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Late Cretaceous granitoids of the Sikhote–Alin orogenic belt, southeastern Russia: implications for the Mesozoic geodynamic history of the eastern Asian continental margin

Igor V. Kemkin¹, Andrei V. Grebennikov¹, Xing-Hua Ma²,³,*, Ke-Ke Sun⁴

¹Far East Geological Institute, Far Eastern Branch, Russian Academy of Sciences, pr. 100-letiya Vladivostok 159, Vladivostok, 690022 Russia

²MNR Key Laboratory of Metallogeny and Mineral Assessment, Institute of Mineral Resources, Chinese Academy of Geological Sciences, Beijing 100037, China

³State Key Laboratory of Geological Processes and Mineral Resources, China University of Geosciences, Beijing 100083, China

⁴Department of Earth and Space Sciences, Southern University of Science and Technology, Shenzhen 518055, China

* Corresponding author: Xing-Hua Ma

Chinese Academy of Geological Sciences, Beijing 100037, China

Tel/Fax: +86-010-68999866

E-mail: maxh@pku.edu.cn

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Abstract

We present new U–Pb age data for granitoids in the Central Sikhote–Alin orogenic belt in SE Russia, which refute the established opinion about the absence of the Late Cretaceous magmatism at the eastern margin of the Paleo-Asian continent. It was previously thought that a period of magmatic quiescence occurred from 88 to 50 Ma, related to subduction of the Paleo-Pacific Plate under the eastern margin of the Paleo-Asian continent, although this is inconsistent with evidence from the Sikhote–Alin, Sakhalin, and Japan regions. Three suites of plutonic rocks with different ages were identified in this study. The first suite has ages of 105–92 Ma and formed in a syn-orogenic setting. The second (86–83 Ma) and third (ca. 73 Ma) suites formed during the post-orogenic stage of the Sikhote–Alin orogenic belt. The second and third suites were coeval with Late Cretaceous granitoids that formed in a suprasubduction continental arc known as the Eastern Sikhote–Alin volcanic–plutonic belt (ESAVPB). However, the studied rocks are located far inland from the ESAVPB. The ages of the studied granitoids coincide with the timing of a change in the angle of convergence between the Paleo-Pacific Plate and eastern margin of the Paleo-Asian continent. This change in motion of the oceanic plate with respect to the continental plate was probably caused by a rupture in the subducted slab (i.e., a slab tear), followed by asthenospheric upwelling and partial melting of the overlying crust, which ultimately generated post-orogenic intrusive magmatism.

Keywords: Zircon U–Pb age, Sikhote–Alin orogenic belt, eastern Asian continental margin, slab tear, post-orogenic magmatism
Introduction

The Sikhote–Alin orogenic belt is located in the extreme southeast of Russia and extends from the northern coast of the Sea of Japan to the southern coast of the Sea of Okhotsk (Fig. 1). This is a key region for investigating the Mesozoic–Cenozoic geodynamic evolution of the eastern margin of the Asian continent. This region contains lithologically variable formations of different ages that are fragments of Jurassic, Late Jurassic–Early Cretaceous, and Early Cretaceous accretionary prisms, an Early Cretaceous syn-strike-slip continental margin basin, an Early Cretaceous island arc, and a Late Cretaceous marginal-continental (magmatic) arc (e.g., Kemkin, 2006; Khanchuk, 2006; Kemkin et al., 2016; Khanchuk et al., 2016). Each of these structural–lithological units is an indicator of a specific geodynamic setting.

The identification and comparison of the various units comprising the orogenic belt and age and geochemical data for igneous rocks intruded them, can constrain the geodynamic setting during the different stages of the formation of the Sikhote–Alin orogenic belt. Such information from the Sikhote–Alin region, eastern China, Korea, and Japan might improve geodynamic reconstructions of the eastern margin of Asia in the Mesozoic and Cenozoic. It has already been established that the Sikhote–Alin orogenic belt formed under a number of geodynamic settings, including an active subduction margin during the Jurassic and an active transform margin during the Early Cretaceous (e.g., Kemkin et al., 2016; Khanchuk et al., 2016, 2019).

The formation of the Sikhote–Alin orogenic belt was accompanied by two stages of syn-orogenic magmatism. These magmatic events are represented by two Early Cretaceous granitoid suites, known as the Khungari and Tatiba series (e.g., Simanenko et al., 1997; Kruk et al., 2014; Grebennikov et al., 2016). Intrusions of the early suite (i.e., the Hauterivian–Barremian Khungari series) are located mainly in Jurassic accretionary prism terranes, and form relatively large massifs (i.e., tens to hundreds of square kilometers in surface area). The ages of the Khungari granitoids, determined by Rb–Sr and K–Ar methods, vary widely between 234 and 107 Ma (Levashev et al.,
1983; Natal’ in et al., 1993; Gvozdev, 2010). Limited zircon U–Pb dating has shown that the age of these rocks varies between 131 and 123 Ma (Kruk et al., 2014; Jahn et al., 2015; Wu et al., 2017).

Unlike the Khungari granitoid suite, granitoid magmatism in the second suite, the Albian–early Cenomanian series (i.e., the Tatiba series), was of much larger scale (e.g., Izokh et al., 1967). These granitoids intruded almost all terranes of the Sikhote–Alin orogenic belt, including a Jurassic accretionary prism and the Early Cretaceous Zhuravlevka–Amur syn-tectonic basin and Kema island arc. The Tatiba granitoids have zircon U–Pb ages of 110–95 Ma (Khanchuk et al., 2008; Jahn et al., 2015; Sakhno et al., 2016; Khanchuk et al., 2019; Kruk et al., 2019).

Late Cretaceous igneous rocks are widespread along the eastern margin of the Eurasian continent, and occur as linear belts of elongate volcanic–plutonic complexes. In the Sikhote–Alin region they are termed the East Sikhote–Alin volcanic–plutonic belt (ESAVPB; Shatsky, 1957; Bykovskaya, 1962; Khanchuk, 2006), and in northeastern Russia are known as the Okhotsk–Chukotka volcanic–plutonic belt (OCVPB; Beliy & Tilman, 1966; Akinin et al., 2020; Tikhomirov et al., 2021, and references therein). These volcanic–plutonic complexes are thought to have formed in a suprasubduction continental arc as a result of subduction of the Paleo-Pacific (i.e., Izanagi) Plate beneath the Paleo-Asian continent (e.g., Khanchuk, 2006; Akinin & Miller, 2011; Valuy, 2014; Jahn et al., 2015; Grebennikov et al., 2016).

Cretaceous granitoids are widespread in the San’yo and San’in zones of Japan, and intruded pre-Cretaceous accretionary complexes that contain regionally metamorphosed rocks (e.g., Nakajima et al., 2016). Similar to the Sikhote–Alin region, the granitoids in southwestern Japan are associated with coeval rhyolites and ignimbrites, although most of the cover rocks have been eroded (Jahn et al., 2015). Whole-rock Sr–Nd isotopic data for the ESAVPB granitoids are similar to those of other igneous rocks of this age in the southwestern Japan (Jahn et al., 2015).
Cretaceous plutonic rocks (i.e., the Bulkuksa granitoids; e.g., Jin et al., 2001) crop out as crop out as scattered minor stocks that are 10–200 km$^2$ in size in the central–southern Korean Peninsula. These plutonic rocks have ages of 100–65 Ma and were associated with subduction of the Izanagi Plate (Kim et al., 2016). However, in adjacent eastern China, as well as western areas of the Sikhote–Alin orogenic belt, Late Cretaceous igneous rocks have not yet been identified. This has led to suggestions that a period of magmatic gap occurred along the eastern margin of the Paleo-Asian continent from 88 to 50 Ma (e.g., Wu et al., 2011; Niu et al., 2015, and references therein). The magmatic quiescence has been explained by the cessation of subduction and Late Cretaceous strike-slip movement of the Paleo-Pacific Plate with respect to the East Asian continent (e.g., Niu et al., 2015; Tang et al., 2016).

Early Paleocene–middle Eocene silicic igneous rocks are widely distributed across the entire Sikhote–Alin region, and indicate the absence of a magmatic gap, at least till ca. 53 Ma (Grebennikov et al., 2020, 2021, and references therein). Moreover, Paleocene–early Eocene igneous rocks that are petrologically and geochemically similar to coeval rocks in the Sikhote–Alin region occur in adjacent areas, including northeastern China (e.g., Wang et al., 2016; Li et al., 2018), the southern Korean Peninsula (e.g., Hwang et al., 2012; Cheong et al., 2013), and the Inner Zone (i.e., the San’in zone) of Japan (e.g., Imaoka et al., 2011; Wakita, 2013; Iida et al., 2015).

Given that the Sikhote–Alin region is geographically located between mainland China, the Korean Peninsula, and Japan, the temporal–spatial distribution of Cretaceous plutonic rocks in this region is key in constraining the Cretaceous tectono-magmatic evolution of East Asia. In this paper, we present new zircon U–Pb age data for several granite massifs located in the inland, central part of the Sikhote–Alin orogenic belt. The data are inconsistent with the proposal of a period of magmatic gap during the Late Cretaceous at 89–74 Ma.

Regional geological background
The studied granitoid massifs are located mainly in the central Zhuravlevka–Amur syn-strike-slip turbidite basin of the Sikhote–Alin–North Sakhalin accretionary orogenic belt (hereafter referred to as the Sikhote–Alin orogenic belt) (Fig. 2). The latter is a tectonic collage of fragments of structural–lithological units (i.e., terranes) of varying ages that initially formed distal from each other in different structural settings. These were amalgamated at the eastern margin of the Paleo-Asian continent during the convergent and transform tectonic interactions between oceanic and continental plates. The Sikhote–Alin orogenic belt is a typical example of a fold belt, and it formed in two geodynamic settings: an active subduction margin during the Jurassic and Late Cretaceous, and an active transform margin during the Early Cretaceous and Paleocene (Khanchuk, 2006; Grebennikov et al., 2016, 2020; Kemkin et al., 2016; Khanchuk et al., 2016).

During Jurassic subduction, thick accretionary complexes formed along the eastern edge of the Paleo-Asian continent. These complexes are intensively faulted and folded on various scales, and occur in the Samarka, Nadanhada–Bikin, Khabarovsk, and Badzhal terranes (e.g., Kemkin, 2008). In the Early Cretaceous, during transform motion of the Paleo-Pacific Plate (also known as the Izanagi Plate), thick strata accumulated in a syn-strike-slip turbidite marginal basin along the eastern margin of the Paleo-Asian continent, which is represented by the Zhuravlevka–Amur terrane. At the southeastern margin of the continent, orthogonal plate motion across the continent–ocean boundary led to the generation of a new subduction zone (e.g., Kemkin et al., 2016) and the formation of latest Tithonian–Hauterivian accretionary complexes. In the Sikhote–Alin orogenic belt, these complexes occur in the Taukha terrane (Fig. 1).

The strike-slip tectonism between the Paleo-Asian and Paleo-Pacific plates, combined with blocking of the southern subduction zone during the Hauterivian (Kemkin, 2006; Kemkin et al., 2016), resulted in large-scale (tens to several hundreds of kilometers), left-lateral, strike-slip movement of continental margin blocks and accretionary complexes along the Tan–Lu fault system. This tectonism caused further folding of the terranes in the Jurassic accretionary prism, which led to
significant crustal thickening (Khanchuk et al., 2013). The increase in lithostatic pressure and
temperature in the lower parts of the folded crustal rocks caused melting of sedimentary rocks and
generation of early syn-orogenic granitoids (i.e., the Khungari series; Khanchuk et al., 2013). At the
end-Early Cretaceous (early to middle Albian), strike-slip movement between the Paleo-Asian and
Paleo-Pacific plates resulted in an Early Cretaceous island arc system being accreted to the eastern
margin of the former plate. Fragments of this island arc in the Sikhote–Alin orogenic belt are
represented by the Kema (rear arc) and Kiselevka–Manoma (accretionary prism) terranes (Kemkin,
2006; Khanchuk, 2006). The collision of the island arc system with the Paleo-Asian continent was
accompanied by reactivated large-scale, left-lateral, strike-slip movement along the Tan–Lu fault
system (Khanchuk et al., 2016). This caused fragmentation of the island arc complexes and
subsequent translation of their blocks along the continental margin, and led to folding of the rocks
of the Jurassic–Cretaceous accretionary prisms and Early Cretaceous turbidite basin into compressed
folds with a range of amplitudes and NE–SW-trending fold axes. This renewed thickening of the
continental margin caused further crustal melting and led to the generation of syn-orogenic
granitoids of the second stage (i.e., the Tatiba series; Khanchuk et al., 2013).

Previous studies have identified isolated Late Cretaceous (85–73 Ma) intrusive bodies in the
inland area of the Sikhote–Alin orogenic belt (Gonevchuk et al., 2015; Jahn et al., 2015; Sakhno et al.,
2016; Tsutsumi et al., 2016; Kruk et al., 2019). However, most of the ages for these Late Cretaceous
granites were obtained by Rb–Sr and K–Ar methods, and may not be reliable. For example, the K–Ar
isotopic system is susceptible to argon loss, and Rb–Sr ages can be disturbed by the mobile nature of
Rb and Sr in hydrothermal fluids. As such, these ages might date younger thermal events that
occurred after crystallization. Accurate geochronological data are key for constraining geological and
geodynamic models. Therefore, in this study we conducted zircon U–Pb dating of granite massifs in
the central Sikhote–Alin orogenic belt in the Bikin–Bol Ussurka region.
The granitoid intrusions dated in this study are mainly exposed in the Berriasian–Valanginian part of the Zhuravlevka–Amur terrane. Lithological units of the eastern Zhuravlevka–Amur terrane, as well as other Early Cretaceous terranes (Kema, Taukha, and Kiselevka–Manoma) are unconformably overlain by Cenomanian–Maastrichtian volcanic rocks of the ESAVPB, which are thought to be the product of a suprasubduction, continental margin (i.e., epicontinental) magmatic arc (e.g., Khanchuk, 2006). Their geochemistry is consistent with the fluid-driven partial melting of metasedimentary and metaigneous rocks under relatively oxidizing conditions, which is typical of suprasubduction settings. The formation of the ESAVPB reflects an Early–Late Cretaceous change in the geodynamic setting along the eastern margin of the Paleo-Asian continent from a transform setting to a subduction margin setting.

Features of the studied granitoid massifs

A summary of the rock types, sample locations, pluton names, petrographic features, and mineral assemblages for the studied granitoid samples is presented in Table 1.

The Pravovalenkuyskiy massif (Fig. 2) is located in the basins upstream of the Pravaya Valinku and Srednyaya Valinku rivers, and is an intrusive body that is oval-like in shape at the surface and crops out over an area of 41 km². The massif occurs within early-middle Albian flysch deposits of the Zhuravlevka–Amur terrane, which show hydrothermal alteration near the contacts with the massif. The massif is zoned, with marginal areas consisting of biotite monzogranite, representing the early phase of the intrusion. The central part consists of quartz monzonite of the second phase.

The monzogranite (sample PrVI-1) consists of plagioclase (An_{30-50}), perthitic alkali feldspar, quartz, and biotite, along with accessory apatite, zircon, monazite, allanite, epidote, and magnetite. The quartz monzonite (sample PrVI-2) consists of plagioclase and opaque minerals, clinopyroxene, orthopyroxene, biotite, amphibole perthitic alkali feldspar, quartz, and accessory magnetite, xenotime, and apatite.
The Lovlyagin Creek massif is located 5 km south of the Vodorazdel’nyy massif (Fig. 2), in the catchment between the upstream part of the Lovlyagin Creek and Vesnyanka Creek, both of which are tributaries on the true left side of the Bol’shaya Ussurka River. The massif covers an area of ~5 km² and intrudes Valanginian terrigenous rocks of the Zhuravlevka–Amur terrane, which have been thermally metamorphosed to hornfels near the intrusion. The intrusion consists mostly of fine- and medium-grained granodiorites that locally grade into diorite. This rock (sample Lvl-1) consists of plagioclase (An₃₀–₅₀), microcline, quartz, amphibole, biotite, and accessory magnetite, titanite, zircon, allanite, and apatite.

The Snezhnyy massif is situated in the catchment of the Bikin and Valinku rivers, ~25 km southwest of the Pravovalenkuyskiy massif (Fig. 2). The pluton crops out over an area of ~20 km², is irregular in shape and slightly elongated in a NE–SE direction. The contact between the pluton and host rocks has multiple embayments. The country rocks are Valanginian terrigenous deposits of the Zhuravlevka–Amur terrane, which are intensively hornfelsed.

The intrusion consists mainly of fine- (at the margins of the massif) and medium-grained (most of the massif) biotite and biotite–amphibole granodiorites (sample Sn-1), which grade in some places to granite or quartz diorite due to variable contents of plagioclase and quartz. The mafic minerals increase in abundance in the southeastern part of the massif. Rocks of the massif consist of plagioclase, perthitic alkali feldspar, quartz, biotite, and amphibole, along with accessory apatite, zircon, allanite, epidote, titanite, and magnetite. In some areas of the intrusion (both in its center and at its margins) granodiorites are cut by small dikes of aplitic microgranite, which appear to be the product of a second (i.e., final) stage of magmatism.

The Pereval’nyy massif (Fig. 2) is wedge-shaped, and its thinner end is located in the northeast. The massif crops out over an area of ~150 km², is located adjacent to the Central Sikhote–Alin Fault in the west, and intrudes cherty terrigenous sediments of the Samarka terrane, but its contacts with the host rocks are mostly tectonic in nature. The sedimentary rocks of the Samarka
terrane were overthrust onto the granitoids. An intrusive contact is only observed around the southern and southwestern parts of the intrusion, where it is a relatively wide (0.5–2.0 km) zone of hornfels and hydrothermally altered rocks (albitization, greisenization, and skarnification).

The Pereval’nyy massif consists of light gray, coarse- and medium-grained biotite leucogranites (samples Prv-1 and -2) that grade into fine-grained variants with a porphyritic texture near the massif margins (Fig. 3). The granitoids consist of quartz, alkali feldspar (including orthoclase, microcline, and albite), biotite, and rare amphibole, with accessory zircon, apatite, monazite, and allanite. The leucogranites are cut by small stocks, dikes, and veins of biotite and biotite–muscovite granites, granite porphyries, aplites, and pegmatites.

The Vodorazdel’nyy massif (Fig. 2) is located between the Perevalnaya and Bol’shaya Ussuka rivers near the Central Sikhote–Alin Fault. Outcrops of the massif form a slightly elongated body that trends NE–SW over an area of ~65 km². The massif intrudes late Tithonian–Berriasian siliceous mudstone and Berriasian–Valanginian sandstone–siltstone of the Zhuravlevka–Amur terrane. The boundary between the intrusion and host rock is marked by a thick zone of hornfels and biotitization of the terrigenous rocks.

The massif comprises two phases. Early melts that formed the main volume of the intrusive body produced coarse- to fine-grained, occasionally porphyritic, biotite leucogranites. During the second phase, dikes of biotite–muscovite leucogranites, aplitic granites, and pegmatites were intruded. The alkali feldspar granite of the first phase (sample Vd-1) consists of quartz, orthoclase, microcline, albite, and biotite, with accessory garnet, zircon, xenotime, tourmaline, and monazite. Monzogranite of the second phase (samples Vd-2 and -3) has a similar mineralogy, but also contain muscovite (Fig. 3). During later deformation the granites in the massif were partially fractured along fault zones, and some blocks were displaced with respect to one another by strike-slip offsets of up to 2–3 km.
The Ust’Malinovskiy massif is located 6 km southeast of the Snezhnyy massif (Fig. 2). The intrusive body only crops out over an area of ~1 km$^2$, and intrudes late Tithonian–Berriasian siliceous mudstones and Berriasian–Valanginian terrigenous rocks (sandstone and siltstone) of the Zhuravlevka–Amur terrane. At the contact, sedimentary rocks are hornfelsed and hydrothermally altered.

This massif is zoned, with a central part that consists of fine- and medium-grained biotite–amphibole granite (hornblende contents of up to 25 vol.%), and marginal facies of alkali feldspar granite that grades into aplitic microgranite and rhyolite (Fig. 3). The alkali feldspar granite (sample Mal-1) consists of quartz, alkali feldspar, plagioclase, and biotite, along with accessory ilmenite, zircon, and allanite.

The Dal’nearminskiy massif (Fig. 2) is located on the right bank of the Bol’shaya Ussurka River, ~20 km north of the Vodorazdel’nyy massif, and adjacent to the Central Sikhote–Alin Fault. The massif has an area of ~450 km$^2$, and it is elongate in a NE–SW direction. The massif intruded Berriasian–Valanginian sandstone–siltstone of the Zhuravlevka–Amur terrane, and there is a thick (up to 2.5 km) contact metamorphic aureole. The western boundary of the massif is the Central Sikhote–Alin Fault.

Two phases also occur in this massif. The first (main) phase consists of fine- to coarse-grained and occasionally porphyritic biotite leucogranite. The central part of the massif comprises coarse-grained rocks, which gradually become finer-grained approaching the margins.

The leucogranites (samples DArm-1 and -2) consist of alkali feldspar (including orthoclase, microcline, and albite), quartz, biotite, and rare amphibole, with accessory zircon, apatite, monazite, allanite, and rare fluorite. The second phase comprises aplitic dikes, leucogranites, and a granite porphyry with a mineralogy similar to that of the first phase rocks.

**Analytical methods**
Heavy liquid and magnetic methods were used to separate zircon grains from crushed unaltered rock samples. Zircons were further handpicked under a binocular microscope. The zircons were mounted in epoxy resin, polished, and covered with a 50-nm-thick coating of high-purity Au. The internal structures of the zircon grains were imaged in transmitted and reflected light and by cathodoluminescence (CL) techniques.

U–Pb dating and trace element analyses of zircon were undertaken simultaneously by laser ablation–inductively coupled plasma–mass spectrometry (LA–ICP–MS) at the Key Laboratory of Regional Geology and Mineralization, Hebei GEO University, Hebei, China. The instrument is a quadrupole ICP–MS (THERMO-ICAP RQ) coupled to a 193 nm ArF excimer laser (RESolution-LR) equipped with a Laurin Technic S155 sample chamber and GeoStarμGIS T software. The laser spot size was set to 29 μm for most analyses, with a laser energy density of 3 J/cm$^2$ and repetition rate of 8 Hz. Each analysis included 10 s of blank measurement, 40 s of sample ablation and measurement, and 20 s of sample chamber flushing. The ablated material was carried into the ICP–MS by high-purity He at 0.6 L/min. Argon was also introduced (0.8 L/min) to increase the signal stability. The dwell time was 20 ms for each element. The zircon analyses were calibrated using the NIST 610 glass as an external standard and Si as an internal standard. U–Pb fractionation effects were corrected using zircon 91500 as an external standard. The zircon standards GJ_1 and Plešovice were also used as secondary standards to assess the data quality. Isotope ratios and elemental concentrations of the zircons were calculated using Iolite v. 3.1 (Paton et al., 2010). Data uncertainties are reported as 2σ. Concordia ages and plots were obtained using Isoplot/Ex (3.0) (Ludwig, 2003).

Results and discussion

We collected 12 granitoid samples from the granitoid massifs located in the central Sikhote–Alin orogenic belt between the Bikin and Bol'shaya Ussurka rivers (Fig. 2; Table 1).
U–Pb dating was conducted on zircons without visible inclusions. CL images of representative zircons are shown in Fig. 4, and U–Pb isotopic data for the zircons are given in Supplementary material (Table S1). Most zircons are euhedral and short prismatic in shape, 50–300 μm in length, and have pyramidal terminations and clear oscillatory zoning, which are indicative of a magmatic origin. Th–U concentrations and U–Pb isotopic data are presented in Table 2. In general, most of the zircons have relatively low U contents of 211.8 ± 8.7 to 1644.0 ± 74.3 ppm. Th/U ratios vary from 0.38 to 0.78 (Table 2), and further support a magmatic origin.

Most of the U–Pb data are concordant (Fig. 4). Quartz monzonites from the Pravovalenkuyskiy massif (samples PrVl-1 and -2) yielded mean $^{206}$Pb/$^{238}$U ages of 104.5 ± 0.5 and 102.4 ± 0.5 Ma (MSWD = 1.9 and 0.03), respectively. The biotite–amphibole granodiorite of the Lovlyagin Creek massif (sample Lvl-1) has a mean $^{206}$Pb/$^{238}$U age of 98.6 ± 1.0 Ma (MSWD = 2.0). The granodiorite sample from the Snezhnyy massif yielded a mean $^{206}$Pb/$^{238}$U zircon age of 92.3 ± 0.5 Ma (MSWD = 1.3). Thirty-five analyses of zircons from alkali feldspar granite and monzogranite of the Pereval’nyy massif (samples Prv-1 and -2) yielded mean $^{206}$Pb/$^{238}$U ages of 85.7 ± 1.0 and 86.3 ± 1.0 Ma (MSWD = 1.5 and 1.3). Forty-four zircon analyses of granitoids in the Vodorazdel’nyy massif (samples Vd-1, -2, and -3) yielded mean $^{206}$Pb/$^{238}$U ages of 83.6 ± 0.5, 85.6 ± 1.0, and 83.7 ± 1.0 Ma (MSWD = 1.0, 0.9, and 1.8). The mean $^{206}$Pb/$^{238}$U age of the alkali feldspar granite in the Ust’Malinovskiy massif (sample Mal-1) is 74.5 ± 1.0 Ma (MSWD = 0.3). Thirty-three zircon analyses of granitoids from the Dal’nearminskiy massif (samples DArm-1 and -2) yielded mean $^{206}$Pb/$^{238}$U ages of 73.4 ± 0.5 and 74.3 ± 0.5 Ma (MSWD = 0.7 and 0.5), respectively.

The U–Pb ages reveal at least three granitoid suites of different ages. One is represented by the late Albian–Turonian Pravovalenkuyskiy, Lovlyagin Creek, and Snezhnyy massifs (105–92 Ma), which corresponds to the late stage of syn-orogenic Sikhote–Alin magmatism (i.e., the Tatiba series). The second includes the Santonian Pereval’nyy and Vodorazdel’nyy massifs (86–83 Ma). The third is represented by the late Campanian Dal’nearminskiy and Ust’Malinovskiy massifs (ca. 73 Ma). The
two younger suites of rocks (i.e., Late Cretaceous) were associated with the post-orogenic stage of
the Sikhote–Alin fold belt and were coeval with intrusion of the ESAVPB. The ESAVPB is interpreted
to be a continental margin magmatic arc that formed at the eastern margin of the Paleo-Asian
continent as a result of subduction of the Pacific Plate (e.g., Khanchuk, 2006, and references
therein). However, the studied rocks are located far (at least 60–80 km) from the ESAVPB.

Although Jurassic–Early Cretaceous granitoids (ca. 190–88 Ma) are widely distributed in
China, Late Cretaceous granitoids have not been identified in the area of accretionary complexes
that are coeval with those in the Sikhote–Alin orogenic belt (i.e., the Nadanhada–Alin Ridge: Shao et
al., 1992; or the Wandashan Ridge: Sun et al., 2015), as well as in eastern China (e.g., Wu et al.,
2011; Niu et al., 2015; Tang et al., 2018, and references therein). Based on a magmatic gap of about
40 Myr from 88 to 50 Ma, some studies have proposed a model for the Mesozoic geodynamic
evolution of the eastern margin of Paleo-Asia that contradicts the available geological data for the
Sikhote–Alin and Sakhalin regions, and Japan (e.g., Zhou & Li, 2000; Li et al., 2012; Niu et al., 2015;
Tang et al., 2016). In particular, the occurrence of Jurassic–Early Cretaceous granitoid magmatism in
eastern China is thought to have been related to subduction of Paleo-Pacific Plate beneath the
eastern margin of the Paleo-Asian continent, which in turn caused basal hydration, weakening and
thinning of the lithosphere, and eventually crustal melting and granitic magmatism. Thus, based on
these data, between 190 and 88 Ma the eastern edge of the Paleo-Asian continent was an active
subduction margin (e.g., Zhou & Li, 2000; Li et al., 2012; Niu et al., 2015; Tang et al., 2016). However,
numerous studies of Japan and the Sikhote–Alin region (e.g., Kemkin, 2006; Khanchuk, 2006; Kemkin
et al., 2016; Moreno et al., 2016, and references therein) have demonstrated that subduction of the
Paleo-Pacific Plate during the Latest Jurassic–Earliest Cretaceous was followed by northward
transform movement of the latter along the continental margin. Consequently, from at least 145
Ma, granitic magmatism at the eastern margin of the Paleo-Asian continent was not related to
subduction of the Pacific Plate.
The transform margin existed until the late Cenomanian (ca. 95 Ma), when subduction of the Paleo-Pacific Plate resumed (e.g., Khanchuk, 2006; Kemkin et al., 2016; Moreno et al., 2016; Khanchuk et al., 2019, and references therein), resulting in formation of structural–lithological complexes along the active convergent (i.e., subduction) margin. These complexes are represented by an epicontinental (i.e., continental margin) magmatic arc (i.e., the ESAVPB), a forearc basin (upper parts of the Western Sakhalin and Sorachi–Yezo terranes), and a Cretaceous–Paleogene accretionary prism (the Shimanto terrane). However, if we only consider the ages of granitoids in eastern China, a magmatic gap occurred along the eastern margin of Paleo-Asia during the Late Cretaceous. Niu et al. (2015) proposed that the gap resulted from subduction termination caused by the trench being jammed by an oceanic plateau or a micro-continent between 100 and 90 Ma. This resulted in a change in the movement direction of the Paleo-Pacific Plate, and a change from subduction to strike-slip tectonism (e.g., Niu et al., 2015; Tang et al., 2016).

The U–Pb ages obtained in this study for granitoids from the central Sikhote–Alin orogenic belt show that magmatism at the eastern margin of the Paleo-Asian continent continued in the Late Cretaceous, even at a substantial distance from the convergent margin. In addition, both Albian–Cenomanian (Khanchuk et al., 2019, and references therein) and Late Cretaceous–early Eocene (Jahn et al., 2015; Tang et al., 2016; Grebennikov & Maksimov, 2021; Grebennikov et al., 2021, and references therein) volcanic and plutonic complexes are widely developed at the eastern margin of Asia (southeastern Russia, South Korea, and Japan). Santonian–Campanian granitoids are located mainly along the continental margin of southeastern Russia and Japan (Jahn et al., 2015; Zhao et al., 2017). Thus, age data from all these regions are necessary for reliable geodynamic reconstructions.

The Late Cretaceous granitoids of this study cannot be easily explained by oblique subduction of the Paleo-Pacific Plate beneath northeastern Asia, given the large distance from the suprasubduction structure (ESAVPB) of the pre-oceanic zone. However, based on tectonic reconstructions (Engebretson et al., 1985), the vector of Paleo-Pacific Plate motion at ca. 85 Ma
(Santonian) changed from 338°NW to 294°NW, at ca. 74 Ma (late Campanian) from 294°NW to 315°NW, and in the middle Paleocene it changed from 315°NW to 358°NNW (Fig. 5). Although these transformations would have occurred over a period of time, they coincide with the emplacement ages of the studied Late Cretaceous granitoids, as well as middle Paleocene–early Eocene A-type silicic magmatism (Grebennikov et al., 2020, 2021; Grebennikov & Maksimov, 2021) across the Sikhote-Alin region (Fig. 5).

A-type intrusive and volcanic rocks have not been identified in other Mesozoic–Cenozoic igneous suites across this region (Khanchuk, 2006), including the Albian–Cenomanian (110–95 Ma) orogenic belt (and coeval Pacific Asia igneous province) that formed in a transform setting at the continental margin (Khanchuk et al., 2019). The formation process of A-type silicic magmas is complex (e.g., Bonin, 2007; Dall’Agnol et al., 2012). However, most studies have concluded that such magmatism occurs during intracontinental extension caused by a change in geodynamic setting (e.g., from transpression to transtension) and cannot occur during orthogonal or oblique subduction or the compressional conditions that exist at a convergent margin (Bonin et al., 1998; Grebennikov et al., 2016; Robinson et al., 2017).

An abrupt change in the motion direction of an oceanic plate with respect to the continental plate can cause slab stagnation and detachment (i.e., a slab window and/or slab tear), and upwelling of sub-slab asthenosphere. The input of reduced and high-temperature mantle melts that are nearly anhydrous can cause partial melting of the overlying crust. The sources of A-type or highly fractionated I-type granites, which are mainly related to post-orogenic events, are mostly associated with large-scale extensional structures (e.g., Grebennikov, 2014; Wu et al., 2017).

Our preliminary geochemical data (Table 3) show that the syn-orogenic (late Albian–Turonian) granitoids of the Snezhnyy, Lovlyagin Creek, and Pravovalenkuyskiy massifs are calc-alkaline, metaluminous, and magnesian (Fig. 6), similar to coeval granitoids in the Tatiba terrane. It has been proposed that these granitoids formed by mixing of anatectic melts derived from
metapelites and, to a lesser extent, metabasites (similar to N-MORB), which occurred in the presence of water in regions of compression-dominated and strike-slip tectonism (Golozubov, 2006; Kruk et al., 2014). The post-orogenic intrusive rocks in the Bikin–Bol’shaya Ussurka area have more differentiated, silicic, calc-alkaline, peraluminous, and ferroan compositions than the Late Cretaceous (86–73 Ma) ESAVPB granitoids (Fig. 6). Based on the classification diagram of Grebennikov (2014), they are highly fractionated granites that formed in a post-orogenic or strike-slip tectonic setting (Fig. 6c).

In summary, the studied Santonian granitoids (ca. 89–86 Ma) are unlikely to have formed in a typical suprasubduction setting. This is also the case for the studied late Campanian (ca. 73 Ma) granitoids located along the Central Sikhote–Alin Fault, far inland from the coeval ESAVPB, even though the vector of Pacific Plate motion changed by only 21° at that time (i.e., from 294°NW to 315°NW; Fig. 5).

The emplacement ages of the studied granitoids coincide with the timing of a change in the angle of convergence between the Paleo-Pacific Plate and eastern margin of the Paleo-Asian continent. This change in motion of the oceanic plate with respect to the continental plate was probably caused by a rupture in the subducted slab (i.e., a slab tear), followed by asthenospheric upwelling and partial melting of the overlying crust, which finally led to the generation of the post-orogenic (Santonian and late Campanian) intrusive magmatism (Fig. 7).

Conclusions

1. New zircon U–Pb age data indicate that granitoid suites of the Bikin–Bol’shaya Ussurka area were emplaced at 105–92 Ma (late Albian–Turonian) during syn-orogenic magmatism in the Sikhote–Alin orogenic belt (southeastern Russia), and at 86–83 Ma (Santonian) and ca. 73 Ma (late Campanian) during its post-orogenic stage.
2. Given that Late Cretaceous granitoids were previously unknown in the inland Sikhote–Alin region, as well as adjacent eastern China, our data provide important constraints on the geodynamic evolution of the eastern Asian continental margin.

3. Emplacement ages of the Late Cretaceous central Sikhote–Alin granitoids coincide with changes in the direction of motion of the Pacific Plate with respect to the eastern Asian continent. This was probably related to destruction of the Paleo-Pacific slab and upwelling of asthenosphere that provided the heat for the generation of melts that were not related to subduction.

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Author contributions IVK: conceptualization (lead), investigation (lead), writing – original draft (lead), writing – review & editing (lead), methodology (lead); AVG: funding acquisition (lead), investigation (equal), project administration (equal), resources (equal), supervision (lead), writing – review & editing (equal); XHM: conceptualization (lead), funding acquisition (lead), project administration (lead), supervision (lead), writing – review & editing (equal); KKS: methodology (equal), writing – review & editing (supporting).

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Sakhno, V. G., Kovalenko, S. V. & Lyzganov, A. V. 2016. Granitoid magmatism of the Arminsky ore district in the Central Sikhote-Alin, Primorye: U–Pb dating, $^3$He/$^4$He isotope characteristics,


Figure captions

Figure 1. Geologic map of the Sikhote-Alin region showing distribution of the different-aged igneous rocks (modified after Grebennikov et al. 2016).

Figure 2. Simplified geological map of the Bikin–Bol. Ussurka area (Central Sikhote-Alin). Unabbreviated terrane names are given in Fig. 1.

Figure 3. Representative field photograph and photomicrographs showing typical textures of the Bikin–Bol. Ussurka granitoid suites. (a) Field outcrop (Pereval'nyy massif); (b) Biotite leucogranite; (c) Biotite-muscovite granite; (d) Leucogranite; (e) Aplitic leucogranites. Q = quartz; Kfs = K-feldspar; Pl = Plagioclase; Bi = Biotite; Ms = Muscovite.

Figure 4.1–4.2. U–Pb concordia diagrams.

Pravovalenkuyskiy monzogranite (PrVl-1) and quartz-monzonite (PrVl-2); Lovlyagin Creek granodiorite (Lvl-1); Snezhnyy granodiorite (Sn-1); Pereval'nyy alkali-feldspar-granite (Prv-1) and monzogranite (Prv-2); Vodorazdel'nyy alkali-feldspar-granite (Vd-1) and monzogranites (Vd-2, Vd-3); Ust'Malinoiskiy alkali-feldspar-granite (Mal-1); Dal'nearminskiy monzogranite (DArm-1) and alkali-feldspar-granite (DArm-2).

Figure 5. Stratigraphic scale and Pacific Plate motion vector after Engebretson et al. 1985 with reported U-Pb age data of the Bikin–Bol. Ussurka granitoid suites (coloured), and literature U-Pb age data of middle Paleocene–early Eocene magmatic rocks in the Southern Russian Far East (gray rectangle).
* - insufficiently reliable data.

Figure 6. Geochemical discrimination diagrams: (a) Agpaitic index (Na_2O+K_2O)/Al_2O_3 (in molar quantities; Liégeois & Black, 1987), (b) FeO/(FeO+MgO) vs. SiO_2 diagram showing the boundary between ferroan and magnesian silicic rocks (in wt.%; Frost et al., 2001), (c)
Variations in (Na₂O+K₂O) vs. Fe₂O₃×5 vs. (CaO+MgO)×5 (in molar quantities; Grebennikov, 2014).

Figure 7. Schematic illustrations of magmatic events in different geodynamic settings at the eastern margin of the Paleo-Asia involving tearing of subducting slab caused by change of the oceanic plate motion in respect to the continent.

**Table captions**

Table 1. Description and position of the Central Sikhote-Alin granitoid samples.

Table 2. Age data of the Central Sikhote-Alin granitoid suites.

Table 3. Whole-rock (major oxides, ICP-AES, wt %) compositions of the Central Sikhote-Alin granitoid suites.
Table 1. Description and position of the Central Sikhote-Alin granitoid suites

<table>
<thead>
<tr>
<th>Sample</th>
<th>Location</th>
<th>Rock type</th>
<th>Petrographic texture</th>
<th>Mineral assemblage</th>
<th>Name of massif</th>
</tr>
</thead>
<tbody>
<tr>
<td>PrVl-1</td>
<td>46°17'07.40&quot;N, 136°42'32.40&quot;E</td>
<td>Monzogranite</td>
<td>Coarse-grained</td>
<td>Pl (45-50%), Kfs (20%), Q (20%), Bi (10%), Hbl (≤1%)</td>
<td>Pravoval’kuyskiy</td>
</tr>
<tr>
<td>PrVl-2</td>
<td>46°15'06.84&quot;N, 136°40'00.84&quot;E</td>
<td>Quartz-monzonite</td>
<td>Coarse-grained</td>
<td>Pl (60-65%), Opq (15-20%): Bi, and Hbl, Q (6%), Kfs (10-15%)</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Accessory: Ap, Zrn, Mnz, Aln, Ep, Mg</td>
<td></td>
</tr>
<tr>
<td>Lvl-1</td>
<td>45°27'43.66&quot;N, 135°15'55.41&quot;E</td>
<td>Granodiorite</td>
<td>Fine- to medium-grained, heterogeneous</td>
<td>Pl (60-65%), Kfs (10-15%), Q (10%), Opq (~10%): Hbl, and Bi. Accessory (~2%): Mag, Tit, Zrn, Ap, Aln</td>
<td>Lovlyagin Creek</td>
</tr>
<tr>
<td>Sn-1</td>
<td>46°27'40.28&quot;N, 136°24'16.38&quot;E</td>
<td>Granodiorite</td>
<td>Fine- to coarse-grained, heterogeneous</td>
<td>Pl (60-65%), Kfs (10-15%), Q (10%), Bi, and Hbl (~10%). Accessory (1%): Ap, Zrn, Aln, Ep, Tit, Mg</td>
<td>Snezhnyy</td>
</tr>
<tr>
<td>Prv-1</td>
<td>45°43'10.10&quot;N, 135°10'43.70&quot;E</td>
<td>Alkali-feldspar-granite</td>
<td>Coarse-grained, equigranular</td>
<td>AF (~65%), Q (30%), Bi (~5%). Accessory: Zrn, Ap, Mnz, Aln</td>
<td>Pereval’nny</td>
</tr>
<tr>
<td>Prv-2</td>
<td>45°41'46.32&quot;N, 135°11'28.64&quot;E</td>
<td>Monzogranite</td>
<td>Medium-grained</td>
<td>Pl (~70%), Q (25%), Bi (~5%). Accessory (1%): are Grt, Zrn, Ep</td>
<td></td>
</tr>
<tr>
<td>Vd-1</td>
<td>45°35'32.56&quot;N, 135°13'17.74&quot;E</td>
<td>Alkali-feldspar-granite</td>
<td>Coarse- to fine-grained</td>
<td>AF (~65%), Q (30%), Bi (~5%), Accessory (1%): Zrn, Ap, Grt</td>
<td>Vodorazdel’nny</td>
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<tr>
<td>Vd-2</td>
<td>45°35'23.05&quot;N, 135°15'23.29&quot;E</td>
<td>Monzogranite</td>
<td>Coarse-grained, heterogeneous</td>
<td>AF (~65%), Q (30%), Bi (~5%), Accessory (1%): Mnz, Zrn, Aln</td>
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<tr>
<td>Vd-3</td>
<td>45°35'21.00&quot;N, 135°15'44.10&quot;E</td>
<td>Monzogranite</td>
<td>Coarse-grained</td>
<td>AF (~65%), Q (30%), Bi (~5%), Accessory (1%): Mnz, Zrn, Aln</td>
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<td>Location</td>
<td>Coordinates</td>
<td>Rock Type</td>
<td>Description</td>
<td>Accessory Minerals</td>
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<tr>
<td>Mal-1</td>
<td>46°21'29.70&quot;N, 136°29'31.02&quot;E</td>
<td>Alkali-feldspar-granite</td>
<td>Fine- and medium-grained, porphyritic</td>
<td>Zrn, Xnt, Trm</td>
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<tr>
<td>Ust’Malinovskiy</td>
<td></td>
<td></td>
<td>AF (~60%), Q (25%), Pl (~5-10%), Bi (~5%)</td>
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<tr>
<td>DAzm-1</td>
<td>45°56'01.33&quot;N, 135°24'41.48&quot;E</td>
<td>Monzogranite</td>
<td>Coarse-grained</td>
<td>Zrn, Ap, Mnz, Aln</td>
<td></td>
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<td></td>
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<tr>
<td>Dalnearminskiy</td>
<td></td>
<td></td>
<td>Q (35%), AF (25%), Pl (30%), Bi (~5%), Hbl (~5%)</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td>Q (40%), AF (25%), Pl (25%), Bi (~5%), Hbl (~5%)</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Note: AF - alkali feldspar (including orthoclase, microcline, and albite); Kfs - orthoclase (microcline); Pl - plagioclase; Bi - biotite; Ms - Mscovite; Q - quartz; Hbl - amphibole; Opq - opaque minerals; Px - pyroxene. Accessory minerals: Ap - apatite; Aln - alanite; Ep - epidote; Flr - fluorite; Grt - garnet; Ilm - ilmenite; Mg - magnetite; Mnz - monazite; Tit - titanite; Trm - tourmaline; Xnt - xenotime; Zrn - zircon.
Table 2. Age data of the Central Sikhote-Alin granitoid suites

<table>
<thead>
<tr>
<th>Sample</th>
<th>Points</th>
<th>U aver., ppm</th>
<th>Th aver., ppm</th>
<th>Th/U\pm2\sigma</th>
<th>MSWD (of concord.)</th>
<th>Probability (of concord.)</th>
<th>Age Ma (\pm2\sigma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>PrVI-1</td>
<td>19</td>
<td>505.9\pm14.0</td>
<td>222.4\pm8.8</td>
<td>0.449\pm0.049</td>
<td>1.9</td>
<td>0.17</td>
<td>104.5\pm0.5</td>
</tr>
<tr>
<td>PrVI-2</td>
<td>20</td>
<td>639.9\pm17.2</td>
<td>347.2\pm15.9</td>
<td>0.776\pm0.056</td>
<td>0.03</td>
<td>0.96</td>
<td>102.4\pm0.5</td>
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<tr>
<td>Lvl-1</td>
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<td>211.8\pm8.7</td>
<td>151.4\pm10.3</td>
<td>0.656\pm0.062</td>
<td>2.0</td>
<td>0.16</td>
<td>98.6\pm1.0</td>
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<td>Sn-1</td>
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<td>0.18</td>
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<td>Prv-2</td>
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<td>1263.4\pm45.9</td>
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<td>Vd-1</td>
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<td>1644.0\pm74.3</td>
<td>519.6\pm21.1</td>
<td>0.346\pm0.065</td>
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<td>0.48</td>
<td>85.6\pm1.0</td>
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<tr>
<td>Vd-2</td>
<td>14</td>
<td>1246.1\pm57.1</td>
<td>431.1\pm18.8</td>
<td>0.428\pm0.073</td>
<td>0.98</td>
<td>0.32</td>
<td>83.6\pm0.5</td>
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<tr>
<td>Vd-3</td>
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<td>1054.7\pm37.2</td>
<td>545.3\pm35.5</td>
<td>0.582\pm0.068</td>
<td>1.8</td>
<td>0.07</td>
<td>83.7\pm1.0</td>
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<td>Mal-1</td>
<td>24</td>
<td>404.3\pm14.0</td>
<td>227.2\pm10.1</td>
<td>0.600\pm0.057</td>
<td>0.30</td>
<td>0.58</td>
<td>74.5\pm1.0</td>
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<td>DArm-1</td>
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<td>489.1\pm10.9</td>
<td>274.7\pm9.3</td>
<td>0.552\pm0.042</td>
<td>0.65</td>
<td>0.42</td>
<td>73.4\pm0.5</td>
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<tr>
<td>DArm-2</td>
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<td>508.7\pm14.4</td>
<td>262.4\pm10.2</td>
<td>0.545\pm0.048</td>
<td>0.50</td>
<td>0.48</td>
<td>74.3\pm0.5</td>
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</table>
Table 3. Whole-rock (major oxides, ICP-AES, wt %) compositions of the Central Sikhote-Alin granitoid suites

| Sample | Age, Ma | SiO$_2$ | TiO$_2$ | Al$_2$O$_3$ | Fe$_2$O$_3$ | Mn O | Mg O | Ca O | Na$_2$O | K$_2$O | P$_2$O$_5$ | H$_2$O | LO | Total | A/ CN | A/ N |
|--------|---------|---------|---------|-------------|-------------|------|------|------|--------|--------|----------|--------|-----|-------|-------|-----|-------|
| PrVl-1 | 104.5±0.5 | 65.4    | 4.07   | 16.35       | 4.01        | 0.07 | 1.40 | 2.69 | 3.69   | 3.91   | 0.22      | 0.27   | 1.22 | 99.7  | 8     | 1.08 | 1.59  |
| PrVl-2 | 102.4±0.5 | 55.6    | 0.97   | 15.34       | 8.91        | 0.16 | 4.53 | 6.30 | 3.33   | 3.05   | 0.63      | 0.12   | 0.77 | 99.7  | 5     | 0.76 | 1.75  |
| Lvl-1  | 98.6±1.0  | 63.5    | 0.50   | 15.30       | 5.43        | 0.10 | 2.70 | 4.72 | 3.18   | 3.29   | 0.17      | 0.18   | 0.63 | 99.7  | 4     | 0.88 | 1.74  |
| Sn-1   | 92.3±0.5  | 67.8    | 0.49   | 14.92       | 4.16        | 0.08 | 1.50 | 3.38 | 3.31   | 2.97   | 0.13      | 0.17   | 0.63 | 99.5  | 7     | 1.01 | 1.72  |
| Prv-1  | 85.7±1.0  | 73.6    | 0.22   | 14.15       | 2.11        | 0.03 | 0.24 | 0.35 | 3.50   | 4.25   | 0.04      | 0.39   | 0.70 | 99.6  | 4     | 1.29 | 1.37  |
| Prv-2  | 86.3±1.0  | 73.4    | 0.32   | 12.65       | 2.65        | 0.06 | 0.54 | 1.58 | 3.73   | 3.74   | 0.10      | 0.34   | 0.66 | 99.8  | 3     | 0.97 | 1.24  |
| Vd-1   | 83.6±0.5  | 76.8    | 0.08   | 12.78       | 1.36        | 0.04 | 0.14 | 0.18 | 3.26   | 4.27   | 0.02      | 0.07   | 0.53 | 99.5  | 3     | 1.24 | 1.28  |
| Vd-2   | 85.6±1.0  | 75.0    | 0.11   | 13.24       | 1.58        | 0.05 | 0.19 | 0.99 | 3.53   | 4.24   | 0.05      | 0.17   | 0.63 | 99.8  | 8     | 1.09 | 1.27  |
| Vd-3   | 83.7±1.0  | 74.4    | 0.14   | 13.27       | 2.48        | 0.09 | 0.24 | 0.79 | 3.30   | 3.89   | 0.06      | 0.30   | 0.95 | 99.9  | 2     | 1.20 | 1.38  |
| Mal-1  | 74.5±1.0  | 75.8    | 0.15   | 12.86       | 1.94        | 0.04 | 0.14 | 0.17 | 3.05   | 4.46   | 0.03      | 0.21   | 0.89 | 99.7  | 6     | 1.27 | 1.31  |
| DArm-1 | 73.4±0.5  | 73.5    | 0.20   | 14.12       | 2.14        | 0.04 | 0.34 | 1.16 | 3.62   | 4.01   | 0.06      | 0.13   | 0.17 | 99.5  | 7     | 1.14 | 1.37  |
| DArm-2 | 74.3±0.5  | 74.1    | 0.20   | 13.78       | 1.93        | 0.04 | 0.25 | 0.39 | 3.69   | 4.52   | 0.06      | 0.20   | 0.70 | 99.8  | 7     | 1.18 | 1.26  |

Note: Total Fe reported as Fe$_2$O$_3$. The geochemical determinations were carried out in the Primorsky Centre of Local Elemental and Isotopic Analysis of FEGI FEB RAS, Vladivostok, Russia. The determination of major elements of the studied samples in oxide equivalent was performed using ICP-AES on iCAP 6500 Duo spectrometer (Thermo Scientific, USA) with the addition of cadmium solution (10 ppm concentration) as internal standard. The determinations of H$_2$O and SiO$_2$ were performed using gravimetry.
Figure 1

- Central Sikhote-Alin Fault;
- Early Paleozoic continental blocks: JM = Jiamusi, KHA = Khanka; BU = Burea;
- Fragment of the Paleozoic continental margin thrust onto Jurassic accretionary prisms: SR = Sergeevka terrane;
- Jurassic accretionary prism terranes containing fragments of Paleozoic ophiolites, cherts, and limestones, and Triassic to Middle Jurassic cherts: SM = Samarka, NB = Nadankhada–Bikin, KH = Khabarovsk;
- Middle Cretaceous accretionary prism terrane containing fragments of Jurassic basalt, chert, and limestones, and rocks derived from Early Cretaceous island arcs: KM = Kiselevka-Manoma;
- Early Cretaceous (Neocomian) accretionary prism terrane containing fragments of Devonian to Triassic limestones, basalts, and late Paleozoic and Triassic–Jurassic cherts and argillites: TU = Tukhka;
- Early Cretaceous turbidite basin terrane: Zh-A = Zhuravlevka-Amur;
- Early Cretaceous island arc system: KE = Kema;
- Late Cretaceous granitoids (a), and volcanic rocks (b);
- Hauterivian–Aptian granitoids.
Figure 2

- Late Paleogene volcanic and sedimentary rocks;
- Middle Paleocene–early Eocene magmatic rocks;
- Late Cretaceous volcanic (a), and intrusive (b) rocks of the East Sikhote-Alin Volcanic-Plutonic Belt (EASVPB);
- Late Campanian (~73 Ma) granitoids: confirmed by U/Pb age dating (a), and assumed age (b);
- Santonian (86–83 Ma) granitoids confirmed by U/Pb age dating;
- Late Albian-Turonian (105–89 Ma) granitoids: confirmed by U/Pb age dating (a), and assumed age (b);
- Zhuravlevka–Amur Early Cretaceous turbidite basin terrane;
- Samarka Jurassic accretionary prism terrane;
- Central Sikhote–Alin Fault.

99.9 = Sample and U-Pb age dating: red colour (this study), black colour (published).
1 = Intrusive massifs: 1 - Vodorazdeľnýy; 2 - Daňnearminsky; 3 - Perevalnyy; 4 - Pravovalenkuyskiy; 5 - Snezhnyy; 6 - Ust'Malinovskiy; 7 - Lovlyagin Creek.
Figure 3
Figure 4.2

Profiles of concordia age of samples Vd-1, Vd-2, Vd-3, Mal-1, DArm-1, DArm-2.

- **Vd-1**
  - Concordia Age: 85.6 ± 1.0 Ma
  - MSWD (of concordance): 0.91
  - Probability (of concordance): 0.48

- **Vd-2**
  - Concordia Age: 83.6 ± 0.5 Ma
  - MSWD (of concordance): 0.98
  - Probability (of concordance): 0.32

- **Vd-3**
  - Concordia Age: 87.3 ± 1.0 Ma
  - MSWD (of concordance): 0.02
  - Probability (of concordance): 0.98

- **Mal-1**
  - Concordia Age: 74.5 ± 1.0 Ma
  - MSWD (of concordance): 0.30
  - Probability (of concordance): 0.58

- **DArm-1**
  - Concordia Age: 73.4 ± 0.5 Ma
  - MSWD (of concordance): 0.65
  - Probability (of concordance): 0.42

- **DArm-2**
  - Concordia Age: 74.3 ± 0.5 Ma
  - MSWD (of concordance): 0.50
  - Probability (of concordance): 0.58
Figure 5

<table>
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<tr>
<th>CRETACEOUS</th>
<th>ALBIAN</th>
<th>346°</th>
<th>100 Ma</th>
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<tr>
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<td>95 Ma</td>
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<td>Santonian</td>
<td>294°</td>
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<td>Maastrichtian</td>
<td>315°</td>
<td>74 Ma</td>
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<td>Selendian</td>
<td>358°</td>
<td>56–53 Ma*</td>
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<tr>
<td>Ypresian</td>
<td>Lutetian</td>
<td>338°</td>
<td>48 Ma*</td>
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<tr>
<td>Bartonian</td>
<td>Priabonian</td>
<td>303°</td>
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<tr>
<td>Eocene</td>
<td>Paleogene</td>
<td>283°</td>
<td>37 Ma</td>
</tr>
</tbody>
</table>

*Ma: Million Years before the present.
Figure 6

= Santonian–Campanian granitoids (reference data from Kruk et al., 2014; Jahn et al., 2015; Grébennikov et al., 2016; Zhao et al., 2017);

= Late Albian–Turonian granitoids (reference data from Kruk et al., 2014; Jahn et al., 2015; Grébennikov et al., 2016; Zhao et al., 2017).
Figure 7

Geodynamic setting of transform margin
Formation of the new continental lithosphere

Geodynamic setting of transform margin
Paleocene–Early Eocene magmatic stage

Geodynamic setting of subduction margin
Early stage of the ESAVPB formation

Model of slab tear formation caused by change of the oceanic plate motion in respect to the continental plate

Geodynamic setting of subduction margin
Late stage of the ESAVPB formation

Legend:
- Pre-Jurassic continental blocks;
- Paleo-Pacific (Izanagi) plates;
- Sikhote–Alin orogenic belt;
- Early Cretaceous island arc system;
- Late Cretaceous Shimanto accretionary prism;
- magmatic events.