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Early Paleozoic High- and Ultrahigh-Pressure Complexes in the Western Part of the Central Asian Orogenic Belt: Ages, Compositions, and Geodynamic Models of Formation

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Abstract—The western part of the Central Asian Orogenic Belt, which comprises folded areas in Kazakhstan, Kyrgyzstan, and northwestern China, includes a number of large Precambrian sialic massifs that are framed by deformed and dismembered Paleozoic ophiolites and by island arc and flysch formations. The basements of the massifs are commonly made up of diverse metamorphic complexes, some of which were metamorphosed under high and ultrahigh pressures in the Early Paleozoic at ~480–530 Ma. These metamorphic formations are the Zerendy Group of the Kokchetav massif in northern Kazakhstan; Akdzhon Group of the Issyk-Kul massif in the northern Tien Shan); Aktyuz, Kemin, and Koyandy complexes of the Chu-Kendyktas and Zheltau massifs in southern Kazakhstan and the northern Tien Shan; and the Kassan Group of the Ishim–Naryn massif in the central Tien Shan. The paper reviews data on the structures, compositions, and metamorphic evolutions of the high- and ultrahigh-, and medium-pressure metamorphic rocks of these massifs. Numerous *P–T* assessments have been made for the near-peak and/or post-peak retrograde metamorphism, and some prograde *P–T* paths have been calculated for the key rock types over the past three decades of the studies. Near-peak and/or post-peak metamorphic ages and some ages of retrograde metamorphism are estimated for most of the high- and ultrahigh-pressure rocks. The paper discusses problems faced by the researcher when building geodynamic models for the high- and ultrahigh-pressure complexes in various massifs of the western part of the Central Asian Orogenic Belt. It is shown that any reliable model shall be underlain by detailed information on the compositions, ages, and formation environments of the protoliths for the ultrahigh-, high-, and medium-pressure rocks and complexes. Moreover, the structures and compositions of Paleozoic complexes surrounding the Precambrian massifs shall also be taken into consideration.

Keywords: high-pressure metamorphism, Central Asian Orogenic Belt, eclogite, geochronology, *P–T* paths, protoliths, geodynamics, Kazakhstan, Tien Shan

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INTRODUCTION

Metamorphic complexes with high- and ultrahigh-pressure rocks have attracted keen interest of many researchers worldwide over the past decades. These are Dabe-Sulu in China, Bohemian massif in the Czech Republic, high-pressure complexes in western Norway, Armorican massif in France, and many others (Medaris et al., 1995; Cong and Wang, 1999; Root et al., 2005; Kotková, 2007; Ballèvre et al., 2009). These complexes were reportedly formed in relation to the subduction of fragments of oceanic or thinned continental crust to significant (>50 km) depths and their subsequent exhumation according to diverse tectonic scenarios. Depending on the composition of the protoliths, high- and ultrahigh-pressure metamorphism produces garnet and spinel peridotites, eclogites, garnet–kyanite gneisses, etc. Tectono-stratigraphic complexes occurring in association with (ultra)high-pressure rocks may vary in composition and metamorphic grades because of the involvement of rocks from various depth levels in the subduction and exhumation processes (Maruyama et al., 1996; Ernst et al., 2007). The reliability of geodynamic models for a given territory is thus much greater if these models involve data on the discovered metamorphic complexes with (ultra)high-pressure rocks, their *P–T–t* evolution, analysis of the chemical and mineralogical composition of the rocks, data on environments in which their protoliths were formed, and information on key rock varieties found in association with the (ultra)high-pressure rocks and their relationships.

The Central Asian Orogenic Belt is one of the principal structures of Eurasia. It was formed during the evolution of the Paleo-Asian Ocean in the Late Neoproterozoic–Early Mesozoic (Fig. 1a). The western part of this belt comprises folded areas in Kazakhstan, Kyrgyzstan, and northwestern China and is characterized by the occurrence of large massifs with Precambrian continental crust, which are surrounded by intensely deformed Paleozoic complexes, including ophiolites, island-arc and flysch complexes (Degtyarev et al., 2017; Yarmolyuk and Degtyarev, 2019) (Fig. 1b). The Precambrian massifs contain variably metamorphosed Proterozoic magmatic and sedimentary complexes and a weakly metamorphosed terrigenous–carbonate cover, which was formed in the Early Ediacaran–Early Paleozoic (Degtyarev et al., 2017).

High- and ultrahigh-pressure metamorphic rocks in the basement complexes of the massifs were traditionally viewed as the oldest ones and were ascribed to the Archean–Early Proterozoic (Nedovizin, 1963; Abdulin et al., 1980; Kozakov, 1993). The basement metamorphic rocks often include eclogites, garnet amphibolites, and garnet–mica gneisses and schists, whose genesis remain largely disputable (Abdulkabirova, 1946; Efimov, 1962; Sobolev, 1977; Kushev and Vinogradov, 1978; Efimov et al., 1983; Dobretsov et al., 1989; Kozakov, 1993). High- and ultrahigh-pressure metamorphic complexes have been studied over the past decades in many Precambrian massifs in the western part of the Central Asian Orogenic Belt. The estimated metamorphic peaks in such complexes in the Kokchetav, Issyk-Kul (northern Tien Shan), Chu-Kendyktas, Zheltau, and Ishim–Naryn (Ishim–Central Tien Shan) massifs occurred in the Cambrian–Early Ordovician (Tagiri et al., 1995; Shatsky et al., 1999; Zhang et al., 1997; Okamoto et al., 2000; Katayama et al., 2001; Togonbaeva et al., 2009; Orozbaev et al., 2010; Kröner et al., 2012; Meyer et al., 2013; Rojas-Agramonte et al., 2013; Klemd et al., 2014; and others). Coesite and diamond microinclusions found in metamorphic minerals of the Kokchetav and Issyk-Kul massifs (Rozen et al., 1972; Sobolev and Shatsky, 1990; Tagiri et al., 1995; Korsakov et al., 1998, 2007; and others), considered together with other petrological and mineralogical–microtextural indications and with newly acquired geochronologic data, indicate that these rocks were formed as a result of subduction of various crustal horizons to depths up to 120 km, and the subsequent exhumation of the high- and ultrahigh-pressure rocks to shallower depths. It has also been established that all high-pressure complexes are characterized by long-lasting crustal prehistories, which imply that the Precambrian rocks were involved in Early Paleozoic subduction and collision processes (Shatsky et al., 1999; Kröner et al., 2007, 2012; Konopelko et al., 2012; Rojas-Agramonte et al., 2014; Tretyakov et al., 2016; Degtyarev et al., 2014, 2017).

This reviewing paper reports the most principal results of structural–geological, mineralogical–petrographic, isotope-geochemical, and geochronologic studies conducted at Early Paleozoic high- and ultrahigh-pressure complexes in the western part of the Central Asian Orogenic Belt over more than past three decades (Table 1).

KOKCHETAV MASSIF

The Kokchetav massif in northern Kazakhstan is one of the largest Paleozoic block in the western part of the Central Asian Orogenic Belt and is dominated by late Precambrian metamagmatic and metasedimentary rock complexes. The massif is surrounded by Early Paleoproterozoic volcano-sedimentary rocks in the west, east, and south. In the north, the complexes of the Kokchetav massif are overlain by Mesozoic–Cenozoic sedimentary sequences of the cover of the West Siberian plate. The massif hosts large Early to Middle Paleozoic granite and granodiorite plutons, which cut all of the older rocks (Fig. 2).

The late Precambrian complexes of the Kokchetav massif host widespread and variably mylonized orthogneisses and gneiss-granites of the Zerendy Group, whose protolith was dated at 1170–1140 Ma. The Sm–Nd and Lu–Hf isotope characteristics of these rocks indicate that they were formed by the recycling of complexes of early Precambrian (2.6–2.1 Ga) continental crust (Tretyakov et al., 2011, 2016; Turkina et al., 2011; Glorie et al., 2015). The orthogneisses and gneiss-granites often host amphibolite bodies and boudins of variable size, which are locally dominant in the vertical sections of the Zerendy Group. Rocks of the Zerendy Group are usually metamorphosed to grades no higher than the amphibolite facies. The moderate-grade amphibolites and gneisses include nappes of rocks metamorphosed up to the eclogite facies (diamond- and coesite-bearing gneisses and schists, eclogites, garnet peridotites, ultrahigh-pressure calc–silicate rocks, etc.). The age of the amphibolite-facies rocks is uncertain and is thought to be close to the age of the high-pressure metamorphism.

The metamorphic complexes with high-pressure (HP) and ultrahigh-pressure (UHP) rocks are overlain by weakly metamorphosed quartzite–schist sequences (Kokchetav Group and its analogues) (Dobrzhinetskaya et al., 1994; Dobretsov et al., 1995; Degtyarev et al., 2017; Kovach et al., 2017). The sources of eroded material for rocks of the Kokchetav Group were dominated by Mesoproterozoic (1.1–1.5 Ga) complexes derived from mostly juvenile source (Kovach et al., 2017).

The high- and ultrahigh-pressure rocks of the Zerendy Group were intensely studied over the past three decades. The reasons for this were the great diversity of the rock types combined within the same structures, finds of rare minerals (such as diamond, coesite, high-K clinopyroxene, K-bearing tourmaline, kumdykolite and kokchetavite, which are orthorhombic and hexagonal albite and K-feldspar, respectively, and others) (e.g., Letnikov et al., 1983; Sobolev and Shatsky, 1990; Shatsky et al., 1999; Schertl and Sobolev, 2013).

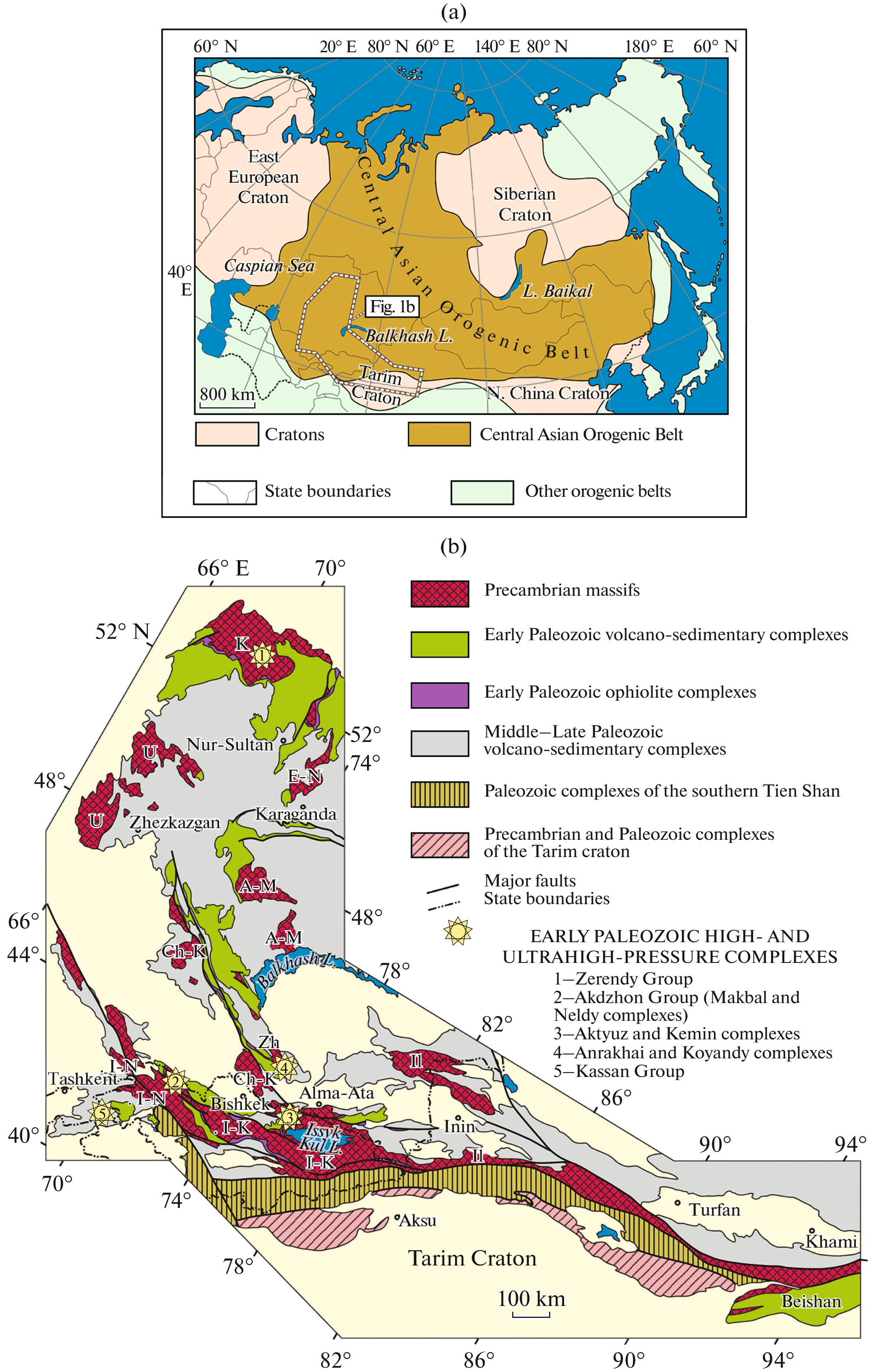

Fig. 1. (a) Location map of the Central Asian Orogenic Belt in Northern Eurasia and (b) a schematic map of the distribution of Precambrian massifs in the western part of the Central Asian Orogenic Belt (Degtyarev et al., 2017): K—Kokchetav, E-N—Erementau—Niyaz, U—Ulutau, Zh—Zheltau, Ch-K—Chu-Kendyktas, I-K—Issyk-Kul, A-M—Aktau—Mointy, I-N—Ishim—Naryn, Y—Yili.

Table 1. Data on the P–T parameters of metamorphism, timing of the near-peak and later stages of metamorphism occurrence, compositions of the protoliths of metamorphic rocks in the Kokchetav, Issyk-Kul, Chu-Kendyktas, Zheltau, and Ishim–Naryn massifs in the western segment of the Central Asian Orogenic Belt

Massif		Group	Locality, complex	Rocks	P–T parameters of peak metamorphism	P–T parameters of retrograde metamorphism
Kokchetav (Fig. 2)	Western domain	Zerendy	Barchi-Kol locality	Eclogites	27–40 kbar; 700–950°C	
				Garnet amphibolites with *Cpx*	12–14 kbar; 700–815°C	
				Garnet amphibolites	12 kbar; 700°C	
				Epidote amphibolites	8.6 kbar; 500°C	
				Diamond-bearing *Grt-Bt* gneisses and calc-silicate rocks	>40 kbar; 950–1000°C	10–12 kbar; 650–800°C
				Coesite-bearing *Ky-Grt* mica schists and *Grt-Bt* gneisses	29 kbar; 800–900°C	
			Kumdy-Kol locality	Diamond-bearing *Grt-Bt* gneisses, metapelites	>40–60 kbar; 800–1000°C	~10 kbar; 740–790°C and ~10 kbar; 650–715°C
				Coesite-bearing eclogites	30 kbar; 780–900°C	
				Diamond-bearing *Grt-Cpx-Qz* rocks	>40 kbar; 920–1050°C	
				Garnet peridotites with *Ti-Chu*	>30 kbar; >740°C	
				Calc-silicate rocks	>40–60 kbar; 800–1000°C	25 kbar; 800°C and <15 kbar; 790°C
				Granitoid veins in diamond-bearing gneisses	Melting at ~1000°C	
	Eastern domain		Kulet and Soldat-Kol localities	Coesite-bearing garnet-mica schists (± kyanite)	34–36 (28–35) kbar; 720–760°C	~8 kbar; 600°C
				(coesite)-talc-garnet-kyanite- phengite schists		
				Phengite eclogites	27–35 kbar, 560–720°C	7–13 kbar; 540–720°C
				Amphibole-garnet-zoisite rocks		
			Sulu-Tobe locality	Eclogites	14–16.5 kbar; 600–860°C	
			Borovoe, Chaikino, and Chaglinka localities	Eclogites (Chaglinka)	620–740°C	
				Mylonitized gneisses		
				Eclogites	18–20 kbar; 800–850°C and 17–18 kbar; 750–800°C	11–12 kbar; 760–790°C, 7–8 kbar; 700–730°C and 5–6 kbar; 570–600°C
				Mica schists		
			Enbek-Berlyk locality	Spinel "harzburgites"	14–15 kbar; 780–840°C	
				Garnet-kyanite- sillimanite -biotite schists	4–7 kbar; 600–700°C	
	Daulet Formation			Andalusite–sillimanite–biotite–cordierite–feldspar - metapelites	2–3 kbar; 500–680°C	
Issyk-Kul (Fig. 3)		Akdzhon	Makbal complex	Eclogites and "glaucophanites"	20–25 kbar; 525–560°C	8–13 kbar; 300–500°C
				Garnet amphibolites after eclogites		
				Quartzites		
				Coesite-garnet-chloritoid–talc schists	28–28.5 kbar; 525–560°C	24 kbar; 580°C
				Quartzite-schists with garnet and coesite	>24 kbar	
			Neldy complex	Eclogites	14 kbar; 620°C and 22–25 kbar; 550–610°C	6.5–12 kbar; 430–630°C
				Garnet-phengite-biotite schists	9–17 kbar; <630°C and 6.5–12 kbar; 430–630°C	
				Schists with garnet and chloritoid relics	12–15 kbar; 485–545°C	>3 kbar; ~ 500°C
				Calc-silicate rocks	11–13 kbar; 600°C	
Chu-Kendyktas (Fig. 4)			Aktyuz complex	Eclogites (garnet amphibolites after eclogites)	16–23 kbar; 550–670°C	10–11.5 kbar at 730, 600–650, 550–570°C
			Aktyuz complex	Garnet-bearing paragneisses	13–15 kbar; 635–745°C	
Zheltau (Fig. 4)			Koyandy complex	Garnet-mica gneisses and schists	15–18 kbar; 750–850°C	580–620°C
				Eclogites	15–18 kbar; 700–800°C	10–14 kbar; 700–750°C; 8–10 kbar; 600–650°C
				Garnet clinopyroxenites	16.5–17.5 kbar; 800–860°C	
				Spinel peridotites		11–14.5 kbar; 580–800°C
			Anrakhai complex	Granitoid veins in gneisses		
Ishim-Naryn		Kassan	Shaldyr complex	Eclogites	16–18 kbar; 490–540°C	11–8 kbar; 560°C
			Semizsai complex	Chlorite–albite schists and two-mica schists		
			Ishtanberdy complex	Kyanite-staurolite-garnet-biotite schists	7.2 kbar; 650°C	

Table 1. (Contd.)

Age of near-peak metamorphism	Age of retrograde metamorphism	Protolith	Geochronologic and isotopic–geochemical constraints on the protolith age	References
		Basalts N-MORB or IAB		Korsakov et al., 1998; Masago, 2000; Hermann et al., 2001; Korsakov et al., 2002; Stepanov et al., 2016
528 ± 3 Ma	503–532 Ma	Clay shales and limestones; calcareous clay shales	Zircon core 207Pb/206Pb = 2867 ± 72 Ma	
528–522 Ma				
530 ± 7 and 537 ± 9 Ma	517 ± 5 ahd 515 ± 3; 507 ± 8; 456–461 Ma	Sedimentary rocks of mixed composition	Detrital zircons ~560–627, 694–767, 906–1003 and 1952–1981 Ma	Jagoutz et al., 1989; Claoue-Long et al., 1991; Shatsky et al., 1993; Shatsky et al., 1995; Borisova et al., 1995; Zhang et al., 1997; Ogasawara, 2000; Katayama et al., 2000; Katayama et al., 2001; Reverdatto et al., 2008; Ragozin et al., 2009; Yui et al., 2010; Shatsky et al., 2018; Skuzovatov et al., 2021
528 ± 7 and 535 ± 3 Ma	508 ± 4 Ma	Basalts N-MORB, less E-MORB or IAB	T_{Nd}(DM) = 1.95–0.67 Ga; T_{Hf}(DM) = 1.02–0.79 Ga	
528 Ma		Chloritized basalts		
		Limestones		
526 ± 2 Ma	512 ± 5 Ma	HP-HP melting of diamond-bearing gneisses	zircon core ^{207}Pb/^{206}Pb = 1.1 Ga	
	519 and 521 Ma; 499 and 505 Ma			Shatsky et al., 1993, 1998; Zhang et al., 1997; Ota et al., 2000; Parkinson et al., 2000; Theunissen et al., 2000; Hacker et al., 2003; Masago et al., 2009; Zhang et al., 2012; 2016
526 ± 9 Ma	498 ± 11 Ma	Metasomatized basaltic rocks		
532 ± 58 Ma	497 ± 5 Ma		Zircon cores 1421 ± 13 Ma	
522 ± 5 Ma				
	512 ± 0.9 Ma	Basalts N-MORB	T_{Nd}(DM) = 1.95–0.67 Ga	Kaneko et al., 2000; Hacker, 2003; Theunissen et al., 2003; Shatsky et al., 1993, 2018; Dobretsov et al., 2006
				Shatsky et al., 1993; Kaneko et al., 2000, 2002; Glorie et al., 2015
537 ± 7 and 524 ± 5 Ma	484 ± 10 and 490 ± 9 Ma	Granitoids	T_{Hf}(DM) = 2.65–2.48 Ga; 1113 ± 24 and 1130 ± 18 Ma; 1111 ± 42 and 1137 ± 36 Ma	
				Zhimulev et al., 2010, 2011
	493 ± 5 Ma			
		Chloritized basalts		Dobretsov et al., 1995; Reverdatto and Selyatitskiy. 2005; De Grave et al., 2006
	476–496 Ma			
	515 ± 5 and 461–516 Ma		1280 Ma; 1138–1143 Ma	Katayama et al., 2001; Kaneko et al., 2002; Terabayashi et al., 2002; Buslov et al., 2010
482 ± 17Ma; 509 ± 7 and 498 ± 7 Ma		Basalts N-MORB (?)/ intraplate basalts (?)	Zircon cores ~820 and ~700 Ma	Tagiri et al., 2010a; Togonbaeva et al., 2010b; Meyer et al., 2013, 2014; Rojas-Agramonte et al., 2014; Klemd et al., 2015; Konopelko et al., 2012, 2016; Bakirov, 2017; Kasymbekov et al., 2020; Alexeiev et al., 2020
470 ± 3 Ma		Basalts N-MORB (?)/ intraplate basalts (?)		
		Terrigenous rocks	Detrital zircons 1600–3780 Ma (with maxima at 1840–2000 Ma)	
509 ± 13, 502 ± 10 Ma and 475 ± 4 Ma		Metasomatized basalts N-MORB (?)/ sedimentary rocks (?)	Detrital zircons 642–2583 Ma	
		Terrigenous rocks		
526 ± 10 Ma				
524 ± 13 Ma				
474 ± 2 Ma	462 ± 7 Ma	Enriched continental basalts		Tagiri et al., 1995; Orozbaev et al., 2007, 2010; Kroner et al., 2012; Rojas-Agramonte et al., 2013; Klemd et al., 2014, 2015
			Detrital zircons 504–2460 Ma with maxima at ~1100–1300 Ma	
Zircon rims 460 ± 11 and 486 ± 11 Ma		Terrigenous rocks	Detrital zircons ~667–834, 868–1051, 1087–1220, 1296–1378 and 2464–2539 Ma and maxima at ~985 and 1151 Ma	Kröner et al., 2007; Tretyakov et al., 2011; Alexeiev et al., 2011; Pilitsyna et al., 2018a, 2018b, 2019; Pilitsyna and Tretyakov, 2020
		Derivatives of intraplate melts		
490 ± 3 Ma				
		Fragments of layered oceanic lithospheric ultramafic-mafic complex		
483–489 Ma		Melting of orthogneisses of the Anrakhai (?) complex	Zircon cores 755 ± 5 Ma and one core 2556 Ma	
Disputable (PZ_1 or PZ_3)		Intraplate basalts		Bakirov et al., 1996, 2003; Ivleva, 2003, 2010; Rojas-Agramonte et al., 2013; Loury et al., 2015; Alexeiev et al., 2016; Mühlberg et al., 2016
			Detrital zircons 510–3000 Ma with maxima at ~1000–1200 Ma	

Mineral symbols: *Grt*—garnet, *Bt*—biotite, *Cpx*—clinopyroxene, *Qz*—quartz, *Ky*—kyanite, Ti-*Chu*—Ti-clinohumite.

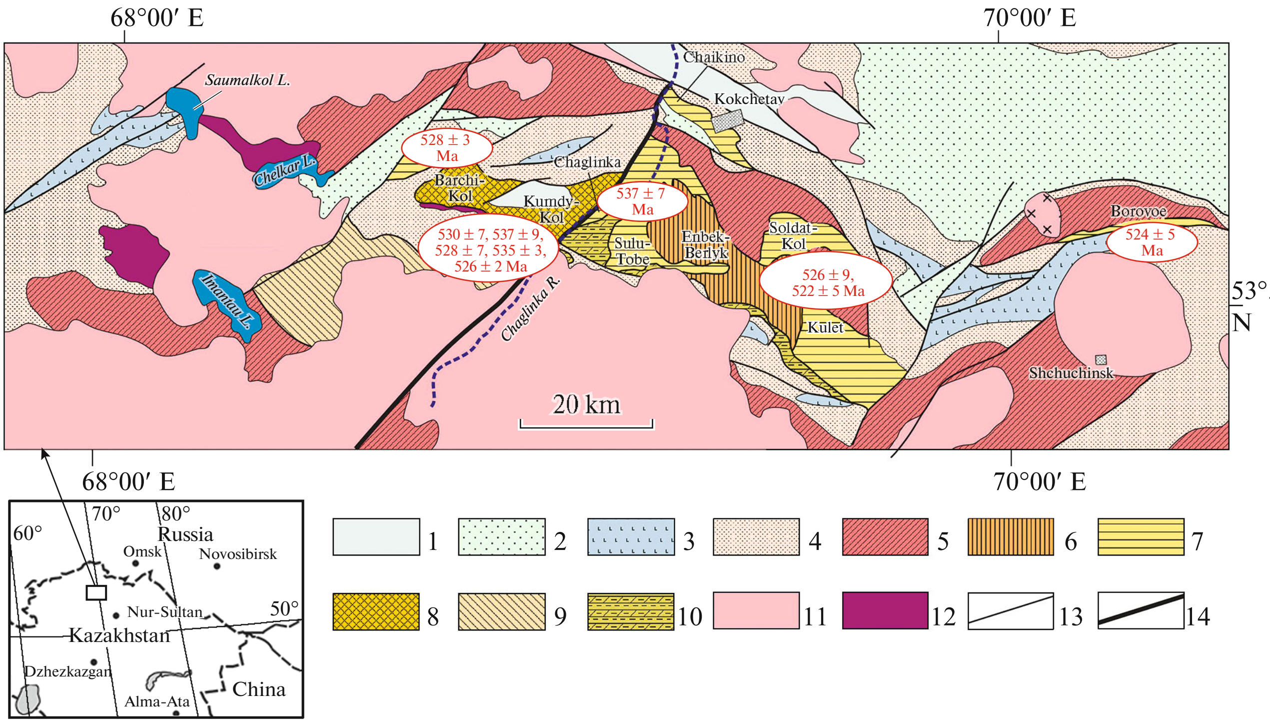

Fig. 2. Schematic geological map of the northeastern part of the Kokchetav massif (modified after Dobretsov et al., 1995, 1998; Kaneko et al., 2000; Degtyarev et al., 2016).
(1) Devonian and Carboniferous terrigenous–carbonate sequences; (2) Ordovician flysch and siliceous–terrigenous sequences; (3) Cambrian volcano-sedimentary sequences; (4) latest Meso- to earliest Neoproterozoic quartzite–schist sequences (Kokchetav Group); (5–8) Zerendy Group: (5) Mesoproterozoic gneiss–granites, orthogneisses, amphibolites, migmatites, blastomylonites after these rocks, including those affected by low-pressure Early Paleozoic metamorphism (no higher than the amphibolite facies), (6) peraluminous schists with amphibolite bodies (Enbek-Berlyk), (7) UHP and HP complexes without microdiamonds (Sulu-Tobe, Kulet, Soldat-Kol, Chaikino, and Chaglinka), (8) UHP diagram-bearing complexes (Barchi-Kol and Kumdy-Kol); (9) intercalating schists, quartzite-schists, quartzites, and amphibolites (compositions of the Beloe Lake domain of uncertain age); (10) Daulet Formation; (11) Early and Middle Paleozoic granitoids; (12) Early Paleozoic mafic and ultramafic rocks, including alkaline; (13) faults; (14) Chaglinka fault. The map shows principal age estimates of near-peak high- and ultrahigh-pressure metamorphism of rocks from the Kokchetav massif.

The most comprehensive information on the structures, chemical and mineral compositions, P–T–t paths, and settings in which the principal varieties of the diamond-bearing ultrahigh-pressure rocks of the Kokchetav massif were formed is presented in (Schertl and Sobolev, 2013).

The high- and ultrahigh-pressure rocks of the Zerendy Group in the Kokchetav massif make up a west-northwestward trending zone that extends for 150 km and is more than 25 km wide. The rocks were most thoroughly studied at a few key sites in two domains, which are separated from each other by the northeast-trending Chaglinka fault (Fig. 2): the western domain, which includes the highest grade rocks, and the eastern one, in which lower grade rocks were found, including those metamorphosed under medium pressures (Dobretsov et al., 1995, 1998; Theunissen et al., 2002; Kaneko et al., 2000). Most researchers focused their studies at the Barchi-Kol and Kumdy-Kol localities in the western domain and the Kulet, Soldat-Kol, Sulu-Tyube, Chaglinka, Borovoe, Chaikino, and Enbek-Berlyk ones in the eastern domain. Low-grade metapelites of the Daulet Formation are found south of the rocks of the Zerendy Group. The two complexes occur in tectonic relationships, but the character of these relationships remains a matter of discussions (Dobretsov et al., 1998; Kaneko et al., 2000; Terabayashi et al., 2002; Buslov et al., 2010).

Western Domain (UHP Rocks)

Barchi-Kol locality. UHP rocks occur there in an area of approximately 5 × 10 km^2 and are tectonically overlain by quartzite-schists of the Kokchetav Group in the northwest, west, and south and are cut by the Krasnomaisky alkaline–ultramafic complex in the south; the Sm–Nd age of the latter rocks is 464 ± 30 Ma (Dobretsov et al., 1998; Masago, 2000; Letnikov et al., 2004; Stepanov et al., 2016) (Fig. 2). The Barchi-Kol metamorphic rocks were produced at various metamorphic grades and, correspondingly, differ in mineral assemblages. The highest grade metamorphic rocks are amphibolized eclogites, garnet pyroxenites, ultrahigh-pressure calc–silicate rocks, mica schists, and various gneisses (Masago, 2000). Minerals of these rocks often contain microdiamond and coesite inclusions or quartz pseudomorphs after coesite (coesite inclusions in zircon from the eclogites, coesite and microdiamond inclusions in zircon and, more rarely, garnet from the clinozoisite gneisses, microdiamond inclusions in zircon from the garnet–biotite gneisses and calc-silicate rocks, and coesite inclusions in zircon from the kyanite–garnet–mica schists), which indicates that the minerals were formed by ultrahigh-pressure metamorphism (Korsakov et al., 1998, 2002; Hermann et al., 2001; Stepanov et al., 2016). The medium-pressure rocks are epidote amphibolites with minor garnet amounts, garnet amphibolites with zoisite, and garnet amphibolites with Na-augite relics (Masago, 2000).

The peak metamorphic parameters of the eclogites are $P = 27–40$ kbar and $T = 700–950°C$, and those of the epidote and garnet amphibolites are 8.6 kbar, 500°C; 12 kbar, 700°C; and 12–14 kbar, 700–815°C (Korsakov et al., 1998; Masago, 2000). The evaluated parameters of the metamorphic peak of the gneisses and schists occurring in association with the eclogites, as well as the Barchi-Kol calc-silicate rocks, are $P > 40$ kbar at $T = 950–1000°C$ (diamond-bearing rocks) and $P = 29$ kbar at $T = 800–900°C$ (coesite-bearing varieties). The retrograde alterations of the rocks during the exhumation of the UHP complexes to lower–intermediate depth levels occurred at 10–12 kbar and 650–800°C and corresponded to a number of transformation episodes at granulite- and amphibolite-facies parameters (Hermann et al., 2001; Korsakov et al., 2002; Stepanov et al., 2016).

The average $^{206}Pb/^{238}U$ zircon age of the diamond-bearing garnet–biotite gneisses and calc-silicate rocks is 528 ± 3 Ma and corresponds to initial decompression after UHP metamorphism and likely to the associated dehydration melting of the rocks (Hermann et al., 2001). Similar age estimates for the prograde, peak, and retrograde metamorphic episodes were obtained for zircons and monazites from the diamond- and coesite-bearing kyanite–garnet–mica schists and garnet–biotite gneisses: 528–521, 528–522, and 503–532 Ma, respectively (Stepanov et al., 2016). The core of a zircon crystal from the schist yielded a Neoarchean age of 2867 ± 72 Ma, which indicates that early Precambrian complexes were involved in processes that produced the protoliths of the Barchi-Kol UHP rocks (Stepanov et al., 2016). The chemical compositions of the rocks suggest that they were partly molten at $T \sim 1000°C$ during initial exhumation (Stepanov et al., 2014).

The ***Kumdy-Kol locality*** (Fig. 2) lies east of the Barchi-Kol one and is characterized by widespread ultrahigh-pressure diamond- and coesite-bearing rocks, which were attributed to a deposit of metamorphic diamond (Lavrova et al., 1999). In the southwest, the ultrahigh-pressure rocks are tectonically overlain with quartzites of the Kokchetav Group. Numerous publications discuss the environments and parameters at which the rocks were formed, the sources and compositions of their protoliths, and the mechanisms responsible for the involvement of the crustal complexes in subduction processes, which brought these rocks to depths of approximately 120 km and deeper and resulted in diamond crystallization in them (Letnikov et al., 1983; Sobolev and Shatsky, 1990; Schertl and Sobolev, 2013). The UHP rocks are mostly various metasedimentary gneisses, schists with eclogite bodies, and calc-silicate rocks (Shatsky et al., 1995; Schertl and Sobolev, 2013). The metapelites are garnet–biotite, garnet–chlorite, and garnet–zoisite rocks containing discernible concentrations of tourmaline (~5 vol %), muscovite, and relict phengite, kyanite, and amphibole. The garnet and amphibole of these rocks contain microinclusions of diamond, clinopyroxene and coesite or quartz pseudomorphs after it (Sobolev et al., 1991; Zhang et al., 1997; Katayama et al., 2000). The gneisses are sometimes migmatized and host granitic bodies and veins, which suggests that the UHP rocks have undergone partial melting (Shatsky et al., 1999; Ragozin et al., 2009). The eclogites occur as blocks and tectonic lenses among variably altered and mylonitized gneisses and schists and consist of the typical assemblage of garnet, albite–augite symplectites with a high K_2O concentration (up to 1 wt %) in the clinopyroxene (the symplectites developed after omphacite), rutile, quartz, and occasional kyanite and zoisite. The garnet grains contain palisade quartz inclusions, which were interpreted as pseudomorphs after coesite (Shatsky et al., 1995; Zhang et al., 1997). Garnet and zircon from the calc-silicate rocks, which consist of garnet, K-feldspar, calcite, and clinopyroxene, host diamond microinclusions (Zhang et al., 1997). Some of the calc-silicate rocks contain diamond inclusions (these are dolomite marbles with diopside, garnet, and phlogopite), while others do not (dolomite marbles with forsterite, diopside, and Ti-clinohumite) (Shatsky et al., 1995; Ogasawara, 2000; Katayama et al., 2000). The Kumdy-Kol locality is noted for the presence of garnet–clinopyroxene–quartz (~40 vol % quartz) and tourmaline–K-feldspar–quartz rocks whose garnet and tourmaline, respectively, host microdiamond inclusions (Zhang et al., 1997; Shimizu and Ogasawara, 2013; Musiyachenko et al., 2019). Garnet peridotites with Ti-clinohumite were also found only at this locality (Zhang et al., 1997; Reverdatto and Selyatitskiy, 2005).

The Kumdy-Kol ultrahigh-pressure rocks were formed at the metamorphic peak under eclogite-facies parameters, but information on the prograde $P–T$ path of their protoliths was almost completely obliterated during the later multiple metamorphic episodes. The $P–T$ evolution of the eclogites involved episodes of prograde metamorphism, whose parameters calculated from the composition of mineral inclusions in the garnet correspond to the epidote-amphibolite facies (Zhang et al., 1997). The peak parameters of UHP metamorphism of the garnet–biotite gneisses, metapelites, and dolomite marbles were $P > 40–60$ kbar and $T = 800–1000°C$ (Shatsky et al., 1995; Zhang et al., 1997; Ogasawara, 2000; Katayama et al., 2001). The $P–T$ parameters of the coesite-bearing eclogites were estimated at 30 kbar and 780°C. At the same time, the peak metamorphic parameters of the assemblages of the eclogites, garnet–calcite–clinopyroxene and garnet–clinopyroxene–quartz rocks at Kumdy-Kol were estimated at $P > 40$ kbar, $T = 920–1050°C$ in (Shatsky et al., 1995). Similar peak metamorphic temperatures of >900°C were calculated for the eclogites in (Skuzovatov et al., 2021). The garnet peridotites with Ti-clinohumite were formed at the metamorphic peak at $T > 740°C$ and $P > 30$ kbar (Zhang et al., 1997). However, it was hypothesized (Reverdatto et al., 2008) that the ultramafic rocks were metamorphosed under ultrahigh pressures similar to those estimated for the host diamond-bearing gneisses. The peak metamorphic pressure of the eclogites and garnet peridotites was estimated, with regard for the K_2O concentration in the clinopyroxene, at as high as 60 kbar at $T \sim 950°C$ (Okamoto et al., 2000). The retrogression of the Kumdy-Kol UHP metamorphic rocks was related mostly to their exhumation and associated decompression. The $P–T$ parameters of retrograde metamorphism of the calc-silicate rocks were estimated at 25 kbar, 800°C and <15 kbar, 790°C (Zhang et al., 1997; Ogasawara, 2000). The diamond-bearing gneisses, schists, and eclogites were retrogressed at parameters of the granulite (740–790°C, ~10 kbar), amphibolite (650–715°C, ~10 kbar), and greenschist (350–420°C) facies (Zhang et al., 1997; Katayama et al., 2001).

Jagoutz et al., Claoue-Long et al., and Shatsky et al. were among the first to date the peak of the ultrahigh-pressure metamorphism at the Kumdy-Kol locality (Jagoutz et al., 1989; Claoue-Long et al., 1991; Shatsky et al., 1993). The Sm–Nd garnet and clinopyroxene mineral isochrons correspond to ages of 533 ± 20 and 528 ± 7 Ma, and the model age of the protoliths of these rocks is 870 Ma (Shatsky et al., 1993, 2018). The $^{206}Pb/^{238}U$ zircon age of 530 ± 7 Ma of the metamorphic peak of the diamond-bearing garnet–biotite gneisses is comparable to the age of the eclogites (Claoue-Long et al., 1991). The metamorphic zircons often contain older cores, which were dated at ~560–627, 694–767, 906–1003, and 1952–1981 Ma (Claoue-Long et al., 1991), which may indicate that the protoliths of the diamond-bearing garnet–biotite gneisses were of sedimentary origin. Further geochronological studies resulted in a more accurate Sm–Nd mineral isochron (garnet and clinopyroxene) for the eclogites, which corresponds to 535 ± 3 Ma. The time span when the diamond-bearing rocks resided under ultrahigh pressures was estimated at 537–524 Ma (Shatsky et al., 1999). The Kumdy-Kol eclogites were the first from which zircons were separated whose age spans a wide range of 533–459 Ma and which characterize the age of the eclogite-facies peak metamorphism and later retrograde episodes of lower metamorphic grade (Skuzovatov et al., 2021). A concordant age of 508 ± 4 Ma was obtained for the most abundant zircon population from the eclogites and pertains to the exhumation of the ultrahigh-pressure rocks to lower crustal levels at parameters of the granulite and amphibolite facies, whereas the Hf model age of zircons corresponds to 1.02–0.79 Ga (Skuzovatov et al., 2021). The Nd model ages of the protoliths of the diamond-bearing gneisses and schists are 2.3–2.2 Ga (Shatsky et al., 1999) and are comparable to the model ages of the Mesoproterozoic orthogneisses and gneiss-granites of the Zerendy Group of the Kokchetav massif (Glorie et al., 2015). The $^{40}Ar–^{39}Ar$ plateau ages of retrograde muscovite and biotite from the Kumdy-Kol diamond-bearing gneiss are 517 ± 5 and 515 ± 3 Ma, respectively, and were interpreted as the exhumation timing of the ultrahigh-pressure rocks to mid-crustal depth levels. The metamorphic rocks were likely exhumed virtually immediately after their protoliths were subducted to depths of ~140 km at 537– 517 Ma (Shatsky et al., 1999). The $^{40}Ar–^{39}Ar$ plateau ages presented in (Hacker et al., 2003) are as follows: ~529 Ma for micas from the diamond-bearing gneisses (which corresponds to the exhumation of the UHP rocks) and ~509 Ma (overprinted metamorphism to the amphibolite and greenschist facies). The morphology of zircons from the diamond-bearing garnet–mica gneisses with kyanite relics provide evidence of multiple metamorphic episodes in the evolution of the rocks (Katayama et al., 2001). The metamorphic zircons usually contain cores with microinclusions of diamond, coesite, omphacite, and kyanite and rims that host inclusions of minerals of lower metamorphic grades (chlorite, plagioclase, and quartz) (Dobrzhinetskaya, 2012). Such a distribution of mineral inclusions indicates that the zircon cores grew under UHP conditions. The growth of the rims was associated with the capture of mineral inclusions stable under much lower $P–T$ parameters and likely occurred when the rocks were brought to shallower mid-crustal depths. The $^{206}Pb/^{238}U$ ages of the cores and rims of the zircons (their inner and outer zones) are 537 ± 9, 507 ± 8, and 456–461 Ma, respectively. Occasional zircon grains contain older cores, which were dated at 1.3–1.4 Ga (Katayama et al., 2001). The Kumdy-Kol diamond-bearing gneisses host bodies and veins, which were reportedly produced by the local partial melting of the UHP rocks during their initial exhumation (Shatsky et al., 1999; Ragozin et al., 2009). The average estimated $^{206}Pb/^{238}U$ age value of 526 ± 2 Ma was obtained from zircon from these granitoids and corresponds to the migmatization of the diamond-bearing gneisses, a process that marked the onset of decompression when the UHP rocks were brought to mid-crustal depth levels. One of the zircon cores yielded an age of 1.1 Ga (Ragozin et al., 2009). Another age estimate for these granitoids is 512 ± 5 Ma (Borisova et al., 1995) and may correspond to another migmatization episode. The average $^{206}Pb/^{238}U$ age of the garnet peridotites with Ti-clinohumite is 528 Ma and corresponds to the age of the ultrahigh-pressure metamorphism (Katayama et al., 2003). Zircon crystallized in the ultramafic rocks likely when they were metasomatized and enriched in HFSE elements, which were brought by the fluid with mantle complexes, under parameters of the eclogite facies (Katayama et al., 2003). The ^{40}Ar–^{39}Ar plateau age of tourmaline from the Kumdy-Kol tourmaline–K-feldspar–quartz rocks is ~492 Ma (Korsakov et al., 2009). This age likely corresponds to the timing of deformations during the final exhumation of the ultrahigh-pressure rocks to mid-crustal depth levels.

Eastern Domain (UHP–HP and MP Metamorphic Rocks)

This domain east of the Chaglinka fault is made up of rocks of the Zerendy Group metamorphosed to the amphibolite facies, medium and ultrahigh- to high-pressure (MP and UHP–HP, respectively) rocks, and low-pressure (LP) schists of the Daulet Formation. The rocks metamorphosed to various grades make up discrete nappes, whose structural settings are disputable.

Some researchers are prone to believe that the lowest structural position is assigned to rocks of the Zerendy Group metamorphosed to the amphibolite facies, and they are overlain by nappes consisting of MP and UHP–HP rocks. Schists of the Daulet Formation, which are overlain (without structural unconformity) by quartzite–schist sequences of the Kokchetav Group, have the uppermost position relatively to the Zerendy Group (Dobretsov et al., 1998; Theunissen et al., 2000). Other scientists think that rocks of the Daulet Formation tectonically underlie the high-pressure complexes of the Zerendy Group, and its metamorphism was induced by contact effects of the UHP and HP rocks when they were brought to middle and upper crustal depths. Structurally above the Daulet Formation, nappes of high- and ultrahigh-pressure rocks and medium-grade amphibolite-facies rocks occur (Terabayashi et al., 2002; Kaneko et al., 2000; Buslov et al., 2010; Theunissen et al., 2002).

The ***Kulet*** and ***Soldat-Kol localities*** east of the Chaglinka fault (Fig. 2) host widespread garnet–mica schists and paragneisses, phengite eclogites, amphibole–garnet–zoisite rocks, and talc–garnet–kyanite–phengite schists with coesite or with quartz pseudomorphs after coesite found as inclusions in garnet and kyanite (Shatsky et al., 1993, 1998; Ota et al., 2000; Parkinson et al., 2000; Masago et al., 2009; Zhang et al., 2012). Orthogneisses and amphibolites metamorphosed to medium grades occur more rarely. It should be mentioned that the coesite-bearing talc schists are typical only of the Kulet and Soldat-Kol localities. Eclogites with retrograde alterations are found as blocks and lens-shaped bodies up to 600 m among garnet metapelites and do not contain coesite inclusions. In the southern part of this locality, UHP rocks occur in tectonic contact with andalusite–cordierite and sillimanite–cordierite schists of the Daulet Formation (Dobretsov et al., 1995, 1998; Parkinson et al., 2000).

The peak metamorphic parameters of the coesite-bearing garnet–mica (±kyanite) and talc–garnet–kyanite–phengite schists corresponded to the eclogite facies: P = 34–36 kbar and T = 720–760°C (Shatsky et al., 1998; Parkinson et al., 2000; Masago et al., 2009). The peak metamorphic pressures of the coesite-bearing talc–garnet schists were evaluated in (Zhang et al., 1997) at 28–35 kbar at temperature values close to those quoted above. The zoning of the garnet grains and the distribution of inclusions in them were used to evaluate the P–T parameters at which the mineral assemblages were formed at the prograde metamorphism: 380–580°C at <10 kbar (Parkinson et al., 2000) and 450–620°C at 8–15 kbar (Zhang et al., 1997). Decompression associated with the exhumation of the UHP rocks first resulted in coesite transition into quartz at P < 26 kbar and the subsequent crystallization of lower temperature assemblages of biotite, muscovite, and chlorite under amphibolite-facies parameters at P ~8 kbar and T = 600°C (Parkinson et al., 2000). Some of the Kulet eclogites possess coronitic and granoblastic textures. The latter contain mineral assemblages corresponding to the peak of the ultrahigh-pressure metamorphism (garnet, omphacite, quartz, rutile, ± phengite), whereas the coronitic eclogites were produced at the transition from the amphibolite to eclogite facies during the prograde evolution of the rocks at T < 500°C, P < 12 kbar and then at T = 500–550°C, P = 19–24.5 kbar (Zhang et al., 2012). At the metamorphic peak, the granoblastic eclogites and amphibole–garnet–zoisite rocks were likely formed within the stability field of coesite, as also were the host metapelites, at 27–35 kbar, 560–720°C. The rocks underwent retrograde transformations at lower metamorphic grades in the course of exhumation, up to the development of garnet amphibolites at 7–13 kbar, 540–720°C (Ota et al., 2000; Zhang et al., 2012).

The Sm–Nd isochron of the Kulet amphibole–garnet–zoisite rocks corresponds to an age of 522 ± 5 Ma, which was the age of the ultrahigh-pressure metamorphism (Shatsky et al., 1993). The ^{40}Ar–^{39}Ar plateau ages of the garnet–mica schists are ~499 Ma (on muscovite) and ~505 Ma (on biotite) and correspond to the retrograde metamorphism to the amphibolite and greenschist facies (Hacker et al., 2003). Other authors (Theunissen et al., 2002) report ^{40}Ar–^{39}Ar mica ages of similar rocks of ~519 Ma (phengite) and ~521 Ma (biotite), with these age values pertaining to the initial exhumation of the UHP rocks. Zircons from the Kulet eclogites can be grouped into two morphological types. One of them comprises zoned zircon grains with high Th/U ratios (1.0–1.5) in the cores and high U concentrations in the marginal zones. The zircons of the other type are much more homogeneous and possess Th/U = 0.1–0.2. The average $^{207}Pb/^{206}Pb$ age of the cores of the type-1 zircons is 1421 ± 13 and corresponds to the age of the protolith. The average $^{206}Pb/^{238}U$ ages of the two populations of the type-2 zircons is 532 ± 58 and 497 ± 5 Ma and reflect the peak and retrograde metamorphism of the rocks (the retrogression was related to the exhumation of the eclogites to mid- and upper crustal levels) (Zhang et al., 2016). Rutile from the talc–garnet–kyanite–phengite schists yielded two age values of 526 ± 9 and 498 ± 11 Ma, which correspond to two retrogression episodes during exhumation (Zhang et al., 2016).

The ***Sulu-Tobe locality*** south of the Kulet one (Fig. 2) is made up of similar metamorphic complexes. However, eclogites and zoisite amphibolites developing after the eclogites at the Sulu-Tobe locality quantitatively dominate over the metapelites and orthogneisses and compose the largest (1×2 km^2) body within the Kokchetav massif (Kaneko et al., 2000). The *P-T* parameters of the eclogites metamorphism were estimated at 600–860°C at 14–16.5 kbar (Dobretsov et al., 1995, 2006; Shatsky et al., 1993). The $^{40}Ar–^{39}Ar$ plateau amphibole age of the zoisite amphibolite is 512 ± 0.9 Ma (Hacker, 2003). The model ages of the Sulu-Tobe eclogites are 1.95–0.67 Ga (Shatsky et al., 2018). The eclogites and garnet amphibolites after eclogites, schists, and gneisses at the ***Borovoe*** and ***Chaglinka localities*** are generally similar to the metamorphic rocks at the Sulu-Tobe, Kulet, and Soldat-Kol localities (Kaneko et al., 2000) (Fig. 2). The Chaglinka eclogites were formed at the metamorphic peak at temperatures of 620 to 740°C (Shatsky et al., 1993), and the peak conditions for the Borovoe eclogites are estimated at 750–800°C, 17–18 kbar (Zhimulev et al., 2010, 2011). The peak parameters of metamorphism of eclogites at the ***Chaikino locality*** were evaluated at 800–850°C and 18–20 kbar (Zhimulev et al., 2010, 2011). The HP rocks were retrogressed during several episodes in the course of their decompression at exhumation to mid- to upper crustal depths at 760–790°C and 11–12 kbar; 700–730°C and 7–8 kbar; and 570–600°C and 5–6 kbar. The $^{40}Ar–^{39}Ar$ muscovite age of the mica schists hosting the eclogites at the Borovoe locality is 493 ± 5 Ma and corresponds to the cooling of the metamorphic rocks at their retrograde evolution (Zhimulev et al., 2010). The variably mylonitized gneisses at these localities contain at least two zircon populations. One of them yielded Mesoproterozoic $^{207}Pb/^{206}Pb$ age values of 1113 ± 24 and 1130 ± 18 Ma (Chaikino) and 1111 ± 42 and 1137 ± 36 Ma (Borovoe), and the average $^{206}Pb/^{238}U$ ages of the other population are 537 ± 7 and 484 ± 10 Ma (Chaikino) and 524 ± 5 and 490 ± 9 Ma (Borovoe) (Glorie et al., 2015). The Mesoproterozoic age values seem to correspond to the protolith age of gneisses at the Chaglinka and Borovoe localities, and the Early Paleozoic ages indicate the timing of ultrahigh-pressure metamorphism and subsequent retrograde alterations related to the exhumation of the high- and ultrahigh-pressure rocks. The Lu–Hf model ages of the Chaglinka gneisses are 2.65–2.48 Ga at $\varepsilon_{Hf}(T) = -11.5$ to -8.4, which suggests that the protoliths were formed at the reworking of complexes of the Neoarchean continental crust (Glorie et al., 2015). The ***Enbek-Berlyk locality*** lies between Kulet and Sulu-Tobe (Fig. 2) and is made up mostly of metamorphic rocks of medium grades: aluminous fine-grained garnet–kyanite–sillimanite–biotite schists that host bodies of garnet amphibolites and coronites (garnet amphibolites with garnet coronas around plagioclase and at contacts with pyroxene). The sillimanite replaces kyanite and was likely formed at retrograde metamorphism. In the northern part of this locality, intensely mylonitized medium-pressure rocks host eclogite bodies (Dobretsov et al., 1995; De Grave et al., 2006). The *P–T* parameters of the Enbek-Berlyk garnet–kyanite–sillimanite schists are 600–700°C and 4–7 kbar (Dobretsov et al., 2006; De Grave et al., 2006). The $^{40}Ar–^{39}Ar$ mica plateau ages lie within the range of ~476–496 Ma (Dobretsov et al., 2006). The time span of 476–496 Ma is interpreted as that of a collision tectonic event related to the transition from continental subduction to the collision of a microcontinent and island arc after ultrahigh-pressure metamorphism (De Grave et al., 2006). Furthermore, some $^{40}Ar–^{39}Ar$ biotite age values close to 447 Ma for mylonitized schists or blastomylonites from the Enbek-Berlyk locality may correspond to the age of younger strike-slip faulting (Theunissen et al., 2000; Dobretsov et al., 2006).

Moreover, "spinel harzburgites" found among schists and quartzites in the southern part of this locality occur in association with orthopyroxenites and anthophyllite schists, whose peak metamorphic parameters were 14–15 kbar, 780–840°C (Reverdatto and Selyatitskiy, 2005).

The rocks of the Zerendy Group in the western domain of the Kokchetav massif are characterized by much higher metamorphic grades and were metamorphosed under ultrahigh pressures, within the diamond stability field at $P > 40$ kbar, $T = 800–1000$°C (at the Barchi-Kol and Kumdy-Kol localities). Rocks of similar composition in the eastern domain show the highest peak metamorphic parameters corresponding to the coesite stability field: 34–36 kbar, 720–760°C (Kulet and Soldat-Kol localities). The rocks of the eastern domain locally host metamorphic rocks of lower (including medium) grades (at the Sulu-Tobe, Borovoe, Chaglinka, Chaikino, and Enbek-Berlyk localities). Although the metamorphic *P–T* parameters of the rocks are different, the ages of high- and ultrahigh-pressure rocks in both domains are similar at ca. 530 Ma.

Daulet Formation

The rocks of the Daulet Formation are constrained to the southern boundary of the field of high- and ultrahigh-pressure complex of the Zerendy Group and show tectonic contacts with the latter (Fig. 2). The andalusite–biotite–cordierite–feldspar and sillimanite (±garnet) metapelites of the Daulet Formation were formed at $T = 500–680°C$ and $P = 2–3$ kbar (Kaneko et al., 2000; Terabayashi et al., 2002).

Zircons from metapelites of the Daulet Formation sampled at the Sulu-Tobe locality (Fig. 2) are zoned and have magmatic cores with oscillatory zoning and outer rims, which likely grew during overprinted metamorphic processes (Katayama et al., 2001). An age estimate for a zircon core is 1280 Ma, two age values for the intermediate zones are 1138–1143 Ma, and the rims were dated at 461–516 Ma. The ages of the rims of zircon from metapelites of the Daulet Formation are comparable to age estimates for the rims of zircon from the ultrahigh-pressure diamond-bearing garnet–mica gneisses with kyanite relics from the Kumdy-Kol locality (507 and 456–461 Ma) (Katayama et al., 2001). The $^{40}Ar–^{39}Ar$ biotite age of schists of the Daulet Formation sampled far away from the Late Ordovician and Early Devonian granitoid massifs were estimated in (Buslov et al., 2010). The oldest age value of 515 ± 5 Ma corresponds to the higher grade metamorphic episode in rocks of the Daulet Formation, and ages close to 480 Ma pertain to a younger metamorphic episode, when andalusite and cordierite were formed (Buslov et al., 2010).

ISSYK-KUL (NORTH TIEN SHAN) MASSIF

The Issyk-Kul massif composes most of the North Tien Shan. Metamorphic rocks are most widely spread in the westernmost part of the massif, where they are spatially constrained to the core of the large Makbal antiform. Its flank parts host fragments of ophiolites, siliceous–basalt and volcano-sedimentary sequences, and olistostrome complexes, which make up tectonic slice packages (Degtyarev et al., 2013) (Fig. 3).

The core of the Makbal antiform consists of intensely deformed and variably metamorphosed Precambrian quartzite–carbonate–schist, schist, and terrigenous–carbonate rock sequences. The complexes of the core of the Makbal antiform are cut across by massifs of Precambrian and Paleozoic granitoids: Mesoproterozoic (1100–1130 Ma) granites of the Karadzhilga complex in the southeastern part of the antiform, Early Cambrian (510–515 Ma) granitic rocks of the Kandzhailyau complex in the central part, and the Middle–Late Ordovician (455–460 Ma) granitic rocks of the Almalyksai complex in the northeastern and southwestern flanks (Apayarov, 2009; Degtyarev et al., 2011; Kröner et al., 2013; Degtyarev et al., 2013; Konopelko et al., 2012).

The Precambrian (mostly metasedimentary) sequences that make up the core of the Makbal antiform are subdivided into the Sharkyrak (upper) and Akdzhon (lower) groups, which differ from one another in the metamorphic grades of their rocks (Bakirov, 2017; Kasymbekov et al., 2020). The bottom part of the Sharkyrak Group consists of dolomite marbles with beds of muscovite–quartz and carbonate schists, and calciphyres (Achiktash complex), and the top of the vertical section is composed of muscovite–chlorite–quartz schists with rare quartzite beds (Kainda complex) (Degtyarev et al., 2013; Bakirov, 2017). The metamorphic grade of the rocks of the Sharkyrak Group is no higher than the greenschist facies. The age of the Sharkyrak Group is currently thought to be older than the Late Mesoproterozoic because these rocks are cut across by granites of the Karadzhilga complex. Structurally above the Kainda complex, quartzites of the Ovva Formation (latest Mesoproterozoic–earliest Neoproterozoic) occur, which contain zircons older than 1030 Ma (Alekseiev et al., 2020). The Akdzhon Group is made up of metasedimentary and less abundant metamagmatic rocks, which were formed at metamorphic peak parameters, including high and ultrahigh pressures (Bakirov, 1978; Demina et al., 2005). The metamorphic rocks of the Akdzhon and Sharkyrak groups occur in tectonic contacts with one another and compose tectonic slices, which were combined when the rocks were exhumed from various depths. The rocks vary in mineralogy and metamorphic grades (Tagiri et al., 2010; Konopelko et al., 2012). The Akdzhon Group consists of the structurally lower Makbal complex, which is mainly composed of quartzites, and the upper Neldy complex, which is dominated by metapelites. Both complexes host variably sized bodies of paragonite-bearing eclogites and garnet amphibolites after eclogites, but the complexes show different metamorphic grades.

The ***Makbal complex*** consists of variably sheared quartzites, whose detrital zircons were dated within the range of 1600–3780 Ma, with a maximum within the range of 1840–2000 Ma (Rojas-Agramonte et al., 2014; Konopelko et al., 2016; Alekseev et al., 2020). Many of the analyzed zircons showed relics of magmatic zoning but no rims providing a record of overprinted processes (including metamorphic ones). The quartzites host nappes consisting of intercalating coarse-grained coesite-bearing garnet–chloritoid–talc schists and quartzite-schists with garnet porphyroblasts, which also contain coesite relics, with boudins of variably altered (amphibolized) eclogites and “glaucophanites” (eclogites with high modal contents of glaucophane). The peak metamorphic parameters

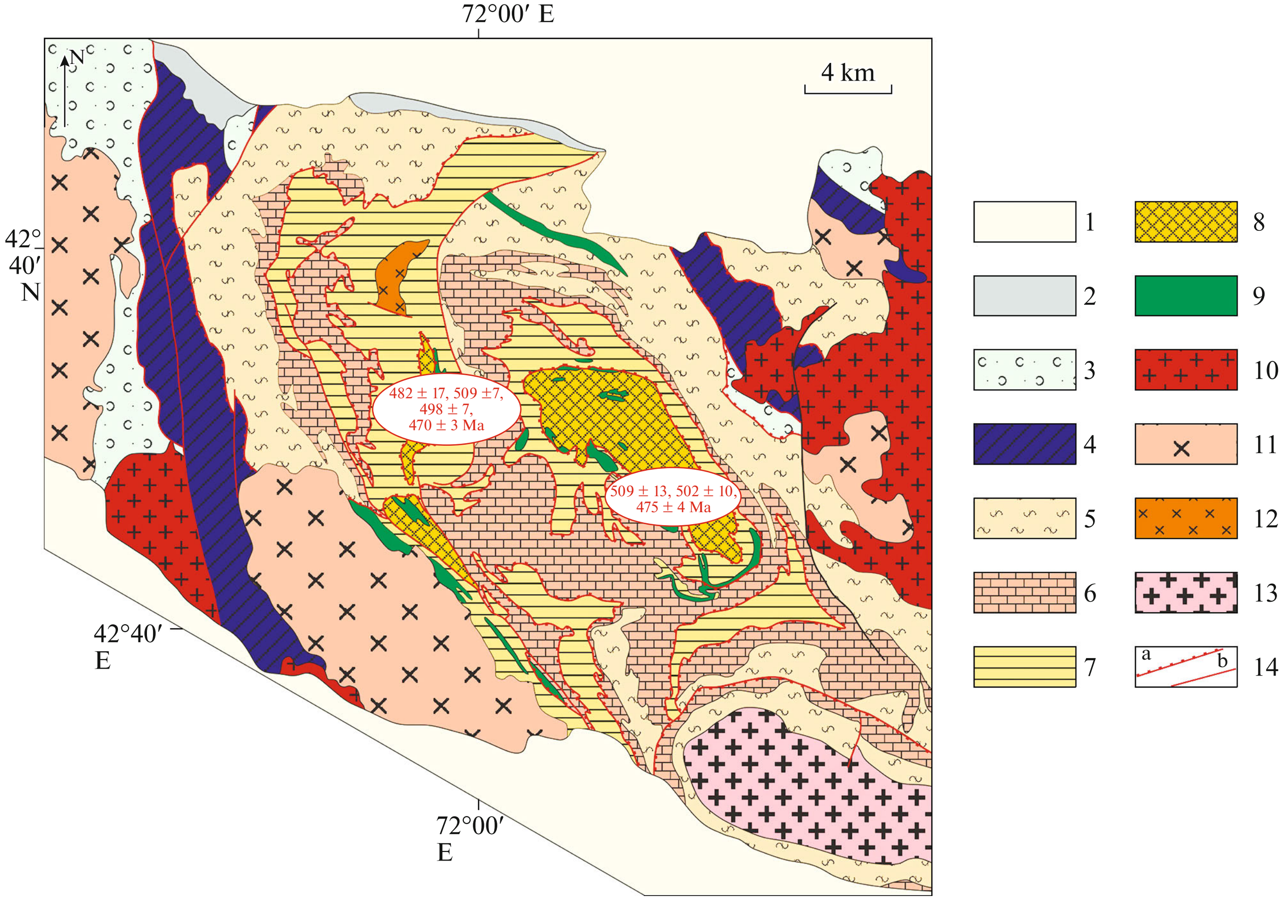

Fig. 3. Schematic geological map of the Makbal antiform and its surroundings (western termination of the Kyrgyz Range) (prepared using data from Apayarov et al., 2006; Degtyarev et al., 2013; Bakirov, 2017).
(1) Cenozoic sequences; (2) Early and Middle Carboniferous terrigenous–carbonate sequences; (3) Early–Middle Ordovician siliceous–tuff sequences; (4) Late Cambrian ultramafic–gabbro and siliceous–basalt complexes; (5–9) metamorphic rocks of the Issyk-Kul massif: (5) muscovite–chlorite schists and quartzites of the Kainda complex, (6) marbles and marmorized limestones with quartzite beds of the Chymynsai complex, (7) garnet–chlorite–muscovite and carbonaceous schists of the Neldy complex, (8) quartzites with beds of garnet–chlorite–talc schists and marbles of the Makbal complex; (9) largest bodies and lenses of amphibolites, garnet amphibolites, and amphibolized eclogites; (10–13) granitoid complexes: (10) Early–Middle Devonian Alamin complex, (11) Middle–Late Ordovician Almalysai complex, (12) Middle Cambrian Kandzhailyau complex, (13) Mesoproterozoic Karadzhilga complex; (14) faults: (a) boundaries of tectonic nappes, (b) others. The map shows principal age estimates of the near-peak high- and ultrahigh-pressure metamorphism of rocks from the Issyk-Kul massif.

of the rocks likely corresponded to ultrahigh pressures (Tagiri et al., 2010; Konopelko et al., 2012; Meyer et al., 2013). In two instances, garnet porphyroblasts in eclogites of the Makbal complex contained quartz pseudomorphs after coesite (Tagiri et al., 2010), and the garnet–chloritoid–talc schists typically contain clinozoisite–kyanite–quartz aggregates, which were formed at the decomposition of lawsonite (Orozbaev et al., 2015).

The ***Neldy complex*** is dominated by garnet–phengite–biotite schists with bodies of calc-silicate rocks, eclogites and garnet amphibolites after eclogites, and subordinate amounts of garnet-free mica schists, marbles, schists with relics of garnet and chloritoid, and quartzites (Tagiri et al., 2010; Togonbaeva et al., 2010b; Kasymbekov et al., 2020). Detrital zircons from the garnet–mica schists are variably rounded grains with both magmatic zoning and structures of metamorphic origin. The age range of the zircons is 2500–1800 Ma (Degtyarev et al., 2013). The metasedimentary rocks of the Akdzhon Group host rare lenses and bodies of winchite schists (Tagiri et al., 2010).

The presence of coesite among the relict minerals of some metasedimentary rocks of the Makbal complex suggests that their peak metamorphic pressure was higher than 24 kbar (Tagiri and Bakirov, 1990). The P–T prograde–peak metamorphic parameters of the garnet–chloritoid–talc schists were estimated at 25.5 kbar and 480°C, the peak parameters are 28–28.5 kbar and 525–560°C, and the retrograde parameters are 24 kbar and 580°C (Tagiri et al., 2010; Meyer et al., 2014). The development of clinozoisite–kyanite–quartz aggregates replacing lawsonite in the schists marks isothermal decompression at 16–20 kbar, 510–580°C, and the peak metamorphic parameters in the lawsonite stability field corresponded to 33 kbar, 530–580°C (Orozbaev et al., 2015). Eclogites *sensu stricto* are very rarely preserved in the rocks, and the high-pressure metabasites are mostly garnet amphibolites or “galucophanites” with relics of minerals of the earlier eclogite assemblage (Rojas-Agramonte et al., 2013; Klemd et al., 2015). The *P–T* parameters of the strongly modified eclogites and “glaucophanites” of the Makbal complex are 18 kbar, 480°C for prograde metamorphism and 20–25 kbar, 525–560°C (Tagiri et al., 1995, 2010; Meyer et al., 2013) near the metamorphic peak. The *P–T* parameters of the later metamorphic episodes, during which the Na- and Ca–Na amphiboles grew, were 8–13 kbar, 300–500°C (Tagiri et al., 2010). Note that the occurrence of garnet-hosted inclusions of two quartz pseudomorphs after coesite found in eclogites of the Makbal complex (Tagiri et al., 2010) may indicate a peak metamorphic pressure of $P > 24$ kbar. The intensely altered eclogites and garnet amphibolites of the Neldy complex were produced at the peak of metamorphism at 22–25 kbar, 550–610°C (Togonbaeva et al., 2010b) or at 14 kbar, 620°C (Rojas-Agramonte et al., 2013). Garnet porphyroblasts in the eclogites host numerous mineral inclusions of Na- and Ca–Na amphiboles, clinopyroxene, epidote, plagioclase, and paragonite, whose distribution is consistent with the prograde zoning of the garnet grains. The *P–T* parameters of the prograde metamorphism of eclogites in the Neldy complex were estimated at $P > 8$ kbar, $T = 410–490$°C (Togonbaeva et al., 2010b). Retrogression of the rocks induced their pervasive amphibolization at 6.5–12 kbar, 430–630°C (Togonbaeva et al., 2010b; Rojas-Agramonte et al., 2013). The *P–T* parameters of the garnet–phengite–biotite schists were estimated at 430–630°C and 6.5–12 kbar, which corresponds to medium-grade metamorphism (Tagiri et al., 2010). High metamorphic pressures for the same rocks were suggested in (Meyer et al., 2014) based on simulation results and some petrographic features of the rocks, and this is partly consistent with data in (Kasymbekov et al., 2020) on compositionally similar rocks (9–17 kbar, <630°C). The calc-silicate rocks with feldspars, diopside, phlogopite, phengite, epidote, and quartz were metamorphosed at peak parameters of 11–13 kbar and 600°C (Tagiri et al., 2010). The peak *P–T* parameters of the schists with garnet and chloritoid relics were 12–15 kbar, 485–545°C, and these parameters for their retrograde metamorphism were >3 kbar, ~500°C (Kasymbekov et al., 2020).

One of the first age estimates for (ultra)high-pressure metamorphism in the Issyk-Kul massif was obtained on paragonite from eclogite sampled in the Makbal complex, which has a K–Ar age of 482 ± 17 Ma (Tagiri et al., 1995). This estimate is generally close to the later U–Pb, Sm–Nd, and Lu–Hf zircon and omphacite age estimates for eclogites from both the Makbal and the Neldy complexes. The concordant $^{206}Pb/^{238}U$ ages of the outermost zones of zircon from two eclogite samples from the Makbal complex are 509 ± 7 and 498 ± 7 and are interpreted as the ages of the (U)HP metamorphism (Konopelko et al., 2012). The intercept between the concordia and discordia corresponds to an age of ~1446–1447 Ma, which may suggest that Mesoproterozoic complexes were involved in processes that produced the protoliths of the eclogites (Konopelko et al., 2012). The cores of zircons from the amphibolized eclogites also yielded concordant ages of ~820 and ~700 Ma, which suggest that the protoliths of the eclogites might have been younger than 700 Ma (Konopelko et al., 2016). A Lu–Hf isochron for garnet from the garnet amphibolite (retrogressed eclogite) from the Neldy complex corresponds to an age of 470 ± 3 Ma, which was interpreted as an age close to that of the peak of the high-pressure metamorphism (Rojas-Agramonte et al., 2013). When this age value is considered, it is necessary to bear in mind that the eclogite shows strong retrograde alterations (and is, in fact, garnet amphibolite), and hence this age estimate can be rejuvenated. A much older Sm–Nd age of the metamorphic peak (526 ± 10 Ma) was obtained for retrogressed eclogites form the Neldy complex (Togonbaeva et al., 2010a). The age of the outer rim of a zircon grain from amphibolite taken beyond the field of the (U)HP rocks of the Makbal complex corresponds to the range of ~465–453 Ma (Konopelko et al., 2016) and is comparable to the emplacement timing of granitoids of the Almalyk complex (Apayarov, 2009; Degtyarev et al., 2013). The $^{206}Pb/^{238}U$ zircon age (zircon rims) of UHP metamorphism of coesite-bearing garnet–chloritoid–talc schists of the Makbal complex is 502 ± 10 Ma, and the cores of the zircons show a broad scatter of their age values within the range of 2583–642 Ma, which likely suggests that the protolith of the UHP schists was of sedimentary origin (Konopelko et al., 2012, 2016). The K–Ar phengite age of the garnet–chloritoid–talc schists is 509 ± 13 Ma (Tagiri et al., 2010) and is consistent with the $^{206}Pb/^{238}U$ age estimate. The Sm–Nd isochron for the garnet corresponds to a younger age of 475 ± 4, which is interpreted as the timing of garnet growth during prograde–peak metamorphism (Meyer et al., 2014). The Sm–Nd characteristics of the rocks ($\varepsilon_{Nd} = -11$) suggest ancient crustal source for the parental melts of the protolith of the schists (Meyer et al., 2014). Similar age estimates (481 ± 26 and 480 ± 56 Ma) were obtained for accessory monazite from the ultrahigh-pressure schists and also pertain to the age of the ultrahigh-pressure metamorphism of these rocks (Togonbaeva et al., 2009). The K–Ar ages of schists with garnet and chloritoid relics and mica schists containing very little garnet (Neldy complex) are 474 ± 12 and 524 ± 13 Ma, respectively (Kasymbekov et al., 2020). The former age value is viewed as an underestimate and reflects the time when the rocks interacted with Ordovician granitoids, whereas the latter age value is interpreted as the age of the HP metamorphism and is well consistent with earlier age estimates of the eclogites and garnet–chloritoid–talc schists (Kasymbekov et al., 2020).

The Makbal and Neldy complexes are thus characterized by uneven metamorphism. The estimated *P–T* parameters of the metamorphic peak indicate that some rocks of the Makbal complexes were metamorphosed under ultrahigh pressures (UHP). The *P–T* parameters of similar rocks in the Neldy complex correspond to high-pressure (HP) metamorphism. Moreover, the Makbal complex includes slices of garnet-bearing quartzites with coesite relics, which were formed under eclogite-facies parameters, and quartzites with detrital zircon that show almost no evidence of high-grade metamorphic transformations (Rojas-Agramonte et al., 2014; Alexeiev et al., 2020).

CHU–KENDYKTAS MASSIF

The Chu–Kendyktas massif is spatially constrained to the southern part of the Paleozoides of Kazakhstan. Much of this massif is overlain by Middle and Late Paleozoic and Mesozoic–Cenozoic rocks, which divide the massif into a number of blocks that are made up mostly of Precambrian complexes. The medium- and high-grade metamorphic rocks occur in the Aktyuz block in the southeastern part of the massif, where these rocks belong to the Aktyuz and Kemin complexes (Fig. 4) (Bakirov, 2003).

The metamorphic rocks of the Aktyuz and Kemin complexes are tectonically overlain with fragments of slightly metamorphosed Early Cambrian (~530 Ma) ophiolites of the Kopurelisay complex: amphibolized gabbro, metabasalts, and dolerites (Bakirov et al., 2003; Kröner et al., 2012). These metamorphic complexes occur in tectonic relationships with the Cambrian and Early Ordovician volcano-sedimentary sequences occurring west and southwest of the Aktyuz block. Within a small area in the southwestern part of this block, the metamorphic rocks are overlain (with a significant unconformity) with terrigenous–carbonate rock sequence of Middle Ordovician age. The metamorphic rocks in the Aktyuz block are ubiquitously cut by large Early, Middle, and Late Paleozoic granitoid plutons (Fig. 4).

The Aktyuz complex is dominated by sheared tonalite–granodiorite gneisses, including those with garnet and phengite, and gneiss-granites. The parental melts of the protoliths of these rocks crystallized at 778 ± 6 and 834 ± 8 Ma (Kröner et al., 2012). The gneisses host volumetrically subordinate beds of marble and quartzite and bodies of garnet and epidote amphibolites ranging from 0.5 to 70 m in thickness and extending for up to 200 m. Their central parts contain preserved mineral assemblages of the eclogite facies. The Kemin complex is dominated by migmatized garnet-bearing paragneisses and schists and to a lesser extent gneiss-granites. The complex also includes intensely migmatized mafic rocks (gabbro-amphibolites) and paragneisses with quartzite, marble, and graphite schist beds (Bakirov et al., 2003). The age estimates for the protoliths of the gneiss-granites of the Kemin complex are 799 ± 6, 810 ± 10, 814 ± 5, and 844 ± 9 Ma (Kröner et al., 2012). The sedimentary protolith of the paragneisses was accumulated no earlier than at 500 Ma, and the eroded rocks belonged to Mesoproterozoic (~1100–1300 Ma) and older (up to 2460 Ma) complexes (Kröner et al., 2012; Rojas-Agramonte et al., 2014). The Sm–Nd and Lu–Hf isotope characteristics of the metamorphic rocks of the Aktyuz and Kemin complexes indicate that the protoliths were formed by the reworking of older crustal complexes dated at ~1.5–2.1 Ga (ε_{Nd} = –5 to –12). At the same time, the intensely altered eclogites have positive ε_{Nd} = +3.1 to +3.7 (Kröner et al., 2012; Klemd et al., 2014). The involvement of the material of older crustal complexes in the protoliths of the metamorphic rocks also follows from the age of a core of a zircon grain from the Neoproterozoic migmatite: 1180 ± 9 Ma (Kröner et al., 2012).

The eclogites of the Aktyuz complex are mostly transformed into garnet amphibolites, and the mineral assemblages (garnet, omphacite, and phengite) corresponding to the *P–T* parameters of the eclogite facies are occasionally preserved only in the central parts of the metabasite bodies (Klemd et al., 2014, 2015). The peak metamorphic parameters of the eclogites were estimated at 16–23 kbar, 550–670°C, and the retrograde assemblages were formed at 10–11.5 kbar and 730, 600–650, and 550–570°C (Orozbaev et al., 2007, 2010; Rojas-Agramonte et al., 2013; Klemd et al., 2014, 2015). Garnet grains in the eclogites often preserve relics of their prograde zoning and host mineral inclusions of glaucophane, Fe-staurolite, paragonite, and Mg-taramite, which are interpreted as the assemblage of the earlier prograde metamorphism to the epidote-amphibolite and blueschist facies at 4–10 kbar, 560–650°C and 8–16 kbar, 330–570°C (Takasu and Orozbaev, 2009; Orozbaev et al., 2010). The mineral assemblages of the garnet amphibolites developing after the eclogites were formed, during prograde and peak metamorphism, at 11–12 kbar, 600–640°C and 14–15 kbar, 675–735°C, respectively (Orozbaev et al., 2010). The *P–T* evolution of the Aktyuz eclogites involved prograde stages of metamorphism, when both the temperature and the pressure increased while the protoliths of the rocks were subducted to depths of >50 km, metamorphic peak, and retrograde transformations at isothermal decompression when the high-pressure eclogites were exhumed. Information on the metamorphic transformations of the host garnet-bearing gneisses is almost completely absent. It was mentioned in (Orozbaev et al., 2010) that the protoliths of the gneisses were also likely metamorphosed during a

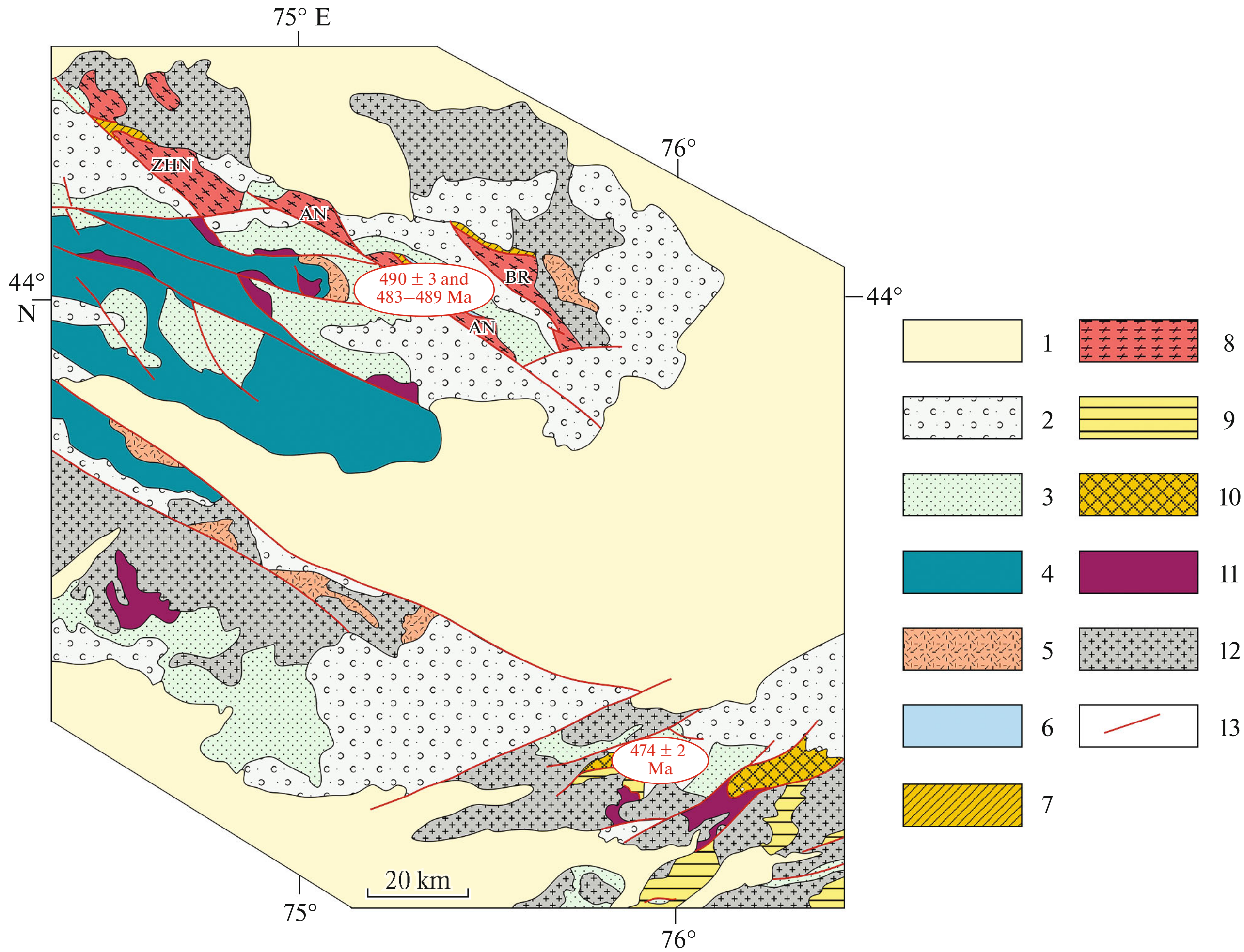

Fig. 4. High-pressure complexes in the Zheltau and southeastern part of the Chu-Kendyktas massifs.
(1) Cenozoic sequences; (2) Devonian and Carboniferous volcano-sedimentary sequences; (3) Ordovician and Early Silurian terrigenous–carbonate and flysch sequences; (4) Early Ordovician flysch sequences; (5) Late Neoproterozoic rhyolites, rhyolite tuffs, and basalts (Kopa Formation); (6–8) complexes of the Zheltau massif: (6) Early Cambrian granodiorites and granites, (7) Koyandy complex, (8) Anrakhai complex; (9, 10) complexes of the Aktyuz block of the Chu-Kendyktas massif: (9) Kemin complex, (10) Aktyuz complex; (11) Cambrian ophiolites; (12) Paleozoic granitoids; (13) faults. Blocks of the Zheltau massif: ZHN—Zhingeldy, AN—Anrakhai, BR—Burly. The map shows principal age estimates of near-peak high-pressure metamorphism of rocks from the Chu-Kendyktas massif.

prograde (10 kbar, 477°C) and peak (13–15 kbar, 635–745°C) episodes and were afterward retrogressed to lower metamorphic grades when exhumed.

The Lu–Hf garnet age of the intensely transformed Aktyuz eclogite is 474 ± 2 Ma and is interpreted as the age of high-pressure prograde to near-peak metamorphism (Rojas-Agramonte et al., 2013). The Sm–Nd age of the eclogite is 462 ± 7 Ma and corresponds to the exhumation of the high-pressure rocks to upper-crustal depths at $P < 10$ kbar, $T = 650$–600°C (Klemd et al., 2014). The ^{40}Ar–^{39}Ar phengite plateau age of the retrogressed eclogites is ~481 Ma and is viewed as averaged because of the excess radiogenic Ar (Klemd et al., 2014). The age estimate of ~749 Ma for the Aktyuz eclogites (Rb–Sr isochron for amphibole, rutile, omphacite, garnet, and the whole-rock) (Tagiri et al., 1995) indicates that the garnet and omphacite are not in equilibrium both with the rutile and the whole rock (Kroner et al., 2012).

ZHELTAU MASSIF

Southeast of the Chu-Ili Mountains in southern Kazakhstan, metamorphic complexes (including high-pressure ones) make up the Zheltau massif, which is situated northeast of the Chu–Kendyktas massif (Fig. 4). The massifs are separated by complexes of the Dzhalair–Naiman ophiolite zone, which consists of fragments of Early Cambrian ophiolites and basalt–rhyolite associations, Late Cambrian siliceous–basalts and tuff sequences, which are tectonically brought close to one another and are overlain by

Early Ordovician flysch and siliceous–terrigenous sequences (Ryazantsev et al., 2009; Degtyarev, 2012). Ordovician and Devonian volcano-sedimentary rocks divide the Zheltau massif into the Anrakhai, Zhingeldy, and Burly blocks (Pilitsyna et al., 2019a). The metamorphic rocks of the Zheltau massif occur in tectonic relations with intensely tectonized ophiolite fragments and unmetamorphosed Early Cambrian (~510 Ma) granitoids, quartz diorites, and granites, which were found exclusively in the Anrakhai block (Alexeiev et al., 2011; Pilitsyna et al., 2019a). The metamorphic rocks of the Zheltau massif are overlain, with a significant unconformity, by Early and Middle Ordovician terrigenous–carbonate sequences (Fig. 4).

According to their composition and metamorphic grades, the Zheltau metamorphic rocks are classified with the Anrakhai and Koyandy complexes. The rocks of the Anrakhai complex are dominant in the Zheltau massif. These are various orthogneisses, whose protoliths were dated at ~790 and 1840 Ma and were formed at the expense of recycled Early Precambrian crustal complexes (Kröner et al., 2007; Tretyakov et al., 2011; Pilitsyna et al., 2019a). The Anrakhai complex also contains widespread garnet and epidote amphibolites, which occur as bodies up to a few hundred meters. The dominant rocks of the Koyandy complex are garnet–mica schists (intensely retrogressed gneisses) with kyanite, phengite, and K-feldspar relics. These rocks are often migmatized. The complexes also include quantitatively subordinate garnet-free biotite–muscovite schists, marbles, quartzites (metamorphosed cherty rocks), and garnet-free amphibolites, which do not contain any index minerals of high-pressure metamorphism. Detrital zircons from the garnet–mica schists of the Koyandy complex possess cores with relict magmatic zoning, which were dated at 667–834, 868–1051, 1087–1220, 1296–1378, and 2464–2539 Ma, with statistical maxima at ~985 and 1151 Ma, and outer zones, which grew during high-pressure metamorphism (Alexeiev et al., 2011; Pilitsyna et al., 2019a). The protoliths of the metapelites were also composed of the recycled material of the early Precambrian continental crust. The Koyandy complex is noted for hosting bodies of high-pressure melanocratic rocks: garnet amphibolites, eclogites, garnet clinopyroxenites, more rare spinel peridotites, talcites and serpentinites (Alexeiev et al., 2011; Pilitsyna et al., 2018a, 2018b; 2019a). These rocks make up variably sized (from a few to a few dozen meters) tectonic lenses and blocks in the garnet–mica schists.

The peak metamorphic parameters of the Anrakhai gneisses and amphibolites corresponded to the amphibolite facies (6.5–10 kbar, 480–550°C). At the same time, the garnet–mica schists of the Koyandy complex contain preserved relict minerals of high-pressure origin. Some garnet grains in the rocks preserve their growth zoning, including bell-shaped profiles of MnO and pyrope concentrations increasing from the cores to the rims. The further transformations of the rocks at increasing temperature and pressure likely involved garnet growth under high pressures and associated dehydration melting of hydrous minerals with the origin of peritectic kyanite and potassic feldspar in association with melt at 15–18 kbar and 750–850°C. The retrograde metamorphism of the garnet–mica schists was related to the exhumation of the rocks and growth of micas under decompression at a temperature of 580–620°C (Pilitsyna et al., 2019a). Some of the garnet grains partly preserve their relict prograde zoning, which corresponded to parameters of the amphibolite facies: 9–9.5 kbar, 600–640°C. The further temperature and pressure increase due to subduction resulted in an eclogite mineral assemblage formed at 15–18 kbar, 700–800°C. The subsequent decompression at the exhumation of the high-pressure rocks was associated with the replacement of the omphacite by diopside–albite symplectites under granulite-facies parameters and the pervasive development of amphibole under amphibolite-facies parameters (Pilitsyna et al., 2018a). The occurrence of relict rims around inclusion-hosting garnet in the structurally similar garnet and epidote amphibolites of the Koyandy complex and similarities in chemical compositions of these rocks suggest that the garnet (and later also the epidote) amphibolites were produced at retrogression of the eclogites. The garnet clinopyroxenites intercalating with the eclogites yield similar peak metamorphic parameters (16.5–17.5 kbar, 800–860°C) and similar parameters of their retrograde metamorphism to the amphibolite facies. However, a prograde episode of metamorphism in the evolution of the rocks is inferred solely from structural similarities and analogous geochemical features of the garnet clinopyroxenites and eclogites and implies that the protoliths of these rocks should have been involved in the subduction processes (Pilitsyna et al., 2018a). An analogous clockwise *P–T* path was suggested for the spinel ultramafics of the Koyandy complex. This evolution scenario includes the involvement of the protoliths in subduction processes and the origin of, first, spinel and then garnet peridotites at a depth of more than 60 km. The later exhumation of the rocks was associated with thermal relaxation and decompression (garnet decomposition and the origin of pyroxene–spinel symplectites with rims around the orthopyroxene and olivine) at 11–14.5 kbar, 580–800°C (Pilitsyna et al., 2018b).

Two spots at outer metamorphic zones of zircon from the Koyandy garnet–mica schists yielded U–Pb ages of 460 ± 11 and 486 ± 11 Ma (Pilitsyna et al., 2019a). The age of the high-pressure metamorphism of garnet clinopyroxenites in the Koyandy complex is 490 ± 3 Ma (Alexeiev et al., 2011). Zircon from the granitoid leucosome of the migmatized gneisses hosting eclogite and spinel peridotite bodies contains cores with clearly discernible magmatic zoning (Th/U = 0.49–0.64) and outer U-enriched rims (Th/U = 0.008–0.069). The average $^{206}Pb/^{238}U$ age of six concordant values for the cores of the zircons is 755 ± 5 Ma, and the $^{207}Pb/^{206}Pb$ age of one of the cores is 2556 Ma. The outer rims of the zircons have a $^{206}Pb/^{238}U$ age of 483 ± 3 Ma. Some zircon grains from the leucosome are zoned and U-rich (Th/U = 0.009–0.012), and their crystallization age is ~490 Ma.

Hence, the Early Paleozoic age estimates (483–489 Ma) for these rims and some individual zircon grains correspond to the time when the granitoid melt of the leucosome was derived in relation to the melting of the gneisses at high metamorphic grades (Pilitsyna and Tretyakov, 2020). These age values are close to the earlier age estimates for high-pressure metamorphism of the garnet–mica schists and garnet clinopyroxenites of the Koyandy complex. The upper age limit for the exhumation of the high-pressure rocks is defined by the fact that they are unconformably overlain by Early Ordovician (~475 Ma) terrigenous–carbonate rocks.

ISHIM–NARYN (ISHIM–MIDDLE TIEN SHAN) MASSIF

The Ishim–Naryn massif is the largest one in the western Central Asian Orogenic Belt, is crescent-shaped, and extends from the Sarydzhaz Range in the eastern Tien Shan to the warp of the Ishim River in northern Kazakhstan. The massif is made up of various, mostly late Precambrian metamorphic, magmatic, and sedimentary complexes. Metamorphic rocks (including high-pressure ones) were found within a relatively small area in the central part of this massif in the southern part of the Chatkal Range and are the most poorly studied rocks of the type in Kazakhstan and the Tien Shan. Northwest of the Chatkal metamorphic rocks, disintegrated Early Paleozoic ophiolites of the Karaterek complex occur. The complex comprises pyroxenites, gabbro-amphibolites, metabasalts, and siliceous rocks (Ivanov et al., 2002; Alexeiev et al., 2016). The metamorphic rocks are cut by Late Ordovician and Silurian (~450–420 Ma) granitoids, and all of the rocks are overlain by Middle to Late Paleozoic sedimentary sequences (Alexeiev et al., 2016).

Metamorphic rocks in the southern Chatkal Range are combined into the Kassan Group (Bakirov et al., 1989, 1996, 2003) of the Kassan (also referred to as Kassansai) metamorphic complex. The Kassan Group comprises a number of rock complexes of various composition, which are metamorphosed to different grades. The structurally lowermost rocks belong to the Shaldyr complex, which consists of garnet–mica schists and gneisses with bodies of garnet amphibolites and amphibolized eclogites and, occasionally, with migmatites. Structurally above them, marbles with quartzite beds are found, which host rare bodies of amphibolites (metadolerites) of the Tereksai complex. The latter complex is overlain by rocks of the Semizsai complex: garnet–biotite, biotite–muscovite, tourmaline, chlorite–biotite, and chlorite–albite schists, gabbro-amphibolites, garnet amphibolites, and retrogressively altered eclogites and biotite–amphibole schists (Bakirov et al., 1996, 2003; Loury et al., 2016). The Semizsai complex includes serpentinite and listwanite bodies (Bakirov et al., 2003). Detrital zircons from metavolcanic rocks (chlorite–albite schists) and biotite–muscovite schists of the Semizsai complex were dated with a maximum at ~460 Ma. The protoliths of the metasedimentary rocks started to accumulate no earlier than at 510 Ma and consist mostly of the eroded material of Mesoproterozoic (~1000–1200 Ma) rocks, with a certain contribution of material from older (up to 3000 Ma) complexes (Bakirov et al., 1996, 2003; Rojas-Agramonte et al., 2014). The Ishtanberdy metamorphic complex is structurally the highest and is made up of metasedimentary rocks, including staurolite, andalusite–sillimanite, and kyanite–sillimanite schists, which are thought to have been produced by overprinted Barrovian-type metamorphism (Bakirov et al., 1996, 2003; Loury et al., 2016; Alexeiev et al., 2016; Mühlberg et al., 2016). The Sm–Nd isotope characteristics of garnet–mica schists of the Semizsai and Shaldyr complexes (ε_{Nd} = –9 to –10) and metapelites of the Ishtanberdy complex provide evidence of the dominance of ancient crustal material in their protoliths, and the Nd model ages are 1.62 and 1.7 Ga (Mühlberg et al., 2016).

The metamorphic evolution of the Kassan high-pressure rocks have been studied fairly cursorily, mostly in the amphibolized eclogites and amphibolites bearing relics of pyroxene–plagioclase and amphibole–plagioclase symplectites of the granulite facies. The peak metamorphic P–T parameters of the retrogressed eclogites were 16–18 kbar, 490–540°C, and these rocks were then metamorphosed to the amphibolite facies at 11–8 kbar, 560°C (Loury et al., 2016; Mühlberg et al., 2016). The prograde metamorphic temperatures of kyanite–staurolite–garnet–biotite schists of the Ishtanberdy complex are ~400°C, and the peak metamorphic P–T parameters were 7.2 kbar and 650°C (Ivleva, 2003, 2010).

The age of high-pressure metamorphism of the Kassan Group is still uncertain and actively debated. Allanite with omphacite inclusions from the garnet amphibolite with a relict eclogite mineral assemblage was dated at the Late Paleozoic (301 ± 15 Ma), and this age value is interpreted as the age of the near-peak high-pressure metamorphism (Loury et al., 2016). The Sm–Nd mineral isochron on garnet from the eclogite with evidence of retrograde alterations corresponds to similar ages (317 ± 4 and 316 ± 3 Ma), which are close to the ^{40}Ar–^{39}Ar plateau ages (314 ± 2 and 313 ± 2 Ma) of muscovite from the garnet–mica schist hosting the eclogites (Mühlberg et al., 2016). However, absolutely no evidence has been reported as of yet that these metasedimentary complexes were affected by high-pressure metamorphism. Based on the age of the granitoids and acid volcanics cutting the metamorphic rocks, it was hypothesized (Alexeiev et al., 2016, 2019) that high-pressure metamorphism of the Kassan rocks is Early Paleozoic, older than ~420–450 Ma.

COMPOSITIONS AND FORMATION SETTINGS OF THE PROTOLITHS FOR THE HIGH- AND ULTRAHIGH-PRESSURE ROCKS

The absence of preserved mineral associations of the protoliths (except only zircon) in the high- and ultrahigh-pressure metamagmatic and metasedimentary complexes does not allow a comprehensive estimation of the composition of the protoliths and a reliable estimation of the sources of the material and environments in which these rocks had been produced before the onset of the metamorphic processes. However, the composition and distribution of microinclusions in index minerals of the eclogite facies (such as garnet and omphacite) and the isotope characteristics of the high- and ultrahigh-pressure rocks, considered together with the geochronologic data, provide widely used criteria for estimating the composition of the protoliths and settings in which they were formed.

Data on high- and ultrahigh-pressure rocks in the western Central Asian Orogenic Belt acquired over the past decades show that these rocks are dominantly metamagmatic and metasedimentary quartz-feldspathic, and these rocks host eclogites or garnet amphibolites that replaced the eclogites or more rare eclogite–blueschist associations (which were found only in the Issyk-Kul and Chu-Kendyktas massifs). *P–T* paths were calculated for almost all of the high- and ultrahigh-pressure complexes, and the age of their high- and ultrahigh-pressure metamorphism was estimated (Table 1). However, the compositions of the protoliths were estimated only in a few publications, as also were the environments in which these protoliths were produced, and this is a challenging problem calling for further studies.

Zerendy Group of the Kokchetav Massif

The metamorphic complexes of the Zerendy Group are made up of various schists, gneisses, and calc-silicate rocks, which are sometimes migmatized and host bodies of eclogites, garnet and amphibole amphibolites, garnet pyroxenites, and garnet and spinel peridotites. The western domain is dominated by diamond- and coesite-bearing metasedimentary complexes with eclogites and garnet peridotites. The eastern domain includes, along with metasedimentary rocks (the highest grade of which are coesite-bearing), also metamagmatic rocks of lower metamorphic grades. The fact that metamorphic zircons from diamond-bearing garnet–biotite gneisses from the Kumdy-Kol locality contain cores dated within a broad age range suggests that the protoliths of these rocks were sedimentary rocks of mixed composition (Claoue-Long et al., 1991; Shatsky et al., 1995, 1999; Ragozin et al., 2009; Stepanov et al., 2016). The Sm–Nd characteristics of the gneisses indicate that the protoliths of these rocks were formed at the reworking of 2.2- to 2.3-Ga continental crustal complexes (Shatsky et al., 1999). The protoliths of metasedimentary rocks at the Kumdy-Kol locality may have been compositionally close to late Precambrian graphite terrigenous–carbonate and quartzite-schist rocks of the Kokchetav Group (Buslov and Vovna, 2008). However, remarkable differences between the isotopic–geochemical characteristics of the quartzites and schists of the Kokchetav Group [$\varepsilon_{Nd}(T)$ = +3.4 to –3.3, $T_{Nd}(DM)$ 1.3–1.8 Ga], on the one hand, and the Kumdy-Kol gneisses and schists [$\varepsilon_{Nd}(T)$ = –12.8 to –5.4, $T_{Nd}(DM)$ 2.2–2.3 Ga], on the other (Shatsky et al., 1999; Kovach et al., 2017), are in conflict with this interpretation. The garnet–kyanite–mica schists at the Barchi-Kol locality, garnet–kyanite–sillimanite–biotite schists at the Enbek-Berlyk locality, and biotite schists at the Sulu-Tobe locality are rich in alumina and compositionally correspond to clay shales, whereas the high K_2O and SiO_2 concentrations of the Kulet mica schists suggest that they protoliths might have been arkoses (Shatsky et al., 1995, 1999). Calcium-rich clay shales were suggested as a protolith for the diamond-bearing gneisses at the Barchi-Kol locality (Korsakov et al., 2002). The diamond-bearing calc-silicate rocks of the Zerendy Group are also of metasedimentary origin (Sobolev et al., 2011) and are likely metamorphosed clay shales and limestones (Shatsky et al., 1995, 1999).

The high-pressure complexes of the eastern domain are made up, along with metasedimentary rocks, also of widespread orthogneisses that host eclogite and garnet amphibolite bodies. The geochemistry of the orthogneisses indicates that their protoliths were felsic magmatic rocks. The orthogneisses were studied in more detail only locally, at the Chaglinka locality (Glorie et al., 2015). The Chaglinka orthogneisses, which host eclogite bodies, contain accessory zircons of Mesoproterozoic (~1100 Ma) and Early Paleozoic (530–490 Ma) age, with these age values corresponding to the time when the protoliths of the rocks were formed and metamorphosed, respectively. The Hf isotope composition of the zircon [$\varepsilon_{Hf}(T)$ = –3.0 to –17.1, $T_{Nd}(DM)$ 1.7–2.6 Ga] shows that the granitic protolith of the orthogneisses was made up of the material of eroded Paleoproterozoic crustal complexes. The Mesoproterozoic granites with zircons of such isotope characteristics are widespread in the Zerendy Group at areas where rocks of this group are metamorphosed to grades no higher than the amphibolite facies (Glorie et al., 2015; Turkina et al., 2011; Tretyakov et al., 2011). Because of this and in spite of the paucity of the information, it is reasonable to hypothesize that the protoliths of orthogneisses in the eastern domain were Mesoproterozoic granitoids, whose sources were early Precambrian complexes.

The composition of the protoliths of the Zerendy Group eclogites and the environments in which these protoliths were formed are disputable. Chemically, most of the eclogites correspond to N-MORB or, more rarely, are close to E-MORB and/or arc basalts (Shatsky et al., 1993, 2018; Yamamoto et al., 2002; Yui et al., 2010). The Nd isotope characteristics of the eclogites [$\varepsilon_{Nd}(T)$ = +7.2 to –12.5, $T_{Nd}(DM)$ 0.67–1.93 Ga] and the fact that many of the samples are enriched in LREE indicate that the protoliths of the rocks were contaminated with crustal material. Contamination could occur either when the mafic magma interacted with continental crustal complexes or when the eclogites were affected by melts derived from metasedimentary rocks during metamorphism at near-peak parameters (Shatsky et al., 2018). The contamination of eclogites showing N-MORB characteristics with LILE is explained in (Yamamoto et al., 2002) by interaction with melts in the mantle wedge separated from the subducted slab. The least contaminated sample among the studied ones is eclogite from the Sulu-Tobe locality, whose normalized incompatible-element patterns and Nd and Sr isotope characteristics are close to those of N-MORB. This eclogite yielded the youngest Nd model age of ~670 Ma, which can be interpreted as the maximum crystallization age of the protolith of this rock. The protoliths of the eclogites are believed to have been small mafic bodies that were emplaced in an extensional environment at the passive margin of the continental block (Zhang et al., 2012). The very low $\delta^{18}O$ (–3.9‰ for the garnet) of the Kulet and Barchi-Kol eclogites may suggest that the protoliths of these rocks have interacted with cold meteoric waters in an environment transitional between continental rifting to oceanic spreading (Masago et al., 2003).

It is thought that garnet peridotites with Ti-clinohumite at the Kumdy-Kol locality and orthopyroxene, anthophyllite, and olivine–spinel peridotites (“spinel harzburgites”) at the Enbek-Berlyk locality were likely formed by metamorphism of near-surface bodies of metasomatized (variably chloritized) basalts, which had occurred in the upper continental crust before being brought to greater depths (Reverdatto and Selyatitskiy, 2005). Other researchers believe that the garnet–Ti-clinohumite peridotites could be formed by metasomatic recycling (Mg metasomatism) of eclogites or their protoliths of basaltic composition (Yui et al., 2010). The probable candidates for protoliths of the coesite–talc–garnet–kyanite–phengite schists are also thought to be basaltic rocks metasomatized by seawater-derived fluid (Yui et al., 2010).

The likely protoliths of the metamorphic rocks of the Zerendy Group were thus Meso- and perhaps also Early Proterozoic sedimentary and magmatic rocks of various composition. In the Early Paleozoic, the protoliths of the high- and ultrahigh-pressure rocks were subducted to various depths, metamorphosed and partly melted there, and then tectonically combined during exhumation (Stepanov et al., 2016; Shatsky et al., 2018).

Akdzhon Group of the Issyk-Kul Massif

The high- and ultrahigh-pressure rocks of the Akdzhon Group are coesite-bearing garnet–chloritoid–talc schists, quartzite–schists, and quartzites with garnet porphyroblasts with coesite relics, amphibolized eclogites, “glaucophanites”, and garnet–phengite–biotite schists that are structurally lowermost in the core of the Makbal antiform. The protoliths of the quartzites were likely terrigenous rocks, which were accumulated in environments of a passive continental margin (Bakirov nad Maksumova, 2001; Togonbaeva et al., 2009; Meyer et al., 2014; Konopelko et al., 2016). The sources of clastic material for the protoliths of the quartzites were Paleoproterozoic and Neoarchean crustal complexes (Degtyarev et al., 2013; Konopelko et al., 2016; Alexeiev et al., 2020). It should be mentioned that ultrahigh-pressure metamorphism affected only some rocks of the Akdzhon Group and resulted in the association of garnet with coesite inclusions, whereas most of the Akdzhon rocks were metamorphosed to lower grades. No information is now available on the compositions and origin environments of garnet–phengite–biotite schists widespread in the Akdzhon Group, but the presence of detrital zircons in these rocks suggests that their protolith consisted of sedimentary rocks (Degtyarev et al., 2013).

One of the most disputable issues is the composition and origin environments of the protoliths of the eclogites and “glaucophanites” developing after them, garnet amphibolites, and garnet–chloritoid–talc schists of the Neldy and Makbal complex of the Akdzhon Group. Some researchers maintain that the protoliths of the eclogites were oceanic crustal fragments produced at spreading centers (Bakirov et al., 2003; Baruleva et al., 2011; Meyer et al., 2013, 2014). The enriched geochemical features of the rocks are reportedly explained by that the protoliths of the eclogites occurred in areas affected by a mantle plume or seamounts (Meyer et al., 2013). Other scientists suggest that the protoliths of the eclogites and garnet amphibolites were mafic dikes, which were emplaced into sedimentary sequences of the continental margin during the final breakup of the Rodinia supercontinent (Konopelko et al., 2012, 2016; Rojas-Agramonte et al., 2013; Klemd et al., 2015). The nature of the protolith of the garnet–chloritoid–talc schists is also uncertain. On the one hand, candidates for their protoliths are mafic magmatic rocks that had been hydrothermally reworked in the oceanic basin and then metamorphosed to garnet–chloritoid–talc schist under ultrahigh pressures during subduction (Meyer et al., 2014). On the other hand, the protoliths of these schists are thought to have consisted of sedimentary rocks, as follows from that the rocks contain detrital zircons and from data on the Nd isotopic composition [$\varepsilon_{Nd}(T) = -11$], which indicate that the protoliths of the garnet–chloritoid–talc schists were sourced from the eroded ancient crustal complexes (Meyer et al., 2014; Konopelko et al., 2012, 2016).

Hence, the protoliths of the high- and ultrahigh-pressure rocks of the Akdzhon Group were Paleo- and Mesoproterozoic sedimentary rocks of various composition. The protoliths of the eclogites were mafic magmatic rocks. These rocks were involved in subduction in the Early Paleozoic and were transformed into coesite-bearing schists, quartzites, and eclogites.

Aktyuz and Kemin Complexes of the Chu-Kendyktas Massif

Data on the composition and nature of the protoliths of high-pressure rocks in the Aktyuz block of the Chu-Kendyktas massif are fragmentary. The affiliation of the most widely spread quartzo–feldspathic rocks of the Aktyuz and Kendyktas complexes (garnet-bearing gneisses and schists) with high-pressure rocks still has not been confirmed. However, currently available isotope geochronologic data indicate that the protoliths of these rocks were formed in the Neoproterozoic by means of recycling of more ancient (~1.5–2.1 Ga) complexes (Kröner et al., 2012). The composition of the amphibolized eclogites and their relationships with the host gneisses suggest that their protoliths were likely mafic dikes, which had been emplaced into continental crustal complexes before the subduction began (Kröner et al., 2012; Rojas-Agramonte et al., 2013; Klemd et al., 2014). The isotopic–geochemical characteristics of the eclogites [$\varepsilon_{Nd}(T) = +3.3$ to $+3.7$] suggest that their protoliths were formed by the melting of the material of subcontinental lithospheric mantle and, likely, also by crustal contamination (Klemd et al., 2014).

Koyandy Complex of the Zheltau Massif

The protoliths of the high-pressure Koyandy rocks were both magmatic and sedimentary rocks. The protoliths of the kyanite–garnet paragneisses of the Koyandy complex were terrigenous rocks accumulated in the Ediacaran–Cambrian. The eroded material was dominated by felsic magmatic rocks ~1 Ga in age, which were derived from a Paleoproterozoic crustal source (Alexeiev et al., 2011; Pilitsyna et al., 2019a). The chemical composition of the eclogites and garnet clinopyroxenites suggests that their protolith consisted of intraplate mafic complexes, which had been emplaced into the continental margin before subduction was initiated (Pilitsyna et al., 2018a). The age of the protolith of the eclogites and garnet clinopyroxenites and, hence, the age of the intraplate magmatism are uncertain. The age estimate of 489 ± 3 Ma earlier obtained for the garnet pyroxenites (Alexeiev et al., 2011) pertains to the timing of the high-pressure metamorphism, which implies that the protolith of the rocks was older than Ordovician. The protoliths of the spinel peridotites seem to have been fragments of a layered ultramafic–mafic complex, which comprised plagioclase-bearing peridotites, troctolites, olivine gabbro, and melanogabbro derived from a depleted mantle source in a suprasubductional environment. During the closure of the oceanic basin, these rocks were brought to significant depths, which corresponded to the garnet stability field, and were then exhumed together with other high-pressure rocks of the Koyandy complex (Pilitsyna et al., 2018b).

Kassan Group of the Ishim–Naryn Massif

The high-pressure rocks of the Kassan Group are more poorly studied than rocks of all complexes discussed above. The age of the high-pressure metamorphism (Loury et al., 2016; Alexeiev et al., 2016) is uncertain, as also is whether most of the metamorphic rocks (except only the amphibolized eclogites) are high-pressure ones. Very little information has been published so far on the composition of the protoliths of the high-pressure rocks and on the environments in which these protoliths were formed. The garnet–biotite, muscovite–biotite, and chlorite–albite schists that host eclogite bodies are metasedimentary rocks, whose protoliths were accumulated no later than at 510 Ma and these were mostly sourced from the eroded Paleo- and Mesoproterozoic complexes (Bakirov et al., 1996; Rojas-Agramonte et al., 2014; Mühlberg et al., 2016). The eclogites are mostly transformed into garnet amphibolites and are classified according to their structural setting into boudins in schists and gneisses of the Shaldyr complex and sill-shaped metadolerite bodies among rocks of the Shaldyr and Tereksai complexes (Bakirov et al., 2003). The chemical compositions of the Shaldyr eclogites indicate that the protoliths of these rocks were formed from an enriched source and were mafic rocks comparable to intraplate basalts (Loury et al., 2016). The protoliths of the Kassan eclogites were likely formed in a continental rifting environment (Mühlberg et al., 2016). The ε_{Nd} values of the eclogites (calculated for an age of ~316 Ma) are +2.4 to +4 and are interpreted as evidence that the mantle source of the protoliths of these rocks was mildly depleted (Mühlberg et al., 2016). For an age of ~460 Ma, which was obtained for these rocks in (Alexeiev et al., 2016), $\varepsilon_{Nd}(460) = +3.6$ to $+5.8$.

Thus, the protoliths of the Early Paleozoic high- and ultrahigh-pressure rocks in the western segment of the Central Asian Orogenic Belt were dominantly late Precambrian quartzo–feldspathic gneisses, schists, and quartzites, which had been derived from the material provided by the reworking of older crustal rocks.

The mafic rocks, which were metamorphosed and transformed into the eclogites and garnet amphibolites, usually made up small rift-related mafic bodies in the continental crust. The high-pressure complexes include very rare fragments of metamorphosed oceanic crust, whose presence is most reliably proved only for the Zheltau massif.

PROBLEMS ENCOUNTERED IN BUILDING GEODYNAMIC MODELS FOR THE ORIGIN OF THE HIGH- AND ULTRAHIGH-PRESSURE COMPLEXES IN THE WESTERN CENTRAL ASIAN OROGENIC BELT

The development of geodynamic models for the origin of high- and ultrahigh-pressure rocks is an important aspect of the studies of the latter. Such models are usually built based on as comprehensive as possible data on the structure, composition, geodynamic environments, and ages of both the high- and ultrahigh-pressure complexes and those found in nearby tectono–stratigraphic zones. Nowadays most researchers adhere to subductional models for the origin of high-pressure complexes, because such models are able to explain how the rocks were brought to depths, corresponding to high temperatures and pressures (Tsujimori et al., 2006; Ernst et al., 2007). According to the models, magmatic and sedimentary rocks in the subducted slab are buried to significant depths (up to 100–150 km) and then rapidly exhumed, along the same pathway, and therewith the rocks at least partly could preserve their high-pressure mineral assemblages. Depending on the types of interaction between the plates, a diversity of high- and ultrahigh-pressure complexes can be formed. At oceanic convergence, oceanic lithospheric complexes are subducted and eclogite–blueschist metamorphic associations are produced. These complexes are commonly spatially constrained to ophiolite zones and are associated with ultramafic rocks, gabbroids, plagiogranites, basalts, and siliceous rocks, as blocks in serpentinite mélange (Maruyama et al., 1996). In the western part of the Central Asian Orogenic Belt, eclogite–blueschist metamorphic complexes are known in the Chara and North Balkhash ophiolite zones (Volkova et al., 2008; Pilitsyna et al., 2019b). If the convergent processes involve slabs of continental crust, the high-pressure complexes are formed in relation to arc–continent (microcontinent) or continent–continent (microcontinent) collision and earlier subduction. Thereby subduction affects a passive margin of the continent (microcontinent), which is brought to mantle depths following the oceanic slab that separates it from the island arc or another continent (Figs. 5a, 5b). Then the high buoyancy of the continental crust and the high density of the subducted ocean slab cause the break-off of the slab, after which the high-pressure complexes are rapidly exhumed to mid- and upper-crust levels (Fig. 5c). According to this scenario, eclogite–schist–gneiss and eclogite–schist–quartzite metamorphic associations are formed, and all of the complexes discussed herein are made up of these associations. The metamorphic rocks are therewith exhumed from different depths, including those in the middle and lower crust, and this is the reason why combinations of rocks metamorphosed to different grades that are often found in a single complex.

This model for the origin of high-pressure rocks, which were produced by metamorphism of continental crustal complexes, implies that a lateral sequence of structures shall occur at the subduction and subsequent collision of the continent (microcontinent) with an island arc or with another continental block. This sequence shall include a continental margin of the continent (microcontinent), a basin with oceanic crust, an accretionary prism, and an island arc or an active continental margin. Upon the termination of the accretion–collision processes and the origin of the thrust–folded edifice, only fragments of the complexes produced in these structures are usually preserved. The complexes of the basin with oceanic crust and those of the accretionary prism can be preserved only very poorly, and these complexes can be sometimes even completely consumed and destroyed at subduction and later collision. The thrust–folded structures contain much better preserved complexes of the continental blocks and their margins, as well as complexes of the island arcs, because these tend not to be buried to greater depths, and their subduction is largely controlled by their positive buoyancy. In view of this, while reproducing the origin of high-pressure rocks, including those in the areas discussed in this paper, it is necessary to distinguish complexes of at least some of the structures of the lateral sequence during subduction and collision. Below we discuss some problem emerging when models are developed for the origin of (ultra)high-pressure complexes for the indicated territories.

Kokchetav Massif

Various models were put forth for the origin of the high- and ultrahigh-pressure rocks in the area and the Early Paleozoic evolution of northern Kazakhstan as a whole. In these models, the tectonic evolution of the area is usually reproduced starting with events in the latest Neoproterozoic, when the largest continental block broke up and a number of microcontinents were produced (including the Kokchetav one), along with basins with oceanic crust and an ensimatic (?) island arc. The most important event in the earliest Cambrian (about 530 Ma) was the subduction of the thinned continental crust of the margin of the Kokchetav massif beneath the island arc to depths of 35–140 km, the origin of high-pressure metamorphic complexes, and the subsequent collision of the massif with an island arc (Dobretsov et al., 1998, 2006;

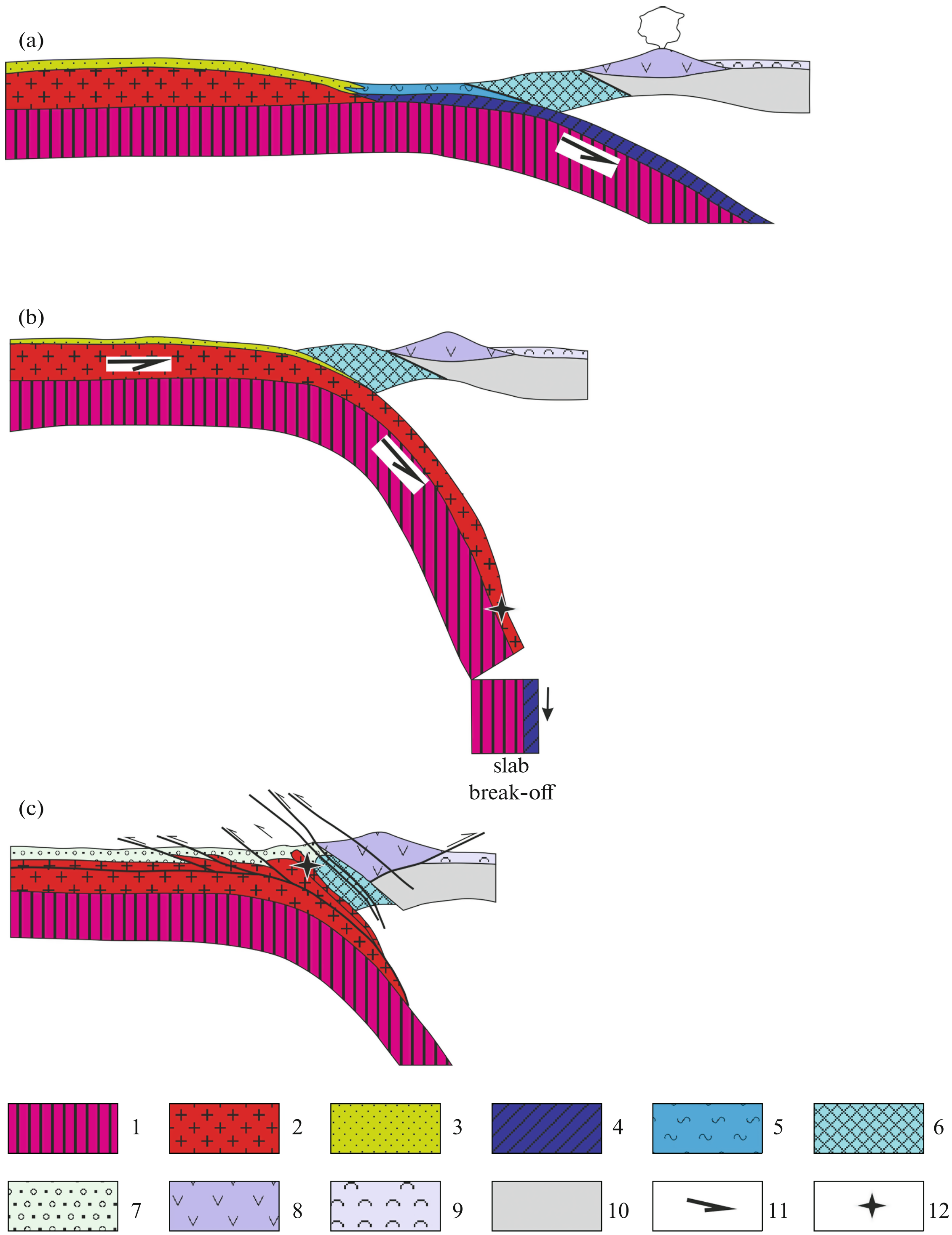

Fig. 5. Simplified geodynamic model of formation of high- and ultrahigh-pressure complexes in the western part of the Central Asian Orogenic Belt. (a) Subduction of the crust of the oceanic basin beneath an ensimatic island arc; (b) collision of the continent (microcontinent) and ensimatic island arc and the formation of (ultra)high-pressure rocks; (c) rapid exhumation of the high-pressure complexes to mid- and upper crustal depth levels.
(1) Lithospheric mantle; (2) continental crustal complexes; (3) terrigenous and terrigenous–carbonate complexes of the passive margin of the continent (microcontinent); (4, 5) complexes of basins with oceanic crust; (4) plutonic component of the ophiolites, (5) siliceous–basaltic and siliceous sequences; (6) complexes of the forearc block; (7) terrigenous sequence, including rudaceous ones; (8–10) complexes of the ensimatic island arc: (8) volcanic rocks, (9) tuff–sedimentary rocks, (10) complexes of the melanocratic basement; (11) directions of the relative motions of the blocks; (12) rocks containing high- and ultrahigh-pressure mineral assemblages.

Maruyama et al., 2000; Glorie et., 2015; Degtyarev et al., 2016).

A principal disadvantage of these models is the absence of reliable data on the oceanic-basin and island-arc complexes, which were formed in the western part of the Central Asian Orogenic Belt in the latest Neoproterozoic (~570–540 Ma). Another uncertain issue is the polarity of the subduction zone to which the origin of the high-pressure complexes was related. Because of this, the models rely on variably justified and warranted assumptions concerning these complexes in the discrete tectono–stratigraphic zones of northern Kazakhstan.

N.L. Dobretsov distinguished the Ishim arc, which developed in the Ediacaran–earliest Cambrian. Its complexes include the Precambrian rocks in the extreme west and northwest of the Kokchetav massif and in the southwest of the Ulutau and Chu-Kendyktas massifs (Dobretsov et al., 1998, 2006). However, no arc complexes of this age are known in either the Ulutau or the Chu-Kendyktas massifs. The latest Precambrian–earliest Cambrian rocks of the massifs make up terrigenous–carbonate and black-shale sequences and tillitoids, which host thick high-Ti basalt flows at various stratigraphic levels and which are correlated with rifting complexes (Zaitsev and Kheraskova, 1979; Abdulin et al., 1980). In the western and northwestern parts of the Kokchetav massif, Cambrian complexes include not only the widespread quartzites and schists of the Kokchetav Group but also rocks of the Efimov Formation, which consists of alternating sericite–chlorite, carbonaceous, and mica–quartz schists with beds of blastopsammitic quartzite–sandstones and marmorized limestones, and the intraplate basalt–rhyolite Imanburluk Formation (Degtyarev et al., 2016). No Neoproterozoic–earliest Cambrian arc rocks that could have been produced in the Ishim island arc have been found in the surroundings of the Kokchetav, Ulutau, and Chu-Kendyktas massifs.

The Selety and Stepnyak island arcs were also considered to be likely candidates for structures characterized by the subduction of the continental crust with which high-pressure metamorphism could be related (Dobretsov et al., 2006; Glorie et al., 2015). Their complexes compose the Selety and Stepnyak tectonostratigraphic zones east of the Kokchetav massif, where arc complexes occur (according to recent results) whose ages are younger than those of high-pressure rocks in the Kokchetav massif (~530 Ma) (Degtyarev, 2012). Hence, the peak of the high-pressure metamorphism occurred before the Selety and Stepnyak island arcs started to develop and was not related to their evolution.

One of the authors of this paper suggested that complexes of the late Precambrian island arc are, perhaps, rocks of the Daut Formation, which were found in the northern part of the Kokchetav massif, in the northern part of the Shatskiy massif. The lower part of the Daut Formation is made up of sheared basalts and mafic tuffs with rare rhyolite flows and felsic tuff beds. The upper part of this formation contains intercalating sheared rhyolite, crystallo- and lithoclastic felsic tuffs, and tuff-sandstones. The total thickness of the formation seems to reach 1500 m, and the composition of its rocks suggests that they were formed in a suprasubductional environment. With regard to its metamorphic grade, the Daut Formation could be provisionally attributed to the Neoproterozoic (Degtyarev et al., 2016). However, our later geochronologic data show that the felsic volcanics of the Daut Formation are of Early Ordovician age (480–485 Ma). Hence, the high-pressure metamorphism of rocks of the Kokchetav massif was not related to the evolution of this arc.

The recognition of complexes of the oceanic lithosphere of the basin that separated the island arc from the Kokchetav massif in the latest Neoproterozoic–earliest Cambrian is also problematic.

N.L. Dobretsov and his coauthors distinguished the domain of Beloe Lake (Fig. 2) south of the zone of high-pressure rocks. The domain consists of amphibolites, amphibole schists, and metamorphosed cherts. They are thought to be crustal fragments of the paleoceanic basin. No data are, however, available on either the composition or the age of these rocks, and it is only known that they are tectonically superimposed by quartzites and schists of the Neoproterozoic Kokchetav Group (Dobretsov et al., 1998).

In later publications, it was suggested that the oceanic lithospheric complexes may include rocks of the Shchuchinsk ultramafic–gabbro complex, which occur as small bodies in the eastern Kokchetav massif (Dobretsov et al., 2006; Zhimulev et al., 2011). The complex is dominated by serpentinized ultramafics and contains quantitatively subordinate gabbro (Mikhailov, 1971). Studies of the largest ultramafic–gabbro body in the eastern part of the Kokchetav massif have resulted in dating the gabbro at the Late Cambrian (~490 Ma) and in determining that these and the metamorphic rocks occur in intrusive relationships with one another (Degtyarev et al., 2016).

No reliable evidence has been found in either the Kokchetav massif itself or its surroundings that most of the aforementioned paleostructures occurred when the high- and ultrahigh-pressure rocks were formed during subduction and collision.

Issyk-Kul Massif

Most researchers are prone to believe that the high-pressure rocks of the Akdzhon Group in the western part of the Issyk-Kul massif were formed in relation to the closure of the Terskey oceanic basin (Degtyarev et al., 2013; Klemd et al., 2015; Bakirov, 2017; Alexeiev et al., 2019). Complexes of this basin occur in the Kyrgyz–Terskey zone of the northern Tien Shan, which now separates the Issyk-Kul and Ishim–Naryn massifs with Precambrian continental crust.

The Kyrgyz–Terskey zone and Issyk-Kul massif includes complexes that were formed within the lateral sequence of structures produced during subduction and collision, when the high-pressure complexes were formed. At the same time, the development of a justified geodynamic model for their origin faces certain problems.

It is thought that the Issyk-Kul microcontinent had a passive margin in the Neoproterozoic–Cambrian, and terrigenous–carbonate rocks were then accumulated on this margin (Mikolaichuk et al., 1997; Degtyarev et al., 2013). However, no reliable data are available as of yet on the age of quartzite–schist sequences in the western part of the Issyk-Kul massif. Some of the schist–carbonate and quartzite sequence, which were previously thought to be of Neoproterozoic–Cambrian age (Degtyarev et al., 2013), were found out to contain detrital zircons dated at older than 1 Ga (Alexeiev et al., 2020). Because of this, these complexes are currently thought to be latest Mesoproterozoic–earliest Neoproterozoic and are correlated with coeval quartzite–schist sequences in northern Kazakhstan (Alexeiev et al., 2020). The timing of the emplacement of the mafic rocks that were then transformed to eclogites and garnet amphibolites that occur as dikes and sills hosted in the terrigenous–carbonate sequence on the passive margin is unknown, however, multiple pulses of intrusion of these rocks were suggested (Degtyarev et al., 2013).

The high-pressure metamorphism of the Akdzhon rocks (~510–500 Ma) is related in the current models to the collision of the Issyk-Kul microcontinent with the Early Cambrian Sultansary ensimatic arc and the closure of the basin that separated them and had an oceanic crust. This event was predated by subduction beneath the island arc (Degtyarev et al., 2013; Alexeiev et al., 2019). Complexes of the Sultansary island arc were found in the central part of the Kyrgyz–Terskey zone, where the Kapkatash Group of differentiated volcanics rocks was ascribed to these complexes. The active evolution of this island arc terminated in the latest Early Cambrian, when it was overlain by a carbonate cover (Mikolaichuk et al., 1997). Relics of complexes of the basin with oceanic crust that separated the island arc and Issyk-Kul microcontinent were found in the western Kyrgyz–Terskey zone: these are Early Cambrian (~520 Ma) ophiolites with suprasubductional characteristics (Degtyarev et al., 2013).

The models do not consider the emplacement of the Kandzhailyau granitoid pluton in the latest Early to earliest Middle Cambrian (~515 Ma), with this pluton cutting across rocks of the Akdzhon Group (Konopelko et al., 2012). The composition of this pluton is known inadequately poorly, and the environments in which it was produced are uncertain. Also, no considerations are given to the Early Ordovician (~470 Ma) age estimates for the near-peak high-pressure metamorphism obtained on garnet amphibolites replacing eclogites in the Makbal complex (Rojas-Agramonte et al., 2013). This metamorphism might have been related to the evolution of the Late Cambrian–Early Ordovician lateral sequence of structures that included the margin of the Issyk-Kul microcontinent, a basin with oceanic crust (basalts of the Late Ordovician Terek Formation), and the Karaarcha ensimatic island arc (basalts, basaltic andesites, and andesites of the Late Cambrian Karaarcha Formation (Degtyarev et al., 2013).

Chu-Kendyktas and Zheltau Massifs

It is worth closely examining side by side geodynamic models for the origin of high-pressure rocks in the Chu-Kendyktas and Zheltau massifs, because high-pressure metamorphism in both massifs was reportedly related to the evolution of the Dzhalair–Naiman ophiolite zone (Alexeiev et al., 2011; Kröner et al., 2012; Klemd et al., 2015; Pilitsyna et al., 2019b). The zone is made up of Early Cambrian (520–525 Ma) suprasubductional ophiolites and a contrasting series, Late Cambrian island-arc complexes (Sulusay Formation), and ophiolites with N-MORB and E-MORB characteristics (Ashchisu Formation) (Ryazantsev et al., 2009; Degtyarev, 2012). These rocks are tectonically superimposed by an Early Ordovician flysch sequence (Degtyarev, 2012).

The high-pressure metamorphism of complexes on the margin of the Chu-Kendyktas microcontinent in the Early Ordovician (~475 Ma) is related in the current models to the closure of the Dzhalair–Naiman basin with oceanic crust and subsequent collision with the Zheltau microcontinent. It is thereby assumed that the margin of the Zheltau microcontinent continued to be active throughout the whole Cambrian (Kröner et al., 2012; Klemd et al., 2015). However, no Cambrian subductional complexes were found in the Zheltau massif. Cambrian (~510 Ma) granodiorites and granites with suprasubductional characteristics in the northeastern part of the Anrakhai block (Fig. 4) are separated from complexes of the Zheltau massif by ophiolites (Pilitsyna et al., 2018b). It thus seems to be more probable that the high-pressure metamorphism of the Aktyuz and Kemin complexes was related to the collision of the Chu-Kendyktas microcontinent with the Sulusay island arc, whose complexes are widespread in the southeastern part of the Dzhalair–Naiman zone. The later collision between the Chu-Kendyktas and Zheltau microcontinents resulted in the southwestward obduction of the ophiolites (Pilitsyna et al., 2018b).

In the model suggested by Alexeiev et al. (2011), high-pressure metamorphism of the Zheltau microcontinent (~490 Ma) is related to the subduction of the oceanic crust of the Dzhalair–Naiman basin beneath the active margin of the Zheltau (Anrakhai)

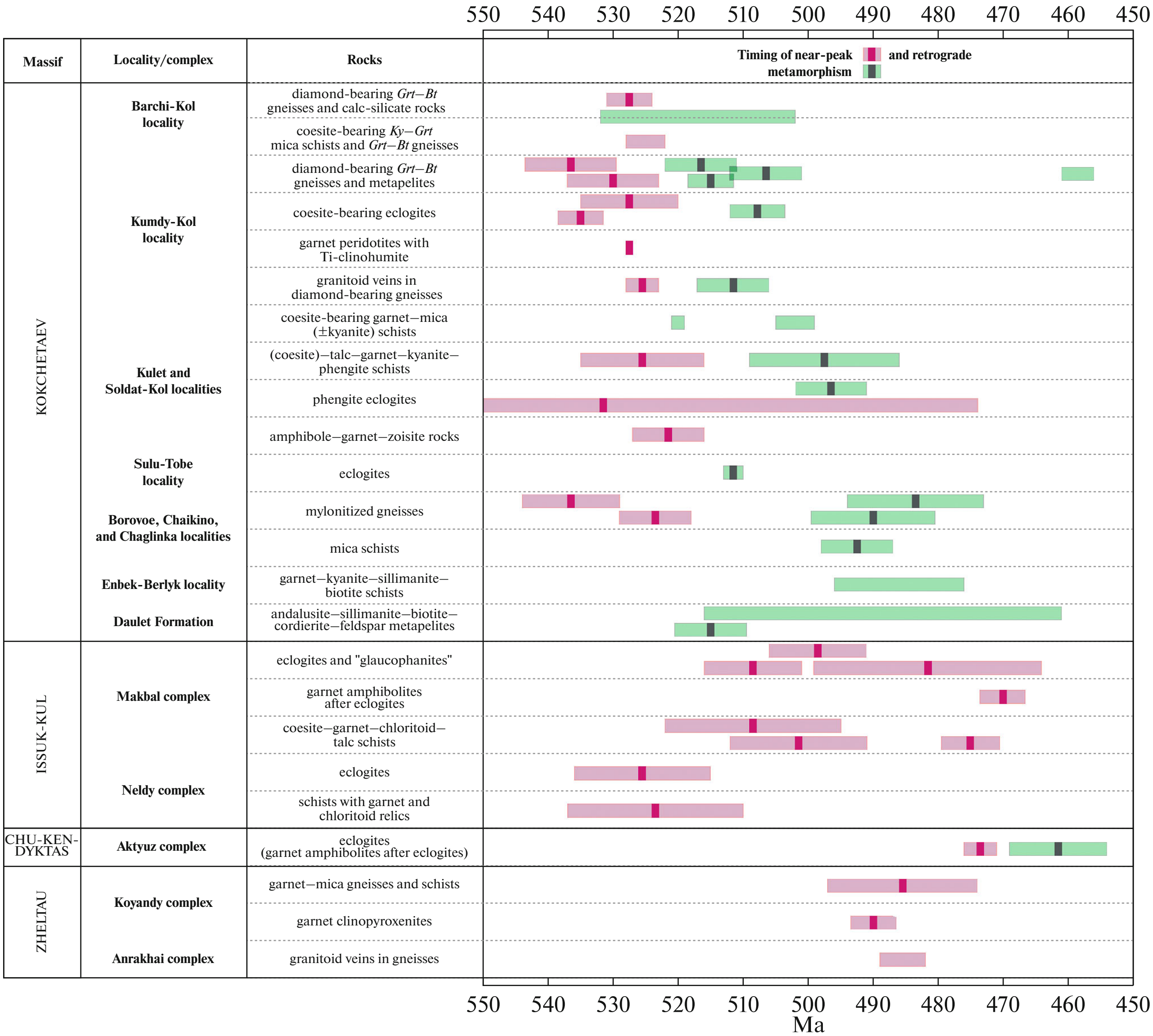

Fig. 6. Estimated timing of the near-peak (pink) and later (green) episodes of metamorphism of ultrahigh-, high- and medium-pressure rocks in the Kokchetav, Issyk-Kul, Chu-Kendyktas, and Zheltau massifs in the western segment of the Central Asian Orogenic Belt.
The dark pink and dark green rectangles are the calculated age clusters, and the pale pink and pale green ones are the calculated deviations or ranges of values.

microcontinent. According to this model, a fragment of the Zheltau microcontinent crust was tectonically detached and involved in subduction process, along with the oceanic slab, and the metamorphic complexes were afterwards exhumed, while subduction continued, along the same channel to upper crustal levels.

According to our model (Pilitsyna et al., 2018b, 2019a), high-pressure metamorphism of complexes of the passive margin of the Zheltau microcontinent was related to the closure of the oceanic basin that separated the Zheltau and Aktau–Yili microcontinents and the subsequent subduction of the oceanic crust and the passive margin of the Zheltau microcontinent beneath the active margin of the Aktau– Yili microcontinent. Rocks of the passive margin of the Zheltau microcontinent are metaterrigenous sequences of the Koyandy complex, which host variably metamorphosed fragments of the oceanic lithosphere (amphibolized gabbro and serpentinized spinel peridotites) occurring in the northeastern part of the Anrakhai block (Fig. 4). Rocks of the active margin of the Aktau–Yili microcontinent are thought to include Cambrian (~510 Ma) granitoids and granites with suprasubductional characteristics in the northeast of the Anrakhai block (Fig. 4).

It follows that geodynamic models of formation of (ultra)high-pressure complexes in the Issyk-Kul, Chu-Kendyktas, and Zheltau massifs can be based on much more extensive information, which enables one to reproduce in more or less detail the lateral sequences of structures that occurred during the subduction and collision. At the same time, the active collision–accretion processes have obliterated the complexes of many of these structures or drastically reduced them. Because data on the high-pressure complexes in the Ishim–Naryn massif are very fragmentary, it is now impossible to suggest any realistic geodynamic model for their origin. Nevertheless, similarities between the compositions and structural positions of the key varieties of metamorphic rocks in the Ishim–Naryn massif and analogous complexes in Kazakhstan and northern Tien Shan make it possible to propose a similar idealized and simplified geodynamic model.

CONCLUSIONS

This paper reviews data acquired over the past three decades by studying Early Paleozoic high- and ultrahigh-pressure complexes in the western part of the Central Asian Orogenic Belt. These complexes make up the basement of a number of large massifs with Precambrian crust: the Kokchetav massif in northern Kazakhstan (Zerendy Group), Issyk-Kul massif in the northern Tien Shan (Akdzhon Group), Chu–Kendyktas and Zheltau massifs in southern Kazakhstan (Aktyuz, Kemin and Koyandy complexes), and the Ishim–Naryn massif in the central Tien Shan (Kassan Group). These rocks (mostly eclogites, gneisses, and schists, which sometimes contain relics of coesite and microdiamonds) were reportedly produced in connection with the subduction of the passive margin of the microcontinents to mantle depth levels. It is thought that the subsequent rapid exhumation of the high- and ultrahigh-pressure complexes, together with rocks metamorphosed to lower grades, from various depths resulted in the formation of the packages of tectonic slices composed of metamorphic rocks of different genesis.

Studies of the high- and ultrahigh-pressure metamorphic complexes made it possible to calculate numerous P–T estimates for the peak metamorphism, retrograde and, to a lesser extent, prograde metamorphism. The timing of the peak metamorphism was assessed for most of the high- and ultrahigh-pressure rocks, as also were the retrogression stages of some of the rocks during their exhumation (Fig. 6). The paucity of geochronological and isotopic–geochemical data on the ages of the protoliths of the key rock varieties and the fragmentary character of information on their compositions, geodynamic settings of formation and the sources still do not make it possible to develop reliable and detailed geodynamic models for the evolution of the high- and ultrahigh-pressure complexes in the Early Paleozoic. Studies of this type are also useful and current from the standpoint of understanding the exhumation mechanisms of such rocks, which are still poorly characterized.

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