

Balkatach hypothesis: A new model for the evolution of the Pacific, Tethyan, and Paleo-Asian oceanic domains

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ABSTRACT

The Phanerozoic history of the Paleo-Asian, Tethyan, and Pacific oceanic domains is important for unraveling the tectonic evolution of the Eurasian and Laurentian continents. The validity of existing models that account for the development and closure of the Paleo-Asian and Tethyan Oceans critically depends on the assumed initial configuration and relative positions of the Precambrian cratons that separate the two oceanic domains, including the North China, Tarim, Karakum, Turan, and southern Baltica cratons. Existing studies largely neglect the Phanerozoic tectonic modification of these Precambrian cratons (e.g., the effects of India-Arabia-Eurasia convergence and post-Rodinia rifting). In this work we systematically restore these effects and evaluate the tectonic relationships among these cratons to test the hypothesis that the Baltica, Turan, Karakum, Tarim, and North China cratons were linked in the Neoproterozoic as a single continental strip, with variable along-strike widths. Because most of the tectonic boundaries currently separating these cratons postdate the closure of the Paleo-Asian and Tethyan Oceans, we are able to establish a >6000-km-long Neoproterozoic contiguous continent referred to here as Balkatach (named from the Baltica–Karakum–Tarim–North China connection). By focusing on the regional geologic history of Balkatach's continental margins, we propose the following tectonic model for the initiation and evolution of the Paleo-Asian, Tethyan, and Pacific oceanic domains and the protracted amalgamation and growth history of the Eurasian continent. (1) The early Neoproterozoic collision of the combined Baltica–Turan–Karakum–South Tarim continents with the linked North Tarim–North China cratons led to the formation of a coherent Balkatach continent. (2) Rifting along Balkatach's margins in the late Neoproterozoic resulted in the opening of the Tethyan Ocean to the south and unified Paleo-Asian and Pacific Oceans to the north (present-day coordinates). This process led to the detachment of Balkatach-derived microcontinents that drifted into the newly formed Paleo-Asian Ocean. (3) The rifted microcontinents acted as nuclei for subduction systems whose development led to the eventual demise of the Paleo-Asian Ocean during the formation of the Central Asian Orogenic System (CAOS). Closure of this ocean within an archipelago-arc subduction system was accommodated by counterclockwise rotation of the Balkatach continental strip around the CAOS. (4) Initial collision of central Balkatach and the amalgamated arcs and microcontinents of the CAOS in the mid-Carboniferous was followed by a bidirectional propagation of westward and eastward

suturing. (5) The closure of the Paleo-Asian Ocean in the early Permian was accompanied by a widespread magmatic flare up, which may have been related to the avalanche of the subducted oceanic slabs of the Paleo-Asian Ocean across the 660 km phase boundary in the mantle. (6) The closure of the Paleo-Tethys against the southern margin of Balkatach proceeded diachronously, from west to east, in the Triassic–Jurassic.

INTRODUCTION

Eurasia, the largest and youngest continent on Earth, was assembled through the Neoproterozoic to the present over a time span of 1 b.y. (Scotese and McKerrow, 1990; Şengör and Natal'in, 1996). The three major oceanic domains of Eurasia—the Paleo-Asian, Tethyan, and Pacific Oceans—represent large Phanerozoic oceans that were (or still are) intimately related to the assembly and evolution of the continent (Fig. 1) (e.g., Şengör 1984, 1992; Şengör et al., 1988, 2008; Zonenshain et al., 1990; Yin and Nie, 1996; Şengör and Natal'in, 1996; Heubeck, 2001; Badarch et al., 2002; Stampfli and Borel, 2002; Sone and Metcalfe, 2008; Biske and Seltmann, 2010; Gehrels et al., 2011; Burchfiel and Chen, 2012; Wu et al., 2017a). Decades of research have generated a wealth of geologic data, yet the tectonic history of the opening and closing of these major ocean systems remains debated. For example, researchers have variably attributed the evolution of the Paleo-Asian Ocean to (1) lateral expansion of juvenile continental crust via protracted trench retreat and strike-slip duplexing of a long-lived continental arc derived from Siberia or Baltica (Şengör et al., 1993; Şengör and Natal'in, 1996; Yakubchuk, 2002), (2) accretion of multiple intraoceanic and/or continental arcs onto the continents bounding the Paleo-Asian Ocean(s) (i.e., a multiple arc model) (e.g., Zonenshain et al., 1990; Hsü and Chen, 1999; Filippova et al., 2001; Badarch et al., 2002; Windley et al., 2007; Xiao et al., 2008), or (3) redistribution and entrapment of preexisting microcontinents within the Paleo-Asian Ocean(s) (e.g., Badarch et al., 2002; Kröner et al., 2013, 2014). Similarly, the paleogeographic origin of the oceanic and continental fragments in the Tethyan orogenic system is also enigmatic: were they derived from (1) the northern margin of Gondwana (Stampfli et al., 2013), (2) (at least partially) the southern margin of Laurasia (Şengör et al., 1993; Yakubchuk, 2002), or (3) a continent located within the Paleozoic–early Mesozoic Tethyan oceans (Şengör, 1984)?

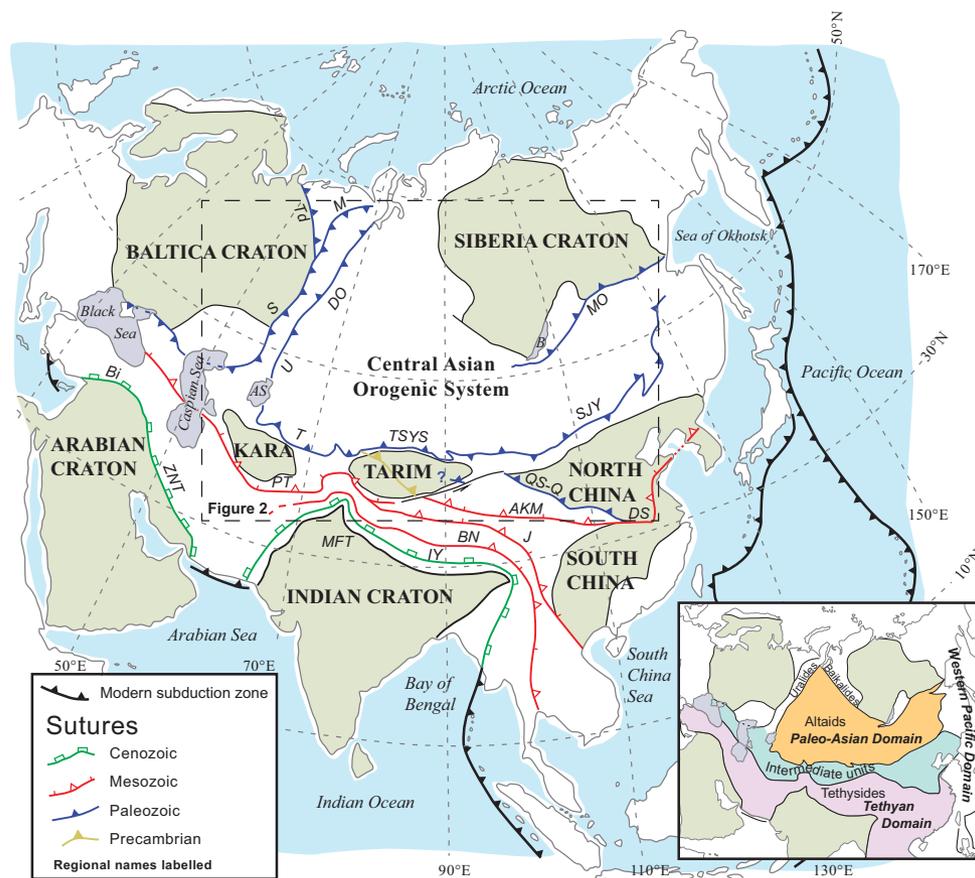


Figure 1. Simplified tectonic map of Asia showing cratons, sutures, and modern subduction zones. Teeth are on overriding plate. For simplicity, the structures and sutures within the Central Asian Orogenic System are omitted. Pre-Neoproterozoic sutures within cratons are not shown. The location of Figure 2 is outlined. Inset map shows major tectonic divisions and oceanic domains. AS—Aral Sea, B—Lake Baikal, Kara—Karakum craton, MFT—Main frontal thrust. Sutures: AKM—Anyimaqen-Kunlun-Muztagh, Bi—Bitlis, BN—Bangong-Nujiang, DO—Denisov-Oktyabrsk, DS—Dabie Shan, J—Jinsha, M—Magnitogorsk, MO—Mongol-Okhotsk, PT—Paleo-Tethys, S—Sakmara, SJY—Solonker-Jilin-Yanji, TSYS—Tian Shan-Ying Shan, T—Turketsan, Td—Timanide, U—Ural, QS-Q—Qilian Shan-Qinling, ZNT—Zagros-Neo-Tethys. Figure after Yin and Nie (1996), Şengör and Natal'in (1996), Natal'in and Şengör (2005), Xiao et al. (2010), and Zheng et al. (2013).

Although extensive research has focused on the evolution of the Central Asian and Tethyan orogens, our understanding of the tectonic history and role of the cratonic units that are between these two orogenic systems, including the Baltica, Karakum, Turan, Tarim, and North China cratons, remains comparatively limited (e.g., Şengör and Natal'in, 1996). This may be attributed to (1) poor bedrock exposure across the region (e.g., Turan, Karakum, and Tarim are extensively covered by Mesozoic–Cenozoic sand-desert deposits) and (2) significant tectonic modification that occurred during and after the closure of the Paleo-Asian and Tethyan Oceans throughout the Phanerozoic (e.g., Windley et al., 1990; Allen et al., 1993, 1995; Yin, 2010). The lack of detailed studies on the Intermediate Units of Şengör and Natal'in (1996) (inset of Fig. 1) has limited our ability to use a geologic process-based approach to reconstruct the tectonic history of the Paleo-Asian and Tethyan orogenic systems in Asia.

In this study we first outline the geology of the Precambrian cratons that separate the Paleo-Asian and Tethyan oceanic domains, including their mutual

tectonic relationships and their deformational history during and after they became individual tectonic entities. Most of the tectonic boundaries currently separating these cratons postdate the closure of the Paleo-Asian and Tethyan Oceans, and we systematically restore the shape of these cratons from their present-day configuration through the Phanerozoic and back into the Proterozoic. By removing the effects of younger tectonic distortion, we show that Baltica, Karakum, Turan, Tarim, and North China were once continuously linked in the Neoproterozoic as a ~6000-km-long continental strip. This contiguous strip of Precambrian continental lithosphere is herein referred to as Balkatach (named after the Baltica–Karakum–Tarim–North China connection of the traditionally defined Precambrian cratons).

The proposed Balkatach hypothesis has implications for the boundary conditions of the development of the three major Phanerozoic oceans dealt with in this study and affects our current understanding of the Proterozoic–Phanerozoic tectonic evolution of Asia and global supercontinent reconstruc-

tions of Columbia-Nuna and Rodinia (e.g., Moores, 1991; Hoffman, 1991; Li et al., 2008; Santosh, 2010). Our analysis of existing data shows how the breakup of Rodinia opened the interconnected Paleo-Asian and Pacific Oceans and how microcontinents were generated within the Paleo-Asian oceanic domain (Fig. 2). These microcontinents may have, in turn, served as nucleation points for subsequent subduction that led to the consumption and eventual destruction of the Paleo-Asian Ocean (e.g., Hsü and Chen, 1999; Briggs et al., 2007; Kelyt et al., 2008). Available geologic data suggest that initial collision of central Balkatach and the amalgamated arcs and microcontinents of the Central Asian Orogenic System (CAOS) in the middle Carboniferous was followed by bidirectional suturing, closure of the Paleo-Asian Ocean, and the development of an independent Pacific Ocean. The closure of the Paleo-Tethys Ocean along the southern margin of Balkatach was diachronous, proceeding from west to east in the Triassic–Jurassic. Our model expands the Wu et al. (2016) Wilson-cycle evolution of the southern margin of Asia into a more global tectonic setting.

METHODS

Paleogeographic reconstructions are based on multiple types of data, including seafloor magnetic, paleomagnetic, geologic, geochronologic, paleontologic, conjugate margin, and geophysical data sets (e.g., Meert, 2014). Although Mesozoic and younger reconstructions can often use reliable seafloor magnetic anomaly data, paleomagnetic poles, and apparent polar wander paths, the lack or limitation of these quantitative tools in the Paleozoic and Precambrian requires a careful analysis and integration of other independent geologic data (e.g., Li et al., 2008). Paleomagnetic data provide important constraints for paleogeography, but basing entire reconstructions of Archean and

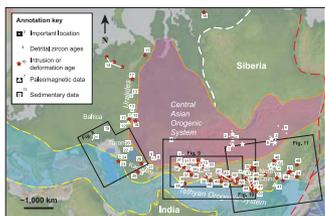
Proterozoic Earth on these data sets is problematic because the temporal and spatial resolution is too low (e.g., Evans, 2009; cf. Evans and Mitchell, 2011).

The purpose of this synthesis and our tectonic reconstruction is to ultimately depart from matching continents and cratons simply by similar basement ages. For example, western and eastern North America belong to the same continent, yet there are very limited Grenville ages in the North American Cordillera basement (cf. Wooden et al., 2013; Powell et al., 2016) and there is no Mesozoic Cordilleran arc signature along the east coast. Thus, expecting to find continent-wide 1.2–1.0 Ga and 200–75 Ma age signatures in present-day North America would not provide useful information about the continuous continent. However, an often discussed correlation between the Tarim and South China cratons is based primarily on the ca. 1.0–0.9 Ga signature observed in both continents. Similarly, it is widely considered that the North China craton could not have been near Tarim in the Paleoproterozoic and Mesoproterozoic because this early Neoproterozoic zircon-age signature (i.e., 1.0–0.9 Ga) is relatively minor in the North China basement. In this contribution, we develop our tectonic reconstruction based on a combination of age constraints, restored continental margin geometries, and plausible known geologic processes that could lead to the observed geology.

Data Synthesis

Our tectonic synthesis incorporates a range of geologic data sets and existing paleogeographic reconstructions, which were plotted together on palinspastically restored base maps. A catalogue of the key data points and sources of information used in our reconstruction can be found in Supplemental Figure¹. Data were organized for each of the geologically significant tectonic domains (e.g., Baltica, Karakum, Turan, Tarim, and North China) (Figs. 2 and 3) and the accretion/collision histories along the margins of these domains were specifically focused on. The locations of domain-specific maps referred to in this work are presented in Figure 3. Events affecting the margins of the cratons are displayed in Figure 4, including collisional orogens, arc magmatism, and passive margin sedimentation.

The reconstruction presented here uses the updated geologic time scale defined by the International Commission on Stratigraphy (ICS) (Gradstein et al., 2012; Cohen et al., 2013). Much of the Chinese literature follows a slightly different time scale put forth by the China Geological Survey (2014; see also Su, 2014). In the Chinese time scale, Phanerozoic periods are similar to the ICS time scale (Cohen et al., 2013), but Proterozoic Era subdivisions differ (e.g., Wang et al., 2013) (Table 1). The Mesoproterozoic is divided into the Changcheng Period (1.8–1.6 Ga), Jixian Period (1.6–1.4 Ga), and a currently unnamed period (i.e., a post-Jixian period) (1.4–1.0 Ga). The Neoproterozoic is divided into the Qingbaikou (1000–780 Ma), Nanhua (780–635 Ma), and Sinian (635–541 Ma) periods. The Nanhua Period, which is broadly correlative to the Cryogenian (Gradstein et al., 2012), is further subdivided into the Lower



Supplemental Figure. Tectonic map of Eurasia showing the outlines of regional domain-specific maps discussed throughout the text, and the locations key data points, regional names, and other information sources used in our reconstruction. Cited references are all listed in the main text.

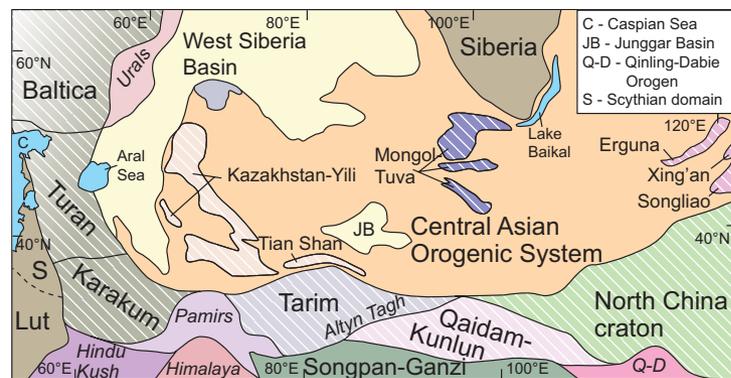


Figure 2. Simplified tectonic map of Central Asia showing the terranes and cratons of our proposed Balkatach (Baltica–Karakum–Tarim–North China; shown with white stripes). Note that some of the Chinese microcontinents (i.e., the Jiamusi, Bureya, and Khanka continents) within the Central Asian Orogenic System are not shown, and are located east of this figure boundary (see Wilde, 2015). See Figure 1 for location.

¹Supplemental Figure. Tectonic map of Eurasia showing key locations, critical observations, and selected references used in this reconstruction. Please visit <http://doi.org/10.1130/GES01463.S1> or the full-text article on www.gsapubs.org to view the Supplemental Figures.

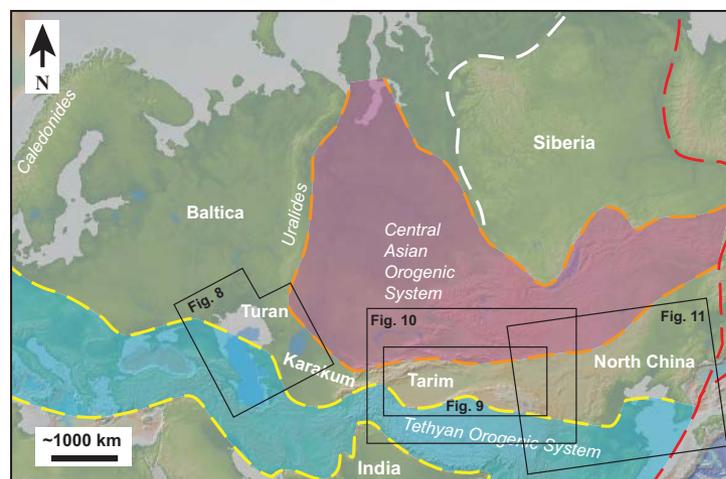


Figure 3. Tectonic map of Eurasia showing the locations of regional domain-specific maps discussed throughout the text, including those in Figures 8–11. The same map overlain with important data points and sources used in our reconstruction can be found in the Supplemental Figure (see footnote 1).

(730–725 Ma), Middle (725–660 Ma), and Upper Nanhua (660–635 Ma) Periods. We convert all of these time units to the ICS convention, as outlined in Table 1.

The tectonic reconstructions and associated kinematic models were developed using Gplates software (www.gplate.org) (Boyden et al., 2010). Eurasian continents were first digitized as undeformable polygons using their present-day shape or estimated geologic boundaries (e.g., Fig. 2) (Ren et al., 2013). Deformable regions were constructed with separate lines that demarcate the geologic boundaries; motion of these lines with respect to time allowed for the restoration of the deformed area along the margins of these polygons, following observed geologic constraints. When applicable, structural information, including the magnitude, timing, and style of deformation, was used to provide a context for the other geological data sets, and when possible, estimated deformation was restored.

Restoring Tectonic Modification of Precambrian Cratons

An important issue for Proterozoic tectonic reconstructions is the effect of later Neoproterozoic–Phanerozoic tectonic modification of the original shape and positions of key tectonic elements within and along the margins of Precambrian cratons. These effects include duplication and elimination of early arc systems (e.g., the Cordilleran arc; see Dickinson, 2004, for discussion), post-Rodinia rifting and drifting of microcontinents, and intracontinental deformation as induced by continental collision and ocean-closure orogeny,

which are particularly significant in modifying the Neoproterozoic shape of the North China, Tarim, and Karakum cratons (Fig. 2). The original configuration of Precambrian continents can be inferred by either (1) recognizing correlative marker horizons and tectonic boundaries within and across deformed regions, or (2) estimating the magnitude and style of later tectonic modification to indirectly restore the initial configuration. For example, Cowgill et al. (2003) used the Permian–Triassic Kunlun arc as offset piercing points to estimate the magnitude of Cenozoic intracontinental deformation (i.e., 475 ± 70 km of left-lateral offset along the Altyn Tagh fault).

The shape and position of the North China craton in popular Rodinia reconstructions (e.g., Zhang et al., 2006b; Pei et al., 2006; Whitmeyer and Karlstrom, 2007; Li et al., 2008; Chen et al., 2013a) is a striking example of how later post-Rodinia deformation is ignored. North China is generally placed outside of the inferred Rodinia supercontinent, and its shape is limited to the current geographically defined northern China, north of the Qinling–Dabie orogenic belt (Fig. 1). Three fundamental issues exist with this configuration: (1) North China was much larger in the Precambrian, extending as far west (in present-day coordinates) as Karakum or Baltica (Fig. 2) (Yin and Nie, 1996; Heubeck, 2001) (this issue is further discussed in this study), (2) Phanerozoic deformation (e.g., accretionary and collisional events in the Paleozoic across Central Asia and large-scale intracontinental deformation associated with the India–Asia collision; Şengör and Natal'in, 1996; Yin, 2010) significantly modified the original shape of the craton, and (3) Paleoproterozoic and Mesoproterozoic structures are truncated by Neoproterozoic rifts and passive margins (Zonenshain et al., 1990; Guo et al., 2005; Zhao et al., 2005; Kusky et al., 2007), which requires the North China craton to fit into a larger continental assemblage prior to this time.

Establishing the Assembly Histories of Tectonic Domains

The recognition and interpretation of suture zones (Dewey, 1977) are critical in evaluating the evolution of the continents on Earth. A suture refers to a zone along which the collision of two preexisting pieces of continental lithosphere and the destruction of an interlying oceanic lithosphere has occurred (e.g., Burke et al., 1976; Şengör and Natal'in, 1996). Implicit in this definition, the convergence and subsequent amalgamation of two continents require the following: (1) a subduction zone (or zones) that consumes the interlying oceanic lithosphere, possibly involving a subduction-accretion complex along the suture zone (e.g., Dewey, 1976; Sample and Fisher, 1986; Fisher and Byrne, 1987; Stern and Bloomer, 1992); (2) a calc-alkaline volcanic-plutonic belt that results from the subduction of the oceanic lithosphere; (3) a collisional orogen, complete with an accretionary wedge-mélange zone, foreland basin, fold-thrust belt, and Barrovian metamorphic core, that results from the convergence of two continents (e.g., the Himalaya; see Yin, 2006, for a review); and (4) a regional unconformity and/or a transition to terrestrial sedimentation due to regional uplift associated with this collision (e.g., Yin and Nie, 1993; Xiao et al., 2009c; Burchfiel and Chen, 2012).

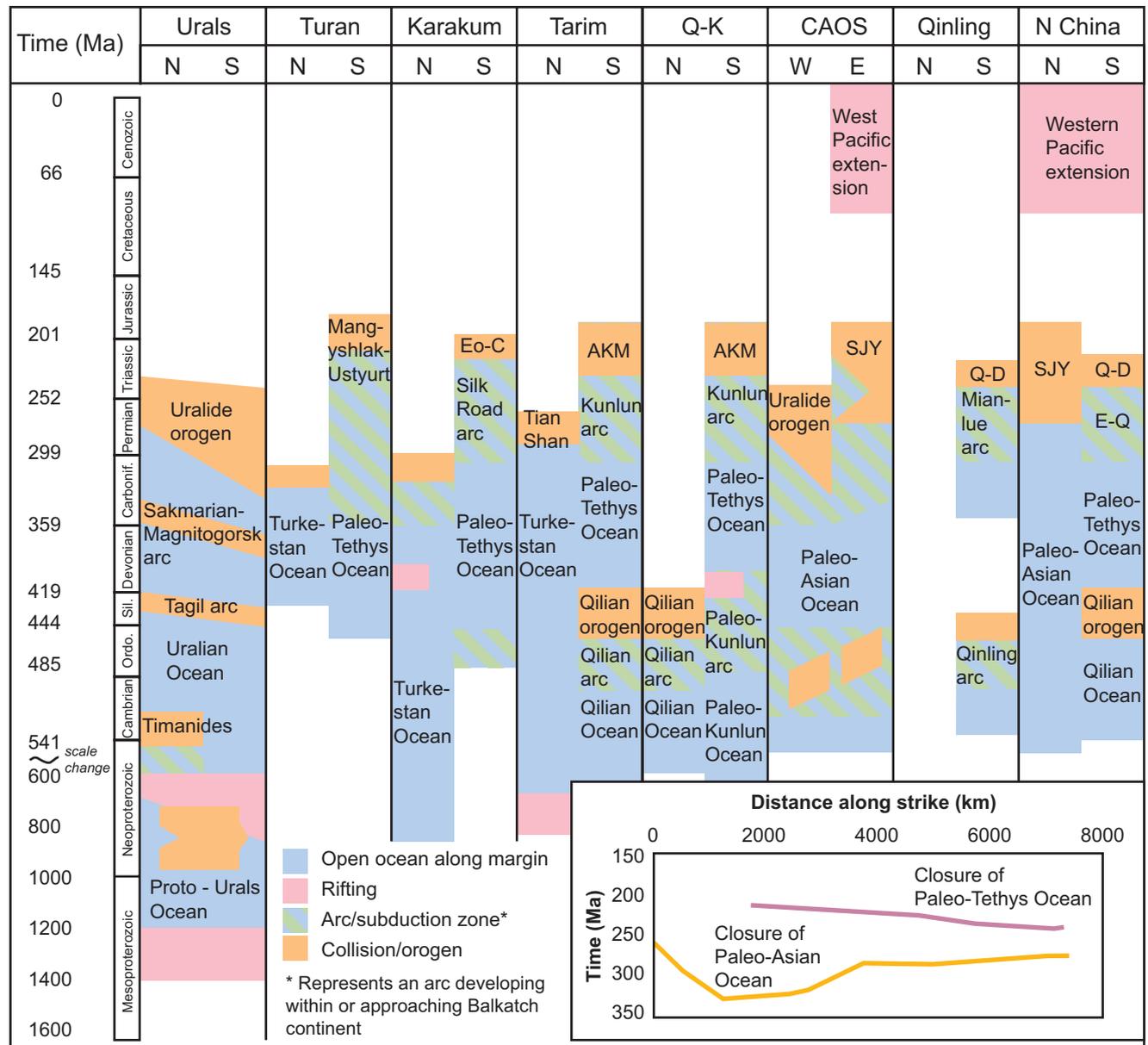


Figure 4. Tectonic correlation chart displaying major tectonic events along the northern (N) and southern (S) margins of the constituent cratons of Balkatach (Baltica–Karakum–Tarim–North China). Inset shows approximate closure time of the Paleo-Asian and Tethyan Oceans, to the north and south of Balkatach, respectively, which are based on the earliest collisional ages observed in the geologic record. Abbreviations: AKM—Anyimaqen-Kunlun-Muztagh, CAOS—Central Asian Orogenic System, E-Q—Earlangping-Qinling arc, Eo-C—Eo-Cimmerian, Q-D—Qinling-Dabie Shan orogen, Q-K—Qaidam-Kunlun, and SJY—Solonker-Jilin-Yanji. Sources are discussed in text.

TABLE 1. COMPARISON OF TIME NOMENCLATURE COMMONLY USED IN THE CHINESE LITERATURE (CHINESE GEOLOGICAL SURVEY, 2014) WITH THE INTERNATIONAL COMMISSION OF STRATIGRAPHY (ICS) TIMESCALE (COHEN ET AL., 2013)

Eon	Era	ICS Period	Age (Ma)	Approximate Chinese Period	
Proterozoic	Neoproterozoic	Ediacaran	635	Sinian	
		Cryogenian		Nanhua	
		Tonian	720	Qingbaikou	
	Mesoproterozoic		Stenian	1000	Currently unnamed
			Ectasian	1200	
			Calymmian	1400	Jixian
				1600	
	Paleoproterozoic			1800	Changcheng
				2500	

Every verified Phanerozoic continent-continent and arc-continent collision consists of these features (e.g., Taiwan, Himalaya-Tibet, Uralides, or Zagros) (e.g., Molnar and Tapponnier, 1975; Suppe, 1984; Yin, 2010; Burchfiel and Chen, 2012). It has been demonstrated that the presence of oceanic material alone is not enough to demarcate a unique suture that juxtaposes distinct pieces of lithosphere (e.g., Yin, 2002; Kapp et al., 2003; Yin et al., 2007a) and the tectonic setting can only be fully realized when the geologic context and geologic history of the region involving oceanic materials are restored and understood. If the package of the aforementioned suture features is missing, an amalgamation of two distinct continental lithospheres cannot be inferred without an adequate alternative explanation. Although it has been debated whether present-day plate tectonics processes operated in the Proterozoic and Archean (e.g., Kröner, 1983; Cawood et al., 2006; Moore and Webb, 2013), we herein assume that the opening and closing of major ocean basins and convergence of distinct pieces of continental lithosphere since at least the late Paleoproterozoic operated via processes similar to those that occurred throughout the Phanerozoic (Glassley et al., 2014; Weller and St-Onge, 2017).

Related to the above issues, there is no geologic evidence for a Phanerozoic collision between the Karakum, Tarim, and North China cratons (Fig. 2). In light of this, we prefer the simplest explanation and infer that these continents were contiguous by the start of the Phanerozoic, rather than make up a collisional history that left no geologic record. Large-scale transform/strike-slip faults and intracontinental shear zones can reposition continental lithosphere (e.g.,

microcontinents and oceanic terranes) without the direct closure of an ocean (e.g., the Cenozoic translation of southeast Asia during the India-Asia collision and the movement of the California borderlands along the San Andreas fault in the western United States), although many of these present-day examples are related to nearby subduction-collision tectonics. Without direct evidence for such structures, they should not be inferred. For example, although an important primary prediction for the Şengör et al. (1993) strike-slip duplication of the Kipchak arc within the CAOS is numerous lithosphere-scale strike-slip faults, faults that were previously conjectured to be strike-slip faults have more recently been reinterpreted as thrust faults (Briggs et al., 2007, 2009; cf. Şengör et al., 2014).

Although our restoration attempts to conform to all existing geologic data sets, the weight of a given geologic data point varies significantly depending on clarity of its geologic content and its compatibility to be integrated into coherent and plausible geologic processes. When contrasting structural models can explain the same sets of geologic observations, the simplest, globally consistent reconstruction was chosen. Because of the nonuniqueness of interpreting the data, we take an iterative, process-based approach to develop our paleogeographic reconstructions so that each step can proceed forward and backward in time and be explained by established geologic processes.

Paleomagnetic Considerations

Many reconstructions consider paleomagnetism as the principal line of evidence for assessing the paleogeographic location of Precambrian continents (e.g., Li et al., 2008; Evans, 2009; Meert, 2014), but the goal of this work is to focus on geologic relationships. Issues of ambiguous hemisphere polarity, unconstrained latitude, paucity of reliable high-resolution data from a particular region, poor temporal resolution, and rigid-block approximation require a conservative approach when using paleomagnetic analyses to reconstruct the internally deformed units of Central Asia back to the Proterozoic. Future work may improve the spatial and temporal resolution of paleomagnetic poles and may lead to better constrained paleogeographic positions. We ensure that our reconstruction fits within the context of existing paleomagnetic data sets (Table 2), but note that significant quantitative conclusions cannot be drawn because there are too few reliable paleomagnetic poles for many of the central Asian cratons.

■ GEOLOGY OF THE MAJOR PRE-MESOZOIC TECTONIC DOMAINS OF CENTRAL ASIA

The assembly of present-day Central Asia has progressed through the Proterozoic and Phanerozoic as an amalgamation of numerous cratons, continental blocks, island-arc fragments, and accretionary complexes (e.g., Şengör et al., 1993; Yin and Nie, 1996; Şengör and Natal'in, 1996; Burchfiel and Chen,

TABLE 2. SELECTED PALEOMAGNETIC POLES AND PALEOLATITUDES FOR CENTRAL ASIAN CONTINENTS

Rock unit	Age (Ma)	Pole		α_{95} (°)	Paleolatitude (°N)	Reference
		(°N)	(°E)			
<u>North China</u>						
Fengfeng, Pingliang Formations	480	37.4	324.3	8.5	−10	Huang et al. (1999)
Changshan, Gushan Formations	500	31.7	329.6	5.1	−13	Huang et al. (1999)
Zhangxia, Xuzhuang Formations	520	37.0	326.7	5.5	−10	Huang et al. (1999)
Dongjia Formation	650	−60.8	97.4	6.7	10	Zhang et al. (2006b)
Huaibi Group	700	−42.9	107.0	5.7	8	Zhang et al. (2006b)
Nanfen Formation	800–780	−16.5	121.1	11.1	35	Zhang et al. (2006b)
Wangshan Formation	890	52.6	330.0	5.3	0–5	Fu et al. (2015)
Jingeryu Formation	950	−41.0	44.8	11.3	10	Zhang et al. (2006a)
<u>North Tarim</u>						
Sugetbrak, Chigebrak Formations	600	19.1	149.7	9.3	27	Zhang et al. (2007c)
Qiaoenbrak Formation	730	30.4	246.1	11.0	6.3 ± 39	Wen et al. (2013)
Baiyixi Formation	740	17	194	4	6	Huang et al. (2005)
Wushi Formation	800–590	−55	53	7.9	8 ± 6	Li et al. (1991)
Aksu dikes	807 ± 12	19	128	6	43	Chen et al. (2004)
Sugetbrak Formation	625	21.1	87.4	7.0	27.4 ± 5.6	Wen et al. (2016)
<u>CAOS microcontinents</u>						
Zavkhan (Baydaric or Dzabkhan)	805–770	n.d.	n.d.	n.d.	47 ± 14	Levashova et al. (2010)
Karatau	766 ± 7	n.d.	n.d.	n.d.	34.2 ± 5.3	Levashova et al. (2011)
<u>Siberia</u>						
Karagas Series	850–740	−12	97	10	22–8	Metelkin et al. (2005)
Nersinsky Complex	740	−37	122	11	2 to −19	Metelkin et al. (2005)
<u>Baltica</u>						
Zigan Formation	550	−15.4	107.7	4.3	7.8 ± 2.5	Levashova et al. (2013)
Average*	560–550	n.d.	n.d.	n.d.	30–40	Meert (2014)
Kildinskaya Formation	750	−26	13	7.2	~0	Torsvik et al. (1996)

Note: CAOS—Central Asian Orogenic System. n.d.—no data provided.

*Average paleolatitude from numerous data sets.

2012). Şengör and Natal'in (1996) divided Central Asia into the Altaids (or CAOS) (e.g., Briggs et al., 2007, 2009; Kelty et al., 2008) and the Tethysides, in the north and south respectively, with the Intermediate Units occupying the space between (Fig. 1). These units are surrounded by three large Phanerozoic oceans: the Paleozoic Paleo-Asian Ocean to the north, Tethyan Ocean to the south, and present-day Pacific Ocean to the east (Fig. 1) (Stampfli, 2000; Zheng et al., 2013). This spatial division is also a temporal one, as the Intermediate Units docked with Eurasia after the predominantly Paleozoic assembly of the CAOS but prior to their collision with the Mesozoic Tethysides.

Below we outline the regional geologic framework of each of the Intermediate Units while focusing on the (1) Precambrian and Paleozoic assembly history of the basement and (2) later tectonic modification of each unit throughout the Phanerozoic. We provide only cursory discussion of Mesozoic and Cenozoic events, as this most recent time period is covered considerably in the

existing literature (e.g., Molnar and Tapponnier, 1975; Tapponnier et al., 1982; Hendrix et al., 1992; Yin and Harrison, 2000; Yin, 2010; van Hinsbergen et al., 2011). The following synthesis begins with the CAOS, continues with the eastern margin of Baltica in western Asia, and progresses eastward to the North China craton (Figs. 1 and 2).

CAOS

The CAOS (Briggs et al., 2007, 2009; Kelty et al., 2008), also referred to as the Altaids (Suess, 1901; Şengör et al., 1993; Şengör and Natal'in, 1996; Xiao et al., 2008, 2009a, 2009b, 2010), the Central Asia Foldbelt (Zonenshain et al., 1990; Filippova et al., 2001), and the Central Asian Orogenic Belt (e.g., Mosakovsky et al., 1993; Jahn et al., 2004; Xiao et al., 2003; Windley et al., 2007),

was the largest Phanerozoic accretionary orogen on Earth. We prefer the name Central Asian Orogenic System (e.g., Briggs et al., 2007, 2009; Kelty et al., 2008) to Central Asian Orogenic Belt (the nomenclature of which is discussed in Xiao et al., 2003) because regardless of which tectonic model is most applicable, the region is clearly not a single well-defined belt, but rather a complex accretionary system of arcs and microcontinents. This immense collage is currently bounded to the north and south by the Siberia and North China–Tarim cratons, respectively, and extends from the Urals and Baltica in the west to the Pacific Ocean in the east (Figs. 1 and 2). Most models for the development of the CAOS incorporate a complicated and long-lived arc-continent collisional process that involves the successive amalgamation of Precambrian microcontinents and island arcs that began ca. 1.0 Ga (Khain et al., 2003), or as early as the late Mesoproterozoic (Kröner et al., 2013), and concluded by the late Permian (e.g., Philippova et al., 2001; Windley et al., 2007; Xiao et al., 2003).

The various tectonic reconstructions of the CAOS can be summarized by two distinct models: (1) duplication and expansion of single long-lived arc (Şengör et al., 1993; Yakubchuk, 2002) or (2) multiple microcontinents and arcs within the Paleo-Asian Ocean (Zonenshain et al., 1990; Philippova et al., 2001; Badarch et al., 2002; Bold et al., 2016). The first model requires Siberia and Baltica to be linked in the Neoproterozoic and the Precambrian basement rocks within the CAOS to be derived from Siberia or Baltica to the north (present-day coordinates), whereas the second requires Neoproterozoic Baltica and Siberia to be separate and that the CAOS microcontinents are predominately derived from Gondwana or Tarim.

Some of the major microcontinents within the CAOS include, from west to east, the Ishim–Middle Tian Shan continent of Kazakhstan, other massifs within the Tian Shan, Tuva–Mongol, Zavkhan (Baydaric or Dzabkhan), Erguna, Xing’an, Songliao, and Bureya–Jiamusi–Khanka (Fig. 2), as well as other less substantiated fragments (e.g., Avdeev, 1984; Windley et al., 2007; Demoux et al., 2009; Biske and Seltmann, 2010; Zhou and Wilde, 2013; Wilde and Zhou, 2015; Wilde, 2015; Bold et al., 2016; Zhu et al., 2017). These microcontinents may have been derived from Siberia, Tarim, North China, South China, or the northern margin of Gondwana (e.g., Zonenshain et al., 1990; Kheraskova et al., 2003, 2010; Khain et al., 2003; Li, 2006; de Jong et al., 2006; Dobretsov and Buslov, 2007; Zhou et al., 2009, 2011; Levashova et al., 2010, 2011; Han et al., 2011; Meert et al., 2011; Rojas-Agramonte et al., 2011, 2014; Kröner et al., 2014; Zhou et al., 2017). Kazakhstan is often considered to be a large conglomeration of several microcontinents that developed toward the end of the CAOS activity (Zonenshain et al., 1990; Philippova et al., 2001; Windley et al., 2007). Treating the collision of Baltica, Tarim, and Karakum against a single Kazakhstan is a useful simplifying assumption in global tectonic reconstructions (e.g., Ziegler, 1989; Philippova et al., 2001; Windley et al., 2007; Xiao et al., 2015).

Similarities between many of the microcontinents include (1) Proterozoic or older basement, (2) regional metamorphism at ca. 1.85–1.80 Ga, (3) magmatic activity at ca. 950 Ma, (4) rift features from 790 Ma to 750 Ma, and (5) the development of circumscribing passive margins by the Cambrian time. Detrital zircon ages older than 750 Ma shown in Figure 5 (Han et al., 2011; Rojas-Agramonte

et al., 2011) highlight similar age signatures among the microcontinents. The oldest exposed rocks of the Lesser Karatau block of Kazakhstan (Levashova et al., 2011) consist of Paleoproterozoic flysch (Sovetov, 1990). The Tuva–Mongol microcontinent consists of several Archean–Paleoproterozoic granulite facies terranes covered by lower grade Mesoproterozoic to Cambrian schist, marble, and quartzite metasedimentary strata (e.g., Badarch et al., 2002; Bold et al., 2016). Tonalitic gneiss of the Gargan massif has a whole-rock Rb–Sr isochron age of 3.2 Ga (Aktanov et al., 1992), and basement rocks of the Bumberger complex have U–Pb zircon ages of 2650–2364 Ma (Kotov et al., 1995; Kozakov et al., 1999). Granulite facies metamorphism occurred at ca. 1840 Ma, as determined by U–Pb zircon rim dating (Demoux et al., 2009). The Zavkhan (Baydaric or Dzabkhan) microcontinent, to the south of the Tuva–Mongol microcontinent, consists of tonalitic gneiss with a U–Pb zircon age of 2646 ± 45 Ma (Kozakov et al., 1993) overlain by Paleoproterozoic metasedimentary rock. This basement is intruded by Paleoproterozoic granite (Kotov et al., 1995) and the region underwent granulite facies metamorphism at 1850–1800 Ma (Khain et al., 2003).

The microcontinents to the east within China appear to have relatively limited Proterozoic or older basement (see review by Zhou et al., 2017, and

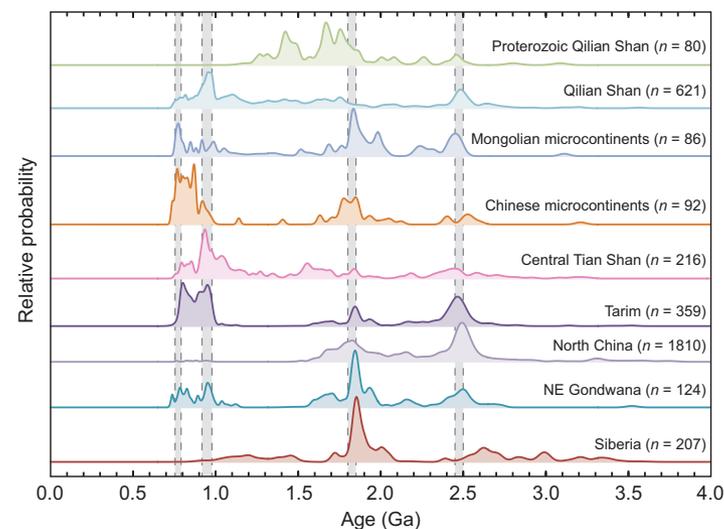


Figure 5. Normalized probability plot of zircon ages older than 750 Ma from the Qilian Shan, Mongolian and Chinese microcontinents, Tarim, North China, northeast Gondwana, and southern Siberia. Note that we excluded Phanerozoic and late Neoproterozoic grains from this plot because our goal was to assess continental links prior to the onset of global rifting at ca. 750 Ma. U–Pb ages were used for zircons younger than 1000 Ma and Pb–Pb ages were used for older zircons. Gray shaded zones denote prominent age peaks at ca. 790–760 Ma, 970–910 Ma, 1850–1800 Ma, and 2500–2450 Ma that correspond to widespread tectonic events as discussed in the text. Chinese microcontinent ages are from Han et al. (2011) and Mongolian microcontinent data are from Rojas-Agramonte et al. (2011). Qilian Shan data are compiled from our own work (see Wu et al., 2016, 2017a) and Xu et al. (2010). Tarim, North China, and southern Siberia data were compiled by Rojas-Agramonte et al. (2011). Northeast Gondwana data were compiled by Squire et al. (2006).

references therein). The Erguna, Xing'an, and Songliao continents (Fig. 2) have yielded Archean–Paleoproterozoic zircons (Wang et al., 2006; Zhou et al., 2012) and have minor confirmed components of Archean–Proterozoic basement (e.g., Wang et al., 2006; Han et al., 2011; Sun et al., 2013; Pan et al., 2014; Shao et al., 2015; Zhou et al., 2017). Further to the east, the Bureya–Jiamusi–Khanka microcontinents represent the easternmost boundary of the CAOS (e.g., Şengör et al., 1993; Jahn et al., 2004). The oldest known rocks in these continents are Mesoproterozoic–Neoproterozoic in age (Wilde et al., 2003), and ca. 1.0–0.9 Ga detrital zircons are prominent within metasedimentary rocks (Zhou et al., 2010, 2017).

Early Neoproterozoic (i.e., Tonian) magmatism and metamorphism (970–910 Ma) observed in many of the CAOS microcontinents (Fig. 5) were followed by late Neoproterozoic (i.e., Cryogenian–Ediacaran) bimodal volcanism (e.g., Zonenshain et al., 1990; Kuzmichev et al., 2005; Zhou et al., 2017) and passive-margin deposition (e.g., Markova, 1982; Alexeiev et al., 2000; Cook et al., 2002) throughout Central Asia (Fig. 6) (e.g., Meert et al., 2011; Rojas-Agramonte et al., 2011, 2014; Bold et al., 2016). The microcontinents are currently surrounded by Neoproterozoic to early Cambrian ophiolite, ophiolitic mélangé, and/or shelf carbonate, which suggests that they were surrounded by an open ocean at the time (i.e., the Paleo-Asian Ocean) (Ren et al., 1999; Khain et al., 2003; Demoux

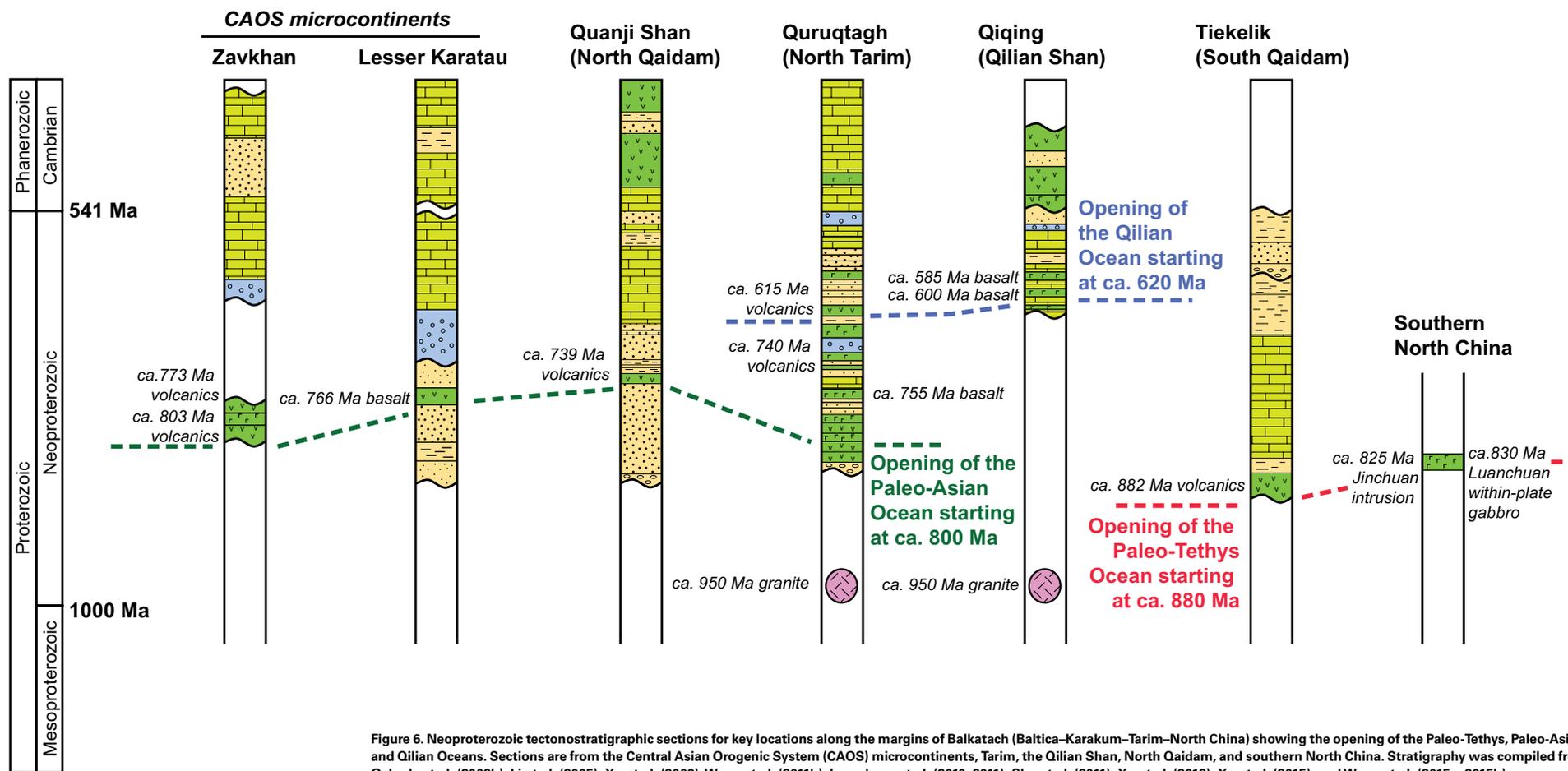


Figure 6. Neoproterozoic tectonostratigraphic sections for key locations along the margins of Baltica (Baltica–Karakum–Tarim–North China) showing the opening of the Paleo-Tethys, Paleo-Asian, and Qilian Oceans. Sections are from the Central Asian Orogenic System (CAOS) microcontinents, Tarim, the Qilian Shan, North Qaidam, and southern North China. Stratigraphy was compiled from Gehrels et al. (2003b), Li et al. (2005), Xu et al. (2009), Wang et al. (2011b), Levashova et al. (2010, 2011), Shu et al. (2011), Xu et al. (2013), Xu et al. (2015), and Wang et al. (2015a, 2015b).

et al., 2009). Cryogenian–Ediacaran glacial tillite-bearing deposits have been documented in the Dzabkhan terrane (Lindsay et al., 1996; Macdonald et al., 2009; Meert et al., 2011) (Fig. 6). Many of the microcontinent basements and their early Paleozoic passive margin cover sequences were modified by later subduction-accretion processes, as evidenced by the intervening Paleozoic accretionary complexes, subduction-related volcanic rocks, and flysch deposits.

Detrital zircon age distributions reveal age populations at ca. 2.5 Ga, 1.85–1.8 Ga, 970–910 Ma, and 790–750 Ma that are common to all of the microcontinents within the CAOS (Fig. 5) (also see Zhou et al., 2017). The peaks at ca. 2.5 Ga broadly correspond to a global period of continental growth at the end of the Archean (e.g., Kröner et al., 2005a, 2005b; Condie et al., 2009; Yao et al., 2011). The 1.85–1.80 Ga and 970–910 Ma (Tonian age) peaks may have correlations with other Asian cratons (e.g., Tarim, Qilian-Qaidam, North China, and South China; Ling et al., 2003; Gehrels et al., 2003a, 2003b; Li et al., 2009; Tung et al., 2013), and the late Neoproterozoic–aged zircons correspond to the rifting of these microcontinents (Fig. 6).

A Precambrian link between Tarim and the Central Tian Shan terrane has been proposed based on stratigraphic and detrital zircon age correlation (Fig. 5) (Shu et al., 2011; Gao et al., 1998; Charvet et al., 2007; Lei et al., 2011; Meert et al., 2011; Ma et al., 2011, 2012) (cf. Hu et al., 2000; Liu et al., 2004; Xiao et al., 2013). Granitic orthogneisses in the Tian Shan yielded U–Pb zircon ages of 942–919 Ma (Tonian age) (Chen et al., 2009; Hu et al., 2010) that may be correlative to the plutonic rocks of similar ages in central-northern Tarim and the Qilian-Qaidam-Kunlun regions, as discussed below.

Baltica

The 3000-km-wide Baltica craton is typically defined with its core as the East European platform (e.g., Zonenshain et al., 1990; Filippova et al., 2001), which consists of a collage of Precambrian cratons commonly referred to as the Fennoscandia, Sarmatia, and Volgo-Uralia blocks (Bogdanova et al., 2008). Its surrounding Neoproterozoic–Phanerozoic margins are marked by the Uralide and Caledonian orogens (Fig. 3; e.g., Torsvik et al., 1996). The craton was affixed to Laurentia first in the Neoproterozoic as part of Rodinia (e.g., Hartz and Torsvik, 2002; Li et al., 2008; Pease et al., 2008) and then at the end of the Paleozoic as part of Laurussia (e.g., Ziegler, 1989; Cocks and Torsvik, 2011).

We do not provide an in-depth review of this craton, a subject well covered in the existing literature (see Bogdanova et al., 2008, and references therein), but rather focus solely on the geology of the eastern margin of Baltica (e.g., Fig. 7), which is most relevant to the tectonic synthesis of this study. This eastern margin has been suggested to extend to the southeast to the Scythian domain and underneath the Peri-Caspian Basin (Pease et al., 2008) (Fig. 8). Geology in much of this region is exposed as a result of the late Paleozoic–early Mesozoic Uralide orogen, which records the collision between Baltica, Siberia, and Kazakhstan during the final closure of the western part of the Paleo-Asian Ocean (Zonenshain et al., 1990; Şengör and Natal'in, 1996).

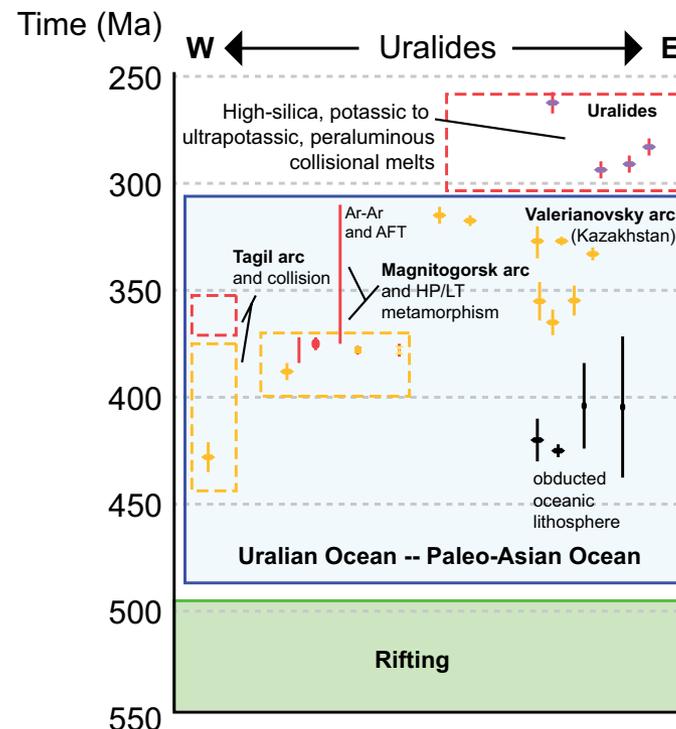


Figure 7. Timing of major Paleozoic events relative to west-east position in the Uralides, based on geochronology, thermochronology, and sedimentological studies. Dashed boxes are timing constraints from the sedimentary record. Primary data sources include Bea et al. (1997, 2002), Fershtater et al. (1997, 2007), Savelieva et al. (2002), and Fershtater (2013). AFT—apatite fission track; HP-LT—high pressure–low temperature.

The 2000-km-long Uralides consist of several longitudinal terranes that extend parallel to the eastern margin of Baltica. These tectonic domains are variably divided, from west to east, as the (1) undeformed foreland basin, (2) foreland fold-thrust belt, (3) Tagil-Magnitogorsk zone, and (4) East Uralian zone (Hamilton, 1970; Berzin et al., 1996; Maslov et al., 1997; Puchkov, 2009; Puchkov et al., 2013; Brown et al., 2011). These zones record Paleozoic subduction, arc magmatism, island-arc amalgamation, and the final collision of Baltica with Siberia and Kazakhstan. Preexisting structures within the Precambrian basement are at a high angle to these tectonics domains. A northwest-southeast structural grain within Archean and Proterozoic rocks is composed of several basement highs and lows, including from north to south respectively, the Mid-Russian aulacogen, Perm-Bashkir Arch, Kaltasin (or Kama-Belsk) aulacogen, Tatar Arch, Sernovodsk-Abdulino aulacogen, and Orenburg Arch (Bogdanova et al., 2008; Brown et al., 2011). The aulacogens are filled with as much as 15 km of Protero-

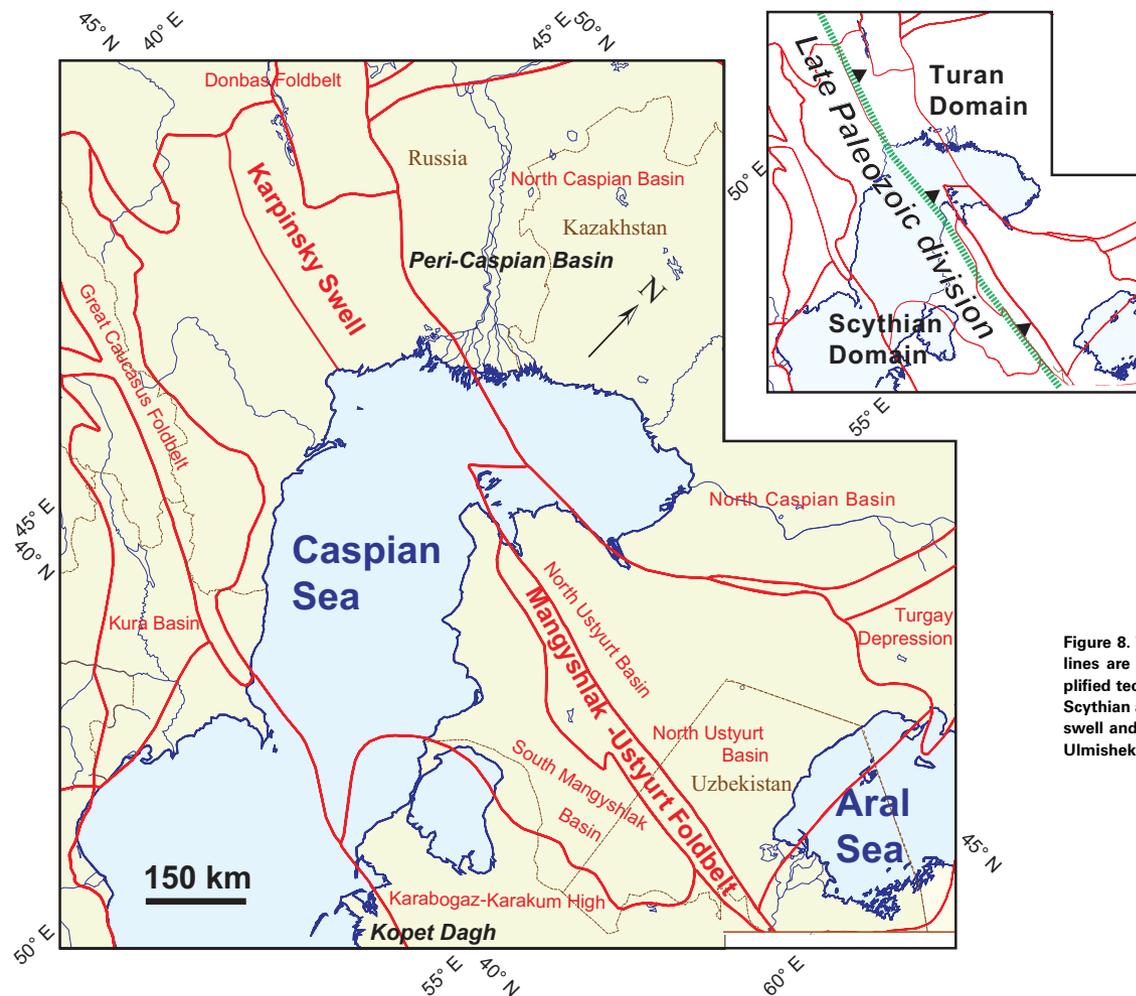


Figure 8. Tectonic units of the Caspian Sea region. Red lines are major tectonic boundaries. Inset shows simplified tectonic map with inferred division between the Scythian and Turan domains, outboard of the Karpinsky swell and Mangyshlak-Ustyurt fold belt. Modified from Ulmishek (2001a, 2001b).

zoic sediments that record ~200 m.y. of variably interrupted intracontinental rifting during the Neoproterozoic (Maslov et al., 1997; Brown et al., 1999).

Early Mesoproterozoic rifting led to initial formation of these aulacogens and the entire region remained an east-facing passive margin along the western boundary of the proto-Urals Ocean (present-day coordinates) (Mossakovsky et al., 1998; Kheraskova et al., 2001). Throughout the Neoproterozoic, several terranes and/or continents collided with and accreted along the eastern margin of Baltica (e.g., Brown et al., 1996; Glasmacher et al., 2001); these events are poorly constrained because the geology recording them was later overprinted by late Neoproterozoic–early Paleozoic rifting and the end-Paleo-

zoic Uralide deformation (Fig. 7) (Brown et al., 1996). Most workers suggest that one or two terranes (e.g., the Bashkirian terrane) accreted to the region during a proto-Ural orogen prior to the Timanian orogen along the north-northeastern margin at 615–575 Ma (Puchkov, 1997; Glasmacher et al., 1999). The geologic relationships of these features are not well understood and it is also possible that these accreted terranes were part of a larger continent that later rifted from Baltica in the latest Neoproterozoic to early Paleozoic.

Late Neoproterozoic–Cambrian rifting (Khain, 1985) and associated bimodal volcanism (e.g., Abdullin et al., 1977; Zonenshain et al., 1984) led to the opening of the Ural Ocean by the early Ordovician (Zonenshain et al., 1984;

Burtman et al., 2000). Ophiolites with ages of 670 Ma and 580 Ma (Khain et al., 1999; Scarrow et al., 2001; Remizov and Pease, 2004) (Fig. 7) have been reported in this region and may represent oceanic crust that was obducted during the later Uralide orogen. Early Paleozoic sediments along this margin are typical shelf passive margin sequences.

The north-northeastern margin of Baltica was affected by the late Neoproterozoic–early Cambrian Timanide orogen, when the inferred composite Arctic Alaska continent (or Arctica; Zonenshain et al., 1990) (Strauss et al., 2013, 2017) collided with Baltica (e.g., Gee and Pease, 2004; Gee et al., 2006; Kuznetsov et al., 2007, 2009, 2010). The Cambrian collision is predated by calc-alkaline magmatism in the Bolshezemel domain of Arctic Alaska (i.e., the Bolshezemel arc), with ages ranging from 700 Ma to 560 Ma (Gee and Pease, 2004; Dovzhikova et al., 2004).

Baltica's eastern margin along the Ural Ocean remained passive throughout much of the Paleozoic, with the exception of several poorly understood arc accretion events (Fig. 7). The Tagil and Magnitogorsk oceanic arcs collided with the eastern margin of Baltica in the Devonian and Carboniferous, respectively (Fig. 7) (e.g., Herrington et al., 2002; Brown et al., 2006, 2011). In the late Carboniferous, the Ural Ocean began to subduct underneath Kazakhstan to the east (Bea et al., 2002). Thick andesitic volcanic deposits in the Balkash area of Kazakhstan (Dercourt et al., 2000) record the initiation of subduction magmatism as part of the Valerianovsky arc (Fig. 7). This subduction system accommodated the convergence of Kazakhstan and Baltica, which led to collision and closure of the Ural Ocean at ca. 320 Ma (Artyushkov and Baer, 1983; Ronov et al., 1984; Zonenshain et al., 1984) (Fig. 4). This collision was diachronous (Zonenshain et al., 1984; Ronov et al., 1984), initiating by ca. 320 Ma in the south (Puchkov, 2000) and propagating northward to the Novaya Zemlya foreland basin by ca. 280 Ma (Heafford, 1988). Syncollisional granite ages also young northward, from 305 Ma to 290 Ma in the south to ca. 265 Ma in the central Urals (Fig. 7) (Fershtater et al., 2007; Fershtater, 2013). An extensive west-migrating foreland basin developed in the late Carboniferous and persisted through the early Permian (Nikishin et al., 1996; Proust et al., 1998; Chuvashov and Crasquin-Soleau, 2000), and Permian flysch transitioned to late Permian–Early Triassic molass and continental deposits. The Taimyr fold belt to the north records the collision of Baltica with Siberia during the same time (Vernikovskiy, 1997; Torsvik and Andersen, 2002), and it is possible that this subduction system was connected to the North American Cordilleran arc-trench system in late Paleozoic–Mesozoic time (e.g., Zonenshain et al., 1987; Miller et al., 2011).

Turan

The Turan Platform is located east-northeast of the Caspian Sea and extends from the southern margin of the CAOS in the north to the Kopet Dagh (Dagh means mountains in Turkish) and Alborz Mountains in the south (Alavi, 1991; Thomas et al., 1999a; Natal'in and Şengör, 2005) (Fig. 8). The pre-Mesozoic basement in the region is known as the Turan domain (Thomas et al.,

1999a), whereas the Turan platform demarcates the extent of the Mesozoic–Cenozoic sedimentary cover. Most of the Turan domain is composed of Paleozoic strata, and Precambrian basement has not been verified. However, Mesozoic–Cenozoic deposits in the North Ustyurt Basin of Kazakhstan and Uzbekistan (Fig. 8) overlie basement rocks that have been variably assigned ages of Paleoproterozoic (Khain, 1977), Neoproterozoic (Milanovsky, 1987), or late Paleozoic (Letavin, 1980). Drilling data suggest that at least some of the basement rocks are Devonian or older in age (Ulmishek, 2001b). The limited basement exposure in this region has led to inconsistencies in the literature regarding the tectonic boundaries and geologic context of the domain. The Karakum craton is to the east-southeast and Scythian domain is to the west (Figs. 2 and 8), and some have suggested that the Turan and Scythian domains were already joined as a single continental block throughout the Paleozoic (e.g., Natal'in and Şengör, 2005). However, as described here, we advocate that the two continental domains were separated by an early Paleozoic ocean (e.g., Brunet et al., 1999; Volozh et al., 2003). The divide between these units is poorly exposed and is often arbitrarily set geographically along the Caspian Sea rather than along a definite geologic boundary (e.g., a suture or transform zone), even though structures appear to connect through the Caspian Sea (Fig. 8).

For the purposes of this synthesis, we suggest that the Permian ophiolites in the Alborz Range (Alavi, 1991), and related outcrops along strike that are associated with the closure of the Paleo-Tethys, demarcate the southern margin of both the Turan and Scythian continental domains. Although a Paleozoic connection between Scythian and Turan domains is still debated, a meaningful divide between the two domains is the northwest-trending Karpinsky swell and Mangyshlak-Ustyurt fold belt that extend to the Kopet Dagh in the southeast (e.g., Zonenshain et al., 1990; Volozh et al., 2003) (Fig. 8). This division is corroborated by a strong fabric of northwest-trending magnetic and gravity anomalies (e.g., Volvovsky et al., 1966; Litvinova, 2000) and documented late Paleozoic thrust faults (Volozh et al., 1999). Following this demarcation, the Turan domain extends from the CAOS in the north-northeast to the Mangyshlak-Ustyurt fold belt in the southwest (Fig. 8). No collision or convergence structures between the Turan domain and Karakum to the southeast have been observed.

Turan was partially covered by the Turkestan Ocean (part of the Paleo-Asian Ocean) in the northeast for much of the Paleozoic, and is covered by regionally extensive late Silurian to Devonian carbonate rocks (Zonenshain et al., 1990; Kurenkov and Aristov, 1995; Garzanti and Gaetani, 2002). Northward subduction of the Turkestan Ocean underneath the CAOS (specifically Kazakhstan and its associated microcontinents) led to its eventual closure in the late Carboniferous (Fig. 4) (e.g., Zonenshain et al., 1990; Filippova et al., 2001). Devonian extension recorded in the Karpinsky swell (Volozh et al., 1999) (Fig. 8) was followed by north-northeastward subduction of the Paleo-Tethys under the Turan domain. Associated arc magmatism began in the Carboniferous and continued into the Mesozoic (Garzanti and Gaetani, 2002), and a backarc basin developed in the Triassic (Gaetani et al., 1998; Thomas et al., 1999a). This subduction accommodated the collision of Turan and Iran-Lut to the south, which

may have been linked with the Scythian domain, as part of the Late Triassic Eo-Cimmerian orogen in northern Iran (Zanchi et al., 2009). This collision may be expressed in the Mangyshlak-Ustyurt fold belt (Fig. 8). Deformed Permian–Triassic flysch in the Mangyshlak anticlinorium is unconformably overlain by Jurassic shallow-marine sandstone (Marcinowski et al., 1996; Ulmishek, 2001a, 2001b). In the Kopet Dagh to the southeast (Fig. 8), strongly deformed Permian volcanic and sedimentary strata are intruded by 207–180 Ma leucogranites (Belov, 1981). In addition, 227–200 Ma K–Ar ages from greenschist rocks have been documented (Lemaire et al., 1997). A regional unconformity in the latest Triassic (Saidi et al., 1997) was followed by deposition of Jurassic deltaic strata that record the unroofing of this orogen (Garzanti and Gaetani, 2002). Cretaceous carbonates covered the region as part of an interior Para-Tethys seaway across the Baltica craton (Marcinowski et al., 1996; Baraboshkin et al., 2003). The Jurassic–Cenozoic deposits to the north of the Mangyshlak-Ustyurt fold belt are important oil reservoirs of the Ushyurt Basin (e.g., Ulmishek, 2001a, 2001b) (Fig. 8).

Karakum

Karakum is mostly covered by thick Mesozoic–Cenozoic sediments, except its easternmost part where Cenozoic thrusting exposes basement rocks in the Baisun and Garm massifs (Burtman, 1976; Biske, 1996; Konopelko et al., 2015). The oldest rocks are gneisses, migmatites, and amphibolites with Pb–Pb isochron ages of 3.0–2.6 Ga (Khoreva et al., 1975; Budanov, 1993) that are covered by Proterozoic metasedimentary rocks consisting of schist and marble. This Precambrian basement underwent rifting and diffuse alkali and tholeiitic basaltic magmatism in the Neoproterozoic (Volkova and Budanov, 1999; Bakirov and Maksumova, 2001; Biske and Seltmann, 2010). Metasedimentary samples from the Garm massif yield detrital zircon ages with significant age peaks at 3.5–3.3 Ga, 2.7–2.4 Ga, 2.2–1.7 Ga, 1.1–0.8 Ga, and 700–575 Ma (Konopelko et al., 2015; Käßner et al., 2016), and a metabasalt (Volkova and Budanov, 1999) has Pb isochron ages of 745–583 Ma (Baratov et al., 1983). Some of the older Proterozoic rocks are intruded by ca. 600 Ma orthogneiss (Käßner et al., 2016) and Permian tonalite (Konopelko et al., 2015). Käßner et al. (2016) suggested that the basement rocks of the Garm massif were linked with the Tarim craton in the Neoproterozoic on the basis of similar inherited zircon ages, zircon Hf-model ages, and whole-rock Nd model ages that require the incorporation of Precambrian material in the melts.

The northern and southern margins of Karakum underwent a relatively uninterrupted Neoproterozoic–Cambrian to early Carboniferous marine transgression (Biske and Seltmann, 2010). Ordovician volcanic rocks in southern Karakum (Mukhin et al., 1991; Dalimov et al., 1993) suggest a south-facing arc system that may have developed along this margin in the early Paleozoic. Ordovician–Silurian seamounts and mid-oceanic ridge basalt (MORB) (Volkova and Budanov, 1999) corroborate this assertion, and may represent relics of the Paleo-Tethys oceanic crust.

Along the northern margin of Karakum, early Carboniferous ophiolitic fragments of the Hissar suture (Portnyagin, 1974; Burtman, 1976, 2006) are thought to represent oceanic lithosphere of the Turkestan Ocean (part of the larger Paleo-Asian Ocean) to the north. Carboniferous to Permian arc-related plutons are found throughout Karakum (Brookfield, 2000; Konopelko et al., 2007; Seltmann et al., 2011). Carboniferous volcanic arc rocks, including low-grade metamorphosed rhyolite, dacite, and andesite, and synorogenic clastic sediments, may demarcate a transition from a passive to an active margin with south-dipping subduction of the Turkestan Ocean at that time (Zonenshain et al., 1990; Brookfield, 2000). This subduction system accommodated the convergence of Kazakhstan and the eventual closure of the Paleo-Asian Ocean (Chen et al., 1999; Charvet et al., 2007). Collision of Karakum with Kazakhstan began at ca. 320 Ma and the intervening Turkestan Ocean was closed by 295–290 Ma (Biske and Seltmann, 2010) (Fig. 4). Northward subduction of the Paleo-Tethys Ocean was initiated during this collision, which explains the presence of seamounts and oceanic lithosphere that accreted onto the southern margin of Karakum. The absence of Permian strata indicates that the region was above sea level for most of this time (Cook et al., 1994), or the rocks were eroded away. Mesozoic northward subduction of the Paleo-Tethys resulted in the development of the Silk Road arc (Natal'in and Şengör, 2005), which may have extended from Turan through the Qaidam-Kunlun continent to the Qinling region in the east (Fig. 2). Early Mesozoic deformation affected the southern margin as Iran-Lut collided with Karakum by the end of the Triassic (Saidi et al., 1997; Zanchi et al., 2009).

Tarim

Although the Tarim Basin, located between the Tibetan Plateau and the Tian Shan (Figs. 1 and 2), is covered by >5 km of Cenozoic sedimentary strata (e.g., Li et al., 1996; Yin et al., 1998), the earliest studies along the margins of the basin recognized cratonic basement underlying these younger sediments (Fig. 9A) (e.g., Argand, 1924; Norin, 1937, 1946). The Precambrian geology of Tarim is inferred from bedrock outcrops along the basin margins (i.e., Quruqtagh, Altyn Tagh, Dunhuang, and Tieklik uplifts) and substantial subsurface data (i.e., well, seismic reflection, seismic refraction, gravity, and magnetic studies) (Fig. 9A). Archean orthogneiss, tonalite-trondhjemite-granodiorite (TTG) gneiss, and amphibolite enclaves are found in the Quruqtagh, Altyn Tagh Range, and Tieklik regions (Fig. 9A). U–Pb zircon ages from these suites indicate a long-lived period of crustal growth at ca. 2.8–2.55 Ga (Lu, 1992; Mei et al., 1998; Lu and Yuan, 2003; Lu et al., 2006; Li et al., 2015). The oldest ages come from the Altyn Tagh Range, where xenocrystic zircons have U–Pb ages of ca. 3.6 Ga (Lu et al., 2008).

Paleoproterozoic metapelites unconformably overlie Archean basement (Gao et al., 1993), and early Paleoproterozoic magmatism occurred throughout Tarim, with observed ages ranging from 2.45 Ga to 2.35 Ga (Lu, 2002; Lu and Yuan, 2003; Lu et al., 2006, 2008). Granitoids with zircon ages of ca. 1.94–1.93 intrude the Quruqtagh region (Ge et al., 2015), and subsequently North Tarim

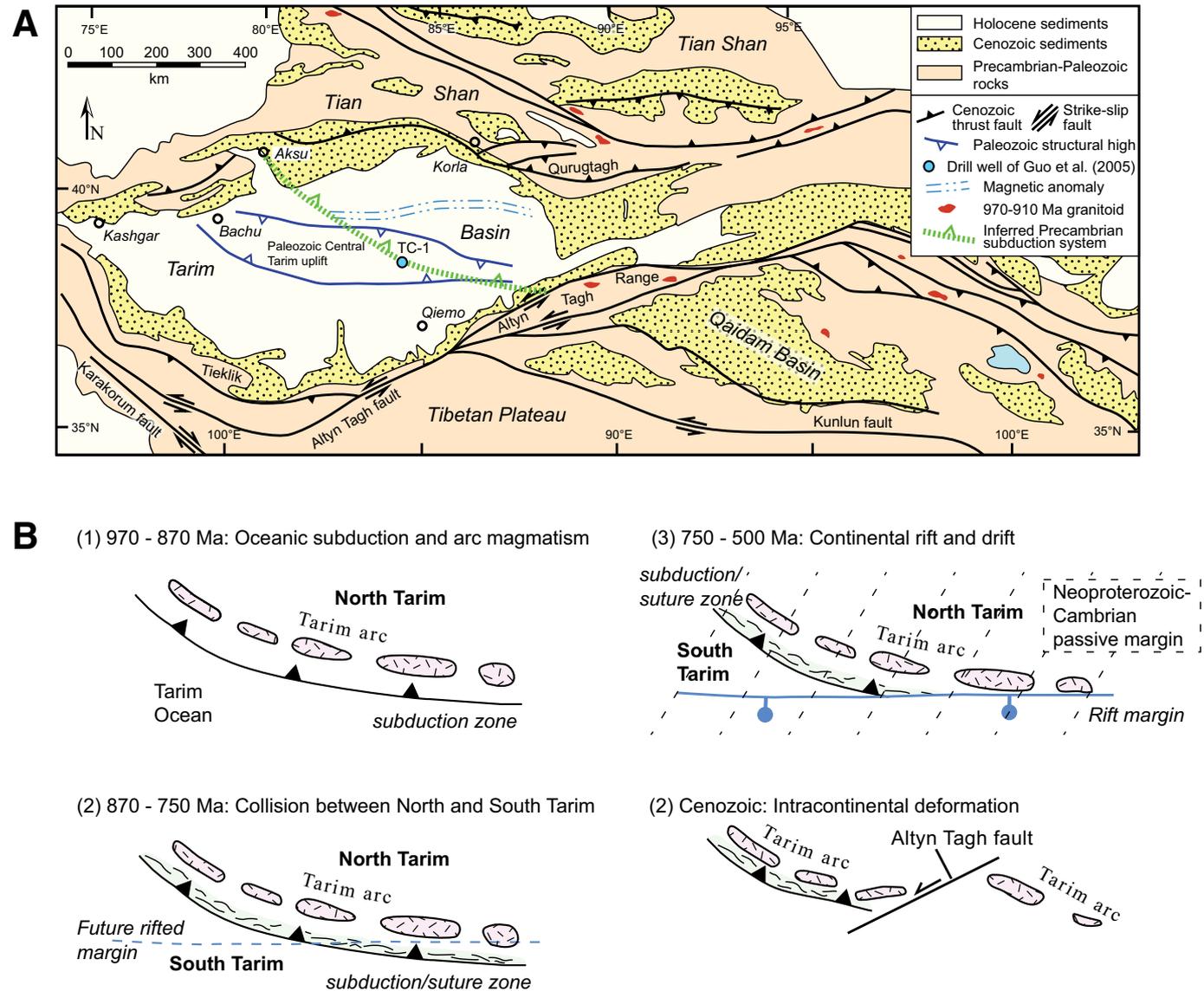


Figure 9. (A) Simplified tectonic map of the Tarim Basin, Qaidam Basin, Qilian Shan, and surrounding areas (modified from Guo et al., 2005). Also shown is the inferred Precambrian subduction system of Guo et al. (2005) and the location of Central Tarim Geologic Survey Well (TC-1). Note the truncated magnetic anomaly and locations of reported 970–910 Ma granitoids (Cowgill et al., 2003; Gehrels et al., 2003a; Wu et al., 2016). (B) Tectonic evolution of the central Tarim Basin and its surrounding regions, originally proposed by Guo et al. (2005) and modified slightly here. (1) Northward subduction under the North Tarim continent formed the Precambrian Tarim arc, which led to the formation of 970–910 Ma granitoids and a blueschist belt along the subduction zone. (2) Closure of the interlying ocean basin led to a collision between North and South Tarim continents. (3) Neoproterozoic rifting and continental breakup was followed by the deposition of passive continental margin sequences over Tarim and surrounding regions. (4) Most recent Cenozoic deformation further modified the geometry of the inferred Precambrian arc.

underwent amphibolite to granulite facies metamorphism at 1.92–1.91 Ga, as documented by U-Pb zircon ages of metamorphic zircons or zircon rims (Lu et al., 2006, 2008; Ge et al., 2015). The timing of these magmatic and metamorphic events is similar to that of the ca. 1.95 Ga Khondalite orogen in North China (e.g., Zhao et al., 2005; Santosh et al., 2006, 2007; Zhao, 2009). The presence of ca. 1.85 Ga mafic dikes (Lu et al., 2008) and ca. 1.77 Ga rapakivi-type granite and mafic dike swarms (Xiao et al., 2003; Lu et al., 2006) suggests an extensional setting at this time (Lu et al., 2006).

The Mesoproterozoic Ailiankate Group of southern Tarim consists of calc-alkaline basalt, andesite, and rhyolite, and is inferred to represent an accreted island arc (Guo et al., 2004). South Tarim was subsequently intruded by 1.4 Ga A-type granites (Ye et al., 2016). Although outcrop exposure is poor, this island-arc collisional event contrasts with the geologic history of North Tarim, where a thick Mesoproterozoic basal conglomerate unconformably overlies Paleoproterozoic rocks, and was subsequently stratigraphically overlain by siliclastic garnet schist and marble (Wang et al., 2004). North Tarim is inferred to have been an undisturbed passive margin throughout most of the Mesoproterozoic.

High-pressure, low-temperature glaucophane-muscovite blueschist and chlorite-rich greenschist exposed near Aksu in northwest Tarim (Liou et al., 1989; Nakajima et al., 1990; Turner, 2010) (Fig. 9A) may represent a relic subduction system between the North and South Tarim cratons (Guo et al., 1999, 2005; Xu et al., 2013). Dating efforts of the blueschist yield a range of ages, including K-Ar glaucophane ages of 718 ± 22 Ma and 710 ± 21 Ma, Rb-Sr phengite ages of 698 ± 26 Ma and 714 ± 24 Ma (Nakajima et al., 1990), and $^{40}\text{Ar}/^{39}\text{Ar}$ phengite ages of ca. 750 Ma (Yong et al., 2013). However, some of these ages must be questioned because undeformed crosscutting dikes have an older U-Pb zircon age of 807 ± 12 Ma (Chen et al., 2004). Unpublished $^{40}\text{Ar}/^{39}\text{Ar}$ ages from glaucophane show peak metamorphism occurring at 872–862 Ma (e.g., Chen et al., 2004). In addition, Ge et al. (2016) reported 830–800 Ma metamorphic zircon ages from granulite facies metamorphic rocks.

The interlying ocean of the Aksu suture is evidenced by the occurrence of metabasalts, which have enriched MORB protoliths with a Sm-Nd isochron age of 890 ± 23 Ma (Zheng et al., 2010). Prior to the collision of the North and South Tarim continents, a subduction system must have facilitated the convergence of these two blocks. This system may have generated the Tonian-aged granitoids (970–910 Ma) northeast of this inferred suture (Fig. 9A) in the Altyn Tagh Range (Cowgill et al., 2003; Gehrels et al., 2003a), the Qilian Shan (Gehrels et al., 2003a; Tung et al., 2007, 2013; Wu et al., 2016, 2017), and the Qinling (Chen et al., 2006), suggesting that this subduction zone was part of a north-east-dipping Tarim arc. Detrital zircon grains from northern Tarim also support arc activity at ca. 900 Ma (Zhu et al., 2011; J. He et al., 2014) (Fig. 5). In addition, a foliated granite from the Quruqtagh region yielded a U-Pb zircon age of 933 ± 33 Ma (Shu et al., 2011), and granodiorite samples recovered from drill cores in central Tarim yield argon ages of 930–890 Ma (Li et al., 2005). This arc would have extended from Tarim through Qaidam and the Qilian Shan to the present-day Qinling area (Fig. 9B) (Wu et al., 2016, 2017). Metamorphic zircon rims

suggest that regional metamorphism may have occurred in northern Tarim from ca. 830 Ma to 790 Ma (He et al., 2012; Ge et al., 2013a, 2014).

Although no Precambrian suture has been reported in the Qaidam, Qilian Shan, or Qinling regions, Neoproterozoic rifting (discussed below) or extensive Cenozoic basin deposits may obscure such observations (Fig. 9B). It is also possible that the numerous ophiolite, eclogite, and mélange exposures found in the Altyn Tagh Range, North Qaidam, and Qilian Shan (e.g., Sobel and Arnaud, 1999; Xiao et al., 2009c; Song et al., 2013, 2014) may contain fragments of this Precambrian suture (Fig. 9) that have been mistakenly grouped with the early Paleozoic Qilian suture(s).

Following these early Neoproterozoic events, Tarim underwent regional rifting and the development of extensive passive margins (Turner, 2010; Zhang et al., 2016). Evidence for Neoproterozoic rift activity along Tarim's northern and southern margins includes mafic dike intrusions, alkaline plutonism, bimodal volcanism, rift-basin development, and passive margin sedimentation (Fig. 6). In northern Tarim, two or three distinct pulses of rift-related magmatic and/or plume activity occurred at 830–800 Ma, 790–740 Ma, and 650–630 Ma (e.g., Zhang et al., 2007a; Lu et al., 2008; Zhu et al., 2008; Shu et al., 2011) (e.g., Fig. 5), which are generally associated with the opening of the Paleo-Asian Ocean. The 830–800 Ma plutons in northern Tarim may either be related to an initial phase of rift development (Lu et al., 2008; Zhu et al., 2008) or arc subduction and collision along the Tarim arc (e.g., Fig. 9), but the lack of geochemical analyses of the observed bimodal volcanic rocks makes this assertion uncertain. The second distinct pulse of magmatic activity, from 790 Ma to 740 Ma, was accompanied by bimodal volcanism and has been related to rifting (Lu et al., 2002; Guo et al., 2005; Xu et al., 2005; Zhang et al., 2006a). The 650–630 Ma Korla dikes indicate that rift-related magmatism continued into the Ediacaran (Zhu et al., 2008; Zheng et al., 2010). This latest stage of volcanism may be related to the opening of the possibly smaller Qilian Ocean to the southeast (e.g., Xu et al., 2015; Wu et al., 2017a).

In southern Tarim, specifically in the Tiekelik region (Fig. 9A), bimodal volcanism and rift-basin development at 900–870 Ma (Wang et al., 2015a, 2015b) indicate that rifting along the Tarim southern margin initiated earlier than in the north. This requires rifting to immediately follow collision along the Tarim suture. This event is associated with the opening of the Tethys oceans.

Following rifting along Tarim's margins, the region was overlain by thick successions of latest Neoproterozoic to Cambrian passive margin strata (Jia, 1997; Zhang et al., 2000; Jia et al., 2004; Huang et al., 2005; Xu et al., 2005; Biske and Seltmann, 2010; Turner, 2010; Shu et al., 2011) (Fig. 6). The northern margin remained a passive continental shelf throughout much the Paleozoic (Graham et al., 1993; Carroll et al., 1995, 2001), but two regional unconformities truncate the sedimentary record. The first occurred in the early Paleozoic, where Silurian–Devonian siliclastic strata, sourced from the southeast, unconformably overlie Ordovician rocks (Carroll et al., 2001). Overlying Carboniferous strata are separated from Silurian–Devonian rocks by a sharp angular unconformity (Lin et al., 2014). These unconformity pulses and siliclastic sediments are related to the Qilian orogen to the southeast, as the Qaidam continent collided

with the southern margin of Tarim and North China (Wu et al., 2017a). The thick deposits of Carboniferous through lower Permian strata are truncated by a major angular unconformity (Carroll et al., 2001). This event is related to the collision of the Tian Shan and CAOS microcontinents along the northern margin of Tarim, and the closure of the Paleo-Asian Ocean.

Late Paleozoic–Mesozoic arc magmatism observed along the southern margin of Tarim in the Western Kunlun Range (Cowgill et al., 2003; Xiao et al., 2005) is related to northward subduction of the Paleo-Tethys Ocean and arc development along the southern margin of Tarim. This magmatic arc represents the westward continuation of the Paleozoic–Mesozoic Kunlun magmatic arc that is located to the southeast in the Eastern Kunlun Range of the Qaidam–Kunlun continent (Jiang et al., 1992; Cowgill et al., 2003; Wu et al., 2016).

Qaidam–Kunlun Continent

The Qaidam–Kunlun continent, which represents the northern margin of the present-day Tibetan Plateau, underwent progressive subduction, arc magmatism, and orogeny throughout the Neoproterozoic, Paleozoic, and Mesozoic (e.g., Yin et al., 2007a, 2008a; Xiao et al., 2009c; Gehrels et al., 2011; Song et al., 2013, 2014; Wu et al., 2016, 2017b), as evidenced by the widespread exposure of ophiolitic mélangé (Wang and Liu, 1976, 1981; Xiao et al., 1978; Feng and He, 1995), ultrahigh pressure (UHP) rocks (Yin et al., 2007a; Menold et al., 2009, 2016; Song et al., 2014), blueschist rocks (Xiao et al., 1974; Liu et al., 2006; Lin et al., 2010), and plutons (Gehrels et al., 2003a, 2003b, 2011; Wu et al., 2016, 2017b) (Figs. 9 and 10). The region has been reactivated by the Cenozoic Qilian Shan–Nan Shan, North Qaidam, and Kunlun–Qimen Tagh thrust belts, which bound Qaidam Basin (e.g., Yin et al., 2007b, 2008a, 2008b; Zuza et al., 2013, 2016).

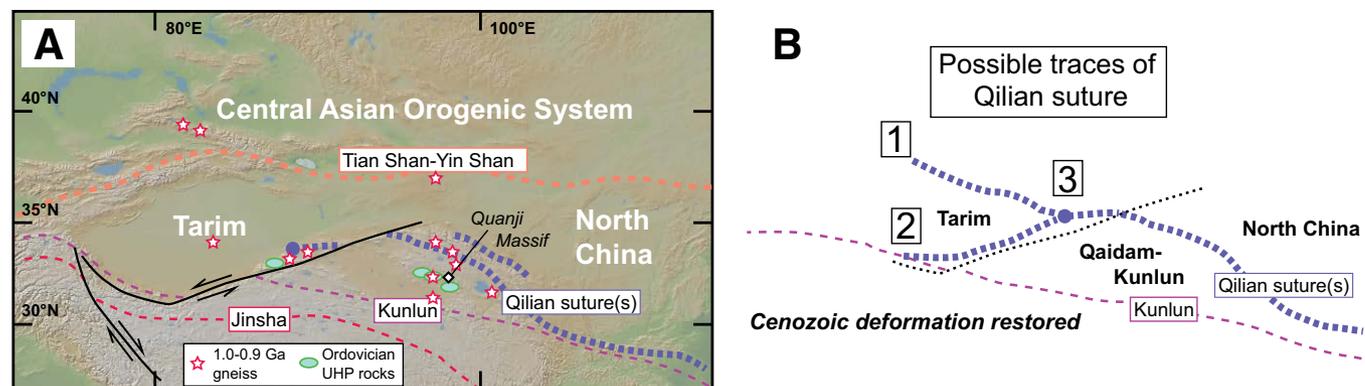
The heterogeneous basement consists of Mesoproterozoic passive margin strata in the west and Archean to Proterozoic crystalline rocks to the east (e.g., Qinghai Bureau of Geology and Mineral Resources, 1991; Guo et al., 1999; Gehrels et al., 2003b; Pan et al., 2004; Yu et al., 2017). The Quanji massif (Fig. 10A), the best-studied Precambrian basement in the Qilian Shan (Lu, 2002; Wan et al., 2006; Wang et al., 2008, 2009; Gong et al., 2012; Chen et al., 2013b; Zhang et al., 2014a), consists of medium- to high-grade metamorphic rocks, which have been referred to as the Delingha, Dakendaban, and Wandonggou groups, unconformably overlain by Neoproterozoic strata (Qinghai Bureau of Geology and Mineral Resources, 1991). The Delingha granitic gneiss yields Paleoproterozoic zircon ages of 2.39–2.37 Ga (Lu, 2002; Lu et al., 2006, 2008). Based on their geochemistry, the granite protoliths are interpreted to have formed in a rift setting, possibly related to the breakup of a Neoarchean continent (Diwu et al., 2007; Gong et al., 2012). The Dakendaban metasedimentary rocks were deposited after ca. 2.47 Ga but before ca. 2.43 Ga, based on detrital zircon analyses and a dated crosscutting pegmatite (Wang et al., 2008), whereas the Wandonggou metasedimentary rocks were deposited between 2.24 and 1.95 Ga (Huang et al., 2011). The Quanji massif underwent amphibolite facies metamorphism at ca. 1.95–1.93 Ga (Hao et al., 2004; Wang et al., 2008; Chen et al., 2009; Chen et al., 2013b) and was subsequently intruded by 1.83 Ga mafic

dikes (Liao et al., 2014) and ca. 1.8 Ga rapakivi granite (Lu et al., 2006; Chen et al., 2012). The tectonic setting of Mesoproterozoic metasedimentary strata is poorly constrained (Lu et al., 2006), and they are overlain by the tillite-bearing siliciclastic- and carbonate-dominated Quanji Group (Shen et al., 2010; Wang et al., 2013) (Fig. 6).

The Qilian Shan, Qaidam, and Kunlun regions were intruded by ca. 970–910 Ma and ca. 850–800 Ma plutons; this suggests that these regions were contiguous by the start of the Neoproterozoic (Fig. 10A) (Gehrels et al., 2003b; Tung et al., 2013; Wu et al., 2016, 2017b). As previously discussed, similarly aged ca. 970–910 Ma plutons have been documented to the west and southeast in the Tian Shan, Altyn Tagh Range, Tarim, and Qinling (Cowgill et al., 2003; Gehrels et al., 2003b; Lu et al., 2008; He et al., 2014b; He et al., 2014a) (Fig. 9A), which suggests the existence of a south-facing Proterozoic subduction zone that extended from Tarim to Qinling (e.g., Guo et al., 2005; Tseng et al., 2009). The ca. 1.0–0.8 Ga plutons and zircon signature within Proterozoic strata of the Qaidam–Kunlun continent (Fig. 5) may correlate with either the Tarim–North China (Gehrels et al., 2003a, 2011; Chen et al., 2006; Peng, 2010; Peng et al., 2011a, 2011b; Wang et al., 2012; Liu et al., 2012; Dan et al., 2014; Yu et al., 2017) or South China cratons (e.g., Tung et al., 2013). Recently reported drilling from Qaidam Basin suggests that ca. 2.4 Ga basement here (Yu et al., 2017) correlates with similar rocks in North Tarim (e.g., Lu et al., 2006, 2008).

The occurrence of 790–750 Ma intrusions in the Qilian Shan has been attributed to the rifting of the Qaidam–Kunlun continent from another continent (e.g., North or South China) and the opening of the Qilian Ocean (Tseng et al., 2006; Song et al., 2013; Wu et al., 2016, 2017b). Alternatively, the ca. 600 Ma basalt interbedded with thick marble rocks in the Qiqing Group may indicate later rifting and opening of the Qilian Ocean (Xu et al., 2015) (Fig. 6). The region was subsequently overlain by Neoproterozoic passive margin strata (e.g., Guo et al., 1999; Wan et al., 2001, 2003a; Cowgill et al., 2003; Gehrels et al., 2003a, 2003b). If the Qaidam–Kunlun continent was originally derived from Tarim and/or North China, the Qilian Ocean must be distinct from the proto-Tethys (or any Tethyan Ocean) (cf. Gehrels et al., 2011) because the Tethyan oceanic domain (Fig. 1) was originally defined as the ocean between Laurasia and Gondwana (Şengör, 1984), whereas the Qilian Ocean opened following the rifting of Qaidam–Kunlun from the Tarim–North China craton.

The collision between the Qaidam–Kunlun continent, interlying arcs, and the Tarim–North China craton is expressed as the early Paleozoic Qilian orogen (Yin and Nie, 1996; Şengör and Natal'in, 1996; Sobel and Arnaud, 1999; Yin and Harrison, 2000; Gehrels et al., 2003a, 2003b; Yin et al., 2007a; Xiao et al., 2009c; Song et al., 2013; Wu et al., 2017a). The Qilian orogen is composed of flysch sequences, arc-type assemblages, ophiolites, and low- to high-grade metamorphic rocks. Unresolved primary problems regarding the Qilian orogen include: (1) how many arcs and what type of arcs (i.e., oceanic or continental) were involved in orogeny (e.g., Xiao et al., 2009c; Yang et al., 2009, 2012a; Song et al., 2013); (2) whether the subduction was north and/or south dipping (e.g., Sobel and Arnaud, 1999; Yin and Harrison, 2000; Gehrels et al., 2003a, 2003b, 2011; Yin et al., 2007a; Xiao et al., 2009c; Yang et al., 2009, 2012a; Yan et



C Geometric configurations of the Qaidam-Kunlun continent and the Tarim-North China cratons

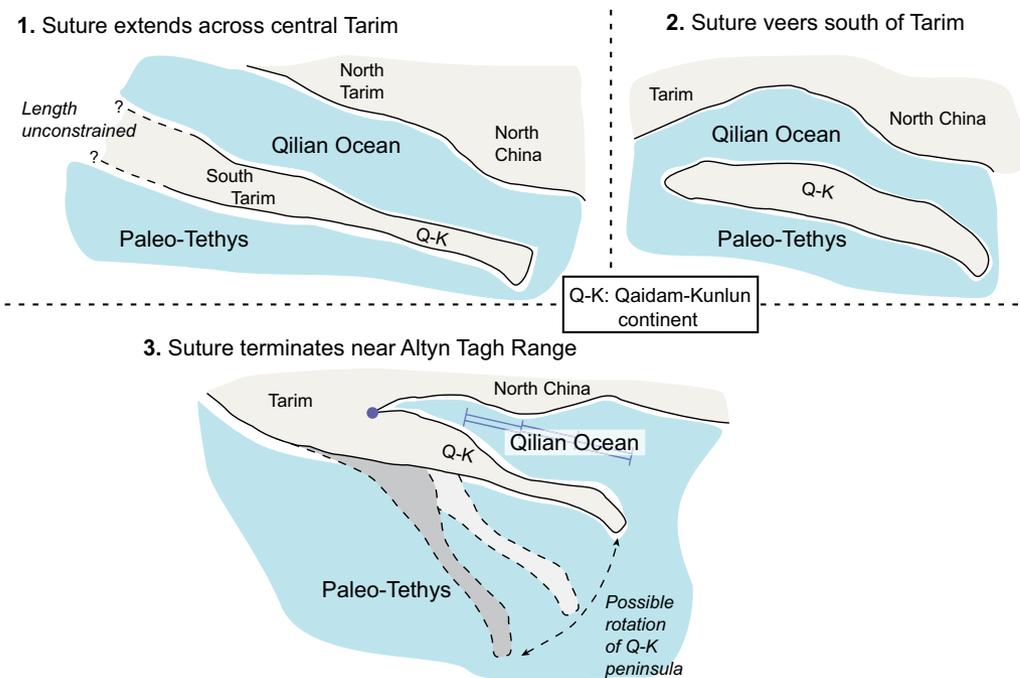


Figure 10. (A) Present-day configuration of the Jinsha, Kunlun, Qilian, and Tian Shan-Yin Shan suture zones, from south to north. Note that the Qilian suture is offset by the Altyn Tagh fault. UHP—ultrahigh pressure. (B) Restored position of the Kunlun and Qilian sutures, after the removal the effects of Cenozoic shortening and offset along the Altyn Tagh fault. (C) Three possible geometric configurations of the Qaidam-Kunlun (Q-K) continent and the Tarim–North China cratons given the restored surface trace of the Qilian suture as shown in B.

al., 2010; Song et al., 2013, 2014); and (3) when the collision and closure of the Qilian Ocean occurred (e.g., Liu et al., 2006; Qi, 2003; Tung et al., 2007; Xiao et al., 2009c; Lin et al., 2010).

Regardless of the aforementioned uncertainties, the following is known about the Qilian orogen. An open ocean or oceans existed from at least 550 Ma to 448 Ma, as evidenced by the widespread distribution of ophiolite fragments (Shi et al., 2004; Smith, 2006; Xiang et al., 2007; Tseng et al., 2007; Xia and Song, 2010; Song et al., 2013). Widespread arc-related plutons indicate that a major subduction system was initiated by ca. 515 Ma and continued throughout the Ordovician; younger accretion-related magmatism may have persisted until 345 Ma (Qinghai Bureau of Geology and Mineral Resources, 1991; Qian et al., 1998; Cowgill et al., 2003; Gehrels et al., 2003a; Su et al., 2004; Wu et al., 2004, 2006, 2010; Hu et al., 2005; Liu et al., 2006; Quan et al., 2006; He et al., 2007; Tseng et al., 2009; Xu et al., 2010; Xiong et al., 2012; Song et al., 2013). The youngest major pulse of magmatism at ca. 445 Ma (e.g., Wu et al., 2016, 2017b), ca. 454–442 Ma $^{39}\text{Ar}/^{40}\text{Ar}$ mica cooling ages (Liu et al., 2006), and Silurian foreland basin deposits (Yan et al., 2007, 2010) suggest that collision between Qaidam-Kunlun and the North China craton probably occurred at ca. 445–440 Ma. Silurian cooling ages and ca. 400 Ma metamorphic zircon rim growth are consistent with intracontinental deformation due to collision at that time (Qi, 2003; Lin et al., 2010).

Based on similar lithological assemblages and geologic histories, the Qilian orogen and sutures likely connect with the Qinling orogen (Fig. 1) to the southeast (e.g., Xu et al., 2008; Tseng et al., 2009; Dong et al., 2011; Wu et al., 2016, 2017b), forming a >1000-km-long early Paleozoic orogenic belt. The western termination of this orogen and suture is not known, and three geometrically plausible models exist for its westward continuation (Fig. 10): (1) it may continue into Tarim and is currently covered by Mesozoic–Cenozoic sediments; (2) it may veer to the south of Tarim and was later overprinted by the Western Kunlun and Pamir thrust belts, although evidence for this suture in the Altyn Tagh Range suggests that the suture cannot be that far to the southwest; or (3) the Qaidam-Kunlun continent may have only partially separated from Tarim–North China, and the abrupt western termination of the Qilian suture represents the western extent of the Qilian Ocean. A previous additional model suggested that the Qilian suture connected to the northwest to the Beishan suture (Zhou and Graham, 1996a), although this idea has mostly been abandoned because the Beishan suture has been related to the CAOS and the closure of the Paleo-Asian Ocean (e.g., Xiao et al., 2010; Guo et al., 2011; Cleven et al., 2015). Each model has implications for the nature of the Qaidam-Kunlun continent, including its size, geometry, origin, and tectonic relationships with adjacent blocks (Fig. 10C). For example, the first model requires (1) the southern portion of Tarim to have linked with Qaidam-Kunlun in the Precambrian and (2) that an early Paleozoic suture extends across the Tarim Basin. The second model is the only one that predicts Qaidam-Kunlun to be an individual entity surrounded entirely by oceans, and therefore currently surrounded by relict and observed suture zones, whereas the third model requires a Precambrian link between the Qaidam-Kunlun continent and Tarim (e.g., Yu et al., 2017).

Northward subduction of the Paleo-Tethys led to the development of the Kunlun magmatic arc along the southern margin of Tarim and Qaidam-Kunlun, which was initiated by the Devonian (Li et al., 1991; Jiang et al., 1992; Wang et al., 1993; Wu et al., 2016) and accommodated the northward convergence of the Qiantang terrane and the South China craton. The Songpan-Ganzi flysch basin developed to the southwest as the North and South China cratons collided (Yin and Nie, 1993; Nie et al., 1994; Zhou and Graham, 1996b; Enkelmann et al., 2007; Pullen et al., 2008). Final closure of the Paleo-Tethys occurred by the latest Triassic (Fig. 4), which is demarcated by the Jinsha and Anyimaqen-Kunlun-Muztagh suture zones (Fig. 1).

North China

The wedge-shaped North China craton (also referred to as the Sino-Korean craton or platform in the literature), is bounded to the west-southwest by the early Paleozoic Qilian-Qinling orogen, to the north by the late Paleozoic CAOS, and to the south by the Mesozoic Qinling-Dabie Shan orogen (Figs. 1 and 2). The craton is traditionally divided into two major Archean–Proterozoic blocks (i.e., the Eastern and Western blocks) separated by the ~1500-km-long north-trending Paleoproterozoic(?) Trans-North China Orogen (e.g., Zhao et al., 1998, 2005, 2012; Trap et al., 2012) (Fig. 11A), which is also referred to as the Central Orogenic Belt (e.g., Kusky and Li, 2003; Kusky et al., 2007). Following this division, the Eastern Block is made up of the Langrim and Lonngang blocks that joined along the Paleoproterozoic Liao-Liao-Ji orogen (e.g., Zhao et al., 2005; Tam et al., 2011), which contains 3.8–2.6 Ga gneiss and greenstone belts overlain by 2.6–2.5 Ga metasedimentary cover. The Western Block consists of the Ordos block and Yinshan block, which coalesced at ca. 1.92 Ga during Khondalite Belt orogeny (e.g., Zhao et al., 2005; Santosh et al., 2006, 2007; Zhao, 2009).

The tectonic framework and timing of the collision between the Western and Eastern blocks (Fig. 11A) have long been debated and two competing hypotheses have emerged. The primary difference between the two groups of models is the collisional age: ca. 1.85 Ga (e.g., Kröner et al., 2005a, 2005b, 2006; Zhang et al., 2007b; Zhao et al., 1998, 1999, 2000, 2001a, 2001b, 2004b, 2005, 2007, 2010, 2012; Faure et al., 2007; Trap et al., 2007, 2008, 2009a, 2009b, 2011, 2012; Xiao et al., 2011; Zhao and Cawood, 2012) or ca. 2.5 Ga (Li et al., 2000a, 2000b; Kusky and Li, 2003; Polat et al., 2005, 2006; Li and Kusky, 2007; Kusky et al., 2007; Kusky and Santosh, 2009; Kusky, 2011). However, it is becoming increasingly clear that both the ca. 2.5 and ca. 1.85 Ga events involved widespread intracontinental deformation that were not restricted to such narrow linear belts as originally envisioned by Zhao et al. (1998) (e.g., Kusky and Li, 2003). Instead, the present-day exposure of these orogens is probably related to more recent tectonics. For example, the apparently north-trending Trans-North China Orogen is primarily exposed along Mesozoic–Cenozoic rift shoulders that are bounded by thick rift-related strata (e.g., Davis et al., 1996, 2001; Graham et al., 2012). Research on Paleoproterozoic orogens in North

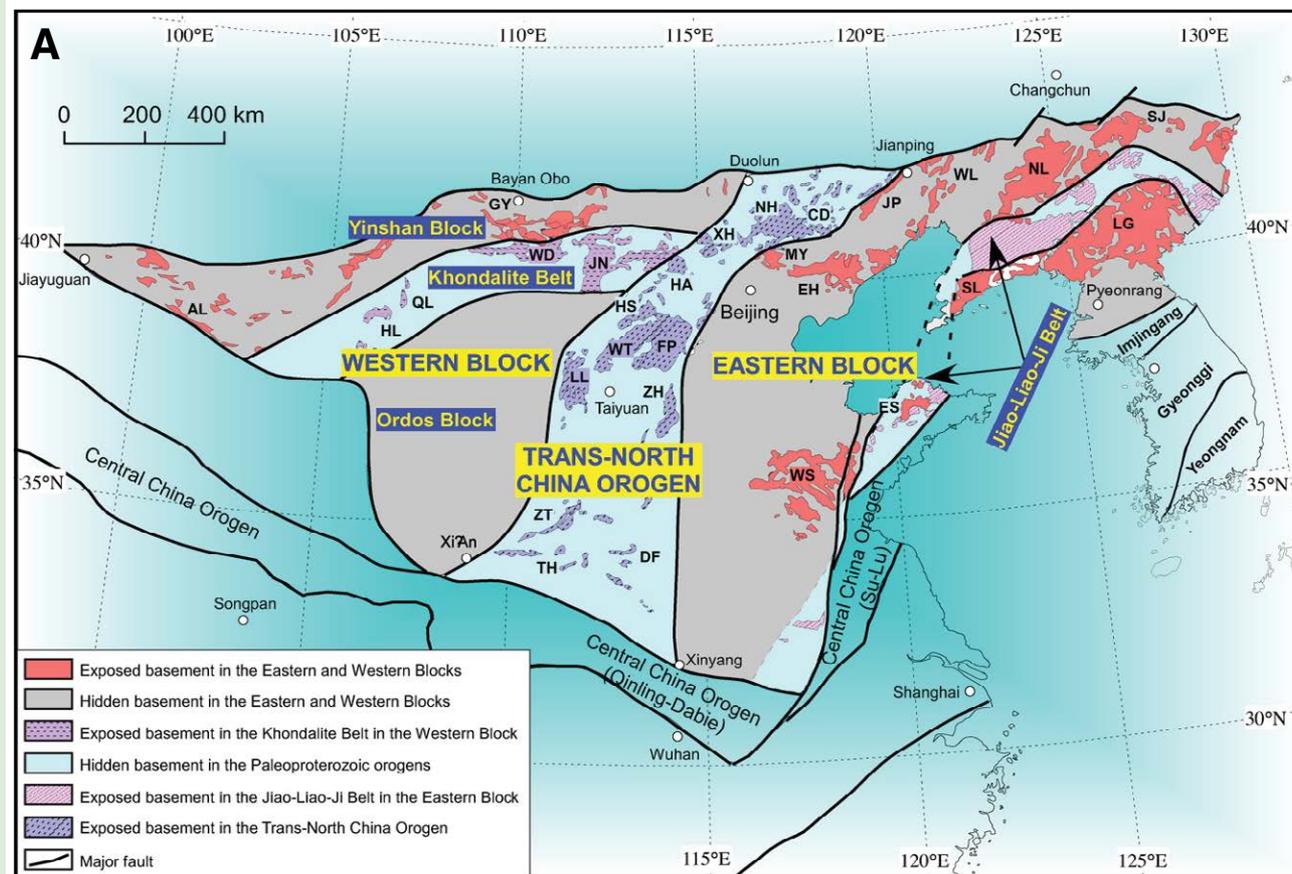
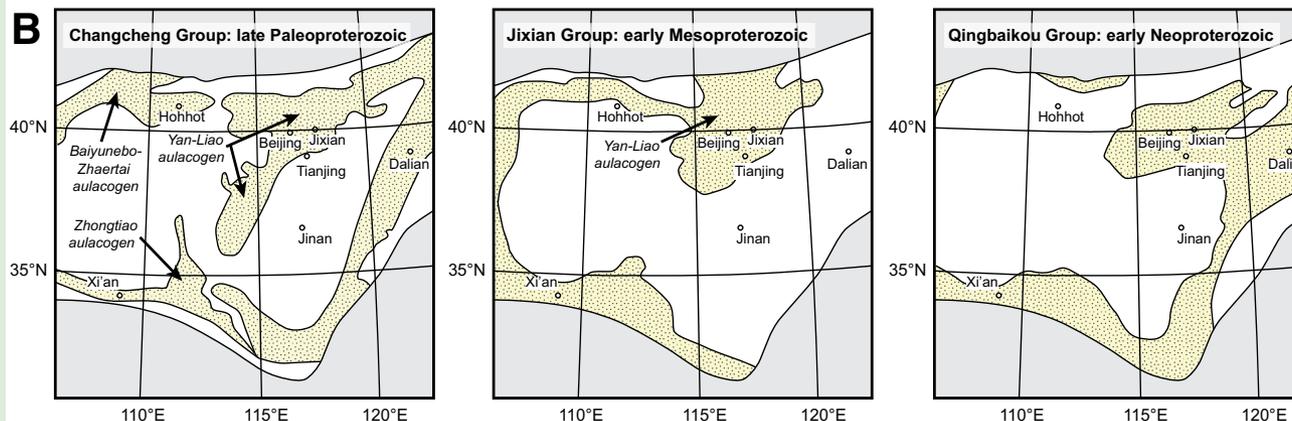


Figure 11. The Precambrian North China craton. (A) Tectonic domains of the North China craton. Abbreviated metamorphic complexes: AL—Alxa; CD—Chengde; DF—Dengfeng; EH—eastern Hebei; ES—eastern Shandong; FP—Fuping; GY—Guyang; HA—Hua’an; HL—Helanshan; HS—Hengshan; JN—Jining; LG—Langrim; LL—Lüliang; JP—Jianping; MY—Miyun; NH—northern Hebei; NL—northern Liaoning; QL—Qianlishan; SJ—southern Jilin; SL—southern Liaoning; TH—Taihua; WD—Wulashan-Daqingshan; WL—western Liaoning; WS—western Shandong; WT—Wutai; XH—Xuanhua; ZH—Zanhuang; and ZT—Zhongtiao. From Zhao et al. (2012; revised from Zhao et al., 2005). (B) Sketch maps showing the distribution of the Proterozoic Changcheng, Jixian, and Qingbaikou Groups within the North China craton. There are no significant late Neoproterozoic (i.e., Ediacaran) deposits in North China. Compiled from Wang (1985), Lu (2002), Wan et al. (2003b), and Meng et al. (2011). Note that intracontinental deformation and other later tectonic modification have not been restored in A or B.



America is undergoing a similar renaissance; orogens that were once considered narrow deformational belts are increasingly being reinterpreted as parts of broader zones of intracontinental deformation (e.g., Vervoort et al., 2016). We do not attempt to further interpret the Archean–Paleoproterozoic North China orogens, and for the purpose of our reconstruction, the North China craton is assumed to have been established by ca. 1.8 Ga.

Following the latest Paleoproterozoic event, near continuous Mesoproterozoic–Neoproterozoic sedimentation is observed in the Jixian County region of North China. The strata can be divided into four distinct successions, the Changcheng, Jixian, post-Jixian, and Qingbaikou (Wan et al., 2011; Sun et al., 2012; i.e., latest Paleoproterozoic, Calymmian, Stenian–Ectasian, and Tonian; Table 1; Fig. 11B), although two more distinct groups may have existed from 1.4 to 1.0 Ga (Li et al., 2013). A late Paleoproterozoic rift event is evidenced by the Yan-Liao aulacogen, continental margin rift deposits (Fig. 11B), and anorthosite-rapakivi granite (Li et al., 2000b; Yang et al., 2005). Within the Yan-Liao aulacogen, the oldest Changcheng Group metasedimentary deposits (i.e., ca. 1.8 Ga zircon ages of Wan et al., 2003b) unconformably overlie Paleoproterozoic orthogneiss, transitioning upward from conglomerate and sandstones at the base to shale and dolomite (Cheng et al., 1981). Calymmian (i.e., early Mesoproterozoic) strata consist of ~4.5-km-thick deposits of dolomite, minor limestone, mudstones, and shale deposited in an active(?) rift basin (Hebei Bureau of Geology and Mineral Resources, 1989; Wan et al., 2011; Ying et al., 2011; Zhu et al., 2016). The Qingbaikou Group unconformably overlies the Jixian sections and consists of <500 m of predominantly siliciclastic rocks (Ying et al., 2011).

There are no middle to late Neoproterozoic (i.e., 850–550 Ma) deposits in the Jixian region of North China, and only minor sections in the Hu-Huai and Lushan-Ruyang regions (Zhang et al., 2006b) to the south and the Liaodong Peninsula (Chang, 1980) to the east have reported strata of this age. The ages of these purported late Neoproterozoic deposits were determined by biostratigraphy, regional correlations, and old isotopic dating techniques, and some formations were shown to be out of stratigraphic order (Zhang et al., 2006b); more modern techniques (e.g., detrital zircon ages or crosscutting igneous rock ages) are needed to verify these findings. The general absence of late Neoproterozoic strata is problematic because there is no regional geologic record of several important events that may have affected the region, including the breakup of the Rodinia supercontinent (e.g., Li et al., 2008) or possible global(?) Neoproterozoic glaciations (e.g., Hoffman et al., 1998; Allen and Etienne, 2008; Li et al., 2013; Rooney et al., 2015). Xiao et al. (2014) documented Tonian stratigraphy along the southern margin of the North China craton. The lack of Neoproterozoic rift deposits along the northern margin of the North China craton has been used as evidence against the craton's involvement in Rodinia, but these deposits may have been eroded away.

Neoproterozoic intrusions have been reported across the North China craton, including ca. 900 Ma sills in the northern Korean peninsula (Peng et al., 2011a, 2011b), 930–910 Ma granites in the western part of the craton (Dan et al., 2014), and ca. 890 Ma dikes in the southern part of the craton (Wang et al., 2012). Tonian rift-related ultramafic or bimodal intrusions and volcanic rocks have

been documented along the southern margin of the North China craton, including the ca. 825 Ma Jinchuan intrusion (Li et al., 2005), the 1.0–0.84 Ga Qin'an dikes (Liu et al., 2012), and the ca. 830 Ma Luanchuan gabbros (Wang et al., 2011b). These ages correlate with rift-related rocks in southern Qaidam (Fig. 6).

Early Cambrian rocks unconformably overlie the uppermost section of early Neoproterozoic rocks in the Jixian region (i.e., the Jing'eryu Formation) (Ying et al., 2011; Sun et al., 2012). Widespread Cambrian shelf limestones (Meng et al., 1997; Myrow et al., 2015) indicate that the North China craton was surrounded to the north and south by open oceans (i.e., the Paleo-Asian and Paleo-Tethys Oceans, respectively). Some suggest that North China and Tarim were linked to the northern margin of Gondwana in the Cambrian (McKenzie et al., 2011; Myrow et al., 2015; Han et al., 2016). However, this configuration is geologically implausible because it does not allow for the collision of the Qaidam-Kunlun continent with the southern margin of the North China craton in the Ordovician–Silurian. For example, in the reconstruction of Han et al. (2016), the opening of the Paleo-Tethys Ocean via the separation of Gondwana and the North China craton is in the same location that the Qilian collisional orogen is inferred to have occurred (e.g., Lin et al., 2010; Wu et al., 2017a). Paleomagnetic reconstructions also do not support North China or Tarim being adjacent to Gondwana in the Cambrian–Ordovician (Cocks and Torsvik, 2013; Torsvik and Cocks, 2013). In addition, a Cambrian Gondwana link with North China's southern margin requires an additional unidentified pre-Mesozoic rifting event to create space for the collision of Songpan–Ganzi–South China with North China (Yin and Nie, 1993; cf. Han et al., 2016). However, passive margin sedimentation along the southern margin of the Eastern Kunlun Shan may have been related to the opening of the Paleo-Tethys (Neo-Kunlun) Ocean (Wu et al., 2016).

Collision between the South China and North China cratons (Yin and Nie, 1993; Zhang, 1997; Hacker et al., 2004, 2006; Wu et al., 2016) (Fig. 1) began in the latest Permian and concluded in the Late Triassic (Zhao and Coe, 1987; Ratschbacher et al., 2003; Hacker et al., 2006), and is marked by the Qinling-Dabie suture. Continued convergence between the North China and South China cratons, expressed by the development of the Daba Shan thrust belt along the southern margin of the Qinling orogen, lasted until the Late Jurassic (ca. 165 Ma) (e.g., Dong et al., 2013). Erosion of this orogen may have resulted in the deposition of thick clastic sediments in the Songpan–Ganzi remnant ocean basin (Zhou and Graham, 1996b; Enkelmann et al., 2007; Pullen et al., 2008). North China's northern margin remained quiescent until the Carboniferous–Permian initiation of south-directed subduction of the Paleo-Asian Ocean, which eventually led to a late Permian to Early Triassic collision against the CAOS that continued into the Jurassic (Xiao et al., 2003, 2009b, 2009c; Eizenhöfer et al., 2014; Wu et al., 2017b; cf. Jian et al., 2016).

■ NEOPROTEROZOIC BALKATACH CONTINENT

Here we describe our tectonic restoration of Central Asia from the present through the Phanerozoic to the Proterozoic. Emphasis is placed on providing evidence for a linked Neoproterozoic continental strip that we refer to as the

Balkatach with its maximum extent ca. 870 Ma when the Western and Eastern domains of this inferred continent collided along the central Tarim suture (Fig. 9). In order to accurately portray the original shape of this continent, we systematically restore the significant continental deformation and tectonic calving of microcontinents that have distorted this Precambrian continent.

We use the most current geological map of Asia (Ren et al., 2013) as a base map to demarcate the present-day shape and boundaries of major tectonic terranes. Although the initial reconstruction presented here (i.e., Fig. 12) makes no attempt to restore the polar region distortion due to the particular projection scheme of Ren et al. (2013), which minimizes map-view distortion of continental China, these distortion effects are not significant because the general trend of the Balkatach is east-west. Note that our final restored shape files

were produced by undeforming polygons using the GPlates software (Boyden et al., 2010). In our reconstruction, we start by restoring the most recent tectonic events that affected the geometry of Asia (i.e., Cenozoic deformation) and progress to older events in the Mesoproterozoic (Fig. 12). Reconstruction information is presented based on present-day geographic location, moving from west to east (i.e., Baltica to eastern North China).

Late Mesozoic–Cenozoic Intracontinental Deformation

We restore intracontinental deformation associated with the Cenozoic India-Asia collision, which initiated at 65–55 Ma (e.g., Le Fort, 1996; Yin and Harri-

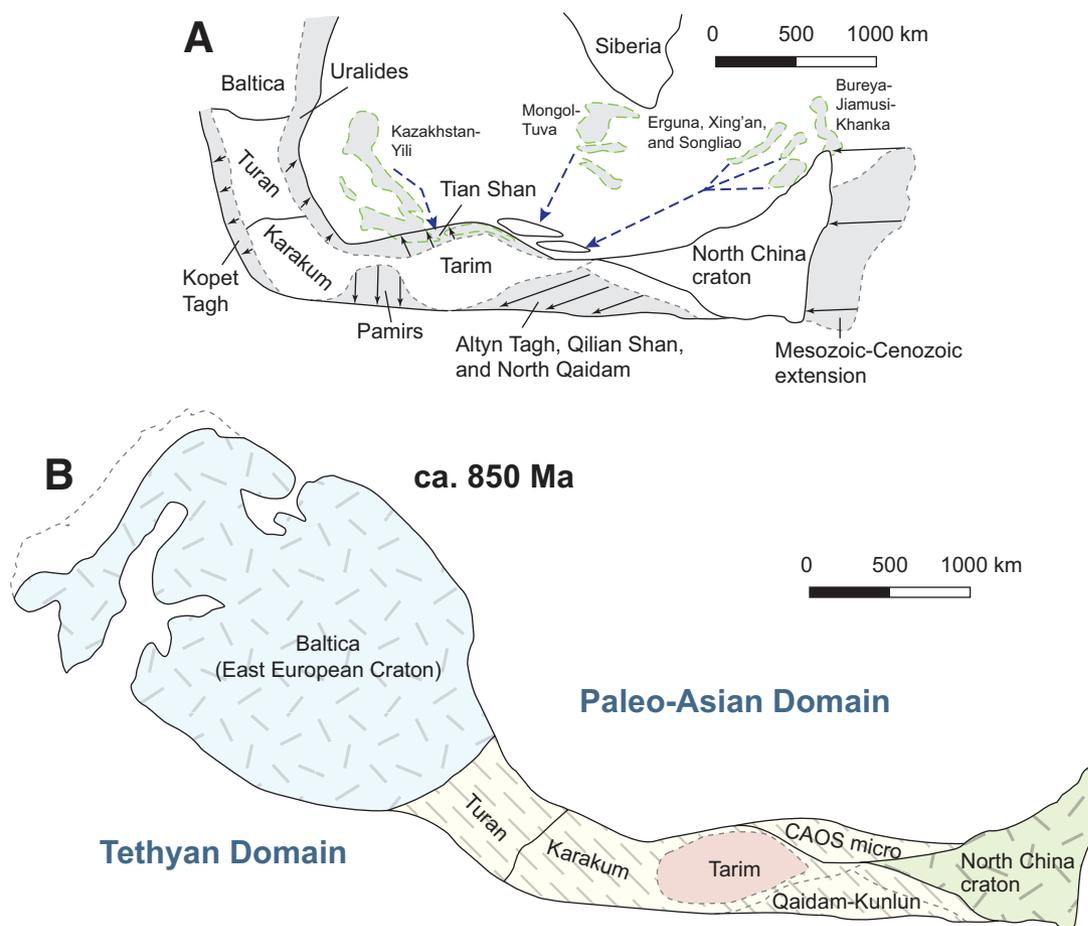


Figure 12. Schematic restoration of Balkatach (Baltica–Karakum–Tarim–North China) based on the geologic map of Asia of Ren et al. (2013). (A) Mesozoic and Cenozoic deformation is restored. Note that the Kazakhstan–Yili microcontinents are reconstructed within the restored Tian Shan thrust belt, and the microcontinents to the east in Mongolia, Russian, and China are restored to a position near the Tarim–North China interface. (B) Paleogeographic reconstruction of a continuous Balkatach continent in the Neoproterozoic. CAOS micro—Central Asian Orogenic System microcontinents.

son, 2000; Zhu et al., 2005; Green et al., 2008; Dupont-Nivet et al., 2010; Najman et al., 2010; Wang et al., 2011a; Hu et al., 2015, 2016), and the Mesozoic–Cenozoic development of the western Pacific subduction system (Figs. 1 and 12A).

Tian Shan, Pamirs, and Kopet Dagh

Cenozoic crustal shortening across the ~2000-km-long east-trending Tian Shan thrust belt decreases from ~200 km in the west to almost zero in the east (Avouac et al., 1993; Yin et al., 1998), consistent with the westward increase of the width of the belt (Fig. 12A). The Pamir and Western Kunlun Shan accommodate southward continental subduction of Tarim (Cowgill et al., 2003) to depths of >100 km (Burtman and Molnar, 1993). Crustal shortening of ~100 km has been estimated by balanced cross-section restoration in the Western Kunlun Shan (Cowgill, 2001) and isostatic equilibrium calculations (Lyon-Caen and Molnar, 1984). The Kopet Dagh range was thrust to the northeast over the Turan domain (Jackson and McKenzie, 1984; Thomas et al., 1999b); crustal shortening estimates are ~70–75 km (Lyberis et al., 1998). We restore this deformation by assuming that the pre-Cenozoic sutures on both sides of the Pamir Mountains were oriented as straight traces along great circles in the east-west direction (present-day coordinates) prior to the India-Asia collision (Fig. 12A). In order to accommodate the restoration of Cenozoic deformation in the westernmost Tian Shan (Avouac et al., 1993), Pamir Mountains (Burtman and Molnar, 1993), and Western Kunlun Shan (Cowgill et al., 2003), the Tarim craton must extend westward, and its geology can be traced through Tajikistan to the Karakum craton (e.g., Biske and Seltmann, 2010; Käšner et al., 2016) (Fig. 12).

Seismic reflection profiles across the Junggar basin to the north (Fig. 2) also suggest that post-Permian crustal shortening of at least tens of kilometers may have occurred (Song, 2006; Yang et al., 2012b). The Neoproterozoic connection between Tarim, Karakum, Turan, and Baltica is largely speculative due to >90% Mesozoic–Cenozoic cover, but all available geologic data sets (Figs. 4, 5, 6, 8, and 9) (e.g., Käšner et al., 2016) are consistent with these continents sharing a common Phanerozoic history (Fig. 12).

Altyn Tagh, Qaidam, and Qilian Shan

Cenozoic left-lateral motion along the Altyn Tagh fault was restored ~400 km based on offset piercing-point estimates in Yin and Harrison (2000), Yang et al. (2001), Gehrels et al. (2003b), and Cowgill et al. (2003) (Fig. 12A). This restoration places the Paleoproterozoic basement of Dunhuang (Zong et al., 2012) adjacent to basement of similar age in the North China craton (Zhang et al., 2012, 2013; cf. Long et al., 2014). In addition, geologic mapping (Yin et al., 2008a; Reith, 2013), analyses of seismic reflection profiles (Gao et al., 2013; Wang et al., 2014; Zusa et al., 2016), and geodetic data (Duvall and Clark, 2010; Zusa and Yin, 2016) reveal a Cenozoic strain gradient across the North Qaidam and Qilian Shan–Nan Shan thrust belts, from ~50% in the west to 25% in the

east, which is restored in this reconstruction (Fig. 12A). We assume that the strain magnitude across the Qimen Tagh thrust belt, south of the Qaidam Basin, is similar (Yin et al., 2007b) (Fig. 12A).

The Haiyuan and Kunlun strike-slip faults, and their terminal thrust belts to the east (i.e., the Liupan Shan and Longmen Shan thrust belts), are prominent features on the Tibetan Plateau today (Molnar and Tapponnier, 1975; Burchfiel et al., 1991; Zusa et al., 2017), but they minimally modify the pre-Cenozoic geology (e.g., Zusa and Yin, 2016). Offsets on the Kunlun fault vary from ~100 km in the west to <10 km in the east (Van der Woerd et al., 2000, 2002; Fu et al., 2005; Kirby et al., 2007), and offsets along the Haiyuan fault vary from ~90 km in the west to <15 km in the east (Burchfiel et al., 1991; Gaudemer et al., 1995; Ding et al., 2004). These faults may result from clockwise rotation of northern Tibet (e.g., Duvall and Clark, 2010; Zusa and Yin, 2016), and because they both parallel Phanerozoic sutures they do not obscure or distort earlier tectonic events or geometries.

North China and Eastern Asia

From the Late Jurassic through the Cenozoic widespread extension affected much of eastern Asia (e.g., Mongolia, northeast China, and the western Pacific) as the western Pacific trench system migrated eastward (e.g., Traynor and Sladen, 1995; Ren et al., 2002; Yin, 2010) (Fig. 1). Subduction rollback, postorogenic collapse, and regional extension began at ca. 120 Ma (Davis et al., 1996, 2001; Graham et al., 2012), expressed by the formation of northeast-to north-northeast-trending rift (or transtensional) basins with synrift volcanic rocks (Traynor and Sladen, 1995; Deng et al., 1999; Zhang et al., 2014b) and marginal sea basins in the western Pacific. Extensional detachment faults have exposed mid-crustal rocks throughout eastern Asia (Zheng and Ma, 1991; Davis et al., 1996, 2001; Webb et al., 1999; Zhu et al., 2012). Based on the widespread exposure of mid-crustal rocks and rift basins, we assume that a total magnitude of 100% extensional strain affected the Huabei–Korea–Japan region (i.e., North China craton) of East Asia (see Yin, 2010) since ca. 120 Ma (Davis et al., 2001); we restore ~550 km of extension during this time (Fig. 12A).

Late Paleozoic–Mesozoic Intracontinental Deformation and Ocean Closure

As the Paleo-Asian Ocean closed against Balkatach in the late Paleozoic (Fig. 4), the northern passive margin rocks of Balkatach were deformed. In the Uralides, an unknown magnitude of the eastern passive margin of Baltica subducted beneath Kazakhstan (Brown et al., 2011); similar events occurred along the Tian Shan–Ying Shan and Solonker–Jilin–Yanji sutures (Fig. 1). The northern continental margin of Tarim may have subducted below a Devonian arc (Charvet et al., 2011), making its original size larger than is exposed today. Restoration of these northern margins along the Paleo-Asian Ocean is highly

speculative, and thus we only restore tens of kilometers that may have been involved in continental subduction and/or underthrusting (e.g., within the Uralides and the Tian Shan) (Fig. 12A).

The magnitude of deformation associated with the closure of the Paleo-Tethys along the southern margin of Balkatach is also poorly understood. Any shortening associated with the collision of the Qiangtang terrane and South China craton with Balkatach is not restored, but we acknowledge that shortening along these margins may be hundreds of kilometers.

Neoproterozoic Rifting and Paleozoic Orogeny

Restoration of Cenozoic intracontinental deformation shows that the Tarim and North China cratons were continuous prior to the Mesozoic. The late Paleozoic–early Mesozoic Kunlun arc suggests that Qaidam-Kunlun continent and Tarim craton were linked at this time, which requires the North China craton to also be linked with the Qaidam-Kunlun continent. The ca. 1.0–0.9 Ga granitoids (and younger 0.9–0.8 Ga metamorphic rocks) found throughout Tarim and Qaidam-Kunlun further support their Neoproterozoic linkage (e.g., Figs. 9 and 10).

The abundant Neoproterozoic rift and Cambrian passive margin strata deposited along the outer margins of Balkatach suggest that the continent was involved in the global-scale breakup of the Rodinia supercontinent (Fig. 6). Rifting was followed by several Paleozoic orogens as separate and distinct microcontinents collided against Balkatach, including the Qaidam-Kunlun, Tagil, and Magnitogorsk microcontinents and arcs. Several microcontinents were involved coevally in the initiation and evolution of the CAOS throughout the Paleozoic. This major transition from rifting to collisional orogeny fundamentally depends on whether the colliding microcontinents were exotic or genetically derived from Balkatach. In the first scenario (i.e., exotic microcontinents), the continents that must have rifted and separated from Balkatach's margins, leading to Neoproterozoic passive margin development, must have drifted far enough away from Balkatach to allow for new exotic microcontinents to travel across the Paleo-Asian Ocean and collide against Balkatach's margins in the Paleozoic. This is envisioned in the reconstruction by Stampfli and Borel (2002) wherein, following Neoproterozoic rifting, Gondwana-derived ribbons of continental crust crossed the Paleo-Tethys and collided with the southern margin of Balkatach and Laurasia (e.g., the Hunia and Galatia continents; see also von Raumer and Stampfli, 2008; Stampfli et al., 2011, 2013).

In the latter alternative scenario, as presented in Hsü and Chen (1999), Neoproterozoic rifting caused several large continents to separate and drift away from Balkatach, whereas some smaller fragments partially and/or completely rifted but remained nearby within the Paleo-Asian and Tethyan Oceans. The rifted microcontinents served as nuclei for the initiation of the CAOS; continental and oceanic subduction within the Paleo-Asian Ocean eventually caused the microcontinents to collide back against Balkatach. Below we provide evidence and rationale for this second model.

Origin of the Qaidam-Kunlun and CAOS Microcontinents

The early Paleozoic Qilian suture indicates that a late Neoproterozoic–early Paleozoic ocean separated North China and Qaidam-Kunlun, but as previously discussed, there is no documented suture that divides the Qaidam-Kunlun continent from Tarim (Fig. 10). The observation that the Qaidam Precambrian basement is similar to the basement of Tarim–North China (e.g., Gehrels et al., 2003b; Wu et al., 2016; Yu et al., 2017) suggests that Qaidam-Kunlun was originally part of Tarim–North China. In this scenario, the Qaidam-Kunlun partially rifted away from the North China margin as a thin, still attached, continental strip in the Neoproterozoic (scenario 3 in Fig. 10C). Other researchers argue that Qaidam-Kunlun was derived from the South China craton based on correlative ca. 950–900 Ma granites, but this is not a unique signature among Precambrian crustal continents, because 1.1–0.8 Ga igneous rocks are found within North China, Tarim, the Tian Shan, and the CAOS microcontinents (Fig. 5) (e.g., Gehrels et al., 2003a, 2011; Chen et al., 2006; Lu et al., 2008; Zhu et al., 2008; Chen et al., 2009; Peng, 2010; Hu et al., 2010; Peng et al., 2011a, 2011b; Wang et al., 2012; Liu et al., 2012; He et al., 2012; Ge et al., 2013a, 2014; Dan et al., 2014).

The Precambrian central Tian Shan microcontinent was probably derived from Tarim. Several workers have also suggested the connection between some of the Mongolian CAOS microcontinents and Tarim–North China on the basis of detrital zircon age distributions (e.g., Rojas-Agramonte et al., 2011, 2014; Levashova et al., 2011). Examination of the pre–750 Ma detrital zircon age distribution reveals similar age peaks at ca. 2.5 Ga, 1.85–1.8 Ga, ca. 970–910 Ma, and 790–750 Ma (Fig. 5). The northeastern Gondwana margin also shares similar age peaks (Fig. 5), and a Gondwana origin for these microcontinents cannot be ruled out, but we follow the suggestion of Rojas-Agramonte et al. (2011) and favor a Tarim origin. Similarly, Han et al. (2011) inferred that the Erguna-Xing'an-Songliao microcontinents in northern China were derived from Tarim–North China on the basis of U-Pb detrital zircon ages (Fig. 5), and Zhou et al. (2017) suggested that most of the microcontinents existed close to the Tarim–North China continent at ca. 750 Ma.

Paleomagnetic Data

The apparent polar wander paths from paleomagnetic poles are commonly used to develop and test plate tectonic reconstructions. The goal of this study is to focus on the geologic histories and relationships of each continent to assess its paleogeographic location. However, preliminary paleomagnetic analysis (Fig. 13; Table 2) was used to constrain the tectonic reconstruction presented herein and to test its viability. Ultimately, there are too few reliable paleomagnetic poles for the individual Balkatach continents to draw significant quantitative conclusions, but available high-quality paleomagnetic data place important paleolatitude bounds that were incorporated into our final tectonic model.

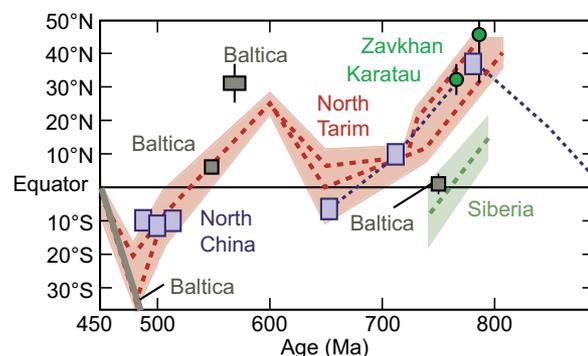


Figure 13. Paleomagnetically determined paleolatitude versus age for the North China, Tarim, Baltica, and Siberia cratons. Two Central Asian Orogenic System microcontinents are also shown: Zavkhan and Lesser Karatau. Note that the hemisphere polarity is ambiguous. Error bars or envelopes are shown, except when not provided or when the error is smaller than the symbol used. Age error in all studies is indiscernible in this figure. See Table 2 for sources used in this compilation.

The connections between the CAOS microcontinents and Tarim–North China discussed herein are supported by paleomagnetic data. The estimated Precambrian paleolatitude of the Karatau and Zavkhan (Baydaric or Dzabkhan) microcontinents (Fig. 2) is similar to North China and Tarim but not Siberia at 805–770 Ma (Fig. 13) (Levashova et al., 2010, 2011). Apparent separations of Baltica, Tarim, and North China at ca. 560 Ma and ca. 750 Ma (Fig. 13) (e.g., Cocks and Torsvik, 2013; Torsvik and Cocks, 2013) are small enough to be explained by the elongate geometry of the Balkatach continent. The $\sim 25^\circ$ latitude separation between Baltica and North China at ca. 650 Ma corresponds to a distance of ~ 2800 km on Earth's surface, and given that the restored length of Balkatach is 4000–5000 km, these paleolatitude data may reveal that the continent had a north-trending orientation. Any minor apparent separation between continents may be the result of the distance between paleomagnetic sample sites. These data sets may provide insights into the orientation of this long continent; the continent was oriented mostly north-south at ca. 750 Ma but rotated to east-west at the start of the Cambrian. Cocks and Torsvik (2013) showed $<5^\circ$ north-south latitude separation between North China and Tarim throughout the Paleozoic, which is within error of the paleomagnetic data and suggests a possible north-south orientation of the eastern arm of Balkatach (i.e., Tarim–North China) at ca. 500–400 Ma. Future examination of paleomagnetic data sets and apparent polar wander paths in the context of the Balkatach hypothesis is necessary to interpret any of these observations.

Were Tarim and North China Linked in the Neoproterozoic?

Critical to our reconstruction is that the Tarim and North China cratons were connected in the Neoproterozoic. Zircon geochemistry and geochronology suggests that both continents may have been tectonically linked in the Ar-

chean–Paleoproterozoic (e.g., Zhao et al., 2001a, 2001b; Ge et al., 2013b; Zhang et al., 2014a; Li et al., 2015), but there is a long-standing view that the North China and Tarim cratons were separate in the Neoproterozoic–early Paleozoic (e.g., Rojas-Agramonte et al., 2011; Gong et al., 2012; He et al., 2012; Zhang et al., 2013; Cocks and Torsvik, 2013; Li et al., 2015). The apparent separation is based in part on the lack of 1.1–0.9 Ga (i.e., so-called Grenville aged) zircon grains in the North China craton (e.g., Zhang et al., 2012; Li et al., 2015), and the broad correlation of Tarim rocks with other Grenville-aged cratons, including South China or parts of Gondwana (e.g., Wan et al., 2001; Tung et al., 2007, 2013; Lu et al., 2008; Song et al., 2013; Xu et al., 2015). These correlations are often based on broad zircon-age time spans involving 1.3–0.9 Ga zircons (i.e., spanning 400 m.y.), not particularly useful for correlating continents, and zircon signatures of this age are found in East Antarctica, India, Australia, Central Asia, South China, and Tarim (e.g., Fitzsimons, 2000; Peng et al., 2011a, 2011b; Chattopadhyay et al., 2015). However, we believe that the following points provide robust evidence that the North China and Tarim cratons may have been contiguous and connected at that time.

1. Restoration of Cenozoic deformation, including shortening and left-lateral strike-slip fault offset, matches Paleoproterozoic and Paleozoic–Mesozoic geology in the North China and Tarim cratons. Specific alignments include Paleoproterozoic basement localities (Zong et al., 2012; Zhang et al., 2012, 2013); early Paleozoic suture, arc, and UHP metamorphic rocks in the Altyn Tagh Range, Qilian Shan, and North Qaidam (e.g., Sobel and Arnaud, 1999; Yin and Harrison, 2000); and arc rocks from the Permian–Triassic Kunlun arc in the Western and Eastern Kunlun Shan (e.g., Cowgill et al., 2003; Xiao et al., 2005; Wu et al., 2016). Thus, by at least the early Paleozoic, the North China and Tarim cratons were contiguous.

2. Given that 1.0–0.9 Ga plutons in the Tarim craton have been ascribed a volcanic arc-subduction system origin (e.g., Ma et al., 2012; Ge et al., 2014), finding similar plutons >1500 km to the east-northeast distributed throughout North China would be unexpected (present-day orientations and distances). The inferred subduction system would only have developed magmatic-intrusive rocks along the margins of the Neoproterozoic continent, not throughout the craton's interior. A lack of 1.0–0.9 Ga zircons across the North China craton is not substantial evidence that this craton was separate from the Tarim craton in the Neoproterozoic.

3. There is growing acceptance that the basement of the Qaidam–Kunlun continent was derived from the North China craton (e.g., Zhang et al., 2013; Wu et al., 2016; 2017a), which requires that Grenville-aged plutons did intrude the broader North China continent, given that there are 1.0–0.9 Ga intrusive rocks in the Qilian Shan (e.g., Cowgill et al., 2003; Gehrels et al., 2003a, 2003b; Wu et al., 2016, 2017a; Fig. 9). We note that this observation is not specifically required for our model to be viable, but it does suggest that this arc system probably developed along northern Tarim and the restored southwestern margin of the North China craton.

4. Mesoproterozoic strata from across the Altyn Tagh Range, the Qilian Shan, and southwestern North China (Longshoushan region) are correlative

based on detrital zircon ages, including zircon age peaks at ca. 1.2 Ga, ca. 1.4–1.5 Ga, 1.7 Ga, and 2.5 Ga (Gehrels et al., 2003a, 2003b; Z. Zhou, 2016, personal commun.; Wu et al., 2017a; C. Wu, 2017, personal commun.). Zircon ages of 1.2 Ga and 1.5 Ga are somewhat rare in Asia (Fig. 5) (Demoux et al., 2009; Rojas-Agramonte et al., 2011), suggesting that these regionally extensive Mesoproterozoic units that extend from the northeastern Tarim Basin to North China shared a similar source (Wu et al., 2017a). Furthermore, a possible source for the ca. 1.5 Ga zircons could have been the 1.52 Ga gneiss from the Baga Bogd massif (Demoux et al., 2009), one of the Central Asia microcontinents, which restores to a position just north of the Tarim–North China junction (Fig. 12A).

5. Conversely, if the Tarim and North China cratons were separate in the Neoproterozoic, then they must have collided sometime in the Paleozoic. Some argue for a late Paleozoic collision (e.g., Cocks and Torsvik, 2013) but the

continuity of the Permian–Triassic Kunlun arc in the Western and Eastern Kunlun Shan (e.g., Cowgill et al., 2003; Xiao et al., 2005; Wu et al., 2016) requires that these continents were connected before this time. It is important to note that there has been no identified late Paleozoic–early Mesozoic collisional belt or suture zone between the two continents. We cannot rule out the possibility that such rocks may be covered by younger sediments.

TECTONIC EVOLUTION OF BALKATACH

We now retrodeform Central Asia, with emphasis on the evolution of Balkatach, from the Proterozoic to the present. Presented here are 11 time slices from the Mesoproterozoic to the present in Figures 14–18, with higher spatial

Mesoproterozoic (shown at ca. 1.4 Ga)

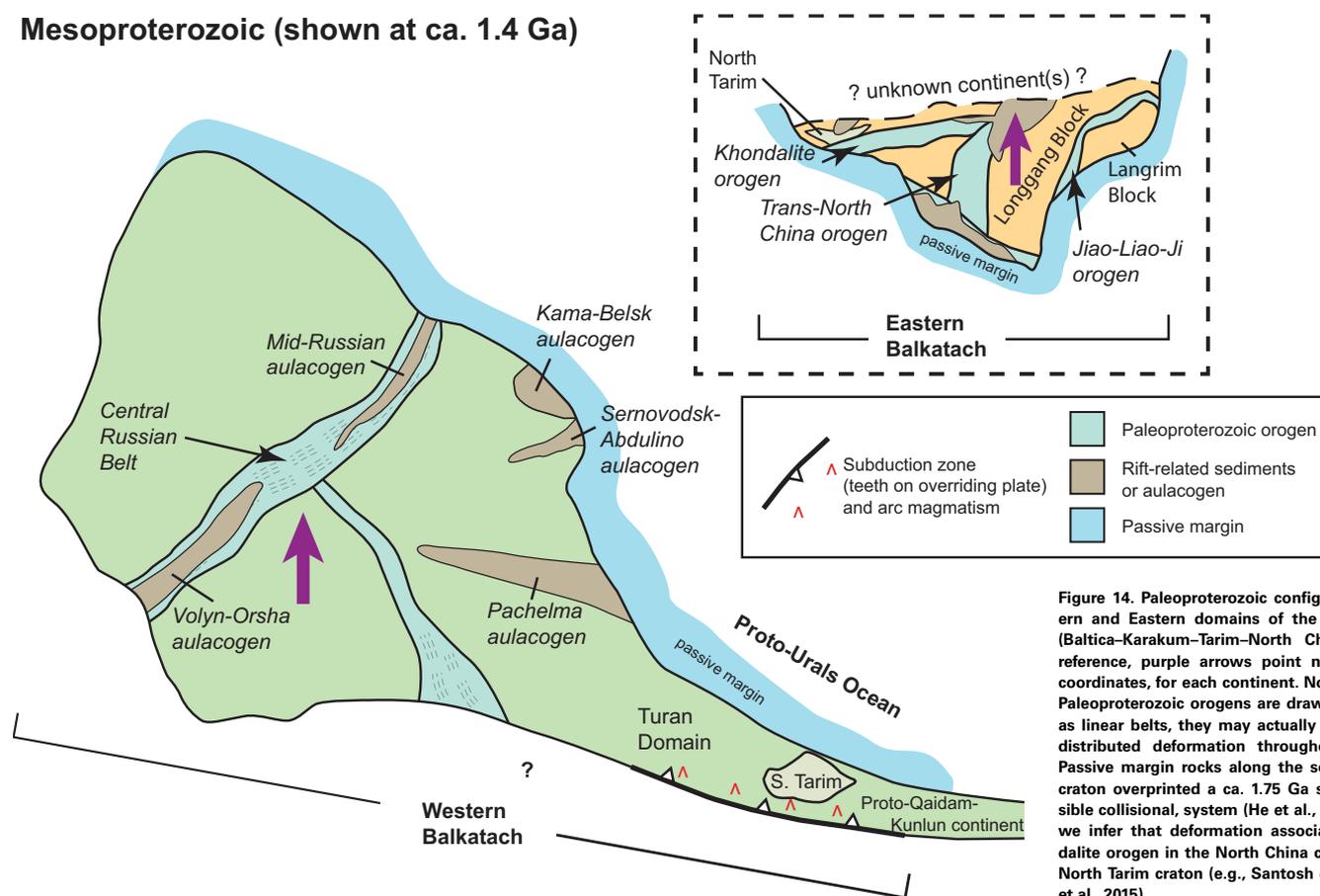


Figure 14. Paleoproterozoic configuration of the Western and Eastern domains of the proposed Balkatach (Baltica–Karakum–Tarim–North China) continent. For reference, purple arrows point north, in present-day coordinates, for each continent. Note that although the Paleoproterozoic orogens are drawn on each continent as linear belts, they may actually involve more widely distributed deformation throughout each continent. Passive margin rocks along the southern North China craton overprinted a ca. 1.75 Ga subduction, and possible collisional, system (He et al., 2008). Also note that we infer that deformation associated with the Khondalite orogen in the North China craton extends to the North Tarim craton (e.g., Santosh et al., 2006, 2007; Ge et al., 2015).

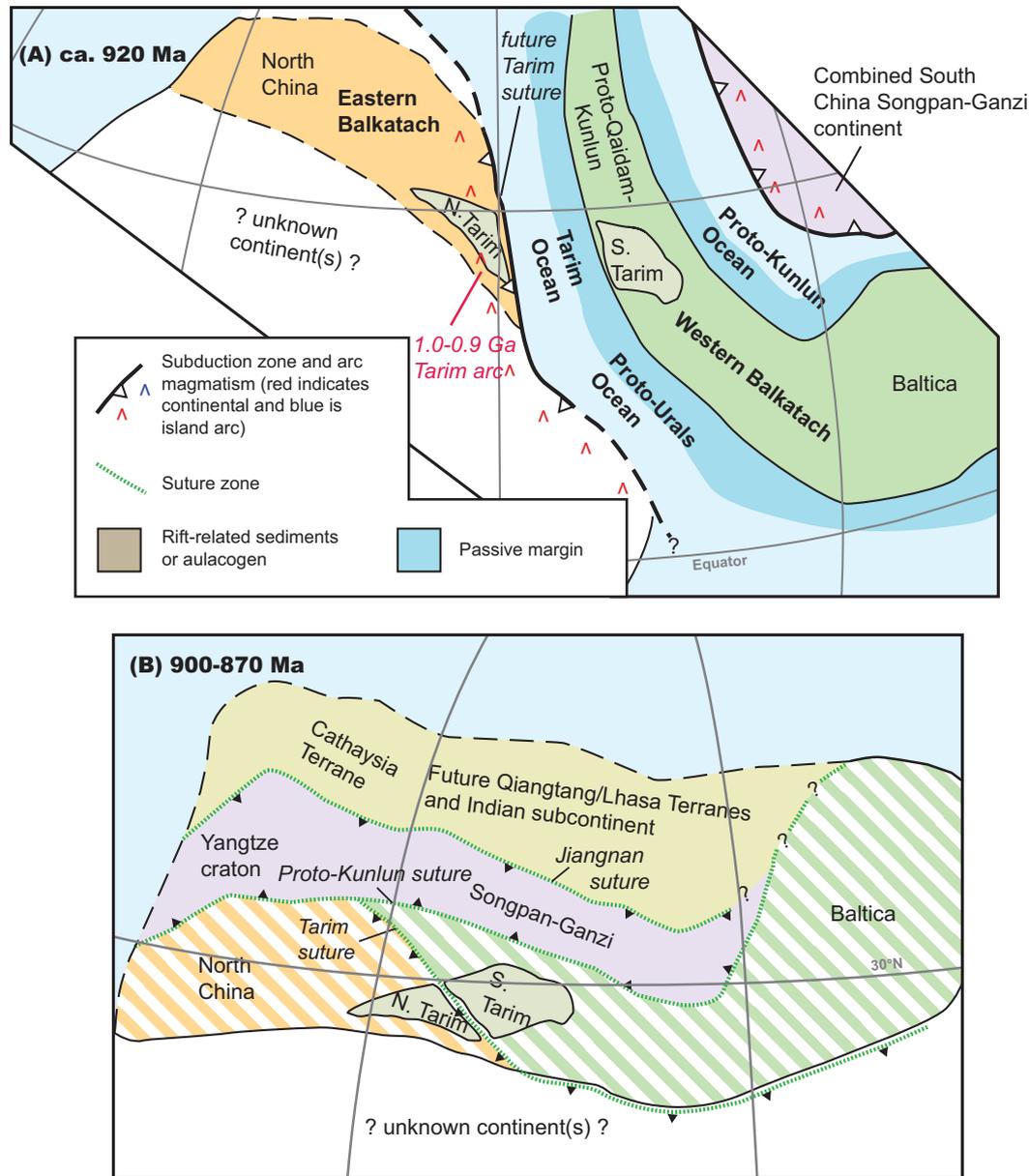


Figure 15. Neoproterozoic–early Cambrian tectonic evolution of Balkatach (Baltica–Karakum–Tarim–North China); note that continents are oriented 180° from Fig. 14. (A) Convergence of Eastern and Western Balkatach was accommodated by the ca. 1.0–0.9 Ga Tarim arc. Eastern Balkatach was likely connected to an unknown continent to the southwest during this time. The combined South China–Songpan–Ganzi continent approached the northern margin of Balkatach (i.e., its present-day southern margin; Wu et al., 2016). (B) By ca. 870 Ma, the Balkatach continent reached its maximum extent (now shown with white stripes as in Fig. 2). The future Qiangtang, Lhasa, and Indian subcontinent terranes were affixed along the northern margin of the Songpan–Ganzi terrane (Wu et al., 2016). (Continued on following page.)

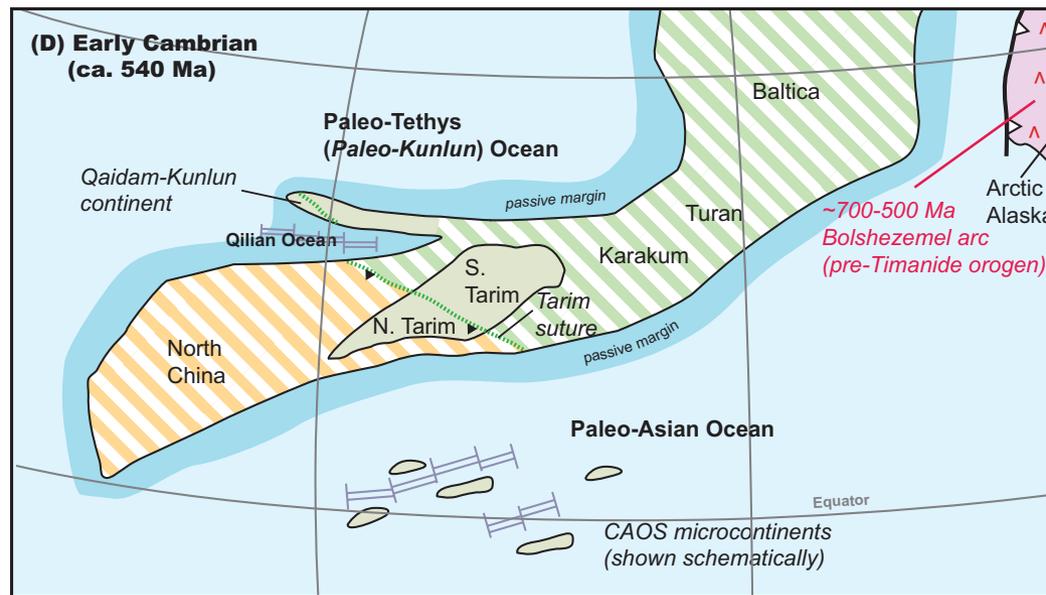
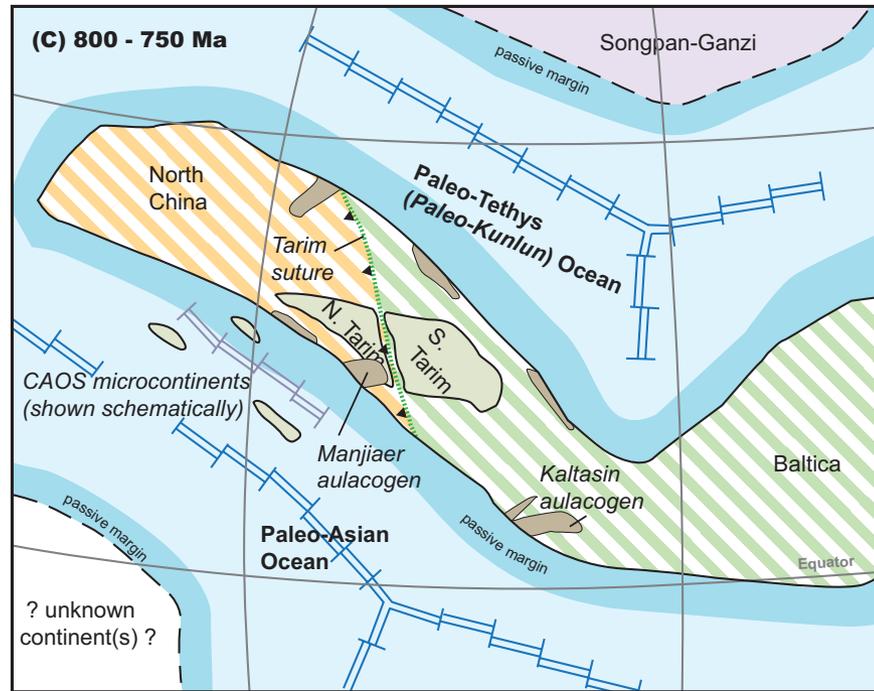


Figure 15 (continued). (C) Rifting along the southwestern and northeastern margins of the Balkatach continent led to the opening of the Paleo-Asian and Paleo-Tethys Oceans, respectively. The Central Asian Orogenic System (CAOS) microcontinents are distributed throughout the Paleo-Asian Ocean. (D) A later stage of rifting caused the Qaidam-Kunlun continent to partially separate from the North China-Tarim continent, causing the peninsular opening of the Qilian Ocean in the late Neoproterozoic-early Cambrian. The Bolshezemel arc accommodated the convergence of the composite Arctic Alaska continent with the eastern margin of Balkatach. Three-dimensional globe projection; latitude and longitude lines are in 30° intervals. Arctic Alaska (Strauss et al., 2013) position and Timanide orogen were adapted from Kuznetsov et al. (2010).

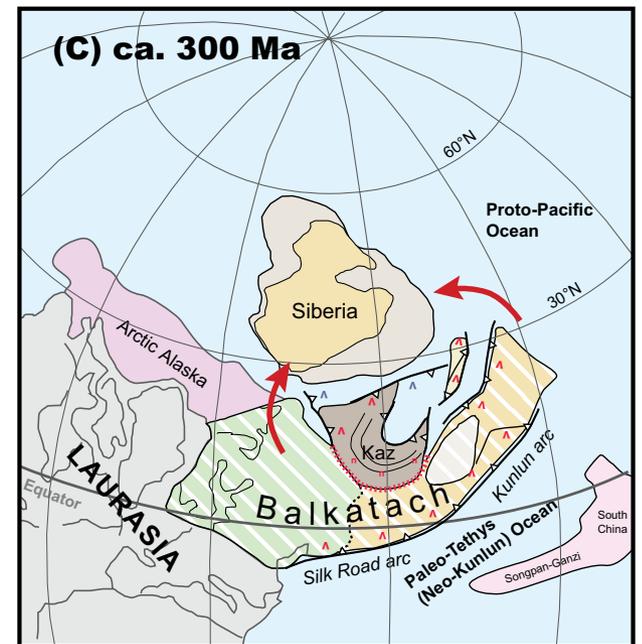
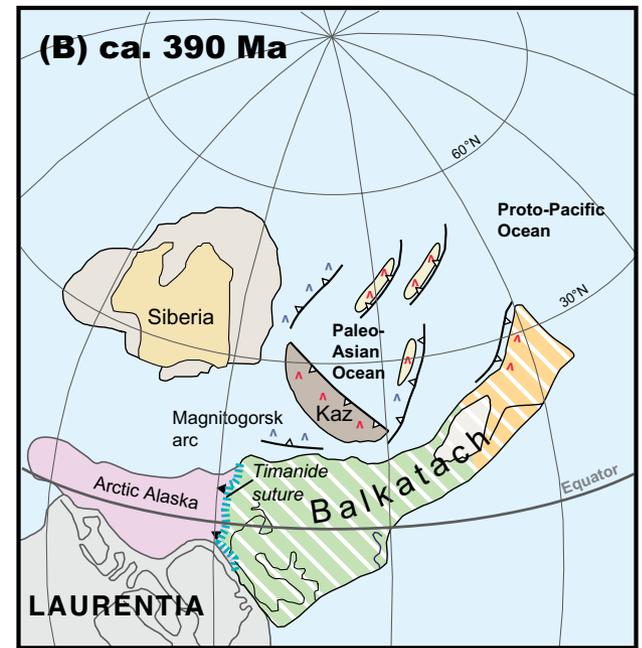
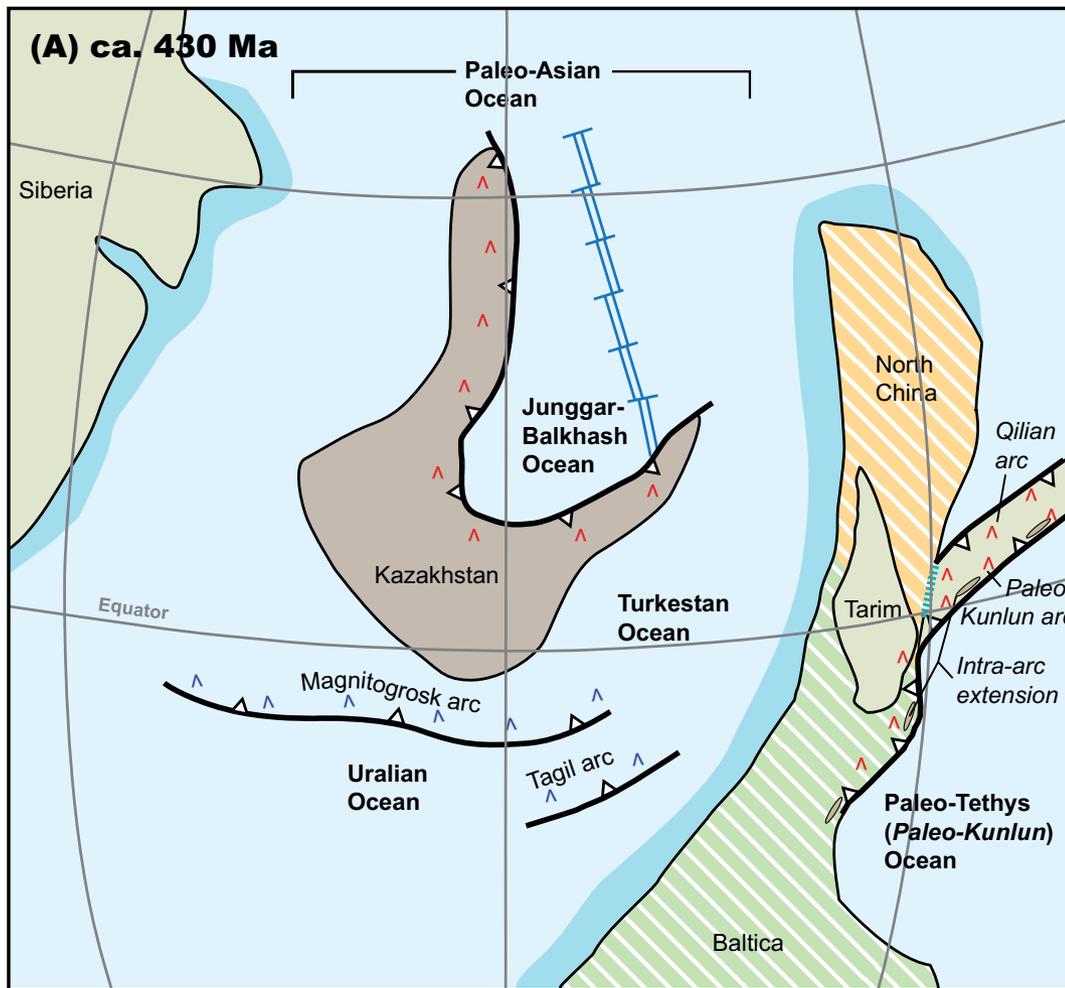


Figure 16. Paleozoic tectonic evolution of Balkatach (Baltica–Karakum–Tarim–North China) and the Paleo-Asian Ocean domain. (A) Subduction of the Paleo-Asian Ocean within the Central Asian Orogenic System caused Balkatach to wrap around the Paleo-Asian Ocean. (B) Continued closure of the Paleo-Asian Ocean accommodated the Balkatach convergence with the Kazakhstan (Kaz) microcontinent conglomeration. (C) Collision between Balkatach and Kazakhstan occurred. This figure is modified from Filippova et al. (2001), Abrajevitch et al. (2008), and Xiao and Santosh (2014). Three-dimensional globe projection; latitude and longitude lines are in 30° intervals.

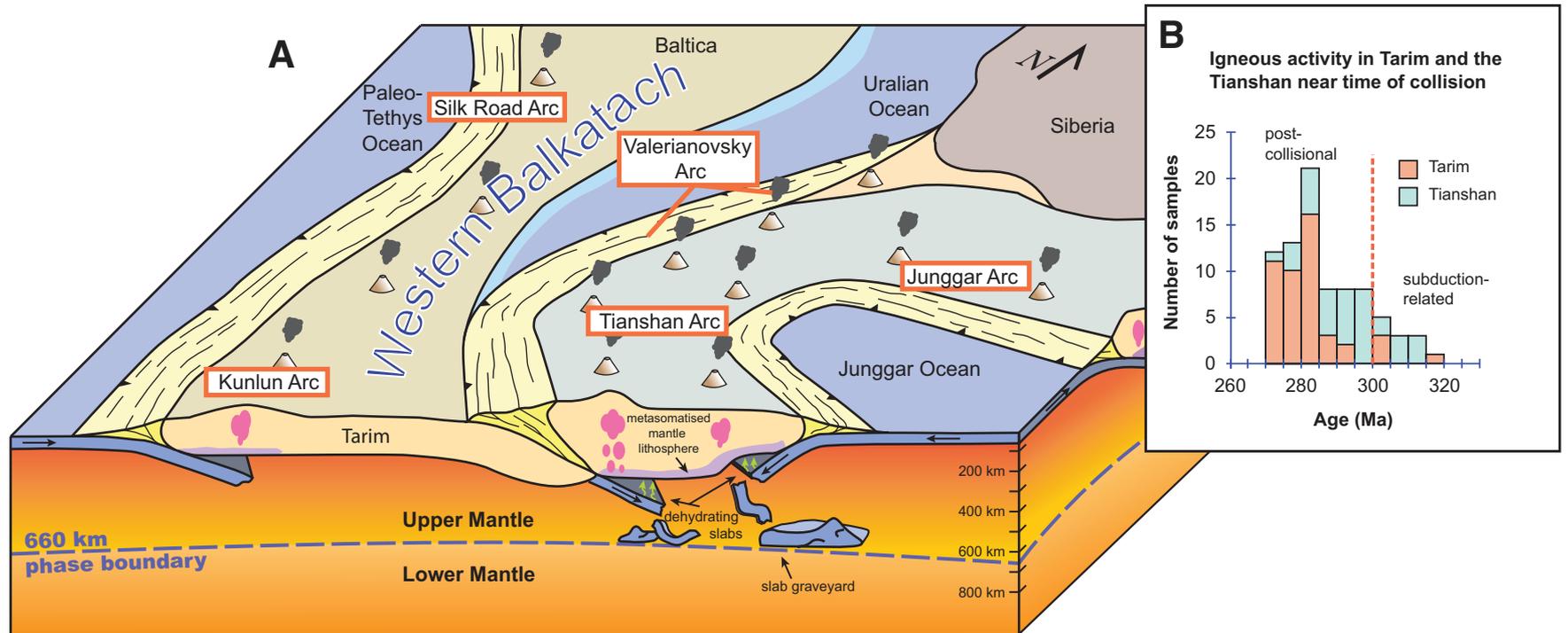


Figure 17. The Kazakhstan orocline and the destruction of the Paleo-Asian Ocean. The large volume of subducted slabs may have collected near the 660 km phase transition, and their eventual sinking across this boundary triggered significant asthenospheric upwelling and the observed Permian magmatic flareup (e.g., Cina, 2011). Not to scale. (B) Histogram of observed granitoid ages in the Tian Shan and Tarim regions, in the area depicted in A (modified after Biske and Seltmann, 2010; Xiao et al., 2013). Note the large amount of igneous activity following the closure of the Paleo-Asian Ocean (signified with red dashed line). Data sources include Mikolaichuk et al. (1997), Kiselev (1999), Zhang et al. (2007a), Konopelko et al. (2008), Apayarov (2010), Glorie et al. (2010), Tian et al. (2010), Yu et al. (2011), Zhang et al. (2010, 2012, 2013), and Seltmann et al. (2011).

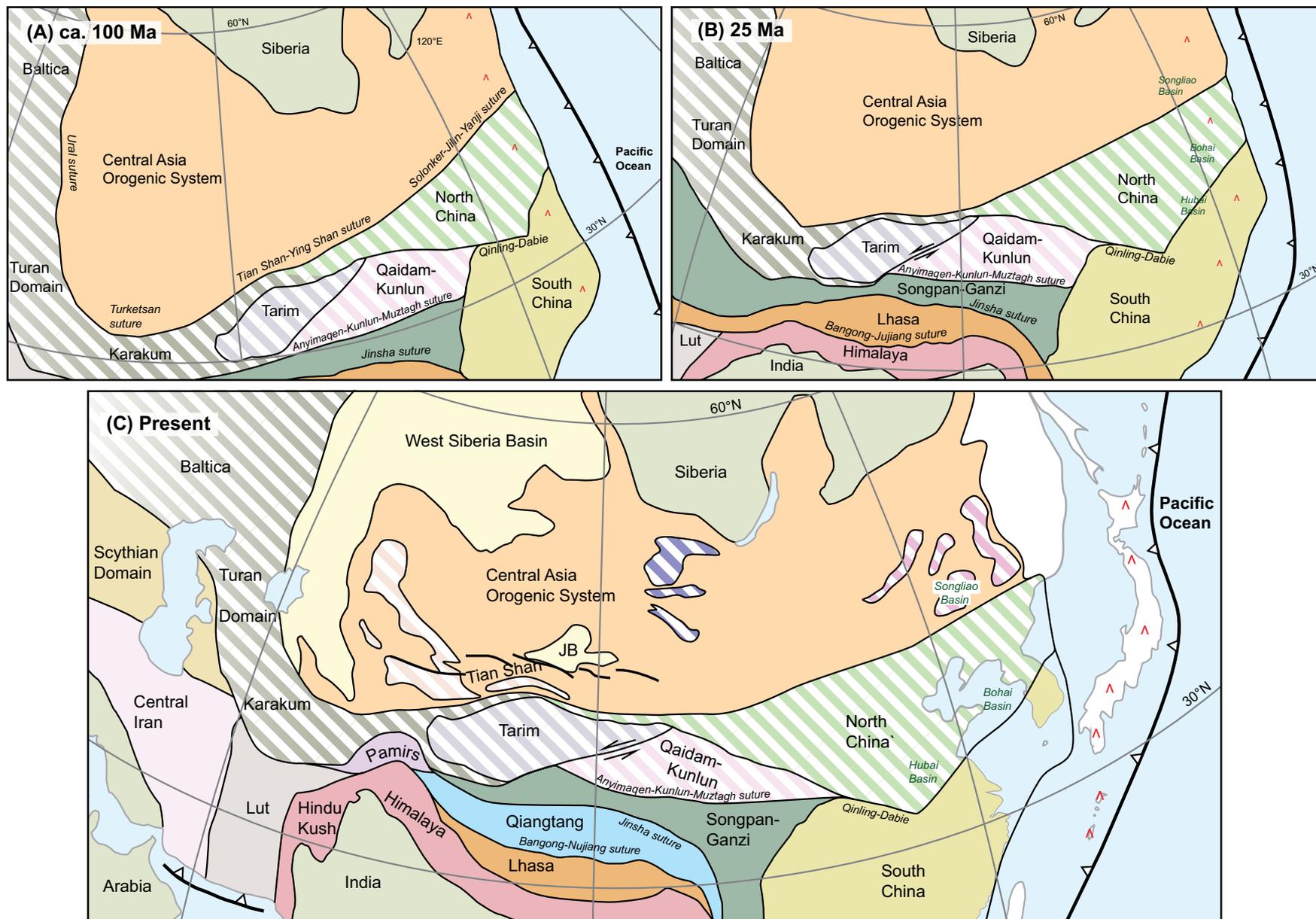


Figure 18. Evolution of Central Asia as restored in our reconstruction. Note that the colors of the tectonic units change to match those in Figure 2, and the present-day boundaries are the same. Three-dimensional globe projection; latitude and longitude lines are in 30° intervals. Proposed Balkatach (Baltica–Karakum–Tarim–North China) continent is shown with white stripes. (A) Middle Cretaceous. (B, C) Cenozoic.

and temporal resolution progressing through time. Note that this reconstruction focuses more heavily on the eastern segment of Balkatach (i.e., east of the Tarim craton), which is due to the scarcity of geologic data in the west. Although there are numerous localized reconstructions that focus on one (or several) specific regions (e.g., Tarim, the Qilian orogen, North China, and the CAOS; Filippova et al., 2001; Biske and Seltmann, 2010; Zhao and Zhai, 2013; Song et al., 2013, 2014; Xiao et al., 2013), none shows how all of these regions relate and evolve through time and space. This model is a preliminary attempt at constraining the evolution of Balkatach and its constituent cratons, and adheres to our current knowledge of later tectonic modification. Quantitative paleopositions are not given for the Mesoproterozoic time, but are for later reconstructions. The evolution of the CAOS is largely after Filippova et al. (2001), with considerations from Windley et al. (2007) and Xiao et al. (2013).

Paleoproterozoic

Eastern Balkatach (North Tarim and North China)

By the Paleoproterozoic, Archean crustal fragments consisting of orthogneiss, TTG gneiss, metamorphosed supracrustal rocks, and greenstone belts as old as 3.8 Ga were amalgamated into the North China craton. The Neoarchean crust of the North Tarim continent formed separately from the North China craton (Long et al., 2010, 2014), although we postulate that they joined by the Paleoproterozoic, because no later collisional events are observed (cf. Yuan and Yang, 2015). At the Paleoproterozoic start of this reconstruction, the North Tarim and North China cratons are together (e.g., Li et al., 2015).

Western Balkatach (Baltica, Turan, Karakum, and South Tarim)

The collision and amalgamation of Samatia, Fennoscandia, and Volgo-Uralia of Baltica was complete by ca. 1.8 Ga (Bogdanova et al., 2006; Shchipansky et al., 2007). The Turan, Karakum, and South Tarim continents were also connected to a unified Baltica craton.

Mesoproterozoic

Intracratonic rifting and aulacogen formation affected both the western and eastern domains of Balkatach throughout the Mesoproterozoic, which suggests that other unknown continents were affixed to their respective margins prior to continental breakup (see Fig. 14). The eastern margin of Western Balkatach (i.e., the Ural Ocean-facing side) and southern margin of Eastern Balkatach underwent a transition to passive margin sedimentation, suggesting that these margins separated from their conjugate continents, although it is unclear if Balkatach entirely separated from other continents at that time.

Neoproterozoic

Subduction of the Tarim–Proto-Ural Ocean underneath east Balkatach led to the development of the 970–910 Ma Tarim arc, which extended from the Qinling–Qilian–Qaidam–Kunlun regions through the composite Tarim continent into an unknown continent (see Fig. 15). This subduction system accommodated the convergence of Western and Eastern Balkatach, which collided along the Tarim suture (e.g., Guo et al., 2005; Xu et al., 2013) by ca. 870 Ma, at which point Balkatach reached its full extent. Intracontinental deformation may have persisted from this collision until ca. 800 Ma (e.g., Ge et al., 2016). The combined Songpan–Ganzi and South China cratons collided against the southern margin (present-day coordinates) of Balkatach at a similar time (Wu et al., 2016), although this collision must have occurred prior to the initiation of bimodal volcanism documented in southern Tarim at ca. 900 Ma (e.g., Wang et al., 2015a, 2015b).

Immediately after and/or during the collision of Western and Eastern Balkatach, rifting commenced along the continent's northern and southern margins. Several aulacogens developed obliquely to the inferred continental margins: in the east the Helan and Manjiaer aulacogens formed on either side of Balkatach (Lin et al., 1995, 2014) and in the west synrift sedimentation was more diffuse and several Mesoproterozoic aulacogens were reactivated as rift basins (e.g., Kaltasin and Sernovodsk–Abdulino aulacogens in Baltica). Rifting of Balkatach from an unknown continent to the north and the Songpan–Ganzi–South China continent to the south (Wu et al., 2016), respectively, opened the Paleo-Asian Ocean at ca. 800 Ma and the Paleo-Tethys Ocean (or Paleo-Kunlun) at ca. 830 Ma.

The rifting process was complex and relatively long lived, leading to protracted rift-related volcanism and volcanic passive margin development. The warmed lithosphere ultimately led to rifting of the CAOS microcontinents and the partial incomplete opening of the Qilian Ocean from 750 to 650 Ma. These rift events may be analogous to the continental fragments that rifted from northern Australia in the Mesozoic–Cenozoic (e.g., Hall, 2011; Metcalfe, 2011).

Early–Middle Paleozoic

The Paleo-Asian Ocean persisted throughout the Paleozoic, and may have been divided into as many as four interconnected oceans by the Kazakhstan continent (see Fig. 16). In this reconstruction we follow the archipelago models of Filippova et al. (2001), and envision Balkatach wrapping around these accretionary arcs (e.g., Abrajevitch et al., 2008; Xiao and Santosh, 2014).

Much of the northern and southern margins of Balkatach, bounding the Paleo-Asian and Paleo-Tethys Oceans, respectively, remained passive throughout most of the Paleozoic (Fig. 6), although Western Balkatach underwent at least two arc-continent collision events as the Tagil and Magnitogorsk arcs collided with the proto-Urals in the Devonian–early Carboniferous. The closure of the Qilian Ocean in the earliest Silurian occurred during the Qilian–Qinling

orogen. This collision progressed from west to east, consistent with the closure of a peninsular Qaidam-Kunlun rotating counterclockwise. The north-dipping paleo-Kunlun arc was constructed on the southern margin of Qaidam-Kunlun and may have extended as far west as Karakum. Intra-arc extension led to the deposition of several Devonian basins in the Karakum and Kunlun regions.

Middle Carboniferous collision of central Western Balkatach with Kazakhstan occurred during the southern Uralide orogeny. Following collision, bidirectional suturing propagated northward (parallel to the strike of the Urals), as Balkatach wrapped around the CAOS toward Siberia, and eastward as the Paleo-Asian Ocean was consumed along the Solonker-Jilin-Yanji suture (Figs. 4, 16C, and 17). This process led to internal oroclinal folding of the central Asian microcontinents, including the Kazakhstan and the Mongolian continents (e.g., Xiao and Santosh, 2014; Kilian et al., 2016). Note that counterclockwise oroclinal folding of these microcontinents is consistent with the inferred bidirectional suturing along Balkatach's northern margin.

Late Paleozoic

The bidirectional closure of the Paleo-Asian Ocean continued along the northern margin of Balkatach into the late Paleozoic (Fig. 4), and by ca. 300 Ma the Pacific Ocean was separated from the Paleo-Asian Ocean (Fig. 16C). The complete closure of the Paleo-Asian Ocean in the Permian was accompanied by a widespread magmatic flare-up, which may have been related to the avalanche of subducted oceanic slabs of the Paleo-Asian Ocean across the 660 km phase boundary in the mantle (e.g., Schubert and Tackley, 1995; Cina, 2011) (Fig. 17). There was an enormous amount of oceanic crust that subducted in the CAOS with the closure of the Paleo-Asian Ocean. If these slabs stagnated at the 660 km phase transition (Schubert and Tackley, 1995) before simultaneously breaking through this phase-transition boundary, there would be a significant amount of asthenospheric upwelling that could melt the metasomatized lithospheric mantle and trigger widespread Permian magmatism across Central Asia (Fig. 17).

Mesozoic–Cenozoic

Collision between Balkatach and the CAOS continued from central Balkatach (i.e., Karakum and Turan) eastward, with continued northward subduction under the CAOS (e.g., Kazakhstan and other microcontinents) (Fig. 18). This led to the progressive destruction of the Turkestan, Asiatic, and Solonker Oceans, all parts of the Paleo-Asian Ocean, that are currently demarcated by the Denisov-Oktyabrsk, Turkestan, and Solonker-Jilin-Yanji suture zones (Fig. 1). Collision and ocean closure were completed by the Triassic.

Northward subduction of the Paleo-Tethys under the southern margin of Balkatach continued in the Permian, as expressed by the Permian–Triassic Kunlun-Yidun-Earlangping-Qinling magmatic arcs, which accommodated the

convergence of the South China craton and numerous other continents (e.g., Qiangtang and Lhasa). Collision between the South China craton and Eastern Balkatach (i.e., North China craton) began in the latest Permian, was concluded by the Triassic, and is demarcated by the Qinling–Dabie suture. The Paleo-Tethys was concurrently subducting to the south under the Qiangtang-Indochina terranes, which collided with Balkatach and the South China craton in the Late Triassic and led to the cessation of arc magmatism in the Kunlun arc (see Wu et al., 2016, for a discussion). By the Late Triassic, the Lhasa block to the south collided with the Qiangtang terrane along the Bangong–Nujiang suture following the closure of the Meso-Tethys (Yin et al., 1994; Murphy et al., 1997). The middle Cretaceous initiation of northward subduction of the Neo-Tethys under Lhasa accommodated convergence of India toward Asia, which led to the formation of an Andean arc and the development of the Gangdese batholith (Allégre et al., 1984; Harrison et al., 1992). The Indus-Yarlung suture zone separates Lhasa from the Himalaya and marks the destruction of the Neo-Tethys Ocean.

Intracontinental deformation that resulted from the collision between India-Arabia and Asia at 65–55 Ma (e.g., Le Fort, 1996; Yin and Harrison, 2000; Zhu et al., 2005; Green et al., 2008; Dupont-Nivet et al., 2010; Najman et al., 2010; Wang et al., 2011a; Hu et al., 2015, 2016) modified the existing configuration of Asia. In addition to the Himalayan orogen, crustal shortening occurred throughout the Tibetan Plateau and Central Asia in the Tian Shan, West Kunlun Shan, Qimen Tagh, North Qaidam, and Qilian Shan–Nan Shan thrust belts. Continental subduction in the Pamirs and strike-slip offset along the Altyn Tagh fault obscured many of the pre-Cenozoic structures. Eastward extension of Asia along the western margin of the Pacific Ocean also continued at that time.

DISCUSSION

This reconstruction conforms to existing geological observations, but in doing so, it raises the following questions.

1. What continents were affixed to the margins of Balkatach in the Proterozoic? This long cratonal strip, with Archean and Proterozoic structures truncated by Neoproterozoic passive margin strata, must have fit into a larger continental assemblage prior to that time (cf. Li et al., 2008).

2. Where does the Precambrian Tarim arc (i.e., 1.0–0.9 Ga granitoids and gneiss) and suture extend to the north or south of the Tarim continent (Fig. 15)? It is not likely that this arc could have laterally terminated within Tarim, and thus there should be evidence of its continuation on another continent that was rifted away from Balkatach in the Neoproterozoic.

3. Although this reconstruction is focused specifically on the tectonic evolution of central Asia, the Baltica craton's proposed connection within Balkatach has implications for the development of both Rodinia and Laurasia. In most reconstructions, Baltica has collided with Laurentia twice since the Neoproterozoic (e.g., Ziegler, 1989; Scotese and McKerrow, 1990). The proposed Baltica-Balkatach connection represents a fundamental departure from the model of Şengör et al. (1993), wherein pre-Uralide and Baykalide subduction systems

(i.e., located along the Urals and southern Siberia) were connected at the end of the Neoproterozoic. Instead, the northern margin of Balkatach may have linked with the North American Cordilleran arc-trench system (e.g., Zonenshain et al., 1987; Miller et al., 2011).

Tectonic Calving and CAOS Microcontinent Formation

A central component of our proposed reconstruction is that the microcontinents in the CAOS and the Qaidam-Kunlun continent are both genetically linked with the margins of Balkatach. This implies that these continents detached from Balkatach during Neoproterozoic rifting when the supercontinent Rodinia was rifting apart. This detachment was either partial, such as what we inferred for the development of the peninsular Qaidam-Kunlun continent, or complete, as for the CAOS microcontinents. However, in both cases, the continents remain near to the Balkatach continent following rifting and collide back with it in the Paleozoic. These rifted microcontinents may be analogous to the continental fragments that rifted from northern Australia in the Mesozoic–Cenozoic (e.g., Hill and Hall, 2003; Hall, 2011; Metcalfe, 2011). We envision that these rifted microcontinents are the result of tectonic calving from continental rifting to drifting, akin to glacial calving.

An issue with microcontinent development during regional rifting and final drifting is that extension should be isolated to the developing mid-ocean ridge, which has a relatively low yield strength compared to the continental crust (Bodine et al., 1981). A rheological weakness must develop to concentrate rifting away from a nearby mid-ocean ridge and allow a microcontinent to separate from the rest of the continental lithosphere. Increasing the temperature of the lithosphere may be one mechanism to generate this weakness (Müller et al., 2001). For Balkatach, especially around Tarim, there are three lines of evidence that suggest that there was a warmed lithosphere in the Neoproterozoic: (1) subduction and arc magmatism from ca. 970 to 910 Ma, collision by ca. 870 Ma along the Tarim suture, and protracted intracontinental deformation until ca. 800 Ma would have led to elevated heat flow; (2) long-lived magmatism (i.e., from 840 to 680 Ma) (e.g., Shu et al., 2011) predated, accompanied, and followed the main stages of rifting; and (3) there is evidence for a possible Neoproterozoic plume beneath the Tarim craton (e.g., Li et al., 2003; Lu et al., 2008; Long et al., 2011).

Continental break-up involving a hotter lithosphere is often associated with the development of a volcanic passive margin or volcanic rift margins (Gernigon et al., 2004; Geoffroy, 2005). If distinct segments of Balkatach's rifted margins involved volcanic-passive margin development, the associated large volume of warmer, weaker continental crust would allow extension and mid-ocean spreading to jump among zones of weakness. During the rifting of several large continents, smaller continental fragments could also rift away but they would remain nearby as local spreading centers shut off and spreading eventually concentrated within a single and central mid-ocean ridge. Proterozoic volcanic passive margins have not been well identified or studied, and

further geophysical examination would be needed to verify the presence of characteristic thick igneous layers in the lower crust or continentward-dipping normal faults (e.g., Gernigon et al., 2004).

It is important to note that while many of the CAOS microcontinents have confirmed Paleoproterozoic and older basement, especially in the western domains (i.e., the Mongolia continents and Kazakhstan), the eastern Chinese microcontinents have limited to no pre-Mesoproterozoic basement (e.g., Wilde, 2015; Zhou et al., 2017). Thus, if the CAOS and subduction within the Paleo-Asian Ocean commenced in the Mesoproterozoic (Kröner et al., 2013), it follows that some of the Chinese microcontinents were generated during intra-Paleo-Asian Ocean arc magmatism, rather than being derived from a Paleoproterozoic Balkatach continent source. There has been one reported Archean granitic gneiss from the Erguna microcontinent (Fig. 2; Shao et al., 2015), and more geochronology is needed to assess whether the Chinese microcontinents were derived from the Balkatach continent or generated within the CAOS.

Neoproterozoic Balkatach in Rodinia

Balkatach should be considered in the context of the supercontinents Columbia–Nuna and Rodinia. For example, in the popular Rodinia reconstruction of Li et al. (2008), the North China, Tarim, and Baltica cratons are positioned separately along the outskirts of the supercontinent, even though Proterozoic structures truncated by rift features along their margins (e.g., aulacogens and prominent transitions to passive margin sedimentation) require their involvement in a larger continental assemblage. For reconstructions of Columbia–Nuna, the North China craton is often connected with India on the basis of linking the Trans-North China orogen with the Central Indian tectonic zone (e.g., Zhao et al., 2002, 2004; Turner et al., 2014; Wan et al., 2015). This connection, although nonunique, may be valid, but the proposed link between the India and North China cratons should be pursued further, considering the much larger dimensions of Eastern Balkatach in the Proterozoic (i.e., a linked North Tarim and North China continent, with the genetically related CAOS microcontinents; Figs. 14–15). The long, relatively low aspect ratio of Balkatach readily allows for other continents to be affixed to either side of Balkatach (e.g., the continent's northern or southern margins, in present-day coordinates). We envision that the Balkatach continent may serve as a thin and previously unrecognized alternative missing link (e.g., Li et al., 1995) between already postulated supercontinent connections, as discussed below.

The well-documented truncation of Archean and Paleoproterozoic structures in western Laurentia (e.g., Taltson–Buffalo Head, Vulcan, and Great Falls tectonic zones) by a thick Neoproterozoic succession of miogeoclinal sediments led early workers to develop the hypothesis that a Precambrian continent had to have rifted away (e.g., Stewart, 1972, 1976; Burchfiel and Davis, 1972; Monger et al., 1972; Sears and Price, 1978). A similar situation exists for Balkatach, yet paleogeographic reconstructions of Balkatach's constituent cratons, especially North China and Tarim, do not consider this issue.

One possibility is that the proposed Balkatach continent was continuous with the western margin of Laurentia in the Proterozoic (Zuza and Yin, 2013, 2014), with Siberia positioned to the north of Laurentia (following Rainbird et al., 1998; see also Evans and Mitchell, 2011; Ernst et al., 2016). The ~6000 km length of Balkatach is equivalent to the north-south length of western Laurentia. This would position Tarim against central-western Laurentia, as recently suggested by Wen et al. (2016) based on paleomagnetic data. The following observations tentatively support this hypothesis.

Paleoproterozoic Subduction and Collision

The ca. 1.85 Ga Trans-North China orogen and Great Falls tectonic zone of western Laurentia (e.g., Gorman et al., 2002; Ross and Eaton, 2002; Mueller et al., 2002; Foster et al., 2006; Vervoort et al., 2016) have similar arc-subduction-collision histories that accommodated the convergence of Archean blocks. Archean crust in the Tarim craton is correlative to those found in western Laurentia (Foster et al., 2006; Long et al., 2010; Ge et al., 2014). The exposed Paleoproterozoic basement in Karakum and Turan would link with the 1.9–1.85 Ga Fort Simpson belt (Ross and Eaton, 2002) in western Canada.

Mesoproterozoic Strata

The Belt-Purcell Supergroup spatially and temporally correlates with Jixian (early Mesoproterozoic; Table 1) deposits in the North China craton (Hebei Bureau of Geology and Mineral Resources, 1989; Ross et al., 1992; Ross and Villeneuve, 2003; Zhai et al., 2015); both groups are thick (i.e., >10 km), and consist of carbonate and mudrocks deposited in active margin-bounded rift troughs. In addition, the Belt-Purcell rocks contain 1610–1500 Ma zircons that are rare in western Laurentia, and it has long been postulated that a separate continent must provide a western source for these zircon grains (Ross et al., 1992; Ross and Villeneuve, 2003; Jones et al., 2015). North China has rapakivi granites with this age signature (e.g., Zhang et al., 2007b).

Neoproterozoic Rift Histories

Rifting and the development of a Neoproterozoic–Cambrian passive margin sequences occurred along the northern margin of Balkatach and the western margin of Laurentia (e.g., Lund et al., 2010; Levashova et al., 2010, 2011; Meert et al., 2011; Shu et al., 2011; Han et al., 2011; Balgord et al., 2013). In addition, diamictites of similar ages (i.e., ca. 710 Ma, ca. 655 Ma, and ca. 630 Ma) have been reported in the Windermere Group of North America (Lund et al., 2003; Balgord et al., 2013), Qurutagh Group of Tarim (Xu et al., 2005; Shu et al., 2011), and Central Asian microcontinents (Levashova et al., 2011; Meert et al., 2011).

■ CONCLUSIONS

By removing the tectonic distortion effects in Central Asia caused by intra-continental deformation and rifting events, we have shown that a continuous continent once extended from the North China craton to Baltica in the west in the Neoproterozoic. We refer to this continent as Balkatach based on the linkage between the Baltica, Karakum, Tarim, and North China cratons. This continent, and the relative motion of its western and eastern arms (in present-day coordinates) in the Paleozoic, played an important role in the tectonic evolution of Asia. Neoproterozoic rifting along Balkatach's margins led to the opening of the Paleo-Asian, Pacific, and Tethyan Oceans. Archipelago development and subduction within the Paleo-Asian Ocean accommodated the oroclinal bending of Balkatach around this ocean. The initial collision of central Balkatach and CAOS in the mid-Carboniferous was followed by bidirectional suturing and closure of the Paleo-Asian Ocean by the Permian. The closure of the Paleo-Tethys Ocean along Balkatach's southern margin proceeded diachronously from west to east during the Permian–Triassic.

The restored ~6000-km-long Balkatach continent must fit in Neoproterozoic Earth and Rodinia reconstructions. The tectonic reconstruction presented here is at odds with current Neoproterozoic models that place each of Balkatach's constituent continents separately along the outskirts of the Rodinia supercontinent (e.g., Li et al., 2008). We tentatively propose that Balkatach was affixed to the western margin of Laurentia in the Proterozoic, with Siberia positioned to the north of Laurentia (Rainbird et al., 1998). More research is needed to explore this hypothesis, and our work presents geologic predictions that can be tested by future investigations.

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