **Fig. 10a**), supporting their cumulate origin. Spinels
467 display compositional zoning characterized by enrichment of the cores in Mg and Al, and of the
468 rims in Cr and Fe, interpreted as a result of fractional crystallization processes (Barnes, 2000). In
469 addition, the spinels show elevated Fe and Ti contents, which are typical of derivatives formed

during low-pressure fractional crystallization (Barnes and Roeder, 2001). At the same time, Cr# values (0.54–0.61) fall within the range characteristic of cumulate spinels from mid-ocean ridge settings (Arai, 1994a). On compositional diagrams for spinels from plutonic rocks, the analyzed compositions plot within the island-arc field (**Fig. 10c**; Arai, 1994a).

Clinopyroxene is one of the most widespread minerals in the studied ultramafic–mafic cumulates, occurring in lherzolites to gabbros. The Mg# of clinopyroxene varies from 0.89–0.91 in lherzolites, 0.83–0.88 in websterites, and 0.76–0.86 in gabbroic rocks. Clinopyroxenes are characterized by low TiO_2 contents, averaging ~0.16 wt.% in lherzolites, ~0.18 wt.% in websterites, and ~0.28 wt.% in gabbros. TiO_2 contents show a negative correlation with increasing Mg#, consistent with fractional crystallization during magmatic differentiation.

Systematic variations in Al_2O_3 and TiO_2 contents in clinopyroxenes provide key constraints for distinguishing mafic rocks formed in mid-ocean ridge versus island-arc settings (Beccaluva et al., 1989). In clinopyroxenes coexisting with plagioclase, decreasing Mg# is typically accompanied by Al_2O_3 enrichment (DeBari et al., 1987; DeBari and Coleman, 1989). Clinopyroxenes from lherzolites and pyroxenites of the Tyulkulam Massif display such enrichment trends, suggesting crystallization before plagioclase accumulation (**Fig. 11a**), a feature also observed in clinopyroxenes from the gabbros associated with lherzolites (sample B2108). The primary clinopyroxene compositions in the rocks define a single evolutionary array, consistent with crystallization from a common parental melt (or compositionally similar melts) under comparable pressure conditions. In contrast, clinopyroxenes from other gabbro occurrences show only minor Al_2O_3 variations with little dependence on Mg# (**Fig. 11a**).

Clinopyroxenes in cumulate rocks derived from tholeiitic melts typically show pronounced Ti enrichment with decreasing Mg#, whereas cumulates crystallized from boninitic melts are characterized by only minor increases in Ti content during differentiation (Ceuleneer and

Le Sueur, 2008; DeBari and Coleman, 1989; Greene et al., 2006; Marchesi et al., 2009; Pirard et al., 2013). As shown in **Fig. 11b**, clinopyroxene from the ultramafic cumulates and associated gabbroids displays a slight increase in Ti content with decreasing Mg#, consistent with crystallization from boninitic melts. In contrast, clinopyroxene from the remaining gabbroids exhibits strong Ti enrichment with decreasing Mg#, suggesting crystallization from tholeiitic melts.

Overall, the low Ti content of clinopyroxene from the ultramafic-mafic rocks of the Tyulkulam Massif points to crystallization from Ti-undersaturated magmas, typical of suprasubduction-zone environments (e.g., Greene et al., 2006). Their low Al and Ti abundances are comparable to those observed in clinopyroxenes crystallized from island-arc tholeiitic and boninitic melts (**Fig. 12b, c**), providing further evidence for formation in a forearc setting (e.g., Beccaluva et al., 2004).

High anorthite content in plagioclase from the Tyulkulam gabbroids (An_{84}–An_{98}) places these compositions near the upper limit of the range typical of plagioclase in oceanic cumulate rocks (An_{28}–An_{95}; Elthon et al., 1992; Bloomer et al., 1989) and mafic to ultramafic island arc assemblages (Jagoutz et al., 2007), consistent with crystallization from relatively primitive arc-related melts. Based on the Mg# of clinopyroxenes (0.76–0.86) and the compositions of coexisting plagioclase (An_{84}–An_{98}), the studied gabbroids plot within the fields of suprasubduction zone ophiolites, particularly those corresponding to island arc and forearc settings (**Fig. 12a**). In contrast, plagioclase from gabbros of back-arc basins typically shows significantly lower anorthite contents (An_{48}–An_{59}; Ohara et al., 2002a).

In summary, the characteristic features of the compositions of primary minerals from the lherzolites, pyroxenites, and gabbroic rocks in the Tyulkulam Massif indicate their formation within a forearc to island arc environment in a suprasubduction zone setting.

CONDITIONS OF THE FORMATION OF ULTRAMAFIC–MAFIC ROCKS OF THE TYULKULAM MASSIF

Compositions of primary minerals in the ultramafic-mafic rocks of the Tyulkulam Massif provide a basis for evaluating the physicochemical parameters of crystallization, including temperature, pressure, and oxygen fugacity (fO_2) conditions. In all calculations, we used data only from the central parts of mineral grains.

Formation temperatures of mineral assemblages were evaluated using multiple geothermometric approaches. Olivine–spinel equilibrium temperatures in the harzburgites were calculated following the work of Ballhaus et al. (1991); they range from 700 to 750±30 °C, whereas the corresponding estimates for the lherzolites and pyroxenites are 710–720±30 °C. In contrast, two-pyroxene thermometry (Brey and Köhler, 1990) yields higher temperatures of 890–990±24 °C for the lherzolites and pyroxenites. Differences between temperature estimates derived from olivine-chromian spinel and orthopyroxene-clinopyroxene equilibria are typical of ultramafic rocks from ophiolite complexes and mantle xenolith suites (Ozawa, 1983; Bazylev and Silantyev 2000). These discrepancies apparently reflect the contrasting diffusion rates of Fe and Mg in the respective mineral pairs, with more rapid equilibration between coexisting pyroxenes and slower re-equilibration between olivine and spinel during cooling.

Application of the REE-in-two-pyroxene geothermometer (Liang et al., 2013) yielded higher temperature estimates of 1100–1200±30 °C for the lherzolites. Temperatures obtained using this method are typically 150–200°C higher than those derived from conventional two-pyroxene thermometry. This discrepancy is likely related to differences in diffusion rates between trivalent rare-earth elements (REE) and divalent cations such as Ca, Mg, and Fe in pyroxenes (Liang et al.,

542 2013). This inference is indirectly supported by the presence of exsolution textures in the
543 pyroxenes. Thus, the conventional two-pyroxene geothermometer likely records temperatures
544 corresponding to the final stages of Fe–Mg exchange equilibration, whereas the REE-in-two-
545 pyroxene and clinopyroxene geothermometers are considered to approach crystallization
546 temperatures of the rocks.

547 Thus, equilibrium temperature estimates for primary minerals from the massif rocks,
548 obtained using different geothermometers, vary over a wide range and likely reflect different
549 stages of mineral re-equilibration during post-crystallization cooling.

550 Amphibole compositions were used to constrain crystallization conditions, including
551 pressure, temperature, and oxygen fugacity, following the approach of Ridolfi (2021). Primary
552 amphiboles in the lherzolites and pyroxenites yield crystallization conditions of 870–900±22°C
553 °C and 2–3 kbar. Amphibole-plagioclase equilibrium temperatures in gabbros occurring as dykes
554 in harzburgite blocks and in gabbros hosting lherzolite-pyroxenite bodies, calculated using the
555 amphibole-plagioclase geothermometer of Holland and Blundy (1994), range from 720–780±40°C
556 and 820–850±40 °C, respectively. Similar temperature ranges (720–770±22 °C and 820–830±22
557 °C) were obtained using the single-amphibole thermometer of Ridolfi et al. (2021). The occurrence
558 of primary amphibole suggests crystallization from water-rich melts. Estimated melt H_2O contents
559 vary from 7.5 to 8.4 wt.% for the lherzolites and pyroxenites, and from 6.5 to 7.9 wt.% for the
560 gabbros.

561 Pressure estimates for the mineral assemblages in gabbroic rocks were derived from
562 amphibole–plagioclase barometry (Molina et al., 2015), falling within a 1.3 to 2.7 kbar (2±0.9
563 kbar) range; single-amphibole barometry (Ridolfi et al., 2021) yields, on the other hand, a
564 somewhat broader range of pressures between 1.2 kbar and 3.2 kbar (2.2±0.4 kbar). Elevated Mg#
565 values (57–68) and relatively high Al^{VI} contents (0.25–0.40 p.f.u.) in the gabbroic amphiboles are

566 characteristic of cumulate assemblages crystallized under low-pressure conditions (e.g., Cooper et
567 al., 2016).

568 Calculated oxygen fugacity values based on the oxybarometer of Ridolfi et al. (2021) yield
569 values of QFM +0.5 to QFM +1.9 for gabbros occurring as dykes within harzburgite blocks and
570 from QFM +0.6 to QFM +1.8 for gabbros associated with lherzolite–pyroxenite bodies. These
571 estimates are consistent with oxygen fugacity conditions characteristic of island arc magmatism
572 (QFM +0.5 to QFM +4; Evans et al., 2012).

573 Oxygen fugacity in the ultramafic rocks was evaluated using the olivine–spinel
574 oxybarometer of Ballhaus et al. (1991), with temperatures constrained by olivine–spinel
575 thermometry and pressures assumed at 1 GPa. Uncertainties in temperature and pressure estimates
576 exert only limited influence on calculated fO_2 values, with deviations of 100 °C and 1 GPa
577 producing changes of ~0.2 and ~0.3 log units, respectively (Wood et al., 1990). Calculated fO_2
578 values for the Tyulkulam harzburgites range from QFM +0.3 to QFM +1.8, whereas higher values
579 of QFM +2.2 to QFM +2.6 were obtained for the lherzolite–pyroxenite cumulates. These values
580 are higher than those typical of mid-ocean ridge peridotites (QFM −2 to QFM +1) (Ballhaus et al.,
581 1991; Pagé et al., 2008) and overlap with the range characteristic of forearc mantle peridotites
582 (QFM −1.1 to QFM +1.81) (Parkinson & Pearce, 1998) and island arc cumulate suites (QFM +0.4
583 to QFM +2.3) (Bucholz & Kelemen, 2019). These findings indicate their formation under
584 relatively oxidizing conditions in a suprasubduction zone setting.

585

586 **CALCULATED EQUILIBRIUM MELT COMPOSITIONS**

587

588 Clinopyroxene compositions from the lherzolites were used to estimate concentrations of
589 trace and rare earth elements in melts in equilibrium with clinopyroxene. Partition coefficients for

590 trivalent REE between clinopyroxene and melt were calculated using the parameterized lattice
591 strain model of Sun and Liang (2012). The calculated partition coefficients are comparable to those
592 reported for melt inclusions in clinopyroxene from boninites of the Troodos Massif (Sobolev et
593 al., 1996).

594 Trace-element patterns of melts reconstructed from clinopyroxene compositions (**Fig. 13a**)
595 define a restricted field of subparallel spectra that closely resemble boninitic melt signatures in
596 both incompatible-element abundances and overall spectral shapes (Kelemen et al., 2003). These
597 features indicate that melts equilibrated with clinopyroxene were most likely originated in a
598 suprasubduction zone setting.

599 The whole-rock chemical composition of cumulate rocks is largely controlled by their
600 modal mineralogy and does not necessarily reflect the composition of the parental melt from which
601 they crystallized. Therefore, the use of bulk composition of cumulates for the reconstruction of
602 their tectonic setting is limited. Nevertheless, the composition of trapped equilibrium melts can be
603 reconstructed using a mass-balance approach involving elemental partition coefficients (Bédard,
604 1994, 2001), based on primary mineral compositions and modal mineral proportions
605 (**Supplementary 3, Table S10** and **Table S11**). A key uncertainty in this approach is the poorly
606 constrained proportion of trapped melt, which may lead to overestimation of trace-element
607 contents in modeled melt compositions. However, the generally low concentrations of the HFSE
608 in the studied cumulate rocks, such as Zr (1.75–49.42 ppm), Y (0.94–38.23 ppm), and Nb (0.02–
609 2.17 ppm) (**Table S1**), indicate that these rocks are dominated by cumulus minerals and that they
610 contain only limited amounts of trapped melt (Parlak et al., 1996).

611 The proportion of trapped melt in the pyroxenites is tentatively estimated at ~1–2 wt.%
612 (e.g., Ledneva et al., 2017), which is generally sufficient to obtain reliable modeled melt
613 compositions. For the lherzolites, trapped melt fractions of ~1–1.5 wt.% were calculated using

614 trace-element data from clinopyroxene and mass-balance formulations of Bédard (1994, 2001). In
615 cumulate gabbros, melt fractions may vary more widely, although a minimum value of ~2 wt.%
616 has been suggested (e.g., Charlier et al., 2005). Reconstruction of equilibrium melt compositions
617 for the harzburgites was not possible due to their extensive serpentinization, incomplete data on
618 primary mineral compositions, and extremely low REE concentrations, commonly below
619 analytical detection limits.

620 Reconstructed trapped melt compositions for the lherzolites, olivine websterites, and
621 gabbro (sample B2108) of the Tyulkulam Massif are consistent with boninitic melt signatures (**Fig.**
622 **13b; Table S11**). The trace element patterns exhibit features typical of suprasubduction zone
623 magmatic assemblages, including negative Nb and positive Sr, U, and Pb anomalies. Boninites are
624 typical volcanic rocks widely distributed in modern island arc systems (Dilek and Furnes, 2014).
625 However, the geodynamic setting of their formation remains debated. They are commonly
626 interpreted to mark the initial episodes of the development of suprasubduction zone systems during
627 the pre-arc stage of their evolution (Ishizuka et al., 2006; Reagan et al., 2010; Dilek and Furnes,
628 2011, 2014; Pearce and Reagan, 2019).

629 At the same time, trace- and rare–earth (REE) element patterns of trapped melts in
630 gabbroids occurring within the harzburgite block, closely resemble those of island-arc basalts (**Fig.**
631 **13b; Table S11**), indicating that the formation of the lherzolites, pyroxenites, and gabbroids in the
632 Tyulkulam Massif was most likely related to the earlier stages of island arc development.

633

634 **TECTONIC AFFINITY OF THE ULTRAMAFIC–MAFIC ROCKS OF THE**
635 **TYULKULAM MASSIF**

636

637 Harzburgites of the Tyulkulam Massif display geochemical characteristics comparable to
638 residual mantle peridotites formed as a result of high degrees of partial melting (17–22%) in a
639 suprasubduction-zone setting (Hellebrand et al., 2001; Parkinson and Pearce, 1998). However,
640 compared to the ultra-depleted harzburgites of the Itmurundy Massif in the western part of the
641 North Balkhash zone, ultramafic rocks of the Tyulkulam Massif display relatively more enriched
642 geochemical signatures, as reflected in both their whole-rock compositions and mineral chemistry.
643 Harzburgites of the Tyulkulam Massif display low HREE concentrations and U-shaped
644 REE patterns marked by enrichment in LILE and LREE, along with positive Zr–Hf and Sr
645 anomalies (**Fig. 7**). These geochemical signatures are characteristic of suprasubduction-zone
646 harzburgites and are comparable to those of forearc peridotites (Parkinson and Pearce, 1998), as
647 well as to the most depleted harzburgites from mid-ocean ridge tectonic settings (Godard et al.,
648 2000).
649 Lherzolites, pyroxenites, and gabbroids of the Tyulkulam Massif also appear to have
650 formed in suprasubduction-zone settings, as indicated by their enrichment in LILE at low
651 abundances of high field strength elements HFSE, the presence of a negative Nb anomaly, and
652 elevated concentrations of U and Ta (**Fig. 7**). Such geochemical characteristics suggest addition
653 and migration of slab-derived components within the zone of interaction between the subducting
654 slab and the mantle wedge in the subduction channel. The chemical compositions of the gabbroids
655 are likewise consistent with their formation in an island arc environment.
656 Based on similarities in structural position and the close spatial association of rock
657 complexes within the serpentine mélange, including tectonic slices of basalts and cherts, the
658 Tyulkulam Massif is considered a part of the North Balkhash ophiolite zone. Geochemical
659 characteristics of dunites and harzburgites from the Itmurundy Massif indicate that they represent
660 mantle residues formed after high degrees of partial melting (19–23%) in the presence of aqueous

661 fluids in forearc settings (Milyukova et al., 2023). During intraoceanic subduction, the most
662 depleted mantle rocks may be tectonically exhumed and incorporated into accretionary prism
663 complexes (Stern, 2010). It is further inferred that rocks of the layered complex, which include
664 peridotites, pyroxenites, and gabbroids with subordinate plagiogranites forming the Arkharsu,
665 Tesiktas, and Eastern Arkharsu Massifs, were also generated in suprasubduction zone
666 environments with participation of slab-derived fluid components released in the mantle wedge–
667 slab interaction zone (Degtyarev et al., 2021). In addition, gabbroids of the Tesiktas and East
668 Arkharsu massifs also exhibit geochemical signatures consistent with suprasubduction zone
669 settings (Degtyarev et al., 2021). At the same time, the geochemical features of the ultramafic–
670 mafic rocks from the Tyulkulam Massif outlined above indicate their genetic relationship with
671 magmatic processes operating during forearc spreading and subsequent island arc magmatism,
672 reminiscent of the tectonomagmatic evolution of the Cretaceous Troodos ophiolite within the
673 Tethyan realm (Dilek et al., 1990).

674 Thus, complexes of the Itmurundy, Arkharsu, Tesiktas, East Arkharsu, and Tyulkulam
675 Massifs are interpreted as fragments representing different structural levels of ophiolite sequences,
676 which were predominantly formed in a forearc tectonic setting during the incipient stages of the
677 evolution of an intraoceanic subduction zone system within the Junggar–Balkhash ocean basin. At
678 the same time, the ultramafic and mafic rock assemblages of the Tyulkulam Massif are
679 characterized by comparatively less depleted geochemical signatures and by the presence of
680 cumulate lherzolites among the ultramafic rocks, in contrast to the massifs of the western North
681 Balkhash ophiolite belt. These features indicate their formation during the progressive transition
682 from forearc to island-arc tectonic settings at the early stages of the evolution of the Junggar–
683 Balkhash Ocean.

684

CONCLUSIONS

1. The Tyulkulam Massif is a large tectonic sliver within the Silurian–Middle Devonian flysch sequence. It consists of a serpentine–matrix mélange containing blocks of residual and cumulate ultramafic rocks and gabbroids. The mélange is intruded by younger porphyritic granodiorite and plagiogranite bodies, as well as by numerous felsic to intermediate dykes and veins dated at ca. 300 Ma, which are not part of the mélange assemblage. This new age constraint provides a minimum age for the mélange formation.

2. Integrated petrological and geochemical data and related interpretations indicate that the studied mantle peridotites, cumulate ultramafic rocks, and gabbroids in the Tyulkulam Massif likely formed in a suprasubduction zone tectonic setting. The data in support of this inference include:

- **a.** Harzburgites reflect high degrees of fluid-assisted partial melting (17–22%) typical of forearc mantle environments;
- **b.** Lherzolites, pyroxenites, and gabbroic rocks correspond to cumulates derived from boninitic to tholeiitic melts that crystallized during the transition from forearc to island arc magmatism at moderate pressures of ~2–3 kbar.

3. Interpretation of the integrated geological, mineralogical, petrological, and geochemical data indicates that the plutonic rocks of the ophiolite sequence in the North Balkhash zone may have formed during different stages of the early development of an intraoceanic subduction zone system within the Junggar–Balkhash Ocean basin.

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FIGURE CAPTIONS

Figure 1. Distribution of ophiolites and spatially and temporarily associated geological complexes in the Junggar-Balkhash region (Degtyarev et al., 2021). Ophiolite zones and mélanges: TK – Tekturmas; AG – Agadyr; NB – North Balkhash; TS – Tastau; TB – Tangbale; ML – Maylie; BELK – Barleik; HT – Hatu; DB – Daerbute; BJT – Baijantan. Volcano-plutonic belt: D – Kazakhstan Devonian; BY – Balkhash-Yili (late Palaeozoic); IZ – Irtysh-Zaissan zone. Main strike-slip faults (in red): CHF – Chingiz; CKF – Central Kazakhstan; JF – Junggar.

Figure 2. Simplified geological map of the Tyulkulam Massif (after Semenov (1960b)).

Figure 3. Photomicrographs of harzburgite samples from the Tyulkulam Massif. Abbreviations: Ol – olivine, Spl – chromian spinel, Tlc – talc, Mag – magnetite, Srp – serpentine. **(a, b)** Harzburgite with a characteristic porphyroclastic texture (BSE images, sample BG2101); **(c, d)** Relict spinels in a serpentinized matrix (BSE images).

Figure 4. Back-scattered electron (BSE) images of a lherzolite (sample B2105). Abbreviations: Ol–olivine, Opx–orthopyroxene, Cpx–clinopyroxene, Spl–chromian spinel, Amp–amphibole. **(a)** Relict grains of orthopyroxene, olivine, and amphibole; **(b)** Amphibole and clinopyroxene inclusions in orthopyroxene; **(c)** Primary grains of olivine and orthopyroxene; **(d)** Olivine and clinopyroxene grains.

Figure 5. Photomicrographs of olivine websterite (sample B2107). Abbreviations: Opx – orthopyroxene, Cpx – clinopyroxene. **(a)** Common panidiomorphic texture and massive structure (XPL); **(b)** Clinopyroxene lamellae in orthopyroxene (BSE image).

Figure 6. Photomicrographs of gabbroic rock textures. **(a)** Gabbro hosting lherzolite bodies (XPL, sample B2108); **(b)** Gabbro-diorite (XPL, sample B2105/1); **(c, d)** Gabbro spatially associated with harzburgite blocks (samples B2111, B2114).

Figure 7. (a) Chondrite (C1)-normalized REE patterns, and **(b)** Primitive mantle (PM)-normalized trace element patterns of the harzburgitic rocks of the Tyulkulam Massif (normalization to C1 and PM are after Sun and McDonough (1989)). Compositions of ultramafic rocks from the Itmurundy Massif are from Milyukova et al. (2023); Fields of abyssal peridotites are after Dick et al. (1984), Dick (1989), Johnson et al. (1990), and Johnson and Dick (1992); suprasubduction peridotite fields are after Parkinson and Pearce (1998). **(c)** C1-normalized REE patterns, and **(d)** PM-normalized trace element patterns of lherzolites, pyroxenites, and gabbroic rocks of the Tyulkulam Massif (normalization to C1 and PM are after Sun and McDonough (1989)). The field of island-arc cumulates is after Greene et al. (2006).

Figure 8. Primitive mantle (PM)–normalized incompatible trace element patterns of clinopyroxene from lherzolites of the Tyulkulam Massif (normalization values are after McDonough and Sun (1989)). The compositional field of clinopyroxene from pyroxenites and websterites of the Voykar–Synya complex is after Batanova et al. (2011) and Belousov et al. (2021).

Figure 9. NiO (wt.%) vs Mg# in olivine from ultramafic-mafic rocks of the Tyulkulam Massif. The field of mantle olivine compositions is after Takahashi et al. (1987).

1023 **Figure 10.** Diagrams for ultramafic-mafic rocks of the Tyulkulam Massif: **(a)** Compositions of
1024 coexisting olivine and spinel (Arai, 1994a). Fields of abyssal peridotites are after Dick and Bullen
1025 (1984); suprasubduction peridotites and passive margin peridotites are after Pearce et al. (2000);
1026 **(b)** Cr# versus Mg# relationships in spinels from the studied rocks. Fields of forearc-basin dunites
1027 and spinel peridotites are after Ishii et al. (1992); back-arc basin spinel peridotites are after Ohara
1028 et al. (2002, 2003); mid-ocean ridge (MOR) spinel peridotites are after Dick and Bullen (1984).
1029 Spinel compositions from ultramafic rocks of the Itmurundy Massif are from Milyukova et al.
1030 (2023); **(c)** TiO_2 (wt.%) vs F#Spl relationships in spinels. Spinel compositions from basaltic rocks
1031 of mid-ocean ridge (MORB), intraplate (intraplate), and island-arc (arc) settings are after Arai
1032 (1992).
1033
1034 **Figure 11. (a)** Al_2O_3 contents (wt.%) and **(b)** TiO_2 contents (wt.%) in primary clinopyroxenes
1035 from ultramafic-mafic rocks of the Tyulkulam Massif. The arc trend is defined for the typical arc
1036 cumulates of the Chilas complex, Kohistan (Jagoutz et al., 2007).
1037
1038 **Figure 12.** Compositional diagrams for clinopyroxene and plagioclase from gabbroic rocks of the
1039 Tyulkulam Massif: **(a)** Ti (a.p.f.u.) vs Al(IV) relationships in clinopyroxenes from the cumulative
1040 rocks of the Tyulkulam Massif. Clinopyroxene compositions from different tectonic settings are
1041 after Beccaluva et al. (1989) and Li et al. (2013): normal mid-ocean ridge basalts (N-MORB),
1042 island-arc tholeiites (IAT), and boninites (BON); **(b)** TiO_2-Na_2O-SiO_2/100 (wt.%) discrimination
1043 diagram. Clinopyroxene compositions from ultramafic-mafic different tectonic settings are after
1044 Beccaluva et al. (1989): normal mid-ocean ridge basalts (N-MORB), ocean island basalts (OIB),
1045 island-arc tholeiites (IAT), boninites (BON), and forearc volcanic rocks (BA-A); **(c)** Compositions
1046 of coexisting primary plagioclase and clinopyroxene from gabbroic rocks of the Tyulkulam

Massif. Fields: Backarc gabbros are after Akizawa et al. (2021); Ohara et al. (2002b, 2003); Mid-Ocean Ridge gabbros are after Bloomer et al. (1995), Dick and Natland (1996); Mariana Forearc and Arc gabbros are after Bloomer et al. (1995).

Figure 13. Primitive mantle (PM)–normalized multi-element diagrams for ultramafic-mafic rocks of the Tyulkulam Massif: **(a)** Melts equilibrated with clinopyroxenes from lherzolites; **(b)** Melts entrapped by lherzolites, pyroxenites, and gabbroic rocks. Normalization to the primitive mantle is after Sun and McDonough (1989). Multi-element patterns of boninites and oceanic island-arc basalts are after Kelemen et al. (2003).

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